Diurnal centroid of ecosystem energy and carbon fluxes at FLUXNET sites

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[1] Data from a network of eddy covariance stations in Europe and North America (FLUXNET) were analyzed to examine the diurnal patterns of surface energy and carbon fluxes during the summer period across a range of ecosystems and climates. Diurnal trends were quantified by assessing the time of day surface fluxes and meteorological variable reached peak values, using the "diurnal centroid" method; the diurnal centroid enabled us to discern whether the peak activity of the variable of interest is weighted more toward the morning or afternoon. In this paper, diurnal centroid estimates were used to diagnose which atmospheric and physiological processes controlled carbon dioxide, water vapor, and sensible heat fluxes across different ecosystems and climates. Sensitivity tests suggested that the diurnal centroids for latent (LE) and sensible (H) heat flux depend on atmospheric resistance, static stability in the free atmosphere, stomatal response to vapor pressure deficit, and advection. With respect to diurnal trends of surface energy fluxes at FLUXNET sites, maximum LE occurred later in the day relative to H at most tall forests with continental climates. The lag between LE and H was reduced or reversed at sites that were influenced by advection or by afternoon stomatal closure. The time of peak carbon uptake of temperate forests occurred earlier relative to the temporal peak of photosynthetically active radiation, as compared to boreal forests. The timing of this peak occurred earlier during periods with low soil water content, as it did during the summer in Mediterranean climates. In this case, the diurnal centroid for the CO_2 flux was influenced by the response of respiration and photosynthesis to increasing afternoon temperature and by afternoon stomatal closure. *INDEX TERMS:* 0315 Atmospheric Composition and Structure: Biosphere/atmosphere interactions; 1818 Hydrology: Evapotranspiration; 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; *KEYWORDS:* FLUXNET, energy flux, sensible heat, latent heat, evapotranspiration, diurnal

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1. Introduction

[2] The diurnal trends of energy and carbon exchange between vegetation and the atmosphere are signatures of important atmospheric and physiological processes that control biophysical fluxes at the surface. The diurnal pattern of partitioning between latent (LE) and sensible (H) heat depends on temporally evolving interactions between atmospheric demand, the net radiation, temperature, and atmospheric vapor pressure deficit (δ_e), and the ability of the vegetation to supply available water. Diurnal variations in the atmospheric demand are largely determined by the diurnal movement of the Sun and the development of the atmospheric boundary layer [*McNaughton and Spriggs*, 1985; *Betts*, 1994; *Lhomme*, 1997]. Alternatively, diurnal changes in the availability of water for evaporation are largely controlled by physiological response to environmental perturbations; they are manifested primarily by changes in stomatal conductance.

[3] Knowledge on the diurnal variation in latent and sensible heat exchange over vegetated surfaces has important implications on the growth of the planetary boundary layer and the exchange of heat across the entrainment zone [*Davis et al.*, 1997]. Daily variations in temperature and vapor pressure deficit are important consequence of the coupling between surface layer fluxes and growth of the convective boundary layer. Even if solar radiation is symmetrical about local solar noon, the higher afternoon δ_e should result in afternoon energy partitioning that favors LE relative to H [*Jarvis et al.*, 1976; K. B. Wilson et al., Surface energy partitioning between latent and sensible heat flux at FLUXNET sites, submitted to *Water Resources Research*, 2001, hereinafter referred to as Wilson et al., submitted manuscript, 2001]. This interaction is likely to be more important over tall vegetation, which is more closely coupled to changes in δ_e [*Jarvis and McNaughton*, 1986]. Over short vegetation, such as an agricultural crop, LE depends mainly on available energy and is less sensitive to changes in δ_e; thus the diurnal lag between LE and H may be less. In other cases, as when the vegetative region is surrounded by water or is an irrigated field [*Rosenberg and Verma*, 1978; *Brakke et al.*, 1978; *Prueger et al.*, 1996], the diurnal variation in atmospheric demand is controlled by advection rather than by the local surface fluxes.

[4] Diurnal variations in physiological processes [e.g., *Hinckley et al.*, 1978; *Schulze and Hall*, 1982; *Dolman and Van den Burg*, 1988; *Turner*, 1991] also influence the diurnal partitioning between H and LE at the canopy scale [*Berbigier et al.*, 1996; *Baldocchi*, 1997; *Anthoni et al.*, 1999; *Goldstein et al.*, 2000]. One example is the occurrence of afternoon stomatal closure in environments with low soil water content and/or high vapor pressure deficit [*Downton et al.*, 1987; *Weber and Gates*, 1990; *Whitehead*, 1998; *Goldstein et al.*, 2000]. As a result, transpiration may peak earlier in the day or reach an afternoon plateau, even though δ_e continues to increase [*Lynn and Carlson*, 1990; *Valentini et al.*, 1995; *Berbigier et al.*, 1996]. Theoretical studies of canopy gas exchange indicate that afternoon stomatal closure and a "plateau" or a decrease in transpiration is an optimal strategy during periods of soil drying [*Cowan*, 1982; *Makela et al.*, 1996].

[5] The diurnal pattern of net carbon flux (F_{CO_2}) depends on interactions between the atmosphere and physiological responses. Afternoon stomatal closure can have a dramatic impact on the diurnal cycle of assimilation, resulting in an early morning peak in F_{CO_2} a response which is often enhanced during periods of low soil water content [*Valentini et al.*, 1995; *Baldocchi*, 1997]. This is consistent with theoretical investigations of optimal stomatal control that indicate that afternoon CO_2 uptake should be increasingly suppressed as soil drying progresses [*Cowan*, 1982; *Makela et al.*, 1996]. The temperature-dependent biochemical reactions involving assimilation and respiration also directly contribute to altering the diurnal cycle of F_{CO_2} Ecosystem respiration, a significant fraction of net ecosystem carbon exchange, typically increases with temperature [e.g., *Hanson et al.*, 1993; *Schmid et al.*, 2000]. Hence higher afternoon temperatures lead to an increase in ecosystem respiration and decrease F_{CO_2} when light levels are unchanging [*Baldocchi*, 1994; *Anthoni et al.*, 1999]. Alternatively, the response of photosynthesis to temperature is not generally monotonic and depends on the biochemical adjustments that can influence the optimum temperature for photosynthesis [*Berry and Bjorkman*, 1980; *Morgan et al.*, 1994]. If leaves are below their photosynthetic optimum temperature in the morning, higher afternoon temperatures may increase assimilation and counteract the simultaneous increase in respiration. Alternatively, leaves that are operating near the optimal temperature early in the day may experience no change or a decrease in assimilation during the afternoon. This latter scenario will enhance the reduction in F_{CO_2} that may already be occurring in the afternoon because of increased respiration or stomatal closure. Differences in morning and afternoon cloudiness and changes in the relative fraction of solar radiation that is diffuse may also affect the diurnal pattern of F_{CO_2} [*Gu et al.*, 1999].

[6] FLUXNET, a worldwide network of stations measuring ecosystem fluxes using the eddy covariance method [*Baldocchi et al.*, 2001], provides a new opportunity to examine the diurnal variability of surface fluxes across a wide range of ecosystems and climates. In this study, we investigate the diurnal patterns of energy and carbon dioxide fluxes and meteorological variables across 20 sites and 57 site-years using the FLUXNET data set. Because of climate and ecosystem variability across the network, we expect differences in diurnal trends of surface fluxes, which represent signatures of important climatological and physiological processes that control surface exchange. This study focuses on relative diurnal trends of the component fluxes and not the magnitude or absolute partitioning of energy. A companion study (Wilson et al., submitted manuscript, 2001) summarizes the magnitude of turbulent energy fluxes and the Bowen ratio at the same sites. The primary goals of this study are to (1) to quantify diurnal trends in surface fluxes at the sites and (2) to evaluate if variability in diurnal trends may be used as a signature of the different important controlling physiological (e.g., afternoon stomatal closure) and climatological (atmospheric boundary layer development, advection) processes across ecosystems and climates. To assist in this second goal, sensitivity tests are performed using a simple big-leaf

