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Estimating Interception from Near-Surface Soil Moisture Response

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Abstract

Interception is the storage and subsequent evaporation of rainfall by above-ground structures, including canopy and groundcover vegetation and surface litter. Accurately quantifying interception is critical for understanding how ecosystems partition incoming precipitation, but it is difficult and costly to measure, leading most studies to rely on modeled interception estimates. Moreover, forest interception estimates typically focus only on canopy storage, despite the potential for substantial interception by groundcover vegetation and surface litter. In this study, we developed an approach to quantify "total" interception losses (i.e., including forest canopy, understory, and surface litter layers) using measurements of shallow soil moisture dynamics during rainfall events. Across 36 pine and mixed forest stands in Florida (USA), we used soil moisture and rainfall data to estimate the interception storage capacity (β_s), a parameter required to estimate total annual interception losses (I_a) relative to rainfall (R). Estimated values for β_s (mean β_s = 0.30 cm; $0.01 \leq \beta_s \leq 0.62$ cm) and I_a/R (mean I_a/R = 0.14; $0.06 \leq I_a/R \leq 0.21$) were consistent with reported literature values for these ecosystems and were significantly predicted by forest structural attributes (leaf area index and percent groundcover), as well as other site variables (e.g., water table depth). The best-fit model was dominated by LAI and explained nearly 80% of observed β_s variation. These results suggest that whole-forest interception can be measured using a single near-surface soil moisture time series and highlight the variability in interception losses across a single forest type, underscoring the need for expanded empirical measurement. Potential cost savings and logistical advantages of this method relative to conventional, labor-intensive interception measurements may improve empirical estimation of this critical water budget element.

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Introduction

Rainfall interception (*I*) is the fraction of incident rainfall stored by above-ground ecosystem structures (i.e., vegetation and litter layers) and subsequently returned to the atmosphere via evaporation (*E*), never reaching the soil surface and thus never directly supporting transpiration (*T*) [*Savenije*, 2004]. Interception depends on climate and vegetation characteristics and can be as high as 50% of gross rainfall [*Gerrits et al.*, 2007; 2010; *Calder*, 1990]. Despite being critical for accurate water budget enumeration [*David et al*., 2005], interception is often disregarded or lumped with evapotranspiration (*ET*) in hydrological models [*Savenije*, 2004]. Recent work suggests interception uncertainty constrains efforts to partition *ET* into *T* and *E*, impairing representation of water use and yield in terrestrial ecosystems [*Wei et al.*, 2017].

When interception is explicitly considered, it is typically empirically estimated or modeled solely for the tree canopy. For example, direct measurements are often obtained from differences between total rainfall and water that passes through the canopy to elevated above-ground collectors (throughfall) plus water that runs down tree trunks (stemflow) during natural [e.g., *Bryant et al.*, 2005, *Ghimire et al.*, 2012, 2016] or simulated [e.g., *Guevara-Escobar et al.*, 2007; *Putuhena and Cordery*, 1996] rainfall events. This method yields the rainfall fraction held and subsequently lost by the canopy but ignores interception by understory vegetation and litter. Alternatively, numerous empirical [e.g., *Merriam*, 1960], process-based [e.g., *Rutter et al.*, 1971, 1975; *Gash*, 1979, 1995, *Liu*, 1998], and stochastic [*Calder*, 1986] models are available for estimating interception. As with direct measurements, most model applications consider only canopy storage despite groundcover (both understory vegetation and litter layers) interception that can exceed canopy values [*Gerrits and Savenije*, 2011; *Putuhena and Cordery*, 1996]. As

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such, it seems likely that conventional measures and typical model applications underestimate actual (i.e., "total") interception.

New field approaches are needed to improve quantification of total interception and refine the calibration and application of available models. A detailed review of available interception models [*Muzylo et al.,* 2009] stresses the need for direct interception measurements across forest types and hydroclimatic regions, but meeting this need will require substantial methodological advances. Throughfall measurements yield direct and site-specific interception estimates [e.g., *Ghimire et al.,* 2017; *Bryant et al.,* 2005], but they are difficult and costly to implement even at the stand scale because of high spatial and temporal variability in vegetation structure. Moreover, comprehensive measurements also require enumeration of spatially heterogeneous stemflow, as well as interception storage by the understory and litter layers, greatly exacerbating sampling complexity and cost [*Lundberg et al.*,1997]. Empirical techniques that estimate total interception, integrate across local spatial and temporal variation, and minimize field installation complexity are clearly desirable.

Here we present a novel approach for estimating total (i.e., canopy, understory and litter) interception using continuously logged, near-surface soil moisture. Prior to runoff generation, infiltration is equivalent to rainfall minus total interception, and the response of near-surface soil moisture during and directly following rain events can be used to inform interception parameters and thus interception losses. Since soil moisture is relatively easy and economical to measure continuously for extended periods, successful inference of interception from soil moisture time series may greatly expand the temporal and spatial domains of empirical interception measurements. As a proof-of-concept, we tested this simple interception estimation method in 36

