Partitioning evapotranspiration across gradients of woody plant cover: Assessment of a stable isotope technique

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[1] In water-limited ecosystems, partitioning ecosystem-scale evapotranspiration fluxes between plant transpiration and soil/canopy evaporation remains a theoretical and technical challenge. We used the Biosphere 2 glasshouse to assess partitioning of evapotranspiration across an experimentally manipulated gradient of woody plant cover using continuous measurements of near-surface variations in the stable isotopic composition of water vapor (δ^2H). Our technique employs a newly-developed laser-based isotope analyzer and the Keeling plot approach for surface flux partitioning. The applicability of the technique was verified by comparison to separate, simultaneous lysimeter and sap flow estimates of *ET* partitioning. The results showed an expected increase in fractional contribution of transpiration to evapotranspiration as woody cover increased—from *T/ET* = 0.61 at 25% woody cover to *T/ET* = 0.83 at 100% cover. Further development of this technique may enable field characterization of evapotranspiration partitioning across diverse woody cover gradients, a central issue in addressing dryland ecohydrological responses to land use and climate change. **Citation:** Wang, L., K. K. Caylor, J. C. Villegas, G. A. Barron-Gafford, D. D. Breshears, and T. E. Huxman (2010), Partitioning evapotranspiration across gradients of woody plant cover: Assessment of a stable isotope technique, *Geophys. Res. Lett.*, *37*, L09401, doi:10.1029/2010GL043228.

1. Introduction

[2] In water-limited ecosystems, evapotranspiration (*ET*) losses can account for more than 95% of all water inputs [*Wilcox and Thurow*, 2006]. It is essential to partition *ET* between transpiration and evaporation in drylands for at least three reasons: 1) Dryland ecosystem dynamics depend on plant water use and plant water use efficiency, which can only be measured at landscape scales by separating transpiration fluxes from soil/canopy evaporation; 2) Dryland regional water scarcity and demographic pressures necessitate quantifying processes that control the relative magnitude of unproductive (e.g., bare ground evaporation) vs. productive water losses (e.g., transpiration) [*Rockstrom et al.*, 2009] in both managed and natural ecosystems; and 3) Determining relative amounts of evaporation and transpiration is necessary to resolve critical uncertainties regarding the coupling of water and biogeochemical cycles in drylands [*Austin et al.*, 2004; *Breshears*, 2006]. However, partitioning of *ET* at landscape scales across different amounts of woody plant cover remains both an observational and theoretical challenge [*Huxman et al.*, 2005; *Caylor et al.*, 2006; *Moran et al.*, 2009], mainly due to the lack of methodologies available to quantify large-scale evaporation or transpiration in an easy and reliable way. For example, estimates of the percentage of annual evapotranspiration attributable to transpiration at similar sites in the Sonoran desert range from 7% [*Sammis and Gay*, 1979] to 80% [*Liu et al.*, 1995].

[3] Common methodologies for the estimation of field-scale transpiration rates include use of individual-tree sap flux [*Jackson et al.*, 2000], whole tree chamber observations [*Wullschleger et al.*, 1998], and paired soil lysimeters [*Scanlon et al.*, 2005]; each of these methods suffer from poor spatial representation. More recently, researchers attempted to partition daily-scale evapotranspiration using time series of soil surface temperature [*Moran et al.*, 2009]. Although this method can be applied over large spatial scales, it depends on consistency in the relationship between soil moisture and transpiration over entire growing season.

[4] Stable isotopes of water vapor hold great potential for resolving transpiration and evaporation fluxes from patch (e.g., 1 m²) [*Newman et al.*, 2010] to landscape scales [*Walker and Brunel*, 1990]. The process of evaporation is accompanied by a high degree of isotopic fractionation that leads to evaporated water with an isotopic composition depleted in heavy isotopes [*Craig and Gordon*, 1965]. Isotopic composition is denoted using δ notation, where $\delta = (R/R_{vsmow} - 1) \times 1000$, where δ is measured water vapor isotope composition (δ^{18}O or δ^2H), R and R_{vsmow} are the heavy/light isotope ratios of samples and the international standard (*VSMOW*). At the same time, the rapid turnover of water in transpiring leaves means that the signature of transpiration is usually similar to the isotopic composition of plant source water, especially during midday [*Ehleringer and Dawson*, 1992]. While some isotopic enrichment can occur in the leaf due to the same kinetic and diffusive effects that lead to evaporative fractionation in soils [*Flanagan et al.*, 1991], these non-steady-state leaf-scale effects usually occur only during early morning hours [*Flanagan et al.*, 1991]. Therefore, the isotopic composition of transpiration (δ_T) is always much heavier than the isotopic composition of evaporation (δ_E) [e.g., *Craig and Gordon*, 1965] and the distinct isotopic signature of these two fluxes can be used

