Root-Water-Uptake Based upon a New Water Stress Reduction and an Asymptotic Root Distribution Function

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ABSTRACT: A water stress–compensating root-water-uptake module was developed based upon a newly proposed water stress reduction function and an asymptotic root distribution function. The water stress reduction function takes into account both soil water pressure head and soil resistance to water flow. It

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requires only physically based parameters that eliminate the need for empirical calibration. The root-water-uptake module, incorporated into a complete Soil–Vegetation–Atmosphere Transfer (SVAT) simulation model, was tested for a variety of soil, crop, and climatic conditions across Canada. Both the proposed water stress reduction and the asymptotic root distribution function performed similarly to existing ones, with the maximum difference in normalized root-mean-square error (NRMSE) between the new and existing water stress reduction function being 0.6%, and between the asymptotic and an exponential root distribution function being 1.2%. The entire root-water-uptake module worked as well as, or better than, published ones. Because the new module uses fewer empirical parameters, it becomes particularly useful in large-scale modeling applications of land surface, hydrology, and terrestrial ecosystems where such parameters are usually not readily available.

KEYWORDS: Root water uptake model; Soil water simulation; Water stress reduction function; Root distribution; Hydrological modeling

1. Introduction

Root water extraction is a key process in the hydrological cycle in that it largely controls the partitioning of infiltrating rainfall into evaporation, transpiration, and leaching. Through its effect on the partitioning of net radiation between latent and sensible heat, it is also an important determinant of climate. Hydrological, ecosystem, land surface, and crop models all require a quantitative description of water uptake by plant roots (Feddes and Raats 2004). Two approaches, a microscopic and a macroscopic methodology, have been developed to simulate water uptake by plant roots. The microscopic approach simulates water movement toward individual roots (see, e.g., Pagès et al. 1989; Nobel and Alm 1993; Steudle 1994; Doussan et al. 1998; Personne et al. 2003). This approach is physically based, but the dynamics and detailed geometry of the rooting system are difficult to measure (Vrugt et al. 2001b). In contrast, the macroscopic approach treats the root system as an abstract object whose vertical distribution is described throughout the soil profile. Commonly, macroscopic water extraction is represented by a depth-dependent sink term that can be incorporated into the Richards (Richards 1931) soil water flow equation. While this approach does not give a complete insight into the physical processes of root-water-uptake, it only needs the soil and plant parameters that are readily available. Consequently, the use of macroscopic root-water-uptake modules is generally favored in many application-oriented hydrological models.

Feddes et al. (Feddes et al. 1978) proposed a macroscopic root-water-uptake scheme in which potential transpiration is first distributed over the rooted zone and then reduced to actual root-water-uptake by a soil water stress reduction function. Based on this conceptualization, numerous root-water-uptake modules have since been developed. While these modules differ in the way the potential transpiration is distributed and/or potential transpiration is reduced by water stress, they all have the same drawback in that empirical parameters are required to run these modules.

The distribution of potential transpiration has been based either on root distribution only (see, e.g., Feddes et al. 1978; Van Dam et al. 1997; Simunek et al. 1998; Li et al. 1999) or on both root distribution and soil water availability (Li et al. 2001a). Various linear (e.g., Hoogland et al. 1981; Hayhoe and De Jong 1988; Prasad 1988) and nonlinear root distribution functions (e.g., Raats 1974; Jarvis 1989; Tiktak and Bouten 1992; Vrugt et al. 2001a) have been used in the root-water-uptake conceptualization. The linear functions that require only minimal data (i.e., rooting depth) are less accurate than the nonlinear functions that require at least one empirical parameter, obtained through fitting with measured root distribution data. Most root distribution functions are derived from local measurements, usually based on a limited number of crops or vegetation types. Consequently, such modules are of limited use at the regional and global scale, where a root distribution function for various biomes and crops is required.

Three water stress functions, proposed respectively by Feddes et al. (Feddes et al. 1978), Van Genuchten (Van Genuchten 1987), and Tiktak and Bouten (Tiktak and Bouten 1992), are commonly used in the soil–plant–water modeling community. All three functions contain empirically determined reduction point parameters and therefore require calibration and validation if no appropriate values can be found in the literature. For simulations at the regional and global scale such parameter calibration and validation are not only time consuming, given the high spatial variability of soil and vegetation, but often impossible in some areas because of unavailability of measured data.

In view of the limitations of the existing root-water-uptake modules in large-scale applications, a more flexible module will be developed based upon an asymptotic root distribution function (Gale and Grigal 1987) and a newly proposed water stress reduction function. In the latter, the empirical reduction point parameters are not required inputs; instead the water stress reduction is determined by dynamically computed soil water pressure heads, soil water retention curves, atmospheric water demand, and the maximum potential transpiration rate. The asymptotic root distribution function involves only one fitting parameter (β) that is available for various biomes (Jackson et al. 1996; Jackson et al. 1997) and will be available at $0.5° \times 0.5°$ grid scales for regional and global simulations (de Fries and Townshend 1999; de Fries et al. 1999; Feddes et al. 2001). The new root-water-uptake module will be tested against soil water content measurements under a variety of soil, crop, and climate conditions across Canada.

2. Methodology

2.1. Development of a root-water-uptake module

2.1.1. General concepts

Like most root-water-uptake modules, this one follows the macroscopic scheme of Feddes et al. (Feddes et al. 1978); that is, it first distributes the potential plant transpiration throughout the entire rooted zone, and then it reduces this at each depth by a water stress reduction function. Hence, the root water extraction rate [$S(h,z)$], usually called the sink term ($L^3\ L^{-3}\ T^{-1}$) when incorporated into the Richards water flow equation, is expressed as

$$S(h, z) = \alpha(h)S_{\max}(h, z), \tag{1}$$

where α is a dimensionless water stress reduction factor as a function of pressure head h (L), $S_{\max}$ ($L^3\ L^{-3}\ T^{-1}$) is the maximum possible root water extraction rate when soil water is not limiting, and z is the soil depth (L).