Table 1. FLUXNET Sites Investigated in This Study

Site	Location	Period	Code	Coordinates	Species	Bowen
			Deciduous Forests			
Harvard	Massachusetts	1992–1999	A	42°32′N 72°11′W	Oak-Maple	0.46
Walker Branch	Tennessee	1995–2000	B	35°58′N 84°17′W	Oak-Hickory	0.34
Hesse	France	1996–1999	C	48°40′N 7°50′E	European Beech	0.42
Vielsalm	Belgium	1996–1998	D	50°18′N 6°0′E	European Beech	0.73
			Coniferous Forests			
Tharandt	Germany	1996–1999	F	50°58′N 13°40′E	Norway Spruce	0.73
WeidenBrunnen	Germany	1996–1998	I	50°09′N 11°52′E	Norway Spruce	1.34
Howland	Maine	1996–1997	K	45°15′N 68°45′W	Spruce-Hemlock	0.91
Duke Forest	North Carolina	1998–1999	L	35°52′N 79°59′W	Loblolly Pine	0.59
Bordeaux	France	1996–1997	M	44°0′N 0°5′E	Maritime Pine	0.82
North Boreas	Manitoba	1994–1997	N	55°54′N 98°30′W	Black Spruce	1.57
Aberfeldy	United Kingdom	1997–1998	O	56°37′N 3°48′W	Sitka Spruce	
Niwot Ridge	Colorado	1999	P	40°02′N 105°33′W	Engelman spruce, Lodgepole pine, Subalpine Fir	0.86
			Mediterranean Climates			
Blodgett Forest	California	1997–1999	Q	38°53′N 120°37′W	Ponderosa Pine	0.61
Metolius	Oregon	1996–1997	R	44°30′N 121°37′W	Ponderosa Pine	1.61
Castelporziano	Italy	1997–1998	S	41°45′N 12°22′W	Quercus. ilex	1.99
Sky Oaks[a]	California/USA	1997–1998	T	33°22′N 116°37′W	Adenostoma-Ceanothus-Chaparral	56.7
			Crops			
Bondville[b]	Illinois	1997–1999	U	40°0′N 88°18′W	Corn/soybean alternate	0.31
			Grasslands			
Little Washita	Oklahoma	1998–1999	Y	34°58′N 97°59′W	Rangeland	1.34
Shidler	Oklahoma	1997	Z	36°51′N 96°41′W	Tallgrass Prairie	0.34
Fort Peck	Montana	1999	&	48°19′N 96°6′W	grassland	0.53

[a]The Sky Oaks site includes two subsites; site-year T1 is an older chaparral in 1997. T2 and T3 are younger chaparral in 1997 (T2) and 1998 (T3).
[b]The Bondville site alternates between corn and soybeans; U1 and U3 are years when corn was grown and U2 was a year when soybeans were grown. The estimated values of the Bowen ratio are from Wilson et al. (submitted manuscript, 2001).

convective boundary layer (CBL) model, and the results are compared to the trends observed from data in the network.

2. Materials and Methods

[7] The FLUXNET methodology and summary studies from the network may be found in the work of *Aubinet et al.* [2000], *Valentini et al.* [2000], *Falge et al.* [2001], and *Baldocchi et al.* [2001]. Files containing eddy covariance and supporting environmental and meteorological data were contributed by individual investigators to the FLUXNET database at Oak Ridge National Laboratory's Data Archive Center (http://daac.ornl.gov/FLUXNET). Each file or "site-year" contributed by the investigators contains half-hourly flux and meteorological data for 1 year at eddy covariance stations across Europe and North America, which includes a range of vegetation types and geographic locations (Table 1). For each site, Table 1 shows the years of data analyzed in this study, along with a one-letter code used to denote the individual sites. The one-letter code and a subsequent numeral were a shorthand method for denoting the site and the year of measurement in this paper. For example, the code "A1" applies to site "A" (Harvard Forest) and "1" denotes the first year of data for this site (1992 in this case). Sites above 60°N were not used because the period of darkness was brief or nonexistent, so the diurnal cycle is less pronounced. Also, shown in Table 1 is the mean Bowen ratio (ratio of H to LE) for each site from Wilson et al. (submitted manuscript, 2001). The sites in Table 1 are classified into six main groups, deciduous forests, coniferous forests, agriculture, tundra, grasslands, and sites in Mediterranean climates. This last group, defined by sites with a distinctive summer dry season, is not a vegetation type, but justification for this group will be shown. Climate information for each site is presented in Table 2.

[8] All sites are in the Northern Hemisphere, and the study period in this research was isolated to the late spring/summer period, between day 165 (mid-June) and day 235 (late August). The summer season was chosen to constrain the scope of the study. Our goal is to isolate the period when solar radiation and temperature, the driving forces for energy, carbon, and water fluxes, are greatest. Furthermore, the study period was limited to daytime. This strategy was chosen because the eddy covariance method is more reliable during the daytime period [*Lavigne et al.*, 1997; *Anthoni et al.*, 1999; *Aubinet et al.*, 2000; *Wilson et al.*, 2002].

[9] The diurnal pattern of flux and meteorological variables was quantified using the "diurnal centroid":

$$C_{gt} = \frac{\int_{t_1}^{t_2} g(t)t\,dt}{\int_{t_1}^{t_2} g(t)\,dt}. \tag{1}$$

The diurnal centroid, C_{gt}, represents the mean daily time (t) weighted by the variable $g(t)$, which is either a meteoro-

Table 2. Climate Conditions at the Sites Used in This Analysis

	Annual Mean Temperature, °C	Annual Rainfall, mm
Harvard	7.8	1066
WalkerBranch	13.9	1372
Hesse	9.2	820
Vielsalm	8	1000
Tharandt	7.5	824
WeidenBrunnen	5.8	855
Howland	5.7	1201
Duke Forest	15.5	1140
Bordeaux	13.5	900
North Boreas	−3.5	419
Aberfeldy	8	1400
Niwot Ridge	0.56	201
Blodgett Forest	10.4	1630
Metolius	7.6	577
Castelporziano	15.3	700
Sky Oaks	12.2	500
Bondville	11.2	953
LittleWashita		
Shidler	14.6	881
Fort Peck		

logical variable or a mass/energy flux density. For example, if variable $g(t)$ is perfectly symmetrical about local noon, C_{gt} would be 12 hours (local noon). If variable $g(t)$ is greater in the morning, C_{gt} would be less than 12 hours, and if variable $g(t)$ is weighted more toward the afternoon C_{gt} would be greater than 12 hours. In this paper, equation (1) was evaluated only during the daylight hours.

[10] Missing data were not gap-filled [*Falge et al.*, 2001] to avoid assuming a priori relationships between environmental variables and flux estimates. Because data were not gap-filled, errors or biases in the flux estimates were still possible because the averaging method in this study ignores periods of missing data. Although this bias was recognized and may be important in some individual cases, it was not expected to significantly alter the general results.

[11] Three indices were calculated in this study to investigate how ecosystems differ in the diurnal patterns of fluxes. The first was the difference in the diurnal centroid between H and LE: $C_{gH} - C_{g\mathrm{LE}}$. Large negative values of $C_{gH} - C_{g\mathrm{LE}}$ suggest that LE is weighted more toward the afternoon relative to H, and vice versa. A second index that was computed was the difference in the diurnal centroid between the carbon dioxide flux ($F_{\mathrm{CO_2}}$) and photosynthetically active radiation: $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$. Large negative values of $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ indicate that CO_2 uptake occurred early in the day relative to PAR, often a major driving variable for CO_2 uptake, and vice versa. A third index, $C_{g\delta_e} - C_{gR_n}$, which is the difference in C_g for δ_e and net radiation (R_n), was computed to characterize the diurnal pattern of climate demand.

