



Environmental drivers of evapotranspiration in a shrub wetland and an upland forest in northern Wisconsin

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[1] To improve our predictive understanding of daily total evapotranspiration (E_T), we quantified the differential impact of environmental drivers, radiation (Q), and vapor pressure deficit (D) in a wetland and upland forest. Latent heat fluxes were measured using eddy covariance techniques, and data from four growing seasons were used to test for (1) environmental drivers of E_T between the sites, (2) interannual differences in E_T responses to environmental drivers, and (3) changes in E_T responses to environmental drivers between the leaf expansion period and midsummer. Two simple E_T models derived from coupling theory, one radiation-based model, and another using mass transfer were used to examine the mechanisms underlying the drivers of E_T . During summer months, E_T from the wetland was driven primarily by Q , whereas it was driven by D in the upland. During the leaf expansion period in the upland forest the dominant driver was Q . E_T from the wetland was linearly related to net radiation using coupling coefficients ranging from a low of 0.3–0.6 to a high of 1.0 between early May and midsummer. Interannually, E_T from the upland forest exhibited near linear responses to D , with an effective reference canopy stomatal conductance varying from 1 to 5 mm s⁻¹. The results show that E_T predictions in northern Wisconsin and other mixed wetland-upland forests need to consider both wetland and upland forest processes. Furthermore, leaf phenology effects on E_T represent a knowledge gap in our understanding of seasonal environmental drivers.

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1. Introduction

[2] In mesic forested regions, evapotranspiration during the growing season can represent a large fraction of the precipitation received during the same period. As such, it is a crucial mechanism for both water and energy cycling, and a significant source of uncertainty in making predictions in watersheds lacking instrumentation [Sivapalan *et al.*, 2003]. Evapotranspiration (E_T) has two primary environmental drivers, radiation (Q) and atmospheric vapor pressure deficit (D). Q and D are inputs to the Penman-Monteith (P-M) combination equation [Monteith, 1965], which is routinely incorporated into large-scale models [Aber and Federer, 1992; Band *et al.*, 1993; Famiglietti and Wood, 1994; Wigmosta *et al.*, 1994; Vertessy *et al.*, 1996; Sellers *et al.*, 1997; Foley *et al.*, 2000]. The relative importance of Q and D as drivers of E_T in landscapes containing both upland forests and bottomland wetlands has generally not been considered, and yet these represent potentially large sources

of nonlinearity in the emergent landscape E_T . Models that are sensitive to these nonlinearities can potentially be used to fill in data gaps or as substitutes for lack of data in remote regions and will begin to provide predictive understanding of the interaction between E_T and climate change. In this paper we analyze the role of Q and D controls over E_T from multiple growing seasons of eddy covariance and micro-meteorological data from a wetland and an upland forest.

[3] E_T from vegetated land surfaces can be predicted from environmental drivers using the P-M combination equation:

$$E_{TPM} = \frac{s \cdot (R_n - G) + \rho_a c_p G_a D}{\rho_w \lambda [s + \gamma \cdot (1 + G_a / G_v)]} \quad (1)$$

where s is the rate of change of saturation vapor pressure with temperature [kPa °C⁻¹], R_n is net absorbed radiation [W m⁻²], G is ground heat flux [W m⁻²], ρ_a is air density [kg m⁻³], c_p is the specific heat capacity of air [J kg⁻¹ °C⁻¹], G_a [ms⁻¹] is aerodynamic conductance, D is vapor pressure deficit [kPa], ρ_w is density of water [kg m⁻³], λ is latent heat of vaporization [J kg⁻¹], γ is the psychrometric constant [kPa °C⁻¹], and G_v is a combination of leaf boundary layer (G_b) and canopy stomatal conductance (G_c):

$$G_v = \frac{1}{\frac{1}{G_b} + \frac{1}{G_c}}, \quad (2)$$

where $G_c = G_S * L$, L is leaf area [m² m⁻²], and G_S is canopy average leaf level stomatal conductance. The

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response of G_S to governing variables has previously been quantified by *Jarvis* [1976]:

$$G_S = G_{S_{\max}} f(R_n) f(D) f(T_A) f(\theta) \quad (3)$$

where $G_{S_{\max}}$ is maximum G_S , Q_o is photosynthetic photon flux density, T_A is air temperature, and θ is soil moisture. Formulations such as equation (3) are not strictly mechanistic, but they highlight potentially important drivers of G_S and consequently E_T . Equations (1) and (3) use R_n and D as direct drivers of E_T and as modifiers of stomatal control. T_A affects G_S through its role on leaf photosynthetic activity. θ affects root water availability and is thus a proxy variable for detecting water stress. Other factors could be included, including soil temperature, which influences root activity and consequently plant water uptake, especially at low temperatures or when soils are freezing.

[4] A more mechanistic view of controls on G_S follows from plant hydraulic theory. As a result of atmospheric dryness and/or high photosynthetic rates, woody plants experience water stress at high transpiration rates due to hydraulic limitations to water transport from the roots to the leaves [*Tyree and Sperry*, 1988; *Sperry et al.*, 1998]. Transpiration rates that exceed the plant's ability to transport water to the leaves cause leaf water content or potentials (Ψ_L) to decrease to a point where stomates close to prevent runaway cavitation. Although transpiration rates respond to D , stomata respond more directly to Ψ_L or transpiration rate rather than D [*Mott and Parkhurst*, 1991]. Available evidence suggests that plants regulate transpiration via changes in Ψ_L resulting from whole plant water status [*Meinzer and Grantz*, 1991; *Saliendra et al.*, 1995; *Cochard et al.*, 1996; *Nardini et al.*, 1996; *Salleo et al.*, 2000; *Ewers et al.*, 2000; *Franks*, 2004]. *Oren et al.* [1999] showed that

$$G_S = G_{S_{\text{ref}}} - m \cdot \ln(D) \quad (4)$$

where $G_{S_{\text{ref}}}$ is reference canopy stomatal conductance defined at $D = 1$ kPa, and $-m$ is the rate of change of G_S with respect to $\ln(D)$. The ratio of $-m$ to $G_{S_{\text{ref}}}$ has been shown using plant hydraulic theory to be between 0.54 and 0.6 when plants are isohydric and regulating minimum leaf water potential to prevent excessive cavitation [*Oren et al.*, 1999; *Ewers et al.*, 2005]. $G_{S_{\text{ref}}}$ incorporates $G_{S_{\max}}$ and all other limits to stomatal conductance other than D . Consequently, an analysis of drivers of E_T should consider all these variables.

[5] Previous researchers have examined a variety of models for estimating E_T in wetlands [*Drexler et al.*, 2004; *Rosenberry et al.*, 2004] and have found no ideal model. However, radiation-based methods outperformed mass transfer methods when the mass transfer component was relatively small and thus had amplified uncertainty associated with it [*Rosenberry et al.*, 2004]. Assuming G is negligible with respect to daily total energy flux [*Amiro and Wuschke*, 1987], and following the 1983 work of *McNaughton and Jarvis* (as discussed by *Monteith and Unsworth* [1990]), equation (1) can be rewritten in the form

$$E_{TPM} = \Omega \frac{s \cdot R_n}{\rho_w \lambda (s + \gamma)} + (1 - \Omega) \frac{\rho_a c_p G_v D}{\rho_w \lambda \gamma} \quad (5)$$

where Ω is a coupling coefficient, defined as $\Omega = \left[1 + \frac{\gamma}{s + \gamma} \left(1 + \frac{G_a}{G_v} \right) \right]^{-1}$, which varies between 0 and 1. The adiabatic term in equation (5) accounts for a lack of equilibrium between the state of the atmosphere at a reference height and the state of the evaporating surface through D . When the atmosphere and evaporating surface are in equilibrium then D approaches 0, and the adiabatic term becomes negligible. For a sufficiently large wetland surface equation (5) can be simplified to

$$E_{TW} = \Omega \frac{s \cdot R_n}{\rho_w \lambda (s + \gamma)}. \quad (6)$$

Note that Ω varies with the ratio of aerodynamic to canopy conductance, and so it retains the physical meaning of these conductances. A low-stature shrub wetland might be considered decoupled or only weakly coupled, in which case Ω tends toward 1 [*Monteith and Unsworth*, 1990]. Since Ω declines with canopy conductance it also retains the physical meaning of G_v , or some counterpart such as soil vapor conductance.

