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Sensitivity of a data-driven soil water balance model to estimate summer evapotranspiration along a forest chronosequence

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Abstract. The hydrology of ecosystem succession gives rise to new challenges for the analysis and modelling of water balance components. Recent large-scale alterations of forest cover across the globe suggest that a significant portion of new biophysical environments will influence the long-term dynamics and limits of water fluxes compared to pre-succession conditions. This study assesses the estimation of summer evapotranspiration along three FLUXNET sites at Campbell River, British Columbia, Canada using a data-driven soil water balance model validated by Eddy Covariance measurements. It explores the sensitivity of the model to different forest succession states, a wide range of computational time steps, rooting depths, and canopy interception capacity values. Uncertainty in the measured EC fluxes resulting in an energy imbalance was consistent with previous studies and does not affect the validation of the model. The agreement between observations and model estimates proves that the usefulness of the method to predict summer AET over mid- and long-term periods is independent of stand age. However, an optimal combination of the parameters rooting depth, time step and interception capacity threshold is needed to avoid an underestimation of AET as seen in past studies. The study suggests that summer AET could be estimated and monitored in many more places than those equipped with Eddy Covariance or sap-flow measurements to advance the understanding of water balance changes in different successional ecosystems.

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1 Introduction

Forested ecosystems are strongly subjected to human disturbances (Bonan, 2008; Hansen et al., 2010) and natural environmental changes (e.g. Kurz et al., 2008). From the year 2000 to the year 2005 more than 1 million km^2 of global forest cover was either converted to agricultural land or altered into developing successional forests (Hansen et al., 2010). Due to the magnitude of recent large-scale disturbances, the long-term effects of ecological succession on the exchanges between the atmosphere and a recovering biosphere are a major concern. Studies in temperate and boreal landscapes have found substantial differences in the carbon cycle (Sitch et al., 2003; Magnani et al., 2007; McMillan et al., 2008) and energy budget (Amiro et al., 2006; Juang et al., 2007) of differently aged forest including post-disturbance forest succession. Forest hydrology studies have long focused on post-disturbance flood increase, but few studies have attempted a quantification of the systematic influence of a recovering forest on the water balance. Jassal et al. (2009) found that in a Douglas fir succession initially reduced actual evapotranspiration (AET) recovered 12 yr after disturbance. Breña Naranjo et al. (2011) detected a recovery effect on AET in forest chronosequences and on the water balance in various watersheds in the North American West. Although changes in AET exhibited a different timing, a gradual recovery was common for at least 60 yr after disturbance. The uncertainty in estimating the timing of such long-time recovery leaves open questions about the complexity models need to predict annual or seasonal AET in a post-disturbance forest cover scenario.

Micrometeorological data obtained from the Eddy Covariance technique (EC) in experimental forested sites that allow to examine the effects of disturbance and succession on water, energy and biogeochemical fluxes are limited to a few research sites along chronosequences (e.g. Stoy et al., 2006, Jassal et al., 2009). However the scale of disturbance calls for more research and in particular for better monitoring of the water balance. In transitional climate zones such as those with a Temperate to Mediterranean Climate Type this research gap concerns particularly the summer AET of changing ecosystems.

Current experimental research in forests is characterized by the holistic observation of the soil-vegetation-atmosphere continuum. Measurements of water fluxes at the ecosystem scale for modelling purposes are, however, constrained by limitations at the spatial scale. The high degree of heterogeneity in soils can significantly affect the dynamics of water movement between the vadose zone and the atmosphere at large spatial scales. At the plot scale, the approximate representative footprint of the soil moisture measurements is in the order of 10^0 m^2 whereas the usual scale from the EC measurements is about 10^4 m^2 (Wilson et al., 2001). Furthermore, EC measurements typically underestimate the latent heat flux because of a bias in the data acquisition instruments, neglected energy storage sinks, losses from high frequency signals and advection (Wilson et al., 2002). With a lack of energy balance closure between 10 and 30 % the surface energy imbalance presents considerable uncertainty. This issue is further discussed in Appendix A1. Despite these challenges, the comparison between AET derived from soil water balance methods and EC observations have been previously explored and validated (Wilson et al, 2001; Schume et al., 2005; Kosugi et al., 2007).

Previous studies using EC observations to validate soil water budget methods are summarized in Table 1. Although the characteristics of each study are quite variable, most of the results are based on short-term periods or field campaigns, especially during the summer season. While the vertical scale of the observations ranged from 0–100 cm $\pm$ 40 cm, the number of spatial measurements varies from two or three soil water content probes in semi-arid ecosystems (Scott, 2010) to a maximum of 100 probes in a temperate mixed forest (Schume et al., 2005). Some studies interrupted the soil water balance computations during rainfall periods as they had no means to account for wetting front propagation, recharge and/or due to inadequate turbulent flux measurements (Wilson et al., 2001; Schelde et al., 2011). Other studies provided AET estimates based on continuous measurements (Cuenca et al., 1997; Moore et al., 2000). Another discrepancy found was the time step considered for modeling. As shown in Table 1 the soil water balance was estimated either using daily time steps (Cuenca et al., 1997; Wilson et al., 2001; Schelde et al., 2011) or shorter temporal resolutions (Schume et al., 2005; Schwärzel et al., 2009; Scott, 2010). Even though the results from the water balance were moderately lower compared to the measurements from the EC technique (with the exception of Schume et al., 2005) the rainfall interception by the canopy was not always explicitly considered in the models (Table 1). Soil water balance models may have additional reasons to underestimate actual evapotranspiration. These include the negligence of interception and root water uptake in shallower layers than the actual root zone depth (Table 1).

Although previous studies were successful in validating the use of the soil water budget to predict AET, none of them has examined systematically: (Eq. 1) a wide range of the few parameters in a data-driven soil water balance model and (Eq. 2) the potential errors arising from computation time steps. This study aims to explore the sensitivity of a data-driven soil water balance model to estimate summer AET along a forest chronosequence in a temperate climate. Besides testing the sensitivity of the model for different age forests and to a range of rooting depths and canopy interception capacity values, we specifically consider the following questions: (a) does the model performance depend on the computation time step?, (b) how much is the aggregation of the computation time step neglecting hydrological processes at the point scale?, and (c) what is the adequate timescale at which simple mass-balance approaches can adequately predict AET?

