Modeling the effects of throughfall reduction on soil water content in a Brazilian Oxisol under a moist tropical forest

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Received 11 September 2006; revised 19 April 2007; accepted 24 May 2007; published XX Month 2007.

[1] Access to water reserves in deep soil during drought periods determines whether or not the tropical moist forests of Amazonia will be buffered from the deleterious effects of water deficits. Changing climatic conditions are predicted to increase periods of drought in Amazonian forests and may lead to increased tree mortality, changes in forest composition, or greater susceptibility to fire. A throughfall reduction experiment has been established in the Tapajós National Forest of east-central Amazonia (Brazil) to test the potential effects of severe water stress during prolonged droughts. Using time domain reflectometry observations of water contents from this experiment, we have developed a dynamic, one-dimensional, vertical flow model to enhance our understanding of hydrologic processes within these tall-stature forests on well-drained, upland, deep Oxisols and to simulate changes in the distribution of soil water. Simulations using 960 days of data accurately captured mild soil water depletion near the surface after the first treatment year and decreasing soil moisture at depth during the second treatment year. The model is sensitive to the water retention and unsaturated flow equation parameters, specifically the van Genuchten parameters θ_s, θ_r, and n, but less sensitive to K_s and α. The low root-mean-square error between observed and predicted volumetric soil water content suggests that this vertical flow model captures the most important hydrologic processes in the upper landscape position of this study site. The model indicates that present rates of evapotranspiration within the exclusion plot have been sustained at the expense of soil water storage.

Citation: Belk, E. L., D. Markewitz, T. C. Rasmussen, E. J. M. Carvalho, D. C. Nepstad, and E. A. Davidson (2007), Modeling the effects of throughfall reduction on soil water content in a Brazilian Oxisol under a moist tropical forest, *Water Resour. Res.*, *43*, XXXXXX, doi:10.1029/2006WR005493.

1. Introduction

[2] Tropical rain forests have a disproportionate importance in the global exchange of carbon, water, and energy between the biosphere and atmosphere [*Schlesinger*, 1997]. While the function of the Amazon river basin in the global water cycle is well recognized, we are only beginning to understand the interaction of factors affecting the below-ground partitioning and availability of water and nutrients to the vegetation in its forest ecosystems. These processes are important for interpreting how humid tropical forests manage to maintain evergreen canopies during the annual dry season and for predicting how these forests might respond to prolonged periods of drought, such as those that result from El Niño–Southern Oscillation (ENSO) events [*Nepstad et al.*, 2004; *Oliveira et al.*, 2005].

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0043-1397/07/2006WR005493$09.00

[3] To study the response of a humid Amazonian forest to severe drought, a partial throughfall exclusion study was initiated in 1998 in the Tapajós National Forest, east-central Amazonia, near Santarém, Brazil [*Nepstad et al.*, 2002]. This experiment compares two 1-ha plots, one of which receives natural rainfall, while the other has plastic panels installed in the forest understory during the rainy season. These panels capture approximately 60 percent of incoming throughfall, channelling the water to a system of gutters and diverting it from the soil. Both the control and exclusion plots are surrounded by a 1.0–1.7 m deep trench, which reduces the ability of trees within the plots to access water from outside the plots [*Sternberg et al.*, 2002].

[4] A variety of processes are being monitored, including: tree growth and mortality, sap flow, litterfall, leaf area index, forest floor decomposition, soil respiration, trace gas emissions, forest floor flammability, and the amounts and chemistry of precipitation, throughfall, litter leachate, and soil solutions. Soil moisture content is also measured by time domain reflectometry using soil shafts that allow access to 12 m depth in both the exclusion and control plots. Soil moisture measurements alone, however, do not describe the magnitudes and rates of water fluxes because two layers may contain the same water volume within a

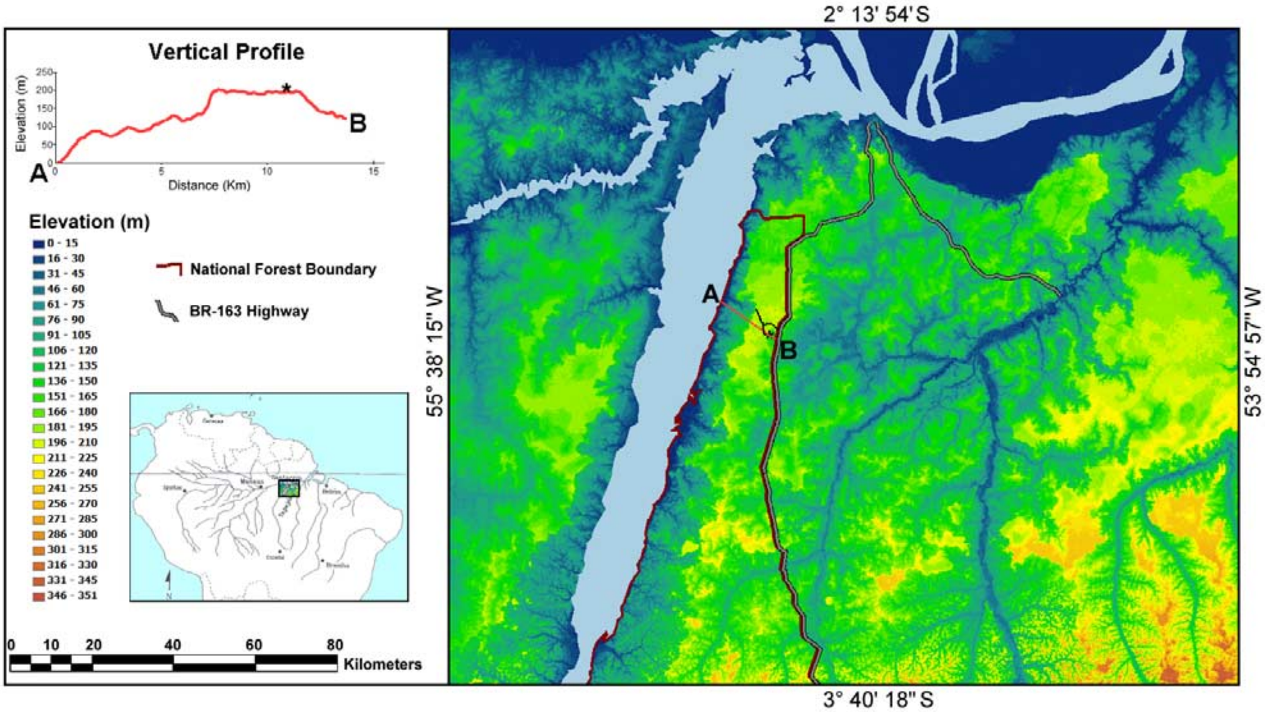

Figure 1. Site location for the throughfall exclusion experiment in the Tapajós National Forest, Brazil. Star in the top left plot indicates the research site location.

given soil volume, but have different rates of fluid movement through them. This means that model estimations of water fluxes are required in order to fully quantify the hydrologic system.

[5] The objective of this component of the throughfall reduction study is to develop an understanding of the physical processes driving the observed soil water dynamics at the site. We will make use of a vertically integrated version of the Richard's mass balance equation to evaluate the sensitivity of various parameters and to compare the hydrologic mass balance of the control and dry-down plots. Hydrologic flux estimates from this model might also be utilized in the future to estimate the advective movement of dissolved chemical components through the soil.

[6] Knowledge of the changes in below-ground storage and partitioning of water enhances our ability to explain other responses of the forest to drought conditions. By quantifying how the ecological functions of tropical forests change during prolonged drought, we hope to better understand the changes that may occur during the annual dry season in functions such as rooting depth or leaf shedding and better predict the ability of these forests to tolerate reductions in precipitation associated with land use conversion as well as long-term climate changes.

2. Tapajós Research Site

[7] The forest being modeled is located in a protected area of Floresta Nacional Tapajós, a Brazilian national forest located in east-central Amazonia, south of the city of Santarém do Pará (2.89°S, 54.95°W), shown in Figure 1. The site is located approximately 150 m above and 13 km east of the Tapajós River [*Nepstad et al.*, 2002]. The study plots are situated on a relatively level, upper landscape plateau position where the soils are predominantly Haplustox (Latasolos vermelhos) dominated by kaolinite clays, and support a terra firme forest, which is a dense, humid, evergreen forest that does not flood annually. The forest at the field site has a continuous canopy that is approximately 30 m tall.

[8] The throughfall reduction experiment was initiated in 1998. After a 1-year pretreatment period, plastic panels were installed at the beginning of the 2000 rainy season that extends from January to May. Panels are removed during the dry season and reinstalled prior to the rainy season of the following year.

2.1. Soil Moisture

[9] Volumetric water contents ($m_w^3\ m_s^{-3}$) were measured using time domain reflectometry (TDR) [*Topp et al.*, 1980] sensors installed to 11-m depth in six soil shafts (two plots; three shafts per plot; yielding six sensors per depth for both plots). Each soil shaft measures 1 m by 2 m in width, and extends to a depth of 12 m. Access is obtained using a system of wooden beams and supports.

