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AN ALTERNATIVE APPROACH FOR CO_2 FLUX CORRECTION CAUSED BY HEAT AND WATER VAPOUR TRANSFER

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Abstract. Energy and CO_2 fluxes are commonly measured above plant canopies using an eddy covariance system that consists of a three-dimensional sonic anemometer and an H_2O/CO_2 infrared gas analyzer. By assuming that the dry air is conserved and inducing mean vertical velocity, Webb et al. (*Quart. J. Roy. Meteorol. Soc.* **106**, 85–100, 1980) obtained two equations to account for density effects due to heat and water vapour transfer on H_2O/CO_2 fluxes. In this paper, directly starting with physical consideration of air-parcel expansion/compression, the author derives two alternative equations to correct for these effects that do not require the assumption that dry air is conserved and the use of the mean vertical velocity. The author then applied these equations to eddy flux observations from a black spruce forest in interior Alaska during the summer of 2002. In this ecosystem, the equations developed here led to increased estimates of CO_2 uptake by the vegetation during the day (up to about 20%), and decreased estimates of CO_2 respiration by the ecosystem during the night (approximately 4%) as compared with estimates obtained using the Webb et al. approach.

Keywords: Air-parcel expansion/compression, Carbon flux, Eddy covariance, Flux correction, Open path CO_2/H_2O infrared gas analyzers.

1. Introduction

Eddy covariance systems are widely used for the measurement of vertical turbulent fluxes of energy, water and CO_2 over various terrestrial ecosystems (e.g., Aubinut et al., 2000; Baldocchi et al., 2001). Such an eddy covariance system usually consists of a three-dimensional sonic anemometer that measures fluctuations of wind speed in three directions and an infrared gas analyzer (IRGA) that measures fluctuations of densities of CO_2 and water vapour. The effect of H_2O/CO_2 density fluctuations arising from heat and water transfer is critical for the interpretation of fluxes made from both open- and closed-path eddy flux systems (Webb and Pearman, 1977; Baken, 1978;

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Jones and Smith, 1978; Smith and Jones, 1979; Webb et al., 1980; Kramm et al., 1995; Paw U et al., 2000; Fuehrer and Friehe, 2002; Massman and Lee, 2002). The basic consideration is that expansion/compression of the total moist air, occurring in the surface layer due to heat and water vapour transfer, leads to constituent density variations. Substantially, this turbulent exchange in the surface layer would lead to a compensatory mean vertical flux of the moist air as a result of air-parcel expansion/compression (Webb et al., 1980; Paw U et al., 2000). Numerous studies have investigated the influence of the mean vertical flux of the moist air on the measurements of vertical fluxes of scalars (e.g., water vapour or CO_2) in the surface layer.

One of the key past issues in developing a measurement approach for scalar fluxes including water vapour and CO_2 (and assuming steady state and horizontal homogeneity), was the estimation of the mean vertical velocity. These initial calculations were sensitive to assumptions about atmospheric processes in the surface layer. For example, Baken (1978) and Jones and Smith (1978) assumed that the mean vertical flux of moist air is zero (i.e., $\overline{\rho w} = 0$, where ρ is the density of moist air and w is the vertical velocity). In contrast, Webb and Pearman (1977) used the assumption that the mean vertical flux of dry air is zero (i.e., $\overline{\rho_a w} = 0$, where ρ_a is the density of dry air). The latter assumption led to the widely-accepted expressions by Webb et al. (1980, hereafter WPL) accounting for the effects of expansion/compression of the dry air (i.e., ignoring of the water vapour part in the total moist air) on the measurement of CO_2 and water vapour fluxes. This treatment of the mean vertical velocity in Webb et al. (1980) (by assuming that the dry air is conserved) may be unrealistic for many ecosystems that have substantial latent heat fluxes since it implies the mean vertical velocity results only from the expansion/compression of the dry air (not the total moist air) in the ascending/descending moist air parcels. This assumption is more likely to be true over extremely dry surfaces like deserts where latent heat fluxes are small and thus there is little water vapour contribution to the air-parcel expansion/compression.

