Surface Evaporative Capacitance: How Soil Type and Rainfall Characteristics Affect Global-Scale Surface Evaporation

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Abstract The separation of evapotranspiration (ET) into its surface evaporation (E) and transpiration (T) components remains a challenge despite its importance for linking water and carbon cycles, for water management, and for attribution of hydrologic isotope fractionation. Regional and global estimates of surface evaporation often rely on estimates of ET (e.g., Penman-Monteith) where E is deduced as a residual or as a fraction of potential evaporation. We propose a novel and direct method for estimating E from soil properties considering regional rainfall characteristics and accounting for internal drainage dynamics. A soil-dependent evaporative characteristic length defines an active surface evaporative capacitor depth below which soil water is sheltered from capillary pull to the evaporating surface. A site-specific evaporative capacitor is periodically recharged by rainfall and discharges at rates determined by interplay between internal drainage and surface evaporation. The surface evaporative capacitor concept was tested using field measurements and subsequently applied to generate a global map of climatic surface evaporation. Latitudinal comparisons with estimates from other global models (e.g., Penman-Monteith method modified by Leuning et al., 2008, https://doi.org/10.1029/2007WR006562 [PML]; Moderate Resolution Imaging Spectroradiometer [MODIS]; and Global Land-surface Evaporation: the Amsterdam Methodology [GLEAM]) show good agreement but also point to potential shortcomings of present estimates of surface evaporation. Interestingly, the ratio of surface evaporation (E) to potential evapotranspiration (ET0) is relatively constant across climates, biomes, and soil types with E/ET0 < 0.15 for 60% of all terrestrial surfaces, in agreement with recent studies.

1. Introduction

Evapotranspiration (ET) from land surfaces is a key component of the hydrologic cycle where 60% (\(\approx 66 \times 10^3 \text{ km}^2 \text{/year}\)) of the rainfall evaporates back to the atmosphere (Oki & Kanae, 2006) in a process that consumes nearly 12% of the radiative energy input (38 W/m\(^2\) of the mean incoming 325 W/m\(^2\); Wild et al., 2015). Hydrological and climate models require reliable estimates of components of ET, namely, the components of surface evaporation including canopy interception losses (E) and plant transpiration (T) for linking the water and carbon cycles, for hydrological isotope interpretation, and for water resource management. Global land surface models often estimate ET using variants of the Penman-Monteith approach (Penman, 1948; Monteith, 1965, for an overview see Table 2 in Mueller et al., 2013) with empirically derived expressions for stomatal conductance and vapor transport resistance (Albergel et al., 2012; Leuning et al., 2008 ; Miralles et al., 2016). In some cases, the contribution of soil evaporation to ET was expressed as a simple function of vegetated fraction, with reduced soil evaporation from densely vegetated regions (Leuning et al., 2008; Miralles et al., 2016). Irrespective of model details, the individual contributions of transpiration and soil evaporation are difficult to quantify separately and estimates vary widely across models (Blyth & Harding, 2011; Lawrence et al., 2007; Miralles et al., 2016; Wei et al., 2017), where the fraction of transpiration varies between 24 and 90% of total ET.

We capitalize on recent advances in soil evaporation physics (Lehmann et al., 2008, 2018) to propose a direct approach for estimating surface evaporation that considers local soil physical properties, precipitation characteristics, internal drainage dynamics, and surface resistance to evaporation. Similar approaches for quantifying surface evaporation range in complexity from simple linear scaling of soil evaporation with relative plant available storage (Albergel et al., 2012) to a more direct (albeit empirical) representation of
soil evaporation by Allen et al. (1998), as part of the Food and Agriculture Organization (FAO) report 56 (referred to here as Surface-FAO 56). The method proposed by Allen et al. (1998) relies on practical experience gained from irrigation water management to estimate surface evaporation from a fixed topsoil layer of a prescribed thickness \( Z_e \) (between 0.10 and 0.15 m). Allen et al. (1998) expressed the surface evaporation rate as a function of the mean water content in the topsoil evaporation layer. Soil water storage for water contents \( \theta \) exceeding field capacity \( \theta_{FC} \) are instantaneously removed by internal drainage. The remaining soil water evaporates initially at a constant potential rate (2 to 12 mm/day) followed by linear decrease in evaporation rate to 0 at a water content below 0.5 \( \theta_{WP} \) (with wilting point \( \theta_{WP} \) assigning a minimal water content before plants permanently wilt, often defined as water content at pressure head of minus 150 m). While the approach proposed in this study shares certain elements with the approach of Allen et al. (1998) and Laio et al. (2001) in considering soil water storage within a prescribed active soil depth available for evaporation, we inject soil-dependent processes including evaporation characteristic depth, drainage dynamics based on climate and precipitation patterns, and water content dependent surface resistance. In certain aspects, the proposed approach is similar to the infiltration-exfiltration model of Eagleson (1978) for the dynamics of water content at the soil surface layer but differs in representing surface evaporation.

Evaporation from wet soil surfaces is often marked by an initially high and relatively constant evaporation rate (stage-I evaporation), in which water is supplied to the evaporating surface by capillary flow from deeper soil layers (Lehmann et al., 2008). This initial stage is followed by a falling rate dominated by water vapor diffusion (stage-II; van Brakel, 1980). Stage-I evaporation, identified with the presence of a vaporization plane at the soil surface, may occur at rates close to potential evaporation or at lower rates due to resistance to vapor transport across vapor shells forming around evaporating pores (Shokri, Lehman, & Or, 2008; Shahraeeni et al., 2012) or due to low hydraulic conductivity of the near-surface unsaturated soil layer (Decker et al., 2017; Haghhighi et al., 2013; Lehmann et al., 2018).

The onset of stage-II evaporation follows the disruption of capillary pathways and the gradual migration of the vaporization plane below the surface (Shokri, Lehmann, et al., 2008). Subsequently, evaporation proceeds by vapor diffusion across a progressively thicker dry soil layer at a rate that is inversely proportional to the square root of time (Brutsaert, 2014; Brutsaert & Chen, 1995; Or et al., 2013). The transition from stage-I to stage-II evaporation has been shown to occur at a characteristic depth or evaporative length that is a function of soil properties and to lesser degree depends on evaporation rate (Lehmann et al., 2008, 2018). The transition in evaporation regime is also associated with a critical surface water content \( \theta_{crit} \) at which capillary pathways are disrupted. The application of this characteristic length (and critical water content \( \theta_{crit} \)) defines the size of the evaporative capacitor in this study.

