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Experimental design for flux measurements: matching scales of observations and fluxes

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Abstract

Practical and conceptual questions of micrometeorological flux measurements over inhomogeneous areas are discussed, especially pertaining to the selection and layout of sites and instrumentation in areas that do not comply with generally accepted standards of fetch and homogeneity. Well-known models for footprint or source area analysis are used to evaluate quantitative criteria for the point-to-area representativeness of each component involved in flux measurements. Based on the acceptance or rejection of these criteria, practical recommendations are formulated about the choice of measurement technique and the height range of instrument deployment for a given situation in the field.

The scope and scale of the practical recommendations for experimental design that are presented concentrate on two principal measurement scenarios: (1) the measurement of a vertical flux, such as water vapour transport, from a specific crop of limited size; (2) the estimate of spatially representative regional fluxes over agricultural, savannah or urban areas. For both scenarios the optimal selection and layout of measurement sites are examined with respect to the choice of measurement technique: by eddy correlation, the profile technique, or via the energy balance approach. It is argued that it is essentially unimportant what method is used, as long as the overriding principle is observed that the scale of areal representativeness of each component of the measurements be matched to the relevant scale of the flux. To achieve this, it is convenient to define a criterion for the point-to-area representativeness of the measurements involved in the determination of the flux. Three different (but related) versions of representativeness criteria are presented in this paper, which differ by their rigour and the amount of effort required for their application. The most pragmatic version, the source area case examination method, is described in some detail, and its use is demonstrated by examples that show that the suitability of observations for flux evaluations varies greatly between the different measurement techniques in practice today. © 1997 Elsevier Science B.V.

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1. Introduction

1.1. Background

Micrometeorologists who intend to measure turbulent fluxes from crops, soil or natural vegetation, or above built-up areas, are confronted with a host of questions of experimental design. These can be summarised in three classes:

1. conceptual questions about the temporal and spatial scale of the study and about the method of observation used. What is the objective of the measurement?
2. Questions about the selection of observation sites and the placement of the instruments. Where, inside the study area, is the best location to set up a meteorological mast, and at what height should the instruments be fixed?
3. Operational and technical questions of instrumentation, measurement physics and data acquisition. Does the signal, measured by the instrument at hand, reflect the behaviour of the physical variable of interest?

Ideally, one would expect to start with the first class of questions and define the scope of the work for which the observations are intended. Are the flux estimates needed as hourly values or as monthly totals? Should the observations reflect the exchange of heat, water vapour or trace gases above a particular field, or are they intended as representative of regional averages? The answer to questions of this kind should provide some guidance for the approach of the second and third classes of questions to determine the number and kind of instruments, masts and data-loggers that are most suitable for a given purpose, where they should be deployed, and at what rate data should be collected.

However, in practice, the method, set-up and operation of flux measurements are often governed by the availability of equipment and by financial, infrastructural and logistical restrictions. The importance of these external factors in determining the character of an observation campaign is always extremely case dependent, making generally valid recommendations nearly impossible. Barr (1987) deplored the fact that these external limitations are often crucial for the realisation of an observation programme, and tend to stand in the way of more objective considerations of experimental design (here, the term 'experiment' is used rather loosely, as is common in meteorology, and does not imply that observations are performed in a controlled system): his short paper is a plea for the principle of 'thinking before measuring' and to deploy instruments to meet experimental objectives.

It is a curious development that, as new instrumentation, recording and processing equipment become more and more specialised and sophisticated, they also tend to become increasingly complex and sensitive to handling errors. Mazzarella (1985) stresses the importance that all components in a measurement system are well matched to each other, to collect 'valid' data. However, a consequence of this notion is that the replacement of a key instrument will usually lead to a new generation of loggers, processors and power supply, etc., and it was already noted by Lettau (1967), at the beginning of the digital era, that the measuring difficulties can become so great that researchers tend to become preoccupied with the mastering of gadgetry and technique and their focus shifts from meteorology to metrology. The relevance of these operational and technical questions of experimental design is well recognised in the micro-meteorological community and is reflected in the abundant literature on observations and experimental design (e.g. Hoehne, 1985; Mazzarella, 1985; Perrier and Tuzet, 1991; Wyngaard, 1991). Thus the discussion of such questions is deliberately omitted here. However, it must be understood that, in any field observation programme, questions of metrology are hardly ever trivial and their careful consideration is imperative for valid measurements. In the following, it is taken as a prerequisite that problems of measurement physics are properly attended to: it is assumed implicitly that the measurements themselves be free of error.

The focus of the present work is on estimates of turbulent exchange, using ground based systems (i.e. meteorological masts or towers) to measure turbulent fluxes directly, by the aerodynamic approach, or in conjunction with the surface energy balance. The arguments presented have only a very limited value for airborne observations (using aircraft or tethersondes), as they are based on the notion of time-averaged point measurements in (quasi-) stationary conditions, and essentially exclude flux estimates inferred from indirect methods (e.g. from SODAR or LIDAR measurements).

The topic of observations from meteorological towers was described by Panofsky (1973) and Kaimal (1986) in chapters in two well-known workshop volumes published by the American Meteorological Society. Panofsky (1973) presented the characteristics of profiles, variances and spectra that can be expected in what he called the 'tower layer' (30–150 m), whereas Kaimal (1986) concentrated on technical matters, such as sensor exposure, data handling and the influence of the tower structure itself on the measurements. Both of these contributions focused on the study of horizontally homogeneous boundary layers or, at the utmost, a one-dimensional land-use change at some distance upstream of the tower.

1.2. Objectives

The present study intends to discuss practical and conceptual questions of flux measurements over complex, heterogeneous surfaces and how they relate to the selection and layout of sites and instrumentation. In most cases, there is more than one possibility to obtain acceptable estimates of surface turbulent fluxes. Here, it is argued that it is essentially unimportant what method is used, as long as the overriding principle is observed that the scale of the measurements be matched to the scale of the fluxes.

The instruments of analysis used to evaluate the relevant spatial scale of a measurement are well-known models of source footprints and source areas for measurements of scalars and scalar fluxes. Rather than reviewing the theory underlying these models, the reader is referred to the recent literature (see below), where the models used in this study are developed and explained in detail. On the contrary, the objective of the present work is to demonstrate the potential for application of these models, in a context that is highly relevant for agricultural meterologists and micrometeorologists, and in a format that is tangible and easily applicable in the field. Thus, the main contribution of this paper lies in the combination of footprint–source area analysis with rigorous criteria for point-to-area representativeness of micrometeorological measurements. This combination is used to formulate practical recommendations for experimental design.

Thus, the question of experimental design is not treated in a comprehensive way, but rather, the focus is on a specific sub-set of problems associated with the observation of surface fluxes in areas that do not comply with generally accepted standards of horizontal homogeneity. It is, however, a set of problems which has been given little attention in the past, but which it is believed will grow in importance as the interest of the boundary-layer research community continues to shift from the well-known, ideal, uniform sites of steppe or prairie grass to non-uniform areas in agricultural, forest or urban environments.

2. Scope and scale

It is obvious that the variety of heterogeneous surfaces is endless. Thus it would be a very ambitious aim to come up with practical recommendations for experimental design that are valid universally for all types of surface variability. To make general and vague principles more tractable, the scope of the argument needs to be limited, and one helpful concept to do this is the notion of scale, which is inseparable from any discussion of heterogeneity (see, e.g. Steyn et al., 1997). The scale of flux studies (and thus of surface variability) considered here is limited to the lower meso-scale (of the order of 1–10 km) or smaller. In terms of Shuttleworth (1988), a surface patchwork in this range of scales is known as 'disorganised variability': individual patches are too small to organise the characteristics of the entire boundary layer and thus, a measurement performed in such a boundary layer is likely to be affected by more than one surface patch.

Two of the most common heterogeneous surface types at the meso- to micro-scale are urban areas and agricultural regions with a patchwork of variable crops. In areas of extensive monocultures or vast natural grasslands it is often possible to perform the micro-meteorological measurements required for flux estimates within the so-called internal equilibrium layer (Perrier and Tuzet, 1991), where fluxes and profiles can be assumed to be well adjusted to local surface conditions. According to the traditionally used rule-of-thumb that the equilibrium sub-layer grows with a slope of approximately 1% downwind of a leading edge, measurements up to a height of 10 m are representative of local surface conditions only if the distance to the nearest field-boundary in the upwind direction exceeds 1 km. Clearly, such comfortable fetch conditions are very rare and most agricultural areas of interest exhibit a smaller-scale field structure that would limit the height of an equilibrium sub-layer to 1 m or less. Although this rule-of-thumb has been shown to be inadequate and outdated (see, e.g. Leclerc and Thurtell, 1990), its use is still widespread, mainly owing to its simplicity. It is one aim of this study to provide recommendations that are simple enough to be used operationally and yet sophisticated enough to be essentially correct.