[12] Coupled surface layer-CBL model simulations were performed to demonstrate how atmospheric and physiological processes affect diurnal centroids. This task was accomplished by sensitivity tests on how diurnal patterns of LE and H responded to specified changes in atmospheric variables, surface exchange characteristics, advection, and stomatal response. The surface exchange module incorporates the big-leaf paradigm:

$$H = \frac{\rho C_p (T_s - T_a)}{R_a}, \tag{2}$$

$$\mathrm{LE} = \frac{\rho C_p \delta_e / \gamma}{R_a + R_c}, \tag{3}$$

where ρ is the density of air (kg m^{-3}); C_p is the specific heat of air (1005 J kg^{-1} K^{-1}); T_s and T_a are the "surface" and air temperatures (C); R_a is the atmospheric resistance (s m^{-1}); δ_e is the vapor pressure deficit (Pa); γ is the psychrometric constant (64 Pa K^{-1}); R_c is the surface resistance to water vapor transport (s m^{-1}). Upwelling longwave radiation was estimated from the simulated surface temperature using Stephan-Boltzmann's law and an assumed emissivity of unity. The downwelling longwave radiation was parameterized as a function of atmospheric temperature [*Monteith and Unsworth*, 1990]. R_a was parameterized as a function of wind speed (U), surface roughness (z_o), and stability from Monin-Obukhov similarity theory following the equations in the work of *Raupach* [1998].

[13] The surface fluxes from equations (2) and (3) served as lower boundary conditions for a CBL model, which simulated the growth of a bulk mixed layer. The mixed layer had a constant potential temperature and specific humidity with respect to height and was topped by a temperature discontinuity (inversion) in the entrainment zone. The equations for potential temperature in the mixed layer (Θ_m), the magnitude of the temperature discontinuity between the mixed layer and free atmosphere ($\Delta\Theta_m$), and the inversion level sensible heat flux (H_i) are calculated from *Tennekes and Driedonks* [1981], but modified to parameterize the horizontal advection of water vapor and heat:

$$\frac{d\Theta_m}{dt} = \frac{H - H_i}{\rho C_p h} + \frac{U}{d}(\Theta_e - \Theta_m), \tag{4}$$

$$\frac{d\Delta\Theta_m}{dt} = \Gamma \frac{dh}{dt} - \frac{d\Theta_m}{dt}, \tag{5}$$

$$H_i = \rho C_p \Delta\Theta_m \frac{dh}{dt}, \tag{6}$$

where Γ is the lapse rate of Θ in the free atmosphere and h is the height of the inversion. Horizontal advection (second term on right-hand side of equation (4)) was parameterized from dimensional analysis, using the mean wind speed (U) and the horizontal distance (d) to a region with a different potential temperature (Θ_e) and specific humidity. As U/d approaches zero, the system becomes closed horizontally, and the developing mixed layer is influenced only by surface fluxes and vertical entrainment from the overlying inversion (one-dimensional system). Alternatively, nonzero values of U/d indicate a system that is open to horizontal advection (parameterizes a two-dimensional system), with larger values of U/d increasing the influence of advection. Although this is a highly superficial treatment of horizontal advection, the parameterization captures some of the responses that are relevant to this study by representing advection from nearby sources with distinctly different temperature and humidity. Equations similar to (4)–(6) were also solved for the specific humidity. In this study, it was assumed that when advection was important, it was

Table 3. Values of Parameters and Boundary Conditions Used in the Surface Exchange and Mixed Layer Model Sensitivity Tests

Parameter	Value
Initial surface potential temperature	17°C
Initial surface specific humidity	10.7 g kg^{-1}
Initial mixed layer height (h)	50 m
Roughness length (z_o)	0.25 m
Surface resistance (R_c)	100 s m^{-1}
Distance to maritime source (d)	∞
Potential temperature lapse rate (γ_Θ)	6.4×10^{-3} K m^{-1}
Specific humidity lapse rate (γ_q)	-2.4×10^{-6} m^{-1}
Mean wind speed above canopy (U)	2.5 m s^{-1}

from nearby maritime sources, with a constant potential temperature and humidity (in this case we assumed Θ_e = 20°C; relative humidity 80%). Advection from other source regions (irrigated field, urban sources, heterogeneous vegetation types) is also likely within FLUXNET, but was not considered in this study. The set of equations was closed using the entrainment coefficient (c = 0.3) [*Tennekes*, 1973; *Raupach*, 2000]:

$$H_i = cH. \tag{7}$$

[14] The parameterized bulk mixed layer model neglects some physical processes, such as subsidence, small scalar gradients above the surface layer [*Wyngaard*, 1987], and the effect of momentum fluxes on the mean wind speed. We also assume the virtual heat flux equals the sensible heat flux. In the surface exchange module, the soil heat flux and heat storage in the biomass were also ignored. These considerations are important in many particular applications, but were not expected to substantially alter the first-order effects of the coupled model system on the relative diurnal trends in H and LE.

[15] Unless specified otherwise, Table 3 shows the model parameters used in each of the model sensitivity tests. Solar radiation was assumed to be 70% of the maximum potential value above the atmosphere at 40°N in June. The roughness length was chosen so that the resulting estimates of R_a were similar to that observed at FLUXNET forest sites (Wilson et al., submitted manuscript, 2001). The value of R_c was between the mean for coniferous and deciduous forests. The lapse rates for potential temperature and specific humidity in the free atmosphere (Γ_Θ and Γ_q) were set to mean values at a midlatitude site in MD, USA [*Wilson et al.*, 1999]. Model tests will show the sensitivity of diurnal trends to Γ_Θ and Γ_q.

3. Results

3.1. Diurnal Centroids of Surface Fluxes

[16] Figure 1 illustrates the conceptual meaning of the index $C_{gH} - C_{g\mathrm{LE}}$ for two forest sites, a primarily oak-maple deciduous forest in the southeastern United States (Walker Branch, TN, USA, site B, Figure 1a) and a Mediterranean oak forest near the Italian coast (Castelporziano, site S, Figure 1b). To emphasize relative diurnal trends and not absolute magnitudes, LE and H in Figure 1 were normalized by the maximum mean half-hourly value for the respective site-years. At Walker Branch, H was centered toward the morning and C_{gH} was 11.1 hours. LE was centered more toward the afternoon with a $C_{g\mathrm{LE}}$ value of 12.7 hours. The difference ($C_{gH} - C_{g\mathrm{LE}}$) was −1.6 hours, a value that was significantly greater than zero on the 5% probability level. At the Italian site, the timing of peak LE was weighted toward morning and H was weighted toward the afternoon ($C_{gH} - C_{g\mathrm{LE}}$ = +0.6 hours); this difference was not significantly different from zero on the 5% probability level. The meaning of the index $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ is similar, but depends on the relative timing between CO_2 uptake and PAR.

[17] The values of the indices $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ (horizontal axis) and $C_{gH} - C_{g\mathrm{LE}}$ (vertical axis) are shown for all site-years in Figure 2. On the left side of Figure 2 are site-years that have CO_2 uptake centered more toward the morning relative to PAR, and the site-years on the right side of Figure 2 generally had $F_{\mathrm{CO_2}}$ centered later in the day, after the centroid for PAR. Site-years on the lower portion of Figure 2 generally have LE centered more toward the afternoon relative to H. Alternatively, the upper portion of Figure 2 shows site-years where the diurnal centroid of H was just before and even after that of LE.

[18] At deciduous forests sites (sites A–E), the timing of peak LE was up to 2 hours after H, and the value of $C_{gH} - C_{g\mathrm{LE}}$ was generally more negative than in other ecosystems; this observation is especially true at the southernmost forest (Walker Branch, site B) (Figure 2). The CO_2 flux density was centered before to just slightly after the diurnal centroid for PAR (Figure 2). Among the different deciduous forests, the relative peak in $F_{\mathrm{CO_2}}$ occurred earliest at the southern forest (about 0.3–0.8 hours before PAR at Walker Branch), and generally occurred later in the more northern forests in North America (Harvard Forest, site A) and in the beech forests of Europe (sites C and D). There was also interannual variation within the individual sites. During years with adequate soil water content at Walker Branch [years B2, B3, B5, and B6; *Baldocchi*, 1997; *Wilson et al.*, 2000], $C_{gH} - C_{g\mathrm{LE}}$ was more negative (LE peaks relatively later) and $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ was more positive ($F_{\mathrm{CO_2}}$ peaks later) than in years with lower soil water content (years B1, B4). The two years of data at the European beech forest (Hesse, France, site C) suggests considerable interannual variation in the value $C_{gH} - C_{g\mathrm{LE}}$, but less variation in $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$.

