ESPM 228, Advanced Topics in Biometeorology and Micrometeorology

Lecture 2 on Micrometeorological Flux Measurement Methods/ Flux-Gradient Theory

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Instructor:
Dennis Baldocchi
Professor of Biometeorology
Ecosystems Science Division
Department of Environmental Science, Policy and Management
345 Hilgard
Baldocchi@berkeley.edu
642-2874

Micrometeorological Flux Methods

"We're not in Kansas anymore, Toto", Dorothy in the **Wizard of Oz**

'Micrometeorologists are like Dorothy, in the Wizard of Oz; they always want to go back to Kansas', quote attributed to David Fitzjarrald, SUNY-Albany

Studies on topics associated with biogeochemistry, ecology, and atmospheric sciences require information on mass, energy and momentum exchange between the ecosystems and atmosphere on a continuous basis. Yet, how can we make such measurements, short of enclosing the ecosystem in a huge cuvette?

Micrometeorological methods provide the primary means of measuring mass and energy exchange across the ecosystem-atmosphere interface. We give homage to Kansas because many of the micrometeorological flux methods in vogue were evaluated in Kansas during pioneering studies in the late 1960s (Kaimal and Wyngaard, 1990), and later were applied to study ecosystem processes during large field campaigns in the 1980s such as FIFE and CASES-99 (Sellers and Hall, 1992).

Micrometeorological methods have many advantages. They are:

1) in situ, so they are non-intrusive;

2) they can be applied on a quasi-continuous time basis;

3) measurements made a point represent an areally-averaged ensemble of mass and energy exchange,with a length scale of 100 m to 2 km across a flux footprint..

While micrometeorological methods have many attributes, these attributes rely on many assumptions on the state of the atmosphere. The basis for formulating any micrometeorological method is the conservation of a trace gas scalar as it is transferred through the atmosphere. In principle, this is a three-dimensional process, which is varying in time. In practice we impose many restrictions upon the application of the conservation equation and force it to describe surface micrometeorology as a one-dimensional processes during steady conditions. Important assumptions on which the simplified method is based include:

1) steady-state conditions
2) horizontally homogeneity in source-sinks
3) flat topography.

Violations in these assumptions must be considered through experimental design and ancillary measurements. These topics will be addressed later when discussing application of the method.

Ergodicity

The expected value of a function, f, distributed across space, is defined by integrating the function across the infinite spatial domain and multiplying by the probability density, p.

Equation 1

$$E[f(x)] = \int_{-\infty}^{\infty} f(x)\, p(x) dx = \int_{-\infty}^{\infty} x\, f\left(x\right) dx$$

It is impossible to measure meteorological properties everywhere. So meteorologists invoke the principle of ergodicity using temporal averaging to represent spatial averaging by invoking Taylor's frozen eddy hypothesis, where x = u t.

Equation 2

$$E[f(x)] = \int_{-\infty}^{\infty} f(x)\, p\left(x\right) dx \approx \frac{1}{2UT} \int_{-T}^{T} ut\, f(t) dt$$

Often ecologists and statisticians criticize micrometeorologists for employing only one tower and claiming their work suffers from pseudo-replication (Hurlbert, 1984). They claim that a measurement with one tower does not allow the experimenter to measure the ensemble average. One rebut from the micrometeorologist is that we can invoke the Principle of Ergodicity, where **space-based** ensemble averages can be substituted with **temporal averaging**.

Equation 3

$$E[f(x)] = \int_{-\infty}^{\infty} f(x)\, p(x) dx \approx \frac{1}{UT} \int_{0}^{T} u(t)\, f(t) dt$$

In recent years, there has been a growing body of literature defending large scale studies (Carpenter et al., 1995; Oksanen, 2001; Schank and Koehnle, 2009). Oksanen (2001), for example, concludes that 'The concept pseudo-replication amounts to entirely unwarranted stigmatization of a reasonable way to test predictions referring to large-scale systems'. The power of micrometeorological flux measurements is their ability to measure the short and long-term dynamics of state variables, and when applied in a paired fashion two contrasting ecosystems, many hundred hectares in area (Hsieh et al., 2000; Schmid, 2002). Micrometeorological methods sample a large footprint, giving us the ability to measure the state of the general population, rather than deducing its expected value statistically with a replicated set of small and infrequent samples. We contend that repeated temporal sampling of meteorological variables is not pseudo-replication, as suggested by Hurlbert (1984), for two reasons. First, the theory of ergodicity tells us we can substitute spatial sampling with temporal sampling (Panofsky and Dutton, 1984). And, second, we find that successive temporal measurements were not auto-correlated; the sub-hourly temperature measurements were uncorrelated after 12 hours and the daily-averaged data became uncorrelated after 90 days. Consequently, if we collect a large number of independent statistical samples of temperature at the two contrasting field sites that were representative of large areas. We also contend that repeated temporal sampling reduces the sampling error and increases the precision in detecting differences between two treatments (Moncrieff et al., 1996).