2 Modelling approach and validation

We will test two hypotheses concerning model behaviour. Our first hypothesis is that by choosing small computation time steps Δt, event-based processes affecting the soil water depletion from transpiration and soil evaporation and evaporation from the canopy will be more frequently observed and taken into account in the model. However, as Δt becomes larger, recharge will be explicitly accounted for resulting in an underestimation of summer AET. Our second hypothesis is that a soil water balance will provide better results in younger than in older forested ecosystems as the lack of dense canopy foliage in young stands diminish the role of interception while at the same time it enhances soil moisture depletion through soil evaporation and root water uptake as the dominant processes in AET. Nevertheless, we acknowledge that the canopy interception component in temperate coniferous forests are primarily a function of the frequency and duration of storms rather than stand structure (Gash and Morton, 1978).

During summer dry periods the soil water balance is proportional to the root water uptake and to evaporation from the soil. Previous research suggested that soil moisture limits evapotranspiration, especially during water stress periods (Teuling et al., 2006a, b). For most forest trees, including Douglas fir, the majority of roots are within the upper 50 cm, and absorbing roots are in the top 20 cm (Hermann, 1977). In regions with high rainfall amounts, compacted soil, or poor drainage, roots may be extremely shallow- in the top few centimetres only (Helliwell, 1989). This was corroborated by Humphreys et al. (2003) by estimating soil matric potential at the MS with the support of soil water retention curves which indicated that the bulk of the roots active in water uptake were in the upper 30 cm of soil. In this study we consider the

Table 1. Previous studies combining soil water balance and EC measurements.

Source	Seasons #	Soil depth range (cm)	Probes /site	Time* (%)	Δt (h)	Model performance	Inter-ception	Slope EB	EBR	EB gap
Cuenca et al. (1997)	1	0–120	5	0	24	Underest.	Yes	NA		
Moore et al. (2000)	2	0–155	NA	0	NA	NA	Yes	0.91	NA	NA
Wilson et al. (2001)	1	0–70	7	57	24	Underest.	No	NA	NA	0.8
Schume et al. (2005)	1	0–100	100	0	0.25	OK	No	NA	NA	0.85
Schwärzel et al. (2009)	1	0–90	51 & 64	NA	1	Underest.	Yes	0.82–0.87	NA	NA
Scott (2010)	13	0–100	2–3	49	0.5	Underest.	No	0.73–0.83	0.96–1.04	NA
Schelde et al. (2011)	1	0–75	8	47	24	Underest.	Yes	NA		

* Amount of time removed because of rainfall and/or flux corrections.

soil moisture dynamics during a summer period of 100 days from the end of June to the beginning of October (DOY 175 to 275). Two parsimonious data-driven soil water balance models with different complexity were tested across the three differently aged stands. In addition to four canopy interception thresholds described in Sect. 3, six different model time steps (Δt): 30 min, 60 min, 180 min, 360 min, 720 min and 1440 min and three different soil depths (Z_r): 0–30, 0–60 and 0–100 cm were used.

The chosen models do not consider physical and biological drivers of evapotranspiration such as net radiation, vapour pressure deficit, leaf area index and canopy conductance, but assume that the soil water balance can estimate AET during dry periods by:

$$Z_r \frac{d\theta(t)}{dt} = P(t) - q(t) - (E_s(t) + T(t)) \tag{1}$$

where Z_r is the active root depth, P is rainfall, q is percolation, E_s is soil evaporation, T is transpiration and θ is the volumetric soil water content. With the exception of Z_r (in mm) and $\theta(-)$, all variables are expressed in mm/Δt. For further simplification, we assume that the value of Z_r and the soil depth where soil moisture limits $E_s + T$ is similar.

This simple water balance model may be valid in young stands where rainfall interception is low. Nevertheless, as forests transition to mature ecosystems, interception will play an important role in evapotranspiration (e.g. Savenije, 2004). A second water balance model (Eq. 2) therefore added the evaporation from water stored in the canopy $I(t)$. A maximum canopy interception capacity I_{max} (mm h^{-1}) is defined to calculate the time variable evaporation from the canopy $I(t) = I_{max}$ when $P(t) \geq I_{max}$ and $I(t) = P(t)$ when $P(t) < I_{max}$. The total AET will hence depend on the evaporation from canopy:

$$\mathrm{AET}(t) = E_s(t) + T(t) + I(t) \tag{2}$$

Positive values of $\frac{d\theta(t)}{dt}$ due to infiltration and hydraulic redistribution of soil water were considered to be equivalent to percolation $q(t)$. The model assumes that q occurs only during rainfall events and up to 2 h after the end of rainfall. During this period we did not calculate AET from the soil moisture observations. For periods before and after this defined period, we assumed that q is equal zero.

The aggregation of computation time steps from 30 min into 1, 3, 6, 12 and 24 h implies to sum the fluxes (P, θ, q, I) so the $I(t)$ values from low-intensity rainfall events, lower that the canopy interception capacity, are set to I_{max}. Likewise, as the time scale of infiltration and redistribution after a storm is smaller than the larger Δt's and for which the soil water content was averaged over Δt values, the impact of summer storms will be implicitly included in the water balance calculations with the longer model time step.

For model validation, observed AET from the flux towers was used. To assess sensitivity of the estimates to the different model choices, modelled and observed AET were compared at two different levels of aggregation:

1. mean total summer AET over the study period

2. mean diurnal cycle.

Furthermore, the sensitivity to the inter-annual climatic variability was assessed based on the total annual summer AET as well as on Nash-Sutcliffe efficiency index (NSE) (Nash and Sutcliffe, 1970) for the simulated time series of each summer.

3 Study site

This study used data from an evergreen forest chronosequence located on the east coast of Vancouver Island, BC, Canada. The Campbell River forest chronosequence, an experimental research site member of the FLUXNET initiative (Baldocchi et al., 2001), consists of three differently aged coastal Douglas-fir (*Pseudotsuga menziesii* (Mirbel) Franco) stands: a young (YS), intermediate (IS) and mature stand (MS). Published data available to the scientific community was obtained online (http://www.fluxnet-canada.ca/). The three sites were harvested in 2000, 1988 and 1949 respectively, followed by slash-burning and planting predominantly Douglas-fir seedlings. Subsequently, several overstory and understory species have also grown. The climate at the sites

Table 2. Characteristics of the successional forest chronosequence experimental sites in Campbell River, Canada.

