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Observed timescales of evapotranspiration response to soil moisture

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Received 28 September 2006; revised 27 October 2006; accepted 1 November 2006; published 9 December 2006.

[1] The sensitivity of evapotranspiration (ET) to soil moisture storage plays an important role in the land-atmosphere system. Yet little is known about its magnitude, or its dependence on vegetation, soil, and/or climate characteristics. Here we relate the sensitivity to the timescale of ET decay in absence of rainfall, and show that it can thus be derived from time series of ET alone. We analyze ET observations from 15 vegetated sites covering a range of climates conditions, yielding timescales of 15–35 days. Longer timescales (weaker ET sensitivity) are found in regions with seasonal droughts, or at sites with woody vegetation. We compare observed values with output of different land surface models (LSMs) from the Second Global Soil Wetness Project, revealing large inter-model differences and significant differences with observations. Our methodology can lead to improved representation of soil moisture effects on ET in LSMs. **Citation:** Teuling, A. J., S. I. Seneviratne, C. Williams, and P. A. Troch (2006), Observed timescales of evapotranspiration response to soil moisture, *Geophys. Res. Lett.*, *33*, L23403, doi:10.1029/2006GL028178.

1. Introduction

[2] Plants play a central role in the global water and energy cycles by regulating the partitioning of energy fluxes at the land surface in response to the soil moisture availability in the root zone and atmospheric conditions. Soil water dynamics in the root zone are dominated by rapid infiltration and subsequent vertical redistribution following rainfall, in contrast to slow drydown due to water uptake by roots for evapotranspiration (*ET*) during interstorm periods. Parameterization of the latter process is complicated by the adaptive behavior of plant water uptake [e.g., *Teuling et al.*, 2006]. Key parameters and states (i.e., root distribution, soil moisture) are difficult to measure at the appropriate scale, and they can show large spatial and temporal variability.

[3] The sensitivity of *ET* to soil moisture is a central parameter within the coupled land-atmosphere system: Several studies with Global and Regional Climate Models (GCMs, RCMs) suggest for instance that it might influence low frequency atmospheric variability in precipitation and temperature [e.g., *Koster et al.*, 2004, 2006; *Seneviratne et al.*, 2006a]; moreover, it is also relevant for the timescales of soil moisture autocorrelation [e.g., *Koster and Suarez*, 2001; *Seneviratne et al.*, 2006b]. However, little is known on the magnitude of this parameter, or on its dependence on vegetation, soil, and/or climate characteristics, due to the lack of concomitant observations of *ET* and soil moisture at similar spatial scales. New ways to analyze data from existing networks are needed. *Salvucci* [2001] for instance present a method to estimate the storage-dependency of losses (including *ET*) in data-limited environments from sparse soil moisture observations and daily rainfall.

[4] Here we propose a method to derive this parameter from *ET* decay during wet-dry transitions in absence of rainfall. We analyze data from different sites with contrasting soil, vegetation, and climate characteristics. For two sites, the derived sensitivity parameters/decay timescales are compared to simulations with different Land Surface Models (LSMs) participating in the Second Global Soil Wetness Project (GSWP-2) [*Dirmeyer et al.* 2006]. Since no calibration has been performed for these sites, the (dis)agreement both among the different LSMs and with the observations is likely to be indicative for the current uncertainty in parameterization of *ET* sensitivity to soil moisture. A larger sample of comparison sites would be desirable for a more comprehensive analysis, but this is limited by the lack of *ET* observations in the GSWP-2 time period (1987–1996).