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92 forest plots spanning a wide range of conditions (e.g., tree density, composition, groundcover,
93 understory management, age, and hydrogeologic setting) across Florida (USA).
94
95 **Methods**
96 **Estimating Interception Storage Capacity from Soil Moisture Data**
97 During every rainfall event, a portion of the total precipitation (*P*) is temporarily stored in
98 the forest canopy and groundcover (hereafter referring to both live understory vegetation and
99 forest floor litter). We assume that infiltration (and thus any increase in soil moisture) begins
100 only after total interception storage, defined as the sum of canopy and groundcover storage, is
101 full. We further assume this stored water subsequently evaporates to meet atmospheric demand.
102 Calculating dynamic interception storage requires first determining the total storage capacity
103 (β_s), which is comprised of the storage capacities for the forest canopy (β_c) and groundcover (β_g)
104 (Fig. 1a).
105 To estimate β_s, we consider a population of individual rainfall events of varying depth
106 over a forest for which high frequency (i.e., 4 hr^{-1}) soil-moisture measurements are available
107 from near the soil surface. Soil moisture content (*SMC*) at the sensor changes only after rainfall
108 fills total interception storage, evaporative demands since rainfall onset are met, and there is
109 sufficient infiltration for the wetting-front to arrive at the sensor. Rainfall events large enough to
110 induce a soil moisture change (ΔSMC) are evident as a rainfall threshold in the relationship
111 between *P* and ΔSMC. An example time series of *P* and *SMC* (Fig. 1b) yields a *P* versus ΔSMC
112 relationship (Fig. 1c) with clear threshold behavior. There are multiple equations whose
113 functional forms allow for extraction of this threshold; here we express this relationship as:
114 $$P = \frac{a}{(1+b*exp^{(-c*\Delta SMC)})} \quad (1)$$

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where P is the total rainfall event depth, ΔSMC is the corresponding soil moisture change, and a, b, and c are fitted parameters. Figure 2 illustrates this relationship and model fitting for observed SMC data from six plots at one of our study sites described below. The x-intercept of Eq. 1 (i.e., where ΔSMC departs from zero) is given by:

$$P_s = \frac{a}{(1+b)} \quad (2)$$

Empirically observed values of P_s represent the total rainfall required to saturate β_s, meet evaporative demands between storm onset and observed ΔSMC, and supply any infiltration required to induce soil moisture response once interception storage has been saturated. This equality can be expressed as:

$$P_s = \beta_s + \int_0^T E dt + \int_t^T f dt = \beta_s + \int_0^t E dt + \int_t^T E dt + \int_t^T f dt \quad (3)$$

where T is the total time from rainfall onset until observed change in SMC (i.e., the wetting front arrival), t is the time when β_s is satisfied, and E and f are infiltration and evaporation rates, respectively. To connect this empirical observation to existing analytical frameworks (e.g., Gash 1979), we adopt the term P_G, defined as the rainfall depth needed to saturate β_s and supply evaporative losses between rainfall onset (t = 0) and β_s saturation (t = t):

$$P_G = \beta_s + \int_0^t E dt \quad (4)$$

Solving for β_s in Eq. 3 and substituting into Eq. 4 yields:

$$P_G = P_s - \int_t^T E dt - \int_t^T f dt \quad (5)$$

Equation 5 may be simplified by assuming that average infiltration and evaporation rates apply during the relatively short period between t and T, such that:

$$P_G = P_s - \acute{f}(T - t) - \acute{E}(T - t) \quad (6)$$

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where $\bar{f}$ is the average soil infiltration rate and $\acute{E}$ is the average rate of evaporation from the forest surface (i.e., canopy, groundcover, and soil) during the time from *t* to *T* (see Gash 1979). The storage capacity β_s can now be calculated following Gash (1979) as:

$$\beta_s = -P_G \frac{\acute{E}}{\acute{R}} ln\left(1 - \frac{\acute{E}}{\acute{R}}\right) = \frac{-\acute{E}}{\acute{R}} \frac{[P_s - (T-t)(\acute{f}+\acute{E})]}{ln\left(1-\frac{\acute{E}}{\acute{R}}\right)} \quad (7)$$

where $\bar{R}$ is the rainfall rate and all other variables are as previously defined. In Eq. 5, $\acute{E}$ is usually estimated using the Penman-Monteith equation [*Monteith,* 1965], setting canopy resistance to zero (e.g., Ghimire et al 2017).

A key challenge in applying Eq. 5, and thus for the overall approach, is quantifying infiltration, since the time, *t*, when P_G is satisfied is unknown. Moreover, the infiltration rate embedded in P_s is controlled by the rainfall rate ($\bar{R}$) and initial soil moisture content (θ_i). It is worth noting that shallower sensor depth placement would likely eliminate the need for this step (see Discussion). However, to overcome this limitation in our study, we used the 1-D unsaturated flow model HYDRUS-1D (Simunek et al., 1995) to simulate the time it takes for the wetting front to arrive (T_w) at the sensor under bare soil conditions across many combinations of $\bar{R}$ and θ_i. As such, T_w represents the time required for a soil moisture pulse to reach the sensor once infiltration begins (i.e., after total interception capacity has been filled), which is *T*- *t* in Eq. 7. For each simulation, T_w (signaled by the first change in *SMC* at sensor depth) was recorded and used to develop a statistical model of T_w as a function of $\bar{R}$ and θ_i. We used plot-specific soil moisture retention parameters from Florida Soil Characterization Retrieval System (https://soils.ifas.ufl.edu/flsoils/) to develop these curves for our six sites, but simulations can be applied for any soil with known or estimated parameters.