Flux-Gradient Method

The flux-gradient technique is an application of Fick's Law of Diffusion to the turbulent atmosphere. It infers the vertical flux density of mass across a virtual plane between the biosphere and the atmosphere is a function of an eddy diffusivity, K ($m^2\ s^{-1}$), times the vertical gradient in the potential, here defined in terms of mixing ratio, c:

Equation 4

$$F = -\rho_a K_c\, \partial\bar{c} / \partial z$$

To have units consistent with a flux density, mol $m^{-2}\ s^{-1}$, we multiple K times the gradient by the density of dry air. The negative sign is imposed so fluxes towards the atmosphere are positive, a gain, and those from the atmosphere are negative, a loss. We can substitute c with temperature times specific heat at constant pressure (C_p) to compute sensible heat exchange. We can substitute c with wind velocity, u, to compute momentum flux density.

Flux-gradient theory borrows from the theory describing fluxes of material by molecular diffusion, which is a function of the molecular weight of the compound. For turbulent transfer, it

is critical to state that the turbulent diffusivity **is not** a function of molecular diffusivity because turbulence motions are many orders of magnitude greater than molecular motions. K has values on the order of 1 while diffusivity has units on the order of 10^{-5}.

So in principle the eddy diffusivities of momentum, heat, water vapor, and other trace gases (K, $m^2\ s^{-1}$) are assumed to be identical, e.g. Reynold's Analogy

Equation 5

$$K_m = K_H = K_v = K_c$$

On the other hand, eddy diffusivities do vary among different scalars when the sources and sinks vary from one another; we will discuss this occurrence below.

The specific equation for the flux-densities of momentum (τ), sensible heat (*H*), latent heat (λE) and a scalar (F_c), such as CO_2, are:

Equation 6

$$\tau = -\overline{\rho_a} K_m \, \partial \bar{u} / \partial z$$

(kg $m^{-2}\ s^{-1}$)

Equation 7

$$H = -\overline{\rho_a} C_p K_h \, \partial \bar{\theta} / \partial z$$

(J $m^{-2}\ s^{-1}$)

Equation 8

$$\lambda E = -\overline{\rho_a} \frac{\varepsilon}{P} \lambda K_v \, \partial \bar{e} / \partial z$$

(J $m^{-2}\ s^{-1}$)

Equation 9

$$F_c = -K_c \, \partial \overline{\rho_c} / \partial z$$

(mole $m^{-2}\ s^{-1}$)

In these equations, u is horizontal wind velocity, $\bar{\theta}$ is mean potential temperature, $\overline{\rho_a}$ is mean air density, C_p is specific heat of air at constant pressure, λ is the latent heat of vaporization, e is

vapor pressure, P is pressure, ε is the ratio of the molecular masses for water vapor and dry air, and $\overline{\rho_c}$ is CO2 mole density (mole m^{-3}).

The sign convention produces positive values of flux densities when they are directed to the atmosphere, as they are adding material to the atmosphere. Negative values for flux densities are associated with losses from the atmosphere. The negative sign is therefore applied to the flux gradient equations as negative gradients are associated with conditions where the state, concentration or mixing ratio of the upper level is less than that closer to the surface.

Application of this method is dependent upon methods of measuring the vertical concentration gradients directly and assessing eddy exchange coefficients indirectly.

Eddy Exchange Coefficients

As mentioned above, one basic assumption of applying the flux-gradient technique is that the sources and sinks of scalars (and possibly momentum) are equal. The basis for this assumption is attributed to Reynolds. If this assumption is true then we can assume Reynolds similarity, $K_v=K_h=K_c=K_m$.

By exploiting this assumption, we can derive several methods for assessing the eddy exchange coefficient. The most notable methods for evaluating K are the:

1) aerodynamic method;
2) energy balance method;
3) 'direct' method.

The need to measure gradients well and to chose the proper method of assessing K cannot be overemphasized. Errors in the flux-gradient method are strongly linked to errors in the gradient measurements, as well as the technique used to assess the eddy exchange coefficients. Verma and Rosenberg (Verma and Rosenberg, 1975) and Sinclair et al. (Sinclair et al., 1975) have conducted detailed error analyzes of the method. They found that errors can reach 30 to 40% when an instrument system is not be able to measure the vertical gradients well, as over forests that are aerodynamically rough, or over vegetation that may not be transpiring much.

Aerodynamic Method:

The aerodynamic method starts with the assessment of momentum transfer, τ, and the measurement of the wind speed gradient.

Equation 10

$$\tau = -\overline{w'u'} = -\overline{\rho_a} u_*^2 = -\overline{\rho_a} K_m \frac{\partial \bar{u}}{\partial z}$$

The eddy exchange coefficient for momentum ($m^2\ s^{-1}$) can be described in terms of the friction velocity, u* and the wind gradient.

Equation 11

$$K_m = \frac{u_*^2}{\frac{\partial \bar{u}}{\partial z}}$$

From boundary layer theory we recall that the wind velocity gradient is also a function of friction velocity.

Equation 12

$$\frac{\partial \bar{u}}{\partial z} = \frac{u_*}{kz}$$

We can now eliminate friction velocity and produce a relation for Km that is solely a function of height and the wind gradient.

Equation 13

$$K_m = k^2 z^2 \frac{\partial \bar{u}}{\partial z}$$

And the coefficient k is von Karman's constant, 0.4. The derivation so far is for neutral thermal stratification and short vegetation.