The specific objectives of this study are to (1) introduce the concept of a surface evaporation capacitor (SEC) that builds upon the evaporation characteristic length and incorporation of evaporation-drainage dynamics to establish bounds on surface evaporation, (2) test the SEC for different soils using field measurements from the literature, and (3) develop new estimates of global surface evaporation (including canopy interception) considering climatic and soil data using the SEC and precipitation characteristics. The formalism used to define the SEC and its water content dynamics is presented in section 2. In section 3 we present case studies for evaluating the SEC including a comparison to Surface-FAO 56 and systematic tests of the SEC for different soil types, biomes, and climates. The SEC concept is applied globally considering spatially resolved soil types, precipitation patterns, and vegetation information combined to delineate global climatic surface evaporation maps and associated quantities in section 4. The results are discussed and compared to other global estimates of surface evaporation (and ET) in section 5 followed by a summary and conclusions.

2. The SEC

The proposed SEC is essentially a leaky topsoil layer extending to a depth defined by the soil-specific evaporation characteristic length \( L_C \) (i.e., the limiting depth for capillary extraction of water during stage-I evaporation). The SEC considers internal redistribution of soil water to deeper soil layers, a dynamic process sensitive to soil properties (retention and drainage rates), and to rainfall characteristics (events and amounts). For simplicity, we assign a uniform mean water content \( \theta_{cap} \) to the evaporation capacitor in which the soil water storage \( S = L_C \cdot \theta_{cap} \) is a time-varying quantity. In the following paragraphs, we describe the dynamics of the soil water storage as function of atmospheric conditions (rainfall patterns and drying
periods characterized by potential evaporation rates) and soil properties. The reader may be interested to compare this deterministic approach with the studies on probabilistic modeling of water balance as described in Laio et al. (2001) and Rodriguez-Iturbe et al. (1999).

### 2.1. Soil Evaporation Characteristic Length

The competition between upward capillary forces that supply water to the surface during stage-I evaporation and the opposing gravity and viscous resistance (i.e., near-surface soil hydraulic conductivity) results in a characteristic length at which these forces are in balance. Lehmann et al. (2008) suggested that when the drying front reaches this soil-dependent depth, the hydraulic continuity for capillary flow is interrupted. A critical capillary pressure \( h_{\text{crit}} \) and a critical water content \( \theta_{\text{crit}} \) at the surface are associated with this state, both related via the soil water characteristic curve, parameterized here by the van Genuchten (1980) model:

\[
\frac{\theta - \theta_{\text{res}}}{\theta_{\text{sat}} - \theta_{\text{res}}} = \left(1 + \alpha h \right)^{-m}
\]

with effective water content \( \theta \); residual and saturated water content, \( \theta_{\text{res}} \) and \( \theta_{\text{sat}} \), respectively; and the empirical parameters \( \alpha, n \), and \( m = 1 - 1/n \).

Lehmann et al. (2008) and Assouline and Or (2014) have estimated the critical pressure \( h_{\text{crit}} \) by linearization of equation (1) between air entry pressure and the critical pressure \( h_{\text{crit}} \) value expressed as

\[
h_{\text{crit}} = \frac{1}{\alpha m^{1+m}}
\]

(2a)

\[
\theta_{\text{crit}} = \theta_{\text{res}} + \left(\theta_{\text{sat}} - \theta_{\text{res}}\right) \left[1 + m \frac{\theta_{\text{crit}}}{\theta_{\text{crit}} - \theta_{\text{air}}}\right]
\]

(2b)

Note that in Lehmann et al. (2008) and Assouline and Or (2014) the expressions are given as function of the parameter \( n \) and are slightly more complex than using the relation: \( m = 1 - 1/n \).

The air-entry value \( h_{\text{air}} \) is defined as the lowest capillary pressure where \( \theta(h_{\text{air}}) = 1 \), given as

\[
h_{\text{air}} = \frac{m^{-m} + (m^2-1) \left(1 + \frac{1}{m}\right)^m}{\alpha m}
\]

(3)

The difference between air entry into the largest pore \( h_{\text{air}} \) and the smallest hydraulically connected pore \( h_{\text{crit}} \) defines the maximum length \( L_G \) for capillary flow against gravity to sustain evaporation from the surface:

\[
L_G = h_{\text{crit}} - h_{\text{air}} = \frac{(1-m) \left(1 + \frac{1}{m}\right)^{(1+m)}}{\alpha}
\]

(4)

Figure 1a depicts the linearization procedure for a sandy soil where \( h_{\text{crit}} = 0.65 \) m and \( h_{\text{air}} = 0.14 \) m. The other ingredient in this derivation is the viscous resistance. The capillary pressure gradient that drives upward capillary flow in fine textured soils \( (L_G = h_{\text{crit}} - h_{\text{air}}) \) may exceed several meters, yet the inherently high viscous resistance associated with low hydraulic conductivity shortens the effective characteristic length to a value \( L_C \) (due to viscous dissipation of the capillary gradient). Viscous dissipation is proportional to the flow rate imposed by potential surface evaporation rate \( E_0 \) (see section 2.2) and by the effective hydraulic conductivity \( K_{\text{eff}} \) within the evaporative capacitor (depth of \( L_C \)).

The resulting characteristic evaporation length \( L_G \) combines gravity and viscous forces (Lehmann et al., 2008):

\[
L_G = L_C + \frac{L_C E_0}{K_{\text{eff}}} \rightarrow L_G = \frac{L_G}{1 + \frac{E_0}{K_{\text{eff}}}}.
\]

(5)

An important variable in equation (6) is \( K_{\text{eff}} \) that has been shown in Haghighi et al. (2013) to be related to average surface water content \( \theta_{\text{surf}} \), and by harmonic averaging yields: \( K_{\text{eff}} = 4K(\theta_{\text{surf}}) \). Considering the
limiting case where capillary flow by evaporation reaches to critical water content \( \theta_{\text{crit}} \) (and \( h_{\text{crit}} \)), we may define \( K_{\text{eff}} = 4K(\theta_{\text{crit}}) \) or in terms of \( h_{\text{crit}} \) for the Mualem (1976) model yielding:

\[
K_{\text{eff}} = 4K(\theta_{\text{crit}}) = 4K_s \sqrt{(1 + (\alpha h_{\text{crit}})^n)^{-m}} \left( 1 - \frac{1}{1 + (\alpha h_{\text{crit}})^n} \right)^m \tag{6}
\]

with the saturated hydraulic conductivity \( K_s \). In fine textured soils, \( K_{\text{eff}} \) would often dominate the evaporation resistance, thereby shortening the evaporative characteristic length \( L_C \) well below the value predicted from capillary-gravity considerations alone \( L_G \). The properties that determine the effective evaporation characteristic length \( L_C \) (i.e., the effective hydraulic conductivity \( K_{\text{eff}} \) and the maximum length for capillary flow \( L_G \)) are both functions of the soil texture as parameterized by the van Genuchten parameters \( n \), inverse of capillary pressure \( \alpha \), and the saturated conductivity \( K_s \). In the following sections, we highlight the interplay of this soil-specific characteristic length and soil texture in the partitioning of rainfall into evaporation and leakage to deeper soil layers as described for example in Fernandez-Illescas et al. (2001). For systematic evaluation of different soil textures, we established continuous empirical relations between soil texture parameterized by \( n \) and the evaporation characteristic length (see Lehmann et al., 2008, for details). The relations depicted in Figure 1b show short characteristic lengths for both coarse-textured soils with high value of \( n \) (due to narrow pore size distribution and small capillary gradient) and for fine-textured soils \( n \rightarrow 1 \) (due to high viscous resistance or low hydraulic conductivity \( K_{\text{eff}} \)). The longest evaporative characteristic length values are associated with intermediate soil textures silt or loam.