[6] For tall stature, closed forest canopies with a high G_a and low Ω , we assume that the mass transfer contribution to equation (5) is relatively large in comparison to the radiation contribution [*Jarvis and McNaughton*, 1986]. If this is the case for our upland forest we can simplify equation (5) to an upland evapotranspiration model (E_{TU}):

$$E_{TU} = \frac{\rho_a c_p G_v D}{\rho_w \lambda \gamma} \quad (7)$$

in which E_{TU} is proportional to D and scaled by G_v . When plants are regulating minimum leaf water potential to prevent excessive cavitation (isohydric plants), the relationship between E_{TU} and D will saturate or even decline at higher D [*Jarvis*, 1980; *Pataki et al.*, 2000; *Ewers et al.*, 2005].

[7] Boreal forests, mixed forests bordering the south shores of the Great Lakes and along the Appalachian Mountains in North America, and similar forests of this climatic regime consist of a continuum of moisture regimes from lowland forested and shrub wetland to mesic, upland forests [*Baldocchi et al.*, 1997; *Fassnacht and Gower*, 1997]. Northern Wisconsin forests lie at the boundary between northern temperate and boreal systems. Their moisture regimes are largely determined by glacial deposits that produced a topography consisting of wetlands surrounding low-rising upland forests at 10–20 m elevation above the wetland areas. This subtle topographic relief has produced, in a first-order analysis, a bimodal pattern consisting of wetland and upland forests, with a deterministic length scale [*Seyfried and Wilcox*, 1995] on the order of 10^2 m [*Burrows et al.*, 2002]. Sparsely vegetated wetlands represent one end-member in which Q is potentially the key driver of E_T , while for closed canopy upland forests that are aerodynamically rough the more important driver may be D .

[8] Given the above theory we tested four hypotheses related to daily E_T . (1) E_T from the wetland site is driven by variations in solar radiation (i.e., $E_T = E_{TW}$), while E_T responds primarily to D in the upland hardwood stand (i.e., $E_T = E_{TU}$). (2) The primary environmental driver of

Table 1. Measured Precipitation and Temperatures for the Study Sites^a

Year	Wetland Precipitation, mm		Forest Precipitation, mm		Temperature, °C	
	May	JJA	May	JJA	May	JJA
2000			54	279	12.8	17.0
2001	88	218	98	210	12.3	18.6
2002	167	260	98	323	7.8	17.7
2003	144	185	106	195	11.2	17.7
2004	74	170	83	218	9.2	15.5

^aJJA is sum of June, July, and August precipitation or average of June, July, and August temperatures.

E_T at either site is invariant among years. (3) Because of phenological changes in leaf area, evapotranspiration in the upland varies between radiation driven and vapor pressure deficit driven. (4) The primary environmental driver of E_T does not change in the wetland during the growing season. To test these hypotheses, we present four growing seasons of eddy covariance latent heat flux data collected on towers situated in a wetland and an upland forest in northern Wisconsin. We then determine the key driver, Q or D , of flux data for each tower year, and on the basis of this analysis we apply the appropriate model to help explain observed intersite, interannual, and interseasonal variations in E_T .

2. Methods and Materials

2.1. Site Descriptions

[9] We made eddy covariance and micrometeorological measurements in two ecosystem types within the Chequamegon-Nicolet National Forest in northern Wisconsin. The area is situated in the Northern Highlands geographic province of Wisconsin. The bedrock is composed of Precambrian metamorphic and igneous rock and overlain by 8 to 90 m of glacial and glaciofluvial material deposited approximately 10 to 12 kyr before present. Topography is slightly rolling with a range of 45 m within the defined study area. Outwash, pitted outwash, and moraines are the dominant geomorphic landforms. Annual precipitation (1971–2000 average) is about 800 mm, with mean January and July temperatures of -12°C and 19°C , respectively [Burrows *et al.*, 2002]. There is no marked dry season, which ensures that even the upland systems in the region are well watered [Ewers *et al.*, 2002; Cook *et al.*, 2004].

[10] The Lost Creek (hereafter called wetland) AmeriFlux site is located at $46^{\circ}4.9'\text{N}$, $89^{\circ}58.7'\text{W}$. Vegetation is a shrub height of 1–2 m, consisting of an overstory of speckled alder (*Alnus regosa*) and willow (*Salix* spp), and an understory of sedge (*Carex*, spp.). Soils consist of poorly drained Totagatic-Bowstring-Ausable complex and Seelyeville and Markey mucks formed on outwash sand, and are composed primarily of sapric material about 0.5 m thick [Natural Resources Conservation Service, 2006]. The Willow Creek (hereafter called upland) AmeriFlux site is a mature, second-growth hardwood forest about 70 years old, which is located at $45^{\circ}48.47'\text{N}$, $90^{\circ}04.72'\text{W}$. Dominant overstory species at this site are sugar maple (*Acer saccharum* Marsh), basswood (*Tilia Americana* L.), and green ash (*Fraxinus*

pennsylvanica Marsh), with an average canopy height of approximately 24 m. Leaf area index of $5.3\text{ m}^2\text{ m}^{-2}$ was measured [Desai *et al.*, 2005] at the site during the period of flux measurement reported here. Soils consist of sandy loam overlying coarse glacial till. A detailed description of the site is given by Cook *et al.* [2004].

2.2. Flux and Environmental Measurements

[11] Three-axis sonic anemometers (Campbell Scientific Inc., Logan, Utah, Model CSAT) and closed path infrared gas analyzers (Li-Cor Inc., Lincoln, Nebraska, model LI-6262) were deployed above canopy at a height of 10.2 m in the wetland and 29.6 m in the upland forest. Continuous measurements have been made since mid-1999 and late 2000 in the upland and wetland, respectively. Latent heat fluxes were calculated using established methods [Berger *et al.*, 2001; Cook *et al.*, 2004]. A detailed discussion of these calculations, spectral corrections, storage fluxes, screening for instrument error and low friction velocity, and quality control per an AmeriFlux relocatable reference system for the upland site are given by Cook *et al.* [2004]. The same methods were applied at the wetland site.

[12] Basic micrometeorological measurements, including air temperature (T_a) and relative humidity, precipitation, irradiance, R_n , and surface soil temperature (T_s) were made at each tower [Cook *et al.*, 2004]. T_s was taken at the soil surface in all years except in 2004 when the wetland peat subsided by approximately 20 cm. Continuous measurements of water table height (Z_w) were made in the wetland using a submerged pressure transducer (Omega Engineering, Stamford, CT, model PX242-005G). Measurements of soil moisture (θ) at 5 cm below the soil surface were made in the upland with a horizontally installed water content reflectometer probe (Campbell Scientific, Logan Utah, model CS615). For the purpose of our analysis we focused on the period from 1 May to 10 September. Data for the upland forest was available for years 2000 to 2003, and for the wetland data was available for 2001 to 2004. Growing season precipitation for all years was within one standard deviation of the 30 year average (48 ± 12 cm) at the National Climatic Data Center station in Minocqua, WI. Among the years in this study, 2000 and 2002 were relatively wet (53.8 cm and 53.0 cm, respectively), 2001 was near average (46.6 cm), and 2003 and 2004 were relatively dry (38.0 and 36.0 cm, respectively). Table 1 summarizes the precipitation and temperatures for May and for June–August for each site.