Water vapour plays an important part in air-mass balance and exchange in the surface layer over most terrestrial ecosystems. The assumption that the flux of dry air is zero may introduce some biases by ignoring the contribution of water vapour in the air mass exchange, especially in regions with high humidity. Although the WPL correction has been validated experimentally by measuring CO_2 flux over a bare and dry underlying surface with no concurrent flux of CO_2 during the daytime (Leuning et al., 1982), or over a parking lot with very small CO_2 efflux (Ham and Heilman, 2003), the influence of water vapour was still not verified over such dry surfaces, in contrast to most current eddy covariance experiments that typically sample over transpiring plant canopies. On the other hand, the Leuning et al. (1982) and Ham and Heilman (2003) experiments demonstrate that the WPL scheme may be a good approximation over such dry surfaces as deserts and parking lots. In the past, the assumption that the dry air is conserved has caused debate (Leuning et al., 1982; Kramm et al., 1995; Fuehrer and Friehe, 2002).

Recently, another issue has been raised by Paw U et al. (2000) and Massman and Lee (2002) concerning the need to use the mean vertical velocity. For some, the mean vertical velocity could be the combined product of different mechanisms such as air-parcel expansion/compression, local circulation, mesoscale processes and even synoptic scale phenomena. Although this mean vertical velocity could be accurately measured, the direct use of this measured vertical velocity could lead to unrealistically large corrections in some cases compared to the WPL correction (Paw U et al., 2000).

Here, we propose a series of density corrections that do not require these assumptions (i.e., that the dry air is conserved as well as the introduction of a mean vertical velocity). Specifically, starting directly from the physical consideration of air-parcel expansion/compression of the total moist air, we derive two equations in Section 2 to explain the influence of air-parcel expansion/compression on the densities of water vapour and CO_2 when open-path infrared gas analyzers are used to obtain CO_2 and water vapour fluxes. We then compare water vapour and CO_2 fluxes from a black spruce forest ecosystem derived from our equations with those obtained from the WPL correction in Section 4. Some possible reasons for the difference in the magnitudes of corrections between two methods are then discussed. In Appendix A, we provide a three-dimensional continuity equation for CO_2 by defining the mixing ratio relative to the total moist air, thereby resulting in our proposed equations.

2. Theoretical Considerations

Before sunrise overland, the atmosphere is often stable and an inversion layer is usually present in the surface layer (eddy covariance measurements are usually made within this layer). After sunrise, surface temperature begins to increase due to the heating from solar radiation, and the surface then heats the adjacent air just above it, leading to the formation of thermal plume (or air parcels). Triggered by mechanically-generated turbulence and driven by buoyancy forces, these thermal plumes move upward with their temperature decreasing at the adiabatic lapse rate. As these plumes rise, their temperature remains higher than the ambient air temperature, leading to positive fluctuations in temperature as measured by fast-response thermometers in eddy covariance systems. Because of the difference in the magnitudes of physical quantities (e.g., temperature T and pressure P) between the plumes and the surrounding air, expansion of the warmer air parcel occurs.

We assume that the ambient atmospheric pressure is constant in the surface layer (at least up to the height of eddy covariance measurements), and that the surface pressure fluctuation, as compared with the water vapour fluctuation, is negligibly small (Webb et al., 1980; Paw U et al., 2000). Under isobaric conditions, the positive perturbation in temperature (i.e., $T' > 0$) will lead to volume expansion, while the negative fluctuation in temperature (i.e., $T' < 0$) will lead to compression. This expansion and compression continually occurs as the temperature varies as the result of the upward and downward motion of plumes. Thus, the densities of water vapour and CO_2 vary with expansion and compression, and implies that, when expansion (compression) occurs, the measured densities of water vapour and CO_2 are lower (higher) than they were at the surface due to the influence of expansion (compression) in the surface layer. Consequently, the measured water vapour and CO_2 fluxes are lower than are expected after Reynolds averaging over a period (e.g., half-an-hour) when the atmosphere is unstable.

Following Webb et al. (1980), applying the ideal gas law to dry air, water vapour and moist air, respectively, we have

$$P_a = \frac{\rho_a}{m_a} RT, \tag{1}$$

$$P_v = \frac{\rho_v}{m_v} RT, \tag{2}$$

$$P = \frac{\rho}{m} RT, \tag{3}$$

where P_a and P_v are the partial pressures of dry air and water vapour, respectively; P is the total atmospheric pressure; ρ_a, ρ_v and ρ are the respective densities of dry air, water vapour, and moist air ($\rho = \rho_a + \rho_v$); m_a, m_v and m are the respective molecular masses of dry air, water vapour, and moist air; R is the universal gas constant; and T is the absolute temperature.

