Chapter 5

LOW FREQUENCY ATMOSPHERIC TRANSPORT AND SURFACE FLUX MEASUREMENTS

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Abstract

We review the issue of turbulent atmospheric transport of scalars or momentum on timescales greater than 30 minutes or 1 hour, regions of the spectrum of atmospheric motion that are not usually sampled by conventional flux measurement methodologies. We first explore what is known about the nature and timescales of turbulent transport structures in the near-surface layers of the atmosphere, and the degree to which this transport is controlled or modulated by the timescales of inner layer (shear) transport and outer boundary layer transport. We then present empirical evidence of the existence of low frequency transport by presenting data from two contrasting field studies, a shear-dominated measurement set-up in Scotland, and a convection dominated measurement set-up in Brazilian Amazonia. It is clear that low frequency transport can transport a significant amount of flux in measurement situations such as towers over forests in anticyclonic conditions, or in the tropics. Thereafter we explore the quantitative implications of undersampling low frequency atmospheric transport. Extending the sampling period of surface flux measurements is desirable under certain conditions, but is not without complications. We conclude by highlighting some of the dangers of extending sampling periods into the low frequency domain.

1 Introduction

The micrometeorological technique of eddy covariance aims to measure the transport of flux via turbulence between the surface and the atmosphere. In practice it samples only a part of the possible spectrum of atmospheric motion, typically time scales between one second and one hour. Transport at timescales less than one second are discussed elsewhere in this volume (Chapter 4). For longer timescales an assumption is made of frequency separation — that a spectral gap exists between the operational timescales of flux transport, and longer-scale atmospheric motions. The presumed existence of such a gap is critical for measured turbulent fluxes, as it provides an upper time limit to measurements averaging and rotation periods to separate "locally meaningful" fluxes from background trends and oscillations. However, it is unclear as to whether this gap really exists. The importance of low frequency motions to energy and carbon balance studies have been highlighted in recent papers (Mahrt 1998a, 1998b, Sakai et al. 2001, Von Randow et al. 2002, Finnigan et al. 2003).

In this Chapter, we first review what is known about the nature and timescales of turbulent transport structures in the near-surface layers of the atmosphere, and the degree to which this transport is controlled by the timescales of inner layer (shear) transport and boundary layer transport. This chapter covers some of the same territory as Mahrt (1998a), but using the perspective of the new turbulence model of McNaughton (2004b). We focus primarily here on daytime turbulent transport, the deeper complexities of nighttime transport are discussed in Chapters 10 and 8 of this book. We then present empirical evidence of the existence of low frequency transport by presenting data from two contrasting field studies, a shear dominated measurement set-up in Scotland, and a convection dominated measurement set-up in Brazilian Amazonia. Thereafter we explore the quantitative implications of undersampling low frequency atmospheric transport, summarizing an analysis presented by Finnigan et al (2003). We conclude by highlighting some of the dangers complications of extending sampling periods into the low frequency domain. The subject of low frequency transport and how to deal with it enters into the still poorly understood fields of self-organized turbulent structures and mesoscale flows, and is clearly an area of ongoing research. We begin by reviewing these issues.

2 Turbulence structure, eddy sizes and sampling times

The question of averaging times for flux measurements clearly depends on the nature, structure and time scales of the fluctuations over which the average is taken. It is possible to approach this purely through a discussion of the statistical character of the signals to be averaged, notably through discussion of their spectra and cospectra. This section is a preliminary to that discussion, providing some insight into the turbulence processes that underlie these spectra. Since the structure of turbulence varies with height we divide the boundary layer into a series

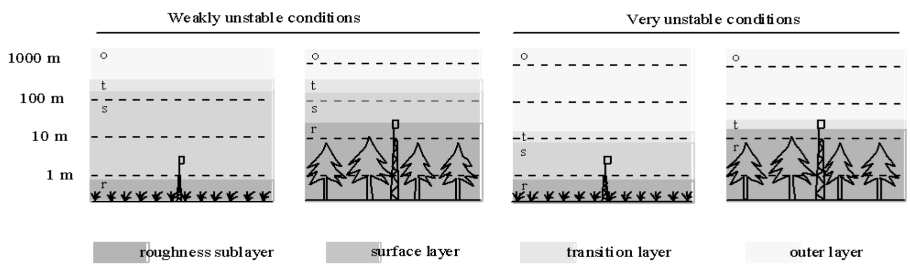

Figure 5.1. Schematic diagram of the location of flux measuring instruments, shown as rectangles on the top of each tower, in relationship to the various layers of the boundary layer in both weakly and strongly unstable conditions. Instruments mounted a few meters above low vegetation are usually within the surface layer. This layer may completely disappear over forests as the outer layer extends downwards into the layer directly influenced by the vegetation (up to at least twice forest height). Measurements from a forest tower may often be in the transition or outer layer during the day. This affects the averaging times needed to make reliable observations of scalar fluxes.