In practice we must consider stability effects and the zero plane displacement, d, so the equations get a bit more complicated

Equation 14

$$\frac{\partial \bar{u}}{\partial z} = \frac{u_*}{k(z-d)} \phi_m((z-d)/L)$$

Further algebraic manipulation yields:

Equation 15

$$K_m = k^2 (z-d)^2 \frac{\partial \bar{u}}{\partial z} \varphi(\frac{z-d}{L})^{-2}$$

Note that this version contains the diabatic stability correction function that depends on the Monin-Obukhov length scale L and information on the zero plane displacement, d.

Alternatively, if one has a direct measure or estimate of friction velocity, from the covariance between w and u, we can apply

Equation 16

$$K_m = k\, u_* (z-d) / \phi_m(\frac{z-d}{L})$$

Equation 17

$$K_h = k\, u_* (z-d) / \phi_h(z/L)$$

The Monin-Obukhov (M-O) Similarity Theory enables us to predict the behavior of wind profiles under conditions of neutral, stable and unstable thermal stratification in the surface boundary layer (Foken, 2006; Hogstrom, 1996), but not in the mixed or planetary boundary layer. Conceptually, a non-dimensional wind velocity gradient is defined and it is a function of a non-dimensional height, *z/L*:

Equation 18

$$\varphi_m(\frac{z}{L}) = \frac{k\, z}{u_*} \frac{\partial u}{\partial z}$$

And for tall vegetation we define the non-dimensional wind shear in terms of the zero-plane displacement:

Equation 19

$$\varphi_m\left(\frac{(z-d)}{L}\right) = \frac{\partial u}{\partial z}\frac{k(z-d)}{u_*}$$

In Equations 18 and 19, *L* is the Monin-Obukhov length scale. It is defined using the turbulent kinetic energy budget or by using scaling arguments (e.g. Buckingham Pi theory). From a physical view point, *z/L* is the ratio between of the buoyant production of turbulent kinetic energy, $\frac{-g\overline{w'\theta_v'}}{\overline{\theta_v}}$, to the shear production ($\overline{w'u'}\frac{\partial \bar{u}}{\partial z}$).

Monin-Obukhov theory says little about the behavior of φ_m with *z/L*. This information must be obtained from experimentation and empirical evidence (Businger, 1971; Foken, 2006; Hogstrom, 1988). It does, however, give us a framework for synthesizing wind and turbulence data.

The functional form of the phi function for momentum is illustrated in Figure 11. The 'phi' function has 3 asymptotic limits:

1. Under neutral conditions *z/L* approaches zero and 'phi' approaches 1.

2. Under unstable conditions *z/L* is less than zero and 'phi' gradually approaches an asymptote near 0.4, as *z/L* becomes more negative.

3. Under stable conditions, *z/L* is positive, but remains less than about 0.25. Values for 'phi' increase rapidly with small changes in *z/L*.

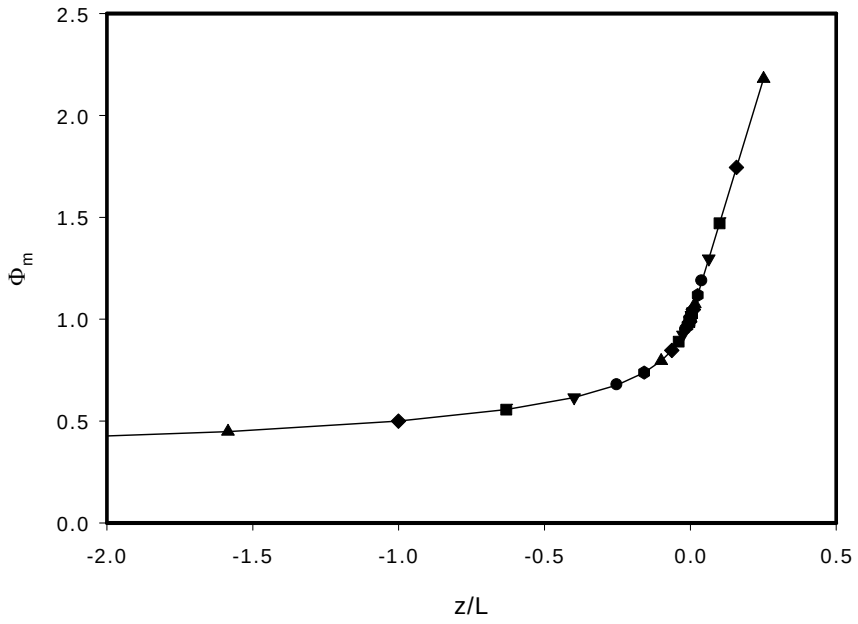

Figure 1

As *z/L* becomes more negative wind speeds decrease and free convection occurs. In this situation friction velocity is not the appropriate scaling velocity. Instead, a convective scaling velocity (w*) is more relevant (Hogstrom, 1996; Wyngaard, 1992).

Equation 20

$$w_* = (z_i (g / \theta_v) \overline{w' \theta_v'})^{1/3}$$

Under stable conditions, the dimensionless groups are independent of height and there is a decoupling between turbulent flow at various layers (Mahrt, 1999); this condition is called z-less stratification. Elevated wind jets are observed a few tens of meters above the surface, where the local wind velocity may be relatively calm. Jets occur because momentum of the free air in the surface boundary layer must find a sink at the ground at night. To accomplish this feat, wind shear must increase (see Figure 1) as stable thermal stratification intensifies, in order to compensate for reduced turbulent mixing under stable conditions (Mahrt, 1999).