Using a simple Taylor series expansion, and ignoring the higher-order terms, we obtain the mean part and perturbation components from Equations (1), (2) and (3),

$$\frac{\rho_a'}{m_a} + \frac{\rho_v'}{m_v} = -\frac{P}{R\bar{T}^2} T', \tag{4}$$

$$\frac{\bar{\rho}_a}{m_a} + \frac{\bar{\rho}_v}{m_v} = \frac{P}{R\bar{T}}, \tag{5}$$

and after re-arrangement of Equations (4) and (5), we obtain

$$\rho_a' = -\frac{m_a}{m_v}\rho_v' - (\bar{\rho}_a + \frac{m_a}{m_v}\bar{\rho}_v)\frac{T'}{\bar{T}}. \tag{6}$$

Note that Equation (6) is the same as Equation (9b) in Webb et al. (1980). It is assumed that the temperature fluctuations as measured by the sonic anemometer or fast thermocouple or thermistor are representative of the temperature fluctuations that occur in the path of the IRGA. For given constituents (e.g., water vapour and CO_2) in the air volume, the mass of water vapour and CO_2 in the volume is conservative. It is reasonable to assume that the variation in the density of water vapour in the air parcel is caused by the variation of its volume due to air-parcel expansion/compression of the total moist air. When this idea is applied for the total density of air, we obtain the relationship between the variation of the total density of air and the variation of the air-parcel volume,

$$\frac{V'}{\bar{V}} = -\frac{1}{\bar{\rho}}\rho', \tag{7}$$

where V' is a perturbation in specific volume, and $\rho' = \rho'_a + \rho'_v$, so that the variation in the total density of a moist air parcel caused by a perturbation in water vapour and temperature can be obtained from Equation (6),

$$\rho' = \rho'_a + \rho'_v = \left(1 - \frac{m_a}{m_v}\right)\rho'_v - \left(\bar{\rho}_a + \frac{m_a}{m_v}\bar{\rho}_v\right)\frac{T'}{\bar{T}}. \tag{8}$$

Substitution of Equation (8) into Equation (7) yields

$$\frac{V'}{\bar{V}} = -\frac{1}{\bar{\rho}}\left\{\left(1 - \frac{m_a}{m_v}\right)\rho'_v - \left(\bar{\rho}_a + \frac{m_a}{m_v}\bar{\rho}_v\right)\frac{T'}{\bar{T}}\right\}. \tag{9}$$

For water vapour in the volume, the perturbation of its density caused by expansion/compression of the total moist air may be estimated by

$$\rho'_{H_2O\,exp} = -\bar{\rho}_v\frac{V'}{\bar{V}}. \tag{10}$$

Similarly, for CO_2 in the volume, the perturbation of its density caused by expansion/compression of the total moist air can also be obtained as

$$\rho'_{CO_2 exp} = -\bar{\rho}_c\frac{V'}{\bar{V}}. \tag{11}$$

Equation (10) with substitution for $V'/\bar{V}$ from Equation (9) gives

$$\rho'_{H_2Oexp} = -\frac{\bar{\rho}_v}{\bar{\rho}}\left\{\left(\frac{m_a}{m_v} - 1\right)\rho'_v - \left(\bar{\rho}_a + \frac{m_a}{m_v}\bar{\rho}_v\right)\frac{T'}{\bar{T}}\right\}. \tag{12}$$

and similarly, Equation (11), with substitution for $V'/\bar{V}$ from Equation (9), gives

$$\rho'_{CO_2 exp} = -\frac{\bar{\rho}_c}{\bar{\rho}}\left\{\left(\frac{m_a}{m_v} - 1\right)\rho'_v + \left(\bar{\rho}_a + \frac{m_a}{m_v}\bar{\rho}_v\right)\frac{T'}{\bar{T}}\right\}. \tag{13}$$

Since the measured density of water vapour by the IRGA (ρ'_v) at the height of eddy covariance measurements is the sum of the real density of water vapour (ρ'_{H_2O}) and the perturbed density caused by the volume expansion/compression ($\rho'_{H_2O\,exp}$), then $\rho'_v = \rho'_{H_2O} + \rho'_{H_2O\,exp}$. Therefore, we obtain

$$\rho'_{H_2O} = \rho_v{}' - \rho'_{H_2Oexp}$$
$$= \rho_v{}' + \frac{\bar{\rho}_v}{\bar{\rho}}\left\{\left(\frac{m_a}{m_v} - 1\right)\rho'_v + \left(\bar{\rho}_a + \frac{m_a}{m_v}\bar{\rho}_v\right)\frac{T'}{\bar{T}}\right\}. \tag{14}$$