2.2. Soil Surface Resistance to Evaporation

The soil evaporation rate is jointly controlled by atmospheric conditions and by the soil ability to supply water to meet atmospheric demand. The concept of potential \( ET_0 \) considers a hypothetical evaporation rate that is not hindered by internal soil or plant resistance (other than the aerodynamic resistance at the surface boundary layer) and is thus driven by energy input and atmospheric conditions (wind, temperature, and humidity) near the surface. Potential evapotranspiration \( ET_0 \) is commonly quantified by the combination method of Penman-Monteith (Monteith, 1965; Penman, 1948) or by the energy-based Priestley-Taylor equation (Priestley & Taylor, 1972). The stringent requirements for wind speed information to estimate surface resistance (aerodynamic) by the Penman-Monteith equation compelled us to seek a simpler (albeit empirical) and easy to implement approach, such as the one from Jensen and Haise (1963) for the estimation of
were incorporated based on Ritchie’s (1972) approach. It simply modifies potential evaporation by the values of vegetation LAI with available potential surface evaporation rate $E_0$ given as

$$E_0 = \exp[-0.398 \cdot \text{LAI}] \cdot ET_0$$

Note that the absolute value in the exponent (0.398) was deduced from five different crop types and may be dependent on plant functional type.

We consider two resistances in series to surface evaporation from unsaturated soil with water content $\theta$ (in addition to the aerodynamic resistance): (1) a resistance due to surface drying and increased spacing between evaporation sites (as described in Shahraeeni et al., 2012) and (2) resistance due to diminishing hydraulic conductivity of the drying soil surface (Haghighi et al., 2013). In a recent study, Lehmann et al. (2018) have shown that for most field conditions, soil capillary flow resistance dominates, and the surface drying diffusive resistance is negligible. In the same study, Lehmann et al. (2018) have shown that the effective unsaturated hydraulic conductivity $K_{eff} = 4K(\theta_{crit})$ and the driving capillary pressure gradient $L_C/L_C$ could be combined to yield estimates of soil evaporation rate $E_s$ from drying soil surfaces:

$$E_s = \frac{E_0 \cdot K(\theta) \left[1 + \frac{E_{0i}}{4K(\theta_{crit})}\right]}{E_0 + K(\theta) \left[1 + \frac{E_{0i}}{4K(\theta_{crit})}\right]}$$

with the driving capillary pressure gradient $L_C/L_C$ given in the square brackets. The applicability of the surface evaporation resistance (equation (9)) to estimate soil evaporation has been recently tested (Lehmann et al., 2018), resulting in very good agreement with 34 flux tower sites (details of the bare soil measurements and soil textures are reported in Merlin et al., 2016). We use $E_0$ to distinguish evaporation from soil surfaces from evaporation of canopy intercepted precipitation ($E_i$); both terms are combined to define overall surface evaporation $E = E_s + E_i$. An important insight gleaned from equation (9) is that soil evaporation rate $E_s$ may drop below potential rate $E_0$ even when the soil is very wet due to low hydraulic conductivity coupled with high evaporative demand. Analysis of equation (9) considering a saturated domain (replacing $K_{eff}$ by $K_c$ and considering a unity gradient) shows that even under saturated conditions $E_s/E_0 < 1$ for soils exhibiting $K_c/E_0 < 8$. Regardless of the initial rate, surface evaporation rate eventually transition to a diffusion-controlled stage-II during which the vaporization plane recedes into the soil. In Appendix B we describe how this transition between evaporation stages is implemented in the surface evaporation capacitance model (especially how $E_i$ during stage II is estimated). In subsequent description of water dynamics in the SEC we focus on stage-I evaporation rate as represented by equation (9).

### 2.3. Water Internal Redistribution Dynamics

An important aspect of the SEC concept pertains to the description of how quickly the surface drains following a rainfall even thereby sheltering that water from evaporation. Following the infiltration of rainwater, the water content of the SEC $\theta_{cap}$ may exceed the critical water content $\theta_{crit}$, thus inducing redistribution of the excess water to deeper soil layers. While the total amount of redistributable soil water storage is easily estimated from $L_C(\theta_{cap} - \theta_{crit})$, the effects on evaporative water losses require estimation of internal drainage rates (redistribution). We assume that the water flux $q$ from the bottom of the SEC is driven by gravity at rates described in Assouline and Or (2014) based on Youngs’ (1960) solution. The redistribution water flux $q$ or volume of drained water $Q$ are given as
\[ q = K(\hat{\theta}_{\text{cap}}) \exp \left( -\frac{t \cdot K(\phi_{\text{cap}})}{L_C(\hat{\theta}_{\text{cap}} - \hat{\theta}_{\text{crit}})} \right) \]  

(10a)

\[ Q = L_C(\hat{\theta}_{\text{cap}} - \hat{\theta}_{\text{crit}}) \left[ 1 - \exp \left( -\frac{t \cdot K(\phi_{\text{cap}})}{L_C(\hat{\theta}_{\text{cap}} - \hat{\theta}_{\text{crit}})} \right) \right] \]  

(10b)

with \( t \) marking the elapsed time since onset of redistribution (arbitrarily marked by the end of the rainfall event) and the hydraulic conductivity of the SEC \( K(\hat{\theta}_{\text{cap}}) \):

\[ K_{\text{cap}}(\hat{\theta}_{\text{cap}}) = K_{\text{sat}} \sqrt{\hat{\theta}_{\text{cap}}} \left[ 1 - \left( 1 - \frac{\hat{\theta}_{\text{cap}}}{\hat{\theta}_{\text{sat}}} \right)^m \right]^2 \]  

(11)

with the definition of effective water content \( \Theta \) given in equation (1).

### 2.4. The SEC Evaporation-Drainage Dynamics

Although the equations defining evaporation and redistribution can be expressed in a continuous form as functions of time, in the following, we adapt a discretized form with daily time steps to replace diurnal changes of evaporation rate with a daily averaged value (to conform with measurements and other global models). Following a rainfall event, whenever the resulting SEC water content \( \hat{\theta}_{\text{cap}} > \hat{\theta}_{\text{crit}} \), the excess water content (and storage) is subsequently diminished by concurrent evaporation and gravity-driven redistribution (Figure 2). Note that in Bartlett et al. (2016) a similar figure of storage capacity has been introduced (Figure 1) with the storage expressed as function of dimensionless curve number tabulated for different hydrologic conditions, land use, and soil types. In the SEC approach, the storage is a soil-specific property defined by the characteristic length \( L_C \) (and modified by evaporation and redistribution rates).

