DRY-AIR BOUNDARY CONDITIONS FOR CORRECTION OF EDDY FLUX MEASUREMENTS

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Abstract. For measurements of eddy fluxes in the atmospheric boundary layer of gases (such as CO_2) whose average concentration is very large compared to the fluctuations, corrections for air density fluctuations are required. With the boundary condition of no flux of *dry* air at the surface, the evaporation correction to eddy fluxes is 2.6 times larger than has been estimated with the boundary condition of no mass flux at all at the surface. The heat flux correction is also increased by a few per cent.

1. Introduction

Bakan (1978) and Jones and Smith (1978) have shown that in the measurement of the average vertical flux of a gas in the atmospheric boundary layer, it is necessary to consider a mean vertical velocity associated with other fluxes. The average vertical gas flux is

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

where the partial density ρ_s and the vertical velocity w are split into mean and fluctuating parts $\langle \rho_s \rangle$ and ρ_s' and W and w', respectively. The first term on the right-hand side of Equation (1) is the 'eddy flux' and the second term is the 'correction', assuming that the fluctuating and mean partial densities are measured. Both Bakan (1978) and Jones and Smith (1978) took the boundary condition at the bottom of the atmosphere to be no flux of air, i.e., $\langle \rho w \rangle = 0$, where ρ is the total density of air. Webb and Pearman (1977) have taken the boundary condition to be no flux of *dry* air at the surface and throughout a constant-flux layer, i.e.,

$$\langle \rho_a w \rangle = 0 \tag{2}$$

where ρ_a is the partial density of dry air. This allows a flux of water vapour at the surface, and we acknowledge this to be a more realistic boundary condition.

2. Correction of Eddy Fluxes Derived from Fluctuations of Partial Density

We shall see that the choice of boundary condition has a large effect on the evaporation correction to eddy fluxes. From Equations 1 and 2, it follows with ρ_a split into mean and fluctuating parts, $\langle \rho_a \rangle$ and ρ_a', that a mean vertical velocity in the constant-flux layer is required to balance a mass flux of dry air due to density

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fluctuations

$$W = -\langle \rho'_a w' \rangle / \langle \rho_a \rangle \,. \tag{3}$$

We now obtain an expression for ρ'_a and $\langle \rho_a \rangle$.

Dalton's gas law relates the total density ρ to the partial densities of dry air and of water vapour, ρ_w, as follows:

$$\frac{\rho}{m} = \frac{\rho_a}{m_a} + \frac{\rho_w}{m_w} + \frac{\rho_s}{m_s} = \frac{p}{Rt} \tag{4}$$

where m, m_a, m_w, and m_s are the molecular weights of moist air, dry air (0.029 kg), water vapour (0.018 kg), and where ρ_s and m_s are the density and molecular weight of the passive scalar which we wish to consider. The gas constant is R, the pressure p, and the absolute temperature t. To calculate W we shall consider ρ_s to be negligible compared to other terms in Equation (4), an assumption quite justifiable for a low-concentration gas such as CO_2. With temperature, pressure, and densities split into mean and fluctuating parts, $t = T + t'$, $p = P + p'$, and $\rho = \langle \rho \rangle + \rho'$, Dalton's gas law becomes

$$\frac{\langle \rho_a \rangle + \rho'_a}{m_a} + \frac{\langle \rho_w \rangle + \rho'_w}{m_w} = \frac{P + p'}{R(T + t')} \simeq \frac{P + p'}{RT}\left(1 - \frac{t'}{T}\right) \tag{5}$$

since $t' \ll T$ in any reasonable situation. The fluctuating part of this equation (neglecting second-order terms) gives

$$\frac{\rho'_a}{m_a} = -\frac{\rho'_w}{m_w} + \frac{p'}{RT} - \frac{Pt'}{RT^2} \tag{6}$$

and within the substitution $P/RT = \langle \rho \rangle / m$ from Equation (4), we can write an expression for the dry-air density fluctuation caused by fluctuations of temperature, water vapour, and pressure,

$$\rho'_a = -\frac{m_a \langle \rho \rangle t'}{Tm} - \frac{m_a \rho'_w}{m_w} + \frac{m_a p'}{RT} \,. \tag{7}$$