of layers, the principle ones being the 'outer layer' or 'mixed layer' where large convective circulations form during the day, and the 'inner layer' or 'surface layer' whose characteristics are dominated by shear. Within the inner layer is a roughness sublayer where the turbulence is directly affected by the surface itself. The inner and outer layers are separated by a transition layer. This has variously been called an 'interaction layer' (McNaughton 2004b), a 'matching layer' (Mahrt 1998a) or a 'free convection layer' (e.g. Garratt 1992). These different names reflect different understandings of what this layer represents. Garratt (1992) sees it as a layer of gradual transition where proposed plume-like structures that develop progressively with height in the surface layer finally break free of shear effects while not yet being constrained by the top of the boundary layer. Mahrt (1998a) takes a more cautious approach and simply notes that layered models typically require that properties in the inner and outer layer should be mathematically matched at this height. McNaughton (2004b) describes it as a layer where two fundamentally different kinds of turbulence interact, so turbulence there is fundamentally less ordered than in the layers above and below. McNaughton's ideas are based on the few direct observations available, and we base our discussion on it.

Figure 5.1 shows four of the identifiable structural layers of a convective boundary layer. Above the boundary layer lies the free atmosphere, with an entrainment layer (another interaction layer) between this and the outer layer. These are not shown because flux measurements are not made at these levels. Figure 5.1 also shows instruments placed on towers above short and tall vegetation in relationship to these structural layers of the boundary layer. The transition layer typically begins at one or two times the Obukhov length, $|L|$, above the ground, where the Obukhov Length is given by

$$L = -\frac{\Theta_v u_*^3}{kg(\overline{w'\theta_v'})_0} \tag{5.1}$$

Here g is acceleration due to gravity, $(\overline{w'\theta_v'})_0$ is the buoyancy flux at ground level, Θ_v is the virtual potential temperature in degrees Kelvin, u_* is the friction velocity, and k is the von Krmn constant. This transition layer height may vary from less than 10m to more than 100 m depending on meteorological conditions, so instruments on a short mast over grassland will usually be in the surface layer while instruments on a tall tower over forest will be variously within the surface layer, the transition layer or even the outer layer, depending on meteorological conditions. This creates some challenging problems for flux measurement over forests.

2.1 Mixed-layer (outer-layer) timescales

Over uniform land the spatial and temporal patterns of eddy motion are not imposed by the boundary conditions, upper or lower, but arise spontaneously through the self-organizing nature of the turbulent motions themselves. During the day this self-organized pattern is usually a set of more-or-less polygonal convective cells that span the whole boundary layer. These cells are about 1.5 z_i wide, where z_i is the height of the capping inversion. The height z_i is typically 1-3 km during the middle of the day over land, being greatest in high heat flux conditions such as continental interiors in summer. These convective cells move along with the mean wind in the outer layer. Their shapes depend on the value of the ratio of the outer and inner velocity scales, w_*/u_* , where the convective velocity scale w_* is related to the standard deviation of the wind velocity in the outer layer and is usually estimated using Deardorff's relationship

$$w_*^3 = \frac{z_i g(\overline{w'\theta_v'})_0}{\Theta_v} \tag{5.2}$$

The ratio w_*/u_* is then related to the Obukhov length through the tautological relationship

$$\frac{w_*}{u_*} = \left(-\frac{z_i}{kL}\right)^{1/3} \tag{5.3}$$

When $z_i/|L|$ is larger than about 25, the cells form a polygonal pattern with no notable alignment. At smaller values of $z_i/|L|$, when drag on the ground is more important, the convective cells become elongated in the direction of the wind and aligned, one with the next, so that they form elongated roll structures aligned with the wind. Water vapor condensing in the updrafts of these structures can form cloud streets (Etling and Brown 1993). These streets, whether made visible by clouds or not, are quasi-permanent in position and many kilometers long. Fixed sensors in the outer layer may then record steady updrafts ($\overline{w} > 0$) or steady downdrafts ($\overline{w} < 0$) over long periods, with obvious consequences for the calculation of fluxes. Very often the decrease in $z_i/|L|$ reflects an increases in $|L|$, so the surface layer grows in thickness to envelope fixed-height sensors above a forest. At $z_i/|L|$ ratios less than about 5 the boundary layer becomes near-neutral, exhibiting no large-scale convective roll structures. We might observe that the combination of deeper surface and transition layers leaves no room for outer convective structures to develop in such situations.