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to partition total *ET* into relative rates of evaporation and transpiration in landscapes.

[5] Traditionally, researchers use cold-trap methods for water vapor sample collection, which attempts to completely condense water vapor contained within an air sample for laboratory analysis. The difficulty regarding collection and analysis of water vapor samples using cold traps has limited most studies to either chamber scales [*Yepez et al.*, 2005], or to temporally coarse observations [*Williams et al.*, 2004]. Recently, laser-based isotope instruments began to make direct and continuous water vapor $\delta^{18}O$ and δ^2H measurements possible, with precision similar to traditional cryogenic based isotope methods [*Lee et al.*, 2007; *Wang et al.*, 2009].

[6] In this study, we develop and evaluate a new technique for evapotranspiration partitioning that is targeted towards field-scale application. The method uses a recently developed laser-based isotope analyzer and a Keeling-plot approach to determine the partitioning of evapotranspiration across a gradient of fractional woody cover obtained through experimental manipulation in the Biosphere 2 facility. This new technique provides, for the first time, evapotranspiration partitioning for across a range of fractional woody cover, and provides important experimental data regarding the effect of woody cover on evapotranspiration partitioning in drylands. We verify the applicability of the technique using independent lysimeter and sap flow measurements (J. C. Villegas et al., manuscript in preparation, 2010).

2. Materials and Methods

2.1. Experimental Setup

[7] Our evapotranspiration experiments were performed within the Biosphere 2 glasshouse in Oracle, Arizona between September and October 2008. The advantage of Biosphere 2 is that the Biosphere 2 facility allows for more control of environmental variables such as temperature, relative humidity, and air circulation; details on the size, environmental control, and gas exchange of the biome can be found elsewhere [*Barron-Gafford et al.*, 2007]. In addition, the experimental framework of altering woody cover would be much more difficult to conduct in a non-greenhouse facility, where logistics of moving and weighing large potted tree containers would be very difficult. Most importantly, this facility ensures the source water for *E* and *T* are the same and that rainfall does not contribute to the water balance during the experiments. The measurements were taken over vegetation arrangements that were comprised of a 10×10 grid of containers (each 60×60 cm with depth of 80 cm). Each container was filled by either bare soil or planted with a 2 meters tall single mesquite tree (*Prosopis chilensis*) on the same soil. Soils were sandy loam texture and were taken from local Sonoran desert soils. We evaluated four arrangements that contained 25%, 50%, 75% and 100% woody plant canopy cover (the remaining canopy windows corresponded to bare soil containers). For each vegetation arrangement, all containers were saturated with tap water and allowed to drain for 16 hours to reach field capacity.

2.2. Continuous Measurements of δ_{ET}, δ_E, and δ_T

[8] We measured δ_{ET} (the δ^2H composition of the evapotranspiration flux) using the "Keeling plot" approach [e.g., *Keeling*, 1958; *Lee et al.*, 2007] applied to data from

the period during which water vapor concentration and δ^2H were most variable and corresponding to when plants were most active (10 am–7 pm). We sampled gas at heights of 0.5, 1.0 and 2.0 m at the center of the container arrangement into a ring-down cavity infrared spectrometer designed for water vapor isotope and water vapor concentration analysis (WVIA, Los Gatos Research Inc., CA), which was covered by a tarp to avoid direct solar radiation and provide temperature stability. The WVIA was calibrated before and after each measurement period using the procedure described by *Wang et al.* [2009]. The WVIA recorded δ^2H and water vapor concentration (ppm) measurements every 2 s during each 90 s measurement interval. Each measurement interval was buffered before and after the sample by a 30 s interval to avoid transient effects of switching among sampling locations, and measurements for each height were repeated every 15 min.