[19] The shift toward a later peak in CO_2 fluxes as one moves north can be explained by the response of photosynthesis and respiration to temperature. At the southernmost site (Walker Branch), photosynthesis is weakly dependent on temperature except at more extreme values [*Baldocchi*, 1997; *Wilson et al.*, 2000]. Ecosystem respiration, on the other hand, is sensitive to temperature [*Hanson et al.*, 1993; *Edwards and Hanson*, 1996]. Consequently, this combination of responses to temperature causes $F_{\mathrm{CO_2}}$ to decrease during the afternoon. In contrast, the slightly later diurnal peak in $F_{\mathrm{CO_2}}$ at some northern deciduous forest site-years (Sites A, C, D) occurs because there is a smaller forcing of ecosystem respiration relative to photosynthesis in these cooler climates.

[20] A group of coniferous forests (most site-years between F and K) experienced peak rates of CO_2 exchange that was centered slightly after PAR (usually within an hour), and the lag between the diurnal centroid for LE and H was noticeably less (less negative $C_{gH} - C_{g\mathrm{LE}}$) than in deciduous forests (Figure 2). There were a number of

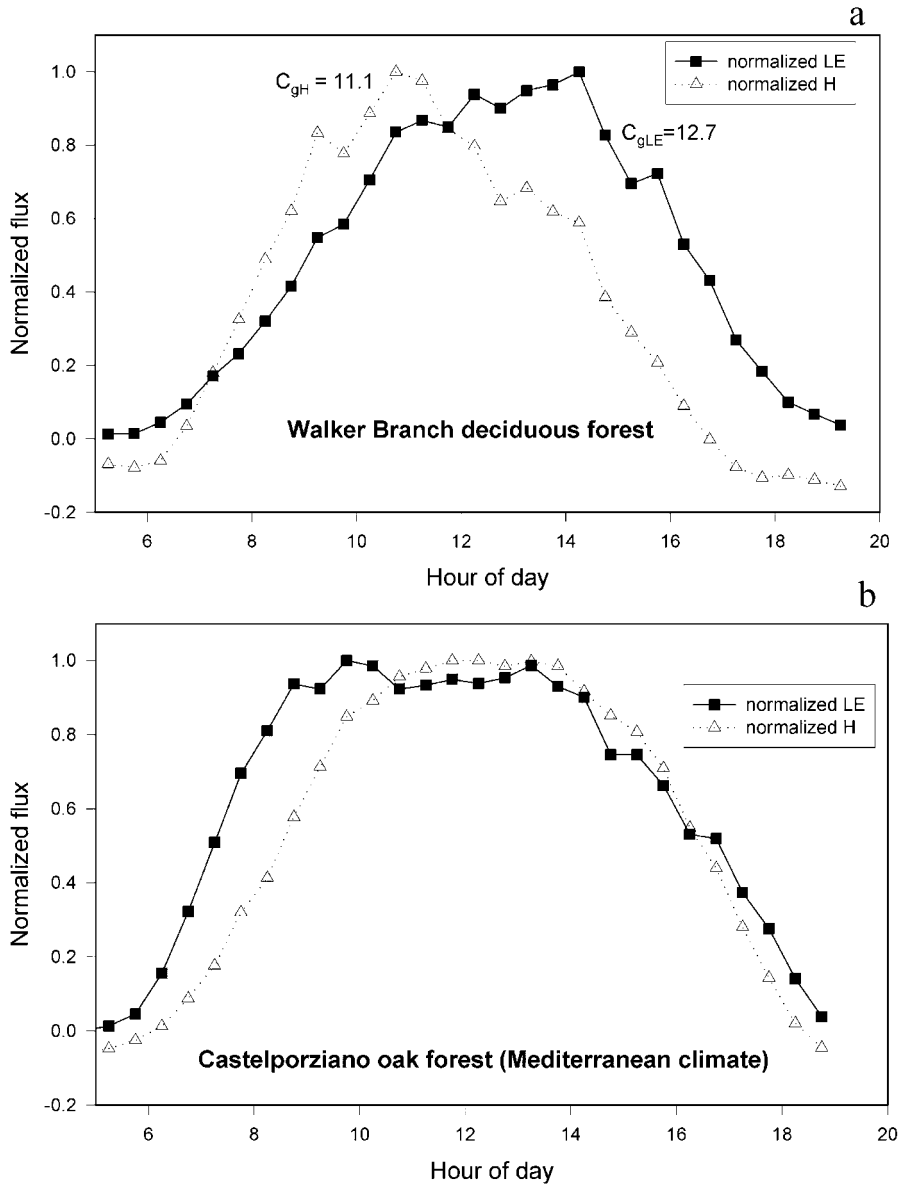


Figure 1. Normalized diurnal trend in latent (LE) and sensible (H) heat fluxes for (a) the deciduous forest at Walker Branch, TN, USA and (b) the oak forest at Castelporziano, Italy. Both LE and H are normalized to the maximum mean half-hourly value for the two respective site-years.

notable exceptions to this general grouping of conifers. The Sitka spruce site in Scotland (site O) had less negative values of $C_{gH} - C_{g\mathrm{LE}}$ and lower values of $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$. The boreal spruce forest in Canada (site N) had the largest positive values of $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ in the study. These results indicate that this forest tends to experience peak CO_2 uptake skewed more toward the afternoon relative to PAR. The montane forest in Colorado (Niwot Ridge, elevation 3050 m, site P) had an especially early relative peak in both $F_{\mathrm{CO_2}}$ and LE. The southernmost coniferous forest in this study (Duke Forest, NC, USA, site L) also experienced an early peak in relative $F_{\mathrm{CO_2}}$ ($C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ most negative), compared to most other sites. Two other coniferous sites that fall outside the conifer group labeled in Figure 2 are included in the group of sites with Mediterranean climates.

[21] The diurnal lag between LE and H was smaller at the coniferous forest sites as compared to deciduous forest sites. However, the reason for this observation cannot be explained by afternoon stomatal closure or development of the planetary boundary layer.

[22] Similar to deciduous forests, the cooler, more northern coniferous forests benefit more from increasing afternoon temperature by increasing photosynthesis more than respiration. This hypothesis is consistent with the conclusions by *Goulden et al.* [1997] for the site with the latest relative center in $F_{\mathrm{CO_2}}$ in this study (the black spruce boreal forest in Canada, site N). *Goulden et al.* [1997] suggested that drought stress was not present, and that assimilation was not limited by δ_e. Because canopy carbon uptake was limited when the temperature was below 14°C, cool mornings may have limited carbon assimilation.

[23] Advection and physiological response resulted in some variability in the indices for conifers. The lack of a late afternoon peak in δ_e at the Scottish Sitka spruce site