An important characteristic of the main convective motions in the outer layer is that they carry most of the flux, with lesser amounts carried by the smaller eddies created by internal friction and breakdown of these larger eddies. An aircraft equipped with flux sensors must fly through many tens of large eddies to sample the flux properly, so an adequate flight path would be of order 50 km long, more or less depending on prevailing conditions and the accuracy required. Flux sensors mounted on a fixed tower or suspended from a tethered balloon must sample a similar number of eddies by waiting as they blow past the instruments, so the sample period should be many tens of times z_i/U_m, where U_m is the mean wind speed in the outer part of the boundary layer. In light winds, as often encountered in the tropics or in mid-latitude anticyclones, the convective structures may pass very slowly so that a good sample is obtained only for times long compared with the evolutionary time scale of the convection cells. Values for this evolutionary time scale have not been reported, but they may depend strongly on any inhomogeneities in roughness, surface heat flux or topography that might cause convective cells to lock into fixed positions on the landscape. Even in favorable conditions it usually takes rather more than an hour to achieve a good average in convective conditions, and much longer if the convective cells are aligned with the wind in quasi-permanent rows or locked onto the landscape. A non-zero mean for vertical velocity may persist for hours. Kanda et al. (2004) use Large Eddy Simulation (LES) models to calculate errors likely to be found in instrumental measurements from very tall towers over uniform ground.

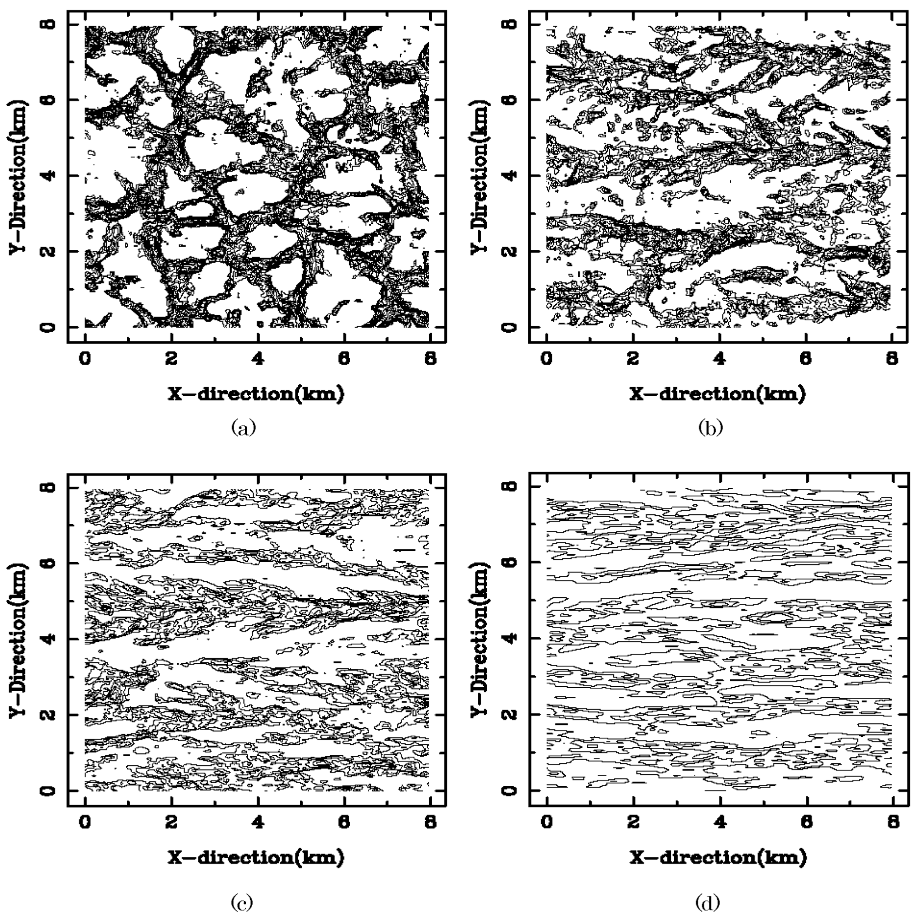

Figure 5.2. Simulations of mean vertical velocity at 100 m height in a typical mid-latitude daytime boundary layer over a flat surface, averaged over 1 hr for different geostrophic wind speeds U_g. Contours (at 0.1 m s^{-1} interval) represent positive (upwards) velocity regions; negative velocity regions are blank. (a) $U_g = 0$ m s^{-1}; (b) $U_g = 1$ m s^{-1}; (c) $U_g = 2$ m s^{-1}; (d) $U_g = 4$ m s^{-1}. Reproduced with permision of Kanda et al. (2004).