Root water uptake modules differ from one another in the way that $\alpha(h)$ and $S_{\max}(h, z)$ are conceptualized. Our proposed water stress function $\alpha(h)$ will be based on Ohm's law analogy of soil–water–atmosphere–plant relations and on an experimental study conducted by Denmead and Shaw (Denmead and Shaw 1962). The conceptualization of $S_{\max}(h, z)$ will be based upon an asymptotic root distribution function that takes into account water stress compensation, that is, the water stress occurring at the densely rooted surface layer can be compensated through increasing water uptake from the sparsely rooted deep layers where more water might be available. It should be noted that many hydrological models, such as the Soil–Water–Atmosphere–Plant Model (SWAP; Van Dam et al. 1997), HYDRUS (Simunek et al. 1998), and SWIF (Tiktak and Bouten 1992), only distribute the potential transpiration into the rooted zone, simply following the root distribution pattern, irregardless of soil water availability. These simple distributions allocate a large portion of potential transpiration to the surface layer even though little water may be available there, while only a small portion is assigned to the deep layers where much more water may be available for uptake. However, this water uptake pattern that follows the root distribution pattern seems to be only appropriate in uniformly wet soil profiles (Olson and Rose 1988), while water compensation is reported to occur when the surface layers are depleted (Arya et al. 1975a; Arya et al. 1975b; Nnyamah and Black 1977; Lai and Katul 2000; Gardner 1983; Gardner 1991; Canadell et al. 1996). Therefore, the distribution of potential transpiration simply following the root distribution pattern is inadequate when dealing with water uptake in a dry period. Similar to a methodology proposed by Li et al. (Li et al. 2001a), which was subsequently tested by Braud et al. (Braud et al. 2005) and recommended for inclusion into large-scale hydrological models due to its robustness, we now distribute the potential transpiration over the rooted zone according to both root distribution and the soil moisture profile. The $S_{\max}(h, z)$ in this study is thus defined as

$$S_{\max}(h, z) = \frac{T_p \alpha(h) f(z)}{\displaystyle\int_0^{d_r} \alpha(h) f(z)\, dz}, \tag{2}$$

where T_p is potential transpiration rate ($L\ T^{-1}$); d_r is the rooting depth (L); and $f(z)$ is a dimensionless specific root fraction at depth z, that is, the root fraction in a unit depth interval. The functions $\alpha(h)$ and $f(z)$ are defined below.

2.1.2. Water stress function $\alpha(h)$

Soil water availability varies with soil water content or soil water pressure head. In this study, the full range of pressure head is divided into three sections: 1) from saturation (pressure head is zero, h_0) to field capacity (h_{fc}); 2) from field capacity to permanent wilting point (h_{pwp}); and 3) below the wilting point.

The soil water content within the pressure head section between h_{fc} and h_{pwp} is usually regarded as the available water content, and hence the water stress reduction function for this section is most important. Employing the analogy of Ohm's law for an electric current, Van den Honert (Van den Honert 1948) described the steady state of water flow as a catenary process through successive sections, expressed as

$$q = -\frac{\Delta h}{R} = -\frac{h_{\mathrm{root}} - h}{R_{\mathrm{soil}}} = -\frac{h_{\mathrm{leaf}} - h_{\mathrm{root}}}{R_{\mathrm{plant}}} = -\frac{h_{\mathrm{leaf}} - h}{R_{\mathrm{soil}} + R_{\mathrm{plant}}}, \tag{3}$$

where h, h_{root}, and h_{leaf} are pressure heads in the soil, at the root surface, and in the leaves, respectively; and R_{soil} and R_{plant} are flow resistances of the soil and the plant. Although steady-state water flow is not common, the equation provides important insights in soil–water–atmosphere–plant relations: the water flux q, or transpiration stream T, is proportional to the pressure head gradient (Δh) and inversely proportional to the resistance (R). Since the water flux through each section of the pathway is of the same magnitude [see Equation (3)], we confine the discussion to the section between soil and root surface. According to Gardner (Gardner 1964), the soil resistance (R_{soil}) is inversely proportional to the hydraulic conductivity (K), expressed as (Hillel et al. 1976)

$$R_{\mathrm{soil}} = \frac{1}{BKL}, \tag{4}$$

where B is an empirical constant, and L is the total length of active roots. Therefore, the water flux (q) is proportional to the hydraulic conductivity (K). Since the hydraulic conductivity (K) is proportional to the soil water pressure head (h_{soil}) or the soil water content (θ), the water flux (q), from the perspective of soil resistance (R_{soil}), is proportional to the soil water pressure head or the soil water content. To simplify the conceptualization, one may use either soil water pressure head or soil water content to reflect the influence of hydraulic conductivity (K) and hence the influence of the soil resistance (R_{soil}) on the water flux (q). From the perspective of soil resistance, it may be conceptually more appropriate to employ soil water content rather than soil pressure head to reflect the water stress impact. For example, as a sandy soil becomes unsaturated, the large pores empty quickly and become nonconductive as the pressure head decreases. Thus, its initially high water content and hydraulic conductivity are decreased sharply. In contrast, in a clay soil with small pores, many of the pores remain full of water and conductive even at a low pressure head, so the soil water content and hydraulic conductivity do not decrease as rapidly and may actually be greater than that of a sandy soil subjected to the same pressure head (Scott 2000). This implies that for a given relatively low pressure head, a sandy soil has a much lower water content and thus much smaller hydraulic conductivity relative to a clay soil. Consequently, the water flux (q) can be viewed as proportional to the potential difference (h-h_{root}) and hence the soil water pressure head (h), but from a perspective of soil resistance, it can also be formulated as proportional to soil water content (θ).

Since the water stress reduction factor (α) is actually a ratio of actual transpiration to potential transpiration, α is proportional to the water flux q (α is 1 when q reaches potential transpiration, while α is 0 when q is 0). Hence, α can be proportional to both soil water pressure head (h_{soil}) and soil water content (θ). To establish the relationship between α and (h_{soil}, θ) for all types of soils, both h_{soil} and θ need to be normalized:

$$H(h) = \frac{h - h_{\text{pwp}}}{h_{\text{fc}} - h_{\text{pwp}}}, \tag{5}$$

$$\Theta(h) = \frac{\theta(h) - \theta(h_{\text{pwp}})}{\theta(h_{\text{fc}}) - \theta(h_{\text{pwp}})}, \tag{6}$$

where H and Θ are the dimensionless normalized soil water pressure head and soil water content, both ranging from 0 at the permanent wilting point to 1 at field capacity. To represent the central tendency of soil water pressure head and soil water content, we use the geometric mean of H and Θ as an index of soil water availability because the properties of H and Θ meet the conditions of using the geometric mean: (i) the key is to determine an "average factor" of H and Θ, (ii) changes in H and Θ occur in a relative fashion, and (iii) both H and Θ are ratios (between 0 to 1). Following the water stress reduction function of Feddes et al. (Feddes et al. 1978) in which α is assumed to be linearly proportional to soil water availability expressed as soil water pressure head, we use the same linear hypothesis, but the soil water availability is expressed as the geometric mean of H and Θ to reflect the roles played by both soil water pressure head and soil resistance. This linear hypothesis is a simplification of the soil–plant–water relationship, and it will be evaluated by using a stand-alone procedure for the water stress reduction function.