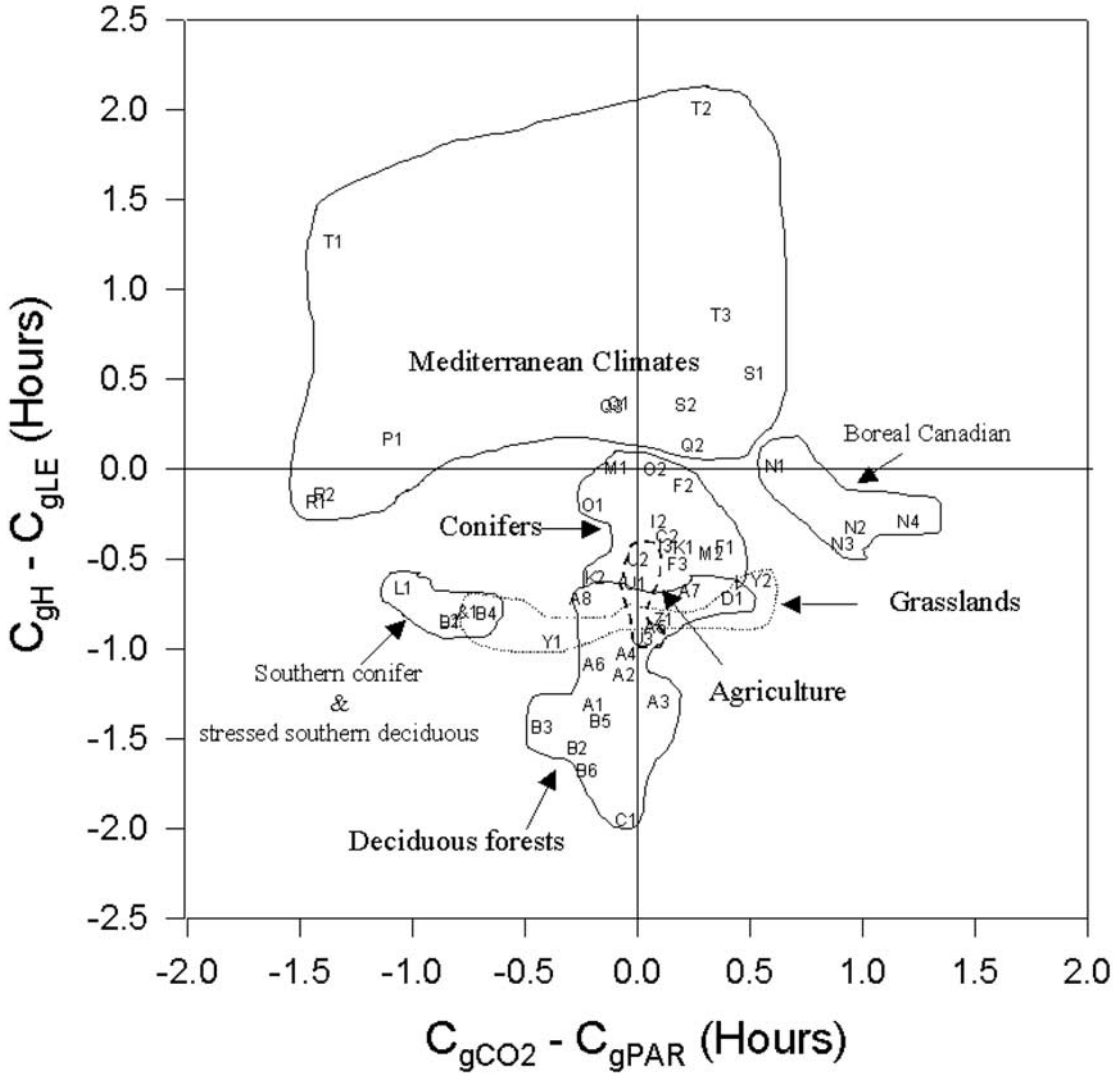


Figure 2. The value of the index $C_{gH} - C_{g\mathrm{LE}}$ versus the index $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ between days 165 and 235 for the FLUXNET sites shown in Table 1. Enclosed circles denote conceptual delineations between different vegetation types and climates, although there are exceptions.

(site O), and near-zero values of $C_{gH} - C_{g\mathrm{LE}}$, is consistent with advection (see section 4). Furthermore, Sikta spruce is sensitive to small changes in δ_e and leaf water potential [*Neilson and Jarvis*, 1975; *Jarvis*, 1976; *Lindroth*, 1985], which also can reduce the lag between LE and *H*. In a different maritime pine site, *Berbigier et al.* [1996] observed that LE peaked well before the peak in δ_e, and stomatal closure was suspected.

[24] Independent of vegetation type and dissimilar to most other sites, $C_{g\mathrm{LE}}$ was often centered only shortly after, or sometimes before, the diurnal centroid in *H* ($C_{gH} - C_{g\mathrm{LE}}$ positive) at sites growing in Mediterranean climates (sites Q–T). There was a wide variation in the relative diurnal timing of $F_{\mathrm{CO_2}}$ at sites in Mediterranean climates. The mature ponderosa pine site at Metolius, OR, USA (Site R) peaks early with regard to CO_2, but the diurnal centroid for $F_{\mathrm{CO_2}}$ is just before and even after that for PAR at the young ponderosa pine plantation at Blodgett, CA, USA (Site Q), which is a wetter site and at the Italian oak forest (Site S). The chaparral site in California (site T) had the most positive value of $C_{gH} - C_{g\mathrm{LE}}$, indicating that LE peaks especially early in the day, but there was considerable interannual variability in $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$.

[25] Diurnal flux patterns at sites in Mediterranean climates, which were exposed to a prolonged dry summer period, were modified by afternoon stomatal closure. Mediterranean species exhibit a range of adaptive physiological responses to drought, and midday stomatal closure has been detected in the chaparral species (*Ceanothus*) present at Sky Oaks [*Tenhunen et al.*, 1994]. Early daytime peaks in $F_{\mathrm{CO_2}}$ and decreasing LE in the afternoon, even when δ_e was increasing, have been observed previously at Metolius [*Anthoni et al.*, 1999] and in a California grassland [*Valentini et al.*, 1995]. *Anthoni et al.* [1999] deduced that stomatal closure was probably a more important response to drought than high temperature. An evaluation of fine-scale process models in these water-limited ecosystems has stressed that accurate stomatal sensitivity is required to capture diurnal

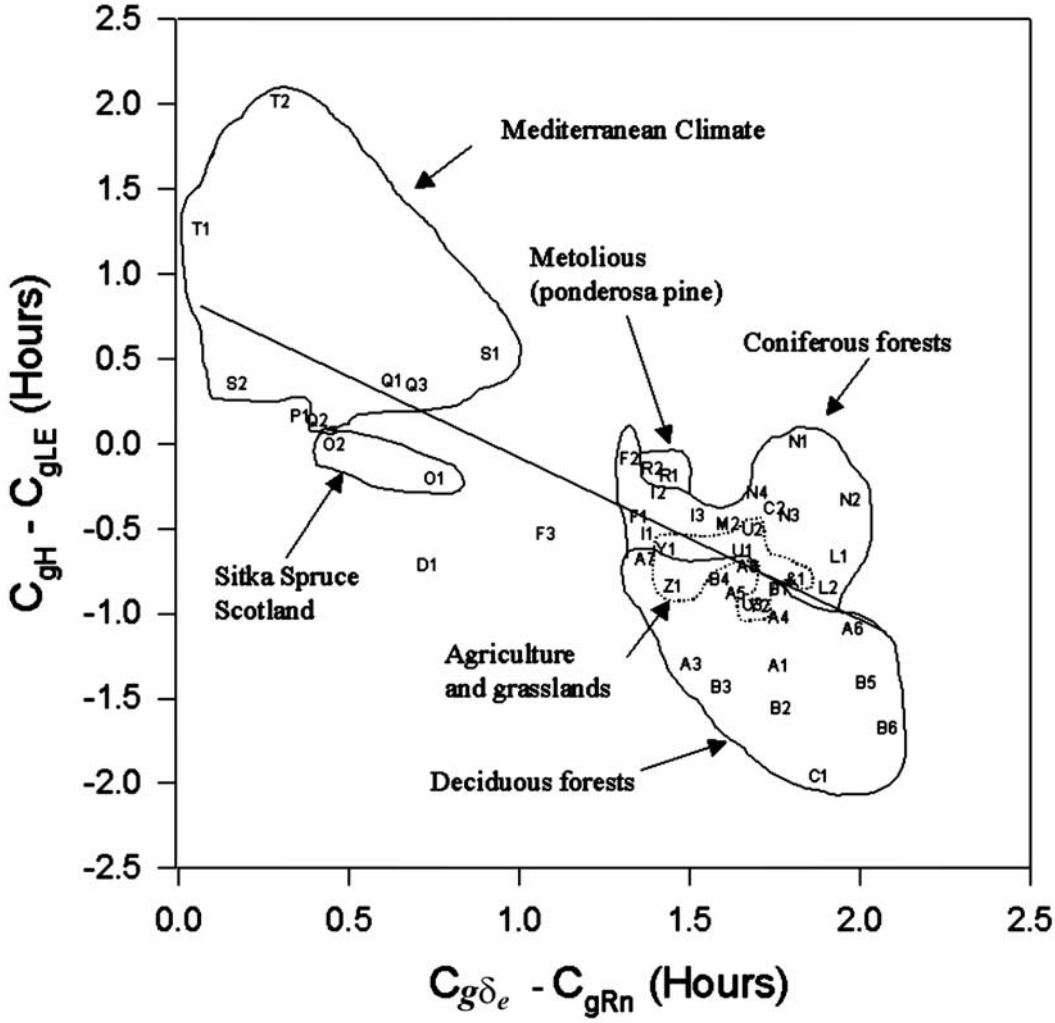


Figure 3. The relationship between $C_{g\delta_e} - C_{gR_n}$ and $C_{gH} - C_{gLE}$ for the FLUXNET sites shown in Table 1. Intercept is 0.87; slope is −0.96 and r^2 is 0.54. Enclosed circles denote conceptual delineations between different vegetation types and climates, although there were exceptions.

variations in photosynthesis and net carbon uptake [*Law et al.*, 2001].