They does not give values of w_*/u_* for their study, but the small values of the geostrophic wind used and the cellular patterns found in their results suggests that conditions are highly convective in all cases. At low geostrophic wind speeds transport in the lower convective boundary layer self-organizes into turbulent organized structures (TOS) consisting of large areas of slow subsidence and smaller areas of updraft (Figure 5.2). As a result, a single point measurement averaged/rotated over a short time period is likely to be biased to the downdraft regions and

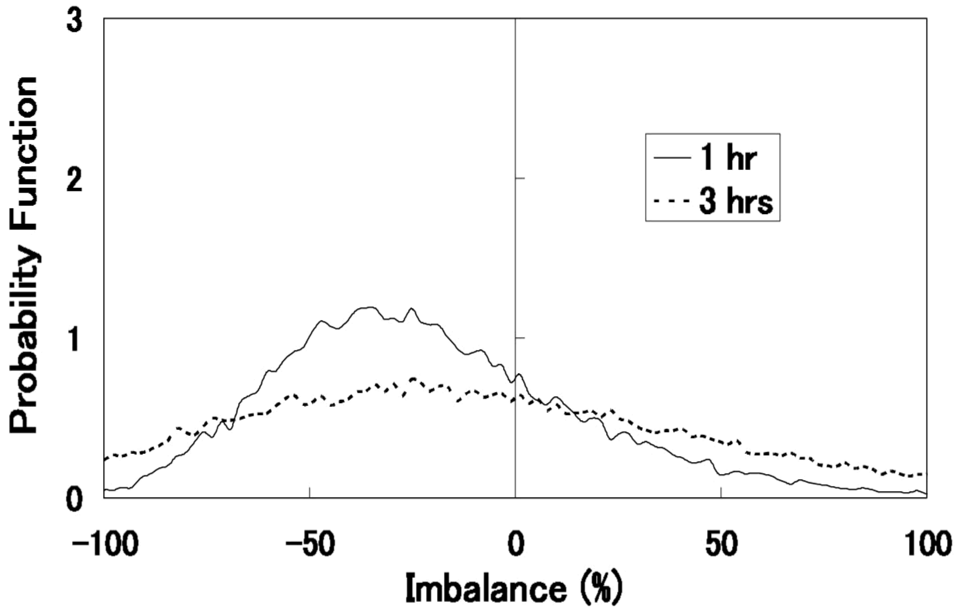

Figure 5.3. The probability distribution of the "imbalance" of fluxes measured at any point in the simulation are shown in Figure **??**a, for geostrophic wind speed $U_g = 0\ \mathrm{m\,s^{-1}}$, height $z = 100$ m. The two curves represent averaging periods of 1 and 3 hr. Measurements at any point are biased to underestimate surface fluxes. Increasing the average period reduces the systematic error in the "imbalance", but broadens the distribution and hence the random uncertainty in any single measurement. Reproduced with permision of Kanda et al. (2004).

hence underestimate the local fluxes. Simulated observations at 100 m above ground display vertical winds that do not average to zero over one-hour periods, with the net upwards or downwards flow typically accounting for half the flux when $u_* = 0.14$ m s^{-1}. This "imbalance" is reduced with (i) increasing wind speed which decreases the cross-wind diameter of the TOS and increases the sampling track in any averaging period (e. g. at $u_* = 0.3$ m s^{-1}, mean flow accounts for about 10% of vertical scalar transport on average over the simulation domain, but with values for particular points having a standard deviation of about 20% about that mean); (ii) lower measurement height increasing the influence of shear turbulence. Hence measurement points located above tall forests, and in particular in low-wind regions such as many tropical and continental interior sites, are particularly likely to be prone to such a measurement "imbalance". In these conditions flux measurements from single point measurements are inherently distorted (Figure 5.3).

2.2 Surface-layer (inner-layer) timescales

Near the ground the flux-carrying eddies have a rather different character, though what that character is has become the subject of debate. Here we base our discussion on the new model of McNaughton (2004b) in which the turbulence consists of large-scale wedge-like structures that are aligned with the wind, along with the detached breakdown products of these. The turbulence is again self-organizing, just like the cellular structures in the outer layer, but here the preferred structures are upscale cascades of TEAL (Theodorsen ejection amplifier-like) structures, which 'compete' for space so that only the best formed and most powerful continue on at each scale. This kind of turbulence is driven by the shear, so its velocity scale is u_*. Though the value of u_* changes with stability it seems that the structure of the shear turbulence does not. This property is again like that of the turbulent outer layer where the velocity scale of the convective cells varies with stability while the polygonal structure and the length scale of turbulence does not (at least while w_*/u_* and z_i are held constant). This model is consistent with-indeed it is based on-the spectral observations from Kansas (Kaimal et al. 1972). It contradicts the basic hypothesis of Monin-Obukhov similarity on which much of our theory of the surface layer is built. Time will judge whether the new model is successful, but for the moment it simplifies our discussion by removing the need to consider the effects of stability acting within the surface layer, at least when we know u_* directly. Turbulence in the surface layer scales simply on z and u_*.