A field experiment by Denmead and Shaw (Denmead and Shaw 1962) shows that, when potential transpiration is very high, water stress occurs immediately when the soil water content falls below field capacity; however, when potential transpiration is low, there is no water stress until a reduction point (a threshold) is reached; the smaller the potential transpiration, the lower the threshold. Based on these experimental results, we (i) propose a base function, for which the potential transpiration is very large and water stress starts as soon as the soil water content falls below field capacity, (ii) find the reduction point given any relatively low potential transpiration rate, and (iii) develop a function for any transpiration rate in terms of the base function and the reduction point.

The base occurs when the potential transpiration rate (T_p) equals the maximum potential transpiration rate (T_m), which is defined as the potential transpiration rate at the "peak" of the growing season. Because T_p is determined by vegetation characteristics (e.g., plant phenology) and climatic conditions, T_m is a long-term average plant- and region-specific parameter. Following our aforementioned assumption, the base reduction function is represented as the geometric mean of $H(h)$ and $\Theta(h)$:

$$\alpha(h) = [H(h)\Theta(h)]^{0.5}, \tag{7}$$

where potential transpiration is so large ($T_p = T_m$) that water stress occurs as soon as H or Θ drops below 1.

The second step is to find the reduction point for potential transpiration rates smaller than the maximum potential transpiration ($T_p < T_m$). After analyzing experimental data of Denmead and Shaw (Denmead and Shaw 1962), we find that the reduction point is highly correlated to the relative potential transpiration rate (T_p/T_m). In this experiment, there were five treatments of the potential transpiration rate: 0.64, 0.56, 0.41, 0.20, and 0.14 cm day^{-1}. The 0.64 cm day^{-1} rate is regarded as the maximum potential transpiration rate (i.e., T_m = 0.64 cm day^{-1}) because, in this case, water stress occurs as soon as the soil water content falls below field capacity. Figure 1, showing the relative potential transpiration rate (T_p/T_m) versus the reduction point P_r, expressed as the geometric mean of H and Θ, that is, $(H\Theta)^{0.5}$, indicates a near-linear relationship:

$$P_r = 1.058(T_p/T_m) - 0.036, \tag{8}$$

with an R^2 value of 0.928. Because in Equation (8) the intercept is close to 0 and the slope is close to 1, the reduction point (P_r) can be approximated by the relative potential transpiration rate (T_p/T_m). Therefore, for a given T_p and T_m, the reduction point value equals T_p/T_m in magnitude (i.e., $P_r = T_p/T_m$), implying that water stress occurs as soon as $(H\Theta)^{0.5}$ falls below a value of T_p/T_m.

Based on our proposed reduction function [Equation (7)] and the reduction point obtained using the above described approach, the water stress reduction function can be conceptualized as follows: 1) α remains 1 (no water stress occurs) when $(H\Theta)^{0.5}$ is equal to or greater than the reduction point value (T_p/T_m), 2) α linearly

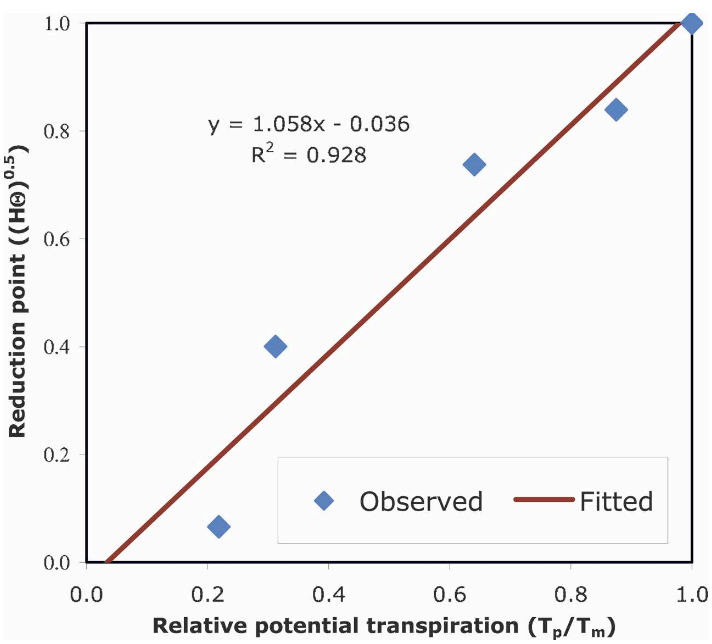

Figure 1. Reduction point ($(H\Theta)^{0.5}$) versus relative potential transpiration rate (T_p/T_m), as obtained from Denmead and Shaw (Denmead and Shaw 1962).

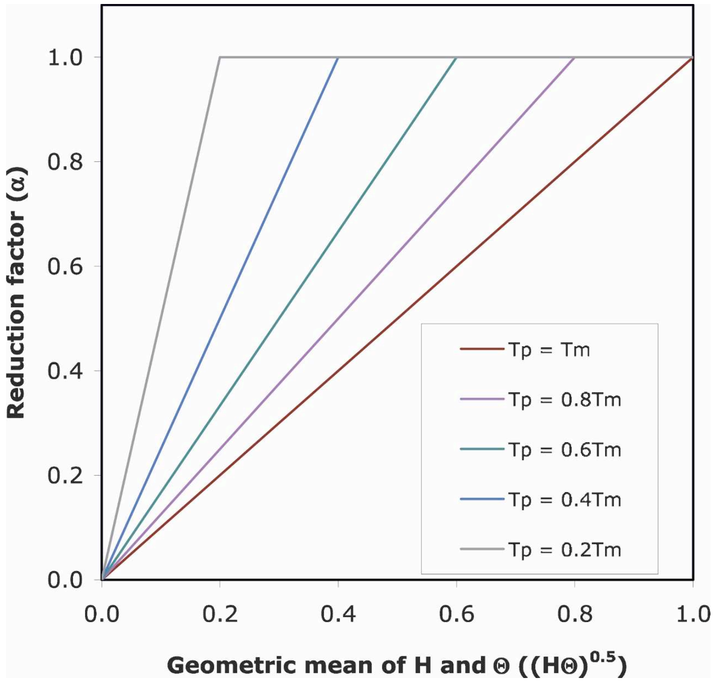

Figure 2. Reduction factor (α) versus geometric mean of H and Θ ($(H\Theta)^{0.5}$) for the soil water pressure head section between field capacity and permanent wilting point.