Variations of surface properties in agricultural areas can affect all kinds of turbulent exchange, be it the flux of momentum, of heat, or of mass (water vapour or trace gases). The vertical transfer of heat or mass can even change sign over small distances. However, in the absence of strong topography, it is possible in many cases to regard individual crop or soil patches as relatively flat (i.e. their vertical structure is on a considerably smaller scale than their horizontal dimension). Hence, it is permissible to assume that the flow over a mosaic of such crop patches is fairly uniform in its mean properties, when viewed at the patch-scale, but that it exhibits considerable variations in its turbulent characteristics. A special case of this category, where changes of surface roughness are relatively small, is a heterogeneous surface of type A in the classification of De Bruin et al. (1991): a non-uniform terrain where primarily the fluxes of heat and water vapour are affected by the irregularities, and the momentum flux is affected only to a much lesser extent. Over an inhomogeneous surface of this type the problem of spatially variable turbulent fluxes is greatly reduced, as it can be assumed that at least the mechanical setting of the exchange processes is roughly uniform. However, variations in soil moisture, plant activity or temperature lead to spatially variable scalar fluxes and associated advective effects that cause problems for the interpretation of flux measurements.

Some difficulties arise with an equivalent assumption over the drastically different conditions of an urban surface (or, to a lesser extent, a mixed forest). Here, the horizontal scale of the principal surface patches (buildings, lawns, streets, groups of trees) is of the same order as their vertical structure. Thus, when viewed at that scale, even the mean flow conditions must be expected to change considerably over short distances. However, in terms of surface morphology, the street, house and lawn scale in an urban setting is equivalent to the scale of individual plants of a crop. One of the foremost concerns of surface flux measurements over urban areas or mixed forests is thus to be far enough above individual surface elements to avoid wake effects, but low enough to still be in the 'constant flux layer'. Schmid et al. (1991) and Schmid and Oke (1992) showed that, in large urban areas, it is possible to define a scale at which the surface may be considered homogeneous. If measurements of turbulent fluxes can be performed at that scale, their meaningfulness for interpretation is greatly enhanced. It is one of the objectives of this paper to present specific criteria for the determination of the scale of a flux measurement.

Rather than attempting to develop such suggestions and recommendations in a rigorous theoretical manner, the underlying principles are presented in an heuristic way, by use of exemplary scenarios of common measurement settings, objectives and techniques. The scenarios to be discussed are: (1) the measurement of water vapour transport to estimate evapotranspiration from a specific crop of limited size; (2) the estimate of regional fluxes over agricultural or urban areas. For both scenarios the optimal selection and layout of measurement sites are examined with respect to the choice of measurement method: by eddy correlation, the profile technique or via the energy balance approach.

3. Scales of observations

The spatial scale of a surface flux estimate depends on the method by which it is obtained. In the measurement scenarios considered here, surface fluxes are either obtained directly by eddy correlation or derived by methods involving measurements of concentration or radiation. All these methods rely on the general assumption that, in the atmospheric surface layer, the exchange of heat, mass and momentum is principally dependent on the capacity of the underlying surface to act as a source or a sink. Over inhomogeneous areas the measured value of an atmospheric variable is thus characterised by those surface patches with the strongest influence on the sensor and varies with position. The scales of observations of such exchange processes are conveniently assessed by considering the relative source strength distribution for a given observation, following Schmid (1994) but generalising that analysis to include surface radiation measurements.

This notion is formally described by a bulk transfer relationship:

$$\eta(\boldsymbol{r}) = \int\int_{\Re} Q_\eta(\boldsymbol{r}') f(\boldsymbol{r}-\boldsymbol{r}')\mathrm{d}\boldsymbol{r}' \tag{1}$$

where η is the value of the measured quantity at point $\boldsymbol{r}$ (i.e. a concentration, a turbulent flux density or a radiation flux density), originating from the source with strength Q_η at $\boldsymbol{r}'$, f is the probability transfer function between $\boldsymbol{r}$ and $\boldsymbol{r}'$, and $\Re$ is the domain of integration. If the source strength distribution is confined to the surface ($z = z_0$), Eq. (1) may be written for an observation point at (x_m, y_m, z_m)

$$\eta(x_m, y_m, z_m) = \int_{-\infty}^{\infty}\int_{-\infty}^{\infty} Q_\eta(x', y', z' = z_0) f(x_m - x', y_m - y', z_m - z_0)\mathrm{d}x'\mathrm{d}y' \tag{2}$$

Here, and in the following, all measures of height refer to the effective height, relative to a potential zero plane displacement. In this case, $f(x_m - x', y_m - y', z_m - z_0)$ relates the value of η at (x_m, y_m, z_m) to the source distribution on the ground and is termed the source weight function (Schmid and Oke, 1988, Schmid and Oke, 1990; Schmid, 1994) or the footprint function (Leclerc and Thurtell, 1990; Schuepp et al., 1990; Horst and Weil, 1992; Leclerc et al., 1992). Given the confusion over these terms in the past, it should be noted that source weight function and footprint function are synonymous and are used interchangeably in this study. However, these terms should not be confused with the source area (defined below), which arises from an integration of the source weight function.

The applicability of Eq. (2) is of course dependent on the surface source strength distribution, Q_η, and on the functional form of the source weight function. One of the fundamental problems in the present context is that the source strength distribution in the area of a prospective measurement site is not known a priori (this being the reason for performing the observation programme). It is thus reasonable to consider only relative source weights here (i.e. the influence of a given point source relative to its source strength). In effect, this is equivalent to assuming a surface consisting of individual marked point sources of unit strength.

The functional form of the source weight distribution, $f(x_m - x', y_m - y', z_m - z_0)$, can be evaluated by considering one such unit point source at a point (x, y, z_0), so that the source strength distribution is written

$$Q_\eta(x', y', z_0) = Q_{\eta,u}\delta(x - x')\delta(y - y') \tag{3}$$

Here, $Q_{\eta,u}$ is a constant of unit source strength to ensure dimensional consistency and δ is the Dirac-delta distribution function. Thus, if the convolution Eq. (2) is performed with Eq. (3), the value of the source weight function, f, is proportional to $\eta(x_m, y_m, z_m)$:

$$\eta(x_m, y_m, z_m) = Q_{\eta,u} f(\Delta x, \Delta y, \Delta z) \tag{4}$$

where $\Delta x = x_m - x$; $\Delta y = y_m - y$; $\Delta z = z_m - z_0$ are the components of the separation vector between the source and the sensor.

Assuming that all individual surface point sources are independent of each other, it thus suffices to compute the distribution of η owing to a unit surface point source with a separation distance of $(\Delta x, \Delta y, \Delta z)$ to evaluate its source weight function. Of course, the problem of the functional form of $f(\Delta x, \Delta y, \Delta z)$ is only deferred and lies now with the determination of the distribution of η owing to a point source. This function is dependent on the nature of η and on the characteristics of its transport between the source and the sensor, and differs greatly between properties that are transferred radiatively and those that are subject to turbulent diffusion. However, if the surface is considered an infinite plane, the source weight function for both radiative and turbulent transport is expected to exhibit a maximum and to fall off asymptotically to all sides as separation distances become large, as illustrated schematically in Fig. 1. For radiative transfer this maximum is located in the nadir of the radiometer, whereas for turbulent transfer the maximum is located in the upwind direction. Thus, if the scale of an observation were understood as the region with a non-zero source weight (i.e. including all sources that can affect the measurement, however small their importance), it would encompass the entire (infinite) plane. Consequently, this interpretation would result in a scale which is itself infinite, although the source weight of the outer reaches is negligibly small.

A more practical concept for the scale of an observation is based on an estimate of what region of the surface is most effectively or most probably influencing the value of η at height z_m. Such a region is defined by considering the source weight function as the distribution of the probability density that a point on the surface has an influence on the measurement. The normal projection into the x–y-plane of any closed curve on the source weight function surface defines a discrete area on the ground. The probability that this so defined portion on the ground exerts a measurable influence on the sensor is proportional to the integral over the source weight function, with the closed curve used as integration limit.

The spatial scale of a given observation of η is then indicated by the smallest possible area to account for a given contribution P (half, say: $P = 0.5$) to the value of η at height z_m. (Note that only the relative source weight is considered here.) This smallest such area, Ω_P, was termed the source area of level P by Schmid and Oke (1988).