Simulations revealed that T_w at a specific depth declined exponentially with increasing θ_i:

$$T_w = ae^{-b\theta_i} \quad (8)$$

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where a and b are fitting parameters. Moreover, the parameters a and b in Eq. (6) are well fitted by a power function of $\bar{R}$:

$$a = a_1 \acute{R}^{a_2}, b = b_1 \acute{R}^{b_2} \quad (9)$$

where a_1 and b_1 are fitting parameters. These relationships are illustrated in Fig. 3 for a loamy sand across a range of $\bar{R}$ and θ_i. The relationship between initial *SMC* and T_w is very strong for small to moderate $\bar{R}$ (< 3.0 cm/hr). At higher values of $\bar{R}$, T_w is smaller than the 15-minute sampling resolution, and these events were excluded from our analysis (see below).

Assuming that $\bar{f}$ equals $\bar{R}$ over the initial infiltration period from t to T (robust for most soils, see below), Eq. 7 can be modified to:

$$\beta_s = \frac{-\acute{E}}{\acute{R}} \left[\frac{P_s - T_w(\acute{R}+\acute{E})}{ln\left(1-\frac{\acute{E}}{\acute{R}}\right)} \right] \quad (10)$$

This approach assumes no runoff or lateral soil-water flow near the top of the soil profile from time t to T. Except for very fine soils under extremely high $\bar{R}$, this assumption generally holds during early storm phases, before ponding occurs (Mein and Larsen, 1973). Moreover, since our goal is to determine β_s, extreme storms can be omitted from the analysis when implementing Eqs. 1-10, without compromising our estimates. Finally, we note that values of β_s from Eq. 10 represent combined interception from canopy and groundcover, but the method does not allow for disaggregation of these two components.

Calculating Interception Loss

Interception storage and resulting interception loss for a given rain event are driven by both antecedent rain (which fills storage) and evaporation (which depletes it). Instantaneous available storage ranges from zero (saturated) to the maximum capacity (i.e., β_s which occurs when the storage is empty). While discrete, event-based interception models [*Gash*, 1979, 1995; *Liu*, 1998] have been widely applied to estimate interception, continuous models more accurately

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represent time-varying dynamics in interception storage and losses. We adopted the continuous, physically-based interception modeling framework of *Liu* [1998, 2001]:

$$I = \beta_s(D_0 - D) + \int_0^T (1 - D)Edt \quad (11)$$

where I is interception, E is the evaporation rate from wetted surfaces, D_0 is the forest dryness index at the beginning of a rain event, and D is the forest dryness index at time T. The dryness index is calculated as:

$$D = 1 - \frac{C}{\beta_s} \quad (12)$$

where C is "adherent storage" (i.e., water that does not drip to the ground) and is given by:

$$C = \beta_s\left(1 - D_0 exp\left(\frac{-(1-\tau)}{\beta_s}P\right)\right) \quad (13)$$

where τ is the free throughfall coefficient. Because our formulation of β_s in Eq. 10 incorporates both canopy and groundcover components (i.e., negligible true throughfall), we approximated τ in Eq. 13 as zero. For single storms or when sufficient time has passed to dry the canopy, D_0 is assumed to be unity [*Liu* 2001]. Between rainfall events, water in interception storage evaporates to meet atmospheric demand, until the dryness index, D reaches unity [*Liu* 1997]. The rate of evaporation from wetted surfaces between rainfall events (E_s) is:

$$E_s = E(1 - D_0)exp\left(\frac{E}{\beta_s}\right) \quad (14)$$

A numerical version of Eq. 9 to calculate interception at each time step, t, is expressed as:

$$I = \beta_s(D_{t-1} - D_t) + \frac{1}{2}[E_{t-1}(1 - D_{t-1}) + E_t(1 - D_t)] \quad (15)$$

Eq. 15 quantifies continuous and cumulative interception losses using precipitation and other climate data (for E) along with β_s derived from soil moisture measurements and corresponding meteorological data.

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203 **Study Area and Data**

204 As part of a multi-year study quantifying forest water use under varying silvicultural
205 management, we instrumented six sites across Florida, each with six 2-ha plots spanning a wide
206 range of forest structural characteristics. Sites varied in hydroclimatic forcing (annual
207 precipitation range: 131 to 154 cm/yr and potential *ET* range: 127 to 158 cm/yr) and
208 hydrogeologic setting (shallow vs. deep groundwater table). Experimental plots within sites
209 varied in tree species, age, density, leaf area index (LAI), groundcover density (%GC), soil type,
210 and management history (Table 1). Each site contained a recent clear-cut plot, a mature pine
211 plantation plot, and a restored longleaf pine (*Pinus palustris*) plot; the three remaining plots at
212 each site included stands of slash pine (*Pinus elliottii*), sand pine (*Pinus clausa*), or loblolly pine
213 (*Pinus taeda*) subjected to varying silvicultural treatments (understory management, canopy
214 thinning, prescribed burning) and hardwood encroachment.

