THE AIR DENSITY CORRECTION TO EDDY FLUX MEASUREMENTS

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Abstract. Under the usual assumptions for the atmospheric surface layer, we show that air density fluctuations, particularly those due to temperature fluctuations associated with a heat flux, result in a small mean vertical wind velocity. Because of this, there can be a significant correction to eddy flux measurements of passive scalars, for example CO_2, whose average concentration is very large compared to concentration fluctuations associated with the eddy flux.

Eddy flux methods have been used frequently in the past 15 years to measure wind stress, heat flux, and moisture flux by analysis of turbulent fluctuations at a point near the lower boundary of the atmosphere (for reviews see Pond, 1971; Kraus, 1972; Businger, 1975). These methods, which require fewer assumptions than do determinations of fluxes from profiles, have been particularly useful over the sea where the presence of waves makes it difficult if not impossible to measure the lower parts of velocity, temperature, or humidity profiles. With the development of fast, sensitive detectors (e.g., Jones *et al.*, 1978), the eddy flux method offers considerable potential for the direct measurement of fluxes of gases or of any passive scalar quantity between the ocean and atmosphere, as exemplified by a recent measurement of CO_2 fluxes (Jones and Smith, 1977). Although this paper deals primarily with examples of flux measurements over the sea, the conclusions are general and applicable to measurements over land. We do not, however, deal with effects of sloping or mountainous terrain.

The average vertical flux through a layer in the atmospheric surface layer of a passive scalar whose concentration is determined by a sensor which measures the partial density ρ_s (mass/vol.) is given by

$$F_s = \langle \rho_s w \rangle \tag{1}$$

where w is the vertical component of the wind velocity and the angle brackets denote time averaging. Following the usual practice, we can split the density and vertical wind velocity into mean values, $\langle \rho \rangle$ and W, and fluctuations, ρ' and w', about their mean values;

$$F_s = \langle \rho_s \rangle W + \langle \rho_s' w' \rangle. \tag{2}$$

This note describes corrections which must be applied to flux measurements when the ambient concentration of the passive scalar is much larger than the concentration fluctuations associated with the flux.

The usual model considered appropriate for the atmospheric surface layer assumes under steady-state conditions of temperature, humidity, and pressure that the mass flux is non-divergent and that the equation of continuity can be written as

$$\nabla \cdot (\rho \mathbf{v}) = 0 \tag{3}$$

where ρ is the density of air and $\mathbf{v}$ is the mean wind velocity. If also, as is usually assumed, the mean flow does not diverge horizontally nor accelerate downstream, equation (3) becomes

$$\frac{\partial}{\partial z} \langle \rho w \rangle = 0 \, . \tag{4}$$

If the surface is level and the rate of solution of air into the surface is negligible, equation (4) further reduces to

$$\langle \rho w \rangle = \langle \rho \rangle W + \langle \rho' w' \rangle = 0$$

or the *mean vertical wind* velocity is

$$W = -\langle \rho' w' \rangle / \langle \rho \rangle \, . \tag{5}$$

Thus, if, for example, updrafts tend to be less dense than downdrafts, a mean upward velocity is required to balance out a downward turbulent mass flux.

The magnitude of W is so small compared to the mean wind velocity U that it cannot be measured directly, particularly over the ocean where waves and current make it difficult to secure a stable platform. However, we can use equation (5) to estimate W from eddy flux measurements. The density of air is determined by the temperature, t, the pressure, p, and the humidity, ρ_w. The air density, ρ in SI units (kg m^{-3}) is

$$\rho = 3.485 \, p/t - 0.3033 \rho_w \, .$$

We represent $p(kPa)$ and t(°K) in terms of mean values, P and T, and fluctuations about their mean values, p' and t'. The air density fluctuation, ρ', is then

$$\rho' = -3.485 P \frac{t'}{T^2} - 0.3033 \rho'_w + 3.485 \frac{p'}{T}$$

where we have assumed the partial pressure of water vapour, P_w, is much less than P. Other terms could possibly be included in this expression for special situations. Thus from equation (5),

$$W = \langle t' w' \rangle / T + 0.087 \langle \rho'_w w' \rangle T/P - \langle p' w' \rangle / P \, . \tag{6}$$

This is a small correction to the usual assumption that $W \equiv 0$ over the ocean. For most sensors used in the eddy correlation technique to determine wind stress and heat flux, it is negligible. Of the terms included in equation (6), the 'pressure flux' term $\langle p'w' \rangle$ will have a relatively small effect on the air density and so this term can be neglected (Elliott, 1972).