	Young stand	Intermediate stand	Mature stand
Abbreviation	YS	IS	MS
Year of establishment	2000	1988	1949
Latitude	49°52′20″ N	49°31′11″ N	49°52′8″ N
Longitude	125°17′32″ W	124°54′6″ W	125°20′6″ W
Elevation (masl)[a]	175	170	300
Average tree height (m)[a]	2.4	7.5	33
Stand density (trees ha^{-1})[a]	1400	1200	1100
Mean summer LAI $\pm$ std. dev ($-$)	1.9 ± 0.5[b]	5.5 to 6.8[c]	8.4[c]
Mean annual ET (mm)[a]	253	362	398
Mean summer ET (mm)	126	161	168
Mean summer P (mm)	112	121	135

[a] from Jassal et al. (2009),
[b] from 2002 to 2005,
[c] from Humphreys et al. (2000)

is characterized by cool and rainy winters and, dry and relatively warm summers. Mean annual precipitation is 1497 mm and the mean annual temperature is 9 °C. The average growing season extends from March to October. From May to October the region has climatic moisture deficits with the largest values in July and August (Moore et al., 2010). Soil texture at all sites is in the range from gravely loamy sand to sand. Humphreys et al. (2006) and Jassal et al. (2009) described further details about the sites. The main characteristics of the forest chronosequence are shown in Table 2.

A meteorological tower sampled the vertical exchanges of latent heat λE above each stand using the EC technique. Half-hourly fluxes of water vapour were provided from three-axis anemometers and infrared gas analysers (for more details see Jassal et al., 2009). The quality of the turbulent fluxes at the three sites was not affected by the wind direction and the energy balance closure did not decrease when wind came from behind each EC tower (Humphreys et al., 2005). Moreover, specific wind directions were not used to remove observed fluxes given the sufficient area of homogeneous stand footprint were within the typical values. At the MS, the fetch was 400 m to the southwest and about 700 to 800 m to the northeast with Douglas fir stands between 25 to 60 yr old surrounding the tower from all sides (Humphreys et al., 2003). At the IS, the fetch was between 400 and 700 m in all directions and surrounded by young and mature forests. At the YS, the fetch was about 400 m in both the northeast and southwest directions surrounded in all directions by 60 yr-old Douglas-fir forests. Footprint assessments suggested that the minimum fetch estimated for all the sites was sufficient for flux observations (Humphreys et al., 2006). Humphreys et al. (2003) established that anabatic–katabatic circulation patterns occurring during most of the year could reduce the expected generation of local circulations patterns developed by the thermal convection due to clear-cut areas surrounded by mature forests. The quality control procedures of measured latent heat fluxes involved despiking, especially during periods of heavy rainfall, and removing portions of the high frequency time series associated with calibrations before computing EC statistics (Humphreys et al., 2006). Fluxes with statistics that did not fall within reasonable limits and/or occurred during instrument malfunction were removed. Flux measurements were also removed when data did not fall within the specified limits for realistic data, when the water vapour mixing ratio was negative and when the non-stationarity ratio was greater than 3.5 (Mahrt, 1998). The underestimation in λE ranged from 0 to 10 %. Any missing λE measurements were replaced using the Penman-Monteith equation. Observed weather variables and canopy conductance were modelled using a Jarvis-Stewart parameterization (see Humphreys et al., 2003).

Precipitation (P) was measured in each stand using two tipping-bucket rain gauges. The frequency of the observations was recorded at 3 s intervals and then averaged for 30 min (model 2501 tipping bucket rain gauge, Sierra Misco, Berkeley, CA or model 4000 Jarek Manufacturing Ltd., Sooke, BC, Canada). In order to avoid outliers, the precipitation records at the MS site and at the nearby Campbell River Airport, British Columbia were compared and considered acceptable since the difference in total rainfall recorded at the two locations within the summer months was less than 5 % (Humphreys et al., 2003). Precipitation data were gap-filled using statistical relationships with related data from the same site or the other sites (http://fluxnet.ccrp.ec.gc.ca). Errors derived from wind factors (gage setup, wind speed and drop size distribution) were not accounted. Scott (2010) found an underestimation of 5 % for semi-arid ecosystems with an annual P between 280 and 400 mm. The summer precipitation across our the study site varied from less than 100 mm to more than 200 mm in dry and wet summers, respectively.

Wind during the summer period is relative low compared to the winter. Therefore we expect an overall error for precipitation during summer in the range of 2–5 %.

Percolation (q) was not measured. However volumetric soil water content data (θ) were available from two locations at each site equipped with time-domain reflectometry (TDR) probes reporting depth ranges averaged over 0–30, 0–60 and 0–100 cm with a time resolution of 30 min (Jassal et al., 2009). Moreover, 11 additional TDR sensors integrating the top 30 and 76 cm of the soil profile were manually measured monthly with a cable length tester and used to observe the spatial variability in θ over the turbulent flux source area at the MS (Humphreys et al., 2003). The probes were established below ground level and beneath the organic layers of leaf litter, near the tower/hut. The high-resolution data from the two locations was averaged and then used as the model input. The original EC and θ data were screened for instrument failure and gap-filled (Humphreys et al., 2005, Jassal et al., 2009).

Humphreys et al (2003) estimated canopy interception with two methods. The first method used measured throughfall and stemflow rates at the MS site. In addition, a second method with the support six leaf wetness sensors was used to determine the minimum depth of water required to saturate the entire canopy surface (1.5 mm), or canopy interception capacity ($I_{\max}$) (Leyton et al., 1965) at different heights of the MS canopy. The canopy was assumed to be completely wet when all the leaf wetness sensors were wet or the calculated water stored on the canopy was greater than 1.5 mm. The canopy was assumed to be dry when all the leaf wetness sensors were dry and the water stored on the canopy was 0 mm. The degree of uncertainty was large, the agreement between estimates of the canopy wetness and, the canopy interception capacity inferred from the sum of throughfall and stemflow ranged between 50 and 70 % (Humphreys et al., 2003). Moreover, the maximum threshold of $I_{\max}$ was roughly approximated from the observed daily AET at the 3 stands before the energy balance closure. The daily AET varied from 2.5 mm to less than 1 mm day^{-1} at the early and late stages of the summer season, respectively.