Empirical algorithms for φ_m typically follow the form:

Equation 21

$$\phi_m (z / L) = (1 - \gamma z / L)^{\beta}$$

Model coefficients for Equation 19 from selected field studies are listed for unstable (Table 1) and stable (Table 2) conditions; these data are extracted from the pioneering study of Businger (1971) and a grand average of studies, based on the reviews and studies of Hogstrom (1988; 1996).

Table 1 Parameters for Phi functions for momentum transfer, unstable thermal stratification

Citation	k	γ	β
(Businger, 1971)	0.35	-15	-1/4
(Hogstrom, 1996)	0.40	-19	-1/4

Table 2 Parameters for Phi functions for momentum transfer, stable thermal stratification

Citation	k	γ	β
(Businger, 1971)	0.35	4.7	1
(Hogstrom, 1996)	0.40	5.3	1

Monin-Obukhov theory can also be used to define dimensionless gradients for potential temperature and other scalars (Foken, 2006; Hogstrom, 1996):

Equation 22

$$\varphi_h(\frac{(z-d)}{L}) = \frac{\partial\theta}{\partial z}\frac{k(z-d)}{\theta_*}$$

When designing an experiment, rules of thumbs are important and helpful for guestimating K, fluxes and the gradients that need to be resolved. A good guess for K can come by applying 0.4 for k, estimating friction velocity as 0.125 times wind speed and assuming that d is about 60 to 70% of canopy height, h.

Energy Balance Method:

Investigators have used the energy balance method for over forty years to assess mass and energy exchange between vegetation and the atmosphere (Baldocchi et al., 1981; Biscoe et al., 1975; Denmead, 1969; Verma and Rosenberg, 1975). The relation starts with the formula for the net radiation balance, where the net radiation flux density (R_n) is partitioned in to flux densities for sensible (H) and latent heat (λE) and conductive heat transfer into the soil (G):

Equation 23

$$R_n = H + \lambda E + G$$

$$(\text{J m}^{-2}\ \text{s}^{-1})$$

Substituting the Flux-Gradient relations, shown at the introduction of this topic, with the energy balance equation yields:

Equation 24

$$K = \frac{R_n - G}{\overline{\rho_a}(C_p \frac{\partial \theta}{\partial z} + \frac{\varepsilon}{P}\lambda \frac{\partial e}{\partial z})}$$

Of course this calculation assumes that $K_v = K_h$. If this relation is applied to forest canopies with substantial standing biomass we must add a heat storage term, S, to the rhs of the energy balance equation.

Error analyzes of the method suggest that errors for computing latent heat and CO_2 flux densities over crops are on the order of 10 and 15%, respectively (Sinclair et al., 1975; Verma and Rosenberg, 1975). The relative error in the Bowen ratio measurement method is a function of the standard deviations of K and the vertical gradients:

Equation 25

$$\frac{\sigma F_c}{F_c} = \sqrt{(\frac{\sigma K}{K})^2 + (\frac{\sigma(\Delta \rho_c)}{\Delta \rho_c})^2}$$

The error associated with deducing K is a function of the errors of the radiation balance and the errors in measuring gradients of temperature and humidity.

Equation 26

$$\frac{\sigma K}{K} \approx \sqrt{(\frac{\sigma(Rn-G)}{Rn-G})^2 + (\frac{\sigma(\Delta\rho_v)}{\Delta T + \Delta\rho_v})^2 + (\frac{\sigma(\Delta T)}{\Delta T + \Delta\rho_v})^2}$$

Pitfalls

Extreme care must be used when applying this method to scalars as many workers (Businger, 1971; Dyer and Hicks, 1970; Pruitt et al., 1973) show that Reynolds analogy fails. Empirical evidence shows that a more common observation is that

$$K_m \neq K_v = K_h = K_c = K_x$$

Under near neutral conditions, one assumption is:

$$1.35K_m = K_v = K_h = K_c = K_x$$

Momentum transfer is affected by pressures forces, which do not play a role in mass transfer, is one argument for the different diffusivity values. Pressure allows momentum to interact with its surroundings more efficiently.

And one can only assume

$$K_v = K_h = K_c = K_x$$

when the sources and sinks are equal.

Eddy diffusivities are not a function of molecular diffusivity, as has been mistaken from time to time (Glotfelty et al., 1983; Hicks et al., 1984).

Differential source-sink locations and processes are a reason why K_m does not equal the scalar values. Investigators have developed algorithms, that are stability dependent, to correct these values for one another. Empirical relations published in the literature include:

Unstable

(Dyer and Hicks, 1970)

$$\frac{K_h}{K_m} \approx \frac{K_w}{K_m} = (1 - Ri)^{0.25}$$

$$\frac{K_h}{K_m} \approx \frac{K_w}{K_m} = (1 - 16 \mid z / L \mid)^{0.25}$$

Stable

$$\frac{K_h}{K_m} \approx \frac{K_w}{K_m} = 1 \text{ (Webb, 1970)}$$

Though the review by Hogstrom (Hogstrom, 1996) indicates that the error noted from the Kansas experiment is not 1/0.75 but closer to 1/0.9.