[13] Our criteria for selecting days for analysis were as follows. Days in which rainfall exceeding 5 mm was recorded between 6 pm of the previous day and 6 pm of the current day were not considered. When a day was missing more than one consecutive midday (8 am to 4 pm) half-hourly flux measurement it was not used. We did not adjust this range to account for changes in day length, as even midsummer before 8 a.m. and after 4 p.m. latent heat fluxes were relatively small in comparison to midday fluxes. More than one consecutive missing observation was accepted before 8 a.m. and after 4 p.m. when latent heat fluxes were on average less than 15% of the fluxes during the midday period and therefore not expected to contribute a large amount of error to the daily sum of evapotranspiration. Single missing observations during midday were corrected

using mean diurnal variations [Falge *et al.*, 2001] when multiple days with similar light and vapor pressure deficit (VPD) conditions were available. When data from similar light and VPD conditions were not available to fill in gaps we replaced the missing observation with the average of the fluxes from one observation prior to and one observation following the missing observation. This approach was used only when increases or decreases in light levels or VPD over the averaging time period were less than 20%.

[14] Half-hourly latent heat flux (LE ; W m^{-2}) was converted to water depth equivalent (mm) flux footprint evapotranspiration as follows:

$$E = \rho_w^{-1} \lambda^{-1} LE \quad (8)$$

where λ is the latent heat of vaporization calculated as a function of air temperature at the respective tower measurement heights on a half-hourly basis. Daily total tower evapotranspiration fluxes, E_D , were aggregated from half-hourly E obtained over daylight hours as follows:

$$E_D = \sum_{i=b}^e E(i) \quad (9)$$

where b and e respectively refer to time at the beginning and end of day, adjusting for day length changes from early May through mid-September. We adjusted b and e so that they delimited a daylight period. For the remainder of the analysis E_D will refer just to these aggregated measurements of E_T .

[15] R_n was measured at the top of the upland and wetland towers using CNR1 radiometers (Kipp and Zonan Inc.), and Q was measured at the top of the upland and wetland towers using a CNR1 radiometer and silicon pyranometer (Li-Cor, Lincoln, NE, model LI-200X), respectively. D was calculated from relative humidity and air temperature measurements [Goff and Gratch, 1946] obtained just below the top of the canopy in the upland forest and about 8 m above the vegetation in the wetland. Where necessary we gap-filled values of D to ensure a more complete data set. However, the use of gap-filled values of D did not influence gap filling of E , since in the cases where D was gap-filled we relied only upon observations of Q and not D at the respective towers to guide gap-filling of E . It should be noted that a large number of such gap-filled values can potentially bias the analysis of E in response to D . However, the number of such gaps was small and values of D among sites were very similar [Mackay *et al.*, 2002]. Gaps were filled using linear fits to the AmeriFlux WLEF tower (30 m above ground) [Davis *et al.*, 2003], to the upland tower in the case of the wetland, to the wetland tower for the upland, to four micrometeorological stations located in red pine, alder, mixed species, and aspen stands (at 1.5 m above ground) near the WLEF tower [Cook *et al.*, 2004; Mackay *et al.*, 2002], or to diurnal average values obtained at each respective tower. To determine daily mean D we retained only days in which either the maximum recorded half-hourly D exceeded 0.6 kPa [Ewers and Oren, 2000], or the daily average D (D_D) was at least 0.1 kPa [Phillips and Oren, 1998]. This screening process generally eliminated only days that were immediately preceded by

nighttime rainfall, and it was applied at both the wetland and upland sites using the same thresholds of D . This conservative approach ensured that our analysis was not based on days with very low D which tend to correspond with erroneous flux measurements [Ewers and Oren, 2000]. D_D was determined as an average of the half-hourly D values and daily Q (Q_D) as the sum of half-hourly radiation values ($\text{W m}^{-2} 30 \text{ min}^{-1}$) between times b and e . In addition, we determined daily average T_A , T_S , and Z_W in the wetland or θ in the upland. The daily variables were used for statistical analyses, but all calculations using equations (1)–(5) were made half-hourly and summed to daily.

2.3. Seasonal Definitions

[16] For each site year we divided the flux values into two groups. The first group (spring) spans a period from preleaf out (1 May) to about mid-June. By mid-June full leaf expansion has generally taken place in northern Wisconsin. The end date was determined partly by breaks in the flux data, with the constraint that the same data was used for each year for a flux tower. For Willow Creek we used 9 June as the leaf out period end date, while for Lost Creek we used 14 June. The second group (summer) extends from the end of spring to about 10 September in any given year.

2.4. Statistical Analysis

[17] Statistical analyses were performed using SAS (version 9.1, SAS Institute, Cary, NC, USA); Proc Reg was used for stepwise multivariate regression. Linear and nonlinear curve fits were performed in SigmaPlot (version 9.01, Systat Software Inc., Richmond, CA, USA). Curve fits were performed on individual groups of flux measurements and then Student's T was used to test for differences in slopes and intercepts among groups.

[18] From equations (2), (4), and (7) it is clear there is nonlinear response of E_{TU} to D , which can be closely approximated by an exponential rise to a maximum (*i.e.*, $E_{TU} \approx a \cdot (1 - \exp(-bD))$, where a and b are fitting parameters [Ewers *et al.*, 2005]). We anticipated that a number of other factors may preclude detecting a nonlinear response of E to D when measured from eddy covariance data. A relatively large, free, unrestricted evaporation source in the flux footprint would demonstrate a linear response of evaporation to D , which could mask or even hide the hydraulically limited signal (equation (4)) of the trees. Also, a set of observations made over a narrow and low range of D can produce a near-linear response of E_T to D because Ψ_L will not be low enough to trigger stomatal closure.

2.5. Modeling Analysis

[19] As further evaluation of the environmental controls on evapotranspiration, we applied equation (6) to the 1 May to 10 September periods for each of the four years of Lost Creek data. Ω was adjusted weekly or when there were gaps in E_{TW} to minimize bias in E_{TW} versus E_D . We chose not to adjust Ω at shorter intervals to reduce the amount by which we were subjecting the fitting procedure to noise in the micrometeorological and flux data.

[20] To evaluate the drivers for the upland forest we employed equation (7). We calculated $G_b = 0.025 \text{ ms}^{-1}$ assuming an average leaf width of 0.06 m and mean sunlit

Table 2. Number of Days, by Year and Period, When Flux Data Were Used for the Analyses^a

	Year	May	June	July	August/September
Wetland	2001	20	16 ^b	5	33
Wetland	2002	22	22	17 ^c	13
Wetland	2003	16	23	18	25
Wetland	2004	19	19	23	21
Upland	2000	15	13	20	24
Upland	2001	13	12	6 ^d	17
Upland	2002	0	10 ^e	14	37
Upland	2003	17	17	6 ^f	13

^aFootnotes b–f indicate where gaps in the data are due to instrument failure. The remaining missing days are due to a relatively short period when meteorological conditions precluded using the flux data.

^bNo data available 21 June to 19 July.

^cNo data available 21 July to 10 August.

^dNo data available 8–30 July.

^eNo data available 18 June to 4 July.

