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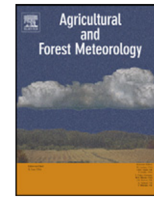

On concentration footprints for a tall tower in the presence of a nocturnal low-level jet

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ABSTRACT

This paper reports on the location of sources contributing to a point concentration measurement in the stable boundary layer. Concentration footprints for sensors located at different heights during the night are examined using a backward-in-time Lagrangian stochastic dispersion model. Simulations of air parcel transport in a non-steady-state atmospheric boundary layer above a bare surface and a forest under different insolation suggest that sources located as far as several hundred kms away contribute to a concentration measurement made at levels as high as 500 m. The origin of the maximal contribution area shifts during the night depending on the wind direction at the sensor location, a feature most prominent in the presence of decoupling between sensor levels and surface sources. Simulations suggest that atmospheric static stability alone is not a sufficient criterion to trigger flow decoupling. The presence of the low-level jet provides vertical mixing of air parcels even for stable boundary layers with Richardson numbers higher than the critical. This finding is in agreement with earlier observations suggesting that the level of the low-level jet nose acts as a strong lid prohibiting vertical gas propagation. These findings have important implications in the interpretation of eddy-flux measurements in nocturnal conditions.

1. Introduction

The estimation of the exchange rate between the biosphere and the atmosphere in nocturnal conditions presents a formidable challenge: low turbulence regime, burstiness, non-stationarity, flow meandering, thermal stratification make the interpretation of a signal obtained using standard meteorological techniques thorny at best. In cases of horizontal homogeneity and in the presence of low-turbulence regime the nocturnal boundary layer (NBL) budget technique can be used, albeit with circumspection, to estimate nighttime CO_2 fluxes (Denmead et al., 1996). The technique consists in measuring vertical CO_2 concentration profiles at different intervals throughout the night inside the stable boundary layer (SBL). Evaluating the accumulation of carbon dioxide inside the stable boundary layer throughout the night enables the experimentalist to determine the vertical flux. However, even with this technique, there are several challenges. One such challenge lies in the estimation of the SBL height itself. That supposes a well defined 'lid', or the presence of its aerodynamic equivalent, as is the case for an atmospheric low-level jet. Furthermore, Mathieu et al. (2005) pointed out that atmospheric static stability alone does not act as a strong lid prohibiting vertical gas propagation and is not sufficient to determine the successful use of the NBL budget technique. They found that the combination of both stable conditions and the presence of either a low-level jet (LLJ) or well-defined wind maximum can lead to the reliable application of this technique. Nocturnal jets are a common feature of the vertical structure of the atmospheric boundary layer (ABL) and impact both the meteorology and the local and regional climate. The presence of a jet impedes vertical gaseous dispersion and facilitates gas accumulation, often to the level of the wind maximum, which can be well below the temperature inversion level (Mathieu et al., 2005; Karipot et al., 2006, 2008, 2009).

However, those point observations do not reflect that, during the night, a concentration partitioning occurs not only vertically but also horizontally. After turbulence decays, the wind in the lower atmosphere changes direction (with those changes more pronounced in the presence of a LLJ (e.g. Baas et al., 2009)). Air parcels holding CO_2 molecules for instance drifting at different heights spread spatially throughout the night. Thus the vertical profile of the tracer concentration measured above the point of interest can be formed by various sources whose upwind position is largely a function of both time and height of measurements. In this case, applying a method such as the nocturnal boundary layer budget

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technique to infer the partitioning of the source areas to the resulting scalar profile (footprint) holds considerable appeal (cf. Schmid, 2002; Vesala et al., 2008). The lack of source attribution information hampers a judicious interpretation of measurements in nighttime conditions.

Since representative measurements focused on spatial scales of concentration and flux measurements at elevated sensors during the night seem to be unlikely possible, because of demanding logistical consideration, the present study resorts to numerical modelling. We apply the backward-in-time Lagrangian stochastic (LS) simulations using flow statistics extracted from an atmospheric boundary-layer numerical model (1) to determine the spatial scale of sources contributing to a passive tracer concentration measured at different heights throughout the night, and (2) to quantify the effect of the low-level jet on this scale. This information will be used further to validate fluxes derived using the nocturnal boundary layer budget technique.

2. Modelling

2.1. Meteorological model

The present study uses the atmospheric boundary layer model SCADIS based on E–ω closure scheme (where E is turbulent kinetic energy and ω is specific dissipation of E) and as such belongs to a one-a-half order closure model class. The full description of the model equations and numerical aspects can be found in Sogachev et al. (2002, 2008). The closure was modified to include a description of the canopy and stratification effects (Sogachev and Panferov, 2006; Sogachev, 2009). With respect to canopy flow, the model was extensively tested against field and wind tunnel experiments (Sogachev and Panferov, 2006) and demonstrated both robustness and universality. Numerical tests showed also the practical applicability of the approach used to describe buoyancy effects. For example, the model has been shown to properly reproduces both the surface-layer wind estimated using Monin–Obukhov similarity theory and the mixing height evolution observed above forested terrain in Southern Finland (Sogachev, 2009). For sake of brevity, we do not include any description of the model or any validation tests.

