

DRAFT

Background & Methodology for measuring CO₂ and H₂O tent fluxes

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Background Carbon flux

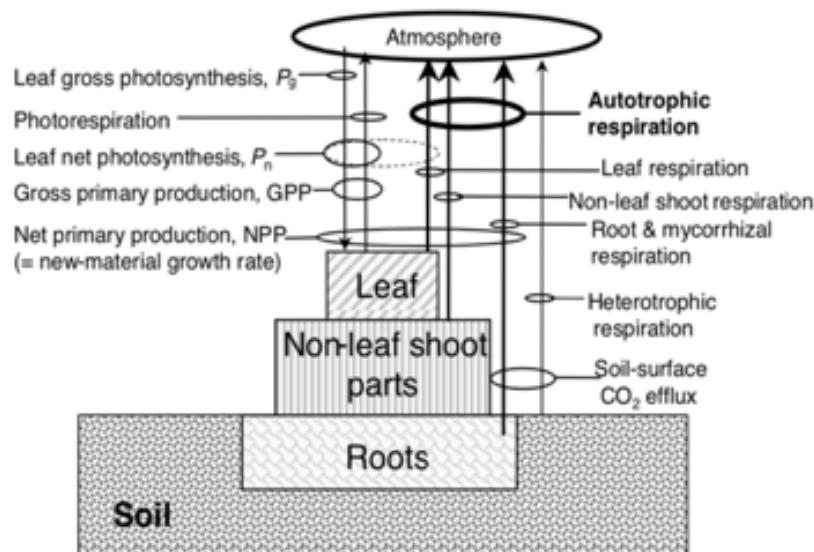


Fig. 1. Diagram of major ecosystem CO₂ fluxes illustrating the fluxes that are combined in defining the composite fluxes of gross primary production, net primary production and autotrophic respiration. Arrow boldness does not reflect the magnitude of the fluxes, rather their pertinence to autotrophic respiration.

An overview diagram of carbon fluxes between vegetation and soil from Gifford 2003. (Gifford, 2003)

The annual accumulation of organic matter per unit of land is a measure of net primary productivity, NPP, often expressed in units of $\text{g m}^{-2} \text{ yr}^{-1}$. GPP is gross primary productivity. NPP is the net amount of primary production after the costs of plant respiration, R , are included.

$$NPP = GPP - R$$

$$GPP = NPP + R$$

Net ecosystem production, NEP is the net amount of primary production after the costs of respiration by plants, heterotrophs and decomposers

From William H. Schlesinger, Emily S. Bernhardt, in [Biogeochemistry \(Third Edition\)](#), 2013

~~Net ecosystem production (NEP) as~~

$$NEP = NPP - (R_h + R_d)$$

~~Where then~~

$$NEP = GPP - (R_p + R_h + R_d)$$

where R_p , R_h , and R_d represent the respiration of plants, herbivores, and decomposers, respectively. Except in unusual circumstances, NEP will be a partial fraction of NPP. NEP is a useful measure to determine carbon balance of ecosystems.

Net ecosystem production is a useful term for a concept different from C accumulation – it is the amount of organic C fixed in an ecosystem that is not respired there and is therefore available for accumulation, export, or nonbiological oxidation (Lovett et al., 2006).

Remember we have NEE, $R_{\text{eco}}(\text{day})$, and $R_{\text{eco}}(\text{night})$. Also, sometimes we hear Net Ecosystem Production which I have seen as $NEE = -NEP$. You can use these nice papers from Luyssaert et al. for references on linking NEE, GPP, R_{eco} , NPP

NEE and NEP is the total flux which represents soil flux, plant flux (respiration and photosynthesis).
 $R_{\text{eco}}(\text{day}) = R_{\text{plant}} + R_{\text{soil}}$

Remember that R_{soil} is both microbial and plant root respiration. One cannot disentangle root respiration from microbial respiration without much more work.

Tent Flux Measures

In our tent, we measure fluxes during the day, at night, and in the dark. Fluxes during the dark are all respiration measures as we shut off photosynthesis and transpiration. These respiration tent fluxes are R_{eco} . Tent respiration measures are a combination of both plant and soil respiration. Remember, soil respiration also includes plant root respiration. Tent flux measurements during the day measure net ecosystem exchange or NEE measure.

$$NEE = GPP - R_{eco}$$

$$R_{eco} = R_{plant} + R_{soil}$$

$$NEE = GPP - (R_{plant} + R_{soil})$$

We do not measure above-ground plant respiration, R_{plant} so that we can approximate it as

$$R_{plant} = R_{eco} - R_{soil}$$

So R_{plant} for the aboveground plant respiration can be estimated by subtracting R_{soil} from $R_{eco}(\text{day})$. Remember, R_{plant} is above-ground plant respiration.

$$GPP = NEE + R_{eco} = NEE + (R_{plant} + R_{soil})$$

$NEP + R_{eco}$, where R_{eco} is again the total ecosystem respiration from the tent flux measure. Another way to state this is $-NEE + R_{eco} = GPP$

GPP is straightforward using our tent measures. To get GPP just either add or subtract the dark day tent flux (depending on if R_{eco} is recorded as a negative or positive value) from the NEE measure.

The trick to get NPP with the tent flux data is to subtract autotrophic respiration, but this is quite difficult as soil respiration contains both heterotrophic and autotrophic.

How to calculate NPP?

$$NPP = GPP - R_{plant}.$$

$$NPP = NEP + R_h$$

Where R_h is heterotrophic respiration or microbial respiration.

$$NPP = GPP - R_a$$

Where R_a is heterotrophic respiration

In Sloat et al. (2015), we do estimate NPP, but NPP is hard. Now, $NPP = GPP - R_a$, but again we can't parse soil respiration into R_a and R_h . We estimated aboveground NPP by subtracting the total soil respiration so that $NPP_{aboveground} \sim NEP - R_{soil}(total)$. To get $R_{soil}(total)$ there are two ways to estimate it. The first step is to calculate the average soil respiration from the soil collars in a plot and then multiply that value by the total area of the plot to obtain the total soil respiration. Make sure the area units are converted to the same.