Based on these observations, $I_{\max}$ is considered not to be larger than 1 mm h^{-1}. Because of differences in LAI, the sensitivity analysis in this study tested values of $I_{\max}$ of 0.25, 0.5 and 1 mm h. Such values also fit with the interception threshold of 0.5 mm h^{-1} suggested by Gash (1979). The value of $I_{\max}$ assumed that the water falling on the canopy during low-intensity rainfall intensity events (up to $I_{\max}$) will mostly be stored and evaporate. Rainfall intensities exceeding this threshold will cause throughfall to the soil surface and infiltration into the soil. Jassal et al. (2009) provides LAI data measured during the summer season in 2002–2005 at the young stand and in 2002 at the intermediate/mature stands. The LAI variability observed for the mentioned periods was not expected to change the $I_{\max}$ range of values. The data is shown in Table 2. The mean annual energy balance

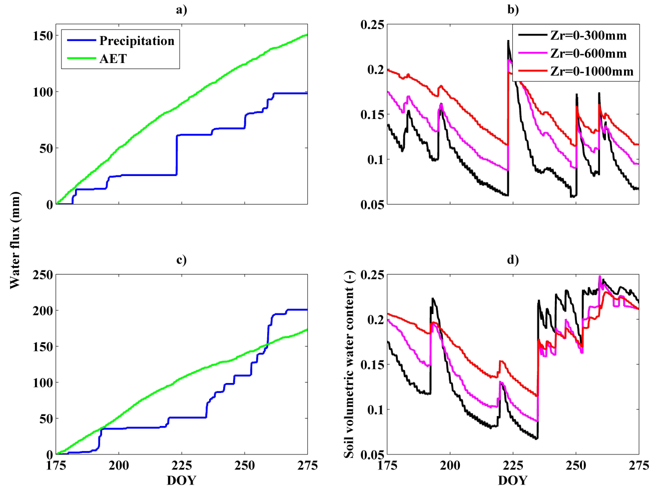

Fig. 1. Measured time series of precipitation (cumulative), evapotranspiration (cumulative) and soil moisture during a dry summer (2003, **a** and **b**) and a wet summer (2004, **c** and **d**) at the mature stand (MS).

closure at the YS, IS and MS was 0.89, 0.83 and 0.88, respectively (Jassal et al., 2009). Our study period covered seven summer seasons at the young and mature stands from June 2001 to September 2008. Observations at the intermediate stand began in 2002.

An example of the water supply-demand in this study site during periods of different water-stress and the importance of the soil water reservoir is shown in Fig. 1. Within the same stand (MS), summer AET will be about 150 mm in a dry summer following a dry winter (Fig. 1a) and about 170 mm in a wet summer (Fig. 1c). While the difference in AET between such contrasting summer seasons in about 20 mm, P in the wet summer was twice as much as in the dry summer. In such cases, the soil water deficit from scarce summer rainfall and water use by vegetation will be compensated by soil moisture stored during spring and winter (Fig. 4, top). Figure 1b, d show the vertical distribution of soil water content for the same contrasting summers. During dry summers (Fig. 1b), the soil water content at shallow soil layers is rapidly depleted compared with the deeper layers whereas this pattern changes at the end of the summer or during wet summers (Fig. 1d).

4 Results

4.1 Sensitivity of the model

Figure 2 (top) shows the mean diurnal cycle of observed and simulated summer AET from 2001 to 2008 using $Z_r = 30$ cm, $I_{\max} = 0.25$ mm h^{-1} and $\Delta t = 60$ min. AET is generally underestimated during daytime: slightly at the young stand

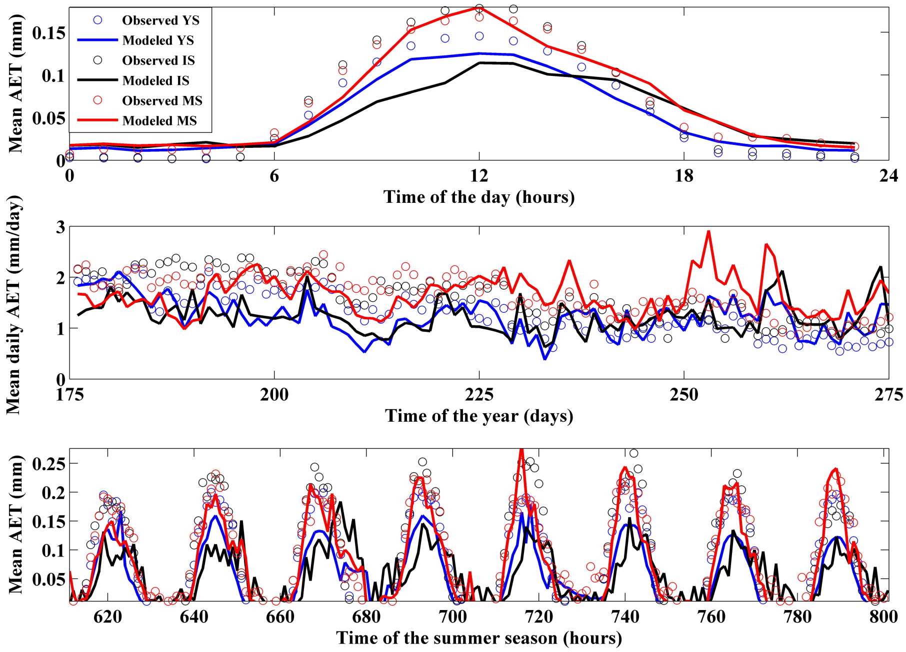

Fig. 2. Modeled and observed summer AET (2001–2008) based on $Z_r = 30$ cm, $I_{max} = 0.25$ mm h^{-1} and $\Delta t = 60$ min: (top) mean diurnal cycle simulated (middle) mean daily AET values simulated (bottom) mean hourly AET values simulated (zoomed in from the whole time series).

(YS) and mature stand (MS), but strongly at the intermediate stand (IS). AET is overestimated during nighttime, again most strongly at IS, where the AET observations show a weaker diurnal variation than at the other stands. With the exception of the YS, the observed soil moisture depletion during nighttime does not reflect the observed strong reduction of latent heat fluxes during the night using EC.