The source area may be interpreted in analogy to the 'field of view' of the instrument. The contributions of individual surface elements within the zone of influence are combined to produce a composite influence of the source area, reflected in the measured signal. Thus, the source area is defined as the area bounded by a source weight function-isopleth $f(\Delta x,\Delta y,\Delta z) = f_P$, such that P is the fraction of the total integrated source weight function, φ_{tot}, contained in Ω_P:

$$P = \frac{\varphi_P}{\varphi_{tot}} = \int\int_{\Omega_P} f(\Delta x,\Delta y,\Delta z)\mathrm{d}\Delta x\mathrm{d}\Delta y \Big/ \int_{-\infty}^{\infty}\int_{-\infty}^{\infty} f(\Delta x,\Delta y,\Delta z)\mathrm{d}\Delta x\mathrm{d}\Delta y \tag{5}$$

where φ_P is the integral of the source weight function over Ω_P. Because, with given measurement height and transfer conditions, f is dependent only on the horizontal separation, Eq. (5) can be simplified, using Eq. (4), by

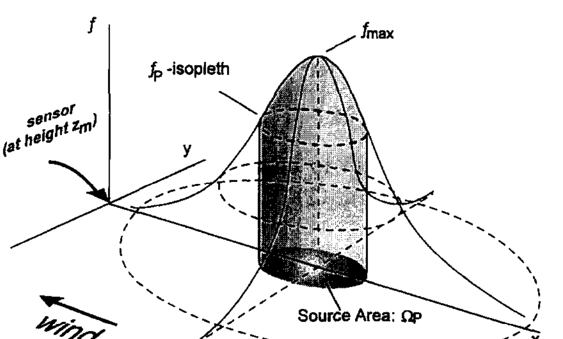

Fig. 1. The source weight function, or footprint function, and its relation to the source area. The source weight is small for small separation distances. It will rise to a maximum with increasing distance and then fall off again to all sides as the separation is further increased. The total volume under the source weight function is φ_{tot}. P is the fraction of this volume bounded by the isopleth f_P, and the cylinder surface below it (hatched). The source area of level P, Ω_P, is the area bounded by the normal projection of the isopleth f_P on the $x-y$-plane. For footprints of diffusing quantities, horizontally homogeneous turbulence is assumed, with the mean wind direction parallel but counter to the x-axis direction. The radiation footprint is centred over the nadir axis of the radiometer (adapted from Schmid, 1994).

considering a measurement at point $(0,0,z_m)$ and reversing the x and y coordinates (compare with Fig. 1). Eq. (5) is then written

$$P = \frac{\varphi_P}{\varphi_{tot}} = \int\int_{\Omega_P} \eta(x,y,z_m)\mathrm{d}x\mathrm{d}y \Big/ \int_{-\infty}^{\infty}\int_{-\infty}^{\infty} \eta(x,y,z_m)\mathrm{d}x\mathrm{d}y \tag{6}$$

The specific functional forms that footprint functions and source area computations take (for radiative fluxes, scalar concentrations, and scalar fluxes, respectively) are discussed in the following. For each species, the equivalent formulations for the source weight density, f, and for Eq. (4) and Eq. (6) are presented. Details of their derivations are deferred to Appendix A or are described in the literature.

3.1. Radiation footprint function and source areas

It is a common assumption that the radiative exchange between an inverted flat-plate radiometer (or the lower surface of a net-radiometer) and the underlying surface is governed by Lambert's cosine law, implying that the surface behaves like a totally diffuse emitter–reflector. In this case, the radiation flux, F_R, received from the surface below by a differential area parallel to the surface and mounted at height z_m (i.e. by the receptor plate of the radiometer) is symmetric relative to the axis normal to the centre of the receptor plate (cylindrical symmetry). The radiative transfer is governed by the geometry of the measurement and can be expressed in terms of the measurement height, z_m, and the radial separation from the nadir point of the measurement, r (refer to Appendix A for an overview of the radiation measurement geometry). Over inhomogeneous surfaces, the absolute amount of radiation received from any given surface element is obviously dependent on its particular radiative properties, in addition to the measurement geometry. For this reason it must be kept in mind that the source weight distribution is a relative transfer function which is superimposed on (or convoluted with) the source strength distribution, as defined in Eq. (1).

Thus, for a radiative flux, the variables in Eq. (4) become

$$\left.\begin{array}{lll} \eta \to \dfrac{\mathrm{d}F_R}{\mathrm{d}A}; & \text{radiation flux density} & \\ Q_{\eta,u} \to I_{n,u}; & \text{unit radiation intensity} & \\ f \to f_R = & \dfrac{\mathrm{d}F_R}{\mathrm{d}A \cdot I_{n,u}} = \left(z_m + \dfrac{r^2}{z_m}\right)^{-2} & \end{array}\right\} \tag{7}$$

where $\mathrm{d}A$ is a differential area on the ground, radiating with $I_{n,u}$. The derivation of Eq. (7) is given in Appendix A (Eq. (A4)). This radial variation of the radiative flux source weight function is plotted in Fig. 2, where f_R is scaled by its value at the nadir point, $f_{R,0}$, where $r = 0$.

Owing to the cylindrical symmetry, isopleths of f_R form concentric circles, with the nadir axis at their centre. The areas contained in these circles are thus the integration domains for the source area computations, as defined by Eq. (5) and Eq. (6). The integration of Eq. (7) according to Eq. (5), or rather, of a form of Eq. (7) that is better suited for this purpose, is presented in Appendix A (leading to Eq. (A8)). The result is an equivalent relation to Eq. (5), for the radiation source area, in terms of r and z_m:

$$\left.\begin{array}{lll} \Omega_P \to \Omega_{P,R}: & \text{radiation source area, level } P & \\ \varphi_P \to \varphi_{P,R}: & \text{fraction } P \text{ of total effect, } \varphi_{tot,R} & \\ P \to P_R = & \dfrac{\varphi_{P,R}}{\varphi_{tot,R}} = \dfrac{r(\Omega_{P,R})^2}{r(\Omega_{P,R})^2 + z_m^2} & \end{array}\right\} \tag{8}$$

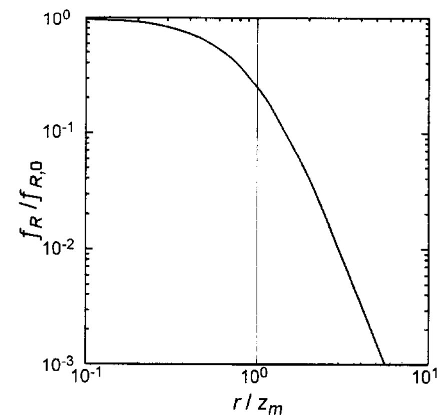

Fig. 2. Radial variation of the radiation source weight function, scaled by the maximum source weight at the nadir point. The functional form of this curve is given by Eq. (7).

For any P_R ($0 \le P_R \le 1$), the ratio of the radius of the P_R-level radiation source area to the sensor height is given by (Schmid et al., 1991)

$$\frac{r(\Omega_{P,R})}{z_m} = \left(\frac{1}{P_R} - 1\right)^{-1/2} \tag{9}$$

As shown in Fig. 3, a useful rule-of-thumb resulting from this simple relation is that half of the radiative surface influence originates from an area with a radius equal to the sensor height.

3.2. Footprint functions and source areas of diffusive scalars

In the case of turbulent transport, the determination of a surface source area is more complex. The temporally averaged 'field of view' of a temperature or humidity sensor is determined not primarily by geometry, but rather by the turbulent diffusion characteristics in the layer between the sensor and the surface. It is constantly changing both its size and position, depending on wind direction and speed and other qualities of the flow.

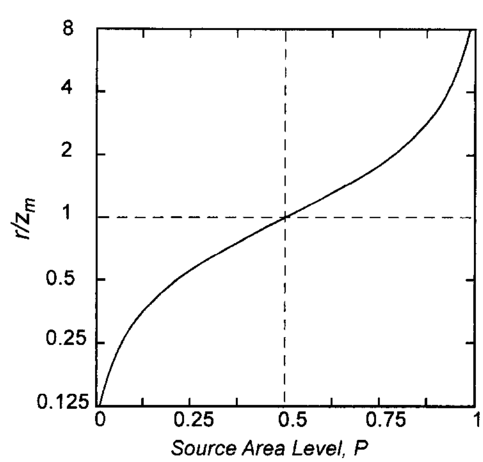

Fig. 3. The size of the radiation source area, dependent on the source area level, P. Half of the radiative surface influence originates from an area with a radius (r) equal to the sensor height (z_m).