To get a good sample of the dominant flux-carrying eddies, with widths about $2z$, we must allow a large number of them to pass our fixed measuring position. A simple estimate is that this should be a hundred times $z/\overline{u}$, which is a very few minutes in most cases. Experience tells us that this is a considerable underestimate and there are several reasons for this. One is that the flux-carrying coherent structures are not randomly dispersed, so our sample must take account of the scale of the aggregates (i. e. whole ramp structures) which are about ten times longer than our estimate of about $2z$ (McNaughton 2004a). Another reason is that flux transport is affected by eddies of a greater range of sizes here than in the outer layer. In the outer layer there are no eddies much wider than $1.5z_i$, while in the surface there are eddies of all sizes, right up to those as tall as the surface layer itself, and all of these transport at least some momentum and scalars. Our averaging period must be long enough to sample not just the dominant eddies but all the significant flux-carrying eddies. As suitable averaging period must be at least several tens of minutes long, perhaps even an hour long for good results.

Another factor is the modulating effects of the outer convection on the turbulence in the surface layer. The outer-layer convection constitutes a variable driving of the whole surface layer. A large-scale gust from the outer layer is equally a large cohort of TEAL structures moving along with enhanced speed and transmitting an enhanced momentum flux towards the ground. The direction, speed and power of the TEAL cascades within the surface layer will therefore all vary with the wind at the top of the surface layer, as will the momentum flux to the ground. Observations of momentum flux from a fixed tower will not represent area means unless sampled on an *outer time scale*, not the inner one. We expect poor agreement in hourly observations of u_* from towers spaced a few hundred meters apart in homogeneous terrain. The situation is easier for the scalar fluxes, most of which are more strongly controlled at source rather than by the varying wind overhead. For example, photosynthesis is controlled by radiation receipt and biochemical and physiological processes within the canopy, both of which are insensitive to wind speed. The flux of water vapor is somewhat sensitive to wind speed, depending on the value of the decoupling parameter Ω calculated with values at reference height z_s (Jarvis and McNaughton 1986), but only for wet canopies is it important. Averaging times for source-limited scalar fluxes depend little on outer time scales. This difference accounts for the reported different averaging times required for momentum and scalars in the Kansas experiment (Wyngaard 1973). Wyngaard shows the variance of heat flux measurements from one-hour runs to be about 8%, with only small dependence on instability ($-z/L$), while the sensitivity of momentum flux ranged from 10% to about 80% with a strong dependence on instability. Wyngaard used $-z/L$ to measure instability while our discussion suggests that $-z_i/L$ should be used.

The layered structure discussed above is not relevant when the outer turbulence dies away at night. In the first part of the night we have, typically, a fully-turbulent state where the turbulence is similar that in the daytime surface layer, but buoyancy now opposes these motions and saps their energy. Momentum transfer then decreases progressively this kind of structure collapses altogether unless the flow is maintained by strong pressure gradients or katabatic forcing. If not then a variety of other phenomena may appear, some creating intermittency in the turbulence with time scales up to an hour or more. These processes are not well described and are discussed in Chapters 9 and 8 of this book. For the weakly stable case flux sampling times can be a little shorter than in the daytime surface layer.

The roughness sublayer is a subdivision of the surface layer and many of the above comments apply equally to it. Near canopy top the turbulence has many of the characteristics of a mixing layer (Raupach et al. 1996), and the main eddies scale on the canopy dimensions $(h - d)$, where h is tree height and d is the displacement height of the forest wind profile. This structure is not well known, but it is known that spectra and cospectra have more peaked shapes than the standard Kansas forms. Following the methods of McNaughton (2004a) we can therefore surmise that tendency of these eddies to align into long wedge-like structures is less pronounced than in the surface layer. Good time samples should be easier to obtain in the surface layer, but the length scale used to calculate these is $(h - d)$ rather than $(z - d)$.

A theme running through the above comments is that eddy flux calculations made on short data runs may not fairly represent true long-term averages. This is so whether the flux itself varies on a rather long time scale, or whether the statistical sampling of the flux-carrying eddies is insufficient. In the former case the actual local flux is a poor sample of the required area average flux; this cannot be remedied by any analysis based solely on the data from that interval. In the latter case the local flux during the measurement interval is not in question, but some of the larger eddies carrying it have been poorly sampled.

3 Empirical evidence of low frequency flux transport

3.1 Wavelet spectral analyses of turbulent fluxes in Scotland and Amazonia

To demonstrate the variable contribution of processes occurring on different scales to the surface layer fluxes, in this study we apply a Haar wavelet transform on the turbulent signals measured at two different sites: a rain forest in south west Amazonia (Rebio Jaru; Von Randow et al. 2002) and a sitka spruce coniferous forest in Scotland (Griffin; Chapter 4). The wavelet transform (WT) is a powerful mathematical analysis tool, which permits an evolutionary spectral study of turbulent atmospheric signals (Daubechies 1998; Farge 1992). The wavelet analyses were done following a similar methodology as Katul and Parlange (1994) and Von Randow et al. (2002).