The other approach is to use the percentage of bare soil area in each plot from the plot cover measures and then multiply the total plot area by the percentage of bare soil to determine the total area that is soil. Take that area and then multiply it by the average soil collar respiration. Again, make sure the area units match. The first method will likely lead to a lower NPP, as not all of the plot area is bare soil, as some plants take up physical space. The second method would likely result in a higher NPP value. The truth lies somewhere in between. Yet another way to get at NPP is to assume that plants' carbon use efficiency (or CUE) is about 40-50% ((DeLUCIA et al., 2007) [Ref](#)) The CUE is the ratio of NPP to GPP. However, CUE varies in time and space. Lastly, we can also calculate the rate of biomass gain in the plots. If you know the season start date, the total biomass at peak, and the change in time between growing season start and peak biomass, you can calculate above-ground NPP as the peak above-ground biomass / change in time.

Sometimes in the literature we also use the gross ecosystem productivity, GEP, is

$$GEP = NEP + R_p$$

From Lovett et al. 2006

“By measuring the vertical net flux of CO₂ above the ecosystem, eddy covariance measurements estimate the total CO₂ exchange, often called the “net ecosystem exchange” (NEE). Net ecosystem exchange is equal to the NEP plus sources and sinks for CO₂ that do not involve conversion to or from organic C (although the NEE is by convention opposite in sign, so that fluxes into the ecosystem are negative):

$$NEE = NEP + \text{inorganic sinks for CO}_2 \text{ inorganic sources of CO}_2.$$

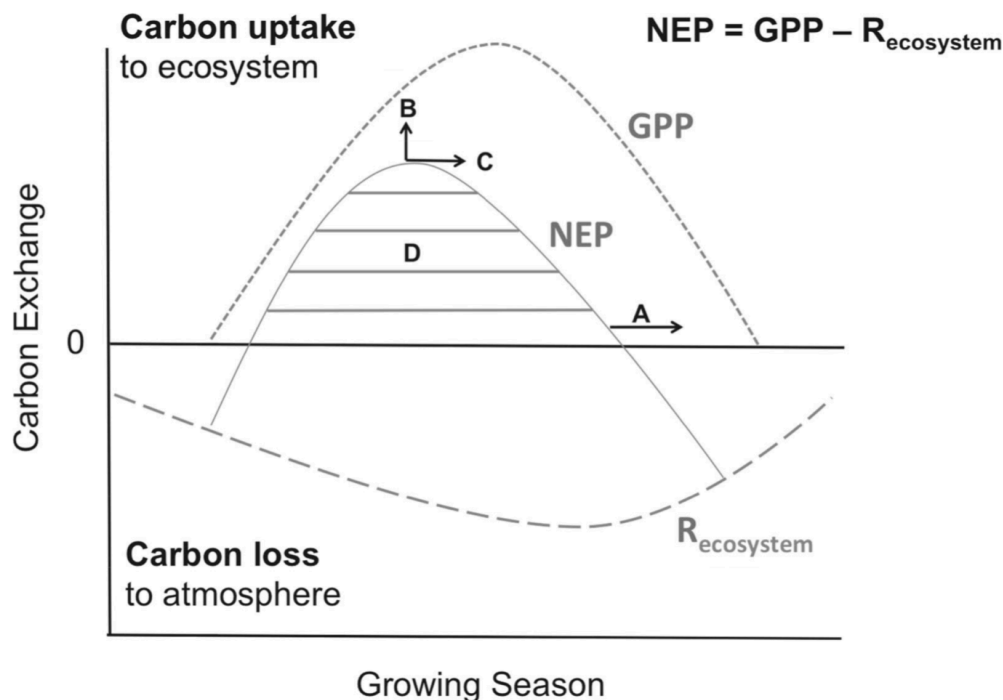
Examples of inorganic sources and sinks are weathering reactions, precipitation or dissolution of carbonates, and atmosphere–water equilibrations. Although they are likely to be minor terms in the CO₂ exchange of a forest (but see Raymond and Cole 2003), these inorganic sources and sinks can be very important in the ocean.”

From Sloat et al. 2015 (Sloat et al., 2015) we provide the following figure and diagram.

Box 1 Net ecosystem productivity (NEP) is metric which includes aboveground and belowground inputs to ecosystem carbon flux.

NEP is the gross primary productivity (GPP) of plants minus respiration by autotrophs and heterotrophs ($R_{\text{ecosystem}}$) (Begon and others 2009). NEP is often equated to NEE, or the inverse of NEE (Kirschbaum and others 2001). At night when plants are respiring NEP would be the same as $R_{\text{ecosystem}}$ (Reay and others 2007). NEP is measured as a rate—in this case it is measured in units of μ moles of CO_2 per m^2/s . This rate can be positive or negative, depending on whether or not the plants and soil together are taking in CO_2 or releasing it into the atmosphere.

The NEP curve over the course of the growing season can be predicted to change in a number of ways in response to experimental or natural variation in available moisture. Plants could remain active later in the season (A). The rate of peak uptake could increase (B) or shift in time (C). Any of these changes have the potential to influence cumulative NEP (D).