Similarly, since the measured density of CO_2 by the IRGA (ρ'_c) at the height of eddy covariance measurements is the sum of the real density of CO_2 (ρ'_{CO_2}) and the perturbed density caused by the volume expansion/compression ($\rho'_{CO_2\,exp}$), then $\rho'_c = \rho'_{CO_2} + \rho'_{CO_2\,exp}$. Therefore,

$$\rho'_{CO_2} = \rho_c{}' - \rho'_{CO_2exp}$$
$$= \rho_c{}' + \frac{\bar{\rho}_c}{\bar{\rho}}\left\{\left(\frac{m_a}{m_v} - 1\right)\rho'_v + \left(\bar{\rho}_a + \frac{m_a}{m_v}\bar{\rho}_v\right)\frac{T'}{\bar{T}}\right\}. \tag{15}$$

Using Equations (14) and (15) multiplied by w′ and then applying Reynolds averaging, we obtain the water vapour flux (E) and the CO_2 flux (F_c) after correction,

$$E = \overline{w'\rho'_{H_2O}} = \overline{w'\rho_v{}'} + \overline{w'\rho'_{H_2Oexp}}$$
$$= \overline{w'\rho'_v}\left\{\frac{\bar{\rho}_v}{\bar{\rho}}\left(\frac{m_a}{m_v} - 1\right) + 1\right\} + \left\{\frac{\overline{\rho}_a}{\bar{\rho}}\overline{\rho_v}\left(1 + \frac{m_a}{m_v}\frac{\bar{\rho}_v}{\overline{\rho}_a}\right)\frac{\overline{w'T'}}{\bar{T}}\right\}, \tag{16}$$

$$F_c = \overline{w'\rho'_{CO_2}} = \overline{w'\rho_c{}'} + \overline{w'\rho'_{CO_2exp}}$$
$$= \overline{w'\rho'_c} + \frac{\bar{\rho}_c}{\bar{\rho}}\left(\frac{m_a}{m_v} - 1\right)\overline{w'\rho'_v} + \left\{\frac{\overline{\rho}_a}{\bar{\rho}}\overline{\rho_c}\left(1 + \frac{m_a}{m_v}\frac{\bar{\rho}_v}{\overline{\rho}_a}\right)\frac{\overline{w'T'}}{\bar{T}}\right\}. \tag{17}$$

Using the same notation as in Webb et al. (1980), Equations (16) and (17) can be rewritten as

$$E = \overline{w'\rho'_v}\left\{\frac{\bar{\rho}_v}{\bar{\rho}}(\mu - 1) + 1\right\} + \frac{\bar{\rho}_a}{\bar{\rho}}\bar{\rho}_v(1 + \mu\sigma)\frac{\overline{w'T'}}{\bar{T}}, \tag{18}$$

$$F = \overline{w'\rho'_c} + \frac{\bar{\rho}_c}{\bar{\rho}}(\mu - 1)\overline{w'\rho'_v} + \frac{\bar{\rho}_a}{\bar{\rho}}\bar{\rho}_c(1 + \mu\sigma)\frac{\overline{w'T'}}{\bar{T}}. \tag{19}$$

To compare our results with the WPL correction, we rewrite the equations of the WPL correction as

$$E_{WPL} = \overline{w'\rho'_v}(1 + \mu\sigma) + \bar{\rho}_v(1 + \mu\sigma)\frac{\overline{w'T'}}{\bar{T}}, \tag{20}$$

$$F_{cWPL} = \overline{w'\rho'_c} + \frac{\bar{\rho}_c}{\bar{\rho}_a}\mu\overline{w'\rho'_v} + \bar{\rho}_c(1 + \mu\sigma)\frac{\overline{w'T'}}{\bar{T}}, \tag{21}$$

where $\mu = m_a/m_v$ and $\sigma = \bar{\rho}_v/\bar{\rho}_a$ and m_a and m_v are the molecular mass of dry air and water vapour, respectively.

Although we start with the same assumptions as in Webb et al. (1980) during our derivation (with the exceptions that we do not assume that the dry air is conserved and that we do not use the mean vertical velocity), our final equations are different from those proposed by WPL. In Appendix A, we also derive a three-dimensional continuity equation for the eddy covariance measurement of CO_2 using the conservation equation for total moist air (instead of using the conservation equation for dry air that was first developed by Paw U et al. (2000) and Massman and Lee (2002)). Equation (A9) in Appendix A includes the term that is consistent with Equation (19). The consistency between two equations obtained from two different physical considerations implies that the expansion/compression of the total moist air (not the dry air only) should be taken into consideration to account for the density effects on measurements of water vapour and CO_2 fluxes in the surface layer.