^fNo data available 13–29 July.

hours wind speeds [Campbell and Norman, 1998]. Although boundary conductance varies with wind speed and leaf display this variation contributes little to total conductance in comparison to variations in stomatal conductance. We adjusted G_{Sref} (reference canopy stomatal conductance; equation (4)) among weekly intervals or where there were extended breaks in E_D . It should be noted that this adjustment of G_{Sref} also accounts for actual changes in L , which would occur through leaf phenology as well as interannual changes in leaf area. The variability of G_v partly reflects changes in both L and G_{Sref} , but there was insufficient leaf area data to adequately separate the effects of both variables and so we adjusted only G_{Sref} . Among years we adjusted the value of m (sensitivity of stomata to the rate of water loss; equation (4)), which has the effect of adjusting the curvature of the relationship between E_{TU} and D . Since equation (7) assumes a fully coupled canopy (i.e., $\Omega = 0$) we tested the validity of this assumption by inverting the full E_{TPM} formulation (equation (1)) to estimate G_v and then solved for Ω in equation (5) at $G_a = 0.2 \text{ m s}^{-1}$. To avoid declines in Ω due to water stress, which would falsely imply stronger coupling [Monteith and Unsworth, 1990], we limited this analysis to well water conditions between 1 May and 31 July. We also compared G_{Sref} values derived using equation (1) to the values derived using equation (7) for both spring and summer periods.

3. Results

3.1. Hypothesis 1: Environmental Drivers

[21] Overall energy balance was 72% at both the upland [Cook *et al.*, 2004] and wetland sites. Some researchers suggest that flux calculations should be corrected on the basis of energy budget errors [Twine *et al.*, 2000]. Such correction was not attempted as it was difficult to confirm that the energy imbalances were not partially due to errors in estimating available energy [Wilson *et al.*, 2002; Mahrt, 1998; Cook *et al.*, 2004]. Moreover, there was no guarantee that the latent and sensible heat fluxes had the same fetch. There was no indication at either site that energy closure was correlated with meteorological conditions, and so we suppose that the relationships between fluxes and drivers would only change in flux amplitude, not shape.

[22] Table 2 summarizes the number of flux days used for the statistical analysis. A small number of data gaps were due to the instruments being off-line. These gaps ranged from 2–4 weeks in length, depending upon when technicians could visit these relatively remote sites (5 hour drive from Minneapolis, Minnesota) to make instrument repairs. Shorter gaps of typically 1–5 days were due to our criteria for selecting days as outlined in section 2.2. With the exception of instrument failure there was generally a balanced sampling of days within and among years at both sites.

[23] The results of a stepwise multiple regression using Q_D , D_D , T_A , T_S , (Z_W or θ), $Q_D * D_D$, and $D_D * D_D$ as predictors of E_D is summarized in Table 3. An additional variable, Julian day (J_{day}), was included in the multivariate analysis to rule out the possibility that additional changes in the system were occurring during the analysis periods. Since we have a comprehensive set of environmental variables covered already, J_{day} can be thought as a proxy for changes in leaf area or at least effective leaf area over time [Samanta *et al.*, 2007].

[24] In all cases the quadratic terms either did not significantly ($P > 0.10$) explain variance in E_D or the variance explained was at most 1%. During the summer periods in all years in the wetland the most significant driver of the variance in E_D was Q_D . Correlation with D_D was indirectly related to correlation with Q_D , with D_D explaining only an additional 2–7% of variance in E_D . During 2003, T_A and T_S explained 9 and 6%, respectively, of the variance in E_D ($P < 0.0001$). However, for the spring period there was no consistent most significant predictor of E_D , with Q_D dominating in 2001 and 2004 and D_D dominating in 2002 and 2003.

[25] For the upland forest most of the summer E_D was best explained as a response to D_D , with less than 6% of the variance explained by adding in Q_D . T_S , T_A , and θ were significant ($P < 0.08$) in 2002, but each contributed less than 2% of the variance. Q_D was the dominant driver of spring E_D in years 2000 and 2001. We could not completely rule out T_S and θ during the spring period in the upland site. T_S was significant ($P < 0.0001$) in 2003 and contributed 51% of the variance, and θ was significant ($P < 0.07$) in 2001 and explained 9% of the variance in E_D during the spring period.

3.2. Hypothesis 2: Interannual Variability

[26] Linear fits for the most significant drivers of E_D in the wetland are shown in Figure 1. The apparent interannual

Table 3. Variance Explained in E_D by Environmental Drivers, Incoming Solar Radiation (Q_D), and Vapor Pressure Deficit (D_D)^a

Year	Lost Creek				Willow Creek			
	Spring		Summer		Spring		Summer	
	Q_D	D_D	Q_D	D_D	Q_D	D_D	Q_D	D_D
2000	-	-	-	-	0.51	0.21 ^b	0.78	0.87
2001	0.86	0.79	0.79	0.72	0.53	0.26 ^b	0.43 ^b	0.81
2002	0.75	0.75	0.84	0.76	-	-	0.66	0.76
2003	0.37 ^b	0.61	0.69	0.67	0.38 ^b	0.40 ^b	0.34	0.79
2004	0.60	0.60	0.89	0.65	-	-	-	-

^aNumbers in bold indicate the most significant driver of E_D for the respective year, site, and season. A dash indicates no data.

^bVariable was not significant ($P > 0.1$).

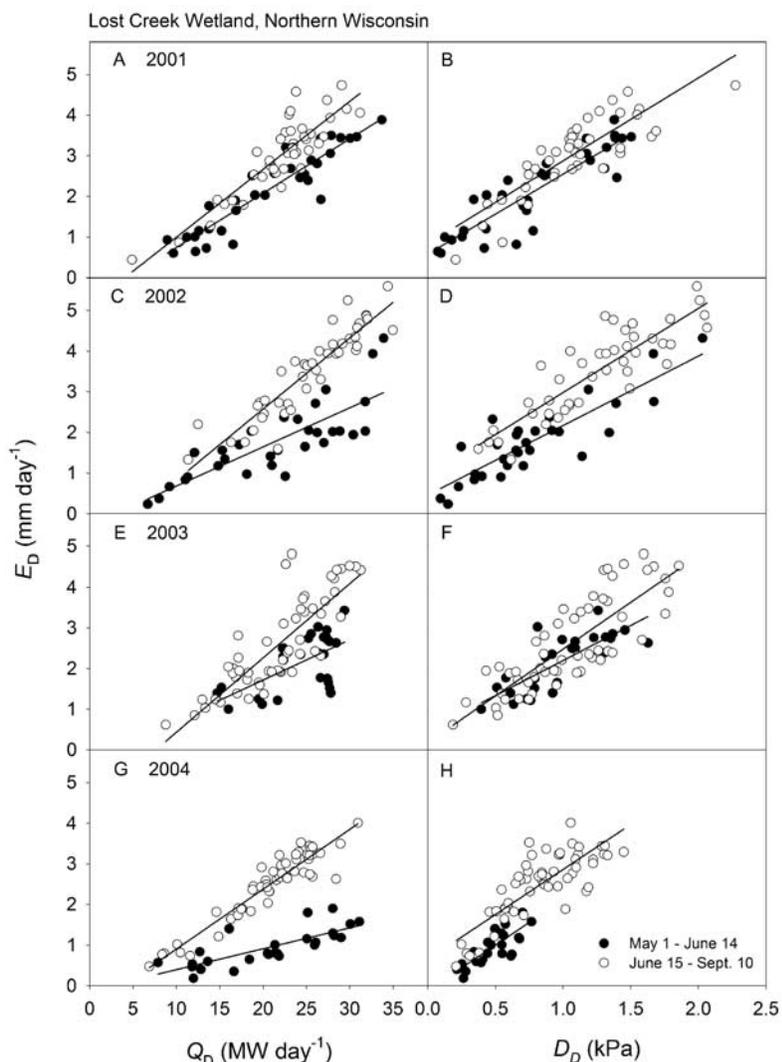


Figure 1. Response of E_D measured from eddy covariance in the wetland to daily above-canopy radiation (Q_D) and vapor pressure deficit (D_D) for years (a, b) 2001, (c, d) 2002, (e, f) 2003, and (g, h) 2004. All regressions are linear fits.

variability in E_D in response to Q_D was negligible among years 2001, 2002, and 2003 ($P > 0.2$ in all combinations). However, the response in 2004 was significantly different from the responses in the other years ($P < 0.1$).