decreases with decreasing $(H\Theta)^{0.5}$ when $(H\Theta)^{0.5}$ falls below the reduction point, and 3) α drops to 0 when $(H\Theta)^{0.5}$ is decreased to 0. These conceptualizations can be formulated in terms of the base case reduction function [Equation (7)] and the reduction point value (expressed as $T_p/\,T_m$):

$$\alpha(h) = \mathrm{Min}\left(\frac{[H(h)\Theta(h)]^{0.5}}{P_r}, 1\right) = \mathrm{Min}\left(\frac{[H(h)\Theta(h)]^{0.5}}{(T_p/T_m)}, 1\right), \quad \text{for} \quad h_{\mathrm{pwp}} \le h \le h_{\mathrm{fc}}, \tag{9}$$

where Equation (9) is illustrated in Figure 2.

It should be noted that although the derivation of this new function employs the analogy of Ohm's law, it has been simplified by eliminating the need for parameters of soil resistance and potential head at the root surface, and hence this function is conceptually based. Physically based functions (see, e.g., Federer 1979; Guswa et al. 2002) require root density, soil, and plant resistance data, which are not readily available for large-scale simulations.

Next, we develop the water stress reduction function for the pressure head section between saturation (pressure head is h_0) and field capacity (pressure head is h_{fc}). We arbitrarily define the water stress as a sigmoid function of the relative soil water content:

$$\alpha(h) = \frac{1}{1 + e^{(12.254\{0.504 - [\theta(h_0) - \theta(h)]/[\theta(h_0) - \theta(h_{fc})]\})}}, \quad \text{for} \quad h_{fc} < h < 0. \qquad (10)$$

Equation (10) is illustrated in Figure 3 and shows that when soil water content is close to field capacity ($[\theta(h_0) - \theta(h)]/[\theta(h_0) - \theta(h_{fc})] > 0.7$), α is larger than 0.9, but when the soil water content increases such that $0.3 < [\theta(h_0) - \theta(h)]/[\theta(h_0) - \theta(h_{fc})] < 0.7$, α dramatically decreases because of an increase in oxygen deficiency; on the other hand, when the soil water content is near the saturation (i.e., $[\theta(h_0) - \theta(h)]/[\theta(h_0) - \theta(h_{fc]})] < 0.3$), α approaches 0. It can be seen that this stress function is conceptually reasonable, but the two empirical constants (12.254 and 0.504) that determine the thresholds effect of the water stress in the Equation (10) are arbitrary. However, this has little impact on the water uptake in the vadose zone, as confirmed by Hupet et al. (Hupet et al. 2003), who demonstrate that water uptake is insensitive to the parameters dealing with water stress due to oxygen deficiency in the water stress function.

Last, we define the reduction factor as 0 when $h < h_{pwp}$:

$$\alpha(h) = 0, \quad \text{for} \quad h < h_{pwp}. \qquad (11)$$

Equations (9), (10), and (11) are three reduction functions corresponding respectively to the three pressure head sections divided by h_{fc} and h_{pwp}. The parameters required to determine the reduction factor involve the water retention curve, h_{fc}, h_{pwp}, and T_m . It should be noted that physically based hydrological models usually already require as model input the water retention curve, either from measured data

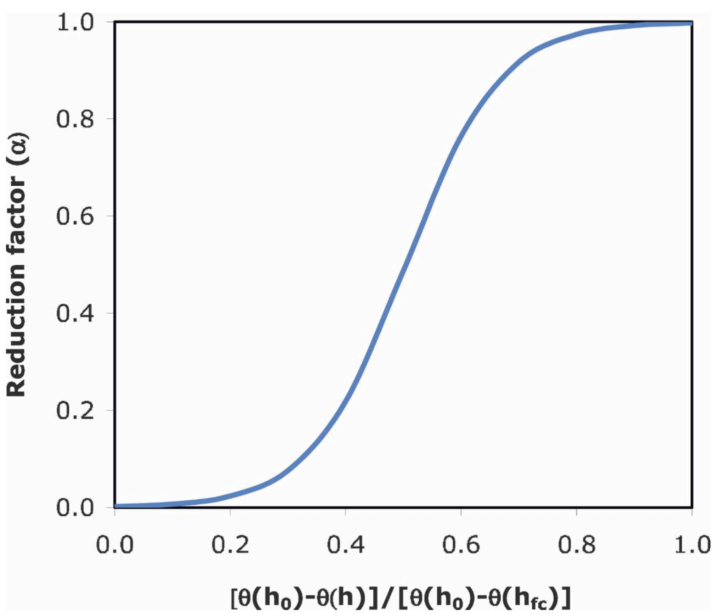

Figure 3. Reduction factor (α) for the soil water pressure head section between saturation and field capacity.

or estimated from a pedotransfer function (see, e.g., Campbell and Norman 1998; Guymon 1994).

2.1.3 Specific root fraction *f(z)*

The specific root fraction $f(z)$, which is the other required component of the root-water-uptake module [Equation (2)], is derived from an asymptotic root distribution proposed by Gale and Grigal (Gale and Grigal 1987):

$$Y = 1 - \beta^z, \tag{12}$$

where Y is the cumulative fraction of roots between the soil surface and depth z (cm). Here, β is an empirical fitting parameter that determines the root distribution with depth. As shown in Figure 4, high β values (e.g., 0.99) give rise to a larger proportion of roots at deeper depths relative to low β values (e.g., 0.90). Use of this asymptotic root distribution has the advantage over other distribution functions that the β values are available for a wide range of natural biomes, including boreal, temperate, and tropical forests; shrubs; grasslands; deserts; and tundra (Jackson et al. 1996; Jackson et al. 1997); furthermore, as aforementioned, a global dataset of β at a spatial resolution of $0.5° \times 0.5°$ is under construction. This function has been successfully applied to the land surface model Integrated Biosphere Simulator (IBIS) across the globe (Foley et al. 1996; Kucharik et al. 2000).