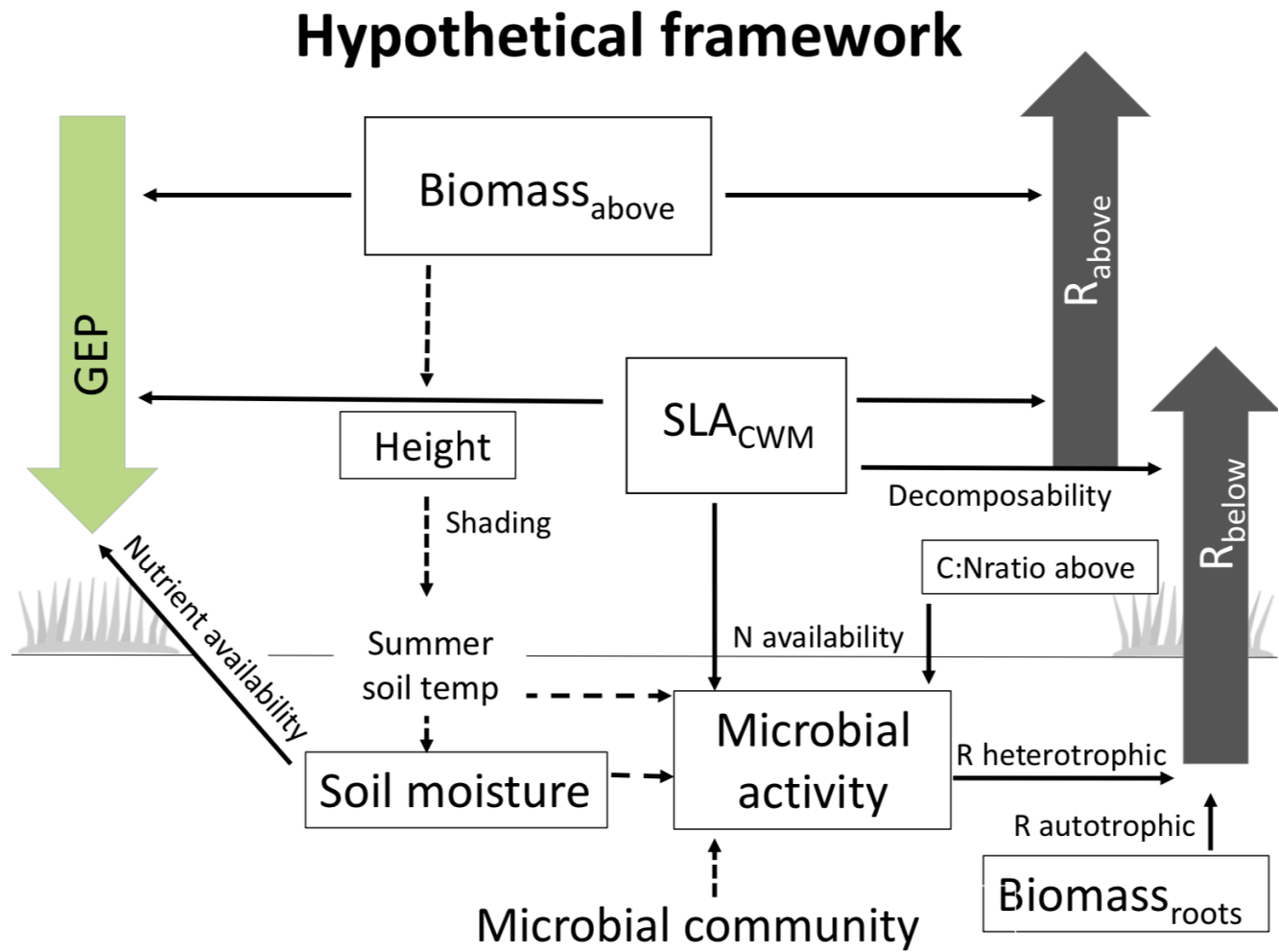


Overview

The balance between the competing processes of photosynthesis and total respiration determines whether an ecosystem is a net carbon absorber or emitter. Net ecosystem exchange (NEE) of CO_2 can be divided into three dynamic processes: GPP, aboveground respiration, R_{above} and belowground respiration, R_{below} . All of these may be affected by the changes in plant and soil microbial community composition associated with shrub expansion via effects on both the abiotic environment and biotic processes.

Here is a nice overview of controls on GEP from our Figure 1 in Sørensen et al. 2019 (Sørensen et al., 2018)

Figure 1



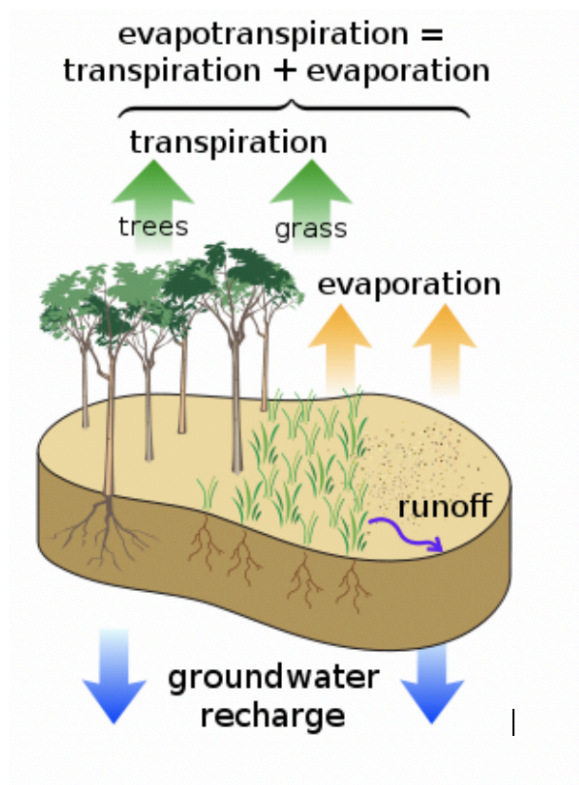
Background H2O flux and Evapotranspiration

Evapotranspiration is a measure of both evaporation of water and transpiration of water.

Evapotranspiration (ET) = sum of rate of evaporation and rate plant transpiration from surface to the atmosphere per unit area per time

$$ET = E + T$$

Evaporation (E) refers to the movement of water from sources such as the soil, canopy interception, and water bodies into the air. Transpiration (T) accounts for water movement *within* a plant and the subsequent loss of water as vapor through stomata in its leaves.