2. Data

[5] Daily time series of *ET* are analyzed for 15 vegetated sites, covering a wide range of geographical, climate, and land cover conditions (see Figure 1 and Table 1). Several sites (AU, BV, LW, IO, RC) are part of FLUXNET. For details see the project website (http://www.fluxnet.ornl.gov). For BV and LW, the data are extracted from the daily gap-filled Marconi Collection [*Falge et al.*, 2005]. At other sites the observations were part of shorter campaigns, e.g., FIFE (KP), SEBEX (SB), or GLOWA-Volta (EJ and TA). Most datasets have already been described in the literature (Table 1). In most cases, *ET* fluxes are derived from eddy covariance measurements. In some cases, they are derived by energy balance closure (SB, EJ, TA). All fluxes were measured above the canopy. At RD, *ET* was calculated from the daily change in lysimeter weight, corrected for drainage. All the data are converted to daily values and appropriate units. Occasional missing (half-)hourly *ET* values are linearly interpolated either from the daily course on the preceding and antecedent days (if not missing) or hours. Days with either too many missing values or minor rainfall are excluded from the analysis. We also utilize daily average observations of global and net radiation.

[6] Model data come from GSWP-2 (http://grads.iges.org/gswp2/). In this project different LSMs produced a Multi-Model land surface analysis on a 1° global grid using

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0094-8276/06/2006GL028178$05.00

Figure 1. Location of the sites mentioned in this study. See Table 1 for abbreviations.

the same forcing and boundary conditions [*Dirmeyer et al.*, 2006]. Here we use the baseline integrations (B0) for RD (51°N, 6°E) and SB (13°N, 2°E) both for the individual LSMs and the Multi-Model analysis. For details on GSWP-2 and the individual models we refer to the project website. Additionally, we also analyzed GSWP-2 simulations with TESSEL (the ERA40 LSM).

3. Theory

[7] Here we consider a vegetated land surface with deep groundwater (i.e., little capillary rise). Since vegetation water storage is much smaller than soil moisture storage in the root zone (order of mm), we neglect its changes during drydown. Under conditions where root water uptake by vegetation for ET is limited by the availability of soil moisture in the root zone, we assume a proportionality between ET and the available soil moisture storage S:

$$ET(t) = c\,S(t), \tag{1}$$

where the proportionality constant c is the sensitivity of ET for S. For convenience, we will consider the inverse of c, i.e., $\lambda = 1/c$, which has the dimension of time.

[8] The (simplified) terrestrial water balance can be written as:

$$\frac{\mathrm{d}S(t)}{\mathrm{d}t} = P(t) - q(t) - ET(t), \tag{2}$$

where P is rainfall and q is drainage. In absence of rainfall ($P = 0$), the soil rapidly drains to field capacity below which water is held against gravity ($q = 0$). In this case, (2) reduces to:

$$\frac{\mathrm{d}S(t)}{\mathrm{d}t} = -ET(t). \tag{3}$$

[9] By combining (1) and (3) it follows that under rainless conditions:

$$ET(t) = ET_0 \exp\left(-\frac{t - t_0}{\lambda}\right), \tag{4}$$

where ET_0 is ET at $t = t_0$. Exponential decay of ET in water-limited conditions has been reported previously from observations [*Williams and Albertson*, 2004; *Dardanelli et al.*, 2004] and has been widely applied in probabilistic soil moisture analysis [e.g., *Rodríguez-Iturbe and Porporato*, 2004]. The e-folding time λ (or $1/c$) is the key parameter controlling the temporal evolution of ET. From (4) it follows that λ can be estimated from observations of ET alone by (linear) regression of $\ln(ET)$ on t.

[10] In practice, the relation between S and ET (or t and $\ln(ET)$) is not perfectly linear. We use R^2 resulting from the regression between t and $\ln(ET)$ to indicate the tightness in the relation between S and ET. Obviously, the hydrological significance of λ not only depends on R^2 but rather on the range in observed ET (and thus soil moisture).