The maximum amount of rainwater intercepted by vegetation canopies (\( I \)) is estimated from the leaf area index (LAI) according to the simple relation:
I = LAI \cdot h^* \tag{12}

with maximum water layer thickness on leaves $h^*$. Estimates of this retained water film thickness on leaves varies between 0.2 and 0.6 mm (Dickinson, 1984; Keim et al., 2006; Klaassen et al., 1998; Link et al., 2004); for purposes of the SEC, we have chosen an average value of $h^* = 0.33$ mm. Canopy intercepted water evaporates at potential rate $E_{T0}$ that determines canopy interception evaporative losses $E_i$.

For a rainfall event in which $P > I$, the cumulative excess amount $P - I$ is assumed to infiltrate into the soil evaporative capacitor. If this amount exceeds the maximum storage capacity of the capacitor $((P - I) + S > L_C \cdot \theta_{sat})$, the following amount is removed as runoff $O$:

$$O = (P-I) + S - L_C \cdot \theta_{sat} \tag{13}$$

Summarizing, the soil evaporation capacitor dynamics follow a daily mass balance according to equation (14a) for a rainy day, and equation (14b) for a dry day (no rainfall):

$$P - I = O + Q + \Delta S \tag{14a}$$

$$Q + E = \Delta S \tag{14b}$$

with precipitation $P$, interception $I$, runoff $O$, internal redistribution to deeper soil layers $Q$ (equations (10a) and (10b)), surface evaporation $E$ (computed with equation (9)), and change in soil water storage of the capacitor $\Delta S$. The SEC water balance is formulated for daily time steps as shown in the flowchart in Appendix B.

2.5. Simplifying Assumptions to Define Upper Bound of Surface Evaporation

The storage dynamics of the SEC are simplified to provide an analytical representation: (i) after each rainfall event we instantaneously assign an average water content to the capacitor, assuming that infiltration time is short relative to daily time step, and (ii) the capacitor depth is fixed and equals to the characteristic evaporative length for the soil type. The characteristic length is based on averaged soil textural properties neglecting layering effects (that could be included as shown in Shokri et al., 2010). Despite these simplifications, the SEC reproduces the salient features of the water redistribution and evaporation dynamics as will be demonstrated in various case studies introduced next.

3. Case Studies for SEC Evaluation and Data Sets for Global SEC Mapping

Before application of the SEC concept at the global scale, we conducted several tests of surface evaporation capacitance using field measurements reported in the literature. For comparison with an alternative method for surface evaporation estimation, we evaluated the Surface-FAO 56 approach for the same experimental conditions.

3.1. Experimental Data From Field Studies

We compiled literature data from evaporation field experiments considering different soil types and external conditions. These include information from a sand-filled lysimeter study in Wisconsin (Black et al., 1969) where drainage information (in addition to surface evaporation) have been provided and enable evaluation of the redistribution flux of the SEC. Black et al. (1969) reported several rainfall events during 70 days (the hydraulic functions are reported in Figures 1 and 2 of Black et al., 1969). Another data set was based on measurements reported in Wallace et al. (1999) for a sandy clay soil in Kenya. This data set provides information on evaporation dynamics from finer textured soils to contrast with results from the coarse sandy soil reported in Black et al. (1969). We considered typical hydraulic properties of the same textural class as listed in Carsel and Parrish (1988). Finally, we considered the measurements of Torres and Calera (2010) of evaporation dynamics from a loamy soil after a rainfall event of 38 mm in the region of Castilla-La Mancha (Spain). We used this data set to test soil evaporation rate under dry conditions using equation (9) and for stage-II evaporation dynamics. The soil hydraulic properties were estimated from information in Table 1 of Torres and Calera (2010) and application of the Rosetta pedotransfer functions (Schaap et al., 2001) for the required soil hydrologic parameters.
3.2. Evaluation of Effects of Climate and Soil Type on SEC Estimates

The SEC predictions of surface evaporation under different climatic conditions (precipitation and potential ET) and vegetation covers were evaluated for three sites: Gobi Desert, Queensland, and Bali with different vegetation and climatic conditions (see Figure 3).

These three regions were selected due to their high potential evapotranspiration (cumulative ET₀ > 1 m/year) to emphasize SEC responses to vegetation cover (LAI) and to systematic variations of soil types. Climatic characteristics of the sites are present as mean monthly precipitation and potential ET in Figure 3 (the SEC was driven by daily precipitation values for the decade of 2002–2012). The annual precipitation amounts varied between 0.06 m (Gobi) to 0.8 m (Bali) and 1.5 m (Queensland). Annual ET₀ at these sites was higher than the respective precipitation amounts (ET₀ = 1.2 m for Gobi, 1.8 m for Bali, and 2.3 m for Queensland). To quantify the sensitivity of surface evaporation with respect to soil properties, we carried out simulations for different soil textural classes for each of these study sites using parameters from Carsel and Parrish (1988).

3.3. Data Sets for Deriving SEC-Based Global Estimates of Surface Evaporation

We established a global map of mean surface evaporation (cumulative evaporation per year) at resolution of 1/4 degrees by application of the SEC to each surface pixel using the average local soil information, daily ET₀, precipitation values. The hydraulic properties (represented by the van Genuchten parameters: $n$, $\alpha$, $\theta_{\text{res}}$, $\theta_{\text{sat}}$, and $K_s$) were computed using the pedotransfer functions of Tóth et al. (2015) and soil textural information from the SoilGrids database (Hengl et al., 2017) available at a resolution of 0.25 km. The rainfall information was obtained from the CMORPH precipitation database (Climate Prediction Center morphing method; Joyce et al., 2004) for the period between 2002 and 2012 (available with a spatial resolution of 0.25° and temporal resolution of 3 hr). For the simulations, we summed the 3-h intervals to obtain daily rainfall amounts. We employed monthly values of mean potential evaporation estimated by the method of Jensen and Haise (1963) using incoming shortwave radiation data from CERES EBAF (Clouds and Earth’s Radiant Energy Systems Energy Balanced and Filled; https://ceres.larc.nasa.gov/) and temperature data provided by the Earth System Research Laboratory (https://www.esrl.noaa.gov/). These monthly data were available at a spatial resolution of 0.5°.

The impact of using different approaches for estimating potential ET rates was analyzed in Appendix A (e.g., comparing SEC predictions for Priestley-Taylor and Penman-Monteith). For vegetation canopy interception estimates (and canopy effects on potential evaporation rate) we have used the LAI data from the Copernicus Global Land Service (https://land.copernicus.eu/global/products/lai) that provides LAI data at 1-km spatial resolution for the period from 1998 to present day. Figure 4 depicts a summary of the global maps of the various input parameters presented above. These data sets were selected for convenience, and similar global data sets could be easily used by the SEC concept.
4. Results

The SEC was first evaluated using small-scale field experimental data sets from the literature (section 4.1). We also performed a theoretical evaluation of the response of the SEC for three different climatic regions, where we systematically examined effects of soil type on surface evaporation for different climatic inputs (section 4.2). Finally, we applied the SEC globally to derive soil and climatic based estimates of surface evaporation and derived several other quantities of interest (section 4.3).