[10] TDR sensors consist of three, parallel, 24-cm stainless steel rods [*Zegelin et al.*, 1989] and were measured with a cable tester (Textronix 1502C, Beaverton, Oregon). Two TDR sensors were installed horizontally in opposing walls at 1-m increments in each soil shaft. Each of the six shafts also has two probes installed vertically from 0 to 0.3 m, and two probes installed horizontally at 0.5 m. Because the shafts were left open to maintain access for root and nutrient studies, sensors were installed into undisturbed soil 1.5 m from the shaft walls. Auger holes were back filled with native soil. This installation method was based on previous work in Oxisols in Paragominas, Pará [*Davidson and Trumbore*, 1995].

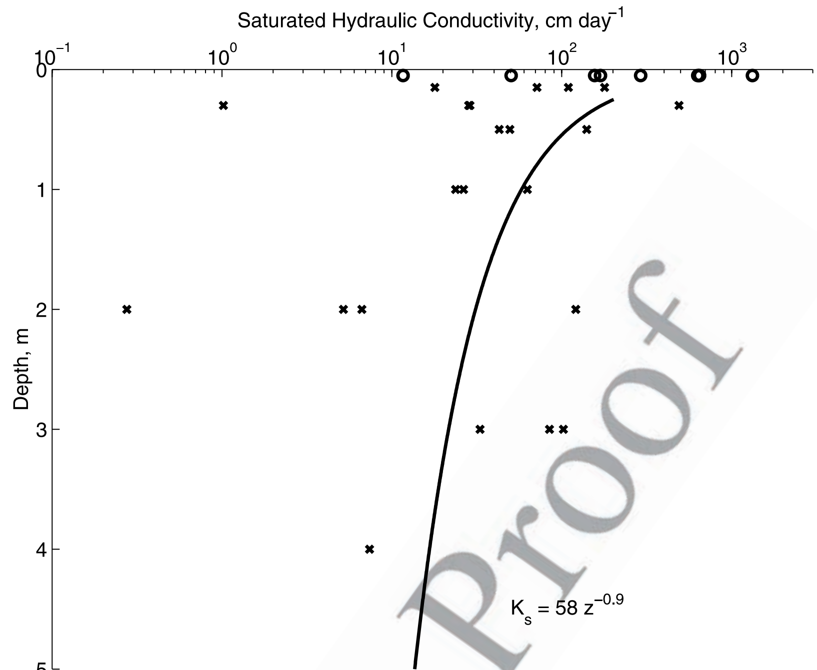

Figure 2. Surface (circles) and subsurface (crosses) measurements of saturated hydraulic conductivity, K_s, in the Tapajós National Forest, Brazil. The K_s values between 0 and 4 m were measured with a Guelph permeameter. Extrapolation to greater depths uses a power function fit.

[11] Waveforms from the TDR sensors were collected approximately once per month. Water contents were estimated following the methods of *Topp et al.* [1980]. The Topp equation has been validated for mineral soils in both surface and deep Oxisols in the Amazon by *Jipp et al.* [1998]. The Belterra clay soil used in the validation study are the same as the soils studied here, and have similar physical characteristics.

2.2. Saturated Hydraulic Conductivity

[12] Saturated hydraulic conductivity (K_s) was quantified using a Guelph permeameter [it *SoilMoisture Equipment Corporation*, 1986]. Seven surface measurements were made in random locations around the plots using a pressure infiltrometer attachment. Figure 2 presents K_s results from the surface to 4 m. Below-ground observations were obtained by augering 6-cm-diameter vertical holes. Three sets of measurements were completed at each of three sites in the study area. The holes were gently brushed before measurements to remove any smearing of the clays that may have occurred during augering.

[13] K_s results from the surface and at 30 cm are presented in Figure 2. Note the large variation in observations, which is consistent with other sites [*Rasmussen et al.*, 1993]. Data were arithmetically averaged at each depth and assigned to the closest layer midpoint. Because our model extends to greater depths, estimates of deeper values are required. We are not aware of any studies that have measured K_s to 11 m depth. It is likely, however, that K_s decreases with depth, because K_s is highly affected by macroporosity and these deep soils become less structured with depth in this region. This hypothesis is supported by the resulting fit of a power function to the observed data (also shown on Figure 2), which indicates a decrease with depth. Point estimates of K_s were extrapolated using this power function for soil layers between 4 and 11 m.

2.3. Rainfall and Throughfall

[14] Rainfall was estimated from three, prism-shaped gauges located in and near the study site. One gauge was installed within each plot on the top of a 28-m tower within a small canopy opening. One additional gauge was located at ground level in an opening approximately 400 m from the plots. Rainfall was monitored daily, except over the weekend; Monday readings include rain that fell over the weekend.

[15] Throughfall samples were collected in 0.16-m-diameter funnels that lead to plastic collection bottles. Each plot has ten throughfall collectors under the canopy. Bottles were at ground level during the pretreatment year. In the following years all bottles were raised approximately 2 m above ground level so that exclusion panels did not interfere with throughfall collection. Sample volumes were measured every two weeks. The ten collectors in each plot were randomly reassigned a location within the sampling grid for that plot after each sampling.

2.4. Fine-Root Biomass

[16] Fine root biomass data ($kg_r\ m_s^{-2}$) were estimated from 24 borings divided into eight depths in each plot (384 samples). Each sample was washed and sorted into live and dead fractions, and then sorted into two size classes (<1 mm and 1–2 mm). The depths at which samples were collected were: 0, 0.5, 1, 2, 3, 4, 5, and 6 m. The fraction of the total fine (live) root biomass (0–2 mm) in each layer was used to estimate a rooting factor, $R(z)$, for each modeled soil layer. The root biomass was considered to be 10 percent

less than the horizon above for estimating root factors below 6 m.

2.5. Soil Moisture Parameters

[17] Soil water retention data were generated by the EMBRAPA-CPATU laboratory in Belém, Pará, Brazil. A standard pressure plate method was used whereby intact soil cores (n = four per depth) were saturated and the water extracted by the application of a steady, constant pressure [*Klute and Dirksen*, 1986].

[18] These data were fit to van Genuchten soil moisture characteristic (SMC) functions using nonlinear regression [*Wraith et al.*, 1993]. The starting values for the nonlinear regressions were the average van Genuchten parameters (α, θ_s, θ_r, and n) reported by *Hodnett and Tomasella* [2002] for tropical clay soils.

2.6. PET and Other Meteorological Data

[19] Potential evapotranspiration (PET) was calculated using the Thornthwaite method [*Thornthwaite and Mather*, 1957]. On-site temperature data were available for estimation with this method and thus avoided additional parameterization (e.g., stomatal conductance) that would have been required with methods such as Penman-Monteith [*Monteith*, 1965]. An eddy flux tower was established in close proximity to the experimental site in 2000 but direct estimates of AET are only available after 2002 [*Hutyra et al.*, 2005]. Data from this eddy flux tower, however, did demonstrate a strong correlation between AET and PET estimated with the Thorntwaite method [*Hutyra et al.*, 2005]. For the current model, temperature inputs utilized were monthly averages of daily daytime air temperatures collected at the canopy level of the control plot with recording Hobo data loggers (Onset Computer Corp., Bourne, Massachusetts). An additional correction was applied to this estimate to adjust for the tendency of the Thornthwaite model to overestimate PET when average air temperature is greater than 26.5°C [*de Amorim et al.*, 1999]. This correction is based on an empiral fit and has the following form:

$$PET \geq 26 = PET*\left(1 - e^{e-0.28*(t-31.1)}\right) \tag{1}$$

where t is the mean monthly temperature. After all corrections, the monthly estimates of PET were divided into equal daily values to be consistent with the time step of the model.

3. Soil Water Model

3.1. Model Structure

[20] The model was designed to simulate daily changes in the distribution of soil water. Vertical water movement through 13 soil layers is driven by the difference in total soil hydraulic head, which integrates the effect of matric and gravitational forces. Plant uptake of water to the forest vegetation is included. Simulations were performed for the control plot with no reduction in water inputs and for the treatment plot using throughfall exclusion during the rainy season.

[21] The model used a daily time step, but changes in VWC were integrated using the Euler method on an hourly basis. The Euler method estimates changes in stocks using the computed flow values. Given larger time steps (i.e., 1 day) this algorithm is preferred. Calibration was performed using soil volumetric water content measured in the control plot on an approximately monthly interval during the first 960 days of the experiment. When the throughfall exclusion treatment switch is selected the model predicts soil volumetric water content for the same time period as the forest undergoes partial throughfall exclusion without any additional calibration of the model.