[9] An estimate of δ_E for soil evaporation was obtained using the Craig-Gordon model [*Craig and Gordon*, 1965]:

$$\delta_E = \frac{\alpha \delta_L - \delta_A h - \varepsilon_K - \varepsilon^*}{(1-h) + 10^{-3}\varepsilon_K}, \tag{1}$$

where δ_E is the isotopic composition of water evaporated from the soil; α is the temperature-dependent equilibrium fractionation factor ($\alpha < 1$ for liquid-vapor transformation), which can be calculated based on soil temperature [*Majoube*, 1971]; δ_L is the isotopic composition of liquid water at the evaporating front; δ_A is the isotopic composition of the background atmospheric water vapor; ε^* is calculated as $(1-\alpha) \times 1000$; ε_K is the kinetic fractionation factor for hydrogen (16.4‰ for non-turbulent conditions and 10.9‰ for turbulent transport [*Cappa et al.*, 2003]); and h is the relative humidity normalized to the soil temperature. The α value was 0.9393 based on Biosphere 2 temperature data (39°C) and following equilibrium equations of *Majoube* [1971]. We estimated δ_L by measuring the isotope composition of irrigated water using a Los Gatos Research liquid water analyzer at the University of Arizona. A value of 16.4‰ was used for ε_K, which corresponds to laminar conditions [*Cappa et al.*, 2003]. The δ_A was measured using WVIA. The h values (0.25–0.30) were obtained from the Biosphere 2 humidity monitoring data.

[10] To directly estimate δ_T for plant transpiration from water vapor, we used two direct approaches, which contrasts with previous approaches that indirectly estimate δ_T from measurements of extracted liquid leaf water or from leaf water enrichment calculations for non-steady state conditions [e.g., *Yepez et al.*, 2005]. Our first approach was to measure transpiration within a customized leaf chamber subjected to a 100% di-nitrogen atmosphere. Leaves used to determine the isotopic signature of transpiration were sealed inside the chamber, which had a small mixing fan, air temperature corresponding to that inside the glasshouse, and was flushed and purged with ultra-high purity nitrogen. Two sets of 20-min (at 0.5 Hz) measurements of δ^2H (1200 samples total) were obtained from each of two different branches; data were averaged by branch. Our second approach was to obtain measurements from branches within a LICOR-6400 standard leaf chamber (6400-02B) exposed to ambient air that had been passed through Drierite. We estimated δ_T of plant transpiration for averages of three different 5-min sampling periods (450 samples total). All

chamber measurements of the isotopic composition of transpiration were obtained under sunny conditions between 1 and 3 pm.

2.3. Evapotranspiration Partitioning Calculations

[11] Assuming a simple 2-source model of total evaporation, the fractional contribution of transpiration (F_T, [0-1]) to total evapotranspiration can be quantified as

$$F_T = \frac{\delta_{ET} - \delta_E}{\delta_T - \delta_E} = \frac{T}{ET}, \quad (2)$$

where δ_{ET}, δ_E and δ_T are the isotope signatures of evapotranspiration, evaporation and transpiration, respectively [*Williams et al.*, 2004]. Because our experimental system consists of soil-filled boxes either with or without single-stemmed mesquite trees, bare soil evaporation can be further partitioned into bare soil evaporation from under tree canopies (E_v), and bare soil evaporation from locations unoccupied by tree canopies (E_b). To determine the relative contribution of bare soil evaporation to total ET, we took advantage of the fact that when the experimental tree cover is 100% the only contributions to ET are T and E_v, so that

$$\frac{T_{100}}{E_{100}} = \frac{T}{E_v} = \eta_v, \quad (3)$$

where T_{100} and E_{100} refer to estimated transpiration and evaporation determined during the 100% tree cover treatment, and η_v is the ratio of transpiration to evaporation within boxes occupied by trees. Because in every treatment $ET = E_b + E_v + T$, we then combined equations (2) and (3) to define the ratio of bare soil evaporation in non-vegetated boxes to transpiration according to