[27] Figure 2 shows linear fits of E_D to Q_D and saturating nonlinear fits to D_D for the upland forest site. Although saturating curves explained slightly more of the variation in E_D than linear fits, this difference amounted to at most 2% of the total variation. However, an examination of the residuals among the linear and nonlinear fits showed a better fit with the saturating fits, which had both small mean residuals and more constant variance. Among linear fits of E_D versus D_D significant interannual differences were found between 2003 and other years ($P < 0.03$).

3.3. Hypotheses 3 and 4: Seasonal Variability

[28] As hypothesized, the driver of upland E_D changed from Q_D in the spring to D_D in the summer (Table 3). To compare curves we tested for significant differences in slopes among the same environmental drivers. With respect to responses to Q_D there were significant seasonal differ-

ences in years 2000 ($P < 0.001$) and 2001 ($P < 0.025$), but not in year 2003 ($P > 0.2$). With respect to D_D there were significant ($P < 0.001$) seasonal differences in slopes for all three years.

[29] The significant driver of spring wetland E_D changed among years, with Q_D being most important in 2001 and 2004, and D_D driving the flux in 2002 and 2003. There were also no consistent patterns in terms of the absolute values of explanatory variables during the spring period, but T_S explained 9% and 13%, respectively, of the variance in E_D during the 2003 and 2004 ($P = 0.002$) spring period. We also could not rule out J_{days} , our proxy for phenology, which was significant ($P = 0.0009$) in 2001 and 2003, and explained 15% of the variance in 2003. Z_W was generally not significant, except in 2002 ($P = 0.03$) when it explained 3% of the variance in E_D .

3.4. Modeling Evaluation of Environmental Drivers

[30] Comparisons of modeled E_{TW} versus measured E_D are shown in Figure 3. The predicted evapotranspiration closely matched the observations in terms of high degree of

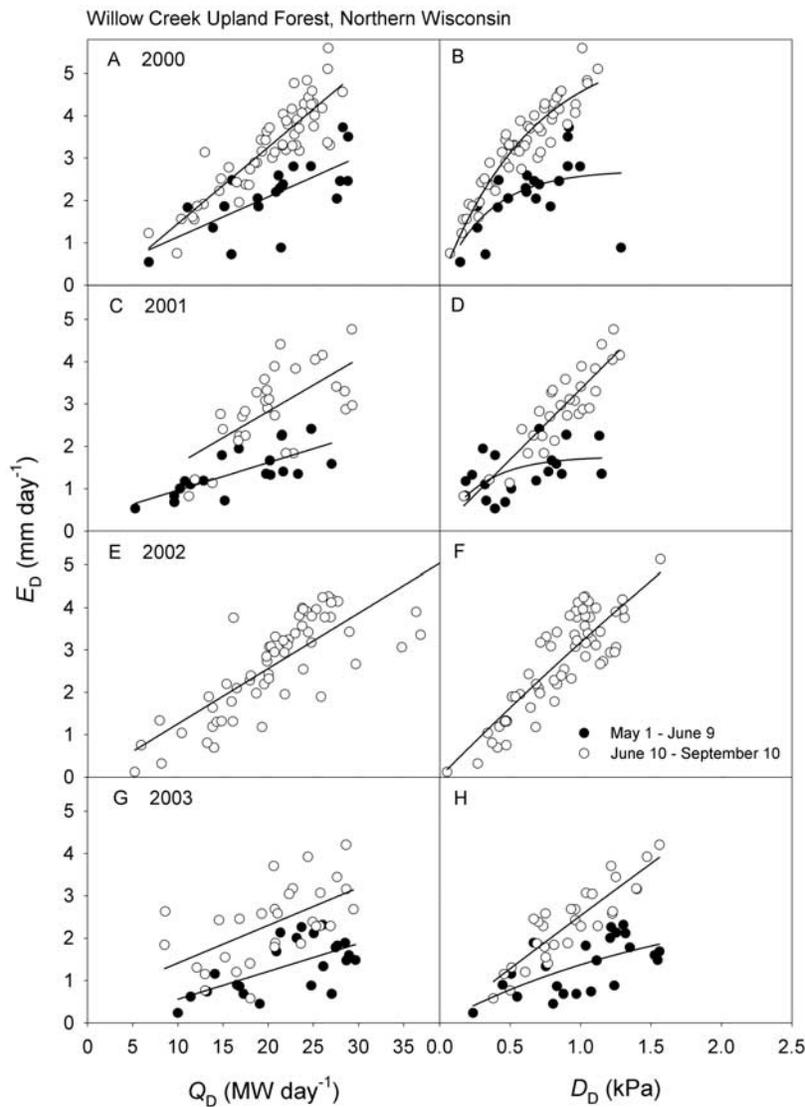


Figure 2. Response of E_D measured from eddy covariance in the upland forest to above-canopy radiation (Q_D) and vapor pressure deficit (D_D) for years (a, b) 2000, (c, d) 2001, (e, f) 2002, and (g, h) 2003. Curves in Figures 2a, 2c, 2e, and 2g are linear fits. Curves in Figures 2b, 2d, 2f, and 2h are exponential saturation curves of the form, $Y = a(1 - \exp(-bX))$.

fit and low bias, although there was slight overestimation of low fluxes and underestimation of high fluxes in 2001 and 2002. Figures 4 and 5 show the values for the coupling coefficient, Ω , with T_a and Z_w , respectively. Midsummer values of Ω generally varied from 0.8 to 1.0, but were lower in the spring and late summer. The lower spring values (0.3 to 0.6) closely followed air temperature, with the lowest values occurring in 2004 during an unusually cool May with mean daily temperatures only slightly above freezing. Values for Ω decreased during periods of water table drawdown (Figure 5), especially in the late summer periods of 2003 and 2004.

[31] The results for the upland forest are shown in Figure 6. Good fits were achieved between the modeled and measured E_{TU} for each year, but there was a significant bias at low flux in 2002. The sensitivity of stomatal conductance to the rate of water loss (m) was $0.6 \cdot G_{Sref}$ in

2000 and 2001, $0.5 \cdot G_{Sref}$ in 2002, and $0.54 \cdot G_{Sref}$ in 2003. These values are within the range reported for a variety of woody species including northern hardwoods [Oren *et al.*, 1999; Ewers *et al.*, 2001; Wullschleger *et al.*, 2002; Addington *et al.*, 2004; Ewers *et al.*, 2005, 2007b]. Figures 7 and 8 show how G_{Sref} varies in comparison to T_a and θ , respectively. G_{Sref} was lowest during May and generally peaked in July. This trend was consistent with leaf phenology during May and early June, during which time the increasing G_{Sref} reflected increases in L as well as increases in stomatal conductance. The continued increase into July could not be explained from the data at the Willow Creek site. G_{Sref} had a lower peak in 2003, which correlated with a steady decline in surface soil moisture (Figure 8). When we inverted equation (1) to obtain G_v we obtained a mean $\Omega = 0.14$ with 90% of the values falling between 0.01 and 0.27. G_{Sref} derived using equation (7) was 1.6 times as