[22] The temporal (Δt) and vertical (Δz) discretization of this model were chosen to be consistent with the scale of the data available for validation (i.e., monthly TDR data for soil layers of 50 to 100 cm). Finer-scale discretization (i.e., $\Delta t \leq$ 1 day and $\Delta z \leq$ 5 cm), however, is often preferred for applications of the Richard's equation particularly with regard to surface soil layers [*Lee and Abriola*, 1999]. To test the affect of these temporal and vertical discretizations HYDRUS 1D was utilized [*Šimuunek et al.*, 2005]. HYDRUS 1D was parameterized utilizing the same data described below although the 13 soil layers over the 11.5 m profile were discretized into 5 cm increments for model solution.

Table 1. Model Inputs[a]

Input	Description	Units
	Soil Water Model Parameters	
Rainfall	daily rainfall rate	mm d^{-1}
PET	daily potential evapotranspiration	mm d^{-1}
Throughfall	rainfall entering soil surface	fraction
Δz(z)	distance between layers	m
H(z)	total hydraulic head	m
D_w(z)	water depth in soil layer	m
K_s(z)	saturated hydraulic conductivity	m s^{-1}
R(z)	root length or biomass present	fraction
	van Genuchten Parameters	
θ_s(z)	saturated water content	$m_w^3\ m_s^{-3}$
θ_r(z)	residual water content	$m_w^3\ m_s^{-3}$
α(z)	water retention	m^{-1}
n(z)	water retention	-

[a]Parameters with (z) are input for each layer of soil.

3.2. Model Inputs

[23] Table 1 contains a list of the inputs required by the model. The depths separating each of the 13 soil layers are: 0, 0.4, 0.75, 1.5, 2.5, 3.5, 4.5, 5.5, 6.5, 7.5, 8.5, 9.5, 10.5, and 11.5 m. These increments were chosen so that the TDR measurements are near the midpoints of each layer. Other site-specific information, such as air temperature for PET estimates, is implicitly incorporated within the model.

3.3. Forest Water Inputs

[24] Rainfall enters the forest system and is partitioned between throughfall and canopy interception (Figure 3). Throughfall was empirically determined at the site to be 88 percent of incoming rainfall; the balance, 12 percent, is intercepted by the canopy. This empirical relationship did not vary by season and data were not available to test a relationship with rainfall intensity. Furthermore, coverage of the canopy, which is usually around 95 percent, did not change in either plot during the simulation period [*Nepstad et al.*, 2002]. When the treatment plot is simulated, 60 percent of the throughfall input is diverted from the soil when the panels are in place. This throughfall exclusion estimate

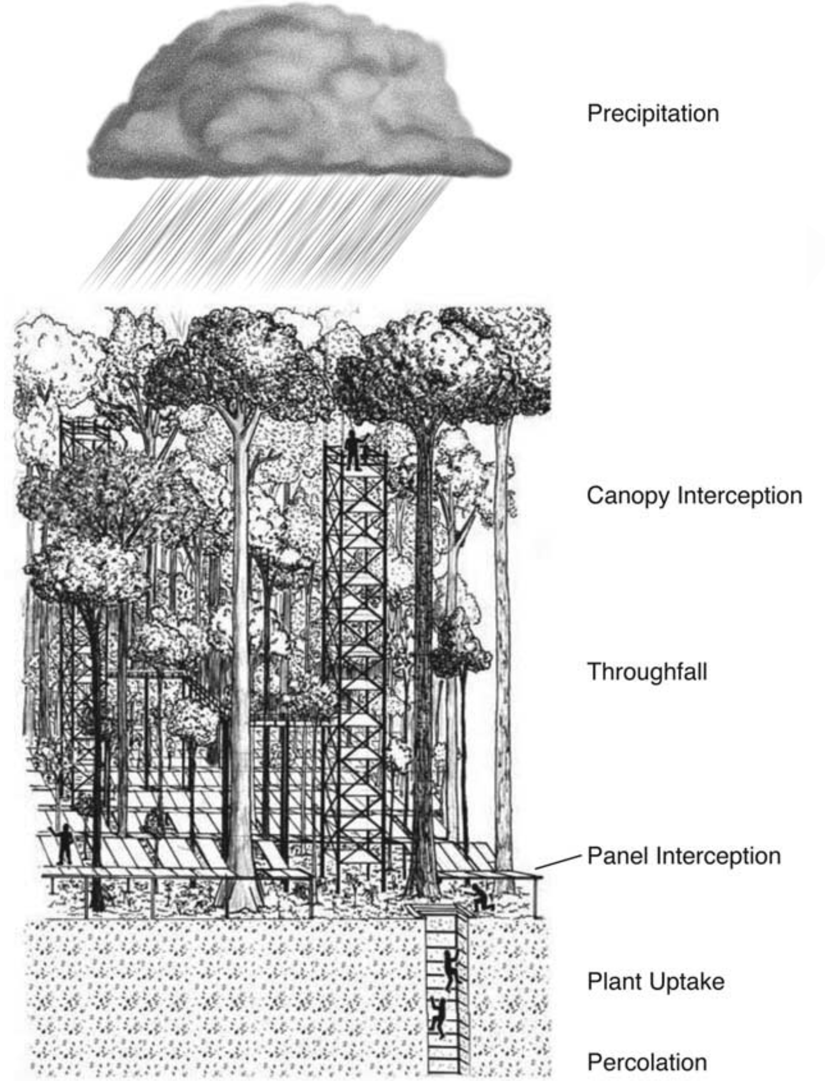

Figure 3. Idealized model structure for water cycling in a deep Oxisol. Precipitation, panel interception, and soil water contents were measured. Empirical functions were used for canopy interception and plant root uptake.

is based on measurements of water volumes collected in the gutters transporting water off the plot [*Nepstad et al.*, 2002].

3.4. Soil Water Movement

[25] Throughfall reaching the soil surface is allowed to infiltrate directly into the uppermost soil layer because the litter layer on the site is thin (approximately 2–4 cm) and the measured surface infiltration rates were high (>30 × 10^{-6} m s^{-1}). All thirteen layers hold a depth of water (D_w, m) equivalent to the soil moisture within that increment of soil. The water depth in each layer was initialized using soil water content data from May 17, 1999, the first day of simulation. The water content of each layer ($\theta(z) = D_w(z)/\Delta z$; $m_w^3\ m_s^{-3}$) is determined using the depth of water (D_w; m) and the soil thickness (Δz; m).

[26] Water flux between soil layers is determined using Darcy's law for one-dimensional (vertical), unsaturated flow [*Muller*, 1999]:

$$q_z = K(\theta)\frac{\Delta H}{\Delta z} \tag{2}$$

where q_z is the vertical water flux (m s^{-1}), $K(\theta)$ is the unsaturated hydraulic conductivity (m s^{-1}), ΔH is the difference in total hydraulic head between two adjoining layers (m) and Δz is the downward directed, vertical distance between the midpoints of the layers (m).

[27] The total hydraulic head of the soil water, $H(z) = h_m + h_z$, in a given layer is the sum of the matric (h_m) and gravitational (h_z) heads. The matric head of the soil water is determined by the van Genuchten equation relating water content to matric head [*van Genuchten*, 1980]:

$$h_m = \frac{1}{\alpha}\left[\Theta^{-1/m} - 1\right]^{1/n} \tag{3}$$

where $\Theta = (\theta - \theta_r)/(\theta_s - \theta_r)$ is the relative saturation of the soil ($m_w^3\ m_s^{-3}$), and where θ_s is the saturated water content, θ_r is the residual water content, and α (m^{-1}), n, and $m = 1 - 1/n$ are fitting parameters.

[28] The soil surface serves as the datum where gravitational head is zero. Unsaturated hydraulic conductivity,

$K(\theta)$, is calculated from saturated hydraulic conductivity, K_s, values according to the equation of *Mualem* [1976]:

$$K(\theta) = K_s \Theta^{1/2} \left[1 - \left[1 - \Theta^{n/(n-1)} \right]^m \right]^2 \tag{4}$$

[29] Changes in soil water storage are modeled using the Richard's (mass balance) equation that accounts for inflows and outflows in each layer:

$$\frac{\partial q_z(z)}{\partial z} \pm U(z) = \frac{\partial \theta(z)}{\partial t} \tag{5}$$

where $U(z)$ are internal sources or sinks within each layer. Root uptake (described below) is the only mechanism for internal water loss within each layer in our model.

3.5. Deep Drainage

[30] Deep drainage out of the lowest layer (m^3 m^{-2} s^{-1}) is calculated using Darcy's law and the assumption that saturated conditions persist at great depth, which is consistent with an observed water table depth of 100 m [*Nepstad et al.*, 2002]. Because the matric head is zero at the water table, i.e., $h_m = 0$, the total head must equal the gravitational head ($H = h_z = z$). This lower boundary condition may have some effect on the simulated drainage rate from the lowest layer, but has less of an influence on the water content of the profile overall.