$$\frac{E_b}{T} = \frac{1}{F_T} - \frac{1}{\eta_v} - 1, \quad (4)$$

[12] Finally, we defined the ratio of bare soil evaporation to transpiration on a per-unit area basis, η_s, which is given by

$$\eta_s = \frac{1-f}{f} \frac{T}{E_b}, \quad (5)$$

where f is the fraction of vegetation cover in each treatment. Equation (2) makes it clear that resolving the relative rates of evaporation and transpiration requires knowledge of the isotopic composition of both end members (δ_E and δ_T) as well as isotopic composition of the total flux itself (δ_{ET}). We determined δ_{ET} using the inverse gradient method (or Keeling plot approach), which has been implemented extensively in CO_2 flux applications, but was also recently used to calculate δ_{ET} at ecosystem level [e.g., *Lee et al.*, 2007]. The Keeling plot approach is based on the conservation of mass and can be expressed as

$$\delta^2 H_a = c_b\big(\delta^2 H_b - \delta^2 H_s\big)(1/c_a) + \delta^2 H_s, \quad (6)$$

where $\delta^2 H_a$, $\delta^2 H_b$ and $\delta^2 H_s$ are the isotope signatures of ambient (observed) water vapor, background water vapor and evapotranspiration respectively, c_a is the ambient water vapor concentration, and c_b is the background water vapor concentrations.

3. Results and Discussion

[13] Temporal dynamics of water vapor concentrations and isotopic composition both exhibited diurnal variation that corresponded to plant activity (Figure 1). Water vapor concentrations increased during the early morning and peaked in the early afternoon before gradually decreasing, regardless of sampling heights and vegetation cover (Figures 1a–1d). These results indicate that there are measurable diurnal changes in evapotranspiration in our study system and that such patterns are mainly driven by the cycles of solar radiation, as expected [*Villegas et al.*, 2010a]. Water vapor concentrations always returned to minimum values at night (Figures 1a–1d), indicating a complete exchange of air with the outside atmosphere over 8–10 hours. The water vapor $\delta^2 H$ values also showed clear diurnal patterns. Regardless of sampling heights and vegetation cover, the water vapor $\delta^2 H$ values increased from early morning and peaked around noon then gradually decreased (Figures 1e–1g). The diurnal changes in $\delta^2 H$ reflect the plant and soil contributions to near surface atmospheric isotopic signatures. Because the total ET was dominated by transpiration and transpiration flux has heavier signals compared with background vapor in this case, when plants start transpiring, the atmospheric isotopic signatures will become more enriched. There were often vertical isotope gradients, particularly during the daytime periods, when $\delta^2 H$ values were higher at 2 m height. Elevated $\delta^2 H$ signatures generally corresponded to increases of vegetation cover, indicating the increasing contributions of transpiration to total evapotranspiration.

[14] The two methods of characterizing plant transpiration $\delta^2 H$ signatures differ in their results. The customized chamber method produced a value of −62.1‰, while the LICOR leaf chamber method produced a value of −74.1‰ (Figures 2a and 2b). Because the irrigation water $\delta^2 H$ value was −63.3 ± 0.1‰, we only used the chamber method results within our evapotranspiration partition calculations, since this method is more consistent with the expectation that plant transpiration should not result in fractionation. The light LICOR result (~10‰) is most likely caused by contamination of a small amount of ambient water vapor, which had a $\delta^2 H$ value of around −110‰. There are very few direct measurements of plant transpiration isotopic composition in the literature [e.g., *Lai et al.*, 2005]. Given the paucity of direct measurements and the inconsistent results obtained from our different approaches, we expect that future refinement of methods capable of accurately measuring transpiration isotopic composition will have substantial contributions to existing theoretical model predications and explanations of leaf water enrichment during transpiration [*Flanagan et al.*, 1991].