Using the same modeling time step of $\Delta t = 60$ min, Fig. 2 (middle) shows the observed and simulated mean daily summer AET from 2001 to 2008. AET is strongly underestimated in early to mid summer and slighty overestimated in late summer. Daily AET at IS and MS is overestimated slightly more during the second half of the summer period when autumn storms start to occur. For all sites, the estimated mean daily AET in the second half remains relatively constant whereas the measurements indicate a decrease. Finally, the mean AET at each hourly time step (Fig. 2, bottom) shows the main weakness of the soil water balance model, which estimates significantly higher or lower values at certain time steps.

Table 3 shows the sensitivity analysis of the model to a multiple combination of rooting depth (Z_r) and canopy interception capacity values (I_{max}) for a fixed time step of $\Delta t = 30$ min. The errors were calculated for the mean total summer AET for the period 2001 to 2008. The results are acceptable when the soil moisture dynamics are considered in the shallow soil layer of 0–30 cm. As the rooting depth is assumed to be larger (e.g. 60 cm and 100 cm), the depletion of the volume of soil water within the layer tended to overestimate summer AET. An exception was observed at the intermediate stand for moderate and high I_{max}. The sensitivity of the model to I_{max} was rather heterogeneous. The differences in the absolute error from $I_{max} = 0$ mm h^{-1} to $I_{max} = 1$ mm h^{-1} were lower for YS than for IS and MS. When the error was expressed relative to the observed AET the differences between YS, IS, and MS are not as pronounced.

The sensitivity to the computational time step is shown in Table 4 and Fig. 3. Overall, the absolute and relative errors increased whereas the NSE decreased as the soil moisture data was aggregated over longer time steps. The step $\Delta t = 30$ min provided the best results for the YS and IS but $\Delta t = 60$ min proved to be more adequate for the MS. Despite an overall decay of model performance as a function of the time step (Fig. 3), an example of model equifinality (i.e. same results with different model parameters) was observed at the YS when a similar error was obtained for $\Delta t = 30$ min and 180 min (Table 4). The main difference in the parameter combination was found to be $Z_r = 30$ cm for $\Delta t = 30$ min and $Z_r = 60$ cm for $\Delta t = 180$ min. Finally, the differences in the error for the least optimal $\Delta t = 1440$ min (= 24 h) at the three stands differed strongly. The largest error was observed at the IS followed by the YS and the MS. Stand age was not an important factor affecting model performance, but its sensitivity to the time step.

Table 3. The sensitivity of the model to soil depth, canopy interception and a fixed $\Delta t = 30$ min. The sensitivity is expressed as the absolute error (AE), relative error (RE) and Nash-Sutcliffe efficiency index from the mean total summer AET (mm) for the period 2001–2008. The lowest errors in each stand are highlighted in bold. Uncertainty derived from the energy imbalance is not considered.

		Young stand (YS)			Intermediate stand (IS)			Mature stand (MS)		
Z_r (cm)	I_{max} (mm h^{-1})	AE (mm)	RE (%)	NSE (–)	AE (mm)	RE (%)	NSE (–)	AE (mm)	RE (%)	NSE (–)
30	0	−8	−7	0.78	−23	−14	0.55	+10	+7	0.68
	0.25	+2	+1	0.81	−12	−7	0.57	+21	+13	0.71
	0.5	+4	+3	0.79	+7	+4	0.55	+40	+25	0.63
	1.0	+11	+8	0.75	+15	+9	0.52	+47	+29	0.58
60	0.0	+22	+15	0.75	−36	−23	0.54	+72	+46	0.07
	0.25	+32	+24	0.76	−26	−17	0.57	+82	+51	0.11
	0.5	+35	+25	0.74	−8	−6	0.56	+102	+63	−0.08
	1.0	+42	+31	0.66	+1	0	0.55	+110	+68	−0.21
100	0.0	NA	NA	NA	NA	NA	NA	+127	+80	−0.91
	0.25	NA	NA	NA	NA	NA	NA	+137	+95	−0.9
	0.5	NA	NA	NA	NA	NA	NA	+157	+97	−1.22
	1.0	NA	NA	NA	NA	NA	NA	+165	+102	−1.42

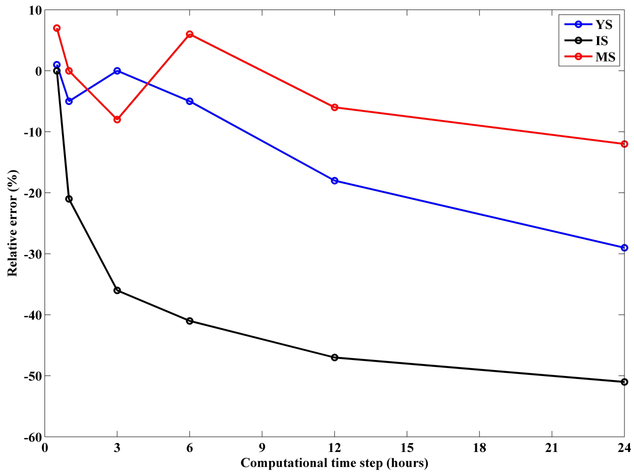

Fig. 3. Sensitivity of the data-driven soil water balance model to the computational time step (Δt). The relative error shown is the lowest error obtained by the combination of Z_r and I_{max} for each time step.

Table 4. Sensitivity of the data-driven soil water balance model to the computational time step (Δt). The absolute error and NSE shown is the lowest obtained by the combination of Z_r and I_{max} for each time step. The relative error is shown in Fig. 3. The absolute error (AE) after the correction from the slope of the energy balance is shown in brackets.