3.6. Soil Evaporation and Plant Uptake

[31] The model assumes that there is no evaporation from the soil surface because only about 1 percent of solar radiation penetrates the forest canopy [*Nepstad et al.*, 2002]. Other researchers have reported that direct evaporation from the soil surface is negligible in Amazonian forests [*Jordan and Heuveldop*, 1981]. Water required for transpiration by vegetation is removed from each soil layer before downward percolation is allowed. It is assumed that when a vapor pressure deficit exists between the forest and surrounding atmosphere, water evaporates from vegetative surfaces more readily than it can be transpired through leaf stomata [*Ubarana*, 1996].

[32] Intercepted water in the canopy is first used to satisfy evapotranspirational demand, which is determined by the PET. Intercepted water is temporarily stored within the canopy and allowed to evaporate directly from it at a rate limited by the PET. If more water is intercepted than can be potentially evapotranspired, then no water is taken from the soil during that time step. When the PET is greater than the amount of water stored within the canopy, then water is removed from the soil in an amount equal to the difference. The fraction of this total uptake extracted from a given layer is

$$U(z) = U_{\max} R(z)\, URF(z) \tag{6}$$

where $U_{\max}$ is the maximum amount of water extracted from the soil (m), $R(z)$ is the proportion of fine root biomass in a given layer, and $URF(z)$ is an uptake reduction factor that restricts plant uptake on the basis of the matric head. URF does not vary with PET, and uptake near saturation is not restricted [*Feddes et al.*, 1978, 2001].

[33] Thornthwaite calculations were performed independent of, and prior to, model simulation and were then provided as a daily input for simulation. Because water content calculations were reported on an approximately monthly basis, the failure to account for intradaily PET variation is not expected to substantially affect model calculations.

3.7. Model Sensitivity and Performance

[34] Sensitivity analysis were performed on the saturated hydraulic conductivity (K_s) and *VG* parameters (i.e., α, θ_r, θ_s, m, and n). The parameter of interest was assigned at least five other values while the remaining parameters were left unchanged. The sensitivity of the model to these changes was quantified by evaluating their effect on the average depth of water stored in that layer, layers above or below, and/or the average depth of water in the entire profile. Model performance was evaluated using the mean difference, root-mean-square error (RMSE), relative root-mean-square-error (RRMSE), and the coefficient of determination (R^2) between measured and predicted volumetric water content.

4. Results and Discussion

4.1. Model Calibration

[35] We endeavored to use only input variables or constants that were determined by measurements made at the site for the initial parameterization of the model (Table 2, but see below). It became apparent during parameterization, however, that the model was unstable when there were large changes in the VG parameters between soil layers. These large differences between adjacent layers may allow one or more layers to wet or dry beyond reasonable ranges. The VG parameters fit to laboratory-generated water retention data for the site demonstrated this characteristic, largely in the upper layers, and thus the model was unstable.

[36] Inconsistencies in physical soil water characteristics between laboratory and field data are not uncommon [*Rasmussen et al.*, 1993]. One reason for the poor correspondence is the alteration of soil structure during sample collection, resulting in an increase in overall macroporosity. Another reason is an artifact of laboratory testing, in that soils are normally tested by drying the samples, yet soil moisture changes under field conditions include both wetting and drying conditions (i.e., hysteresis effects). Spatial variability of soil properties is another possible explanation. Finally, it is possible that the laboratory data is correct but that the numerical method utilized in the model was insufficient to adequately represent the true variation.

[37] VG parameters for each soil layer were calibrated iteratively using data from the control plot until RMSE between the measured and predicted volumetric water content for all depths over all dates was minimized (Figure 4). The resulting RMSE is 1.88 percent water content, which is a RRMSE of 5.1 percent.

[38] The soil moisture characteristic (SMC) curves that result from the optimized VG parameters are displaced below the laboratory data (Figure 5). In other words, calibrated water contents are drier than the laboratory values when compared at the same matric suction. In all cases, laboratory data have the lowest average range of water content between the saturated, θ_s, and residual, θ_r, water content (0.216 m$_w^3$ m$_s^{-3}$). The laboratory SMC curve has the

Table 2. Parameter Values and Initial Values for Stocks Input for Each Model Layer[a]

		Initial D_w			van Genuchten Parameters				
Layer	Depth, m	Control, m	Treatment, m	K_s, μm s^{-1}	θ_s, $m_w^3\,m_s^{-3}$	θ_r, $m_w^3\,m_s^{-3}$	α, m^{-1}	n	R
1	0–0.4	0.155	0.125	31.69	0.44	0.24	0.040	1.2	0.688
2	0.4–0.75	0.106	0.109	8.98	0.41	0.23	0.040	1.5	0.103
3	0.75–1.5	0.247	0.225	4.35	0.40	0.21	0.040	1.5	0.050
4	1.5–2.5	0.339	0.307	3.85	0.35	0.20	0.055	1.3	0.030
5	2.5–3.5	0.370	0.325	8.50	0.46	0.22	0.050	1.5	0.022
6	3.5–4.5	0.378	0.354	0.86	0.44	0.23	0.055	1.6	0.019
7	4.5–5.5	0.415	0.374	1.64	0.47	0.23	0.050	1.4	0.019
8	5.5–6.5	0.423	0.394	1.40	0.49	0.23	0.045	1.5	0.017
9	6.5–7.5	0.433	0.399	1.23	0.49	0.22	0.045	1.4	0.015
10	7.5–8.5	0.418	0.373	1.10	0.49	0.19	0.040	1.4	0.014
11	8.5–9.5	0.414	0.386	0.98	0.47	0.21	0.045	1.4	0.012
12	10.5–11.5	0.410	0.392	0.90	0.47	0.21	0.045	1.4	0.011
13	11.5–12.5	0.406	0.413	0.82	0.46	0.20	0.040	1.4	0.010

[a]Initial throughfall fraction is 0.88. Saturated hydraulic conductivity (K_s) was measured in the field for layers 1–6 but were extrapolated below layer 6 using the power function shown in Figure 1. Van Genuchten parameters (θ_s, θ_r, α, and n) obtained from model calibration. Fraction of total, fine (0–2 mm) live root biomass (R) were measured in layers 1–8 but were extrapolated below layer 8 by assuming a reduction of 10 percent in each subsequent layer.

highest water content, primarily due to a smaller average n value. This higher laboratory SMC curve for the surface soil yields a larger θ_r value than optimized values. In all cases, the optimized θ_s values are lower than the porosities measured in the laboratory. Values for α are also moderately higher, which reflects the presence of pores that empty with small changes in matric head.

[39] For comparison, Figure 5 also presents an average SMC curve for tropical soils with clay textures, as well as an average SMC curve for Ferralsols [*Hodnett and Tomasella*, 2002]. Both soils contain kaolinite clays which do not swell and tend to have higher α values because they drain from saturation quickly [*Hodnett and Tomasella*, 2002]. Note that the optimized SMC curve resembles the average for tropical Ferralsols. The soils at the study site being modeled are classified as Latosols in the Brazilian taxonomy, which is similar to the FAO definition of a Ferralsol [*Richter and Babbar*, 1991]. The most notable difference between the parameters is that the average range of water content ($\theta_s - \theta_r$) for the optimized parameters is much lower, 0.234 $m_w^3\,m_s^{-3}$ compared to an average of 0.322 $m_w^3\,m_s^{-3}$ for the Ferralsols.

[40] Regardless of these discrepancies, using the difference between the water contents at 30 and 1500 kPa to represent the maximum plant-available water (PAW), it is clear that all SMC curves contain 6.2–6.9 percent PAW.

[41] For the calibrated simulation, the top two layers have poorer fits than the others, with RRMSEs of 9.8 percent or greater (Figure 6). Except for the third layer, which has an RRMSE of 5.4 percent, the errors in the other horizons are all below 4.6 percent. The poorer fit in the top horizons, did not result simply from the coarse vertical discretization of the model as evidenced by comparison to the 5-cm discretization of the HYDRUS 1D model (Figure 7). Simulations from both models demonstrate similar seasonal patterns and

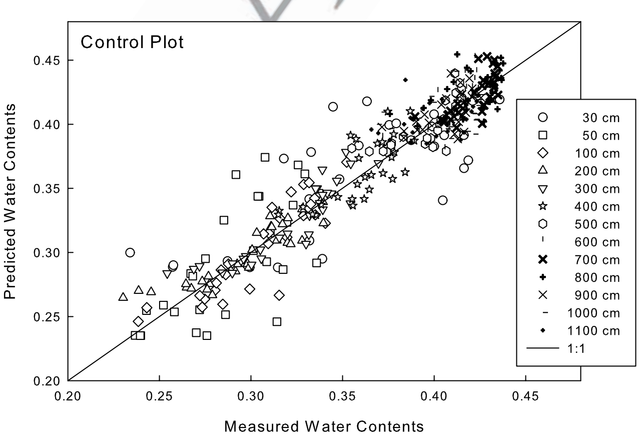

Figure 4. Scatterplot of measured and predicted volumetric water contents (θ) in the control plot, Tapajós National Forest, Brazil. The comparison is for 13 depths and 29 dates between May 1999 and December 2001 on which water content (θ) was measured at the field site.