[26] At the agricultural site (site U), the value of $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ was nearly zero. LE was centered later than H, but $C_{gH} - C_{g\mathrm{LE}}$ was generally less negative than in deciduous forests. The value of $C_{gH} - C_{g\mathrm{LE}}$ at grassland sites was similar to that at the agricultural site, but the value of $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ varied considerably (sites Y, Z and &, but not shown as a group in Figure 2). $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ was negative at the most northern grassland site (Fort Peck, MT, USA, site &) and 1 year at Little Washita, OK, USA [*Meyers*, 2001, site-year Y1].

[27] Theoretical studies [*Jarvis and McNaughton*, 1986] indicate that evaporation in short vegetation tends toward an equilibrium rate that is sensitive primarily to R_n and not to the diurnal variations in δ_e and R_c. The presence of a slight afternoon lag in LE relative H in agriculture and grasslands, both in the observed data and model sensitivity tests, is indicative of the fact that these shorter canopies were not totally decoupled from the atmosphere and that equilibrium evaporation increases slightly with temperature. It is uncertain why the lag between H and LE was at least as great in the shorter vegetation of agricultural fields and grasslands as in the taller coniferous forests, but it may indicate that the vegetation types respond differently to δ_e, or that there were important differences in advection or atmospheric stability.

[28] The nearly zero lag between PAR and CO_2 fluxes at the agricultural site is consistent with especially strong control of assimilation by PAR in agricultural species [*Larcher*, 1973] and with minimal afternoon stomatal closure and minimal temperature stress. The tendency for CO_2 uptake to peak early at the northern grassland site in Fort Peck, MT, USA (site &1), is likely the result of afternoon depressions in assimilation at temperatures above 25°C (T. Meyers, personal communication, 2002). The extreme difference in $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$between the 2 years at Little Washita, OK, USA (site Y) is likely the result of more severe water deficits in 1998 (Y1), and earlier daytime peaks in $F_{\mathrm{CO_2}}$ compared to 1999 (Y2) [*Meyers*, 2001].

[29] The effect of cloudiness was not considered explicitly in this study, only implicitly through the calculations of $C_{g\mathrm{PAR}}$. However, with some individual site-year exceptions, the general pattern in Figure 2 was similar when the difference between $C_{g\mathrm{CO_2}}$ and local noon was on the horizontal axis instead of $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ (not shown).

3.2. Diurnal Centroids of Atmospheric Demand

[30] Some of the variation in the diurnal patterns of energy partitioning between sites results from differences in how the atmospheric boundary layer and vapor pressure deficits develop. The horizontal (x) axis of Figure 3 illustrates site-year differences in the diurnal evolution of δ_e and net radiation (R_n) above the canopy, using the index $C_{g\delta_e} - C_{gR_n}$, or the difference in the diurnal centroid between δ_e and R_n. For site-years on the left side of Figure 3, there was not a significant diurnal lag between δ_e and R_n. Alternatively, for site-years on the right of Figure 3, δ_e was highest several hours after the peak in R_n. There was less of a tendency for δ_e to be highest in the late afternoon ($C_{g\delta_e} - C_{gR_n}$ tends toward zero) at sites in Mediterranean climates (sites Q–T), with the possible exception of the more inland Metolius, Oregon, USA site (site R), which is semiarid. Other sites without relatively late afternoon peaks in δ_e included the montane coniferous forest in Colorado (site P), and the Sitka Spruce site in Scotland (site O). Although all sites and vegetation types do not appear to respond identically, Figure 3 shows that a very general relationship exists between the climate index $C_{g\delta_e} - C_{gR_n}$ and the flux-partitioning index $C_{gH} - C_{g\mathrm{LE}}$ ($r^2 = 0.54$), which appears to indicate that the late peak in δ_e also results in a later peak in LE.

3.3. Effect of Soil Water Content on Diurnal Indices

[31] The effect of low soil water content on the indices $C_{gH} - C_{g\mathrm{LE}}$ and $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ was analyzed in more detail using data during published cases of developing drought in three different ecosystems at FLUXNET sites, a deciduous forest in the southeastern United States (Walker Branch, site B) [*Baldocchi*, 1997; *Wilson and Baldocchi*, 2000], a Ponderosa Pine plantation at Blodgett, CA (site Q) [*Goldstein et al.*, 2000], and a grassland in the southcentral United States (site Y, Little Washita, OK) [*Meyers*, 2001]. Values of $C_{gH} - C_{g\mathrm{LE}}$ and $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ were estimated during documented periods of both high and low soil water content for each site within a single season (Figure 4). Although the initial index values at high soil water content differ for each site, decreased soil water content consistently resulted in a more negative value of $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ and a more positive value of $C_{gH} - C_{g\mathrm{LE}}$, indicating that carbon and water fluxes occur earlier in the day during drought. During these particular cases, low soil water content had more effect on the timing of maximum carbon fluxes ($C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$) in the grassland and had more effect on the relative timing of energy fluxes ($C_{gH} - C_{g\mathrm{LE}}$) in the two forests.

4. Discussion

[32] To support our explanation of differences in the $C_{gH} - C_{g\mathrm{LE}}$ among ecosystems and climates, several sets of sensitivity tests were calculated using the coupled model. The variables chosen to analyze for sensitivity were R_a, Γ_Θ, Γ_q, U/d (by changing d), and R_c (δ_{es}) (the response of the surface resistance to "leaf" surface vapor pressure deficit). Each test consisted of 15 simulations, in which the variable of interest was altered incrementally.