215 Within each plot, three banks of TDR sensors (CS655, Campbell Scientific, Logan, UT,
216 USA) were installed to measure soil moisture at multiple soil depths (Fig. 1a). Only data from
217 the top-most sensor (15 cm below the ground surface) were used in this study. Soil-moisture
218 sensor banks were located to capture representative variation in stand geometry (i.e., below the
219 tree canopy and within inter-canopy rows), and thus capture variation in surface soil moisture
220 response to rainfall events driven by forest canopy and groundcover differences. Within each
221 clear-cut plot at each site, meteorological data (rainfall, air temperature, relative humidity, solar
222 insolation, wind speed and direction) were measured using a weather station (GRSW100,
223 Campbell Scientific, Logan, UT; Fig. 4c) every 3 seconds and used to calculate hourly *E* by
224 setting the canopy resistance to zero [*Ghimire et al.*, 2017; *Gash*, 1995; *Monteith*, 1965].
225 Growing season forest canopy LAI (m^2 m^{-2}) and groundcover (%) were measured at every 5-m

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node within a 50 m x 50 m grid surrounding soil moisture measurement banks. LAI was measured at a height of 1 m using a LI-COR LAI-2200 plant canopy analyzer, and %GC was measured using a 1 m^2 quadrat.

To estimate β_s, mean ΔSMC values from the three surface sensors were calculated for all rainfall events separated by at least 72 hours. Storm separation was necessary to ensure the canopy and groundcover surfaces were mostly dry at the onset of each included rainfall event. Rainfall events were binned into discrete classes by depth and plotted against mean ΔSMC to empirically estimate P_s (e.g., Fig. 2). For each rainfall bin, mean θ_i, $\overline{R}$ and $\overline{E}$ were also calculated to use in Eq. 10, which was then applied to calculate β_s. Subsequently, we developed generalized linear models (GLMs) using forest canopy structure (site-mean LAI), mean groundcover (% GC), hydrogeologic setting (shallow vs. deep groundwater table), and site as potential predictors, along with their interactions, to statistically assess predictors of β_s estimates. Because models differed in fitted parameter number, the best model was selected using the Akaike Information Criteria (AIC; Akaike, 1974). Finally, we calculated cumulative annual interception loss (I_a) and its proportion of total rainfall for each study plot using the mean β_s for each plot (across the 3 sensor banks), climate data from 2014 to 2016, and Eq. 15. All analyses were performed using R statistical software [*R Core Team*, 2017].

Results

Total Storage Capacity (β_s)

The exponential function used to describe the *P*-ΔSMC relationship (Eq. 1) showed strong agreement with observations at all sites and plots (overall $R^2 = 0.80$; $0.47 \leq R^2 \leq 0.97$; Table 1) as illustrated for a single site in Fig. 2. This consistency across plots and sites suggests

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249 that Eq. 1 is capable of adequately describing observed *P*-Δ*SMC* relationships, enabling
250 estimates of β_s across diverse hydroclimatic settings and forest structural variation. Estimates of
251 β_s ranged from 0.01 to 0.62 cm, with a mean of 0.30 cm. Plot-scale LAI was moderately
252 correlated with plot-mean β_s, describing roughly 32% of observed variation across plots (Fig.
253 4a). This relatively weak association may arise because LAI measurements only characterize
254 canopy cover, while β_s combines canopy and groundcover storage. The best GLM of β_s (Fig. 4b)
255 used %GC and an interaction term between site and LAI (R^2 = 0.84 and AIC = 253.7, Table 2).
256 The best GLM without site used LAI and hydrogeologic setting (shallow vs. deep water table)
257 but had reduced performance (R^2 = 0.55 and AIC = 338.3; Table 2).

258 **Annual Interception Losses (I_a)**

259 Despite having similar rainfall regimes (mean annual precipitation ranging from 131 to
260 154 cm yr^{-1} across sites), mean annual interception losses (I_a) differed significantly both across
261 sites (one-way ANOVA $p < 0.001$) and among plots within sites (one-way ANOVA $p < 0.001$).
262 Estimates of I_a/P across all plots and sites ranged from 6 to 21% of annual rainfall (Table 1) and
263 were moderately, but significantly, correlated with mean LAI, explaining approximately 30% of
264 variation in I_a (Fig. 5a). Correlations among I_a/P and LAI were stronger for individual sites than
265 the global relationship ($0.51 \leq R^2 \leq 0.84$), except for site EF, where I_a losses were small and
266 similar across plots regardless of LAI (Fig. 5b; Table 1). This suggests that additional site-level
267 differences (e.g., hydroclimate, soils, geology) play a role in driving I_a, as expected following
268 from their effects on β_s described above.

269 **Discussion and Conclusions**

270 When combined with local rainfall data, near-surface soil moisture dynamics inherently
271 contain information about rainfall interception by above-ground structures. Using soil moisture

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data, we developed and tested an analytical approach for estimating total interception storage capacity (β_s) that includes canopy, understory, and groundcover vegetation, as well as any litter on the forest floor. The range of β_s given by our analysis (mean β_s = 0.30 cm; $0.01 \leq \beta_s \leq 0.62$ cm) is close to, but generally higher than previously reported canopy storage capacity values for similar pine forests (e.g., 0.17 to 0.20 cm for mature southeastern USA pine forests; Bryant et al. 2005).