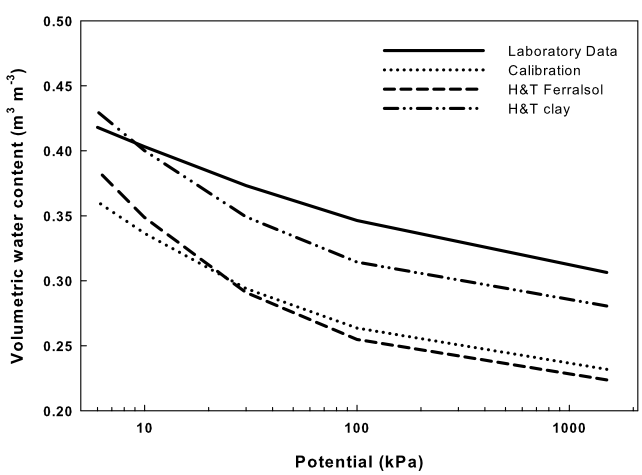

Figure 5. Soil moisture characteristic curves described by the van Genuchten parameters fit to laboratory pressure plate data and by the parameters resulting from model calibration. For comparison, the curves described by the average parameters for tropical soils with clay textures and for tropical soils in the Ferralsol soil group as reported by *Hodnett and Tomasella* [2002] are shown. All curves are for upper surface soils (<10 cm).

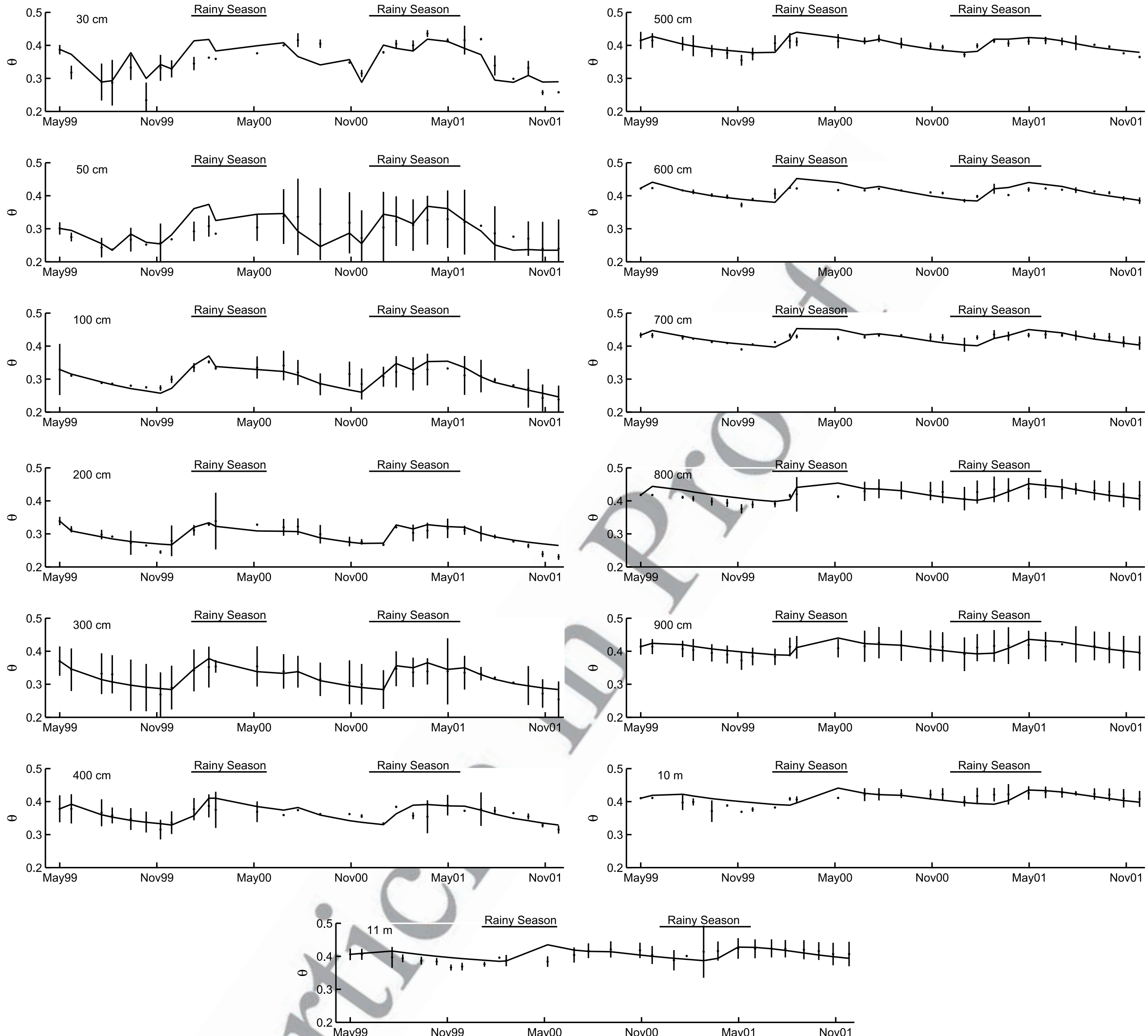

Figure 6. Monthly measured (dots) versus predicted (lines) volumetric water contents (θ) in the control plot, Tapajós National Forest, Brazil. Measured θ are averages of six TDR sensors per depth. Also shown are vertical bars representing measurement standard deviations, where available.

both tend to underestimate the wettest measurements while over estimating the driest measurements. Surface soils clearly undergo a great deal of variation in VWC and both model discretizations struggle to capture this variance where the soil moisture conditions are more dynamic. Limited discretization of inputs to the model (e.g., daily rainfall or daily average PET) may also limit the ability to capture surface soil dynamics. In the lower depths where VWC is more static both models preform well.

[42] The calibrated STELLA model does succeed in capturing important seasonal trends and shows the expected delay in recharge and depletion responses with increasing depth. The timing of these delays, however, are about a month or two slow in the model predictions. This slower response is consistent with observations in other moist systems where empirical estimates of the hydraulic velocities are greater than estimates based on SMC functions, [*Rasmussen et al.*, 2000].

4.2. Sensitivity Results

[43] We tested the sensitivity of the model to the input parameters using optimized parameters from the calibrated model (Table 2). Analyses show that the model is more sensitive to θ_s, θ_r, and n, but less sensitive to α and K_s (Figure 8). The sensitivity of the model to the VG parameters is not unexpected given that they are used in both the equation that determines matric heads and the equation for unsaturated hydraulic conductivities. For the parameters to which the model is most sensitive, however, the effect of a change in one layer is largely confined to that layer. For example, raising the θ_s in a layer from 0.40 to 0.60 (40–60 percent water content) increased the average soil moisture of that layer by 10.6–13.6 percent, but the average water

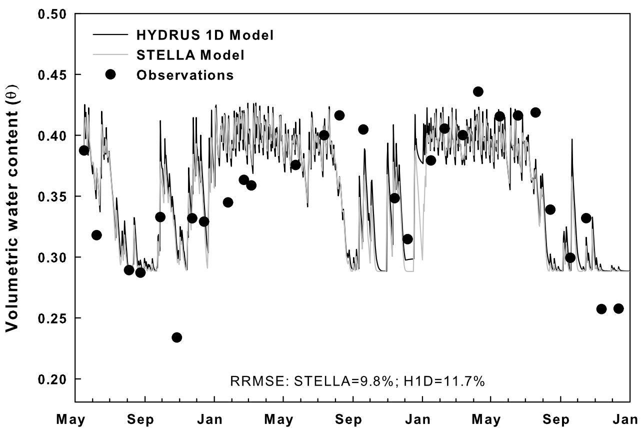

Figure 7. Comparison of monthly measured (dots) versus predicted (lines) volumetric water contents (θ) in the control plot using a coarse (one 40 cm layer in STELLA) or finer (eight 5 cm layers in HYDRUS 1D) model discretization. Measured θ are averages of six TDR sensors per depth in the Tapajós National Forest, Brazil.

content of the layer above or below generally decreased by 1 percent or less. The relative patterns in sensitivity remain the same for the average water content of the entire profile, but the increase is only 0.2–1.1 percent.

[44] Changing θ_r in a layer from 0.10 to 0.30 (10–30 percent water content) increased the average soil moisture of the layer by 3.8–8.5 percent. In contrast, a hundredfold increase in K_s resulted in only a 1–2 percent decrease in the water content of a layer. The water content is somewhat sensitive to K_s when the value is low because K_s represents the maximum flow rate. Thus the soil water content is affected whenever K_s is less than the water flux, but increases in K_s above the water flux have little effect on the water content.