[15] Our calculated evaporation isotopic signature ($\delta^2 H$) was −137‰, which is slightly lower than prior studies in arid environments (e.g., −131‰ [*Williams et al.*, 2004]). As described above, our treatment-level δ_{ET} values were calculated using the Keeling plot approach (Equation 6). Notably, our results support the expectation that as woody
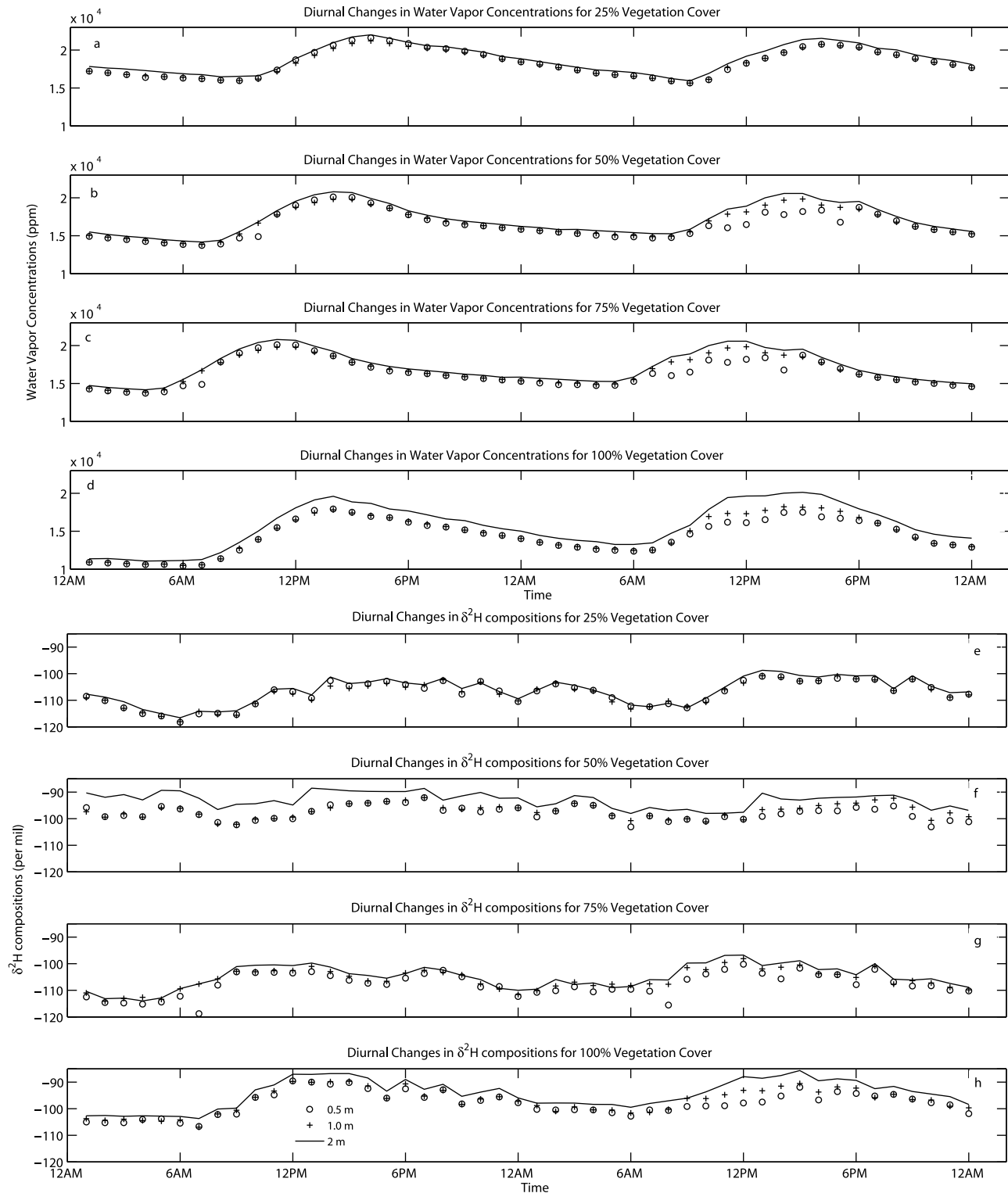
Figure 1. Diurnal variations of water vapor concentration (ppm) at different height (0.5 m, 1 m and 2 m) for vegetation cover (a) 25%, (b) 50%, (c) 75% and (d) 100%. Diurnal variations of $\delta^2 H$ values at different height for vegetation cover (e) 25%, (f) 50%, (g) 75% and (h) 100%.