An important distinction between our method and previous interception measurement approaches is that the soil moisture-based method estimates composite rainfall interception of not only the canopy, but also of the groundcover vegetation and forest floor litter. Rainfall storage and subsequent evaporation from groundcover vegetation and litter layers can be as high, or higher than, canopy storage in many forest landscapes [*Putuhena and Cordery*, 1996; *Gerrits et al.*, 2010]. For example, *Li et al.* [2017] found that the storage capacity of a pine forest floor in China was between 0.3 and 0.5 cm, while maximum canopy storage was < 0.1 cm. *Putuhena and Cordery* [1996] also estimated storage capacity of pine forest litter to be approximately 0.3 cm based on direct field measurements. *Gerrits et al.* [2007] found forest floor interception to be 34% of measured precipitation in a beech forest, while other studies have shown that interception by litter can range from 8 to 18% of total rainfall [*Gerrits et al.*, 2010; *Tsiko et al.*, 2012; *Miller et al.*, 1990; *Pathak et al.*, 1985; *Kelliher et al.*, 1992]. A recent study using leaf wetness observations [*Acharya et al.* 2017] found the storage capacity of eastern redcedar (*Juniperus virginiana*) forest litter to range from 0.12 to as high as 1.12 cm, with forest litter intercepting approximately 8% of gross rainfall over a six-month period. Given the composite nature of forest interception storage and the range of storage capacities reported in these studies, the values we report appear to be plausible, and consistent with the expected differences between canopy-only

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295 and total interception storage. As such, our results support the general applicability of the soil
296 moisture-based approach for developing forest interception estimates across a wide range of
297 hydroclimatic and forest structural settings.

298 Interception losses vary spatially and temporally and are driven by both β_s and climatic
299 variation (i.e., P and E). Our approach represents storage dynamics by combining empirically
300 derived β_s estimates with climatic data using a previously developed continuous interception
301 model [*Liu* 1998, 2001]. Cumulative I_a estimates in this study ranged considerably (i.e., from 6%
302 to 21% of annual rainfall) across the 34 plots, which were characterized by variation in canopy
303 structure ($0.12 < \text{LAI} < 3.70$) and groundcover ($7.9 < \%\text{GC} < 86.2$). In comparison, interception
304 losses by pine forests reported in the literature (all of which report either canopy-only or
305 groundcover-only values, but not their composite) range from 12 to 49% of incoming rainfall
306 [*Bryant et al.*, 2005; *Llorens et al.*, 1997; *Kelliher and Whitehead*, 1992; *Crockford and*
307 *Richardson*, 1990]. Notably, most of the variation in this range is drive by climate rather than
308 forest structure, with the highest I_a values from more arid regions (e.g., Llorens et al. 1997).
309 Broad agreement between our results and literature I_a values supports the utility of our method
310 for estimating this difficult-to-measure component of the water budget. Additionally, the
311 magnitude and heterogeneity of our I_a estimates across a single forest type (southeastern US
312 pine) underscores the urgent need for empirical measurements of interception that incorporate
313 information on both canopy and groundcover storage in order to develop accurate water budgets.
314 This conclusion is further bolstered by the persistent importance of site-level statistical effects in
315 predicting β_s (and therefore I_a), even after accounting for forest structural attributes, which
316 suggests there are influential edaphic or structural attributes that we are not currently adequately
317 assessing.

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318 Generally, estimated I_a losses in clear-cut plots were smaller than plots with a developed
319 canopy, as expected. One exception was at EF where the clear-cut plot exhibited the highest I_a of
320 the six EF plots (8.4%, Table 1). Notably, differences among EF plots were very small (I_a ranged
321 only from 7.9 to 8.4 % of annual rainfall), an annual interception rate consistent with or even
322 slightly lower than other clear cuts across the study. This site is extremely well drained and has
323 dense litter dominated by mosses and nutrient-poor sandy soils, highlighting the potential for
324 additional local measurements to better understand how forest structure controls observed
325 interception.

326 There are several important methodological considerations and assumptions inherent to
327 estimating interception using near-surface soil moisture data. First is the depth at which SMC is
328 measured. Ideally, soil moisture would be measured a few centimeters into the soil profile,
329 eliminating the need to account for infiltration when calculating P_G in Eqs. (4-6). Soil moisture
330 data used here were leveraged from a study of forest water yield, with sensor deployment depths
331 selected to efficiently integrate soil moisture patterns through the vadose zone. While the extra
332 step of modeling infiltration may increase uncertainty in β_s, infiltration was extremely well-
333 described using wetting front simulations of arrival time based on initial soil moisture and
334 rainfall. As such, while we advocate for shallower sensors in future efforts, our solution here
335 given the depths that were available seem tenable for this and other similar data sets. Second, in
336 contrast to the original Gash (1979) formulation, Eq. 5 does not explicitly include throughfall.
337 While throughfall has been a critical consideration for rainfall partitioning by the forest canopy,
338 our approach considers total interception by aboveground forest structures (canopy, groundcover,
339 and litter). A portion of canopy throughfall is captured by non-canopy storage and thus
340 intercepted. Constraining this fraction is not possible with the data available, and indeed our soil

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moisture response reflects the "throughfall" passing the canopy, understory and litter. Similarly, estimation of β_s using Eqs. 1-7 cannot directly account for stemflow, which can be an important component of rainfall partitioning in forests (e.g., Bryant et al., 2005). We used the mean soil moisture response across three sensor locations (close to a tree, away from the tree but below the canopy, and within inter-canopy rows), which lessens the impact of this assumption on our estimates of β_s. Finally, Eqs. (3-10) assume the same evaporation rate, E, for intercepted water from the canopy and from the understory. Evaporation rates may vary substantially between the canopy, understory, and forest floor [*Gerrits et al.*, 2007, 2010], especially in more energy-limited environments. Future work should consider differential evaporation rates within each interception storage, particularly since the inclusion of litter as a component potentially accentuates these contrasts in E.