[45] While the sensitivity of the model to individual changes in VG parameters may be important, it is also important to examine how the four parameters work together to define the water retention and unsaturated flow rates. A full factorial analysis of the interaction between θ_s, θ_r, and n for the 1.5–2.5 m layer confirms that the model is also sensitive to the difference between θ_s and θ_r. This difference is more important than absolute values because it indicates the range of water content expected in the soil (and the range for which the van Genuchten and Mualem models are valid). The difference between the average water content of the layer when θ_s is high (0.6 m^3 m^{-3}) and θ_r is low (0.1 m^3 m^{-3}) versus when θ_s is low (0.4 m^3 m^{-3}) and θ_r is high (0.3 m^3 m^{-3}) is about 2 percent, but when both parameters are low or high the difference in water content was 19.5 percent.

4.3. Treatment Plot Predictions

[46] Using the VG parameter values calibrated within the control plot, we simulated the soil water content in the throughfall exclusion plot over the 960 days of available data (Figure 9). The treatment plot shares similar throughfall inputs as the control plot during the first eight months of the simulation in May through December 1999. In 2000 and 2001, 895 and 817 mm of throughfall were excluded from the treatment plot by the model, values slightly greater than the 890 and 794 mm estimated empirically by *Nepstad et al.* [2002].

[47] Throughout this period of simulation the RMSE in soil moisture is 3.1 percent water content. This is a RRMSE of 9.2 percent. The mean difference is -0.65 ± 0.16 percent water content. Any loss of soil contact with the TDR sensors, due either to compaction during installation or to later soil drying, could cause low soil moisture reading [*Baker and Lascano*, 1989; *Knight*, 1992, 1994].

[48] Overall, the treatment plot simulation model was able to explain about 73 percent of the variability in the volumetric water content data (Figure 10). The model overpredicts lower TDR readings and slightly underpredicts the wetter ones. The seasonality and timing of soil moisture depletion and wetting of the treatment plot simulations below 1 m also seem delayed by 1–3 months. Additionally, from 6 to 11 m the model simulates a greater drawdown of water than the TDR data indicate, especially during the second posttreatment rainy season (Figure 9).

[49] The greater simulated drawdown in the deeper soil layers could be due to incorrect assumptions regarding the K_s or the root distribution or function in those layers. The K_s values estimated by the power function may be too high, which would drain these deeper layers too fast. The model also lacks a mechanism to account for a change in the distribution of fine roots. Fine root biomass down to 6 m at this site was first estimated in August 2000. A second series of root sampling was performed in July 2001, over 2 years after the start of the experiment and about 1.5 years after the treatment panels were first installed. Samples were collected only between 0 to 2 m, corresponding to the depth where most of the soil moisture depletion had occurred. The results show no significant difference in root biomass between the treatment and control plots [*Nepstad et al.*, 2002]. Deeper depths were not sampled, however, so it is not known whether or how root biomass changed below 2 m. As the surface layers continue to dry, increased fine

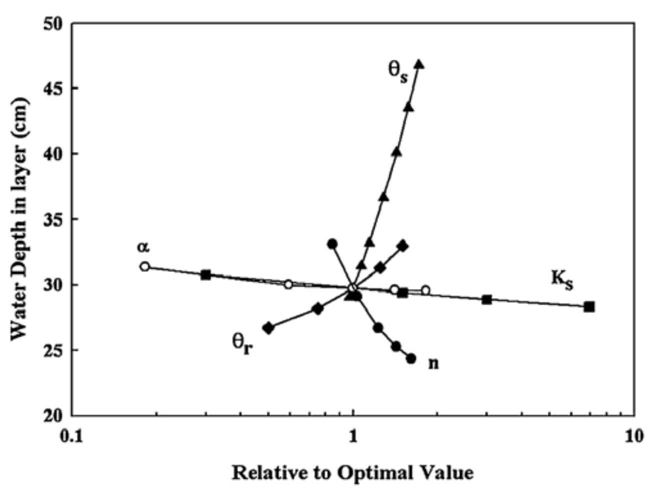

Figure 8. Sensitivity analysis of van Genuchten parameters (θ_s, θ_r, α, and n) and K_s for 1.5–2.5 m showing the effect of a change in the parameter values on the average depth of water in that layer over the 960-day simulation period. The change in parameter values is relative to the default values listed in Table 2.

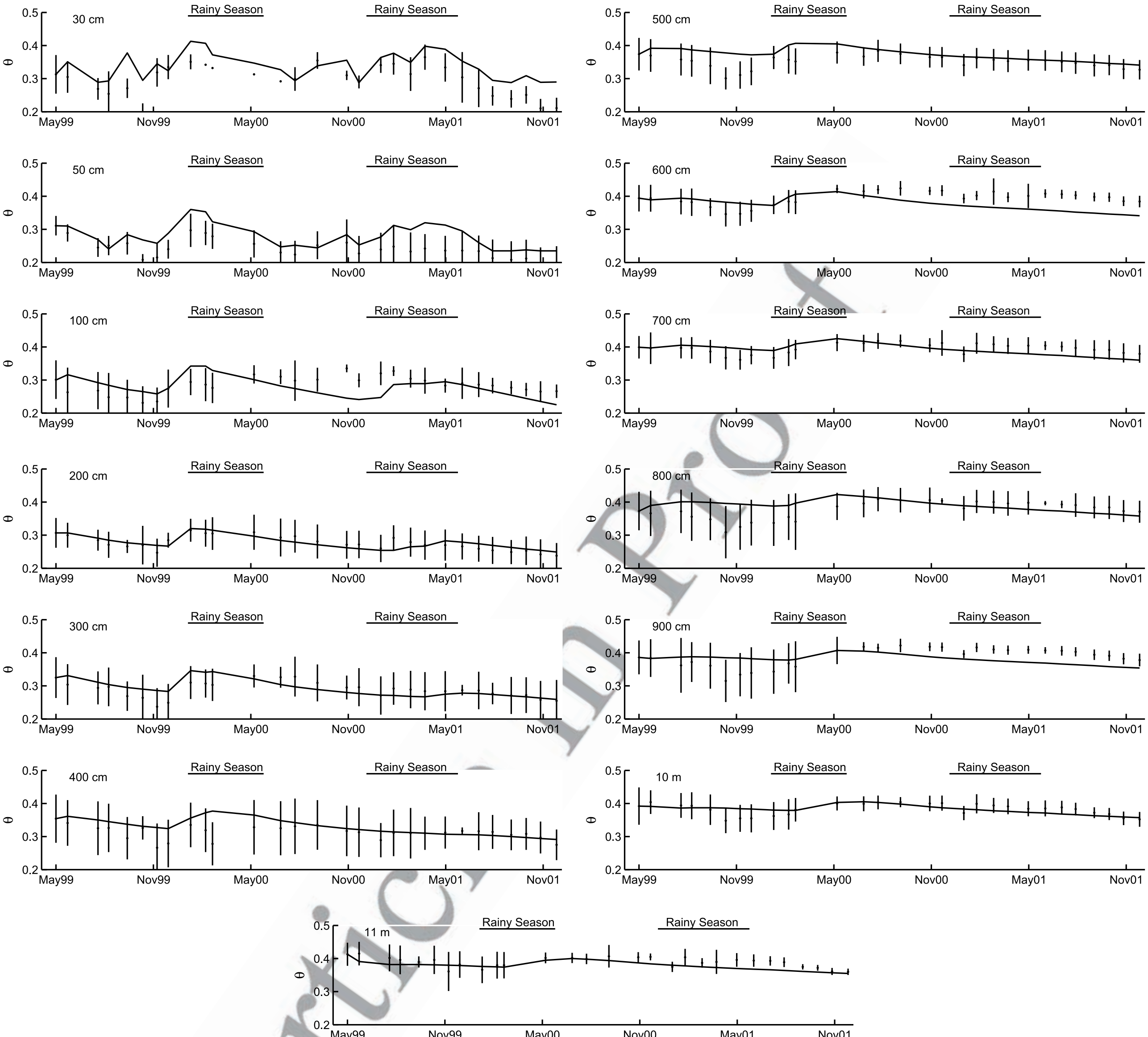

Figure 9. Monthly measured (dots) versus predicted (lines) volumetric water contents (θ) in the treatment plot, Tapajós National Forest, Brazil. Measured θ are averages of six TDR sensors per depth. Also shown are vertical bars representing measurement standard deviations, where available.

root growth at depth has been hypothesized [*Nepstad et al.*, 2002].