The specific root fraction function $f(z)$ is the derivative of Equation (12) with respect to soil depth z, expressed as

$$f(z) = \frac{dY}{dz} = \frac{d}{dz}(1 - \beta^z) = -\beta^z \ln\beta. \tag{13}$$

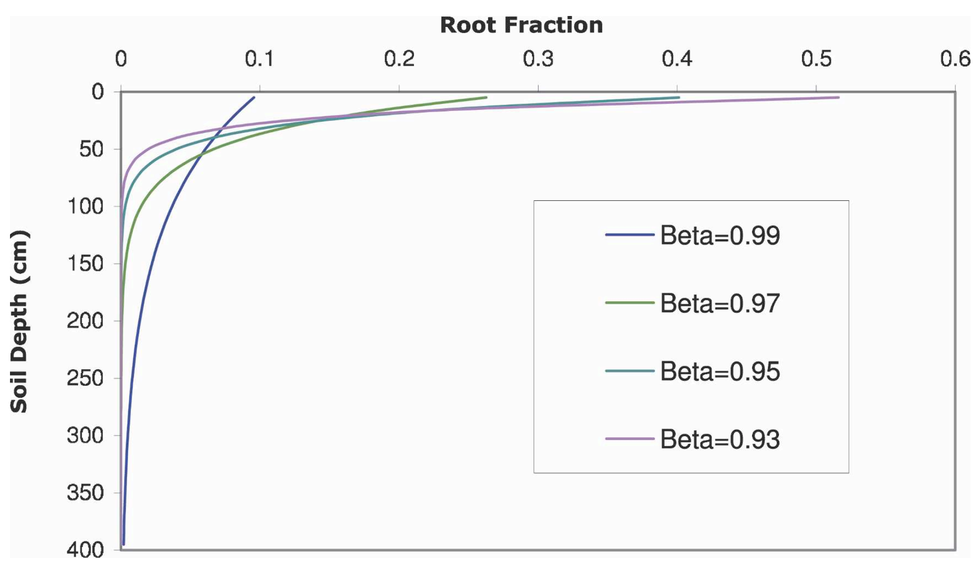

Figure 4. The asymptotic root distribution function with different β values.

In practice, the β value may not be available on a day-to-day basis during the growing season. We propose to estimate the β value from the rooting depth d_r (cm):

$$\beta = 0.01^{\frac{1}{d_r}}. \tag{14}$$

Equation (14) is derived from Equation (12) based on the assumption that the total root fraction from the soil surface to the rooting depth d_r is 0.99 [because Equation (12) is an asymptotic one, the β value cannot be derived if the total root fraction is exactly 1.0].

2.2. Procedures for model performance and sensitivity evaluation

The above-described root-water-uptake module [Equation (2)] was incorporated into the widely used Soil–Plant–Atmosphere Continuum (SPAC) simulation model: SWAP (Van Dam et al. 1997). The new water stress reduction function was compared by itself with the existing reduction function of Feddes et al. (Feddes et al. 1978); and the asymptotic root distribution function was compared by itself with the exponential root distribution function (Dwyer et al. 1988) that was employed in the root-water-uptake modules developed by Li et al. (Li et al. 1999; Li et al. 2001a). Both the performance and the sensitivity of the root-water-uptake module as a whole were quantitatively evaluated against a range of field soil water content measurements under different crops across Canada, including 1) spring wheat (*Triticum aestivum* L.) grown on a silty loam soil at Swift Current, Saskatchewan, Canada (50.3°N, 107.7°W); 2) soybeans (*Glycine max* L.) grown on a sandy loam soil near Simcoe, Ontario, Canada (42.9°N, 80.3°W); and 3) grass grown on a clay soil near Ottawa, Ontario, Canada (45.4°N, 75.7°W). The vastly different soil, crop, and climate conditions at these sites ensure that the model will be properly tested, at least under agricultural conditions in Canada. Parameterizations for the root distribution and the water stress reduction functions for each site are described in Tables 1 and 2, respectively.

2.2.1. Field measurements

2.2.1.1. Swift Current spring wheat site. Full details of the field work were published by Campbell et al. (Campbell et al. 1983); consequently, we provide only a brief summary. Swift Current is located in the semiarid region of western Canada. The experiment was conducted on a Swinton silty loam soil (Orthic Brown Chernozem), with physical properties such as saturated hydraulic conductivity and water retention curve measurements reported by Cameron et al. (Cameron et al. 1978) and Campbell et al. (Campbell et al. 1984). These measurements were used to estimate soil water retention curves (Table 3) following the Van Genuchten (Van Genuchten 1980) methodology.

During the study period from 1967 to 1984, spring wheat was generally seeded in early May and harvested during the second half of August. Soil water contents were measured at emergence, three-leaf, five-leaf, shot blade, soft dough, and harvest time under the wheat phase of a wheat fallow rotation. Gravimetric sampling took place from the 0–15-, 15–30-, 30–60-, 60–90-, and 90–120-cm depths.

The weather data, including daily precipitation, maximum and minimum air

Table 1. Parameterization for the root distribution functions.

Root distribution function	Swift Current, spring wheat	Simcoe, soybeans	Ottawa, grass
Asymptotic	β: estimated from rooting depth (d_r) using Equation (14); d_r: measured value from Campbell et al. (Campbell et al. 1977).	β: estimated from rooting depth (d_r) using Equation (14); d_r: derived from Allmaras et al. (Allmaras et al. (1975).	β: estimated from rooting depth (d_r) using Equation (14); d_r: 40 cm (De Jong et al. 1992).
Exponential	Extinction coefficient b: dynamically estimated from rooting depth (d_r) and the fraction of top 10% of the root zone (Li et al. 2001b); [d_r and root fraction: measured values (Campbell et al. 1977)].	Extinction coefficient b: 0.11 cm^{-1} (Dwyer et al. 1988), fixed through entire growing season.	Extinction coefficient b: 0.11 cm^{-1}, fixed through entire growing season.

temperatures, solar radiation, wind speed, and relative humidity, were measured about 1 km away from the experimental plots. Net radiation was estimated from solar radiation (Feddes et al. 1978) and open water evaporation was calculated with the Penman (Penman 1948) formula. Potential evapotranspiration was then estimated by applying the growing season variable crop coefficients of Doorenbos and Pruitt (Doorenbos and Pruitt 1977). Seasonal changes in soil cover, used to partition potential evapotranspiration into potential transpiration (T_p) and potential evaporation, were taken from Li et al. (Li et al. 2001a). Based upon the 18-yr potential transpiration rates during the peak of the growing season (June and July), T_m was selected to be 0.75 cm day^{-1}.