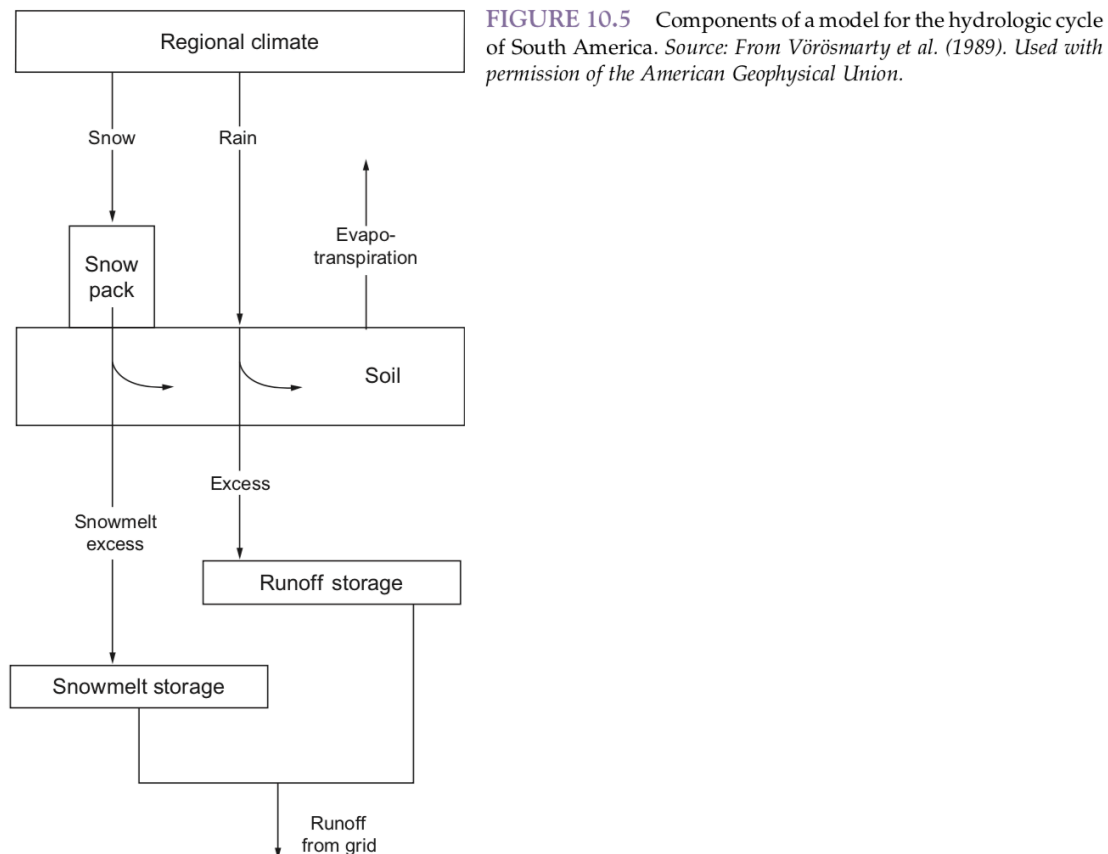
<https://en.wikipedia.org/wiki/Evapotranspiration>

From Schlesinger and Bernhardt 2013

“Evaporation removes about 413,000 km³ /yr of water from the world’s oceans (compare Syed et al. 2010). Thus, the mean residence time of ocean water with respect to losses to the atmosphere is about 3100 years. Only about 373,000 km³ /yr of this water returns to the oceans in rainfall; the rest contributes to precipitation on land, which totals 113,000 km³ /yr. Plant transpiration and surface evaporation return 63,000 +/- 13,100 km³ /yr from soils to the atmosphere (Ryu et al. 2011). The relative importance of plant transpiration (T) varies regionally (Table 10.1), but globally plants appear to be responsible for about 60% of total terrestrial evapotranspiration (ET) to the atmosphere (Choudhury et al. 1998, Alton et al. 2009), with the fraction T/ET increasing from 61% at 25% vegetation cover to 83% at 100% cover in greenhouse experiments (Wang et al. 2010a). With respect to precipitation inputs or evapotranspiration losses, the mean residence time of soil water is about 1 year.”

(Schlesinger & Bernhardt, 2013)

Figure 10.5 from Schlesinger and Bernhardt 2013 showing how ET fits into the overall water cycle.



0.5 from E

Estimating ET

Traditionally, ET is approximated using the Penman–Monteith equation. This calculated approximation of (ET), requiring as input the daily mean temperature, wind speed, relative humidity, and solar radiation. The Priestley–Taylor equation was developed as a substitute to the Penman–Monteith equation to remove dependence on observations. For Priestley–Taylor, only radiation (irradiance) observations are required.

Below is a wikipedia entry giving an overview of these ways to approximate ET

https://en.wikipedia.org/wiki/Penman%E2%80%93Monteith_equation#Equation

From Wikipedia

Like the [Penman equation](#), the **Penman–Monteith equation** (after [Howard Penman](#) and [John Monteith](#)) approximates net [evapotranspiration](#) (ET), requiring as input daily mean temperature, wind speed, relative humidity and solar radiation. Other than radiation, these parameter are implicit in the derivation of Δ , c_p , and δ_g , if not conductances below.

The United Nations [Food and Agriculture Organization](#) (FAO) standard methods for modeling evapotranspiration use a Penman–Monteith equation.^[1] The standard methods of the [American Society of Civil Engineers](#) modify that Penman–Monteith equation for use with an hourly time step. The [SWAT model](#) is one of many [GIS-integrated hydrologic models](#)^[2] estimating ET using Penman–Monteith equations.

Evapotranspiration contributions are very significant in a watershed's [water balance](#), yet are often not emphasized in results because the precision of this component is often weak relative to more directly measured phenomena, e.g. rain and stream flow. In addition to weather uncertainties, the Penman–Monteith equation is sensitive to vegetation specific parameters, e.g. [stomatal](#) resistance or conductance.^[3] Gaps in knowledge of such are filled by educated assumptions, until more specific data accumulates.

Various forms of [crop coefficients](#) (K_c) account for differences between specific vegetation modeled and a *reference evapotranspiration* (RET or ET_0) standard. Stress coefficients (K_s) account for reductions in ET due to environmental stress (e.g. [soil saturation](#) reduces [root-zone](#) O_2 , low [soil moisture](#) induces [wilt](#), [air pollution](#) effects, and salinity). Models of native vegetation cannot assume crop management to avoid recurring stress.