	Young stand		Intermediate stand		Mature stand	
Δt (min)	AE (mm)	NSE (–)	AE (mm)	NSE (–)	AE (mm)	NSE (–)
30	+2 [−11]	0.81	+1 [−10]	0.55	+10 [−12]	0.68
60	−6 [−19]	0.81	−33 [−44]	0.55	−2 [−24]	0.75
180	+2 [−11]	0.81	−56 [−67]	0.33	−14 [−36]	0.44
360	−4 [−17]	.55	−65 [−76]	−0.05	+8 [−14]	−0.29
720	−21 [−34]	−1.1	−74 [−85]	−1.45	−12 [−34]	−2.85
1440	−35 [−48]	NA	−81 [−92]	NA	−22 [−44]	NA

4.2 Interannual variability

Observed summer AET shows considerable inter-annual variability with differences along the chronosequence: AET at the YS and IS generally varies more from year to year than AET at the MS (Fig. 4). For the different stands the performance of the model of each summer shows a different sensitivity to this climatic variability (Fig. 5). The summers in 2003 and 2006 were dry compared to historical observations while 2004 and 2007 were wet summers. Figure 4 shows that the observed AET in those years and the simulated $E_s + T$ (dashed line) at the YS and IS followed the expected variation of ET from a dry to a wet summer and subsequently to a dry one. The behaviour at the MS was more complex. The soil water balance indicates an increase in $E_s + T$ until the occurrence of the extreme dry summer of 2003 and then started a steady decrease until 2008 despite the subsequent wet summers. The inclusion of evaporation from $I(t)$ (bold line) has hardly any effect on the magnitude and interannual variability of AET at the YS. However, at the IS the inclusion of $I(t)$ improved the prediction of AET during five years out of six.

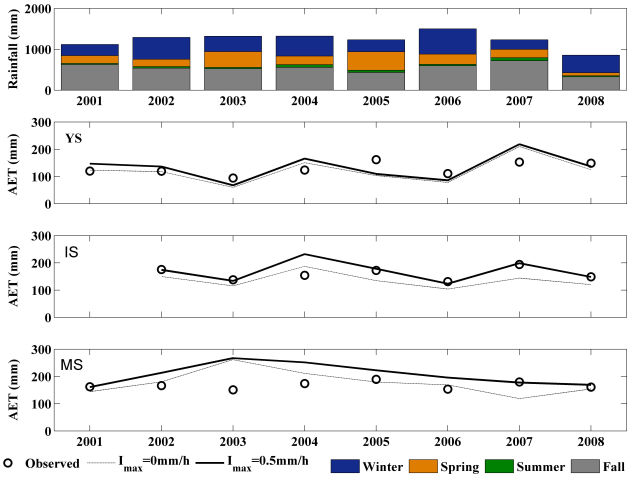

Fig. 4. Inter-annual variability of seasonal precipitation (top) and, observed and modeled summer AET at the different successional forest stands. AET modeled with $\Delta t = 30$ min.

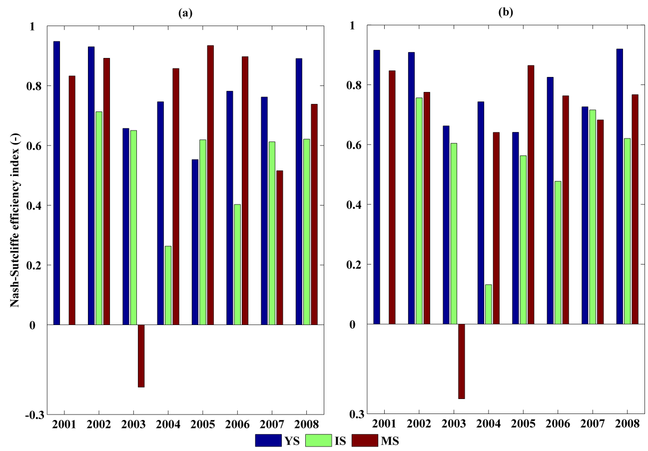

Fig. 5. Nash-Sutcliffe efficiency for seasonal simulations with **(a)** $I_{max} = 0\,mm\,h^{-1}$ and **(b)** $I_{max} = 0.5\,mm\,h^{-1}$. AET modelled with $\Delta t = 30$ min.

The standard deviation of the simulated time series is twice as large as for the observed time series because the inter-annual variability of soil moisture dynamics is higher (not shown). At the young and intermediate stands, the fluctuations of θ are higher during the analysed period whereas at the mature stand the fluctuations of θ are mainly affected during the drought year in 2003.

In a similar way, the model performance can be summarized by NSE calculated from the half-hourly values for each summer period. The indices revealed considerable variation among years (Fig. 5). However, the NSE values are overall high. At the young and mature stands they were mostly above 0.7 while at the intermediate-aged stand they were mostly in the order of 0.6. They are proof for the predictive power for diurnal AET estimates as the NSE values are biased to the high values of a time series. The drought summer in 2003 presents an outlier with a negative NSE for the mature stand. The effect of drought may also have had a lagged effect on the modest results at IS in 2004. However the ability of the model to capture AET patterns at YS and MS recovered during the subsequent years. At the YS the model approach seems most resilient to seasonal drought. Finally, the mean NSE was more sensitive to Δt (see Tables 3 and 4) than to the parameters Z_r and I_{max}. The higher Δt, the lower was the obtained NSE. Low diurnal values of modelled AET for larger time steps suffer from recharge in the model mistaken for soil moisture depletion due to AET.

5 Discussion

The combination of three rooting depths (Z_r), four canopy interception capacity thresholds (I_{max}) and six computational time steps (Δt) yielded 72 possible results of the mean total summer AET at each stand for the period 2001–2008. This period of 2300 days in total years was longer than that of previous studies. Based on the literature we expected that the adequate Z_r was 30 cm and that the optimal I_{max} thresholds would be in the range between 0.25 and $0.5\,mm\,h^{-1}$. However, the model was tested with more parameter combinations to explore if more improbable physical parameters could lead to a better agreement between the simulations and the observations.

The results indicated three factors systematically influencing the results, one for each model parameter. The first was related to Z_r. As Z_r became larger, the mean summer AET was increasingly overestimated. The second was related to Δt. The choice of longer time steps resulted into a stronger underestimation of AET. Similar underestimations were previously observed for $\Delta t = 24$ h (Cuenca et al., 1997; Wilson et al., 2001; Schelde et al., 2011). The third factor related to I_{max} proved less influential than Z_r or Δt. Larger I_{max} were related to higher AET estimates and vice versa. The young stand was less sensitive to I_{max} than the other stands (Table 3). The combination of overestimated (due to Z_r) and underestimated (due to Δt) summer AET across the summer seasons thus influences model equifinality. As shown in Table 4, the absolute error of +2 mm for the mean total summer AET at YS was simultaneously found for $\Delta t = 30$ min and $\Delta t = 180$ min. Their respective Z_r and I_{max} values were 30 cm and $0.25\,mm\,h^{-1}$ and 60 cm and $0.25\,mm\,h^{-1}$, respectively. The good agreement between observed and modelled summer AET from 2001 and 2008 was in fact influenced by a compensation of errors, in this case an underestimation from the parameter $\Delta t = 180$ min and an overestimation from the parameter $Z_r = 60$ cm.