The problem of the 'view factor' and the source area of a scalar that is subject to turbulent diffusion was addressed by Pasquill (1972), and more recently by Schmid and Oke (1990) and Schmid (1994). In this work, K-theory is used to describe the diffusion of a passive scalar in the surface layer. The source weight function of a scalar concentration is found by applying Robert's solution to the advection–diffusion equation, as described by Gryning et al. (1987). This model has been verified against measured concentration data, as reported in the Appendix of Schmid and Oke (1990).

According to Eq. (4), the scalar concentration source weight for a reference point at $z = z_m$, and sources at $z = z_0$ and horizontal separation (x, y), is proportional to the concentration value $C(x, y, z_m)$ owing to a unit point source $Q_{C,u}$. Thus, if the defining relation for the source weight function, Eq. (4), is adapted to a measurement of scalar concentration, the variables become

$$\left.\begin{array}{lll} \eta \to C: & \text{scalar concentration} & \\ Q_{\eta,u} \to Q_{C,u}: & \text{unit surface point source of } C & \\ f \to f_C(x, y, z_m - z_0) = & \dfrac{C(x, y, z_m)}{Q_{C,u}} = \dfrac{D_y(x, y) D_z(x, z_m)}{U(x)} & \end{array}\right\} \tag{10}$$

In Eq. (10), D_y and D_z are the crosswind and vertical concentration distribution functions, respectively and $U(x)$ is the effective speed of plume advection, where the notation of Horst and Weil (1992) is used. Gryning et al. (1987) provided a closed set of equations for $D_y(x, y)$, $D_z(x, z)$ and $U(x)$, dependent on the standard surface layer scaling parameters (see Schmid (1994) for further details).

The scalar concentration-source area is evaluated by substituting Eq. (10) into Eq. (6) to give

$$\left.\begin{array}{ll} \Omega_P \to \Omega_{P,C}: & \text{scalar concentration source area, level } P \\ \varphi_P \to \varphi_{P,C}: & \text{fraction } P \text{ of total effect, } \varphi_{\mathrm{tot},C} \\ P \to P_C = & \dfrac{\varphi_{P,C}}{\varphi_{\mathrm{tot},C}} = \displaystyle\iint_{\Omega_{P,C}} \frac{D_y(x, y) D_z(x, z_m)}{U(x)} \mathrm{d}x\mathrm{d}y \bigg/ \int_0^{\infty} \frac{D_z(x, z_m)}{U(x)} \mathrm{d}x \end{array}\right\} \tag{11}$$

Details of the numerical evaluation of this relation have been given by Schmid (1994), who presented also the characteristic dimensions of the 50% source area, as functions of the measurement height above the roughness length, z_m/z_0, the stability z_m/L (where L is the Obukhov length), and the level of crosswind turbulence, σ_v/u_* (u_* is the friction velocity, and σ_v denotes the standard deviation of lateral wind fluctuations).

Schmid (1994) also presented a parametric regression model which approximates the full source area model for moderate input values. However, it turns out that in most situations of interest, the parameterised version is invalid or very inaccurate. For this reason it is strongly recommended that the full version of the model is used at all times. For convenience, the model details and a form to submit model run requests have been made available on the world-wide-web by the present author. The internet address of this web-site is:

http://www.indiana.edu/~climate/SAM/SAM-ESAM.html

3.3. Footprint functions and source areas of scalar fluxes

The problem of some form of footprint function for scalar fluxes (or precursors thereof) has been addressed in the literature by numerous workers (e.g. Gash, 1986; Leclerc and Thurtell, 1990; Schuepp et al., 1990; Wilson and Swaters, 1991; Horst and Weil, 1992; Leclerc et al., 1992; Schmid, 1994). For convenience of notation, the present work follows the study by Schmid (1994), which was based on that of Horst and Weil (1992) and Horst and Weil (1994).

Using K-theory as in Eq. (10), the vertical flux of a scalar C, F, is expressed as

$$F(x, y, z) = -K_C(z)\frac{\partial C}{\partial z} = -K_C(z) Q_{C,u} \frac{D_y}{U} \frac{\partial D_z}{\partial z} = D_y(x, y)\overline{F^y}(x, z) \tag{12}$$

where $K_C(z)$ is an eddy diffusivity and $\overline{F_y}$ is the crosswind integrated flux. $\overline{F_y}$ can be expressed in terms of the mean wind speed profile and the crosswind integrated concentration, by the two-dimensional advection–diffusion equation (see, e.g. Schmid, 1994). After integration of the advection–diffusion equation, the two-dimensional flux-source weight function $f_F(x,y,z_m\text{-}z_0)$, for $x>0$, follows according to Eq. (4), using the following substitutions:

$$\left.\begin{array}{ll} \eta \to F: & \text{vertical turbulent flux of } C \\ Q_{\eta,u} \to F_u: & \text{unit surface point source for flux of } C \\ f \to f_F(x,y,z_m\text{-}z_0) & = \dfrac{F(x,y,z_m)}{F_u} \\ & = \dfrac{1}{F_u}\overline{F^y}(x>0,z_m)\,D_y(x,y) \end{array}\right\} \quad (13)$$

It follows from a continuity argument that $\varphi_{tot,F}$ becomes unity, and thus, the substitutions to adapt Eq. (6) for a scalar flux source area are

$$\left.\begin{array}{ll} \Omega_P \to \Omega_{P,F}: & \text{scalar flux concentration source area, level } P \\ \varphi_P \to \varphi_{P,F}: & \text{fraction P of total effect, } \varphi_{tot,F} \\ P \to P_F = & P_F = \dfrac{\varphi_{P,F}}{\varphi_{tot,F}} = \dfrac{1}{F_u}\displaystyle\int_{\Omega_{P_F}}\!\!\int F(x,y,z_m)\,\mathrm{d}x\,\mathrm{d}y \end{array}\right\} \quad (14)$$

Again, details of the derivation of Eq. (14) and of its numerical evaluation have been given by Schmid (1994), together with the characteristic dimensions of $\Omega_{P,F}$, for $P_F = 50\%$ as functions of z_m/z_0, z_m/L, and σ_v/u_*.

The parameterisation equations for the 50% flux source area model (FSAM) presented by Schmid (1994) are fraught with the same restrictions as for the scalar concentration version of the model (see above). In addition, the 1994 version of FSAM contained an error with the effect that the measurement height z_m needs to be approximately doubled. For this reason, it is recommended that the full version of FSAM is used, which is also available at the world-wide-web site given above. The flux footprint model used here and at the web-site is the analytical model presented by Horst and Weil (1994). This model has been tested against experimental data of trace gas fluxes by Finn et al. (1996).

It turns out that scalar flux source areas are smaller than scalar concentration source areas, by approximately an order of magnitude (Schmid, 1994). The question about the physical reason for this difference was put to the author and discussed at every occasion where source area modelling results were presented over the last few years. However, although the difference is very clear mathematically, it is harder to demonstrate it by physical arguments.

The flux of a scalar C is related to the concentration of C through the advection–diffusion equation. By the integrated form of this equation, the flux at a given height varies with the vertical integral (up to this height) of the horizontal concentration gradient (see Schmid (1994), Eq. (16)). At any given height, the along-wind roll-off of the concentration profile becomes increasingly flat with distance, although the concentration level itself may not yet have dropped drastically below its maximum value. This finding indicates that horizontal concentration gradients (and thus the vertical flux) decrease more steeply with distance compared with the concentration magnitude, which would explain a smaller source area for the flux at least qualitatively.

As will be discussed below, the differences between the scalar flux and the scalar concentration source areas can have serious consequences for the interpretation of energy balance or Bowen-ratio profile results, if they rely on a combination of direct flux and concentration measurements.

4. Spatial representativeness

It is a well-recognised requirement for the usefulness of results from micrometeorological field programmes that they be based on a representative dataset. Thus, the spatial representativeness of surface flux measurements can serve as a rationale for the formulation of recommendations for their experimental design. However, the criteria of what comprises representativeness are largely subjective and must be tuned to the objectives of the application at hand. In view of the judgmental nature of representativeness, Nappo et al. (1982) attempted to summarise previously published formulations. For convenience, their terminology is adopted in this work. They concluded that there are two fundamental kinds of representativeness, spatial and temporal, and that representativeness is a function of scale. In search of a consensus, they proposed to define representativeness as "the extent to which a set of measurements taken in a given space–time domain reflects the actual conditions in the same or different space–time domain taken on a scale appropriate for a specific application".