Our tent flux method offers a more direct way to measure ET (no approximation is required) via quantifying changes in water concentrations.

CO₂ and H₂O Flux Protocol

The flux of CO_2 and H_2O is measured by mass balance of CO_2 and H_2O in a flux chamber. The mass balance is given by $CO_2 \text{ In} = \text{Storage} + CO_2 \text{ Out}$ and $H_2O \text{ In} = \text{Storage} + H_2O \text{ Out}$. As is described in the Appendix, the mass balance is given by treating the diffusion as the sole mechanism for transport of gases from the soil and vegetation into the flux chamber. The flux of gases is described the diffusion equation. With no chamber, the rate of gas exchange can be described by Fick's Law. A non-steady state chamber restricts the volume of air available for exchange across the land surface (the plot) that is covered with the chamber. The net exchange of any gas can then be determined by the rate of change in the chamber concentration.

CO_2 flux was measured in a closed system composed of a xxxx tent chamber and a LI-7500 CO₂/H₂O infrared gas analyzer (LI-COR Inc., Lincoln, NE, USA). The Carbon flux was measured as instantaneous day- time peak uptake (ca. 10:00) and nighttime peak respiration (ca. 22:00) (see Saleska and others 1999 for diurnal measurements). Ambient CO₂ and H₂O was measured by a Li-Cor 7500 infrared gas analyzer for 30 s, and then the tent was put in place over the plot and the CO₂ concentration within the tent was measured for 90 s (Jasoni and others 2005).

Daytime measurements were only taken under cloudless conditions. The tent was designed to allow 75% of photosynthetically active radiation (as measured by the tent fabric from Shelter Systems). The air inside the tent was well-mixed by fans, and the tent chamber was sealed using a long, canvas skirt along the base of the tent, which was covered with a heavy chain.

The volume of the tent used along the gradient was 2.197 m³, whereas a smaller tent of volume 0.125 m³ was used in the watering experiment (see Sloat et al. 2015). Although the tent flux method (described by Jasoni and others 2005) does not offer the same spatial coverage as continuous flux measurements such as eddy flux towers, the portable design does allow the ability to rapidly assess carbon flux at a wider variety of locations within and across experimentally manipulated and control plots (Arnone and Obrist 2003).

The tent chamber was sealed with a canvas skirt along its base, which was covered with a chain weighing XX kg. Two fans mixed the air inside the chamber for 30 seconds before and during each measurement.

The CO₂ concentration was recorded every second throughout a 120 s light measurement followed by a 120 s dark measurement. The tent was taken off for 1 minute in between measurements so to equilibrate.

Figure from the Licor 2003 soil flux manual showing that within the tent (here soil chamber) the fluxes monitored by the Licor are water fluxes, f_w , and carbon fluxes f_c .

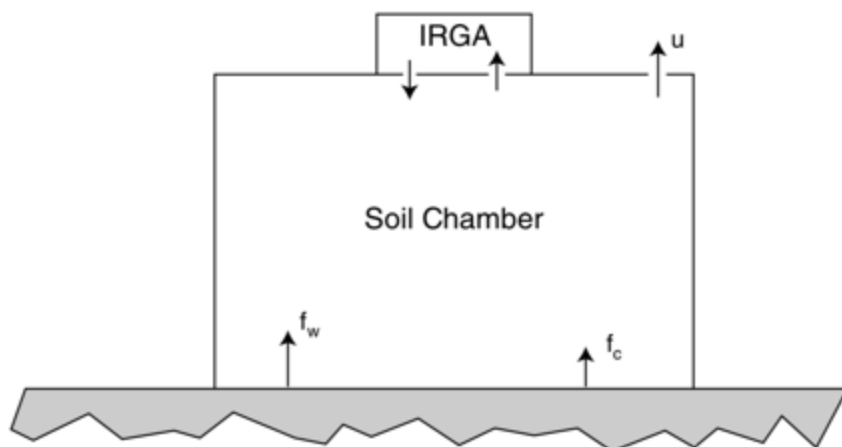


Figure 6-1. Schematic of the soil chamber. Soil evaporation f_w and soil CO₂ flux f_c add mass to the chamber, which is balanced by a flow u of air, water, and CO₂ out of the chamber.

Determining Net Ecosystem Exchange of carbon

The flux of carbon in the chamber (tent), f_c is calculated by mass balance (see Fig above and appendix equation). Within the chamber the evaporation of water, f_w , and the flux of CO₂ from soil and plant respiration and/or plant photosynthesis add mass to the chamber. If there is no leak, u , or if the leak is very small, we can calculate f_c .

From the light (daytime) measurement net ecosystem exchange (NEE) was determined, and from the dark measurements ER was determined. We then correct the CO₂ concentration for water content (C) and then used linear regression to find the CO₂ flux (Jasoni, Smith, and Arnone 2005; see also 6400-09 Soil CO₂ Flux Chamber Instruction Manual, 2003; Healy et al. 1996). Here

$C' = C / (1 - (W/1000))$, where C is the mole fraction of CO₂ (mmol mol⁻¹) and W the mole fraction of H₂O (mmol mol⁻¹).

$$C' = \frac{[CO_2](\mu mol mol^{-1})}{[H_2O](mmol mol^{-1})}$$

$$f_w = \frac{V \cdot P}{R \cdot T_{air} \cdot S} \cdot \frac{dC'}{dt}$$

Where V = volume chamber (m^2), P = air pressure (kPa; estimated to be 90 kPa at our sites at 1,100 m elevation), R = the ideal gas constant ($8.314 \text{ J mol}^{-1} \text{ K}^{-1}$), T_{air} = average air temperature ($^{\circ}\text{C}$) during the measurement, S = surface area of the plot (m^2), dC'/dt = the slope of linear regression of C' on time. Further details on these static chamber flux calculations can be found in the soil flux calculation procedure (see 6400-09 Soil CO_2 Flux Chamber Instruction Manual, 2003 <https://www.licor.com/documents/igpbvjolxbhrf52mlimp> and the [original Licor 1990 6200 Instruction Manual](#)). Equations are copied below in the Appendix.