[11] In order to link λ (or c) to LSM soil physical and root parameters, it is useful to estimate the size of the

Table 1. Summary of the Datasets

Site	Abbreviation	Latitude	Longitude	Vegetation/Land Use
Audubon	AU	31.59°N	110.51°W	Desert grassland
Bondville	BV	40.01°N	88.29°W	Cropland
Ejura[a]	EJ	7.33°N	1.27°E	Transitional savanna
Ghanzi[b]	GZ-g	21.51°S	21.74°E	Grassland
Ghanzi[b]	GZ-m	21.50°S	21.75°E	Mixed (shrub/trees)
Ione-Tonzi Ranch[c]	IO-T	38.43°N	120.97°W	Oak savanna
Ione-Vaira Ranch[c]	IO-V	38.41°N	120.95°W	Annual grassland
Konza Prairie[d]	KP	39.08°N	96.56°W	Tallgrass prairie
Little Washita[e]	LW	34.96°N	97.98°W	Grassland
Rheindahlen[f]	RD	51.18°N	6.43°E	Grassland
Roccarespampani	RC	42.39°N	11.92°E	Deciduous broadleaf forest
SEBEX[g]	SB-s	13.25°N	2.28°E	Fallow savanna
SEBEX[g]	SB-t	13.25°N	2.28°E	Tiger-bush
Tamale[a]	TA	9.48°N	0.91°E	Guinea savanna
Twizel[h]	TW	44.23°S	170.15°E	Tussock grassland

[a]See *Schüttemeyer et al.* [2006].
[b]See *Williams and Albertson* [2004].
[c]See *Baldocchi et al.* [2004].
[d]See *Brutsaert and Chen* [1995].
[e]See *Meyers* [2001].
[f]See *Xu and Chen* [2005].
[g]See *Verhoef et al.* [1999].
[h]See *Hunt et al.* [2002].

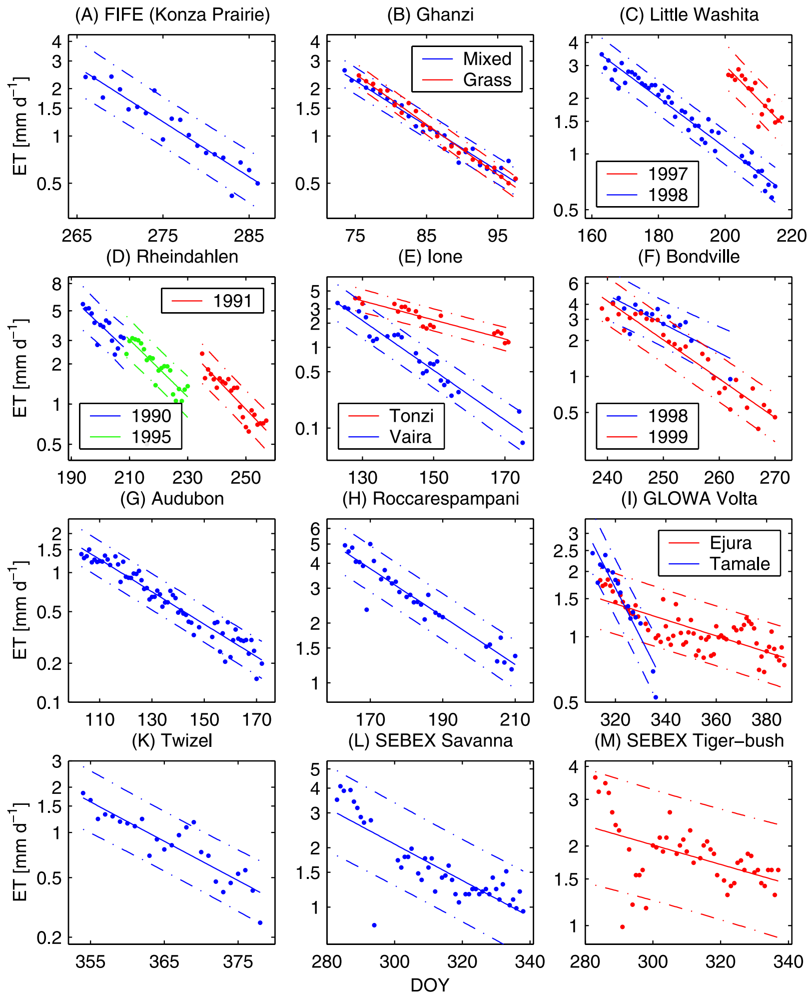

Figure 2. Daily *ET* versus day of year (DOY) for different sites under rainless conditions. Years and regression statistics are listed in Table 2. Note that the *y*-axis scale is logarithmic and that the scale of both axes can vary. Dashed lines correspond to the 95% prediction intervals.