4.1. Evaluation of the SEC Concept Using Data From Field Experiments

Results from bare sandy soil (no vegetation) evaporation reported by Black et al. (1969) were used to evaluate SEC estimates as depicted Figure 5a for cumulative evaporation. The SEC estimates captured 60% of the measured 65 mm of evaporation during the 70 days of the experiment. Although details of evaporation dynamics following rainfall events exhibited some discrepancies between model and measurements, the SEC evaporative losses per event were in good agreement with observations. In the results in Figure 5a, we estimated that soil evaporation during stage II contributed about 75% of total evaporative loss. This observation suggests that despite large differences in rates between stages I and II (Lehmann et al., 2008; Shokri et al., 2009) the contribution of stage II becomes significant for extended periods between rainfall events. For the data in Figure 5a, the duration of stage II was 10 times that of stage I (with average evaporation rates of 2.7 and 0.9 mm per day for stage I and stage II, respectively).

Figure 5a also shows comparisons between measured and estimated internal drainage (redistribution) Q. We note that Black et al. (1969) reported measurements of drainage flux from the bottom of their 1.5-m lysimeter, whereas the SEC estimates the redistribution flux from the bottom of the SEC at a depth of $L_C = 0.35$ m. This depth difference could explain the rapid and more abrupt redistribution rates predicted by the SEC in Figure 5a relative to lysimeter measurements (due to the longer time required to reach a depth of 1.5 m).

In Figure 5b we present an evaluation of the Surface FAO 56 approach (Allen et al., 1998) and associated internal drainage. The initial surface evaporation rates were overestimated by the Surface FAO 56 probably due to the assumption of constant stage-I evaporation rate at potential rate ($ET_0$) irrespective of soil properties (e.g., viscous resistance of capillary flow). The evaporation fluxes from the sand were similar to the SEC approach, but the redistribution rates were underestimated due to differences in the critical water contents for onset of drainage and absence of drainage dynamics (as manifested more clearly for fine-textured soils). The experiments of Wallace et al. (1999) for evaporation from sandy clay enable evaluation of the SEC (and Surface FAO 56) for a fine-textured soil as shown in Figure 5c. Unlike the estimates for sand, the incomplete
account of water redistribution dynamics and ignoring viscous limitations to capillary flow in the sandy clay result in overestimation by the Surface FAO 56 approach, whereas the SEC estimates were in a closer agreement with measurements. Finally, the experiments reported by Torres and Calera (2010) for loam soil shown in Figure 5d further highlight limitations of the Surface FAO 56 relative to the SEC approach. Even with complete surface wetting by rain, the ensuing soil evaporation does not occur at potential rate as assumed in Surface FAO 56; it gradually decreases due to surface resistance as suggested by equation (9) and implemented in the SEC (see also Lehmann et al., 2018).

4.2. Effects of Climatic Conditions and Soil Types on Surface Evaporation

The rainfall amounts and intervals that define discrete capacitor recharge events and the potential evaporation rates that determine capacitor depletion rates vary considerably across climatic zones (Figure 4) and are expected to affect surface evaporation. To evaluate the potential role (if any) of soil type under different climatic conditions, we evaluated surface evaporation dynamics predicted by the SEC for three soil textural classes (sand, loam, and clay) under three climatic scenarios (regions). We simulated a period of 11 years, starting in all cases with a soil at its own critical water content. For the desert soil (Gobi), it took more than 2 years to reach its own typical (climatic) dry water content values; we thus analyzed the SEC performance for the last 8 years only as shown in Figure 6.

Despite large differences in cumulative rainfall amounts (for the considered period) from 0.3 m in the Gobi desert to more than 13 m in Queensland, the differences in predicted surface evaporation were relatively
small (between 0.3 m in Gobi and about 1 to 2 m in Bali and Queensland). We note that the three regions share high potential ET$_0$ (>8 m for the period of 8 years) with surface evaporation representing only a small fraction of ET$_0$. In Queensland and Gobi the surface evaporation was mainly bare soil evaporation; however, for Bali the evaporation from canopy interception contributed 40% of total surface evaporation. The reasons for the relatively small surface evaporation are different for the three regions: in the Gobi desert, low rainfall amounts limit surface evaporation (Figure 6). For Bali, the high vegetation cover intercepts a fraction of the rainfall and simultaneously reduces the potential evaporation rate acting at the soil surface by nearly 40% (i.e., without vegetation shading effect, surface evaporation would be 40% higher). Another important process for the resulting low surface evaporation in Bali and Queensland is the role of internal drainage that removes nearly 50–80% of the infiltrated rainwater to deeper water layers and thus sheltering this fraction of soil water from direct evaporation (noting however that plant roots may intercept some of the redistributed soil water; a process not considered in the SEC that quantifies nontranspiration surface evaporation only).

Considering the effects of soil type, in the three climatic regions we find that surface evaporation is highest for loamy soils (that have the highest $L_C$ or deepest evaporation capacitor; see Figure 1b). Soil type effect on
Surface evaporation was highest for Queensland with evaporation losses for loam almost 100% higher than losses from a clay soil. Soil type effects were smaller for the other two climatic regions due to the limited rainfall in Gobi where nearly all rainfall evaporates regardless of soil type, and the dominance of canopy interception and attenuation of radiation in Bali that reduces the role of surface evaporation capacitance.

For a more systematic evaluation of soil type effects on surface evaporation, we calculated SEC dynamics for the soil textural classes reported in Carsel and Parrish (1988) with values $n$ ranging between 1.1 (clay) and 2.7 (sand). In addition, we systematically varied related soil hydraulic properties ($\alpha$, $K_s$) as shown in Lehmann et al. (2008) using $\alpha = 8.7(n - 1)$ (units of 1/m) and $K_s = 0.0077n^{7.35}$ (m/day). The mean annual surface evaporation from different soil types for the three regional climates are depicted in Figure 7, confirming that soil type had only a small influence for regions limited by low rainfall amounts (Gobi). The results for Bali show higher influence of soil type; however, the largest effect was manifested for Queensland conditions. The ratio between surface evaporation $E$ and potential evapotranspiration $ET_0$ is relatively small ranging between 0.05 and 0.15 across soil textures and climatic regions. The highest $E/ET_0$ values were found for Bali where canopy interception contributed about 40% to total surface evaporation. We will revisit the $E/ET_0$ ratio and associated regional climatic influences in the context of global surface evaporation estimates analyzed next.