After application of the Haar wavelet to the data from the two sites, the scale covariances of vertical wind velocity and scalars were calculated and the partial contribution of each scale to the total covariance was determined at each record. The results are presented in Figure 5.4 (daytime) and Figure 5.5 (nighttime). The x-axes represent the spatial

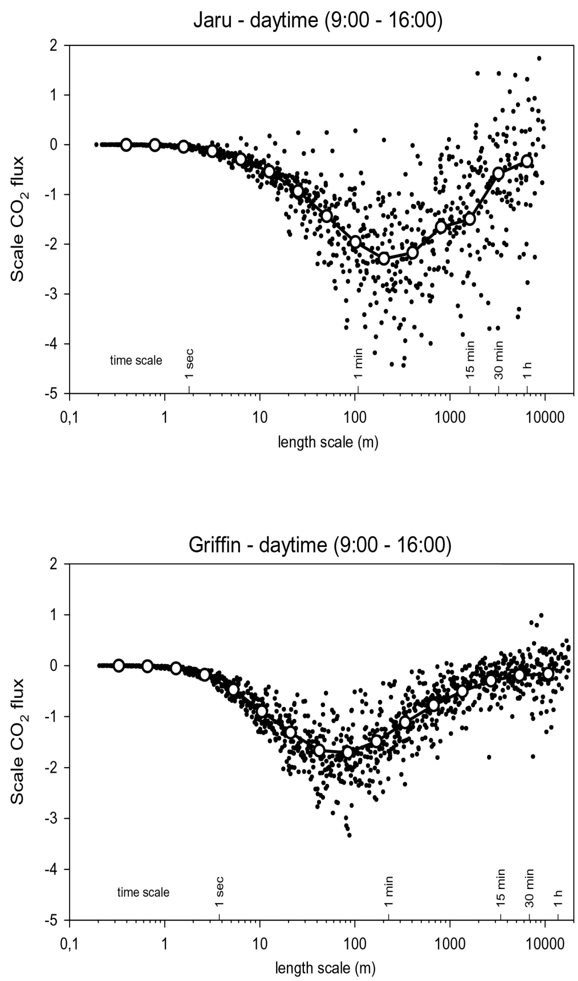

Figure 5.4. Haar wavelet cospectra of daytime CO_2 fluxes at (a) a tropical site (Jaru, Brazil) and (b) a maritime mid-latitude site (Griffin, Scotland). The solid line represents binned averages; length and time scales are indicated on the x-axis.

scales, which are estimated using the average wind velocities and the assumption of Taylor's Hypothesis, similar to the method applied by Von Randow et al. (2002). Above the x-axes, approximate time labels are also included to illustrate the time scales of the processes.

Comparing the average daytime results from Jaru and Griffin (Figure 5.4), it is apparent that the contribution from the largest scales (lower frequencies) are more important at Jaru than at Griffin. At Re-

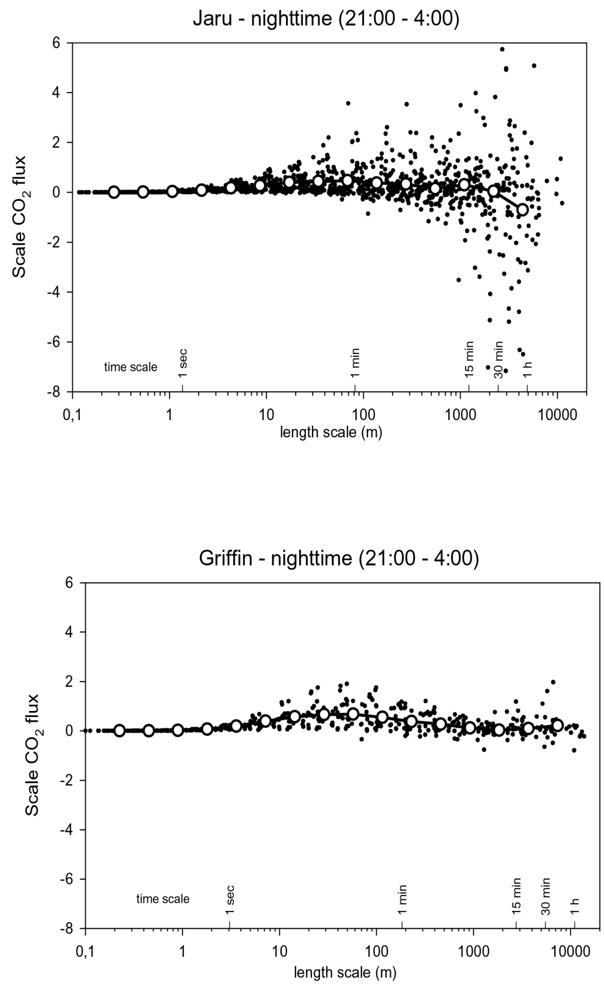

Figure 5.5. Haar wavelet cospectra of nighttime CO_2 fluxes at (a) a tropical site (Jaru, Brazil) and (b) a maritime mid-latitude site (Griffin, Scotland). The solid line represents binned averages; length and time scales are indicated on the x-axis.