Thus the magnitude and variability of a set of observations at a given location and time may be representative of the behaviour of a meteorological phenomenon in a different location, or of a process taking place at a different time. However, it is important that the temporal and spatial scales of the observations match the scales of the phenomenon. Both temporal and spatial representativeness require that the physical setting (or boundary conditions) of the observations correspond to the physical setting of the phenomenon, i.e. to the prevailing atmospheric conditions at a larger temporal and spatial scale than that of the phenomenon itself. In effect, if the measurements should be representative of fair-weather, daytime evaporation rates in summer, they must be taken in similar conditions.

Nappo et al. (1982) differentiated between several types of representativeness. Among these, the so-called point-to-area representativeness is of primary interest here. It describes the extent to which the measured conditions in a point reflect the actual (average or aggregated) conditions over a surface.

In the context of the two principal surface flux measurement scenarios discussed here, the instruments are exposed to atmospheric conditions at a certain height above the physical surface. Yet, the aim of both scenarios is to obtain an estimate of the exchange of mass, heat or momentum between the (physical) surface and the atmosphere. Thus the conditions measured by the instruments on the mast are supposed to represent a process at the surface, or rather, a specific portion of the surface. Taken a step further, the scenario (2) for the estimate of regional fluxes requires in addition that the specific portion of the surface with a direct influence on the sensor is representative of the entire region. This example indicates that there is often a logical catena linking the different types of representativeness: surface-to-surface, point-to-surface, and, ultimately, the measurement–physics representativeness.

Nappo et al. (1982) stated that there is no quantitative method to determine representativeness absolutely, but in practice specific criteria for representativeness and intervals within which they can be accepted need to be defined. Criteria for point-to-area representativeness are most easily defined, if the temporal and spatial variability of a measurement field is known, e.g. (Nappo et al., 1982)

$$\Pr\left\{\frac{(\eta-\overline{\eta})^2}{\overline{\eta}^2} \le \delta^2\right\} = \Pi \tag{15}$$

i.e. there is a probability Π that the point measurement, η, lies within $\pm\delta$ (in per cent) of the area-average value, $\overline{\eta}$. In the present context, η corresponds to the measured value of the flux, and $\overline{\eta}$ depends on the scenario. For Scenario (1) $\overline{\eta}$ is the 'true' average flux of the specific surface patch in question, and in Scenario (2) $\overline{\eta}$ is the 'true' spatially aggregated flux of the region. Considering the inherent uncertainty of flux measurements even over homogeneous areas, a δ of 10% should be acceptable in most cases.

Keeping the present objective of experimental design criteria in mind, the variability of the measurement field is not a priori known, as the representativeness of measurements needs to be established before the start of observations. Simple and easily verified representativeness criteria that are based on existing observations, such as Eq. (15), are thus not directly applicable to the problem of planning an observation programme with a view to obtaining data that meet their objectives. However, for the temporally averaged measurement of a surface flux in the lowest 10% of the boundary layer (i.e. the surface layer which, over homogeneous conditions, is often called the constant flux layer), it is assumed that the variability of the measurement field is primarily influenced by the spatially variable surface conditions. The problem in this case is that the specific portion of the surface which is expected to influence the measurement most prominently must be estimated by a model of the source area, or alternately, the source weight distribution or footprint. As demonstrated in Section 3, suitable source weight distribution models depend on the type of sensors involved in evaluating the flux.

Thus, to make it tractable, the representativeness criterion given by Eq. (15) is modified in two steps here. The first approach (the 'footprint method') is used to estimate η and $\overline{\eta}$ by a source weight function model, and by assuming that the diffusion from each point on the surface is independent from that of its neighbours. The second step (the 'source area method') uses an integrated form of the source weight function, and is a practical and economical method to estimate the spatial representativeness of flux measurements that leads to generally applicable design criteria. The concepts of both these methods are developed below.

4.1. The footprint method

The basis of the footprint method is previous knowledge of the distribution of source strengths from each point of the surface in the region, $Q_\eta(\boldsymbol{r}')$, as indicated by Eq. (1). In practice, such information is not available and thus the source strengths must be estimated based on, for example, surface temperature, surface moisture availability, non-dimensional vegetation index (NDVI) or trace gas production rates.

The source weight function, $f(\boldsymbol{r}-\boldsymbol{r}')$, assigns a relative weight to each of the source strengths Q_η, depending on the separation between the measurement and the source, $\boldsymbol{r}-\boldsymbol{r}'$. The ensemble of these weighted sources makes up the 'footprint', or what the sensor 'sees' of the surface sources. Assuming that interaction between the sources is negligible, the expected value measured by the instrument, $\eta(\boldsymbol{r})$, is then given by Eq. (1). For the evaluation of Eq. (15), the formulations of η and $\overline{\eta}$ need to be evaluated separately for the cases where η is a flux or a concentration.

If η is a flux (i.e. $\eta \to F$, and $Q_\eta \to F_0$), and neglecting any vertical flux divergence between the surface and the measurement height, the average value, $\overline{F}^{\mathrm{A}}$, is easily obtained by

$$\overline{F}^{\mathrm{A}} = \frac{1}{A}\int\int_A F_0(\boldsymbol{r}')\mathrm{d}\boldsymbol{r}' \tag{16}$$

where A is the area of reference, for which the measurement should be representative. In Scenario (1) A is the area of a specific surface patch, and in Scenario (2) A is the entire region.

If η is a scalar concentration (C), the evaluation of $\overline{C}^{\mathrm{A}}$ at height z_{m} is not so simple. The average surface flux of $\overline{F}_0^{\mathrm{A}}$, is related to the average concentration at height by an effective flux–profile relation, based on an effective surface parameter, which, in itself, is a very contextual and not easily evaluated quantity (for a discussion of this topic see, e.g. Schmid and Bünzli (1995)). In view of this rather involved procedure to evaluate $\overline{C}^{\mathrm{A}}$, the use of more simplified representativeness criteria for scalar measurements, such as described below in Section 4.2, is advisable.

If η and $\overline{\eta}$ (for a concentration or a flux) is nevertheless specified by the footprint method, using the formulations of Sections 3.1, 3.2 and 3.3, the expected difference between the two, δ, can be evaluated according to Eq. (15). The probability distribution of δ can be obtained by evaluating η (and thus δ) over a range of stabilities and wind directions, or by varying the surface source strength distribution randomly over an estimated uncertainty interval.

As is indicated in Eq. (2), and the specific formulations in Eq. (7), Eq. (10), and Eq. (13), the source weight distribution for both fluxes and scalars, and therefore the expected representativeness of a measurement, depends strongly on the measurement height and on the horizontal separation between the surface sources of interest and the sensor. Thus, the representativeness criterion Eq. (15) can be applied to estimate an optimal height range and site location for a measurement to meet the objectives at hand.

The above analysis serves primarily to demonstrate that the formulation of precise experimental design criteria, based on a standard concept of spatial representativeness, such as Eq. (15), is possible in principle. However, it equally demonstrates that the effort required to perform the computations for the complete footprint method is immense and can hardly be expended or justified in applied cases of experimental design.

4.2. The source area method

To overcome the difficulties encountered in the footprint method, the formal representativeness criterion Eq. (15) needs to be relaxed and converted to a more qualitative argument. Here, as anywhere, practicality is achieved at the expense of universality and formal rigour. Nevertheless, it is proposed that the approximate source area method to evaluate the point-to-area representativeness of a measurement presented here, is well suited for on-site questions of experimental design and as a tool to examine the plausibility of measured data.

In scenario (1) the interest is to determine the flux from a specified homogeneous surface patch of limited area. Thus, a flux measurement is representative of the conditions inside that patch to the extent that the surface source area of the sensor is located inside the patch. Scenario (2), on the other hand, aims at aggregating the variable surface influence of a region. The extent to which a flux measurement is representative of the regional average is given by the fraction of the regional surface variability contained inside the source area. In the following, this concept is brought into a functional form that serves as a basis for a formal representativeness criterion, similar to Eq. (15).

The source area of a measurement effectively separates the most dominant surface area of influence from those regions on the ground with relative source weights that are too low to be of importance. In general, the 50% source area boundary of a diffusive scalar concentration measurement, or of a scalar flux, corresponds to a source weight function isopleth with a value on the order of 10–20% of the maximum source weight (the exact value depends on the surface roughness and on atmospheric stability). As a consequence, a point source located on or outside the 50% source area boundary must be at least 5–10 times stronger than the point source at the maximum source weight location (in the centre of the source area), to achieve a similar response in the sensor. For a radiative flux, the drop-off of the relative source weight from the maximum is slower: the source weight corresponding to the 50% source area boundary is still over 20% of the maximum, but drops off quickly afterwards (see Fig. 2).