Calculating carbon dioxide flux, dC'/dt , is the slope of the linear regression of C' on t . Flux values are then standardized for V, P, T_{air} and area S by multiplying dC'/dt by $VP/(RT_{air}S)$.

Determining chamber evapotranspiration - H2O flux is measured similarly

We corrected the H2O concentration for CO2 content (W') and then used linear regression to find the H2O flux. Here $W' = W/(1 - (C/1000))$, where W is the mole fraction of H2O ($mmol \text{ mol}^{-1}$) and C the mole fraction of CO2 ($mmol \text{ mol}^{-1}$).

The flux of water inside the tent is a measure of evapotranspiration as the increase in water vapor in the tent is due to both transpiration (T) and evaporation (E). Water flux measurements in the dark thus represent just evaporation rates, E.

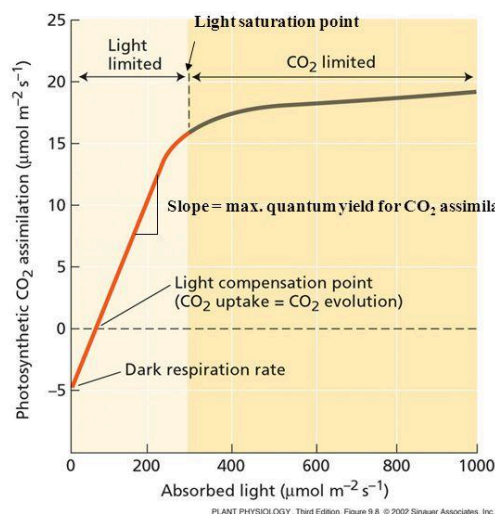
To calculate the flux of carbon dioxide dW'/dt is the slope of the linear regression of W' on t . Flux values are then standardized for V, P, T_{air} and area S by multiplying dW'/dt by $VP/(RT_{air}S)$.

Light Curve measurements

One light-curve measurement consisted of one measurement in full light, measurements at three increasing levels of shading, and one measurement in full darkness (Williams et al. 2006; Street et al. 2007). There was a gap of at least 30 s between each measurement period. The shading was done with three layers of black tulle. For dark measurements, we used an opaque hood to block out the light (Street et al. 2007).

<https://slideplayer.com/slide/3375192/>

Light response curves



- Light saturation for an individual leaf is $\sim 1/3 - 1/2$ photon flux of full sunlight

BUT

- At the whole plant level P/S is rarely saturated even in full sunlight

More Formalized Protocol for Fluxes

Once the quality of the measurements has been checked, fluxes can be calculated. The calculation of ecosystem gas fluxes within the field chamber is governed by flux theory given by (refs). The flux is calculated from the slope of the change in gas concentration inside the flux chamber. This slope, estimated by the flux_fitting function, represents the rate of change of gas concentration over time, t ,

denoted as $\frac{d[g]}{dt}$, where $[g]$ is the concentration of the target gas species.

The flux, denoted as f_{gas} , is described by the following relationship:

$$f_{\text{gas}} = -N_{\text{air}} \cdot \frac{d[g]}{dt} \quad (1)$$

Here, the rate of change of gas concentration will be negative when gas exits the chamber (e.g. CO₂ is assimilated into plants via photosynthesis), the negative sign ensures a positive flux. This equation represents a variation of Fick's Law of Diffusion, where we consider a bounded region of space (e.g., a flux chamber) and analyze the diffusion-driven flux of gas into or out of this region.

To calculate the flux specifically for carbon dioxide (CO₂) or other gases, we can extend the equation by incorporating the ideal gas law:

$$PV = NRT \quad (2)$$

where P is the air pressure (Pa), V is the chamber volume (m³), N is the number of moles of air, R is the ideal gas constant (8.314 J mol⁻¹ K⁻¹), and T is the air temperature (K). Substituting the ideal gas law, the number of moles of air N can be expressed as:

$$N = \frac{PV}{RT} \quad (3)$$

This leads to the equation for the flux of CO₂, expressed in moles per unit area per unit time, as:

$$f_{\text{CO}_2} = -\frac{1}{A} \cdot \frac{PV}{RT} \cdot \frac{d[\text{CO}_2]}{dt} \quad (4)$$

Here, A represents the area of the chamber frame base (m²). Substituting the observed slope of gas concentration change over time, slope , the flux equation simplifies to:

$$\text{flux} = \text{slope} \cdot \frac{P \cdot V}{R \cdot T \cdot A} \quad (5)$$

This equation is implemented in the `flux_calc` function. Different input and output units can be specified, and the function includes conversion factors as needed. The chamber volume can be specified as a constant or a variable, depending on the field setup. The `flux_quality` function provides the `f_slope_corr` column, which directly incorporates quality flags to either retain the slope, discard it, or replace it with zero, ensuring the robustness of flux calculations.

Field chamber with a Leaky Seal

Leaks likely influence field flux measures in the field chamber. Below, following Saleska et al. (1999 ; Saleska, S. R., Harte, J., & Torn, M. S. (1999). The effect of experimental ecosystem warming on CO₂ fluxes in a montane meadow. *Global Change Biology*, 5(2), 125-141.) calculating fluxes in a closed chamber with a leak A flux chamber with a leak can be modeled from first principles as a modified version of a chamber. The exchange of either carbon dioxide out of the chamber or tent and into the soil and plants or water vapor out of the soil and plants and into the chamber or tent can be modeled by an additional gas exchange between the interior of the tent and the outside atmospheric environment due to an imperfect seal along the floor of the tent, a hole in the tent fabric itself, or deviations in photosynthetic uptake that can occur.