It is clear that both eddy-flux observations and the soil moisture data-driven model will produce results with uncertainty due to measurement errors, model assumptions and parameterization. Unfortunately, the eddy-flux community does not provide a continuous estimation of the error for the estimated latent heat flux, which could then be used to derive an overall error for the different aggregation levels. Assuming that the error was evenly distributed between the sensible and latent heat flux, we can calculate from the energy closure an underestimation of summer AET between 7 and 13 % (Appendix A, Table A1). Part of the energy closure imbalance (error in AET) can be attributed to the contribution of low-frequency turbulent fluxes by the EC technique. Differences in the fluxes calculated with the 20 and the 8 Hz signals (from different sonic anemometers) were found to be negligible by Humphreys et al. (2005) but recently Sánchez et al. (2010) analysed the energy balance ratio over a boreal forest for different time steps and found out that the closure could be significantly improved when aggregating the original turbulent fluxes over longer time steps than the standard 30 min available from the FLUXNET database. Unlike the simulated AET, the observed latent heat fluxes were not aggregated into larger time steps.

We showed that summer soil moisture depletion at the point scale can be used to estimate summer AET at the footprint scale as observed with eddy-flux measurements. Despite important differences in the main canopy, understory species and root structure, stand age at different stages of succession did not systematically affect the agreement between the observations and model estimates in most years and for the long-term average. Interannual climatic variability somewhat affected the estimates. The response of the soil moisture balance to the interannual variability was consistent and proportional to the observed changes in the summer AET, but appeared to show a closer coupling for the young and intermediate stands. In the mature stand summer AET shows little inter-annual variability and during an extreme dry year, AET of the mature stand appeared decoupled from soil moisture in the top soil layer.

The main limitation of the presented model is the simplified evaporation from canopy interception. Evaporation from the canopy to the atmosphere relies on the boundary-layer meteorology (Brutsaert, 1982). Hence, variables as net radiation, temperature aerodynamic conductance and vapour pressure deficit are able to provide more detailed simulations of AET, but also considerable uncertainty (Van der Tol et al., 2009). As shown in Fig. 2, the main weakness of applying a soil water budget method results in misestimates for summer AET at the sub-daily scale but not at the seasonal scale. We acknowledge that using a canopy interception-evaporation threshold to describe the exchange of water vapour from the canopy to the atmosphere will not provide the best estimate. However, since the main aim of the manuscript is to estimate summer AET only with precipitation and soil moisture data, i.e. without an extensive meteorological dataset, which is often missing, and with the most parsimonious model possible and within an acceptable error range (Tables 3 and 4) the achieved results are promising with best estimate for the relative error of less than 1 % and 10 % when the energy balance closure correction was applied.

A potential improvement of the modelled AET through the consideration of rainfall interception for predicting summer AET was inconclusive. However, since there is not a clear relation between LAI and the interception parameter, it should be kept in mind that due to the measurement scale of the soil moisture data, throughfall driving soil moisture may have a large spatial variation. This spatial variability of throughfall shows a consistent temporal persistence as for example studied by Keim et al. (2005) for different forest stands. This characteristic may explain that the mature stand can have the smallest interception threshold due to the spatial variability being highest and therefore a high chance that the soil moisture measurements are at locations with a small local interception.

The choice of a computation step of $\Delta t = 30$ min and an active root depth $Z_r = 30$ cm proved generally superior to other model options. Larger computation time steps imply that the effects of recharge during rainfall events are taken in account so it can notably reduce root water uptake and hence underestimate summer ET by more than 50 % according to field observations. The negligence of tree rooting strategies during periods of water stress by using soil moisture from a fixed depth of 30 cm for the model resulted in an overestimation during extreme dry years at the mature stand. The incorporation of soil water to root depth dynamics may reduce the error (Teuling et al., 2006a; Schymanski et al., 2008), but requires at least one additional parameter. Nevertheless, the role of hydro-climatic seasonality on vertical root dynamics is relevant as the combination of warm and wet conditions can result in deeper rooting depth. Schenk and Jackson (2005) found that in a cool-temperate forest (a biome similar to our study), the 50 % and 95 % rooting depths (the depths above which 50 % or 95 % of all roots were located) was at 21 and 104 cm, respectively. Moreover, as pointed out by Schenk and Jackson (2002), rooting depth strongly depends on climatic properties and to a minor extent on soil properties. The geographical distribution of roots will require a different model parameterization for Z_r based on climatological and vegetation properties. Generally, shallow roots occur mainly in cold and wet regions due to the lack of water stress whereas deeper roots have been observed in warm and dry regions.

6 Conclusions

During the recovery of ecosystems, measurements of heat fluxes, rainfall interception, stomata conductance and root depth, among others, show constant change. Under such circumstances, gross simplifications are needed in order to avoid a model over-parameterization at each successional stage. In this paper, we tested the hypothesis that a data-driven soil water balance model using soil moisture and precipitation measurements can provide a reasonable approximation of summer AET when choosing an appropriate computational time step, rooting depth and canopy interception. The model was validated in a successional chronosequence characterized by long-term physiological changes that directly affect the water balance partition in forests. We found that stand age does not affect the performance of the model.