3. Experimental Data

We used observations that were made at a black spruce forest near Delta Junction (63.9 °N, 145.74° W), located about 100 km to the south-east of Fairbanks, Alaska. This site was dominated by relatively homogeneous black spruce *(Picea mariana)* that was approximately 80 years old; the mean canopy height was approximately 4 m with occasional trees reaching up to 6 m, and the canopy coverage was about 60%. The terrain was generally flat on a glacial outwash in the Tanana River drainage of interior Alaska (Chambers and Chapin, 2002). From the micrometeorological tower there was an adequate fetch from all directions with the shortest fetch to the east (approximately 300 m). The dominant ground cover species were feather-moss (*Pleurozium schreberi* and *Rhytidium rugosum*) and lichen (*Cladonia* spp. and *Stereocaulon* spp.)" that reached a depth of 0.1–0.15 m.

Eddy covariance and microclimate measurements have been made at this site since September 2001. The eddy covariance system consisted of a three-dimensional sonic anemometer (CSAT3, Campbell Sci Ltd., Utah, U.S.A.) and an open-path IRGA (LiCor 7500, LICOR Inc., Lincoln, NB, U.S.A.) that were mounted at a height of 9.5 m. The sonic anemometer was used to measure wind velocity and temperature fluctuations. The IRGA was used to measure the fluctuations of water vapour and CO_2 density.

We also measured microclimate variables as 30-min averages of 1-s readings, including net radiation (REBS Q-7.1, Seattle, Washington, U.S.A.), incoming and reflected shortwave radiation (Precision Spectral Pyranometers, Eppley Laboratories, Providence, RI, U.S.A.), photosynthetic photon flux density (PPFD: LI190SB, LICOR Inc., Lincoln, NB, U.S.A.). Air temperature and relative humidity were measured at 10, 6 and 2 m with temperature/humidity probes (HMP45C, Vaisala, Helsinki, Finland). Wind speed and wind direction were measured at 12 m using a wind sentry (Model 03001, RM Young, Inc., U.S.A.) and wind speed was measured at 6 m using a cup anemometer (Model 03101, RM Young, Inc., U.S.A.). Additionally, thermocouples, soil moisture probes (CS615, Campbell Sci Ltd., Utah, U.S.A.) and soil heat flux plates (Model HFT3, REBS, Seattle, Washington, U.S.A.) were buried at 0–0.34 m below the surface in a variety of microenvironments ranging from mostly sunlit and mostly shaded sites. The 30-min heat storage in the soil above the soil heat flux plates was estimated from the combination method (Oke, 1987), using volume fractions of mineral, organic and water content of the surface soil for calculating surface heat capacity (Chambers and Chapin, 2002).

Note that, because the sonic temperature, which is approximately the virtual temperature, is used for obtaining sensible heat flux in our measurements, correction is needed to explain the influence of water vapour. Additionally, across-wind effects should be taken into account (Schotanus et al., 1983; Kaimal and Gaynor, 1991; Kaimal and Finnigan, 1994). For this purpose, the equations proposed by Liu et al. (2001) are used. For the purpose of comparison, we only used the data that were measured in the summer season from June 15 to August 31, 2002. For the growing season net ecosystem exchange (NEE) estimates, we used the data obtained from April 23 to October 21, 2002.

4. Results and Discussions

4.1. Energy and CO_2 fluxes over the black spruce forest

Evaluation of energy closure is accepted as an important factor in accessing data quality by eddy covariance system. Usually linear regression coefficients (slope and intercept) from ordinary least squares between (H + LE) and (Rn − G − S) are used to evaluate the energy balance closure, H being the sensible heat flux, LE the latent heat flux, with L the latent heat, Rn is net radiation, G is soil heat flux and S is the storage. However, because the data are subject to measurement error not only in the turbulent fluxes but also in the available energy, we used straight-line regression with errors in both coordinates for our purpose (Press et al., 1992). Using this method and the data for the summer season, the slope and intercept between (H + LE) and (Rn−G−S) are 0.87 and 4.1 W m^{-2} for the site ($r^2 = 0.89$).