storage volume S_0 that would be depleted during full drydown ($t \to \infty$) after the last rainfall event (at t_0):

$$S_0 = \int_{t_0}^{\infty} ET(t)\mathrm{d}t = \lambda ET_0, \tag{5}$$

where ET_0 is estimated from the regression.

4. Results

4.1. *ET* Observations

[12] The results of the linear regression between ln(ET) and t are shown in Figure 2 and are summarized in Table 2. At all sites, there is a strong relation between ln(ET) and t. High R^2 values (often >0.9) at many sites suggest the existence of a tight linear relation between S and ET. For three sites, namely BV, RD, and LW (Figures 2c, 2d, and 2f), we have data for multiple years. This allows us to test the hypothesis that c is different for every drydown. Interestingly, even with different ET_0, the interannual differences in λ at are non-significant at the 95% level. This means the hypothesis cannot be rejected based on the observational evidence we present. This implies that the assumption of constant (though site-specific) sensitivity of ET for S is reasonable.

Table 2. Results of the Regression Analysis[a]

		ET[b]		ET/R_g[c]		ET/R_n[d]		
Site	Year	λ, d	R^2	λ, d	R^2	λ, d	R^2	S_0, mm
KP	1987	12.4	0.89	15.8	0.85	22.3	0.80	31.5
BV	1999	13.4	0.92	15.5	0.85	18.9	0.86	61.5
IO-V	2001	14.1	0.94	13.4	0.94	15.1	0.94	49.8
GZ-g	2002	14.2	0.97	13.8	0.92	20.1	0.91	43.9
GZ-m	2002	16.2	0.96	16.7	0.95	21.3	0.92	52.3
TW	1998	16.6	0.80	15.3	0.82	16.7	0.74	28.1
BV	1998	18.0	0.68	18.8	0.56	22.5	0.73	81.7
TA	2002	19.0	0.88	20.8	0.96	25.7	0.87	51.3
RD	1991	19.1	0.84	26.5	0.80	33.8	0.61	109.
RD	1995	20.3	0.80	25.3	0.82	29.8	0.77	38.4
RD	1990	21.0	0.63	36.9	0.63	37.9	0.58	64.8
LW	1997	23.7	0.76	26.0	0.78	37.2	0.70	68.1
LW	1998	32.3	0.96	29.7	0.97	39.9	0.93	110.
AU	2004	34.7	0.93	32.9	0.94	34.7	0.89	53.7
RC	2003	36.0	0.91	36.8	0.92	35.2	0.91	164.
IO-T	2001	36.8	0.86	35.0	0.79	-	-	146.
SB-s	1989	47.2	0.69	50.7	0.64	78.3	0.46	141.
EJ	2002	117.	0.59	116.	0.50	−228.	0.22	175.
SB-t	1989	118.	0.26	136.	0.21	758.	0.01	274.

[a]The entries are sorted in order of decreasing sensitivity of evapotranspiration (ET) for soil moisture (increasing λ).
[b]Regression of daily ET with time.
[c]Regression of daily ET normalized by global radiation.
[d]Regression of daily ET normalized by net radiation.