Each year measured soil water contents at emergence were used as the initial soil water content condition to run the SWAP model. A free draining profile with a unit hydraulic gradient at 120-cm depth was assumed to be the lower boundary condition.

2.2.1.2. Simcoe soybean site. The experimental data used from this site were collected by Bailey and Davies (Bailey and Davies 1981) on a free draining

Table 2. Parameterization for the water stress reduction functions.

Water stress reduction function	Swift Current, spring wheat	Simcoe, soybeans	Ottawa, grass
Proposed function	h_{fc}: −300 cm; h_{pwp}: −30 000 cm (Li et al. 2001a); T_m: 0.75 cm day^{-1}.	h_{fc}: −200 cm; h_{pwp}: −16 000 cm (Clemente et al. 1994); T_m: 0.63 cm day^{-1}.	h_{fc}: −300 cm; h_{pwp}: −16 000 cm (Clemente et al. 1994); T_m: 0.61 cm day^{-1}.
The function of Feddes et al. (1978)	h_1: −10 cm; h_2: −25 cm; h_3^{high}: −500 cm; h_3^{low}: −1000 cm; h_4: −30 000 cm (Li et al. 2001a).	h_1: −10 cm; h_2: −25 cm; h_3^{high}: −500 cm; h_3^{low}: −800 cm; h_4: −16 000 cm (Clemente et al. 1994).	h_1: −10 cm; h_2: −25 cm; h_3^{high}: −500 cm; h_3^{low}: −800 cm; h_4: −16 000 cm (Clemente et al. 1994).

Table 3. Soil hydraulic parameters obtained by fitting Van Genuchten's function (Van Genuchten 1980) to measured water retention and K_{sat} data, where θ_r is residual soil water content; θ_s is saturated soil water content; K_{sat} is saturated hydraulic conductivity; and α and n are empirical parameters.

Depth (cm)	θ_s (cm^3 cm^{-3})	θ_r (cm^3 cm^{-3})	K_{sat} (cm day^{-1})	α (cm^{-1})	n
Swift Current wheat site					
0–15	0.54	0.0177	40.1	0.0386	1.2890
15–30	0.51	0.0140	49.9	0.0228	1.2329
30–60	0.47	0.0126	90.0	0.0268	1.2239
60–90	0.40	0.0102	100.1	0.0073	1.2085
90–120	0.40	0.0000	35.0	0.0062	1.1999
Simcoe soybean site					
0–25	0.25	0.0000	299.5	0.0171	1.3597
25–50	0.45	0.0650	299.5	0.0234	1.9469
50–100	0.38	0.0416	299.5	0.0309	2.0455
Ottawa grass site					
0–25	0.48	0.0000	175.4	0.0401	1.1160
25–50	0.49	0.0000	70.3	0.0249	1.1151
50–80	0.48	0.2551	3.1	0.0030	1.5062
80–120	0.50	0.0000	3.1	0.4329	1.0517

Caledon sandy loam soil (Gray Brown Podzolic) near Simcoe. Soil hydraulic parameters (Table 3) were obtained by fitting Van Genuchten's functions (Van Genuchten 1980) to the water retention and saturated hydraulic conductivity data reported by Clemente et al. (Clemente et al. 1994).

Soybeans were planted on 6 June and harvested during the middle of September 1974. Soil water contents were measured on a daily basis at 0–25-, 25–50-, and 50–100-cm depth using gravimetric sampling (0–25-cm depth) and a neutron probe (>25 cm depths). Precipitation was monitored at the site throughout the growing season, and potential evapotranspiration was estimated using the Priestley and Taylor (Priestley and Taylor 1972) method. Soil cover data were derived using the same procedure as reported by Li et al. (Li et al. 1999). Based upon the potential transpiration rates during June, July, and August, T_m was estimated to be 0.63 cm day^{-1}.

A measured initial soil water condition and a free draining lower boundary condition were used to run the SWAP model for the 1974 growing season.

2.2.1.3. Ottawa grass site. The experiment was conducted in 1982 on a well-structured Rideau clay soil (Gleyed Melanic Brunisol) covered with grass hay. Triplicate undisturbed soil cores were collected and used to measure the water retention curves. Saturated hydraulic conductivity was measured in situ using a Guelph permeameter (Elrick et al. 1989). The water retention curves were obtained by fitting the Van Genuchten (Van Genuchten 1980) function to the measured data (Table 3).

Soil water contents were measured weekly at 7.5-, 22.5-, 40,- and 65-cm depths using time domain reflectometry (TDR) technology. Daily precipitation was measured using an on-site recording rain gauge, and potential evapotranspiration was estimated using the Priestley and Taylor (Priestley and Taylor 1972) formulation. Water table elevation data were obtained by using the average water level in three standpipes, each of which extended to a depth of 300 cm below the soil surface.

Measured soil water contents on 1 May and weekly interpolated groundwater table depths were used as the initial soil water content and the lower boundary condition, respectively. Following De Jong et al. (De Jong et al. 1992), soil cover was also held constant at 90% throughout the growing season. Based upon the potential transpiration rates during June, July, and August, T_m was estimated to be 0.61 cm day^{-1}.

2.2.2. Sensitivity and performance evaluation

Sensitivity coefficients, $\lambda(z, t, b_j)$, for the new root-water-uptake model were calculated using an approach proposed by Simunek and Van Genuchten (Simunek and Van Genuchten 1996):

$$\lambda_h(z, t, b_j) = h(\boldsymbol{b} + \Delta \boldsymbol{b} \boldsymbol{e_j}) - h(\boldsymbol{b}), \tag{15}$$

$$\lambda_\theta\,(z, t, b_j) = \theta(\boldsymbol{b} + \Delta \boldsymbol{b} \boldsymbol{e_j}) - \theta(\boldsymbol{b}), \tag{16}$$

where $\lambda_h(z, t, b_j)$ and $\lambda_\theta\,(z, t, b_j)$ are, respectively, the changes in h (soil water pressure head) and in θ (soil water content) corresponding to a 5% increase in parameter b_j,; e_j is the jth unit vector, and $\Delta b = 0.05b$. To combine all time and space steps, the sensitivity coefficients were represented by

$$\lambda_h^* = \frac{1}{nm} \sum_{t=1}^{n} \sum_{z=1}^{m} \lambda_h(z, t, b_j), \tag{17}$$

$$\lambda_\theta^* = \frac{1}{nm} \sum_{t=1}^{n} \sum_{z=1}^{m} \lambda_\theta(z, t, b_j), \tag{18}$$

where λ_h^* and λ_θ^* are time- and space-combined sensitivity coefficients in terms of h and θ, respectively; n and m are the total number of time and space steps, respectively.