Rainfall interception by forests is a dynamic process that is strongly influenced by rainfall patterns (e.g., frequency, intensity), along with various forest structural attributes such as interception storage capacity (β_s) [*Gerrits et al.*, 2010]. In this work, we coupled estimation of a total (or "whole-forest") β_s parameter with a continuous water balance model [*Liu*, 1997, 2001; *Rutter et al.*, 1975], providing an integrative approach for quantifying time-varying and cumulative interception losses. We propose that soil moisture-based estimates of β_s have the potential to more easily and appropriately represent combined forest interception relative to existing time- and labor-intensive field methods that fail to account for groundcover and litter interception. Soil moisture can be measured relatively inexpensively and easily using continuous logging sensors that require little field maintenance, facilitating application of the presented approach across large spatial and temporal extents and reducing the time and resources that are needed for other empirical measures [e.g., *Lundberg et al.*, 1997]. Finally, while direct

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364 comparisons with other empirical measures of forest canopy interception should be treated
365 cautiously, this approach yields values that are broadly consistent with the literature, and provide
366 an estimate of combined canopy and groundcover storage capacity that has the potential to
367 improve the accuracy of water balances models at scales from the soil column to watershed.

368

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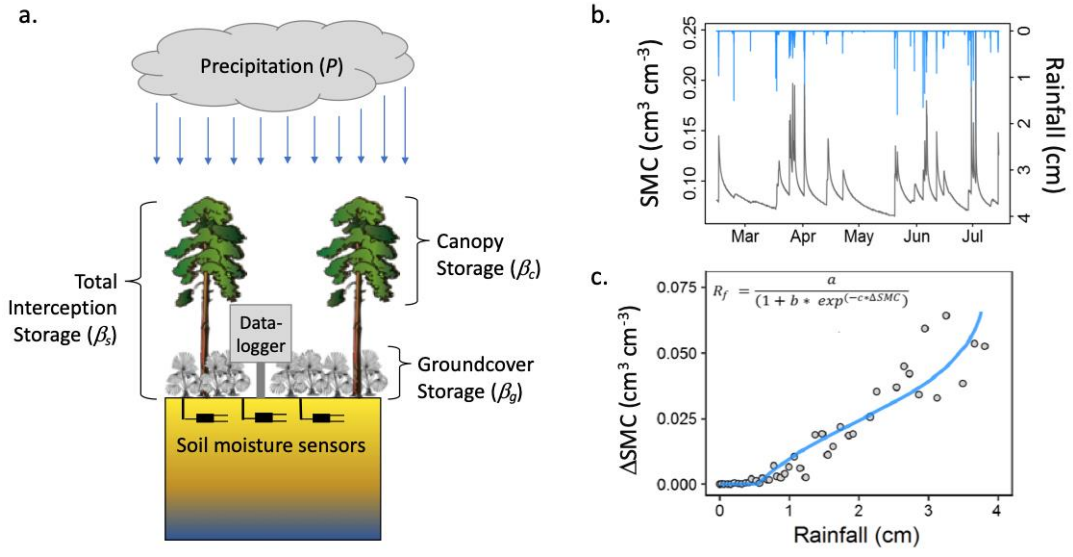

Figure 1. (a) Schematic illustration of experimental setup and interception water storages, where total interception storage (β_s) is the sum of canopy storage (β_c) and groundcover (understory and litter) storage (β_g). (b) Example time series of rainfall (blue lines) and corresponding near-surface soil moisture content (*SMC*, black line; observed at 15 cm in this study). (c) Resultant relationship between rainfall and change in soil moisture *ΔSMC* during rainfall, along with fitted model to extract the x-intercept (i.e., P_s).

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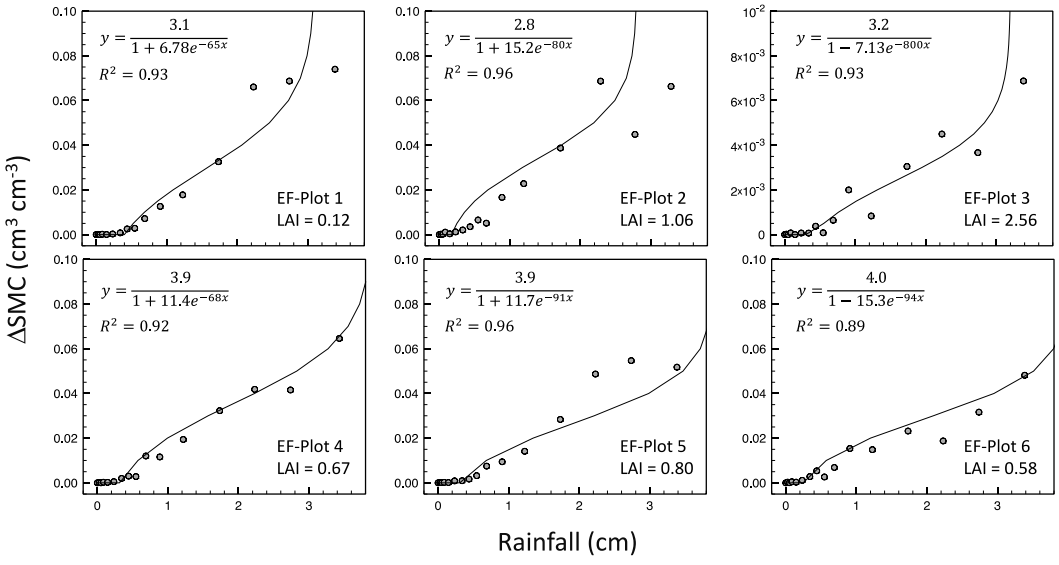