Using the source area concept as an estimate for the spatial scale of the measurement, the spatial representativeness of a measurement can be evaluated by comparing the surface characteristics inside the source area with those of the area of reference. These surface characteristics are given by the distribution of source strengths in the area (Q_η), and may take the form of radiation temperatures, soil moistures (or similar), depending on the type of measurement involved. The average conditions in the source area, Ω_P, are evaluated as

$$\overline{Q}_\eta^{\Omega_P} = \frac{1}{\Omega_P} \int\int_{\Omega_P} Q_\eta(\boldsymbol{r}')\mathrm{d}\boldsymbol{r}' \tag{17}$$

where Ω_P takes the place of the radiative, scalar concentration, or scalar flux source area, as the case may be. In contrast to the footprint method, where the full convolution of the source strength distribution with the source weight function must be computed, the source area method evaluates $\overline{Q}_\eta^{\Omega_P}$ in a bulk approach. The crudeness of this approximation is compensated by the fact that the tedious convolution of the source strength distribution with the asymptotic footprint function in Eq. (2) is replaced by a simple averaging process in Eq. (17).

The average surface conditions in the area of reference, A, are given by a generalised form of Eq. (16), without requiring that η be a flux:

$$\overline{Q}_{\eta}^{A} = \frac{1}{A} \int\int_{A} Q_{\eta}(\boldsymbol{r}')\mathrm{d}\boldsymbol{r}' \tag{18}$$

Again, A is the area of the specific surface patch of interest for Scenario (1), and indicates the entire region for Scenario (2).

With Eq. (17) and Eq. (18) taking the place of η and $\overline{\eta}$ in Eq. (15), a modified representativeness criterion for the source area method can be formulated as

$$\mathrm{Pr}\left\{ \frac{\left(\overline{Q}_{\eta}^{\Omega_P} - \overline{Q}_{\eta}^{A}\right)^2}{\left(\overline{Q}_{\eta}^{A}\right)^2} \le \delta^2 \right\} = \Pi \tag{19}$$

i.e. there is a probability Π that the average of the surface conditions inside the source area, $\overline{Q}_{\eta}^{\Omega_P}$, lies within $\pm\delta$ (in per cent) of the average value of the area of reference, $\overline{Q}_{\eta}^{A}$. In principle, the probability distribution of this expression is obtained in an analogous process to that of the footprint method: the input parameters of the source area model are varied over an expected range of stabilities and wind directions, or the source strength distribution is varied over an expected uncertainty interval. However, in many instances it may be sufficient to examine the surface cover inside the source area for a few cases over a range of stabilities, sensor heights and wind directions, and come up with a subjective estimate of the point-to-area representativeness of a measurement. This case examination method lends itself well to ad hoc representativeness checks in the field and is described in the next section, using the scenario examples introduced in Section 2.

5. Illustrative examples: the source area case-examination method

5.1. Scenario 1: evapotranspiration measurement of a given crop

The aim of Scenario 1 is to examine the optimal placement of sensors, with a view to obtaining a valid estimate of the turbulent evaporative flux from a given field of limited size (an alfalfa crop, say). The setting of this scenario is an imaginary agricultural area with a mixture of short crops and tilled fields, resulting in an average surface roughness length of $z_0 = 10^{-2}$ m. The state of the boundary layer is moderately unstable, with an Obukhov length of $L = -10^2$ m, and cross-wind turbulence near the surface characterised by $\sigma_v/u_* = 1.5$. The proposed site for an instrument mast in our alfalfa field is located about 100 m downwind (along the main wind axis) of the field boundary, as indicated in Fig. 4a,b. For convenience, it is assumed that all questions of measurement technique and of instrumental precision are attended to in a satisfactory manner, so that the placement of the sensors, and the particular method to derive the flux are the only remaining questions of experimental design.

Evaporation can be measured in various ways. The most direct micrometeorological method is by eddy-correlation, using a fast response humidiometer in association with a sonic anemometer for the vertical wind speed fluctuations. As the turbulent flux of moisture is measured directly, the relevant surface source area can be estimated using the flux source area model described in Eq. (14). The 50% source area outlines for measurement heights of $z_s = 2$ m and 5 m are plotted in Fig. 4a, together with the respective maximum source weight locations. The source areas indicate that eddy correlation measurements at both heights are influenced mainly by the local surface patch. Thus, they both are expected to give reasonable estimates of the alfalfa evaporation. Obviously, for eddy correlation measurements, the choice of measurement height in the range 2–5 m is not critical for the given conditions. However, a change in wind direction, or a thermal regime tending more

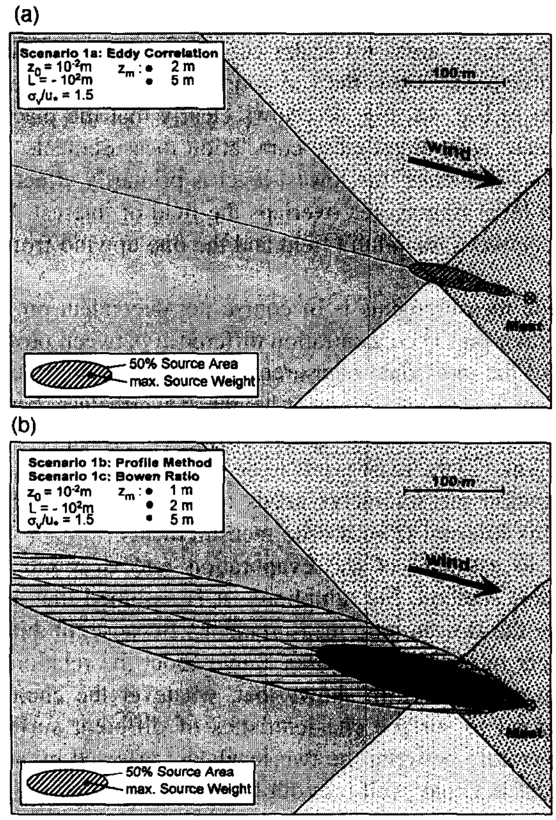

Fig. 4. (a) Scenario 1a, eddy correlation method, plan view of the setting. The four different field types are indicated by variations in shading. The instrument mast is located in the field of interest, about 100 m downwind of the leading edge. The 50% flux source areas for both measurement heights of 2 m and 10 m are largely inside the field of interest. The points of maximum source weight are indicated by dots inside the respective source areas: the one closest to the mast refers to the 2 m source area. (b) Scenario 1b, profile method. (See (a) for an explanation of the setting.) The points of maximum source weight are indicated by dots inside the respective source areas: with increasing distance from the mast they refer to the 1 m, 2 m, and 5 m source areas, respectively. The 50% scalar concentration source areas are larger than the corresponding flux source areas. With height, the source areas are increasingly overlapping areas outside the field of interest. The measured profile can thus not be expected to be in equilibrium with the alfalfa evaporation flux. Scenario 1c: Bowen ratio–energy balance combination method. For the two components of the Bowen ratio measurement, similar reservations as for the profile method apply. The upwelling part of radiation measurements is influenced only very locally, with 50% radiation source areas for sensor heights up to 5 m that are too small to be plotted in this figure.

towards stable conditions (resulting in a source area that is shifted upwind and has increasingly elongated proportions), will move a considerable portion of the $z_s = 5$ m source area outside the crop of interest.

If eddy correlation equipment is not available, the use of the aerodynamic or profile method to arrive at an evaporation estimate is still widespread, even over inhomogeneous areas. With this method, measured profiles of wind speed and specific humidity are related to the vertical flux by one of the well-known flux–profile relationships (e.g. the Dyer–Businger relations; Businger, 1988). Over areas that are patchy with respect to the thermal or humidity properties of the surface, a source area analysis of each instrument level of the humidity profile can shed some light on the consistency of the profile and on the fetch conditions of each level. As it is the humidity concentrations that are measured in the profile method the scalar concentration source area model Eq. (11) is relevant here. However, it should be noted that the resulting source areas do not refer to the flux inferred from the measurements, but to the (scalar) measurements themselves (see also below).

Fig. 4b shows the 50% source areas and maximum source weights for humidity sensors mounted at 1, 2 and 5 m. As mentioned in Section 3, source areas for scalar concentrations are about an order of magnitude larger than the corresponding flux source areas. Thus, the 1 m concentration source area in Fig. 4b is almost the same size as the 5 m flux source area in Fig. 4a. Fig. 4b shows clearly that the placement of sensors in the profile method is much more critical than with direct eddy correlation measurements. The set of humidity sensors is exposed to an inconsistent surface influence: the lowest level is primarily affected by local conditions, whereas only a minor portion of the 5 m level source area overlaps the field of interest. Depending on the magnitude of the surface moisture difference between the alfalfa field and the one upwind from it, the resulting profile can be difficult to interpret.