To account for this scenario, we need an additional term on the right-hand side of Eq. (1) that represents this exchange of gas into or out of the tent:

$$f_{\text{gas, mol}} = -\frac{PV}{RT} \frac{d[g]}{dt} + \beta([g] - [g]_{\text{amb}}) \quad (6)$$

where we have replaced N_{air} with the ideal gas substitution $N = \frac{PV}{RT}$, and $[g]_{\text{amb}}$ represents the ambient concentration of the gas outside of the tent. The factor β is a diffusion leak "error" coefficient meant to capture leaky processes that are linearly dependent on the difference between the internal gas concentration ($[g]$) and the external ambient concentration $[g]_{\text{amb}}$. The factor β will vary depending on environmental conditions and the species composition being sampled. Thus, we must solve for β from our data to measure flux.

Eq. (4) is now a linear, first-order differential equation that can be solved exactly. The general solution is expressed as:

$$[g](t) = ([g]_0 - [g]_{\text{ss}})e^{-t/\tau} + [g]_{\text{ss}} \quad (7)$$

Here, $[g]_0$ is the concentration of gas at time $t = 0$ (interpreted as the beginning of the measurement window). The term $[g]_{\text{ss}}$ is the steady-state concentration of gas within the tent if the exchange were to continue indefinitely and is equal to:

$$[g]_{\text{ss}} = [g]_{\text{amb}} - \frac{f_{\text{gas, mol}}}{\beta} \quad (8)$$

Despite having an exact expression, $[g]_{\text{ss}}$ is typically solved numerically from a dataset. Finally, τ is called a time constant, and it signifies the time scale over which diffusion processes occur. It is equal to

$\tau = \frac{PV}{\beta RT}$. Like the steady-state concentration, the time constant is solved numerically (and simultaneously) from the data.

As Eq. (5) gives us a relationship for the instantaneous gas concentration, we can approximate the gas

flux using the instantaneous rate of change of the gas concentration $\frac{d[g]}{dt}$ evaluated at ambient gas concentration ($[g] = [g]_{\text{amb}}$). The flux we are interested in calculating now takes the form:

$$f_{\text{gas, mol}} = \frac{1}{A} \cdot \frac{PV}{RT} \cdot \frac{\partial}{\partial t}[g](t) \Big|_{[g]=[g]_{\text{amb}}} \quad (9)$$

Substituting Eq. (5) into the above expression results in an exact formulation of flux that accounts for leaky tent processes.

Unit Conversion

The units of the Ideal Gas constant $R = 8.314 \text{ Joules/mol/Kelvin}$ require that we convert all LiCOR data output into SI units. Examining some sample LiCOR 7500 data output in Figure 2, we can infer that the temperature is measured in degrees Celsius, given its approximate numerical value of 18, and that pressure is measured in kiloPascals, given its approximate numerical value of 74. Thus, both of these values will need to be converted by adding 273.15 to 18 to convert to the Kelvin scale and multiplying 74 by 1000 to convert to Pascals.

Conversion from 'wet' to 'dry' units

One final correction that must be made to the measured water vapor H_2O and carbon dioxide CO_2 is related to the fact that these measurements are occurring in wet air as opposed to dry air.

Phenomenologically, when water vapor is present in the air being analyzed by the gas spectrometer within the LiCOR machine, it introduces an error. This error occurs because water vapor absorbs heat from the light used to analyze the gases. To correct for this effect, both the CO_2 and H_2O concentrations need to be transformed as follows:

$$[CO_2]_{\text{dry}} = \frac{[CO_2]_{\text{wet}}}{1 - \frac{[H_2O]_{\text{wet}}}{1000}} \quad (10)$$

$$[H_2O]_{\text{dry}} = \frac{[H_2O]_{\text{wet}}}{1 - \frac{[H_2O]_{\text{wet}}}{1000}} \quad (11)$$

where $[CO_2]_{\text{wet}}$ and $[H_2O]_{\text{wet}}$ correspond to raw LiCOR outputs.

Appendix (Equations other resources)

The R package `gpp` From `flux v0.3-0` by Gerald Jurasinski

Model GPP From `CO2 Closed Chamber Flux Data` and includes light correction equations with citations.

<https://www.rdocumentation.org/packages/flux/versions/0.3-0/topics/gpp> see also

<https://rdr.io/cran/flux/man/flux-package.html>

Details

The function models the relationship between CO_2 uptake by plants (gross primary production, GPP) and irradiation using one out of 4 methods (Falge et al. 2001). Per default the Michaelis-Menten kinetic (e.g., Schmitt et al. 2010) is used. The following models can be fitted to the data:

$$GPP = (GPmax * alpha * PAR) / (alpha * PAR + GPmax) \text{ (Michaelis-Menten)}$$

$$GPP = (alpha * PAR) / (1 - (PAR/2000) + (alpha * PAR/GPmax)) \text{ (Falge)}$$

$$GPP = (alpha * PAR * GPmax) / \sqrt{GPmax^2 + (alpha * PAR)^2} \text{ (Smith)}$$

$$GPP = GPmax * (1 - \exp(-(alpha * PAR)/GPmax)) \text{ (Misterlich)}$$

with PAR the incoming light (irradiation). Note, that irradiation can be given in PAR or in PPFD although the equation states PAR . $GPmax$ and $alpha$ are the parameters that are fitted. $GPmax$ refers to the maximum gross primary production at saturating or optimum light whereas $alpha$ refers to the ecosystem quantum yield and gives the starting slope of the model.

Transparent closed chamber measurements in the field typically capture net ecosystem exchange (NEE), which is the sum of the two opposing processes ecosystem respiration (R_{eco}) and GPP. Therefore, it is necessary to subtract modeled R_{eco} from the measured NEE to obtain GPP that can be used for the modelling against irradiance.

Real R_{eco} at the time of the NEE measurement is typically unknown because dark and light measurements cannot be taken at the same spot at the same time. Therefore, R_{eco} has to be modelled based on dark chamber or nighttime measurements (see $reco$). For modelling GPP from NEE chamber measurements, gpp just needs measured NEE , the associated irradiance (PAR) and temperature ($PAR.Temp$) values and the R_{eco} model(s) ($Reco.m$). The R_{eco} model(s) can derive from a longer period of time than the NEE data, which is often better to get more reliable models. In contrast, $gpp2$ extracts R_{eco} fluxes from actual measurements.