4.3. Spatially Resolved Global Estimates of Surface Evaporation

The SEC model was applied globally at 1/4 degree resolution at daily time steps for a period of 11 years between 2002 and 2012. The mean annual values of surface evaporation and derived quantities are summarized in four global maps in Figure 8 (note that surface evaporation $E$ combines soil evaporation $E_s$ and canopy interception evaporation $E_i$). The maps reveal that estimated surface evaporation values are relatively small for high latitudes (due to low potential ET) and for low rainfall amounts (in arid regions).

In general, as expected, soil evaporation is highest for regions with warm and wet climate and surfaces with low LAI. Evaporation of canopy interception is high for warm regions with high LAI (i.e., tropical rain forests). A comparison of the maps in Figures 4 and 8 shows that surface evaporation is nearly an order of magnitude smaller than either rainfall or potential ET. In dry regions, most of the precipitation evaporates to the atmosphere with the ratio of annual surface evaporation $E$ to precipitation $P$ close to 1. The sheltering effect of internal drainage to deeper soil layers becomes important in regions with high precipitation. The ratio of surface evaporation to potential ET is relatively small with a global median value $E/ET_0 = 0.13$ (for the soil component only without canopy interception, the value is slightly smaller:$E_s/ET_0 = 0.11$). The global maps of climatic annual surface evaporation $E$ and the $E/ET_0$ ratio shown in Figure 8 are consistent with the results from the regional case studies (section 4.2) showing relatively low surface evaporation irrespective of soil type, in other words, climatic conditions dominate the process.
Interestingly, we find that for about 60% of the terrestrial surfaces, the values of $E/ET_0 \leq 0.15$, similar to recent findings of Fatichi and Pappas (2017) obtained for 60 Fluxnet sites. The temperature color scheme in Figure 9 suggests that regions exhibiting large $E/ET_0$ ratios are typically cold, often with low potential rate $ET_0$ and low precipitation. It is for conditions of low $ET_0$ that surface viscous resistance due to capillary flow (or hydraulic conductivity limitations; Lehmann et al., 2018) is smaller and thus could potentially permit a larger surface evaporation fraction $E/ET_0$ to be lost. Additionally, for regions with low precipitation amounts, the water sheltering mechanism of redistribution is expected to be smaller; hence, more of the rainfall water remains accessible to surface evaporation.

5. Discussion

In the following we evaluate predictions by the SEC model with established global models and discuss the potential of well-constrained surface evaporation in improving estimation of actual ET and interpretation of observed isotope fractionation trends.


To compare the surface evaporation calculated with SEC with other established global models, we have selected the data from the PML (Penman-Monteith method modified by Leuning et al., 2008) and GLEAM model (Global Land-surface Evaporation: the Amsterdam Methodology). For the PML model we used data presented in Zhang et al. (2016; downloaded from https://doi.org/10.4225/08/5719A5C48DB85). The PML model estimates soil evaporation rate at time $t$ as fraction of potential rate $E_0$ (not $ET_0$, due to radiation sheltering by vegetation canopy) as determined by precipitation and potential rate during 32 days before and after time $t$ (Zhang et al., 2010). The GLEAM considers remote sensing data to estimate surface water content. Soil hydrological processes are simulated using a multilayer bucket model. Because GLEAM assigns all evaporative losses to transpiration when vegetation is present (i.e., it does not distinguish between soil evaporation and transpiration), information from pixels with bare soil only is provided as soil evaporation product by GLEAM. This assumption limits the comparison of SEC and GLEAM for bare soil and introduces a bias in GLEAM estimates of this component as discussed in Appendix C. Thus, we focus in the main text on a comparison between SEC and PML.

The soil evaporation computed with SEC model predicts higher $E_s$ values near the equator (Figure 10a). The PML predicts low soil evaporation in most tropical areas, possibly due to the sheltering functions of canopy...
affecting radiation reaching the soil surface. The description of the interception evaporation (Figure 10b) is more detailed in GLEAM and PML (describing different types of canopy cover and using monthly data of canopy cover), but the resulting interception evaporation values are similar in order of magnitude. Note that PML model describes two additional local peaks (Mexico and Southern Asia on Northern Hemisphere and New Zealand and Argentina in Southern Hemisphere). The total surface evaporation (sum of soil evaporation and interception evaporation) is presented in Figure 10c. The values for SEC are higher around equator due to the higher soil evaporation compared to PML. The values for SEC and PML are higher than GLEAM due to the attribution of surface evaporation by GLEAM to bare soil only (with the implications discussed in Appendix C).

5.2. A Speculative Estimate of Actual ET

Recent studies show that the ratio of $T/ET$ is remarkably constant across many regions and biomes with a value about $T/ET \approx 0.7$ (Fatichi & Pappas, 2017; Schlesinger & Jasechko, 2014; Wang et al., 2014; Wei et al., 2017). Considering the independent estimate of $E$ provided by the SEC, we were intrigued by the possibility of using the constancy of $T/ET$ (for biomes with vegetation at climatic equilibrium, not agricultural regions; Paschalis et al., 2018) to propose an independent and somewhat speculative estimate for climatic actual ET. Based on SEC-derived surface evaporation $E$, we express actual ET as $ET = E/(1 - T/ET)$ considering $T = ET - E$. In Figure 11a we present latitudinal values of the climatic annual values of $E$ and ET.
We then compare this simple estimate of climatic ET with other global products based on detailed models of transpiration (MODIS; Mu et al., 2011; ERA-interim, Berrisford et al., 2011; and the PML model; Zhang et al., 2016). Despite the speculative nature of using \( T/ET \sim 0.7 \), the ET values by the SEC approximation are in surprising agreement with the detailed ERA-interim product and in reasonable agreement with other well-established global estimates (Figures 11b and 11c). For clarity, we are not plotting the MODIS-based ET; however, the latitudinal values and the global map follow closely the ET values by the PML model. The global mean annual ET calculated with the SEC was 515 mm/year and is similar to estimates provided in the review of Mueller et al. (2013) that evaluated various ET models and obtained average ET = 493 mm/year.

5.3. Potential Applications for Interpreting Hydrologic Isotope Fractionation

The focus of this study was on quantifying surface evaporation, yet, the SEC concept offers a window into improving estimates of hydrologic isotope fractionation associated primarily with isotope enrichment during surface evaporation (but not during transpiration). Specifically, within the simplifying assumptions of the SEC, we proposed an estimate for the ratio \( Q/P \) that could serve (with certain assumptions) as an independent metric for hydrologic isotope fractionation potential due to soil evaporation (Allison & Hughes, 1983; Good et al., 2015; Schlesinger & Jasechko, 2014). We envision a complete accounting of the convection of the fractionated soil water amount left in the evaporation capacitor during the previous cycle of rainfall-evaporation by the internal drainage \( Q \) yielding a net flux of enriched water of different levels. Provisions could be made to reset the fractionation when the surface layer dries completely and the isotope enriched water evaporates back to the atmosphere. To apply the SEC concept to estimating isotope fractionation, the effect of instantaneous soil water mixing assumption following rainfall (section 2.5) must be revisited. For the purposes of this study, we have not yet pursued the potential of this metric; however, considering the dominance of soil evaporation in hydrologic isotope enrichment, we expect that processes encapsulated by the SEC with provisions for internal drainage flux opens the door for a mechanistic estimation and attribution of these important hydrologic tracers. Another potential application of the SEC is for estimation of near surface soil water content dynamics as input to various eco-hydrological models, due to the intrinsic sensitivity to effects of soil texture on drainage dynamics and evaporative losses (Fernandez-Illescas et al., 2001; Laio et al., 2001; Rodriguez-Iiturbe et al., 1999).