The flux inferred from profile measurements is of course not dependent on absolute concentrations, but on the concentration gradient, or rather on the concentration difference between profile levels. It is thus not directly the concentration source areas of the individual measurement levels that are relevant for the flux, but rather the rate of change of these source areas with measurement height. Nevertheless, based on the concentration source area estimates, the conclusion for experimental design in the present example is that (1) the profile needs to be restricted to heights lower than 3 m, (2) the mast needs to be moved further away from the upwind leading edge, or (3) a change to eddy correlation measurements needs to be considered. Horst (1997) presents the direct estimate of the footprint of a flux inferred from profile measurements.

A third method to obtain an estimate of the evaporation rate over our alfalfa field is by an energy balance–Bowen ratio combination method. For simplicity, it is assumed that any two of the three instrument levels in Fig. 4b can be used for the Bowen ratio measurements. Because, in this conventional form, the Bowen ratio method is a (bulk-) gradient method, the remarks above about the relationship between the concentration source areas and their gradient apply. Fig. 4b shows that, whatever the choice, the two components of the Bowen ratio are likely to respond to surface characteristics of different surface patches, and thus the bulk gradients of humidity and temperature between the two levels are affected by advection, in addition to vertical transport. However, this inconsistency may well turn out to be irrelevant, owing to the common finding that the partitioning of energy between sensible and latent heat is similar over various types of vegetation, leading to similar Bowen ratios.

As long as horizontal homogeneity in the partitioning of energy can be assumed, it is thus much more important that the net-radiation measurement (that forms the core of the energy balance part of the technique) reflects conditions that are local and representative of the crop of interest. The diameter of the 50% radiation source area is equal to the sensor height, and thus the upwelling part of the net radiation measurement is influenced almost entirely by local conditions, even if the sensor is mounted at 5 m on the mast in Fig. 4b. Indeed, the radiation source area is so small that it cannot be shown in the figure.

On the other hand, if the patchiness is characterised by considerable differences in surface moisture (e.g. by selective irrigation) the above assumption needs to be discarded and the same restrictions as for the profile method apply.

5.2. Scenario 2: estimate of a regionally representative heat flux over patchy terrain

Whereas the setting of Scenario 1 is purely imaginary, the basis for Scenario 2 is an observation programme performed in summer 1985 over a suburban area of Vancouver, Canada (Schmid, 1988; Schmid et al., 1991). The aim of that study was the examination of the spatial scales of variability of sensible heat flux over the highly variable, patchy suburban area. Application of the principles and methods presented here to the data of that observation programme gives an indication of the validity of these methods. Because this observation programme also included a calibration period for the source areas of the two sensors involved, the effect presented below could be clearly attributed to changes in source area. This observation programme thus constitutes a direct test of the source area model against measured data.

Thus, Scenario 2 is concerned with the determination of a regionally representative sensible heat flux over patchy suburban terrain, where the thermal conditions of the surface vary at scales of 10–200 m (corresponding

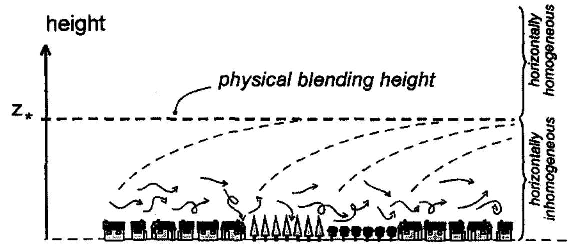

Fig. 5. Schematic diagram of the surface layer structure over a rough, patchy area. The physical blending height is defined as the level above which profiles and fluxes attain horizontal homogeneity by turbulent mixing, or blending.

to the average house spacing and the length of a 'block'; see Schmid and Oke (1992)). Over such a very rough and patchy suburban surface the structure of the surface layer needs to be considered in two parts, as illustrated schematically in Fig. 5. The level of the physical blending height, above which the structure of turbulence has attained horizontal homogeneity by turbulent mixing, is expected to vary with the mixing activity (i.e. essentially stability).

The degree of inhomogeneity at 30 m height was determined by two sets of eddy correlation instruments mounted on 30 m towers. Whereas one tower remained fixed at one site, the other was a telescopic mobile tower that was operated at five different sites, 1 week at the time. The separation of the mobile sites and the fixed tower was on the order of approximately 1 km. At the end of this 'inter-site period', the two sensors were mounted side by side in a 'calibration period' of similar duration, to examine the sensor-induced differences in the flux measurements (see Schmid et al., 1991).

The role of the self-induced spatial averaging of turbulent transport may be examined by analysing the measured variability of sensible heat flux vs. the size of the flux source area: if the 'averaging power' of the flow is large, the flux source area is also large and the remaining spatial variability of flux measurements is expected to be reduced. Thus, source area analyses of the observations were applied to the sensible heat flux variability (ΔQ_H spread) for both the inter-site period and the calibration period (Schmid et al., 1991). The term 'ΔQ_H spread' denotes the variation (with source area size) of the median of the Q_H-difference distribution between the two sensors (see Schmid, 1988). It is determined by a non-parametric, locally weighted regression scheme LOWESS (Chambers et al., 1983). The results are summarised in Fig. 6. Although the curve for the inter-site period shows a marked reduction of spatial variability towards larger source areas, an overall trend is

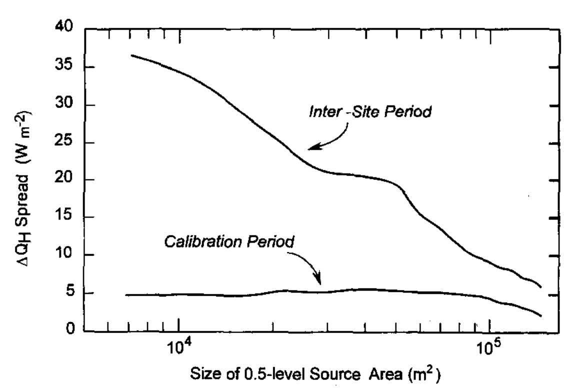

Fig. 6. Dependence of the spatial variability of sensible heat flux (ΔQ_H Spread) on the size of the flux source area. With increasing source area the measurements of sensible heat flux are representative of a larger area and their spatial variability is reduced. This effect is absent when two instruments mounted on the same tower are compared during a calibration period (modified from Schmid et al., 1991). The term 'ΔQ_H spread' denotes the variation (with source area size) of the median of the Q_H-difference distribution between the two sensors (see Schmid, 1988).

virtually absent in the calibration period. This latter result comes as no surprise: during the calibration period the two sensors were mounted close together and were thus affected by the same source area. However, the dependence of the measurable spatial variability on the size of the source area during the inter-site period is clearly demonstrated by the upper curve in Fig. 6. When the source area is small, the individual measurements may not be representative of the dominant morphological scales of the urban surface and are subject to considerable spatial variability. Larger source areas correspond to flow conditions with a more 'efficient spatial averaging power'. Thus, the flux measurements are representative of a large area, leaving less room for variation.

Thus, with small source areas (i.e. in strongly unstable conditions) the blending height is high, and the surface layer is inhomogeneous up to more than 30 m. With less buoyancy production, the 'field of view' of the instruments is increased and the measured values are representative of a larger area: depending on its definition, the blending height has then descended to below 30 m and the measurements at 30 m may be accepted as regional averages.

6. Summary and conclusions

The overriding principle of experimental design put forward in this paper is that the scale of the observations be matched to the scale of the phenomenon they are aimed at. If the aim is to obtain an estimate of the surface flux of some scalar quantity over an inhomogeneous area, the spatial scale at which it is wanted needs to be defined: the problem is to obtain either the surface flux over a single patch, or an aggregated flux value over a region of several patches.

Commonly, the measurements that are used to infer such surface flux estimates are conducted at height, with instruments mounted on a mast. Owing to the combination of horizontal advection and vertical turbulent diffusion (or the transfer geometry for radiative fluxes), such instruments are exposed to the surface influence not only of one point, but of a potentially large area: the surface source area. Thus every micrometeorological point measurement represents some kind of spatial average of surface conditions. The scale of this spatial average, and thus of the observation, depends on the type of quantities involved in the measurements (radiation, scalar flux, scalar concentration profile), the measurement height, stability and the intensity of cross-wind turbulence. This scale can be determined by footprint models or source area models designed for the relevant method of measurement.

To match the scale of the observation to the scale at which the flux is wanted, it is convenient to define a criterion for the point-to-area representativeness of the measurements involved in the determination of the flux. Three different (but related) versions of representativeness criteria are presented in this paper, using the source weight distribution or footprint function of the measurements or its integrated form, the source area model. The three methods to examine point-to-area representativeness differ also in their rigour and the amount of effort required for their application. The most pragmatic version, the source area case examination method, is described in some detail, and its use is demonstrated by examples in Section 5.