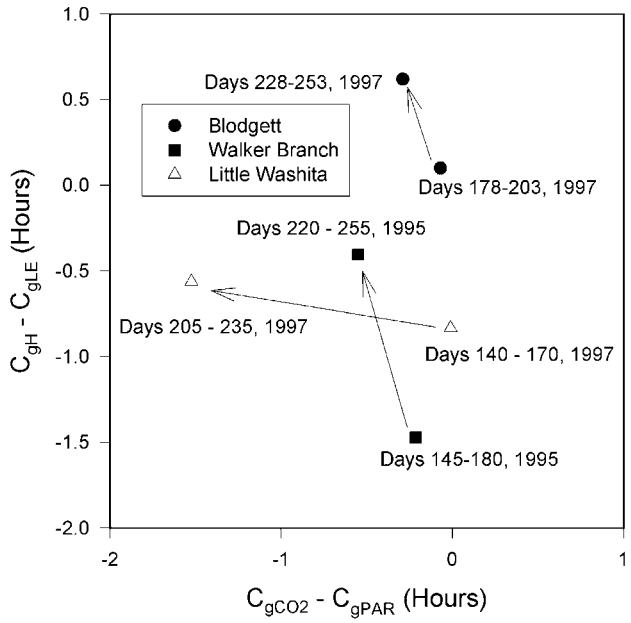
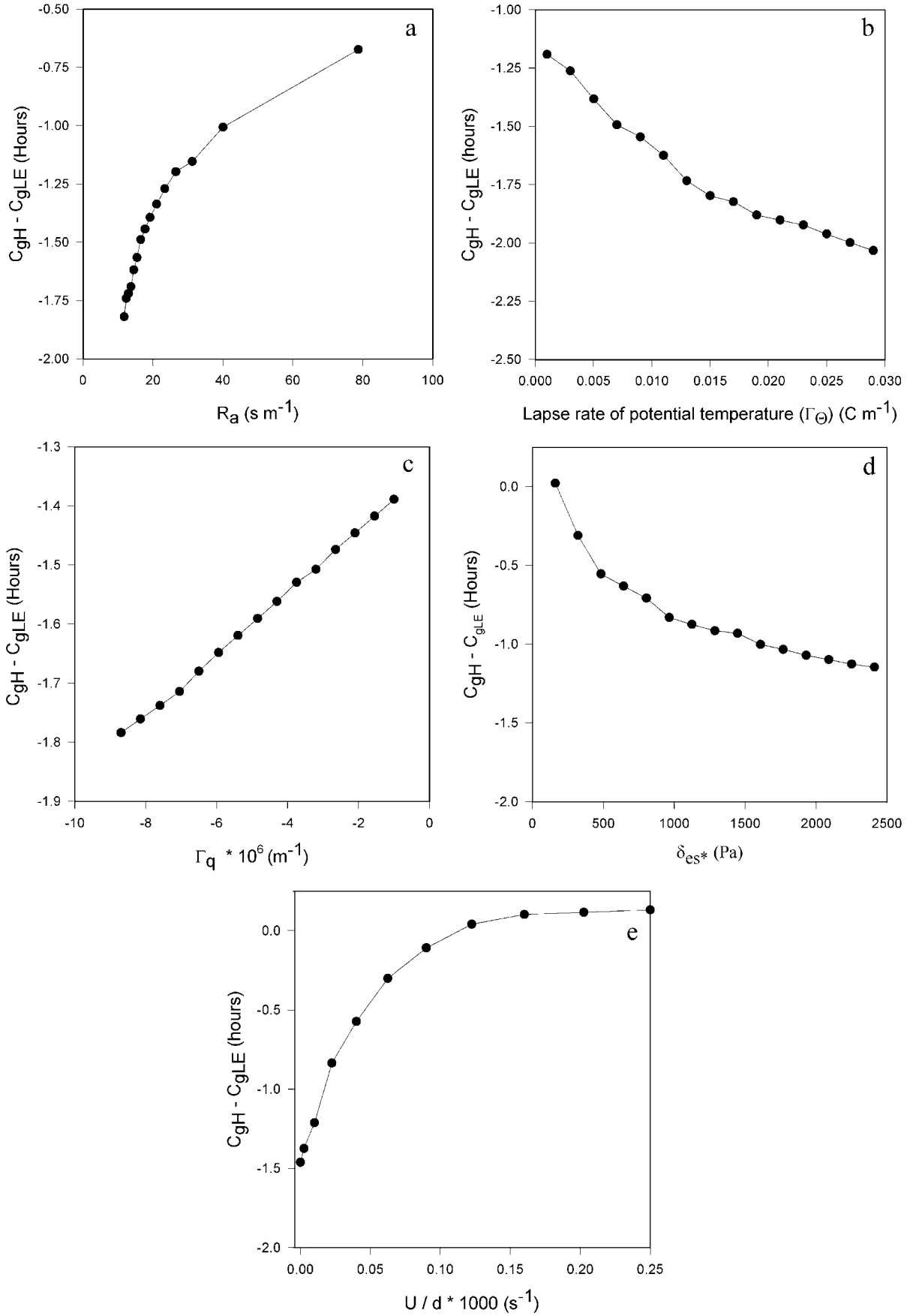


Figure 4. The seasonal evolution of $C_{gH} - C_{g\mathrm{LE}}$ and $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ at three different ecosystems during years with documented periods of decreasing soil water content. The sites are a deciduous forest at Walker Branch Watershed [*Baldocchi*, 1997], a ponderosa pine forest at Blodgett, CA, USA [*Goldstein et al.*, 2000] and a grassland at Little Washita, OK, USA [*Meyers*, 2001]. The arrows denote the change in the indices within single seasons from periods of adequate soil moisture to periods of low soil moisture.

[33] Aerodynamic resistance (R_a), as mediated by different levels of canopy roughness (e.g., forests versus crops), has a significant influence on $C_{gH} - C_{g\mathrm{LE}}$ (Figure 5a). The time difference between the time centroids for H and LE was more negative for tall rough vegetation (small R_a values) than for short, smooth vegetation (large R_a values). The lapse rate of potential temperature (Γ_Θ) influenced the height, temperature, and humidity of the mixed layer, and also the diurnal partitioning between H and LE. $C_{gH} - C_{g\mathrm{LE}}$ was more negative at large lapse rates (Γ_Θ) (high static stability) in the free atmosphere (Figure 5b). The maximum mixed layer height (h) decreased from nearly 5000 m at the weakest stability to only 700 m at the strongest stability (i.e., $C_{gH} - C_{g\mathrm{LE}}$ was more negative when the mixed layer height was lower). $C_{gH} - C_{g\mathrm{LE}}$ became more negative as the magnitude of the lapse rate of specific humidity increased (Figure 5c), likely because the steeper lapse rate resulted in more dry air entrainment from above the mixed layer, which progressively dried the mixed layer during the afternoon. The sensitivity to both Γ_Θ and Γ_q suggests that $C_{gH} - C_{g\mathrm{LE}}$ becomes more negative as the overlying atmosphere becomes more stable and dry.

[34] The potential effects of a stomatal response to δ_{es} on $C_{gH} - C_{g\mathrm{LE}}$ were explored by quantifying R_c as a function


a
$C_{gH} - C_{gLE}$ (Hours)
R_a (s m^{-1})
b
$C_{gH} - C_{gLE}$ (hours)
Lapse rate of potential temperature (Γ_Θ) (C m^{-1})
c
$C_{gH} - C_{gLE}$ (Hours)
Γ_q * 10^6 (m^{-1})
d
$C_{gH} - C_{gLE}$ (Hours)
δ_{es*} (Pa)
e
$C_{gH} - C_{gLE}$ (hours)
U / d * 1000 (s^{-1})

of δ_{es}, the vapor pressure deficit at the leaf surface, following *Lohammer et al.* [1980] and *Leuning* [1995]:

$$R_c = R_{\min}\left(1 + \frac{\delta_{es}}{\delta_{es*}}\right), \tag{8}$$

where $R_{\min}$ was the minimum surface resistance (100 s m^{-1}), and δ_{es*} determines the sensitivity of R_c to δ_{es}. Figure 5d shows the sensitivity of $C_{gH} - C_{g\mathrm{LE}}$ to specified values of stomatal sensitivity (δ_{es*}). Increasing the sensitivity to vapor pressure deficit (reducing δ_{es*}) enhanced afternoon stomatal closure, which resulted in values of $C_{gH} - C_{g\mathrm{LE}}$ that approached zero. Typical values of δ_{es*} for several species range from about 700 to 1500 Pa [*Leuning*, 1995]. The value of $C_{gH} - C_{g\mathrm{LE}}$ varies by one-half hour over this range of δ_{es*}.

[35] The effect of horizontal advection from an assumed maritime source on $C_{gH} - C_{g\mathrm{LE}}$ is shown by the response to U/d (Figure 5e). A horizontally closed system or site with limited advection ($U/d \rightarrow 0$), which might represent extensive vegetation at a continental site, gives the most negative value of $C_{gH} - C_{g\mathrm{LE}}$. As the maritime source is assumed to be closer (decreasing d) and/or mean wind speed is increased (increasing U), advection becomes more important. As a result, $C_{gH} - C_{g\mathrm{LE}}$ approaches zero. Also, shown in Figure 5e are rough estimates of U/d for two FLUXNET sites believed to be influenced by maritime sources. The estimates of U/d are based on mean wind speed at the sites and the approximate distance to large bodies of water. In both cases, the wind direction was often, but not always, from nearby a maritime source and the effects of advection would be less when it was not.

[36] Sensitivity studies suggested that advection probably influenced diurnal patterns at the Italian forest (Site S), which was within about 15 km of the Mediterranean Sea, by reducing the magnitudes of $C_{g\delta_e} - C_{gR_n}$ and $C_{gH} - C_{g\mathrm{LE}}$. The effect of advection at other sites with Mediterranean climates is less certain because they were more distant from large bodies of water (at least 75 km). The small values of $C_{g\delta_e} - C_{gR_n}$ and $C_{gH} - C_{g\mathrm{LE}}$ at the Blodgett, CA, USA, site would not appear to be the result of proximity to the coast (over 200 km), but possibly the result of high elevation (1315 m asl). This site and another high elevation site (Niwot Ridge, CO, USA, site P, 3050 m asl) both had reduced diurnal lags between δ_e and R_n and between H and LE, although the role of elevation on advection and the calculated indices in this study are not known.