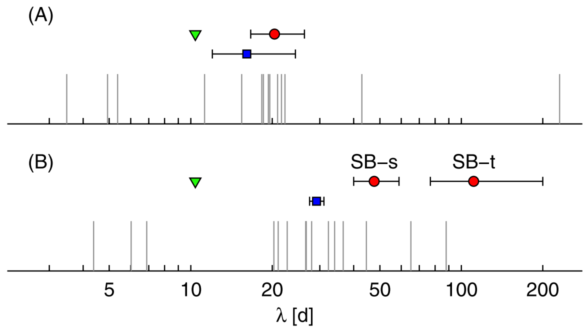

Figure 3. Distribution of λ for the GSWP-2 LSMs (vertical lines) at (a) Rheindahlen during the 1995 drought, and (b) SEBEX. Axis is logarithmic. Observations (circle) and GSWP-2 Multi-Model analysis (square) are both plotted with 95% confidence limits. Triangle indicates the *Dardanelli et al.* [2004] model.

[13] At GZ (Figure 2b), two sites only 2 km apart but with different land cover show similar drydown. At IO (Figure 2e), Vaira and Tonzi Ranch are also less than 2 km apart but show a distinct difference in drydown, with the woodland being more persistent. Note that in both cases, both sites would likely be assigned very similar root parameters. *Schenk and Jackson* [2002] report 50% and 95% rooting depths for temperate savannah of 23 and 140 cm, respectively, and for semi-desert scrubland 28 and 130 cm, respectively. AU (Figure 2g) has a low ET_0, likely due to low vegetation cover. Through its multiplicative effect on *ET*, the fraction of vegetation cover affects ET_0 rather than λ.

[14] There appears to be a stronger relation between λ and vegetation cover than between λ and soil type or depth. Small λ (<20 d) are found at grassland sites (KP, IO-V, TW), while larger λ (>30 d) are found at sites with tree cover (RC, IO-T) and/or sites that experience seasonal droughts (SB, AU, EJ). *Dardanelli et al.* [2004] report *c* to be 0.096 d^{-1} for several agricultural crops (corresponding to λ = 10.4 d). This is smaller than any of the values in Table 2, even for the cropland site BV. At the Sahelien sites SB and EJ, the large λ suggests that the perennial vegetation has adapted to the seasonal droughts by developing deep roots that prevent a rapid decay of *ET* during the dry season. The effect of deep roots on soil moisture and *ET* has been well studied for the Amazon [*Nepstad et al.*, 1994; *Bruno et al.*, 2006]. Interestingly, the large λ are in contrast with the very shallow soils (<0.5 m) that were reported for the SB sites [*Verhoef et al.*, 1999]. This suggests that the depth to which roots penetrate is not constrained to the shallow soil layer.

[15] To test whether the decay in *ET* is a soil moisture and not a radiation effect, we normalize *ET* both by daily average global and net radiation (R_g and R_n, respectively). R_g can be considered as the main external driving force of *ET*, since it does not depend directly on conditions at the land surface. In general, *e*-folding times of ET/R_g are close to those obtained by analysis of *ET* alone, with comparable R^2. Considerably larger *e*-folding times are found for ET/R_n, indicating a dependence of the denominator on *ET* (surface temperature and albedo increase during drydown). The RD lysimeter data enables us to check the validity of the zero drainage assumption. Indeed, drainage only accounted for a negligible 1.03% (1990), 0.02% (1991), and 0.29% (1995) of the total water loss. Not surprisingly, we find similar λ when *ET* is regressed directly against the lysimeter storage *S*: 20.0, 19.3, and 21.5 d, for the years 1990, 1991, and 1995, respectively.

[16] Table 2 also lists estimates of S_0 made by means of (5). These range from as little as 30 mm for shallow rooting grass vegetation to over 150 mm for sites with deep rooting trees. Note that S_0 should not be interpreted as the total storage in the root zone. Ideally, it is only the part in which storage limits *ET*. In most cases however, soil moisture might already have been limiting *ET* before the last rainfall. For TW, *Hunt et al.* [2002] report that 53% of the available storage was utilized before *ET* was reduced.