[50] Finally, recent work at this same site has demonstrated a potential for hydraulic redistribution of water through roots [*Oliveira et al.*, 2005]. Hydraulic redistribution can move water passively through roots either upward or downward whenever a gradient in soil water potential exists among soil layers which is stronger than the overall gradient between soil and atmosphere. Hydraulic redistribution has been well documented in drier ecosystem but only with this work has it been demonstrated in moist tropical forest ecosystem. In fact, there was increased evidence for downward hydraulic redistribution in the dry-down plot of this study relative to the control [*Oliveira et al.*, 2005]. Unfortunately, estimating the mass of water that may move through these hydraulic processes is difficult. The estimate for this site suggests as much as 10 percent of rainfall inputs my be transported to deeper soils through this process [*Lee et al.*, 2005].

5. Hydrologic Budgets

[51] We compared the simulated hydrologic budgets for both plots to further elucidate the mechanisms driving the soil draw down observed in the treatment plot. Over the 960-day simulation period, there was an average of 5.3 mm d^{-1} rainfall, 4.6 mm d^{-1} throughfall, and 0.63 mm d^{-1} interception. On an annual basis, the 1925 mm of rainfall is near the average of 2000 mm reported by *Nepstad et al.* [2002] for this site. This 2-year average belies the fact that in 2000 the rainfall was about 24 percent above normal (2469 mm) and in 2001 rainfall was 10 percent below normal (1798 mm).

[52] We estimated an average of 12 percent interception of gross rainfall on the basis of data for our site. This value

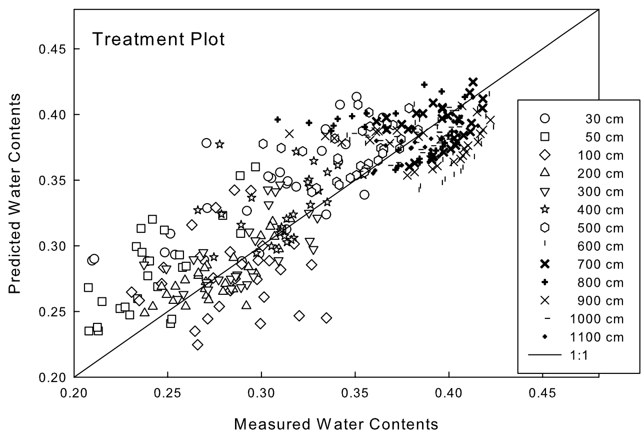

Figure 10. Scatterplot of measured and predicted volumetric water contents (θ) in the treatment plot, Tapajós National Forest, Brazil. The comparison is for 13 depths and 29 dates between May 1999 and December 2001 on which water content (θ) was measured at the field site.

is less than the 20 percent interception that *Nepstad et al.* [2002] reported for the site during the 2000 rainy season. In two other terra firme forest sites near Marabá and Ji-Paraná in Brazil, *Ubarana* [1996] reported 13–14 percent interception, while in the Columbian Amazon, *Marin et al.* [2000] report an interception of 13–18 percent.

5.1. Evapotranspiration

[53] In the control plot simulation the average evaporation of 0.63 mm d^{-1} intercepted rainfall plus the average transpiration of 3.07 mm d^{-1} plant uptake yielded an actual evapotranspiration (AET) rate of 3.7 mm d^{-1}. Water balance studies have estimated AET rates of 4.15 mm d^{-1} for an eastern Amazonian forest [*Jipp et al.*, 1998], 4.1 mm d^{-1} for the central Amazon [*Leopoldo et al.*, 1995], and 3.59 and 3.65 mm d^{-1} for 2 years of field eddy correlation measurements near Manaus [*Shuttleworth*, 1988; *da Rocha et al.*, 1996]. Two recent eddy flux tower studies in close proximity to our site within the Tapajós National Forest measured AET at 3.45 mm d^{-1} for July 2000 to 2001 [*da Rocha et al.*, 2004] or 3.1 mm d^{-1} for January 2002 to 2004 [*Hutyra et al.*, 2005]. The average control plot AET rate also equals the value *Klinge et al.* [2001] simulated for an eastern Amazonian forest from a model using the Penman equation for PET and a matric head–dependent reduction function.

[54] PET is typically higher in the July to December dry season (5.0 mm d^{-1}) compared to the wet season (4.1 mm d^{-1}) because of a higher vapor pressure gradient between air and leaf surfaces. The model predicts that AET is equal to PET for most of the year, except during the dry season when soil moisture becomes limiting (Figure 11). On average, AET was 80 percent of PET, which is calculated to be 4.6 mm d^{-1} using the modified Thornthwaite model.

[55] In the treatment plot simulation, AET declined by 0.125 mm d^{-1}. Considering the exclusion of throughfall by the panels, only 2.85 mm d^{-1} water reached the soil in the treatment plot, as opposed to 4.64 mm d^{-1} in the control plot. Although less water is returned to the atmosphere in the treatment plot simulation, AET is 25 percent higher than the inputs that arrived at the soil surface. While evapotranspiration may exceed inputs for brief periods of time, the water storage in the soil would become depleted if this were to continue.

[56] Except for the top two layers, where uptake is restricted during the dry season, the fraction of actual uptake coming from each layer in the control plot strongly follows the assumed root distribution. The same is true for the treatment plot, although the uptake from the top two layers is more restricted. The layers at 1 m and below became slightly more important contributors to uptake as the exclusion treatment continued. This interpretation excludes hydraulic redistribution, which indicates that water might be redistributed from lower layers to upper layers allowing plant uptake for evapotranspiration to continues from upper layers [*Oliveira et al.*, 2005].

5.2. Water Movement and Storage

[57] K_s varies from 3.2×10^{-5} m s^{-1} in the surface layer to 8.2×10^{-7} m s^{-1} in the deepest layer (Figure 2). Observed K_s values are large for clay-rich soils [*Hillel*, 1998], but are within the 2×10^{-7} to 6.4×10^{-5} m s^{-1} range measured by researchers at another terra firme forest site in Paragominas with similar, deeply weathered Oxisols [*Moraes et al.*, 2006]. In fact, the range of variation, even for a given depth, is not atypical for K_s measurements, which can cover many orders of magnitude [*Rasmussen et al.*, 1993].

[58] Not surprisingly, unsaturated hydraulic conductivity ($K(\theta)$) values are markedly reduced from the maximum rates achieved at saturation. The simulated $K(\theta)$ for the control plot are on the order of 10^{-7} m s^{-1} at the surface to 10^{-9} m s^{-1} at other depths, while in the treatment plot the simulated $K(\theta)$ are on the order of 10^{-8} m s^{-1} at the surface to 10^{-9} m s^{-1} at other depths.

[59] The lower rates in the treatment plot mean that less water drains past each layer than in the control, where water fluxes are three to four times greater (Table 3). Before the treatment was applied, similar amounts of water drained through the profiles in both plots. After the panels were first installed in early February 2000, the average control plot fluxes went up because of the arrival of the rainy season. However, in the treatment plot, the average fluxes decreased at that time even with the increased rainfall.

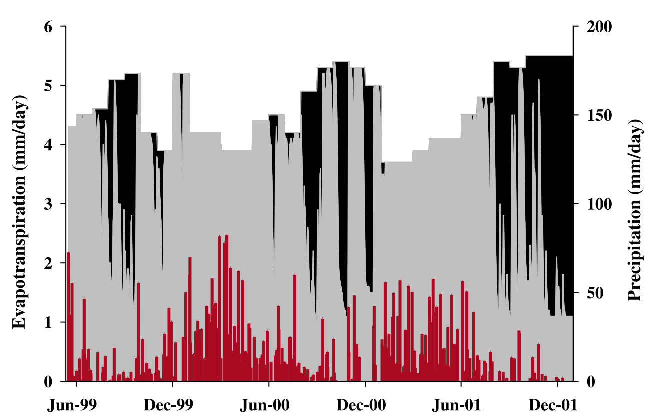

Figure 11. Potential evapotranspiration (PET) and simulated actual evapotranspiration (AET) for the Tapajós National Forest, Brazil. The dark areas represent periods of water deficit when AET is less than PET.