cover increases, δ_{ET} signatures generally increase, due to expected greater contributions of transpiration relative to evaporation [*Breshears*, 2006]. Specifically, the average δ_{ET} signatures (10 am–7pm, cf. Methods) are −90.8‰ at 25% woody plant cover, −84.7‰ at 50% cover, −78.4‰ at 75% cover, and −74.7‰ at 100% cover (ANOVA, p < 0.001

(Figure 2c)). Because δ_{ET}, δ_E and δ_T were all measured/calculated independently, the contribution of transpiration to total evapotranspiration (F_T) for each level of woody plant cover can be determined (equation (2)). Our results showed the expected increase in *T/ET* as woody plant cover increases, and F_T rose from 61% to 83% as vegetation cover
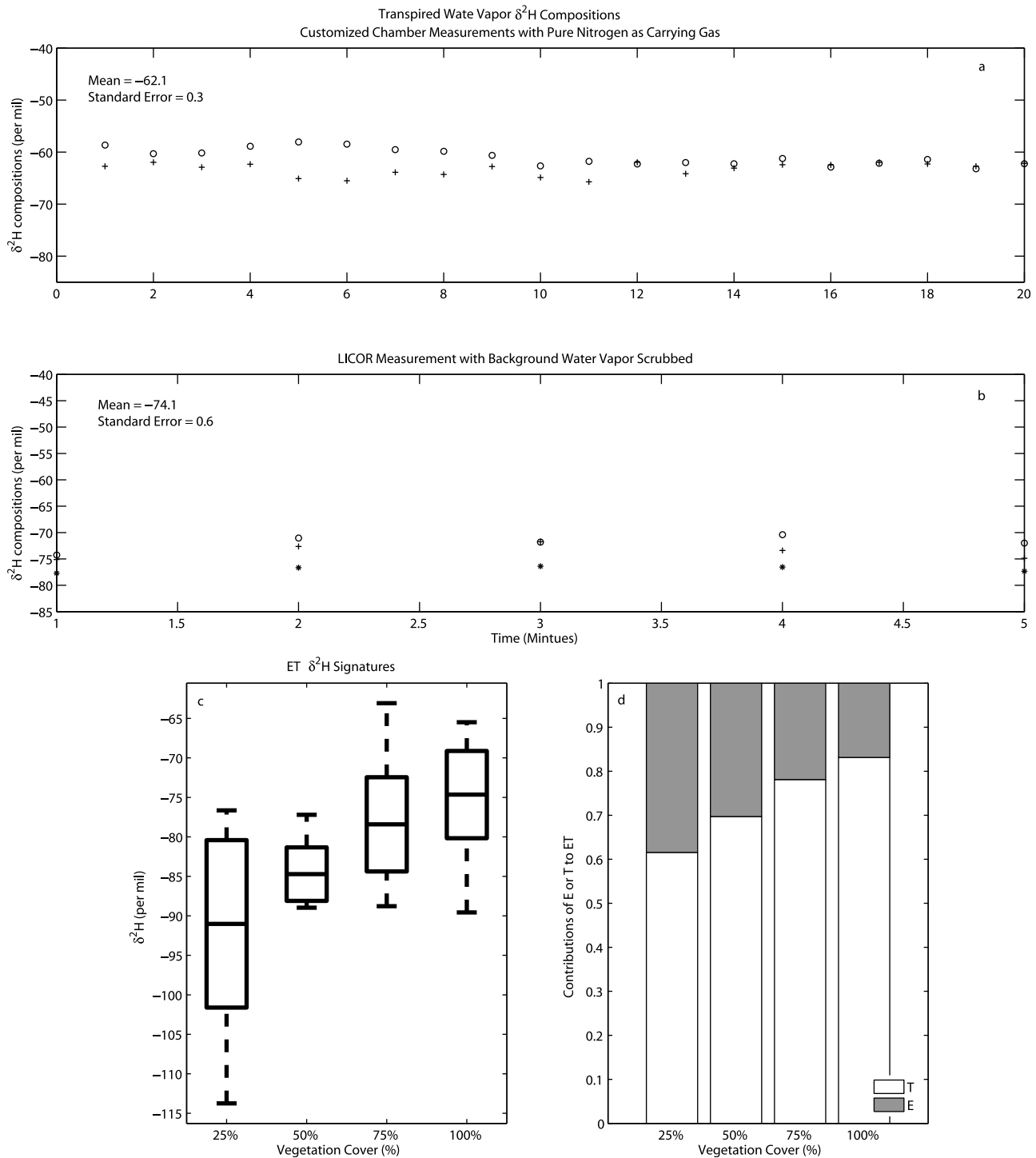
Figure 2. (a) Transpired δ^2H signatures measured using a customized chamber with ultra-high purity nitrogen gas as carrying gas and (b) LICOR 6400 leaf chamber with water vapor scrubbing. The circle, cross and asterisk symbols indicate different measurements. (c) Calculated evapotranspiration δ^2H signatures (box and whisker plot) for different vegetation covers and (d) the contributions of transpiration to evapotranspiration for different vegetation covers. In the box and whisker plot, the box lines represent means and standard deviations of the observations, and whiskers represent maximum and minimum values of the observations.