Figure 1 shows the mean 30-min net radiation, and sensible and latent heat fluxes measured over the black spruce forest in Alaska from June 15 to August 31, 2002. The average midday net radiation was 380 $\pm$ 170 W m^{-2} with the peak value of 740 W m^{-2}. The average midday sensible heat flux was 165 $\pm$ 90 W m^{-2} with its peak value of 370 W m^{-2}, while the average midday

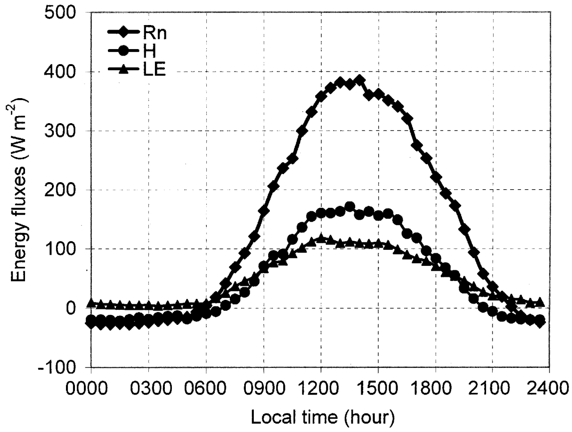

Figure 1. Mean 30-min surface energy fluxes from a black spruce forest in interior Alaska during the period from June 15 to August 31, 2002. Net radiation (Rn) is denoted by diamonds; sensible heat flux (H) by circles, and latent heat flux (LE) after the WPL correction by triangles. Data from this same site and time period are presented in subsequent Figures 2–7.

latent heat flux approached 110 ± 45 W m^{-2} with its peak value of 260 W m^{-2}. Generally, the sensible heat flux exceeded the latent heat flux, giving a Bowen ratio greater than 1. Note that we still use the WPL correction (i.e., Equation (20)) for calculating the latent heat fluxes shown in Figure 1.

Figure 2 shows the average daily course of net ecosystem carbon dioxide flux after the WPL correction (i.e., Equation (21)) for the same period as in Figure 1; we also show the raw CO_2 flux for comparison. Compared with the CO_2 fluxes after the WPL correction, the raw CO_2 fluxes show a much larger diurnal cycle with more positive values during the night and more negative

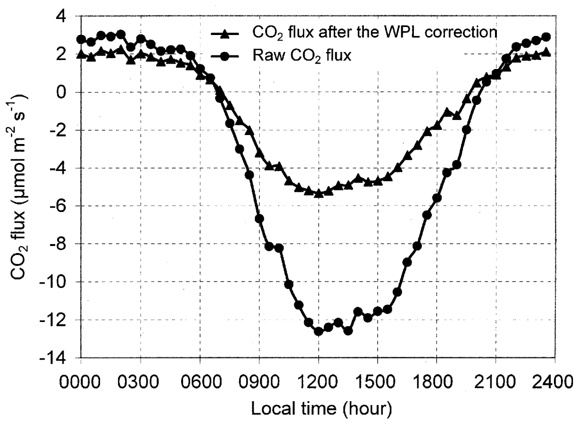

Figure 2. Mean 30-min raw CO_2 flux and CO_2 flux after the WPL correction using Equation (21).

values during the day. After the WPL correction, there was still a strong diurnal pattern of daytime net CO_2 uptake and nighttime net CO_2 release; the averaged midday CO_2 flux was about -5.0 ± 2.9 μmol m^{-2} s^{-1}. Average nighttime fluxes resulting from ecosystem respiration were about 1.8 μmol m^{-2} s^{-1} for this period. In general, the WPL correction for density effects reduced the magnitude of the raw CO_2 flux during both night and day. The raw CO_2 fluxes were on average about 1.4 times larger (more positive) than those after the WPL correction during the night, and 2.5 times larger (more negative) than those after the WPL correction during the day, indicating approximately 40% and 60% adjustment in magnitudes relative to raw CO_2 fluxes (Figure 2). Because the adjusted magnitudes in the WPL correction are proportional to sensible and latent heat fluxes, larger corrections are expected during the day according to Equation (21). When sensible and latent heat fluxes are close to zero (i.e., when the atmosphere is neutral) during the morning and evening transition periods, there are no adjustments and thus raw CO_2 fluxes are equal to fluxes after correction (Figure 2).