Figure 2: Change in soil moisture content (*ΔSMC*) versus binned rainfall depths for six plots at one of the study sites used in the study (Econfina; EF). The x-intercept of the fitted relationships were used to derive P_s in Eq. 2. Note different y-axis scale for EF-Plot 3.

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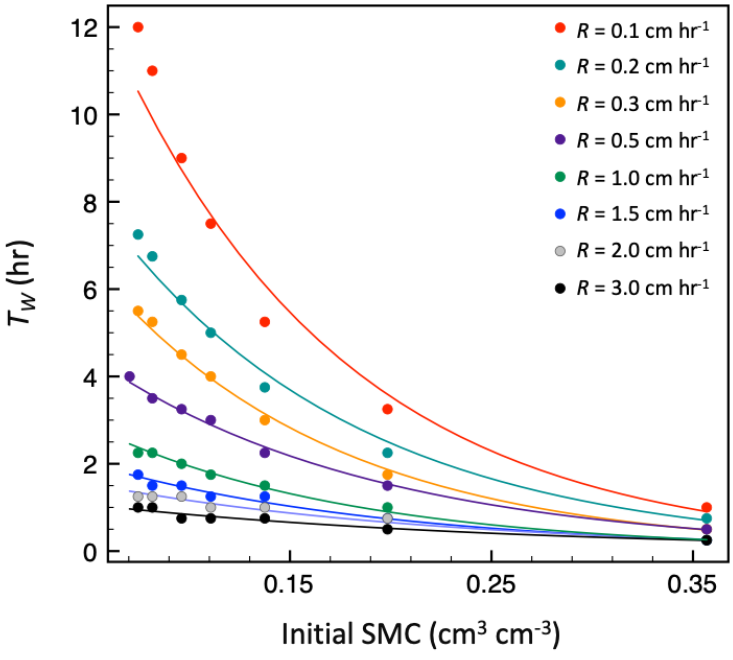

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473 Figure 3: Initial soil moisture content (SMC) versus time of wetting front arrival (T_w) for a loamy
474 sand soil. Dots are simulated results from HYDUS-1D simulation, and lines are the exponential
475 model given in Eq. 8, fitted for each rainfall rate, R.

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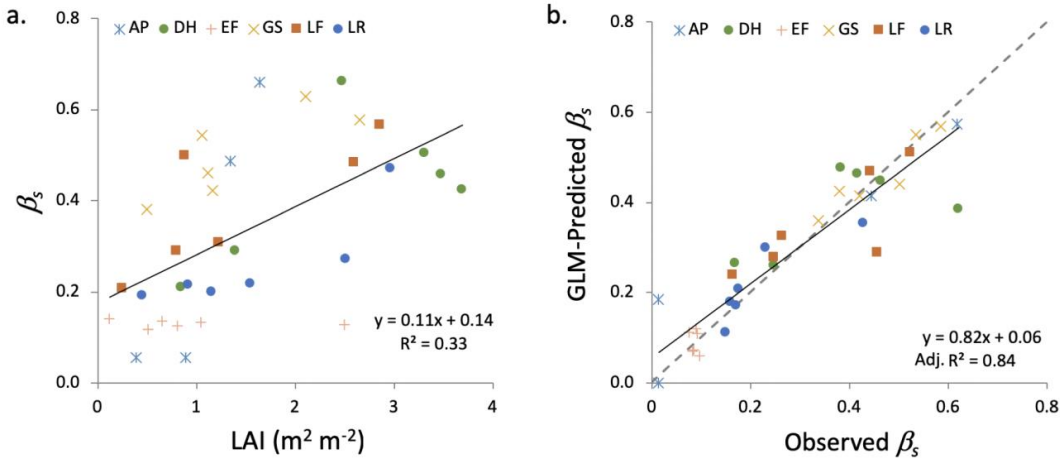

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477 Figure 4. (a) Interception storage capacity (β_s) versus leaf area index (LAI) for all sites and plots.

478 (b) Modeled versus observed β_s using the best GLM, which included % groundcover vegetation

479 and an interaction term between site and LAI. The dashed line is the 1:1 line.