With all three methods discussed in this study, the position of the instrument mast, the height of sensor deployment, and the choice of measurement method can be optimised by maximising the degree of point-to-area representativeness. The examples presented demonstrate clearly that: (1) the spatial context of micrometeorological flux observations depends on the method of measurement (eddy correlation or accumulation, profile method, or Bowen ratio–energy balance combination); (2) eddy correlation (or eddy accumulation) equipment should be used whenever possible over inhomogeneous areas, as the spatial context of this method is most clearly defined; (3) for energy balance studies, care needs to be taken to avoid a mismatch between radiation and turbulent flux source areas; (4) application of footprint or source area modelling is useful to organise the spatial variability in flux data and to estimate the physical blending height.

In closing, it is important to add some cautionary remarks and to point out some of the limitations of the models and methods discussed in this work. Source area or footprint modelling relies on the assumption that the mean flow and turbulent mixing are stationary and homogeneous, even over areas where the distribution of passive scalar sources is not. However, it is a common finding that variations in vegetation or other surface cover types that control the emission or uptake of scalars are usually associated with changes of roughness and geometry. Thus, the validity of the above assumption needs to be carefully examined with every application of these methods. In addition, the theoretical basis of these models is restricted to the surface layer, and thus they should not be used for measurement heights above the surface layer, or in free convection conditions, where mixed layer and entrainment effects must be expected to influence a flux measurement. The source area models used in the present work are available over the world-wide-web at the following URL-address:

http://www.indiana.edu/~climate/SAM/SAM_FSAM.html

It should again be pointed out that the discussion of scalar concentration footprints and source areas in conjunction with profile measurements refers only to the area of influence of the concentration measurements themselves and not to the flux inferred from them. A model for the footprint of a flux inferred by profile measurements is presented in forthcoming work by Horst (1997).

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Appendix A. Specific formulations of radiation footprint function and source areas

The geometry of a surface radiation measurement (i.e. the upwelling part of a net radiation measurement) is illustrated in Fig. 7. A differential surface area, dA, is seen by the radiometer fixed at height z_m, at a nadir angle of θ. The radius, r, is the horizontal distance of dA from the nadir point, whereas $R\ (=\sqrt{r^2+z_m^2})$ is the distance from the radiometer. The azimuthal angle around the nadir axis is denoted as ϕ.

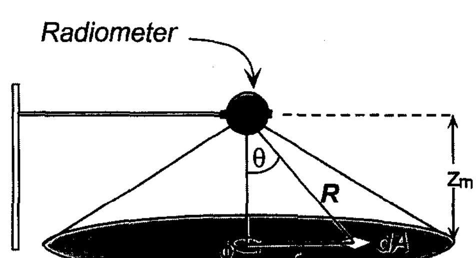

Fig. 7. The geometrical arrangement of a radiometer above a flat, horizontal surface. (Refer to the text for definitions of the geometrical elements.)

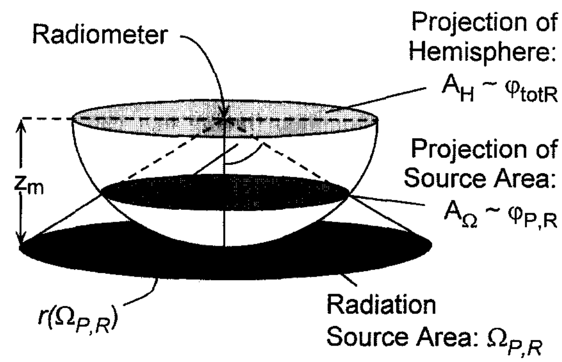

Fig. 8. The geometrical derivation of the radiation source area (Eq. (A8)). The source area level P is the ratio of the projection of the source area onto the lower hemisphere and then onto the radiometer plane (A_Ω) to the projection of the entire hemisphere onto the radiometer plane (A_H).

Lambert's cosine law of radiation may be expressed as

$$\mathrm{d}F_R = I_{n,u} \cos\theta \,\mathrm{d}\omega \tag{A1}$$

where $I_{n,u}$ is the normal component of the unit radiation intensity originating from $\mathrm{d}A$, and $\mathrm{d}\omega$ is a differential solid angle, with its centre located θ away from the nadir. If $\mathrm{d}\omega$ is the solid angle at which $\mathrm{d}A$ is seen from the centre of the radiometer plate, $\mathrm{d}\omega$ is expressed in terms of $\mathrm{d}A$, θ, and z_m as

$$\mathrm{d}\omega = \mathrm{d}A \frac{\cos^3\theta}{z_m^2} \tag{A2}$$

Substitution of Eq. (A2) into Eq. (A1) gives

$$\mathrm{d}F_R = I_{n,u} \frac{\cos^4\theta}{z_m^2} \mathrm{d}A \tag{A3}$$

and, as $\cos^2\theta = [1 + (r/z_m)^2]^{-1}$, the radiation footprint function, f_R, follows as

$$f_R = \frac{\mathrm{d}F_R}{\mathrm{d}A \cdot I_{n,u}} = \left(z_m + \frac{r^2}{z_m} \right)^{-2} \tag{A4}$$

To compute the radiation source area of a given level, it is convenient to take a slightly different approach, following Schwerdtfeger (1976). He gave the contribution to a radiative flux measurement, from a differential annulus on the surface with radius r around the nadir axis, in polar coordinates, as

$$\mathrm{d}F_R = 2\pi I_{n,u} \sin\theta \cos\theta \,\mathrm{d}\theta \tag{A5}$$

where θ is the nadir angle of the annulus. The total flux from the entire (plane) surface is given by integrating Eq. (A5) over the lower hemisphere, so that $F_{R,tot} = \pi I_{n,u}$. The total integrated source weight for radiation follows as $\varphi_{tot,R} = F_{R,tot}/I_{n,u} = \pi$. According to Lambert's law and Eq. (A4), the source weight function isopleths for radiative transfer are concentric rings around the nadir axis for a given sensor height (owing to its cylindrical symmetry, the azimuthal distribution is uniform and thus irrelevant). It follows that a source area of level P is defined by the radius $r(\Omega_{P,R})$, and $\varphi_{P,R}$ is found by integrating Eq. (A5) to an angle corresponding to $r(\Omega_{P,R})$ and dividing by $I_{n,u}$:

$$\varphi_{P,R} = F_R(\Omega_{P,R})/I_{n,u} = 2\pi \int_0^{\theta(\Omega_{P,R})} \sin\theta \cos\theta \,\mathrm{d}\theta \tag{A6}$$

With $\theta(\Omega_{P,\mathrm{R}}) = \sin^{-1}[r(\Omega_{P,\mathrm{R}})/\sqrt{r(\Omega_{P,\mathrm{R}})^2 + z_\mathrm{m}^2}\,]$, the solution of Eq. (A6) is given by Schwerdtfeger (1976)

$$\varphi_{P,\mathrm{R}} = \pi \frac{r(\Omega_{P,\mathrm{R}})^2}{r(\Omega_{P,\mathrm{R}})^2 + z_\mathrm{m}^2} \tag{A7}$$

The source area fraction P_R for radiative transfer thus follows from the application of Eq. (A7) and φ_{totR} in Eq. (6) and is determined by the geometric configuration of the measurement alone:

$$P_\mathrm{R} = \frac{\varphi_{P,\mathrm{R}}}{\varphi_{\mathrm{tot},\mathrm{R}}} = \frac{r(\Omega_{P,\mathrm{R}})^2}{r(\Omega_{P,\mathrm{R}})^2 + z_\mathrm{m}^2} \tag{A8}$$

In this formulation P_R is equivalent to the view factor of the radiometer as discussed by Reifsnyder (1967).

Reifsnyder (1967) also provided an instructive geometric derivation of Eq. (A8), as illustrated in Fig. 8: the inverted radiometer 'sees' the entire ground-surface as projected to the lower hemisphere. However, by Lambert's law, the total energy received ($\varphi_{\mathrm{tot,R}}$) is proportional not to the surface of the hemisphere, but only to the area of its normal projection onto the radiometer plane (A_H in Fig. 8). Similarly, the energy received from a finite circular source area is proportional to its central projection onto the hemisphere, and then the normal projection onto the radiometer plane (A_Ω in Fig. 8). Thus, the source area level (or view factor) is defined by the radius of the source area, $r(\Omega_{P,\mathrm{R}})$, and by the sensor height, z_m, in Eq. (A8).

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