Eddy exchange coefficients will differ from one another when there is significant spatial separation in sources and sinks. Model computations of K_w, K_h and K_c over a tall temperate forest demonstrate the impact of this point. The sources of heat and vapor tend to be co-located, sunny leaves in the upper portion of the canopy. So assuming K_w equals K_h is a good assumption. CO_2 exchange, however, involves leaf photosynthesis by these active leaves and soil respiration, about 20 m away. In this case K_c does not equal K_w and is smaller by 10 to 20 %. One will also expect K_w and K_h to deviate from one another after a rain event causes the soil to be wet. Then most sensible heat will be generated by the sunlit leaves while evaporation will occur from the leaves and the soil. In this case K_c and K_w may approach one another.

There is some theoretical appeal to estimating exchange coefficient corrections in terms of Ri rather than z/L due to the effects of autocorrelation in assessing turbulent fluxes and z/L (hicks). This linkage and criticism should be explored whenever you are comparing two derived variables.

Another factor that may cause the sources and sinks of heat and water transfer to be different is sensible heat advection, as occurs over irrigated fields in arid and semi-arid environments. There is considerable controversy on this topic. A team of scientists from Nebraska (Motha et al., 1976) have reported that K_h does not equal K_w and exceeds it when there are inverted temperature gradients over actively transpiring crops. Another group of worker, studying irrigated rice in Australia (Lang et al., 1983) report divergent results, too. But they find relation between K_h and K_v is reversed. The Australian group explains their results as arising from the effect of large eddies being associated with the downward transport of sensible heat and moisture. Since the profiles of these two scalars are different due to insufficient adjustment in an evolving boundary layer, the mechanism for transport differs. Their hypothesis is supported with model calculations derived from Rao et al. (Rao et al., 1974) ($K_h < K_v$) and observations that the correlation between T and q fluctuations is weak.

In advective conditions one can assume the flux is the sum of contributions for the dry and wet areas, with proper weighting considered:

Equation 27

$$\frac{\overline{\rho_a} C_p K_h \partial \overline{T} / \partial z}{\overline{\rho_a} \lambda K_v \partial q / \partial z} = \frac{\overline{\rho_a} C_p K_{h,dry} \partial \overline{T_{dry}} / \partial z + \overline{\rho_a} C_p K_{h,wet} \partial \overline{T_{wet}} / \partial z}{\overline{\rho_a} \lambda K_{v,dry} \partial q_{dry} / \partial z + \overline{\rho_a} \lambda K_{v,wet} \partial q_{wet} / \partial z}$$

Assuming that the upwind dry area has no evaporation and the upwind humidity gradient is zero allows a simplification of this relation to yields:

Equation 28

$$\frac{K_h}{K_v} = \frac{1 + \frac{K_{dry}}{K_{wet}} \left(\frac{\partial T_{dry}}{\partial z} / \frac{\partial T_{wet}}{\partial z} \right)}{1 + \left(\frac{\partial T_{dry}}{\partial z} / \frac{\partial T_{wet}}{\partial z} \right)}$$

The ratios of the temperature gradients are negative in an advective situation. The dry area is more convective so it will be associated with larger eddies, causing $K_{dry}/K_{wet} > 1$. This results in $K_h/K_v < 1$.

This theory does not explain the Nebraska results, however. McNaughton and Laubach revisited the problem recently using a theory that divided eddy fluctuations into active and inactive components:

$$x = \overline{x} + x_a' + x_i'$$

The active and inactive components are defined by applying high and low pass filters to the data. The inactive component is the long term, slow component and is associated with planetary boundary layer motions.

They found that $K_h > K_v$ when the Bowen ratio is small and negative, as was observed by the Nebraska team. McNaughton and Laubach (1998) conclude that 2 mechanisms account for differences between K_h and K_w. One factor is the lack of complete boundary layer adjustment, as when one is in an evolving and advective boundary layer. The second occurs from non-steady winds that are associated with high evaporation rates and large saturation deficits.

Another factor to consider is the co-location of sources and sinks. With the CANVEG model, that considers non-local turbulent transfer, we can see that the exchange coeffient for CO2 which has a soil source and vegetative sink, does not equal the exchange coefficient for water, where most of the source is transpiration from the vegetation. The exchange coefficient for heat is collocated with water as energy is absorbed by the upper vegetation.

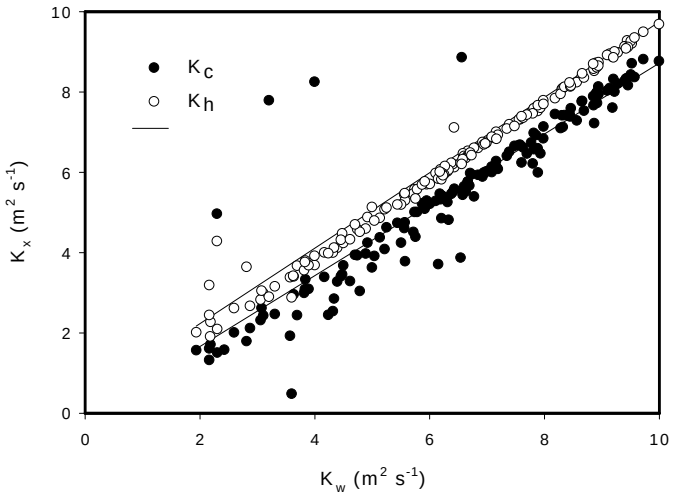

Figure 2 Calculations of the eddy exchange coefficients for water, heat and CO2 over a tall temperate forest with a Lagrangian random walk model. Note how Kh is similar to Kv because the sources of heat and vapor come from the upper foliage. In contrast, Kv and Kc differ because there is a strong source of CO2 at the soil, as well as a sink in the upper foliage.