The second term on the right-hand side represents dry air displaced by water vapour, while the equivalent term used by Jones and Smith (1978), $(m_w - m_a)\rho'_w/m_w$, allowed only for the lighter weight of moist air. From the mean part of Equation (4),

$$\langle \rho_a \rangle = \langle \rho \rangle \frac{m_a}{m}\left[1 - \frac{m}{m_w}\frac{\langle \rho_w \rangle}{\langle \rho \rangle}\right] \tag{8}$$

and substituting for $\langle \rho_a \rangle$ and ρ'_a in Equation (3),

$$W = \frac{\langle t'w' \rangle}{T\left[1 - \frac{m}{m_w}\frac{\langle \rho_w \rangle}{\langle \rho \rangle}\right]} + \frac{(m/m_w)\langle \rho'_w w' \rangle}{\langle \rho \rangle\left[1 - \frac{m}{m_w}\frac{\langle \rho_w \rangle}{\langle \rho \rangle}\right]} - \frac{\langle p'\omega' \rangle}{P\left[1 - \frac{m}{m_w}\frac{\langle \rho_w \rangle}{\langle \rho \rangle}\right]} \,. \tag{9}$$

The mean vertical velocity is now expressed in terms which may be measured in micrometeorological experiments. Substitution of W from Equation (9) into Equation (1) gives the flux correction. Webb and Pearman (1977) give the evaporation correction (second term in Equation (9)) for eddy fluxes and a similar correction for gas fluxes obtained from vertical profiles of mean concentration.

The heat flux correction differs from the corresponding term of Jones and Smith (1978, Equation (6)) by the denominator in square brackets, which increases the correction at most by a few per cent, particularly in hot, humid conditions. The evaporation correction, in addition to this denominator, differs by a factor $m_a/(m_a - m_w) = 2.6$ and so it is much more important than we realized previously. In general, both heat flux and evaporation will have to be known to calculate atmospheric boundary-layer fluxes of gases such as CO_2.

3. An Example

The contributions of the terms on the right-hand side of Equation (9) can be evaluated for the mean of three heat and CO_2 flux measurements reported by Jones and Smith (1977), also used as an example by Jones and Smith (1978). Suppose that at a height of 10 m the wind speed $U_{10} = 7.4\ \mathrm{m\,s^{-1}}$, the air temperature $T_a = 14.8\ °\mathrm{C}$, the water temperature $T_s = 12.7\ °\mathrm{C}$, and the water vapour density $\rho_w = 0.0111\ \mathrm{kg\,m^{-3}}$ corresponding to 87% relative humidity at a 10-m height and saturation at the surface water temperature. The value for $\langle t'w' \rangle = -0.011$ is an average of the three measurements reported by Jones and Smith (1977). Under these conditions, the evaporation term is negligibly small, and from Equation (9), we get

$$W = -0.011/(12.7 + 273.2)[1 - \frac{29}{18} \times \frac{0.0111}{1.23}] = -3.9 \times 10^{-5}\ \mathrm{m\,s^{-1}}. \quad (10)$$

Substituting in Equation (1), we obtain the corrected CO_2 flux as reported by Jones and Smith (1977) to be 30% less than the uncorrected flux. This new correction differs from the one applied previously only by the denominator in the square brackets, $[0.985]^{-1}$, a negligible change under the conditions of those measurements. However, if as in the second example used by Jones and Smith (1978), we suppose that the air at 10-m height was reduced to an unusually dry 50% relative humidity ($\rho_w = 0.0063\ \mathrm{kg\,m^{-3}}$) with other conditions remaining the same, then the second term in Equation (9) must also be included. The rate of evaporation can be estimated using an evaporation coefficient $C_E = 10^{-3}$ (Kondo, 1975),