One can estimate the contribution of the remaining two terms using bulk parameterization formulas with typical values which may be encountered (e.g., Smith, 1974). Jones and Smith (1977) reported heat and CO_2 flux measurements, wind speed of 7.4 m s^{-1}, water and air temperatures of 12.7 and 14.8 °C, respectively, during a period of high humidity. From that data and equation (6), $W = -3.8 \times 10^{-5}$ m s^{-1}, and the corrected CO_2 flux values are about 30% less than the uncorrected values. These measurements were made in early summer when a rising water temperature was expected to cause a relatively large upward CO_2 flux. Under similar conditions but with unusually dry air of 50% relative humidity at 10-m height and with an evaporation coefficient $C_E = 1.2 \times 10^{-3}$, the rate of evaporation would be

$$\langle \rho'_w w' \rangle = C_E U \Delta\rho_w = 3.9 \times 10^{-5}\ \mathrm{kg\ m^{-2}\ s^{-1}}$$

which from equation (6) adds 1.0×10^{-5} m s^{-1} to the vertical wind velocity, W. In temperate or arctic conditions, one would thus expect W to be determined primarily by the heat flux. In tropical conditions $\Delta\rho_w$ can be much larger and the evaporation term may dominate. The correction $\langle \rho_s \rangle W$ can be fairly large, and under many circumstances may even be larger in magnitude than $\langle \rho'_s w' \rangle$.

In the usual steady-state, non-divergent, non-accelerating boundary layer, the heat flux and evaporation are nearly *constant with height* and therefore so also is W. If the mean flow is horizontally divergent or accelerating, then the mean vertical wind must vary with height, starting with zero value just above a hypothetical viscous sublayer which typically would be a fraction of a millimetre thick. Such height-dependent vertical velocities might in some circumstances be comparable to W calculated from equation (6) at a typical measuring height of 10 m. However, the flux of a passive scalar *at the surface* where the height-dependent mean vertical velocity vanishes should *not* be affected by these terms. Similarly, vertical velocities at the measurement height associated with mean flow divergences caused by diurnal or synoptic changes in mean temperature, humidity, or pressure do not form a part of the surface flux.

The situations for which one must include W in the determination of a flux of a passive scalar can be determined from equation (2). Roughly, since in our example $W \approx 10^{-4} w'$, the correction must be applied unless $\rho'_s \gg 2 \times 10^{-4} \langle \rho_s \rangle$, assuming a correlation of the order of 0.5 between ρ'_s and w'. The correction certainly cannot be ignored for CO_2, the only gas flux between the ocean and atmosphere yet measured by the eddy correlation method. It would almost certainly apply also to the other relatively abundant atmospheric gases such as N_2, O_2 and Ar whenever the sensor used in an experiment measures the partial density, ρ_s, in the atmosphere.

The correction applies in principle for water vapour flux measurements. In the example presented earlier, $\langle \rho_w \rangle W = 0.2 \times 10^{-6}$ kg m^{-2} s^{-1}, only about 0.6% of the total flux. However if the air is humid and the temperature difference is large, this correction can become more important.

Finally, let us consider a class of sensors which measures a relative concentration, c, of a passive scalar in the atmosphere rather than ρ_s such that

$$\rho_s = c\rho\,.$$

As before, we can represent these in terms of their mean values $\langle\rho_s\rangle$, C, and $\langle\rho\rangle$, and fluctuations about their mean values, ρ_s', c', and ρ', so that

$$\rho_s = C\langle\rho\rangle + C\rho' + c'\langle\rho\rangle + c'\rho'\,.$$

The average vertical flux is (Bakan, 1978)

$$F_s = W(C\langle\rho\rangle + \langle c'\rho'\rangle) + C\langle\rho' w'\rangle + \langle\rho\rangle\langle c' w'\rangle$$

where terms involving only one fluctuating quantity are by definition equal to zero after averaging and a term which is of third order in fluctuating quantities has been neglected. Substituting W from equation (4) and neglecting a term fourth order in fluctuating quantities gives simply

$$F_s = \langle\rho\rangle\langle c' w'\rangle\,.$$

Thus to second order, fluctuations in air density do *not* contribute to a flux of a passive scalar whose concentration is determined by a sensor which measures the relative concentration, c, rather than the absolute concentration, ρ_s. Such a sensor would have the considerable advantage of eliminating the requirement for separate determinations of heat and moisture fluxes, thus simplifying measurements of F_s and perhaps, since the density fluctuation corrections can be large, leading to more accurate determinations of F_s.

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