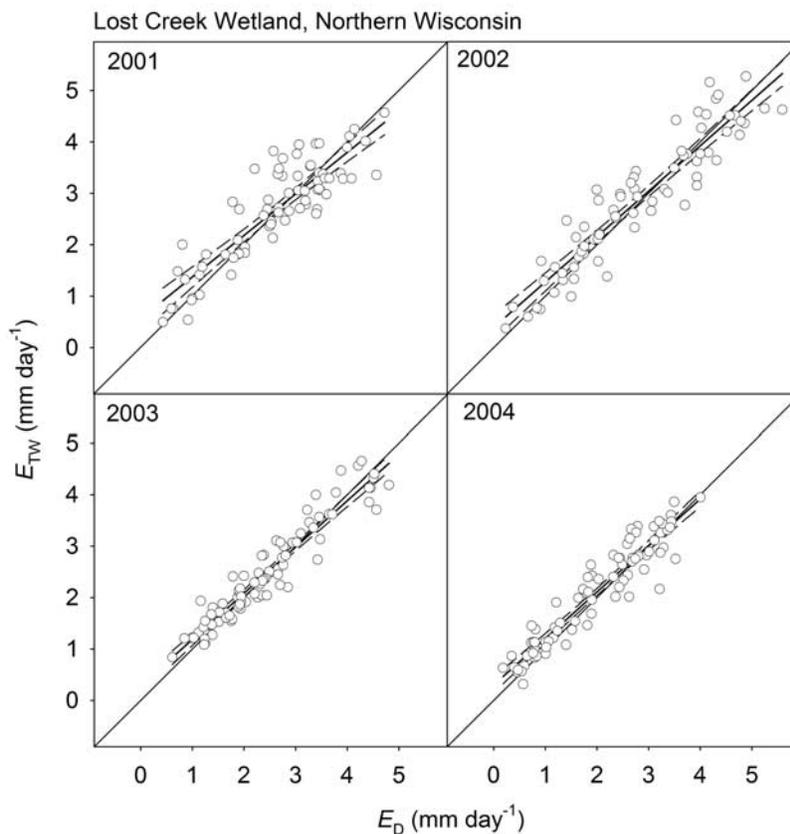


Figure 3. E_{TW} modeled on the basis of equation (3) versus E_D from eddy covariance in the wetland. Shown are linear regressions with dashed lines representing the 95% confidence intervals. Solid lines are one-to-one relationships.

large as G_{Sref} derived using equation (1) during summer months, and 2.4 times as large during spring.

4. Discussion

4.1. Hypothesis 1: Environmental Drivers of E_T

[32] Wetlands and upland forests represent end-members along edaphic gradients in northern Wisconsin, with transpiration dominating the evaporative signal in upland forests and soil evaporation dominating in wetlands. Hydrologic models for these types of systems may utilize just a single E_T formulation, a single environmental driver (e.g., radiation, vapor pressure deficit, temperature), or a full combination method to estimate E_T without knowing which environmental driver is dominant. We used four hypotheses to better understand the nonlinearities associated with these different environmental drivers in northern Wisconsin. Our first hypothesis, E_T in the wetland and upland forests is driven by variations in Q and D , respectively, was not rejected. The statistical analysis and simulations with equation (6) support the claim that the wetland E_T is driven primarily by Q , as has been demonstrated in other studies [Drexler *et al.*, 2004; Rosenberry *et al.*, 2004].

[33] The statistical analysis and simulations with equation (7) support the claim that upland E_T is driven primarily by D . The mean value of Ω (=0.14) supports the assumption of strong coupling for the upland stand [Jarvis and McNaughton, 1986]. However, by assuming fully coupled ($\Omega = 0$) instead

of fully P-M conditions, G_{Sref} was forced to compensate by increasing by a factor of 1.6 during the summer months and a factor of 2.4 during spring. This is due to the weaker E_T response to changes in G_S at higher values of Ω [McNaughton and Jarvis, 1991]. When G_{Sref} is determined using E_{TPM} then the values we obtain are similar to values reported for other sugar maple stands in northern Wisconsin and the upper peninsula of Michigan [Mackay *et al.*, 2003; Ewers *et al.*, 2007a, 2007b]. We note that an analysis using just equation (1) would mask the effects of drivers seen here. Moreover, an analysis with equation (5) requires simultaneous adjustments to partially correlated parameters, Ω and G_{Sref} , which potentially masks the relationships found here.

4.2. Hypothesis 2: Interannual Variability

[34] Our second hypothesis, that the primary driver at each site is invariant among years, was also supported by the available data. The Ω values for the wetland generally varied from about 0.6 to 1.0 in response to variations in soil surface temperature, water table depth, and potentially leaf phenology. This variability is consistent with changes in surface conductance [Monteith and Unsworth, 1990; L'Homme, 1997]. At peak water table levels (positive Z_W in Figure 5) or even shallow depths to the water table the surface soil may be considered above field capacity. One issue to consider further is that our wetland site is not a true well watered bare soil, an irrigated crop or grassland, a bog,

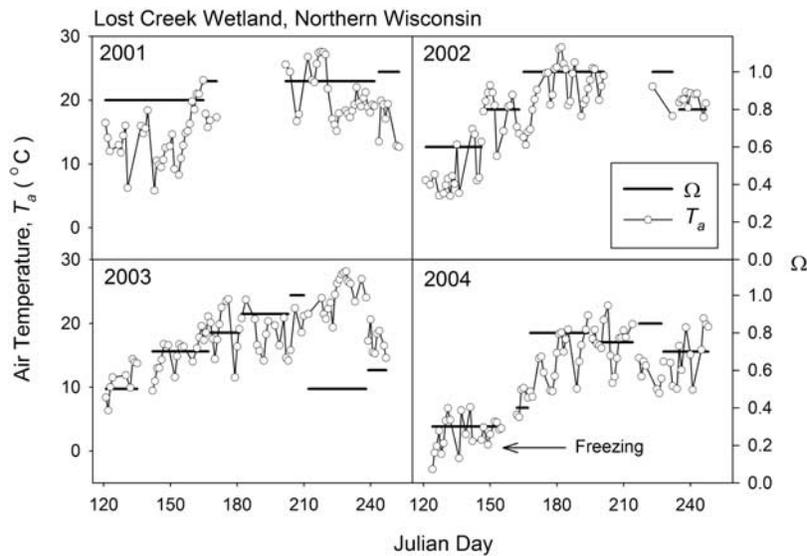


Figure 4. Adjustments to the coupling coefficient over time for each season at the wetland site. Also shown is air temperature at the site.

or a true forest. With its shrub height vegetation this system is perhaps more analogous to a rice paddy than these other systems. *Gao et al.* [2003] showed that rice growth affected aerodynamic roughness properties, but not energy partitioning patterns. Although our system is not a bog it shares some of the vascular plant characteristics reported by *Lafleur et al.* [2005], who show weak relationships between E_T and water table height in a system with a wider range of water table fluctuations than observed here. In our shrub wetland we could be seeing differences in the timing and rates of sedge, willow and alder leaf expansion among years, which would impact both aerodynamic and stomatal conductances. Further research, especially seasonal dynamics of leaf area and physiology, is needed to determine how leaf phenology may contribute to differences in Ω in such shrub wetlands.