Given the underlying surface properties (e.g. vegetation structure, photosynthetic characteristics, ground albedo, etc.) and specified synoptic conditions (e.g. geostrophic wind speed, air mass characteristics) the solution of the model equations provides diurnal variations of meteorological variables depending only on the diurnal course of net radiation and as such on incoming solar radiation (insolation). In the present study, the evolution of key boundary layer parameters is estimated by SCADIS for different underlying surface properties (bare surface or forest) and insolation (clear sky or overcast). The ground aerodynamic roughness was taken as 0.1 m for both types of surfaces. The forest was presented by trees with a height, h, of 20 m, a total leaf area index of 2 $m^2\,m^{-2}$, and a vertical distribution of leaf area density described by the beta-function with parameters (α, β) = (7, 3) (with a maximum in canopy foliage was located at 3/4 h) (e. g. Sogachev et al., 2008).

Incident solar radiation driven the diurnal cycle of the surface heat budget was computed on July 1 at 50° latitude for clear sky (cloudless) and overcast (full cloud cover) conditions according to Sogachev et al. (2002). The initial profile of temperature was given by a gradient of −0.006 $K\,m^{-1}$ with a temperature of 267 K at the upper boundary of the model domain which was set to 4 km. The initial values of air moisture were constant and equal to 0.003 $kg\,m^{-3}$. The geostrophic wind speed, G, was taken as 10 $m\,s^{-1}$. Evaporation and transpiration were calculated with properties typical for loam soil and for a spruce forest, respectively (Sogachev et al., 2002). It was assumed that the model with these initial conditions approximates a typical variation of wind, turbulence and temperature within the course of a summer day for temperate latitudes. To adequately resolve these variations, a model grid resolution consisting of 120 nodes was arranged with variable distances non-linear with height (about 0.2 m near the ground and 270 m near the upper border of the modelling domain).

2.2. Dispersion model

Transport and diffusion of a tracer were simulated using a random-walk model which has the advantage in calculating turbulent diffusion over a long-range distance. The diffusing particles in the model are immersed in a flow field in which turbulent kinetic energy, mean wind velocity, and Lagrangian time scales determine the particle trajectories. Particle position at time $t+\Delta t$, where Δt is the time step, is given by the Lagrangian relations (Ley, 1982; Legg and Raupach, 1982):

$$x_i(t+\Delta t) = x_i(t) + (\bar{u}_i + u'_i)\Delta t \tag{1}$$

x_i is the coordinate of the particle in i-th direction ($i = x, y, z$ states for the longitudinal, lateral and vertical directions, respectively), $\bar{u}_i$ and u'_i are its mean (averaged in space and time) and instantaneous components of the velocity along x_i, respectively. The components u'_i of the turbulent velocity are computed using a Markov-chain simulation at each time step:

$$u'_i(t+\Delta t) = R_i u'_i(t) + \sqrt{(1-R_i^2)}\sigma_{u_i}\xi_i + T_{L_i}(1-R_i)\frac{\partial \sigma_{u_i}^2}{\partial x_i}. \tag{2}$$

Herein R_i are the Lagrangian autocorrelation functions defined as

$$R_i(\Delta t) = \exp\left(-\frac{\Delta t}{T_{L_i}}\right), \tag{3}$$

where T_{L_i} denotes the Lagrangian time scale, ξ_i is a random number taken from a Gaussian normal distribution with zero mean and unit variance. The velocity variances σ_{u_i} are obtained from simple relation for an isotropic turbulent field as

$$\sigma_{u_i}^2 = \frac{2E}{3}. \tag{4}$$

The time scale T_{L_i} is obtained from the combination of the Taylor theorem with regularities of Fickian diffusion

$$T_{L_i} = K_H \sigma_{u_i}^{-2}. \tag{5}$$

Besides $\bar{u}_i$ and E, the eddy conductivity, K_H, is also supplied by the SCADIS model. In the present study, all meteorological input parameters assumed to be horizontally homogeneous were predefined and available for every 30 min beginning at midnight. Their values at particle positions (x, y, z) (tracked using Δt = 0.5 s) were derived by a simple extrapolation between adjacent values in time and space. We calculated the trajectories of the Lagrangian model (Legg and Raupach, 1982) in a backward time frame (cf. Thomson, 1987; Flesch et al., 1995; Flesch, 1996; Kljun et al., 2002). In this case, the trajectories are initiated at the measurement point itself and tracked backward in time, with a negative time step, from the measurement point to any potential surface source. Using a numerical spin-up procedure (e.g. Kljun et al., 2002), we released scalar particles at different moments in time (considered as the time of measurement) from the receptor point $(0, 0, z_m)$. A reflective boundary condition has been prescribed at the top of the model domain, but in fact no particle reached that height. At the surface, particles were reflected upward at the assumed source location levels and the vertical velocity was reversed (Wilson and Flesch, 1993). Effective source locations were at ground roughness height z_0 = 0.1 m for