Roughness sublayer

K theory is notorious for its failure in a zone called the roughness sublayer. It is a region in the internal surface boundary layer, immediately adjacent to the vegetation. The roughness sublayer can extend 2 to 3 canopy heights, as it is directly affected by the influence of local trees or plants. In this zone Monin-Obukhov similarity theory fails.

When applying Flux-gradient theory, one can in principle avoid errors that are introduced by working in the roughness sublayer by making measurements above it. Two extra problems arise with this approach. One, we need larger fetch. Two, the gradients become weaker and harder to detect.

Investigators commonly examine the roughness enhancement factor. It is the ratio of the observed K to that computed from Monin-Obukhov scale theory

Equation 29

$$\gamma_e = \frac{K_{obs}}{K_{mo}} = \frac{\phi(z/L)_{mo}}{\varphi_{obs}}$$

Classic and pioneering work by Raupach (Raupach, 1979) indicated that the enhancement factor was on the order of 2. Others (Cellier and Brunet, 1992; Simpson et al., 1998) revisited the problem with modern instrumentation and were able to study the problem over the course of a growing season show a distinct effect of leaf area and thermal stratification.

Table 3 Enhancement factors for Kco2, Forest with foliate canopy.

Derived from Simpson et al. 1998

height	Unstable	Near neutral	Stable
1.9-2.2h	0.92	1.18	1.18
1.6-1.9h	1.23	1.27	1.52
1.4-1.6h	1.64	1.31	1.49
1.2-1.4h	1.60	1.57	1.66

Table 4 Enhancement factors for Kco2, Forest with senescent leaves

Simpson et al. 1998

height	Unstable	Near neutral	Stable
1.9-2.2h	0.90	1.38	1.27
1.6-1.9h	0.84	1.52	1.49
1.4-1.6h	0.93	1.35	1.47

1.2-1.4h	1.20	1.88	1.92

They conclude that the roughness sublayer extended to about 2 times canopy height.

Cellier and Brunet (Cellier and Brunet, 1992) evaluated data over a corn canopy. They found that the depth of the roughness sublayer is a function of the heterogeneities. They scaled the enhancement factor for heat and vapor as:

Equation 30

$$\gamma_e = \frac{K_{obs}}{K_{mo}} = \frac{z_*}{z}$$

They report that the normalized length scale, z*/δ□□ ranges between 3 and 4, where δ is the distance between plants.

Below, we use the conservation equation for a scalar flux covariance, $\overline{w'c'}$, to arrive at a theoretical understanding why K theory fails in the roughness sublayer.

Equation 31

$$\frac{\partial \overline{w'c'}}{\partial t} = 0 = -\overline{w'^2}\frac{\partial \bar{c}}{\partial z} - \frac{\partial \overline{w'w'c'}}{\partial z} - \overline{c'\frac{\partial p'}{\partial z}} + g\frac{\overline{\theta'c'}}{\theta}$$

Terms on the rhs are due to shear production, transport, pressure production and buoyancy production.

The steady state balance in a scalar covariance budget is a function of shear production, turbulent transport, pressure production and buoyancy production. A common parameterization of the pressure term is as a function of the flux covariance and a time scale:

Equation 32

$$\overline{c'\frac{\partial p'}{\partial z}} = \frac{\overline{w'c'}}{\tau}$$

On re-substitution (and assuming near neutral thermal stratification) we have a new equation for the flux covariance:

Equation 33

$$\overline{w'c'} = \frac{-\overline{w'^2}}{\tau}\frac{\partial \bar{c}}{\partial z} - \frac{\partial \overline{w'w'c'}}{\tau \partial z}$$

The first term on the RHS is equivalent to K dc/dz. The second term is a transport term. In essence, conventional flux gradient theory fails when the turbulent transport term is non-zero. Our data shows this to be true in the layer up to about 1.5h. High above vegetation canopies the turbulent transport term is nill (Finnigan, 2000; Paw U and Meyers, 1989)

As a closing note, I want to stress that one should never attempt to apply Flux-Gradient theory by placing one instrument in the canopy and another above it. I have seen this done by colleagues with faint acquaintance of micrometeorology and it is a violation of the concepts discussed so far. K theory is not valid in the mixed layer either.

'Direct' Methods

The direct method uses a direct measure of a flux density and its vertical gradient to develop an eddy exchange coefficient for another scalar. Classic examples employ the use of lysimeters to measure water vapor fluxes and the eddy covariance method to measure sensible heat transfer (e.g. the modified Bowen ratio, (Meyers et al., 1996). In principle, flux densities are computed as a function of the flux density of a known quantity, x, and the concentration gradients for c and x:

Equation 34

$$F_c = F_x \frac{\Delta c}{\Delta x}$$

It is important to stress, NEVER evaluate this equation measuring concentrations differences above and within vegetation. Concentration gradients should only be measured in the constant flux layer and preferably above the roughness sublayer.

Sampling protocol.