[35] It is encouraging that with the exception of the dry 2003 the maximum G_{Sref} among years varied by only about 1 mm s^{-1} , suggesting that at least under optimal conditions this parameter is robust. Maximum G_{Sref} among years appears to be related to surface soil moisture, with the drier conditions in 2003 having the lowest peak G_{Sref} . *Ewers et al.* [2001] showed that G_{Sref} also declined with decreasing soil moisture in *Pinus taeda* while retaining the same ratio between $-m$ and G_{Sref} . Another possible interpretation of the relationship between G_{Sref} and soil moisture is that soil evaporation is contributing significantly to the midsummer peak flux, although it is generally relatively small under closed forest canopies [*Baldocchi et al.*, 2000] as has been shown in other sugar maple systems in northern Wisconsin [*Mackay et al.*, 2002]. We cannot rule out the possibility of additional sources of moisture. In particular, there is a

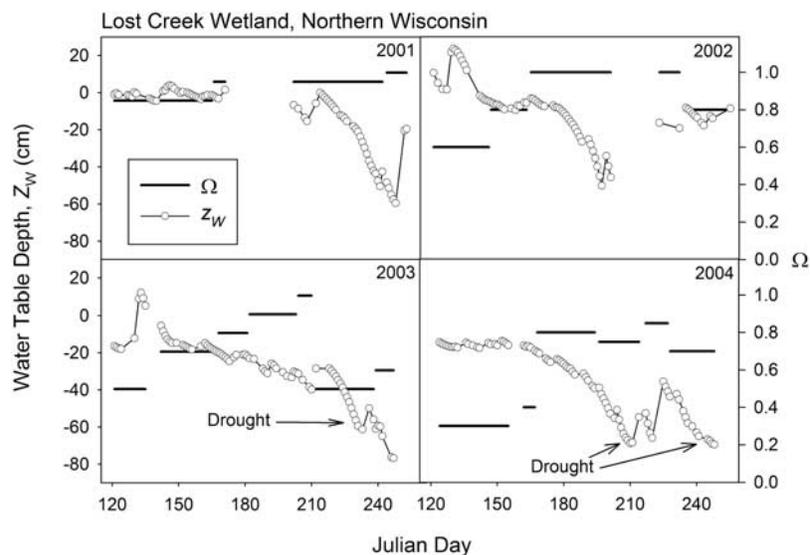


Figure 5. Adjustments to the coupling coefficient over time for each season at the wetland site. Also shown is water table height at the site.

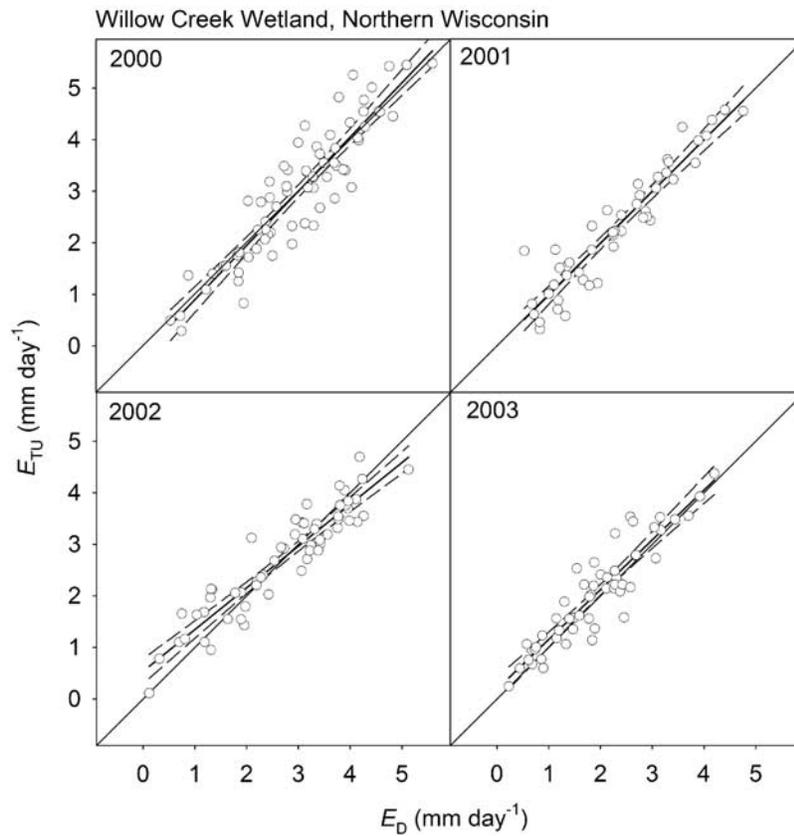


Figure 6. E_{TH} modeled on the basis of equation (4) versus E_D from eddy covariance in the upland forest. Shown are linear regressions, with dashed lines representing the 95% confidence intervals. Also shown are one-to-one lines.

wetland that is sometimes within the fetch of the Willow Creek tower depending on wind direction [Cook *et al.*, 2004; Desai *et al.*, 2005]. As such, it is possible that the G_{Sref} values here reflect the flux contributions from the

adjacent wetland. Further analysis of the flux footprint is needed to test this hypothesis.

[36] The peak E_T in 2001 was smaller than that for 2000 and 2002. During June 2001, a widespread outbreak of tent

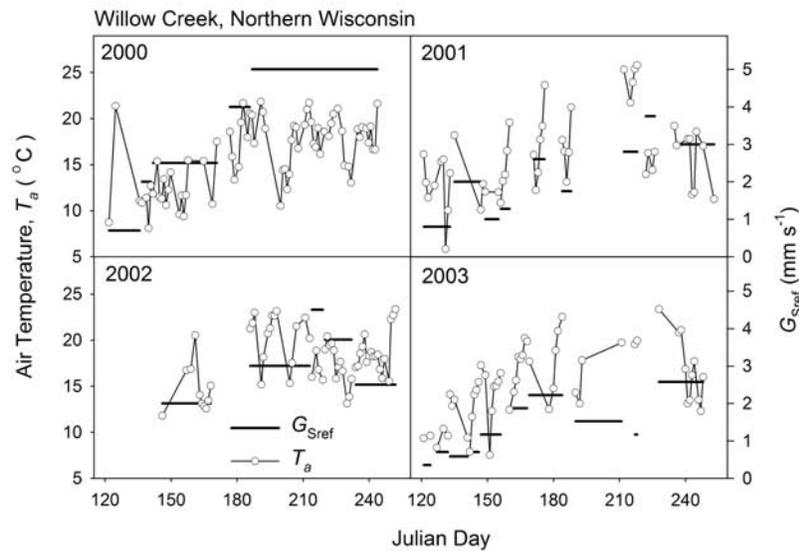


Figure 7. Variability in parameterized G_{Sref} over time for each season at the upland site. Also shown is air temperature at the site.