Table 3. Simulated Fluxes of Water in the Control Plot and Throughfall Exclusion Treatment Plot, Tapajós National Forest, Brazil[a]

	Control						Treatment					
	Pre	W1	D1	W2	D2	Total	Pre	W1	D1	W2	D2	Total
Rainfall	1.21	1.50	0.59	1.55	0.25	5.10	1.21	1.50	0.59	1.55	0.25	5.10
Interception	0.14	0.18	0.07	0.19	0.03	0.61	0.14	0.18	0.07	0.19	0.03	0.61
Exclusion	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.80	0.12	0.81	0.00	1.73
Throughfall	1.07	1.32	0.52	1.36	0.22	4.49	1.06	0.52	0.40	0.55	0.22	2.75
Actual ET	0.98	0.63	0.75	0.72	0.51	3.59	0.98	0.61	0.65	0.72	0.47	3.43
Flux												
0.40 m depth	0.51	1.08	0.18	1.02	0.03	2.82	0.51	0.27	0.08	0.16	0.02	1.04
0.75 m depth	0.39	0.93	0.12	0.85	0.01	2.31	0.39	0.23	0.02	0.11	0.00	0.75
1.50 m depth	0.34	0.93	0.13	0.82	0.03	2.24	0.34	0.25	0.00	0.06	0.00	0.65
2.50 m depth	0.35	0.93	0.15	0.76	0.05	2.23	0.35	0.27	0.01	0.02	0.00	0.65
3.50 m depth	0.34	0.93	0.18	0.69	0.09	2.22	0.34	0.30	0.02	0.00	0.00	0.66
4.50 m depth	0.32	0.92	0.20	0.63	0.12	2.19	0.32	0.31	0.04	0.00	0.00	0.67
5.50 m depth	0.32	0.90	0.22	0.58	0.15	2.16	0.32	0.30	0.05	0.01	0.00	0.68
6.50 m depth	0.34	0.85	0.24	0.53	0.19	2.14	0.34	0.27	0.07	0.02	0.00	0.70
7.50 m depth	0.36	0.80	0.25	0.48	0.21	2.11	0.36	0.24	0.09	0.02	0.00	0.71
8.50 m depth	0.36	0.76	0.28	0.44	0.24	2.07	0.36	0.21	0.11	0.03	0.01	0.72
9.50 m depth	0.37	0.71	0.29	0.40	0.27	2.05	0.37	0.18	0.12	0.04	0.01	0.74
10.50 m depth	0.38	0.67	0.31	0.36	0.30	2.02	0.38	0.16	0.14	0.05	0.02	0.75
11.50 m depth	0.40	0.63	0.34	0.33	0.33	2.03	0.40	0.14	0.16	0.06	0.03	0.80
Δ Soil Storage	−0.43	0.45	−0.45	0.50	−0.60	−0.53	−0.21	−0.04	−0.33	−0.05	−0.26	−0.89

[a]The pretreatment period (Pre) is May 1999 to January 2000; the first wet season (W1) is February to June 2000; the first dry season (D1) is July to December 2000; the second wet season (W2) is January to June 2001; and the second dry season (D2) is July to December 2001. All units are in meters.

[60] The fraction of water lost to deep drainage is smaller under the treatment. In the control plot, about 45 percent of water input to the soil is drained past 11.5 m, compared to 17 percent in the treatment plot. The negative change in water storage in the control plot is an artifact of the 960-day time span for which the simulated fluxes of Table 3 are reported. This period covers both pretreatment and posttreatment periods and includes three dry seasons but only two wet seasons.

[61] The measured water content in the control plot over the entire simulation period clearly demonstrates that soil moisture was recharged during the 2001 rainy season (Figure 12). The measured water contents in the control plot also show that the soils at depth are wetter than near the surface. During the dry season, water is withdrawn from the entire profile, especially in the upper profile where there is a higher concentration of roots. By the middle of the rainy season, the surface soils rewet and the water storage below 4 m recharges.

[62] The soils in the treatment plot were drier than the soils in the control plot even before the exclusion panels were first installed in February 2000 (Figure 12). Because the dry season preceding the first treatment period and the treatment period itself were wetter than average, the panels did not divert sufficient water to invoke drought stress in the vegetation within the treatment plot [*Nepstad et al.*, 2002]. However, during the second treatment period, the soil near the surface dried more extensively and recharge at depth was not complete. The predicted water contents show similar patterns with depth, the most notable difference being that the current model predicts that the soil below 5 m in the treatment plot dries out more than measured during the second treatment year.

6. Conclusions

[63] A soil water model using Darcy's law and Richard's equation is presented for the purpose of evaluating unsaturated water fluxes and storage in a moist tropical forest soil. Model predictions are compared with soil water content estimates.

[64] The one-dimensional model used in this study predicts soil volumetric water content within 3 percent of water content measures obtained using TDR probes in six 11-m-deep soil shafts for the first 960 days of a throughfall reduction experiment under a moist tropical forest. This accuracy of prediction is quite impressive and indicates that physical processes of soil water movement in the ecosystem are captured by the model even despite the relatively coarse vertical and temporal scale of modeling. Landscapes with more complex terrain may require models with additional dimensions, but one-dimensional, vertical flow seems appropriate for this well-drained plateau site, a common feature in the Amazon basin.

[65] The model is sensitive to the van Genuchten parameters, θ_s, θ_r, and n, but less sensitive to K_s and α. These parameters are used to translate water content to head and to determine the unsaturated water flow function. Theoretically, the water retention properties these parameters describe are physical properties of the soil that can be quantified, although in our model we needed to calibrate these parameters to reproduce the observed soil moisture data.

[66] During the first year of throughfall exclusion, the measured water contents demonstrate, and the model predicts, mild soil water depletion near the surface. Persistence of the drought into a second year leads to more extensive drying of the surface soils and prevented complete recharge of water stored deeper in the soil. The model predicts that evapotranspiration declined during this period, and that water drainage was diminished. More importantly, however, our model shows that decreases in evapotranspiration were marginal while decreases in water flux were substantial.

[67] Clearly, soil water stores were being depleted by a reduction in soil moisture inputs. Over prolonged periods of drought this imbalance between water inputs and evapo-

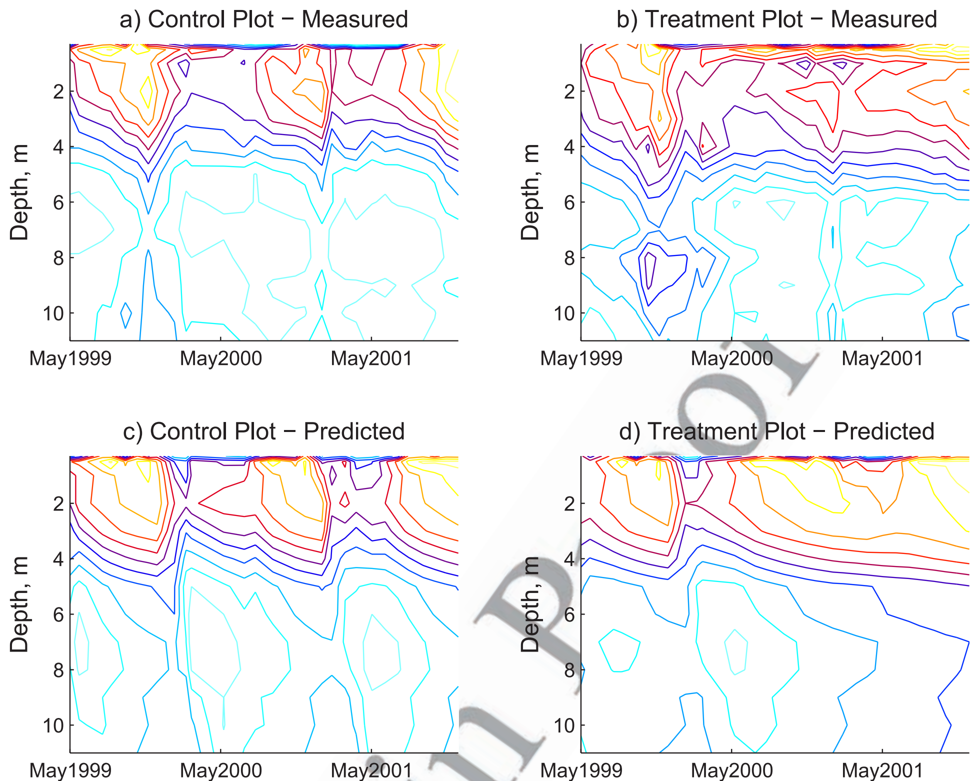

Figure 12. Measured water contents ($m^3\ m^{-3}$) in the (a) control and (b) treatment plots and simulated water contents in the (c) control and (d) treatment plots, Tapajós National Forest, Brazil. Each depth is an average of six TDR sensors. Range of water contents is from 0.18 to 0.46 in increments of 0.02. Drier soils are marked in red and wetter soils in blue. Throughfall reduction in the treatment plot commenced at the start of the 2000 rainy season.

transpiration are unsustainable once soil moisture reserves are exhausted. In fact, it is exactly such an increase in drought severity that has been predicted in response to global climate change that could exceed the limit of drought tolerance of these moist tropical forests, which the continuation of this experiment is designed to test.

[68] **Acknowledgments.** We wish to thank David Radcliffe, Miguel Cabrera, John Dowd, A. Marysol Schuler, David Redman, and Paul Lefebvre, as well as the researchers at the Woods Hole Research Center and the Instituto de Pesquisa Ambiental da Amazônia for their various contributions to this project. Field aspects of this research were funded by NSF grant DEB 9707566. The UGA Graduate School provided salary support for the lead author during model development. We also thank the Instituto Brasileiro do Meio Ambiente e dos Recursos Naturais Renováveis for logistical support on the Tapajós National Forest.

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