4.2. GSWP-2 *ET* Simulations

[17] Figure 3a shows λ derived from regressions between ln(*ET*) and *t* for all models for the 1995 drought at RD. The results for 1990 and 1991 are similar (not shown). For most of the models, λ compares well to the observations. This is also true for the Multi-Model analysis. Some models have low λ, even below the lowest value in Table 2. In contrast, some other models hardly show an effect of soil moisture depletion on *ET*. Since all models have the same forcing, inter-model differences in λ are likely (although not necessarily) caused by differences in soil moisture parameterization. Most R^2 values are smaller than those in the observations, suggesting that most models have a less strong coupling between *S* and *ET*. The *Dardanelli et al.* [2004] model significantly underestimates the observations at RD and is among the models with smallest λ.

[18] At SB, the situation is different (Figure 3b). Although most LSMs simulate a near-perfect exponential decay of *ET* (not shown), the variability in λ between the models is large. Some models reduce *ET* by 63% (1/*e*) in several days, while in other models this takes months. Although most models (and the Multi-Model analysis) exhibit larger λ for SB than for RD, there seems to be a structural underestimation of the observed λ for both the savanna (SB-s) and tiger-bush (SB-t) site. This structural difference can be explained by the fact that most models do not account for water uptake by deep roots.

[19] The inter-model variability is comparable to that found in previous LSM intercomparison studies. *Lohmann and Wood* [2003] reported composite *e*-folding times for ET/R_n for LSMs participating in the PILPS phase 2(c) Red-Arkansas River experiment. From their results, *e*-folding times can be derived ranging from 4.3 to 42.0 d with an inter-model coefficient of variation (*CV*) of 0.75. Even when only the significant regressions in our analysis are considered, the *CV* among the models is comparable: 0.59 and 0.72 for RD (1995) and SB, respectively. Although we test the GSWP-2 results only for 2 sites, the problems of some current generation LSMs under conditions where soil moisture limits *ET* are likely to be typical for many other regions.

5. Discussion and Conclusion

[20] The use of *ET* observations to determine the sensitivity *c* of *ET* to soil moisture storage has important

advantages. Firstly, the estimates are constrained by the water balance (3). Secondly, no soil moisture observations and root parameters are required. Thirdly, *ET* observed at a flux tower will reflect soil moisture dynamics at the scale of a footprint. Although both the size and the location of the footprint vary with atmospheric boundary layer conditions, the derived *c* will represent a much larger area than a point-scale soil moisture observation. A disadvantage of the method is that wet-dry transitions without significant rainfall might not occur frequently. Minor incidental rainfall (<1 mm) will be intercepted and readily evaporated, and will not influence the results as long as these days are excluded from the regression. More research is needed to test the robustness of our results against factors such as vegetation dynamics, seasonality, and landscape complexity.

[21] The timescales in Table 2 can serve as benchmarks for LSM performance. The confrontation of state-of-the-art GSWP-2 LSM simulations with observations for two sites suggests that the representation of soil moisture effects on *ET* can still be improved in many LSMs. For many regions, estimates of the parameter *c* from Table 2 might be used directly to diagnose *c* in LSMs. For regions such as the Sahel, it might prove necessary to include a parameterization that captures the effect of deep roots on *ET*. Calibration/validation of LSMs under more extreme conditions (like the wet-dry transitions studied in this paper) can lead to improved model-dependent parameters or states (e.g., "effective" rooting depth, leaf area index). This, in turn, might lead to more robust simulations of land surface hydrology under a range of (changing) climatic conditions.

[22] **Acknowledgments.** We thank John Albertson, John Hunt, Dirk Schüttemeyer, and Anne Verhoef for sharing their data with us. We thank the FLUXNET and GSWP-2 partners for making their data available on the internet. Marita Voogt is acknowledged for providing the TESSEL simulations. This research is supported by the Wageningen Institute for Environment and Climate Research (WIMEK) and the EU-project Development of a European Land Data Assimilation System to predict Floods and Droughts (ELDAS, EVG1-CT-2001-00050).

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