The parsimonious data-driven model for the estimation of summer AET presented here has a main advantage over more data-intensive modelling schemes to study hydrological changes in recovering forested ecosystems. For example, LAI, soil properties, boundary-layer meteorology and detailed root depth parameterization are not required. The current approach proved to be useful for predicting evapotranspiration during the water limited period of the year in a disturbed and successional temperate ecosystem and its differences in their magnitude caused by different patterns of interception, soil evaporation and transpiration. An eight-year comparison of both models' output showed that in order to provide acceptable AET estimates it is essential to account for rainfall in the soil water balance by choosing an adequate calculation time step (Δt) and rooting depth (Z_r). However, we would like to emphasize that methods solely based on vadose zone measurements cannot capture detailed processes occurring at the vegetation-atmosphere continuum and therefore are not suitable to describe AET at temporal resolutions shorter than several days. The model showed to be less sensitive to the canopy interception threshold (I_{max}) than to the former parameters. Overall, the combination of a Δt between 30 and 60 min and a Z_r between 0 and 30 cm provided the best estimates for AET. Parameter values outside this optimal range were likely to enhance error estimates leading to an over- or under-estimation of the summer AET. Acceptable errors using non-optimal parameters were observed but it was found to be the consequence of error compensation from the sum of over- and under-estimated periods. The error measurements derived from the energy balance non-closure did not considerably affect the agreement between observed and simulated mean summer AET.

The results provided here showed that one single model structure can be sufficient to make an appropriate estimation of hydrologic change in a gradually evolving landscape and thus does not constitute a problem for modelling of transient hydrological systems. However, the model should be tested across a wide range of forested and non-forested landscapes in different climates to evaluate applications including estimates at sites where Eddy Covariance measurements are limited or difficult due to topographic complexity and in sites where long-term soil moisture data are available.

Table A1. Parameters of the linear regression between the available energy ($R_n - G - S_b - S_{LE} - S_H$) and the turbulent fluxes (LE + H) at the three stands from 2001 to 2008.

Site	EBR	Slope	Intercept ($W\,m^{-2}$)	R^2	RMSE ($W\,m^{-2}$)
YS	–	0.79	6	0.93	39
IS	–	0.85	14	0.95	39
MS	–	0.74	4	0.92	47
YS (08:00 am–05:00 pm)	0.85	0.79	9	0.84	53
IS (08:00 am–05:00 pm)	0.82	0.90	−4	0.88	55
MS (08:00 am–05:00 pm)	0.71	0.79	−15	0.85	63

Appendix A

Energy balance closure

Any model for prediction purposes relies on good quality input data. The validation of any model further relies on the quality of the measurements it is validated against. In this case, the model validation is affected by the energy balance closure of the EC derived AET observations. It is known that the EC technique overestimates the available energy and underestimates the turbulent fluxes and therefore AET (Wilson et al., 2002). The energy imbalance also depends on local factors such as topography, wind direction and friction velocity, among others. In order to estimate the uncertainty in the observed AET, the surface energy imbalance was estimated using two different methods following Wilson et al. (2002): the first is the Energy Balance Ratio (EBR) (Sánchez et al., 2010) over a certain time period. EBR is defined as the ratio of the available energy to the turbulent fluxes:

$$\mathrm{EBR} = \frac{R_n - G - S_b - S_{LE} - S_H}{\mathrm{LE} + H} \tag{A1}$$

where R_n is the net radiation, G is the soil heat flux, S_b is the biomass heat storage flux, SLE is the latent heat storage flux, SH is the sensible heat storage flux, LE is the latent heat flux above the canopy and H the sensible heat flux above the canopy. All the units are expressed in $W\,m^2$. The second method used to estimate the lack of closure of the energy balance is an ordinary linear regression between the available energy and turbulent fluxes, numerator and denominator in Eq. (A1), respectively.

Since the energy imbalance is larger during nocturnal periods (Wilson et al., 2002), the EBR and the linear regression parameters (slope, intercept, R^2 and root mean square error) were estimated for 24 h periods as well as during the day (Table A1). Figure A1 shows the results for the EBR and linear regression. With ERB between 0.71 and 0.85, slopes between 0.75 and 0.85, and R^2 between 0.84 and 0.95, the energy imbalance values for the three stands are consistent with previous studies (Wilson et al., 2002, Sánchez et al., 2010). The IS was the site with the lowest imbalance, the YS ranks second. The imbalance is larger at the MS. The improvement

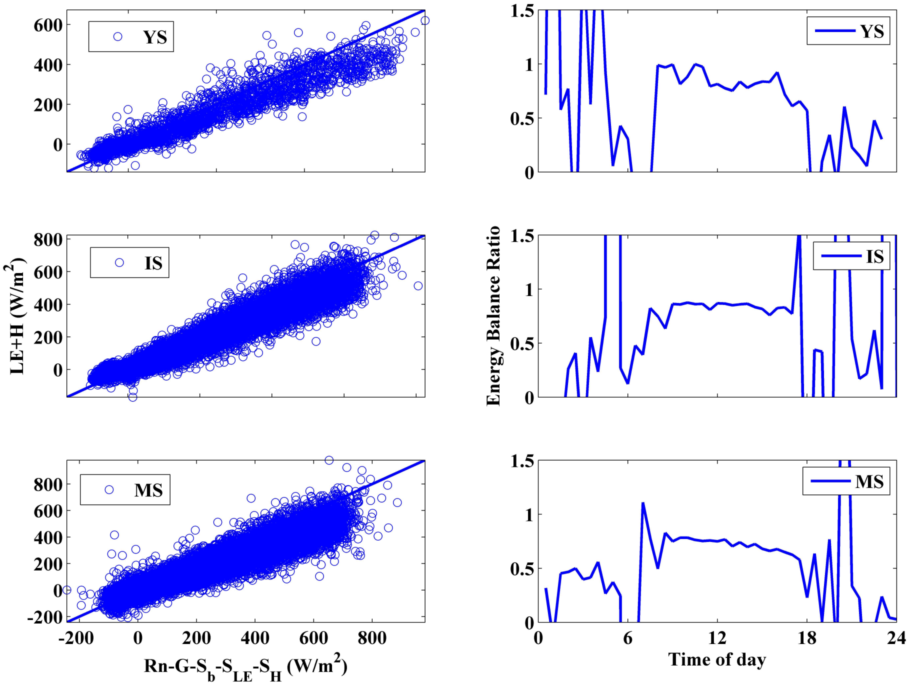

Fig. A1. Left side: relationship between the measured turbulent fluxes (y-axis) and the available energy (x-axis) at the three stands YS (top), IS (middle) and MS (bottom) from 2001 to 2008. Right side: variability of the average energy balance ratio during the day.

of the closure when excluding the late evening/nocturnal periods (Fig. A1, right side) was moderate at the IS and MS but did not have any effect at the YS (Table A1).

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