4.2. Raw water vapour flux and its correction

We present a comparison between the raw latent heat fluxes and those after our proposed correction (i.e., Equation (18)) in Figure 3. The magnitudes of correction for the measurements of latent heat flux are about +10%. The

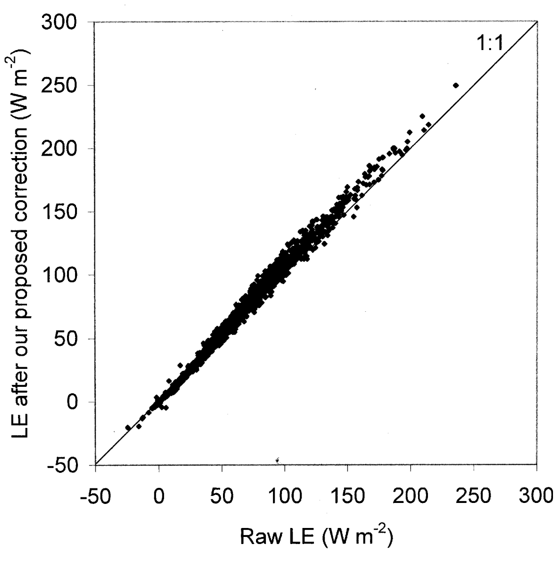

Figure 3. Comparison between raw LE and LE estimated from our proposed correction using Equation (18).

comparison of the latent heat fluxes as estimated using the WPL correction and our proposed correction is shown in Figure 4. There is a very small difference between the two approaches for latent heat flux (< 1–2%), indicating that the expansion/compression of water vapour has little influence on the density of water vapour itself over this high-latitude spruce forest with small evapotranspiration rate. During the day, our proposed correction leads to slightly smaller estimates of latent heat fluxes as compared with the WPL correction. Larger differences may occur in temperate and tropical regions that have higher rates of evapotranspiration.

4.3. Raw CO_2 flux and its correction

During the day, the minimum raw CO_2 fluxes can be as low as -24 μmol m^{-2} s^{-1} (Figure 5). After our proposed correction (i.e., Equation (19)), the magnitudes of CO_2 flux were about -10 μmol m^{-2} s^{-1}. Thus more than half of the daytime raw CO_2 flux over this spruce forest was caused by heat and water vapour effects on air density due to air-parcel expansion/compression. Around 72% of the correction can be attributed to the influence of sensible heat flux (according to the third term on the right-hand side of Equation (19)) and approximately 38% to latent heat flux (according to the second term on the right-hand side of Equation (19)). During the night, the raw CO_2 fluxes were as large as 4 μmol m^{-2} s^{-1} for this site. After our proposed correction,

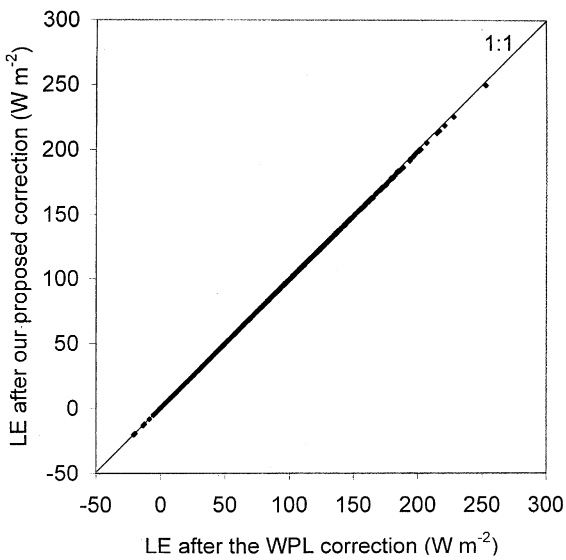

Figure 4. Comparison between LE after the WPL correction using Equation (20) and LE after our proposed correction using Equation (18).

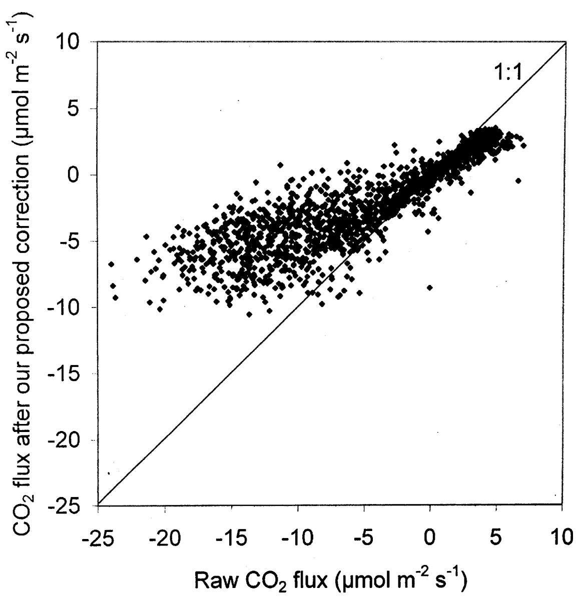

Figure 5. Comparison between raw CO_2 flux and CO_2 after our proposed correction using Equation (19).

the maximum magnitudes of the ecosystem respiration rates were around 3.0 μmol m^{-2} s^{-1}.