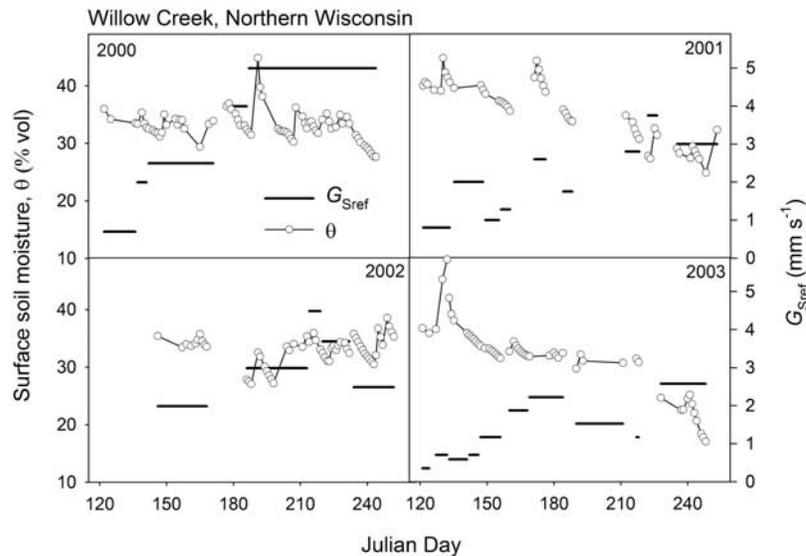


Figure 8. Variability in parameterized G_{Sref} over time for each season at the upland site. Also shown is soil moisture in the top 5 cm of the soil at the site.

caterpillars defoliated aspen and partially defoliated some other upland stands, including the Willow Creek stand. Reduction in G_{Sref} in 2001 is at least partly attributable to a reduction in L associated with the defoliation without a compensating effect from remaining foliage [Pataki *et al.*, 1998; Ewers *et al.*, 2007b]. In addition, volumetric soil moisture in the lower half of the rooting zone declined nearly monotonically from $0.30\ m^2\ m^{-2}$ in early July to $0.24\ m^2\ m^{-2}$ by late August [Cook *et al.*, 2004], whereas soil moisture remained above $0.30\ m^2\ m^{-2}$ for the whole 2000 growing season. The decline in rooting zone soil moisture during 2001 likely contributed to a small decline in G_{Sref} . Carryover effects of the defoliation and subsequent use of resources to grow new leaves in the same summer on reduced radial growth increment both during summer 2001 and early spring growth in 2002 could explain why total reference canopy stomatal conductance in 2002 did not recover to year 2000 levels.

[37] Year 2003 was the driest summer of the study period, and volumetric soil moisture in the top 100 cm declined steadily from an average of $0.30\ m^2\ m^{-2}$ in early July to $0.20\ m^2\ m^{-2}$ by late August. This drop in soil moisture can represent a 0.1 to 0.2 MPa [Clapp and Hornberger, 1978] decline in soil water potential, which may partly account for the lower E_{TU} response to D in comparison to the other years (Figure 2). This result is consistent with Ewers *et al.* [2007a] who showed a 25% decline in 2003 sugar maple G_{Sref} in comparison to 2002 G_{Sref} derived from inverting a canopy model driven by sap flux data inputs, although they found soil moisture to be a significant but small factor in this decline.

[38] The variability in m from 0.5 to 0.6 times G_{Sref} is well within that expected given the relatively short range of D over which most of the flux data values are distributed in some years, and is not likely attributed to changes in plant function. Even under a host of conditions affecting the interannual variability of G_{Sref} for a variety of northern Wisconsin species m was approximately equal to $0.6 \cdot G_{Sref}$ [Ewers *et al.*, 2007b]. One possible source of variability in

m is that nonstomatal sources of water are included within the flux signal. Although the near linear response of E_D to D_D (Figure 2) gives the appearance of nonstomatal sources of water, such linearity is also not found in species regulating leaf water potential. Ewers *et al.* [2005] found that old black spruce exhibited linear flux responses to D and Ogle and Reynolds [2002] found similar results in desert shrubs due to a lack of minimum leaf water potential regulation. If one adds to this open water sources and bryophytes without stomata, then systems such as the northern Wisconsin forests are especially complicated. A way around this problem is to distinguish these linearly responding systems from the nonlinear ones, by making a more thorough mapping of the component fluxes along moisture gradients.

4.3. Hypotheses 3 and 4: Seasonal Variability

[39] Our third hypothesis, that upland forest E_T is driven by Q during spring and D in the summer due to phenological changes was not rejected. During the spring phenology period upland E_T was explained by radiation. Three potential factors during phenological changes could explain the response at the upland site. First, with a more open canopy prior to leaf expansion a greater proportion of the total flux is expected to occur from below the canopy as a response to a relatively larger penetration of radiation to the forest floor [Baldocchi *et al.*, 2000]. Second, leaf budburst follows shortly after snowmelt, and so surface soil moisture content is high. During summer months soil evaporation in the hardwood stands comprises a relatively small (<10%) proportion of total evapotranspiration within the Chequamegon forest [Mackay *et al.*, 2002] and in other studies [Moore *et al.*, 1996; Kelliher *et al.*, 1995; Wilson *et al.*, 2000]. A third potential factor is that G_{Sref} during the period of leaf expansion may have been limited by low daytime air temperatures (Figure 7), low soil temperatures, nighttime freezing, or by limited development of gas exchange and photosynthetic capacity [Gratani and Ghia, 2002].

[40] To achieve a good fit between simulated and observed E_T at Willow Creek required us to make adjust-

ments to G_{Sref} . Seasonal variations in G_{Sref} generally followed leaf phenology for the region rather than tracking air temperature during the May–June period. An improved understanding of leaf phenology for the region would therefore reduce the variability of G_{Sref} . However, an explanation for why the increasing trend continued well into middle to late summer remains elusive. Ewers *et al.* [2007b] and Mackay *et al.* [2003] found a similar unexplained trend in sap flux data at another sugar maple stand in northern Wisconsin.

[41] Our fourth hypothesis, the primary environmental driver of E_T at the wetland site is always Q , was rejected because E_T was better explained by D in two of the springtime periods. However, there was insufficient data or evidence of changes in environmental conditions to fully explain why D was the dominant spring flux driver in 2002 and 2003. D_D may be statistically the more significant driver during these periods due to the relatively small sample sizes, 22 and 16 days, respectively. Moreover, evaporation was equally correlated to Q and D in 2002. This underscores the importance of recognizing that if D_D is the dominant driver, Q_D may also be strongly correlated. It also shows a need for information on how evapotranspiration from contrasting wetland and upland sites responds to changing environmental conditions during the spring-to-summer transition period.

[42] It is clear that the upland flux during May is lower than the JJA fluxes and seasonal wetland fluxes, suggesting that an improved understanding of leaf phenology and its effects on evaporative response to environmental drivers in the upland forests is needed. While measurements of eddy covariance have shown that water loss and carbon uptake at the stand level are correlated with leaf phenology [Goulden *et al.*, 1996; Granier *et al.*, 2000], these types of correlations have not been extensively tested and mechanistic connections have not been established [Turner *et al.*, 2003]. There is physiological evidence that such correlations between water and carbon fluxes and leaf phenology are not robust. For instance, Gratani and Ghia [2002] found increases in stomatal conductance and photosynthesis through the leaf expansion period. Furthermore, the susceptibility of larger xylem conduits to cavitation from freezing is well known [Sperry, 1995], but the repair of freezing induced cavitation is poorly understood [Hacke and Sperry, 2001]. Given the importance of phenology in regional and global-scale modeling [Myneni *et al.*, 1997; Schwartz, 1998; Menzel and Fabian, 1999; Schwartz and Reiter, 2000], and the limited ability of current models to predict phenology [Botta *et al.*, 2000], this represents an important area of ecohydrologic research.

5. Conclusions

[43] Our results show that estimates of forest evapotranspiration in northern glaciated and similar systems should be made using an understanding of both wetland and upland processes, and their responses to environmental conditions. Evapotranspiration fluxes from our two end-member sites representing upland closed forest and short stature vegetation-dominated wetland were more sensitive to vapor pressure deficit and radiation, respectively, during summer months. Our analyses also show that these results are primarily dependent on season, and the key drivers can change between the leaf expansion period and summer.

Evapotranspiration fluxes from our upland forest responded to radiation during the leaf expansion periods, whereas the key driver for the wetland site varied among years during the same periods. The results of this study show that when trying to determine evapotranspiration across a wetland-upland mosaic landscape, it is important to select models that are sensitive to the key drivers of evapotranspiration across the span of environmental conditions from upland to wetland sites.

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