$$\begin{aligned} \langle \rho'_w w' \rangle &= C_E(\rho_{ws} - \rho_w)U_{10} \\ &= 10^{-3}(0.0111 - 0.0063)7.4 \\ &= 3.6 \times 10^{-5}\ \mathrm{kg\,m^{-2}\,s^{-1}} \end{aligned} \quad (11)$$

where ρ_{ws} is the saturation water vapour density at the surface. Using the first and second terms in Equation (9), we get

$$W = -3.9 \times 10^{-3} + \frac{(29/18)3.6 \times 10^{-5}}{1.23[0.99]} = (-3.9 + 4.7) \times 10^{-5}\ \mathrm{m\,s^{-1}} \qquad (12)$$

and the evaporation correction has become larger than the heat flux correction. For warmer waters, which evaporate much more rapidly, the evaporation correction could be much larger than the heat flux correction, as envisaged by Webb and Pearman (1977), but in general both these terms should be considered.

From measurements over land at similar wind speeds (Elliott, 1972; Table 3) the third term in Equation (9) amounts to $\langle p'w' \rangle / P \simeq 0.6 \times 10^{-10}$ m s^{-1} and can safely be neglected in this case and in general.

4. Correction of Water Vapour Flux Measurements

The same corrections apply to measurements of evaporation E, as pointed out by Webb and Pearman (1977), in which case $\rho_s = \rho_w$ and Equations (1) and (9) simplify to

$$E = \langle \rho'_w w' \rangle + \langle \rho_w \rangle W = \frac{\langle \rho'_w w' \rangle}{\left[1 - \dfrac{m}{m_w} \dfrac{\langle \rho_w \rangle}{\langle \rho \rangle}\right]} + \frac{\langle \rho_w \rangle \langle t' w' \rangle}{T\left[1 - \dfrac{m}{m_w} \dfrac{\langle \rho_w \rangle}{\langle \rho \rangle}\right]} . \qquad (13)$$

The measured water vapour flux is increased typically by a few per cent, using the denominator term in square brackets. From the form of Equation 10, we can see that had we not neglected ρ_s after Equation (4), there would be additional denominator terms $[1 - (m \langle \rho_s \rangle)/(m_s \langle \rho \rangle)]$ in Equations (9) and (10). These would be significant if $\langle \rho_s \rangle$ were not small, and would be important in correction of flux measurements of relatively abundant gases such as oxygen. In the example given above, the heat flux correction to the evaporation (last term in Equation (10)) is less than 1% of the flux, but this term may become more significant in humid, tropical conditions with larger values of $\langle \rho_w \rangle$.

5. Correction of Gas Fluxes with Sensors of Relative Concentration

The above analysis applies to eddy fluxes computed from measured fluctuations in partial density (mass per unit volume). Jones and Smith (1978) and also Brook (1978), using a 'zero mass flux' boundary condition, pointed out that no gas flux corrections would be required if relative concentration (mass of gas or water vapour per unit mass of air) could be measured. Similarly, no gas flux correction would be required with the 'dry air' boundary condition if a sensor were devised which measured relative concentration per unit mass of dry air. In the more probable case of a hypothetical sensor measuring relative concentration c per unit mass of moist air with mean and fluctuating parts C and c', the appropriate flux correction is due only to the difference between the two boundary conditions. Thus the required correction to a measure flux $\langle\rho\rangle\langle c'w'\rangle$ is the difference between the correction using Equations 1 and 9 here and the equivalent correction from Jones and Smith (1978). Neglecting denominator terms in square brackets in Equation (9) ($m \simeq m_a$), this amounts to a fraction $m_w/(m_a - m_w) = 62\%$ of the evaporation correction from Equations (1) and (9) for a partial density sensor,

$$F_s \simeq \langle\rho\rangle\langle c'w'\rangle + \langle\rho\rangle C\langle\rho'_w w'\rangle \tag{14}$$

and such a sensor would have the advantage of not requiring a heat flux correction, although the need for an evaporation correction remains.

Brook (1978) has pointed out that there is also a sensible heat flux, equal to about 10% of the latent heat flux, associated with humidity dependence of the specific heat of air. This flux correction is in addition to and independent of the gas flux corrections discussed here.

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