6. Summary

Reliable estimates of surface evaporation are important for the separation of ET to its components and links with the carbon cycle. The study proposes a simple and direct model for surface evaporation estimated (termed the surface evaporation capacitance SEC). The SEC relies on recent findings of soil evaporation dynamics and surface resistance as applied to estimation of spatially resolved surface evaporation at global scale. The SEC approach defines a soil specific maximum depth for capillary-driven evaporation (stage-I). Water redistributed by gravity flow to deeper soil layers become largely sheltered from evaporation (with provisions for plant interception not considered here, and stage-II diffusion that becomes important with infrequent rainfall and is included here). The SEC concept combines soil evaporation, canopy interception, internal redistribution, and runoff, with soil-mediated dynamics and surface resistance. The SEC has been tested for different soil types and climatic regions, and the main findings are summarized as follows:

1. Evaporation occurs in different stages with high (but water content dependent) stage-I rates followed by lower and diffusion limited stage-II rates.
2. Evaporation losses are higher for soils with intermediate textures (due to the larger characteristic length) and smaller for coarse (narrow pore size distribution) and fine-textured media (limiting viscous losses).
3. The soil evaporation losses for three different regions with high potential ET (>1 m per year) were small (~0.2 m per year) for different reasons: small precipitation rates in Gobi desert, canopy reducing the incoming radiation in Bali, and large redistribution fluxes to deeper soil layers in Queensland.
4. The ratio between surface evaporation and potential ET was small at global scale with 60% of terrestrial area with ratios <15%.
5. Higher ratios (>15%) were found in cold and dry regions.
Latitudinal comparisons of surface evaporation by SEC were in reasonable agreement with well-established global models (PML and GLEAM); however, \( \bar{E}_s \) estimates by SEC reveal potential shortcomings of assumptions regarding \( \bar{E}_s \) representation in GLEAM as discussed in Appendix C. We explored the potential of using the independent estimates of \( \bar{E} \) and the relative constancy of the ratio of \( T/ET \) to propose a speculative constraint on actual ET. The estimates were in reasonable agreement with more detailed global models (ERA-interim, MODIS, and PML). The consideration of hydrologic processes in SEC such as internal drainage offers the possibility of delineating regions with respect to their hydrologic isotope fractionation potential. The independent estimates of surface evaporation by the SEC approach link soil properties and rainfall patterns and yield a new avenue for the separation of \( \bar{E} \) and \( T \) and better constraints on the water and carbon cycles.

Appendix A: Sensitivity of Soil Evaporation Rates on Potential ET Rates

Evaporation dynamics of the SEC are not only determined by the simulated soil physical processes (depth equal to the characteristic length \( L_{\text{GC}} \), transition of evaporation stages defined by a critical water content \( \theta_{\text{crit}} \), viscous resistance of capillary flow, and redistribution to deeper soil layers) but also by the choice of the approach to determine the potential ET rate \( ET_0 \). For the SEC model we used the relatively simple approach of Jensen and Haise (1963) to compute potential ET rate. In this appendix we explore how the surface evaporation estimates computed with SEC change with the choice of potential rate. Usually, the approach of Penman (1948) and Monteith (1965) using the FAO formulation proposed by Allen et al. (1998) is applied to estimate potential rate. However, in the GLEAM model that was used to test our SEC approach (Figure 10 of the main text and Appendix C), the Priestley and Taylor (1972) approach was chosen. For completeness, the three equations (Jensen Haise, Penman Monteith, and Priestley Taylor) are given below in equations (A1a), (A1b), and (A1c), respectively.

\[
ET_0 = \frac{R_S}{L_{\text{Heat}} \rho_w} [0.025(T_{\text{air}} - 273.15) + 0.08] \quad (A1a)
\]

\[
ET_0 = \left( \frac{\beta}{\Delta} \right) \left( R_N - G \right) + \frac{p_{\text{sat}}(T_{\text{air}}) - p_{\text{sat}}(T_{\text{m}})}{L_{\text{Heat}} \rho_w (\beta + 1)} \quad (A1b)
\]

\[
ET_0 = \beta \frac{R_N - G}{L_{\text{Heat}} \rho_w} \quad (A1c)
\]

with latent heat of evaporation \( L_{\text{Heat}} \), water density \( \rho_w \), incoming solar radiation \( R_S \), air temperature \( T_{\text{air}} \) (°K), aerodynamic resistance term \( R_{\text{AH}} \) (see details in Merlin et al., 2016, and Allen et al., 1998), net radiation \( R_N \), soil heat flux \( G \), vapor pressure at saturation \( p_{\text{sat}} \), and in air \( p_{\text{air}} \), psychrometric constant \( \gamma \) and the derivative of the saturated vapor pressure with temperature \( \Delta \). The coefficient \( \beta \) was set to \( \beta = 1.26 \frac{\Delta}{(\Delta + \gamma)} \) in Priestley and Taylor (1972). Global estimates of potential ET rates were compared in several studies (Maes et al., 2018; Trambauer et al., 2014), and a systematic comparison is not in the focus here. Instead we will compare the potential evaporation rate \( ET_0 \) that we computed at the global scale, with two other products providing global maps: the potential evaporation data provided by GLEAM model (Miralles, De Jeu, et al., 2011, Miralles, Holmes, et al., 2011) that is based on the Priestley Taylor approach and the CRU data (Climatic Research Unit) that were used by Harris et al. (2014) to compute the potential rate with the Penman Monteith approach. Note that the differences between GLEAM and CRU model differ not only with the method to compute \( ET_0 \) ( Priestley Taylor vs. Penman Monteith) but also with respect to the underlying satellite data. The results are shown in Figure A1a and highlight that the Jensen Haise approach is similar to the CRU data (Penman Monteith) but differs for high rates from the GLEAM data ( Priestley Taylor). Interestingly, these large difference in \( ET_0 \) do not have a very high effect on the actual surface evaporation rate computed with the SEC approach as shown in Figure A1b. The ratio between SEC computations using Priestley Taylor (from GLEAM) with SEC computations using Jensen Haise is about 0.95. To test the sensitivity of surface evaporation on potential ET rate in a more systematic way, we changed the potential rate...
computed with Jensen Haise by 50%. The results are shown as a function of latitude in Figure A1b. The effect is minor in higher latitudes and larger for regions with large precipitation amounts (latitudinal bands between −20 and 20°). In average, the surface evaporation rate as a function of latitude changes from 90 to 110% by changing the potential rate from 50 to 150%. We can thus conclude that the SEC surface evaporation data are relatively independent from the underlying map of potential rate.