480

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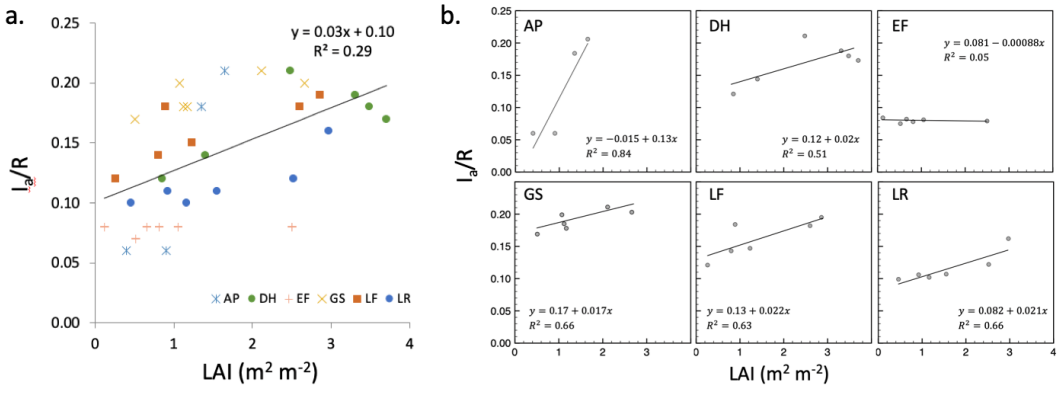

Figure 5. (a) Annual proportion of rainfall that is intercepted (I_a/R) intercepted versus LAI for all sites and plots. (b) Site-specific I_a/R versus LAI relationships. The relationship is generally strong except for the EF site, where the overall storage capacity is small across all values of LAI.

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Table 1. Summary of storage capacity (β_s) and annual interception losses (I_a) for all sites and plots, along with plot characteristics (mean annual precipitation, P; leaf area index, LAI; percent groundcover, %GC; and species). Note that the AP site only had three plots with the data required for the analysis.

Site	Plot	LAI	%GC	Species	β_s (cm)	R^2 (ΔSMC-R)	P (cm)	I_a/P
AP	2	1.65	47.6	SF Slash	0.620	0.31	145.0	0.206
AP	3	0.90	62.8	SF Slash	0.014	0.78	145.0	0.06
AP	4	1.35	49.1	SF Slash	0.445	0.67	145.0	0.184
AP	6	0.40	73.4	Longleaf	0.014	0.57	145.0	0.06
DH	1	0.85	86.2	Loblolly	0.170	0.90	131.5	0.121
DH	2	2.48	51.2	Slash	0.621	0.68	131.5	0.211
DH	3	1.40	39.2	Slash	0.249	0.49	131.5	0.144
DH	4	3.31	35.8	Slash	0.464	0.71	131.5	0.188
DH	5	3.70	27.1	Loblolly	0.383	0.69	131.5	0.173
DH	6	3.48	32.9	Slash	0.418	0.40	131.5	0.18
EF	1	0.12	13.6	Clearcut	0.099	0.93	153.8	0.084
EF	2	1.05	56.9	Slash	0.092	0.96	153.8	0.081
EF	3	2.50	11.8	Sand	0.086	0.93	153.8	0.079
EF	4	0.66	50.9	Slash	0.094	0.92	153.8	0.082
EF	5	0.81	17.9	Sand	0.085	0.96	153.8	0.078
EF	6	0.52	52.0	Longleaf	0.076	0.89	153.8	0.075
GS	1	1.07	67.9	Clearcut	0.502	0.84	132.4	0.199
GS	2	2.66	7.9	Slash	0.535	0.88	132.4	0.203
GS	3	2.11	71.5	Slash	0.587	0.82	132.4	0.211
GS	4	1.12	42.4	Slash	0.421	0.90	132.4	0.185
GS	5	1.17	45.6	Slash	0.382	0.76	132.4	0.178
GS	6	0.51	55.2	Longleaf	0.339	0.78	132.4	0.169
LF	1	0.26	43.5	None	0.166	0.85	136.3	0.121
LF	2	2.86	23.1	Slash	0.525	0.64	136.3	0.195
LF	3	1.23	24.9	Slash	0.266	0.72	136.3	0.147
LF	4	0.80	25.7	Slash	0.248	0.64	136.3	0.143
LF	5	2.60	12.3	Slash	0.443	0.63	136.3	0.182
LF	6	0.89	25.9	Longleaf	0.458	0.69	136.3	0.184
LR	1	0.46	34.0	Clearcut	0.151	0.96	144.5	0.099
LR	2	2.97	38.1	Slash	0.429	0.84	144.5	0.162
LR	3	0.92	47.0	Slash	0.173	0.95	144.5	0.106
LR	4	2.52	26.7	Slash	0.232	0.92	144.5	0.122
LR	5	1.55	28.1	Slash	0.177	0.96	144.5	0.107
LR	6	1.16	35.5	Longleaf	0.160	0.96	144.5	0.102

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492 Table 2. Summary of generalized linear model (GLM) results for interception storage capacity
493 (β_s). LAI is leaf area index, GC is groundcover, and WT is water table (shallow vs. deep). The
494 best model (by AIC) is shown in bold.

Model #	Variable(s)	AIC	R^2
1	LAI	378.1	0.32
2	LAI + site	318.5	0.66
3	LAI * site	255.9	0.83
4	**LAI * site + GC**	**253.1**	**0.84**
5	LAI + WT	338.3	0.55
6	LAI * WT	339.8	0.55
7	LAI * WT + GC	341.8	0.55
8	LAI + WT + GC	340.3	0.55

495