The performance of the SWAP model, with the incorporated proposed root-water-uptake module, was evaluated by comparing simulated soil water contents with measurements, using commonly used error statistics:

$$\mathrm{RE} = \frac{100 \sum_{i=1}^{n} (S_i - M_i)}{n M_{\mathrm{mean}}}, \tag{19}$$

$$\mathrm{NRMSE} = \frac{100 \left[\dfrac{\sum_{i=1}^{n} (S_i - M_i)^2}{n} \right]^{0.5}}{M_{\mathrm{mean}}}, \tag{20}$$

where RE (%) is the average relative error and NRSME (%) is the normalized root-mean-square error. Here, S_i is the simulated soil water content, M_i is the corresponding measured value, and M_{mean} is the average of the measurements.

3. Results and discussion

3.1. Comparison of root distribution functions

The asymptotic root distribution function was compared with the exponential root distribution function (Dwyer et al. 1988), which performed satisfactorily when incorporated into the root-water-uptake modules of Li et al. (Li et al. 1999; Li et al. 2001a). The comparison of the two functions was based on the same water uptake scheme as expressed in Equation (2) and using the water stress function of Feddes et al. (Feddes et al. 1978). None of the two functions yielded consistently lower NRMSE and RE values than the other (Table 4), indicating that the asymptotic root distribution function performed as well as the exponential root distribution function. For both root distribution functions, the NRMSE ranged from 4.4% to 22.4% with a mean less than 16% for all three sites. It should be noted that the NRMSE in the surface layer was greater than 15%, markedly higher than those for the layers below this depth. The relatively large simulation errors for the surface layer were also reported in other modeling studies (Clemente et al. 1994; Li et al. 1999; Li et al. 2001a). They were probably not associated with the root distribution function per se, but, at least in part, might be attributed to the fact that the surface layer was frequently wetted and dried and hence hysteresis might repeatedly occur in the relationship between soil water content and soil water pressure head, as well as between water content (or pressure head) and the hydraulic conductivity. Soil

Table 4. NRMSE and RE of soil water content simulations using the proposed root-water-uptake module with the asymptotic and the exponential root distribution function.

Depth (cm)	NRMSE (%)		RE (%)	
	Asymptotic	Exponential	Asymptotic	Exponential
Swift Current wheat site				
0–15	19.5	19.3	−0.6	1.1
15–30	19.6	19.6	11.2	11.1
30–60	11.4	11.7	−3.8	−4.3
60–90	13.5	12.9	6.6	6.1
90–120	13.3	12.9	4.2	3.7
0–120	8.8	8.7	3.0	2.8
Mean	15.5	15.3	3.5	3.5
Simcoe soybean site				
0–25	22.4	21.4	16.7	14.4
25–50	8.2	9.5	1.8	4.0
50–100	12.6	8.8	−5.2	1.2
0–100	7.6	7.8	1.7	4.9
Mean	14.4	13.2	4.4	6.5
Ottawa grass site				
7.5	15.2	15.0	−4.1	−3.7
22.5	13.0	13.3	−1.6	−2.6
40	6.9	7.1	1.3	0.6
65	11.0	10.8	10.1	9.6
100	5.2	5.2	4.6	4.6
0–100	4.6	4.4	3.7	3.3
Mean	10.3	10.3	2.1	1.7

water hysteresis was not taken into account in the current modeling approach. The absolute value of RE ranged from 0.6% to 16.7% with a mean below 7%, implying that that both root distribution functions did not systematically over- or underestimate the measured soil water contents.

3.2. Comparison of water stress reduction functions

The comparison was made between the proposed water stress function and the function developed by Feddes et al. (Feddes et al. 1978). Since the asymptotic root distribution function performed satisfactorily (see section 3.1.), the simulations for the two water stress functions were made with the asymptotic root distribution function, as incorporated in the water uptake scheme expressed by Equation (2). Both NRMSE and RE for the two water stress functions were similar at all soil layers (Table 5). The difference in mean NRMSE (over all layers) between the two functions was less than 1% and the difference in mean RE was less than 2%. Similar to the comparison of the root distribution functions, the relatively large NRMSE in the layer of 0–30 cm for both water stress functions might be attributed to soil water hysteresis occurring in the surface layers. The good consistency in both the NRMSE and RE between the proposed water stress reduction function and the function of Feddes et al. (Feddes et al. 1978), one of the most commonly used, strongly suggests that the proposed water stress reduction function is behaving properly. Like many other functions, the function of Feddes et al. (Feddes et al. 1978) contains empirical reduction threshold parameters, and its performance

Table 5. NRMSE and RE of soil water content simulations using the proposed and the water stress reduction function of Feddes et al. (Feddes et al. 1978).

Depth (cm)	NRMSE (%) Proposed	NRMSE (%) Feddes et al.	RE(%) Proposed	RE(%) Feddes et al.
Swift Current wheat site				
0–15	18.7	19.5	0.5	−0.6
15–30	20.2	19.6	12.5	11.2
30–60	11.1	11.4	-2.9	−3.8
60–90	14.4	13.5	7.5	6.6
90–120	13.9	13.3	4.8	4.2
0–120	9.3	8.8	3.9	3.0
Mean	15.6	15.5	4.5	3.5
Simcoe soybean site				
0–25	21.7	22.4	12.2	16.7
25–50	8.3	8.2	2.2	1.8
50–100	13.4	12.6	−5.6	−5.2
0–100	8.6	7.6	0.7	1.7
Mean	14.5	14.4	2.9	4.4
Ottawa grass site				
7.5	13.2	15.2	−2.2	−4.1
22.5	15.3	13.0	−0.8	−-1.6
40	9.1	6.9	1.6	1.3
65	11.6	11.0	10.1	10.1
100	5.4	5.2	4.7	4.6
0–100	5.7	4.6	4.1	3.7
Mean	10.9	10.3	2.7	2.1

depends largely on them. The advantage of our water stress reduction function is that it does not require empirical threshold reduction parameters, but nevertheless its performance remains comparable to the Feddes et al. (Feddes et al. 1978) function. Therefore, our function will increase the applicability and feasibility of the root-water-uptake module, in particular for regional and global simulations. The good performance of the proposed water stress function signifies that the linear assumption and geometric mean index are acceptable in this derivation.