[37] The simulated response of $C_{gH} - C_{g\mathrm{LE}}$ and $C_{g\delta_e} - C_{gR_n}$ to changes in R_a, Γ_Θ, Γ_q, U/d, and δ_{es*} are summarized in Table 4. Most negative values of $C_{gH} - C_{g\mathrm{LE}}$ will occur at sites with: (1) tall vegetation and/or high wind speeds (low R_a), (2) in continental regions with limited affects of advection (large d), (3) where static stability is large (large γ_Θ), and (4) where the stomatal response to vapor pressure deficit (large δ_{es*}) is weak. Alternatively, sites with shorter vegetation (high R_a), prominent advection from maritime sources (large U/d and $\Theta_e - \Theta_m$), weak static stability (small Γ_Θ), and strong stomatal response to vapor pressure deficit (small δ_{es*}) will likely have a $C_{gH} - C_{g\mathrm{LE}}$ much closer to zero. Although any of these factors may result in a value of $C_{gH} - C_{g\mathrm{LE}}$ near zero, U/d (advection) is the factor that had the most substantial impacts on $C_{g\delta_e} - C_{gR_n}$ including simulated $C_{g\delta_e} - C_{gR_n}$values less than 1 hour (not shown).

Table 4. The Response of $C_{gH} - C_{g\mathrm{LE}}$ and $C_{g\delta_e} - C_{gR_n}$ (Two Columns on Right Side) to Changes in Six Model Parameters (Left Column)

Change in Model Parameter	$C_{gH} - C_{g\mathrm{LE}}$	$C_{g\delta_e} - C_{gR_n}$
Increase R_a	less negative	small decrease
Increase stability (γ_Θ)	more negative	increase
Decrease humidity of atmosphere (γ_q)	more negative	increase
Increase advection (U/d)	less negative	decrease
Increase sensitivity to δ_e (decrease δ_{es*})	less negative	small increase

5. Summary

[38] Figure 6, which shows the conceptual phase space of Figure 2, illustrates the physiological and meteorological conditions that affect the relative diurnal timing of carbon dioxide and turbulent energy fluxes. Site-years on the relative upper portions of Figures 2 and 6 ($C_{gH} - C_{g\mathrm{LE}}$ slightly negative to positive) are most likely either (1) at locations where the diurnal variability of temperature and humidity in the atmospheric boundary layer are dampened by advection from large bodies of water or (2) are subject to afternoon stomatal closure, or both. The relative effects of these two processes can be evaluated to some extent by evaluating $C_{g\delta_e} - C_{gR_n}$. The sites in upper "left" portion of Figures 2 and 6 ($C_{gH} - C_{g\mathrm{LE}}$ slightly negative to positive and negative $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$) have early CO_2 and water fluxes and are most likely subject to afternoon stomatal closure, responses consistent with optimal gas exchange strategy during soil drying [*Makela et al.*, 1996]. Sites on the upper "middle and right" side of Figures 2 and 6 typically have relatively small diurnal variability in δ_e (small $C_{g\delta_e} - C_{gR_n}$) and afternoon stomatal closure may be less important than the influence of advection, most likely from large bodies of water.

[39] The bottom portions of Figures 2 and 6 generally contain sites with more continental atmospheric boundary layers and late afternoon peaks in δ_e (larger $C_{g\delta_e} - C_{gR_n}$). The bottom "left" portion of Figures 2 and 6 ($C_{gH} - C_{g\mathrm{LE}}$ fairly large negative and $C_{g\mathrm{CO_2}} - C_{g\mathrm{PAR}}$ negative) contains sites that probably experience some level of afternoon stress, either because of limited stomatal closure or increases in afternoon respiration because of temperature. Site-years in this quadrant are primarily from southern forests. Although there are no site-years on the extreme

Figure 5. (opposite) Response of the simulated value of $C_{gH} - C_{g\mathrm{LE}}$ to changes in model parameters; (a) atmospheric resistance (R_a), (b) atmospheric stability (γ_Θ), (c) lapse rate of specific humidity; (d) sensitivity of surface resistance to surface vapor pressure deficit (δ_{es*}); and (e) sensitivity to advection (U/d). The value of R_a in Table 5a was a midday mean value and incremented R_a values were obtained by altering z_o. Values of model parameters, if not changed to evaluate sensitivity, are shown in Table 3.

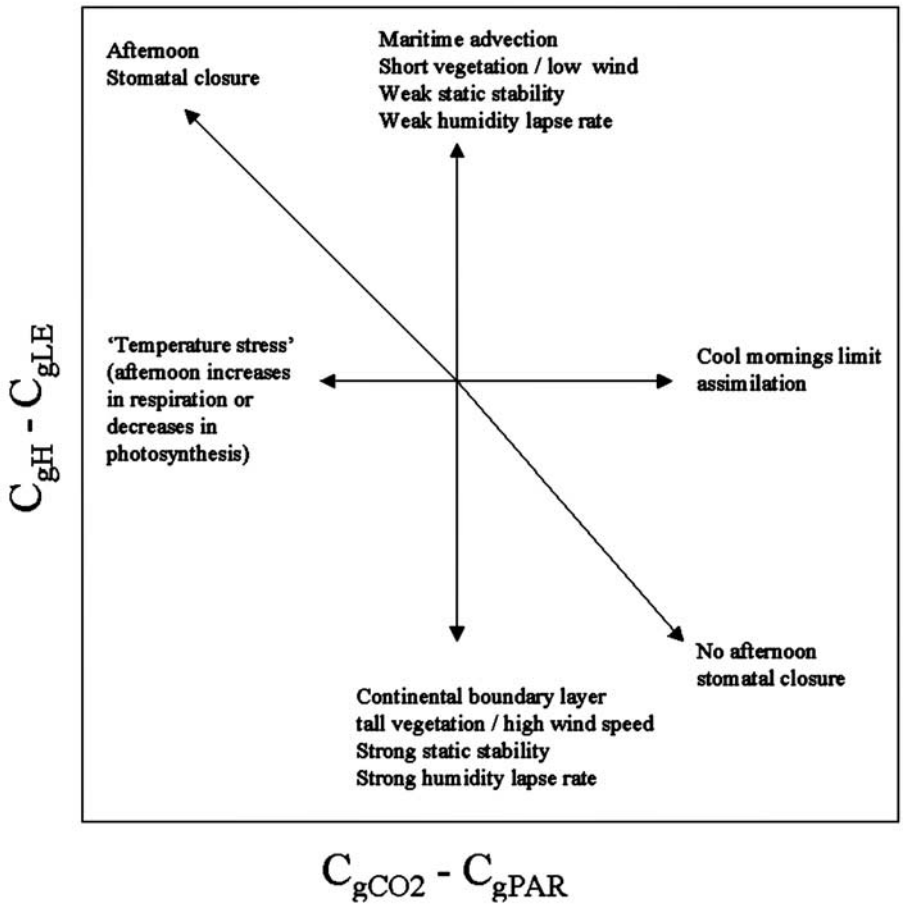


Figure 6. The conceptual phase space of Figure 2. The arrows indicate how particular processes tend to characterize sites within the phase space.

bottom right portion of Figure 2, movement toward this portion of the graph probably indicates almost no afternoon stomatal closure nor afternoon temperature stress and CO_2 assimilation rates that probably increase with afternoon temperature. Middle and high-latitude coniferous forests are generally closest to this region, but $C_{gH} - C_{g\mathrm{LE}}$ was only moderately or slightly negative in those forests, indicating that there may have been moderate afternoon stomatal closure.

[40] **Acknowledgments.** This work is result of the FLUXNET (NASA) Workshop in Marconi, CA, USA (June 2000). It is also a contribution to the Ameriflux program (U.S. Department of Energy's Terrestrial Carbon Program, TCP). DOE (TCP and NIGEC) and NOAA (GAPP) funded several of the sites in this study.

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