When applying the flux-gradient method, one needs a gradient large enough to resolve. This is a challenge because vigorous turbulent mixing causes concentration gradients to be small. Typical values for precision required to measure meaningful concentration gradients is 0.5%!

$$\frac{\Delta C}{C} = -\frac{F}{ku_* C}\ln(\frac{z_2}{z_1}) + \psi(z_1 / L) + \psi(z_2 / L)$$

The ratio, F/C, is defined as the deposition velocity. With perfectly responding sensors, the best one can usually measure this value is on the order of 0.01 cm s^{-1}. So one needs to resolve relative concentration gradients to a level better than +/- 0.3% to apply the method across a range of conditions. A summary table is listed below.

Table 5 ***The measurement requirements for assessing relative concentration gradients as a function of turbulent mixing the the underlying flux density. Values of □C/C for near neutral conditions (after Wesely et al., 1989). (Wesely et al., 1989)These computations are for near neutral stratification.***

u*	**V_d = 0.1 cm/s**	**V_d= 0.5 cm/s**	**V_d= 1.0 cm/s**
0.1	1.7%	8.7%	17.3%
0.2	0.9	4.3	8.7
0.3	0.6	2.9	5.8
0.6	0.3	1.4	2.9

For more general application one can use:

$\frac{\Delta C}{C} = 56V_d / u_*$, for unstable thermal stratification

$$\frac{\Delta C}{C} = 173 V_d / u_*$$, near neutral stratification

$$\frac{\Delta C}{C} = 611 V_d / u_*$$, stable stratification

How does one sample gradients to achieve the high level of precision that is required to apply flux-gradient methods? We have two choices. Either use two instruments, placing one at each level, or use a single instrument and switch its sample port back and forth. If we use separate instruments we can measure two height simultaneously and continuously. But there is great potential for bias errors due to drifting of the instrument zero and its calibration. The effect is magnified if the sensor is non-linear or suffers from temperature or humidity effects. To apply this method one must rezero and intercompare the instruments often. In the fields of evaporation, scientists used temperature/psychrometer systems that systematically and periodically reversed or were brought to a common level.

Another approach is to use on sensor and switch between heights at distinct intervals. But if this method is used then one does not measure each level simultaneously. There is potential for sampling error if one level is systematically measured during turbulent sweeps, when the air is well mixed, and the other level is measured during quiescent periods, when concentration differences are magnified.

How long should we sample each level and how often should we reverse the measurements? Woodruf (1986) report that the sampling error is proportional to the cycling time, T_c and the turbulent time scale, τ:

Equation 35

$$\varepsilon = 6\left(\frac{T_c}{\tau}\right)^{0.8}$$

If we sample for 30 s at each level, then T_c is 60 s, for example. The turbulent time scale can be assessed as the inverse of the frequency associated with the peak of the scalar power spectrum. Typical time scales are on the order of 100 to 200 seconds. A more rapid switching time reduces the sampling error. In some circumstances, this may not be practical if one is measuring a profile at multiple points. If we measure 5 levels and sample 30 seconds at each level it takes 2.5 minutes to complete a cycle. We also have to consider passing air through the cell of the sensor, in some circumstances.

Error estimates as a function of the sampling recycling time and the turbulence time scale.

ε□□□□	Tc	τ
7	60	50
4	60	100
3	60	200
44	600	50
25	600	100
14	600	200
105	1800	50
60	1800	100
35	1800	200

Some chemical sensors require tens of minutes to make a single measurement, as in the case of hydrogen peroxide. An approach used by Hall et al, is to fit a regression through the intermittent time series and take the differences from this means. Otherwise one has a sampling bias on the order of 40-60%, as in the case when sampling period is 30 minutes

References

Baldocchi, D.D., Verma, S.B. and Rosenberg, N.J., 1981. Mass and energy exchange of a soybean canopy under various environmental regimes. Agronomy Journal., 73:: 706-710.

Biscoe, P.V. et al., 1975. Barley and Its Environment .1. Theory and Practice. Journal of Applied Ecology, 12(1): 227-&.

Businger, J.A., 1971. Flux-profile relationships in the atmospheric surface layer. Journal of Atmospheric Science, 28: 181.

Carpenter, S.R., Chisholm, S.W., Krebs, C.J., Schindler, D.W. and Wright, R.F., 1995. Ecosystem Experiments. Science, 269(5222): 324-327.