bio Jaru, the peak of contribution to the covariances happens on scales of 100-500 m (that correspond approximately to scales of 1 to 5 min), while at Griffin scales less that 100 m (usually corresponding to scales of less than 1 min) dominate the transport of carbon dioxide. One other noticeable difference between the two sites is that at Jaru the variation from the average scale dependence is higher, especially at longer scales (low frequencies). On scales longer than 1 km, the low frequency mo-

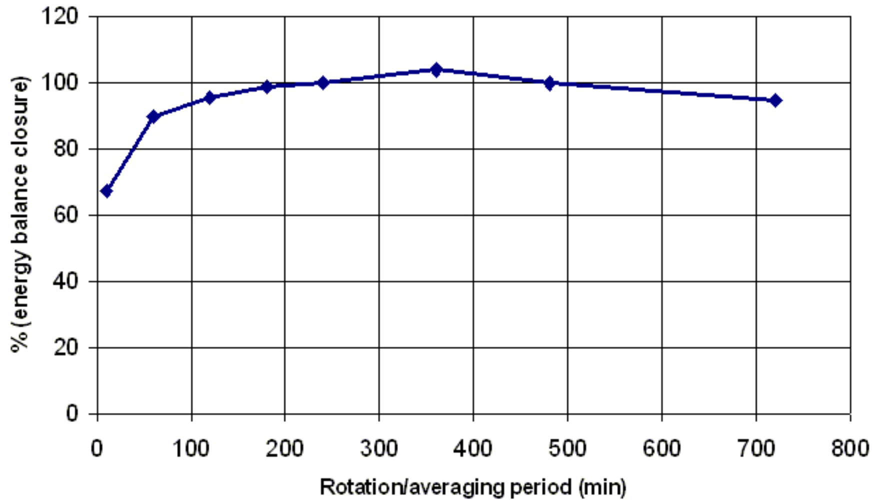

Figure 5.6. The relationship between long-term energy balance closure and the averaging/rotation period of the flux calculations, for a site over a tropical rainforest near Manaus, Brazil. Derived from Malhi et al. (2002).

tions influences can be of either sign (Figure 5.2a), clearly not related to processes of the surface layer only. These low frequency processes can include deep convection, large roll vortices and local circulations induced by topography or surface heterogeneity. At the Brazilian site, a modest but significant about of turbulent transport occurs at time scales beyond 30 minutes, or even beyond one hour.

During stable conditions (nighttime) there is no clear timescale for flux transport at the Brazilian site (Figure 5.5 top panel), and a great degree of variance in the spectra. Boundary-layer and mesoscale flows clearly dominate the nighttime turbulent flux. In contrast, the spectra are fairly consistent at the relatively windy Scottish site (Figure 5.5 bottom panel).

3.2 Low frequency transport and energy balance

The lack of energy balance closure in eddy covariance studies is a widespread feature of flux measurements over forests (Wilson et al. 2002), suggesting the presence of a widespread problem, whether instrumental or methodological. One possible explanation is that latent and sensible heat flux is being missed by the flux measurements, through either inadequate spatial sampling or inadequate sampling of the frequency domain.

Malhi et al. (2002) explored this phenomenon for flux measurements above a forest near Manaus in central Amazonia (Figure 5.6). They found that extending the rotation/averaging period of the measurements from 1 hr up to 4 hr improved the energy balance closure to about 100%; increasing the rotation/averaging period further resulting in no further increase in mean fluxes, although variance increased substantially. This result suggests that, for some sites at least, low frequency transport may "solve" the energy balance problem, for ensemble-averaged data at least. The general applicability of this approach remains uncertain, however, as Kruijt (pers comm.) and Malhi and Iwata (unpublished data) find that the low frequency component at other sites is not sufficient to account for all the "missing" flux.

To summarize this section, there is evidence of significant flux transport at low frequencies at some measurement sites; in particular sites in the tropics with tall measurement towers, low mean wind speeds and deep convective boundary layers. In the next section we examine how undersampling these low frequency fluxes using standard eddy covariance analysis can affect measured fluxes.

4 What effect do standard flux calculations have on low frequency flux terms?

In this section we explore the effect that the coordinate rotation used in eddy covariance analyses has on calculated fluxes. Finnigan et al. (2003) recently presented a revision of the theory of measurement of turbulent fluxes in terms of mass balance equations. First they considered an idealized case, where the long-term ensemble-averaged flow is horizontally homogeneous, the mean wind vector is always confined to the $x - z$ plane, so that $\overline{v} = 0$ and only the inclination of the velocity vector, $\alpha = \tan^{-1}(\overline{w}/\overline{u})$ changes from period to period. To transform the long-term vertical covariance $\overline{wc}_{LT}$ in any period to short-term 'rotated every period' coordinates in which $\overline{w} = 0$, coordinates are rotated through an angle α given by,

$$\alpha = \tan^{-1} \frac{\overline{w}\,'}{<\overline{u}> + \overline{u}\,'} \tag{5.4}$$

where $<>$ is the ensemble mean over all periods, $\overline{w}\,' = \overline{w} - <\overline{w}>$, and $\overline{u}\,' = \overline{u} - <\overline{u}>$.