increased from 25% to 100% (Figure 2d). The partition values are similar to an independent, concurrent lysimeter and sap flow based measurement that reported F_T values of 0.36, 0.42, 0.70 and 0.79 for 25%, 50%, 75% and 100% vegetation cover after removing night evaporation (J. C. Villegas et al., manuscript in preparation, 2010). The differences between these two methods range from 4% to 26% with an average of 15.6%. Considering the uncertainties in both sap flow and isotope measurements, the reasonable agreement between these two methods demonstrate the

credibility of our new technique. The E_b/T ratios were 0.43, 0.22 and 0.08 for 25%, 50% and 75% cover, indicating that the relative contribution of bare soil evaporation compared to transpiration rapidly decreases as woody plant cover increases [*Villegas et al.*, 2010b]. However, the relative effectiveness of bare soil evaporation per unit area (η_s) varied only slightly as cover increased (0.15 at 25% cover, 0.22 at 50%, and 0.23 at 75%), which suggests that in our experiment, the occurrence of sparse and low-LAI canopies have a minimal shading effect on bare soil evaporation. This is consistent with field observations [*Villegas et al.*, 2010a].

[16] Our results yield initial insights into how *ET* partitioning can change with woody plant cover, although more general and diverse relationships of *ET* partitioning with woody plant cover remain uncertain and likely vary depending on climate, soils, leaf area and species, among other factors [*Huxman et al.*, 2005; *Breshears*, 2006]. Our experimental design isolated individual containers, precluding below-ground connectivity of patches associated with woody plant roots that extend into neighboring patches, as occurs in the field. Such connectivity affects *ET* partitioning at the patch scale between canopy patches of woody plants and the intercanopy patches that separate them [*Caylor et al.*, 2006; *Newman et al.*, 2010], so our reported *ET* partition values at different cover may not be exactly the same as in the field setting.

[17] Our experimental results illustrate the utility of a technique for continuous δ_{ET} measurements that enables *ET* partitioning in landscapes. In our experiment, evaporation fluxes only came from bare soil, whereas in natural environments rainfall interception by the vegetation canopy and subsequent evaporation may constitute a significant part of total evaporation. Because evaporation from soil and canopy surfaces are governed by the same principle and has similar signals, this new technique will be able to capture the partitioning of *ET* across many different ecosystems. In areas where soil water evolves much different isotopic composition than rainwater, it may be possible to even further partition evaporation fluxes between canopy and soil evaporation. Our study also includes development of an approach that directly measures plant transpiration signals, which in the past have been largely estimated by measuring plant source water and modeling water enrichment under non-steady state conditions. Although our technique provides new frameworks and represents important progress, we believe it will be necessary to directly couple high-frequency observations of isotope measurements to eddy covariance systems in order to eliminate the dependence of Keeling plot approaches (need strong near-surface gradients in water vapor isotopic composition).

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