Cellier, P. and Brunet, Y., 1992. Flux Gradient Relationships above Tall Plant Canopies. Agricultural and Forest Meteorology, 58(1-2): 93-117.
Denmead, O.T., 1969. Comparative micrometeorology of a wheat field and a forest of Pinus radiata. Agricultural Meteorology, 6(5): 357-371.
Dyer, A.J. and Hicks, B.B., 1970. Flux-Gradient Relationships in Constant Flux Layer. Quarterly Journal of the Royal Meteorological Society, 96(410): 715-&.
Finnigan, J., 2000. Turbulence in Plant Canopies. Annu. Rev. Fluid Mech., 32(1): 519-571.
Foken, T., 2006. 50 years of the Monin-Obukhov similarity theory. Boundary-Layer Meteorology, 119(3): 431-447.
Glotfelty, D.E., Taylor, A.W. and Zoller, W.H., 1983. ATMOSPHERIC DISPERSION OF VAPORS - ARE MOLECULAR-PROPERTIES UNIMPORTANT. Science, 219(4586): 843-845.
Hicks, B.B., Rao, K.S., Durham, J.L., Gifford, F.A. and Lenschow, D.H., 1984. ATMOSPHERIC DISPERSION OF PESTICIDE VAPORS - ANALYTICAL METHODS QUESTIONED. Science, 224(4651): 905-905.
Hogstrom, U., 1988. Non-dimensional wind and temperature profiles in the atmospheric surface layer-a re-evaluation. Boundary Layer Meteorology, 42: 55-78.
Hogstrom, U., 1996. Review of some basic characteristics of the atmospheric surface layer. Boundary-Layer Meteorology, 78(3-4): 215-246.
Hsieh, C.I., Katul, G. and Chi, T., 2000. An approximate analytical model for footprint estimation of scaler fluxes in thermally stratified atmospheric flows. Advances in Water Resources, 23(7): 765-772.
Hurlbert, S.H., 1984. Pseudoreplication and the Design of Ecological Field Experiments. Ecological Monographs, 54(2): 187-211.
Kaimal, J.C. and Wyngaard, J.C., 1990. The Kansas and Minnesota Experiments. Boundary-Layer Meteorology, 50(1-4): 31-47.
Lang, A.R.G., Mcnaughton, K.G., Fazu, C., Bradley, E.F. and Ohtaki, E., 1983. Inequality of Eddy Transfer-Coefficients for Vertical Transport of Sensible and Latent Heats During Advective Inversions. Boundary-Layer Meteorology, 25(1): 25-41.
Mahrt, L., 1999. Stratified atmospheric boundary layers. Boundary Layer Meteorology, 90: 375-396.
Meyers, T.P., Hall, M.E., Lindberg, S.E. and Kim, K., 1996. Use of the modified Bowen-ratio technique to measure fluxes of trace gases. Atmospheric Environment, 30(19): 3321-3329.
Moncrieff, J., Malhi , Y. and Leuning, R., 1996. The propogation of errors in long-term measurements of carbon and water. Global Change Biology, 2: 231-240.
Motha, R.P., Verma, S.B. and Rosenberg, N.J., 1976. Turbulent Transport under Advective Conditions. Bulletin of the American Meteorological Society, 57(11): 1406-1406.
Oksanen, L., 2001. Logic of experiments in ecology: is pseudoreplication a pseudoissue? Oikos, 94(1): 27-38.
Panofsky, H. and Dutton, J.A., 1984. Atmospheric Turbulence: Models and Methods for Engineers. John Wiley, New York.
Paw U, K.T. and Meyers, T.P., 1989. Investigations with a higher-order canopy turbulence model into mean source-sink levels and bulk canopy resistances. Agricultural and Forest Meteorology, 47(2-4): 259-271.
Pruitt, W.O., Morgan, D.L. and Lourence, F.J., 1973. Momentum and Mass Transfers in Surface Boundary-Layer. Quarterly Journal of the Royal Meteorological Society, 99(420): 370-386.
Rao, K.S., Wyngaard, J.C. and Cote, O.R., 1974. Structure of 2-Dimensional Internal Boundary-Layer over a Sudden Change of Surface-Roughness. Journal of the Atmospheric Sciences, 31(3): 738-746.
Raupach, M.R., 1979. Anomalies in Flux-Gradient Relationships over Forest. Boundary-Layer Meteorology, 16(4): 467-486.
Schank, J.C. and Koehnle, T.J., 2009. Pseudoreplication is a pseudoproblem. Journal of Comparative Psychology, 123(4): 421-33.

Schmid, H.P., 2002. Footprint modeling for vegetation atmosphere exchange studies: a review and perspective. Agricultural and Forest Meteorology, 113(1-4): 159-183.

Sellers, P.J. and Hall, F.G., 1992. Fife in 1992 - Results, Scientific Gains, and Future-Research Directions. Journal of Geophysical Research-Atmospheres, 97(D17): 19091-19109.

Simpson, I.J., Thurtell, G.W., Neumann, H.H., Den Hartog, G. and Edwards, G.C., 1998. The validity of similarity theory in the roughness sublayer above forests. Boundary-Layer Meteorology, 87(1): 69-99.

Sinclair, T.R., Allen, L.H. and Lemon, E.R., 1975. An analysis of errors in the calculations of energy flux densities above vegetation by a Bowen ratio profile method. Boundary Layer Meteorology, 8: 129-139.

Verma, S.B. and Rosenberg, N.J., 1975. Accuracy of Lysimetric, Energy-Balance, and Stability-Corrected Aerodynamic Methods of Estimating above-Canopy Flux of Co2. Agronomy Journal, 67(5): 699-704.

Webb, E.K., 1970. PROFILE RELATIONSHIPS . LOG-LINEAR RANGE, AND EXTENSION TO STRONG STABILITY. Quarterly Journal of the Royal Meteorological Society, 96(407): 67-&.

Wesely, M., Lenschow, D.H. and Denmead, O.T., 1989. Flux Measurement Techniques. In: D. Lenschow and B. Hicks (Editors), Global Tropospheric Chemistry: Chemical Fluxes in the Global Atmosphere. NCAR, Boulder, CO, pp. 31-46.

Wyngaard, J.C., 1992. Atmospheric-Turbulence. Annual Review of Fluid Mechanics, 24: 205-233.