Additional citations

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6 Equation Derivation

The mass balance of CO₂ for the 6400-09 soil flux chamber (Figure 6-1) is given by

$$\text{CO}_2 \text{ In} = \text{Storage} + \text{CO}_2 \text{ Out}$$

$$sf_c = \rho v \frac{\partial c}{\partial t} + uc \quad 6-1$$

where s is the soil surface area (m²) enclosed by the chamber, v is the volume (m³) of the chamber and IRGA, f_c is the flux of CO₂ coming from the soil surface (mol CO₂ m⁻² s⁻¹), ρ is the density of the air (mol m⁻³), c is the CO₂ concentration (mol CO₂ mol⁻¹), and u is the flow rate (mol s⁻¹) of escaping air from the system, largely due to soil evaporation into the system.

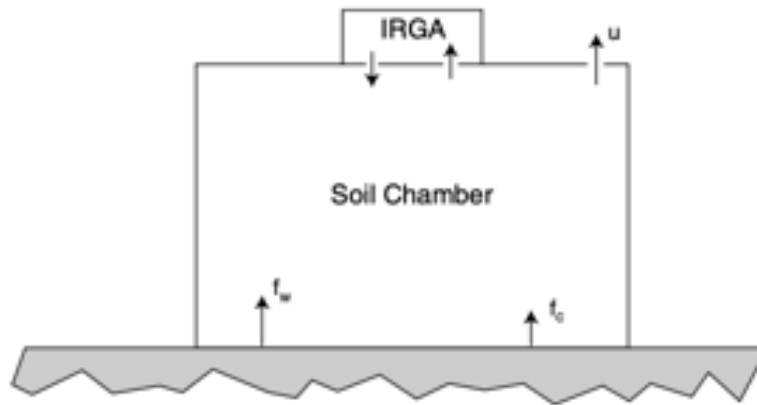


Figure 6-1. Schematic of the soil chamber. Soil evaporation f_w and soil CO₂ flux f_c add mass to the chamber, which is balanced by a flow u of air, water, and CO₂ out of the chamber.

The mass balance of water vapor is given by

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Water In = Storage + Water Out

$$sf_w = \rho v \frac{\partial w}{\partial t} + uw \quad 6-2$$

where f_w is the flux of H₂O coming from the soil surface (mol H₂O m⁻² s⁻¹), and w is the water vapor concentration (mol H₂O mol⁻¹).

If we assume that evaporation is the sole cause of the leakage, then $u = sf_w$, and we can write

$$\begin{aligned} sf_w &= \rho v \frac{\partial w}{\partial t} + sf_w w \\ &= \frac{\rho v}{(1-w)} \frac{\partial w}{\partial t} \end{aligned} \quad 6-3$$

Substituting this for u in (6-1) leads to

$$sf_c = \rho v \frac{\partial c}{\partial t} + \frac{\rho v c}{(1-w)} \frac{\partial w}{\partial t} \quad 6-4$$

Collecting terms leads to

$$f_c = \frac{\rho v}{s} \left(\frac{\partial c}{\partial t} + \frac{c}{(1-w)} \frac{\partial w}{\partial t} \right) \quad 6-5$$

Equation (6-5) takes a slightly different form as implemented in the LI-6400, since the measured and entered parameters have different units, and density must be computed from temperature and pressure. The terms are defined in Table 6-1.

$$c \left(\frac{\text{mol CO}_2}{\text{mol air}} \right) = C \left(\frac{\mu\text{mol CO}_2}{\text{mol air}} \right) \times 10^{-6} \left(\frac{\text{mol}}{\mu\text{mol}} \right) \quad 6-6$$

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$$w\left(\frac{\text{mol H}_2\text{O}}{\text{mol air}}\right) = W\left(\frac{\text{mmol H}_2\text{O}}{\text{mol air}}\right) \times 10^{-3} \left(\frac{\text{mol}}{\text{mmol}}\right) \quad 6-7$$

$$\rho\left(\frac{\text{mol}}{\text{m}^3}\right) = \frac{P(\text{kPa}) \times 10^3 \left(\frac{\text{N / m}^2}{\text{kPa}}\right)}{8.314 \left(\frac{\text{Nm}}{\text{mol K}}\right) (T_c + 273(\text{K}))} \quad 6-8$$

$$\frac{v}{s} \left(\frac{\text{m}^3}{\text{m}^2}\right) = \frac{V(\text{cm}^3) \times 10^{-6} \left(\frac{\text{m}^3}{\text{cm}^3}\right)}{S(\text{cm}^2) \times 10^{-4} \left(\frac{\text{m}^2}{\text{cm}^2}\right)} \quad 6-9$$

Substituting Equations 6-6 through 6-9 into 6-5 yields

$$F_c = \frac{kPV}{S(T_c + 273)} \left(\frac{\partial C}{\partial t} + \frac{C}{(1000 - W)} \frac{\partial W}{\partial t} \right) \quad 6-10$$

where $k = 10/8.314 = 1.2028$.

Note, the above derivations likely stem from Healy et al. 1996

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Table 6-1. The terms of the LI-6400's soil flux equation (6-10).

Symbol	Description	Units	Soil Config LABEL
F_c	Soil flux of CO ₂	$\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$	EFFLUX
P	Atmospheric pressure	kPa	Press_kPa
V	Total system volume	cm ³	Vtot
S	Enclosed soil area	cm ²	Area
T_c	Air temperature in chamber	°C	Tsch_C
C	CO ₂ concentration	$\mu\text{mol CO}_2 \text{ mol}^{-1}$	CO2S_μml
W	H ₂ O concentration	$\text{mmol H}_2\text{O mol}^{-1}$	H2OS_mml
$\frac{kP}{(T_c + 273)} \left(\frac{\partial C}{\partial t} + \frac{C}{1000 - W} \frac{\partial W}{\partial t} \right)$			dc'/dt

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