3.3. Evaluation of the new root-water-uptake module

In general, the proposed module performed reasonably well, with mean NRMSE values ranging from 10.9% to 15.6% and mean RE values from 2.7% to 4.5% (Table 6). Comparison of this study with the previous study by Li et al. (Li et al. 2001a) indicates that the newly proposed root-water-uptake module performed comparably well or better than existing ones. The module takes water stress compensation into account, which significantly improved the simulation at Swift Current compared to the one without water stress compensation in Li et al. (Li et al. 2001a; in which, the simulations both with and without water compensations were made), where water stress occurring in the surface layer was compensated by increasing water uptake from the deeper and wetter layers. For example, for the Swift Current wheat site, the mean NRMSE and mean RE for the new module are, respectively, 15.6% and 4.5%, which are similar to the error statistical indices of the exponential module with water stress compensation (the mean NRMSE is

Table 6. NRMSE and RE of soil water content simulations using the proposed root-water-uptake module.

Depth (cm)	NRMSE (%)	RE (%)
Swift Current wheat site		
0–15	18.7	0.5
15–30	20.2	12.5
30–60	11.1	−2.9
60–90	14.4	7.5
90–120	13.9	4.8
0–120	9.3	3.9
Mean	15.6	4.5
Simcoe soybean site		
0–25	21.7	12.2
25–50	8.3	2.2
50–100	13.4	−5.6
0–100	8.6	0.7
Mean	14.5	2.9
Ottawa grass site		
7.5	13.2	−2.2
22.5	15.3	−0.8
40	9.1	1.6
65	11.6	10.1
100	5.4	4.7
0–100	5.7	4.1
Mean	10.9	2.7

16.7% and the mean RE is 5.7%), while they are markedly lower than those of the exponential module without water stress compensation (the mean NRMSE is 19.3% and the mean RE is 9.0%; Li et al. 2001a). Evidently, the new module has better or comparable performance but requires fewer empirical parameters than the existing ones.

3.4. Model sensitivity

The sensitivity tests of the module were carried out on four parameters, namely, the soil water pressure heads at field capacity (h_{fc}) and at the permanent wilting point (h_{pwp}), the rooting depth (d_r), and the maximum potential transpiration (T_m). Table 7 shows the average sensitivity of both soil water pressure head (λ_h^*) and soil water content (λ_θ^*) to a 5% increase in these parameters at all three sites. With regard to λ_h^*, the model is sensitive to all four parameters, but no one parameter is consistently more important than the others. For example, in terms of λ_h^*, T_m is the most sensitive parameter for the Swift Current wheat site, while h_{fc} and h_{pwp} are the most sensitive parameters for the Simcoe soybean site and Ottawa grass site, respectively; in terms of λ_θ^*, the most sensitive parameter is T_m for the Swift Current wheat site and the Simcoe soybean site, while it is d_r for the Ottawa grass site. In general, λ_θ^* varies in a narrow range (0.0002 to 0.0014 $cm^3\ cm^{-3}$) among different parameters, while λ_h^* varies in a wider range (5.0098 to 666.8630 cm water). The differences in the sensitivity between different sites may be attributable to the differences in soil type, weather conditions, and crop type.

4. Summary and conclusions

A macroscopic root-water-uptake module that considers water stress compensation was developed based upon a newly proposed water stress reduction function and an asymptotic root distribution function (Gale and Grigal 1987). The water stress reduction function was formulated according to the analogy of Ohm's law, taking into account both soil water pressure head and soil resistance to water flow; this methodology was conceptually more satisfactory than the ones that consider only soil water pressure head. Moreover, the water stress reduction function eliminated the need for reduction-point empirical parameters that are required in the existing functions, thereby increasing the model's applicability, but maintaining acceptable performance. The asymptotic root distribution function can be applied

Table 7. All time- and space-combined sensitivity coefficients for soil water pressure head (λ_h^*) and soil water content (λ_θ^*) of the proposed root-water-uptake module.

Site	Sensitivity coefficient	h_{fc}	h_{pwp}	d_r	T_m
Swift Current wheat	λ_h^*	28.6620	128.2276	128.2441	518.2569
	λ_θ^*	0.0002	0.0003	0.0010	0.0014
Simcoe soybean	λ_h^*	666.8630	223.3969	171.2854	629.1468
	λ_θ^*	0.0002	0.0001	0.0004	0.0007
Ottawa grass	λ_h^*	5.0998	27.6489	19.5430	21.7709
	λ_θ^*	0.0003	0.0002	0.0007	0.0003

not only to agricultural crops, but also to herbaceous and woody plants; the root distribution parameter β, the only parameter required in addition to the rooting depth, is available for a wide range of natural biomes. If the β value is not available, it can be estimated from the rooting depth as proposed in this study. Similar to a root-water-uptake model proposed by Li et al. (Li et al. 2001a), the current module takes water stress compensation into account.

The performance of the root-water-uptake module was evaluated against measured soil water contents at various depths at three different sites with different crops across Canada. The results show that the water stress function and the asymptotic root distribution function behaved similarly to existing ones, with the maximum difference in mean NRMSE and RE between the new and the existing water stress reduction functions being, respectively, 0.6% and 1.5%, and between the new and existing root distribution functions being, respectively, 1.2% and 2.1%. The root-water-uptake module performs equally well or better than the published modules, with mean NRMSE values ranging from 10.9% to 15.6% and mean RE values from 2.7% to 4.5%.

The root-water-uptake module was sensitive to all four parameters investigated and, comparing the three sites, no one parameter was consistently more sensitive than the others. Compared to existing ones, the new module contains only one empirical parameter (i.e., the root distribution parameter β), while the other ones are physically based and readily available. This module is therefore particularly useful when integrated into large-scale regional and global modeling efforts. Further evaluation of the model on other soil, climate, and plant types remains to be done, as well as more detailed sensitivity analyses.

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