Appendix B: Stage II Evaporation

In the main text the description of water dynamics of the soil evaporation capacitance (SEC) was given for conditions with water content \( \theta > \theta_{\text{crit}} \) (with critical water content \( \theta_{\text{crit}} \) when capillary flow pathways become disconnected) because only for these conditions water is percolating to deeper soil layers and evaporation rates are high and close to potential evaporation rate \( E_0 \) (that may be smaller than the potential ET rate \( ET_0 \) due to attenuation by plant cover). However, for soils in arid regions (and fine textured soils with limiting capillary flow) the time window with stage-I evaporation rates close to the potential rate is relatively narrow and water evaporating at stage-II evaporation has to be quantified as well (in Figure B1 the simulation scheme including first and second stages of evaporation is shown). For coarse-textured media (i.e., for soils without flow limitations by capillary flow) stage-I evaporation ends when the drying front depth (interface between wet and partially dry soil) is at depth \( L_C \) and the capillary-driven water supply pathways disconnect from the surface with the vaporization plane receding in the porous medium. In stage-II the evaporation rate is controlled by diffusion from the receding vaporization rate through a dry porous medium (we set this minimum water content as \( \theta_{\text{res}}/2 \) with residual water content \( \theta_{\text{res}} \)) with a diffusion coefficient \( D_{\text{soil}} \) of the porous medium set according to Moldrup et al. (2000) as

\[
D_{\text{soil}} = D_{\text{bin}} \left( \frac{\theta_{\text{sat}} - \theta_{\text{res}}/2}{\theta_{\text{sat}}} \right)^{10/3} \tag{B1}
\]

with the binary diffusion coefficient of water vapor \( D_{\text{bin}} \) and saturated water content \( \theta_{\text{sat}} \). The dynamics of the receding vaporization plane and its effect on drying rate as a function of time \( t \) was defined in Or et al. (2013). By integrating the secondary stage evaporation rate over time the cumulative evaporation loss since onset of stage-II at time \( t_{II} \) equals
with change \( \Delta \theta \) in water content across the vaporization plane, evaporation rate \( E_s \), and front depth \( \xi \) at onset of stage-II evaporation. Following Shokri and Or (2011) the initial depth of the vaporization plane \( \xi \) in equation (B2) is of the order of 5 to 15 mm independent of soil type, yielding evaporation rates at onset of stage-II \( E_s \) between 1.0 and 2.0 mm/day. We considered the initial stage-II value to be the minimum between 1.0 mm/day and \( \frac{E_0}{2} \) with \( \xi = 10 \) mm at onset of stage-II. Usually at the end of stage-I the capacitor is at critical water content \( \theta_{\text{crit}} \) and the change in water content across the vaporization plane is \( \Delta \theta = \theta_{\text{crit}} - \theta_{\text{res}}/2 \). But there are two exceptions: first, for fine-textured media the stage-I evaporation rate drops with decreasing water content due to viscous capillary flow (equation (9) in main text). The soil evaporation rate \( E_s \) can become smaller than diffusion controlled rates for \( \theta > \theta_{\text{crit}} \) and the transition to stage-II occurs earlier; for this case we differentiate between the water volume in the capacitor \( \theta \cdot L_C - \theta_{\text{crit}} \cdot L_C \) that still can drain to deeper layers but evaporation is controlled by secondary stage. A second exception occurs under dry conditions with \( \theta < \theta_{\text{crit}} \) and a vaporization plane below the surface. When rainfall occurs, we assume a fast redistribution within the capacitor resulting in a new homogeneous water content \( \theta \) and vaporization plane at the surface that recedes after rainfall with a change in water content \( \Delta \theta = \theta - \theta_{\text{res}}/2 \) across the vaporization plane. In Figure B1 these rules are integrated in the flowchart of the SEC model. It includes as well the evaporation of water intercepted by canopy.

**Appendix C: SEC and GLEAM Comparison of Bare Soil Evaporation \( E_s \)**

To compare surface evaporation estimates by the SEC with other global-scale estimates, we evaluated results from the GLEAM presented in Figure 5 of Miralles, De Jeu, et al. (2011). Because GLEAM makes no distinction between soil evaporation and transpiration for regions with vegetation cover, a comparison with
Figure C1. Comparison of bare soil evaporation $E_s$ computed with Global Land-surface Evaporation: the Amsterdam Methodology (GLEAM; shown in red) and with the surface evaporative capacitor (SEC) model (black) using different averaging methods. (a) Bare soil evaporation in GLEAM is restricted to certain regions (the inset in (a) is taken from Miralles, De Jeu, et al., 2011, and shows in red regions with bare soil evaporation $E_s > 0$). These areas correspond to regions with low leaf area index LAI < 0.15 (shown more clearly in the inset in (b)). By attributing $E_s$ from these bare soil regions to the total terrestrial area (to compare to Figure 5 of Miralles, De Jeu, et al., 2011), the resulting surface evaporation values are relatively small (these low values also reflect a bias of low vegetation cover associated with dry regions). In (b) $E_s$ estimates with SEC from areas with LAI < 0.15 were averaged over bare soil area only to provide the correct estimates of annual bare soil evaporation. The difference in $E_s$ estimates in (b) highlights the general underestimation of $E_s$ component in GLEAM.

GLEAM is limited to regions with spare vegetation only. As in Miralles, De Jeu, et al. (2011), we present in Figure C1 (and in Figure 10 of the main text) bare soil evaporation ($E_s$) for the different longitudinal bands (a band consists of entire terrestrial area for that latitude). As shown in the inset of Figure C1a, bare soil evaporation computed by Miralles, De Jeu, et al. (2011) is significant only for relatively small regions of the globe. The extent of these bare soil evaporation regions overlaps with regions characterized by low LAI < 0.15 (see inset of Figure C1b).

To compare the estimates of $E_s$ by GLEAM with SEC, we computed soil evaporation from this area. We thus distinguish two types of latitudinal averaging using the SEC: (i) cumulative evaporation from bare soil area divided by the entire terrestrial area as done in GLEAM (Miralles, De Jeu, et al., 2011, Figure 5) and shown in Figure C1a and (ii) cumulative evaporation from bare soil divided by bare soil area only in Figure C1b. The results in Figure C1a illustrate that GLEAM and SEC yield similar estimates for the same assumptions but underestimate the $E_s$ component for the latitude. The $E_s$ estimate is easily corrected by the second type of areal averaging shown in Figure C1b for the SEC. The same averaging of bare soil $E_s$ component in GLEAM would result in similarly corrected $E_s$ values.

**References**


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