Hence, the horizontal and vertical components of the long-term, ensemble-averaged flux vector, denoted by subscript LT, can be expressed in terms of components, denoted by subscript, R, in the 'rotated-every-period' coordinate frame as (Finnigan et al. 2003, equation 26)

$$
\begin{aligned}
(\overline{uc})_R &= (\overline{uc})_{LT} \cos\alpha - (\overline{wc})_{LT} \sin\alpha \\
(\overline{wc})_R &= (\overline{wc})_{LT} \cos\alpha + (\overline{uc})_{LT} \sin\alpha
\end{aligned}
\tag{5.5}
$$

Even in horizontally homogeneous terrain, the flow field is only horizontally homogeneous when averaged over a period much longer than that of any significant temporal perturbation to the flow. When the duration of an individual averaging period, T, is comparable to that of significant atmospheric motions, then in any one period, horizontal flux divergence may be important. Many periods must be averaged to ensure the canceling out of transient vertical advection events that in reality contribute nothing to the vertical flux because they merely balance simultaneous but unmeasured transient horizontal advection events. To ensure this averaging is done rigorously, mass balances over any time period must be calculated in a single coordinate system that applies to the entire period.

What is the precise effect of rotating the vector basis so that $\overline{w}$ is forced to zero in each period? In particular, what is the effect on the measured covariance in any one averaging period of rotating coordinates so that $\overline{v} = \overline{w} = 0$? *Dividing the signals into a set of periods of length T, block averaging in each period and then subtracting the block averaged values $\overline{w}$ and $\overline{c}$ is equivalent to high-pass filtering the signal with a boxcar filter function of width T.* This is precisely what is achieved when we rotate coordinates each period so that $\overline{w}$ is forced to zero. In effect we have thrown away the part of the covariance carried by motions of frequency lower than the (rather leaky) cut-off frequency of the boxcar filter function.

However, in addition, the coordinate rotation itself distorts the covariance carried by frequencies higher than the boxcar cut-off by folding into $(\overline{wc})_R$ some of the streamwise and lateral fluxes $(\overline{wc})_{LT}$ and $(\overline{vc})_{LT}$, as we have shown in Equation 5.5 for the case where the low frequency flow is confined to the $x - z$ plane. A simple algebraic example of the distortion to be expected is presented in Appendix 2 of Finnigan et al. (2003).

In summary, rotating coordinates every period effectively high-pass filters the signal so that contributions to the covariance from atmospheric motions of period longer than T are lost but the rotation itself folds some of the $(\overline{wc})_{LT}$ and $(\overline{vc})_{LT}$ flux into $(\overline{w'c'})_R$ in an essentially unpredictable way.

5 Complications

The energy balance closure in Figure 5.6 suggests that in some sites consideration of low frequency transport can improve the flux measurements to the point of full energy balance closure. If this were a general principle that could be applied to all sites, then it would appear that the problem of poor energy balance closure at eddy covariance sites would have been "solved". However, there are good grounds for skepticism about the wider applicability of this result.

The problem revolves around whether these low frequency fluxes are "locally meaningful" or represent features of the wider landscape that are not related to the local surface. Kanda et al. (2004) demonstrated in their LES model simulation that, although the systematic bias decreased when turbulent fluxes are averaged over longer time periods, the variance increases greatly (Figure 5.3). Hence any single measurement period is vulnerable to random advective fluxes, and becomes increasingly difficult to interpret in terms of local surface fluxes.

Another problem arises in complex terrain, or in the presence of fixed pressure fields caused by local inhomogeneities (e. g. land-water interfaces, or crop-forest mosaics). In this situation, a variation in wind direction can result in a low frequency covariance that has nothing to do with flux transport (Finnigan et al. 2003). Interestingly, Kanda et al. (2004) showed that moderate inhomogeneity (ca 5%) can actually reduce bias by dampening the self-organization of TOS, but greater degrees of inhomogeneity generate local circulations and enhance the bias.

6 Conclusions

It is important to consider transport at time scales up to at least $10z_i/U_m$, where U_m is the mean wind speed in the outer part of the boundary layer. This corresponds to length scales of about 10-50 km) but in complex topography or variable landscapes, it is tricky to disentangle what we want (low frequency CBL turbulent transport) from what we do not want (wind direction covariance). There is clearly transport of flux at these low frequencies which can explain the failure of eddy covariance systems to fully capture fluxes. However, separating the locally meaningful fluxes from wider-scale advection may prove a challenge.

7 References

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