We compared the corrected CO_2 fluxes using our proposed correction and the WPL correction (see Figure 6). During the daytime, the magnitudes of CO_2 fluxes with our proposed correction are consistently on average about 10–20% more negative than those obtained using the WPL correction, indicating up to about 0.9 μmol m^{-2} s^{-1} more CO_2 uptake by vegetation. During the night, the CO_2 flux estimated using our proposed correction gives consistently about 4% (or about 0.3 μmol m^{-2} s^{-1}) lower respiration flux than that obtained from the WPL correction.

For the implication of long term NEE estimation, our proposed correction gives consistently more negative NEE than the WPL correction, implying a larger carbon sink during the growing season. Using the observations at the spruce forest stands from April 23 to October 21, 2002, the total integral of NEE using the WPL correction is −188 g C m^{-2} while the integral of NEE using our proposed correction is −237 g C m^{-2}, indicating up to 26% more carbon sink for our site during this period when our proposed correction is applied. Note that, in this seasonal flux estimate, no correction was made for the low nocturnal CO_2 flux as a result of inadequate turbulent mixing. However, it can be expected that this effect should be the same for the two methods.

Taking into account the effect of mean vertical velocity that is induced by air parcel expansion/compression when air parcels move upward and

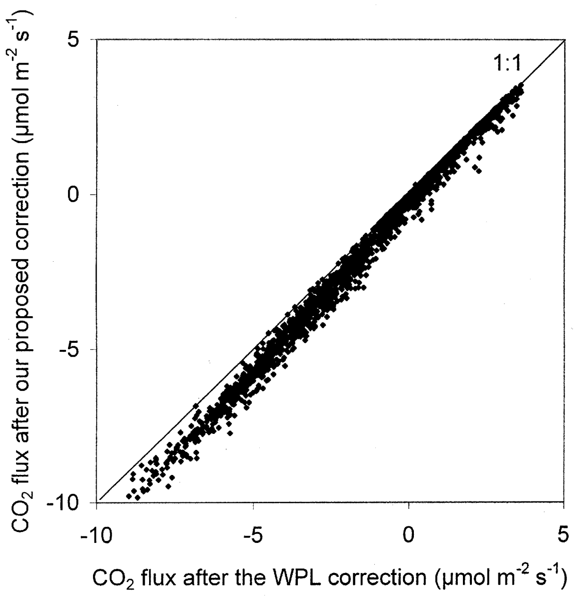

Figure 6. Comparison between CO_2 flux after the WPL correction using Equation (21) and CO_2 flux after our proposed correction using Equation (19).

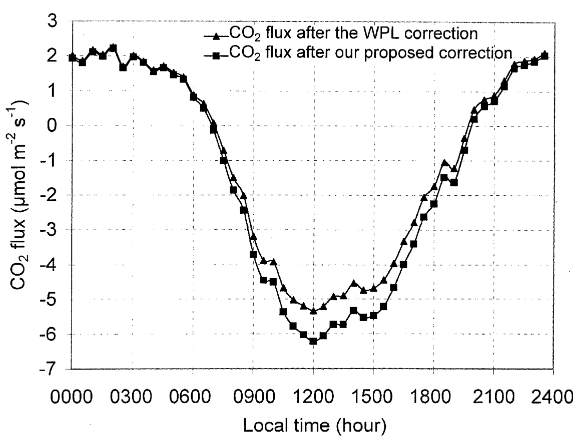

Figure 7. Mean 30-min CO_2 flux after the WPL correction using Equation (21) and CO_2 flux after our proposed correction using Equation (19).

downward in the surface layer, the WPL correction is obtained by assuming that the dry air is conserved (Webb et al., 1980). This assumption by Webb et al. (1980) actually implies that the mean vertical velocity is caused by the expansion/compression of the dry air only, while ignoring the effect of the expansion/compression of the water vapour part in the total moist air.

Although the mass of water vapour is quite small, there are still considerable contributions of water vapour expansion/compression to the mean vertical velocity over vegetated surfaces with strong evaporation. Our proposed approach, directly based on the physical mechanism of expansion/compression concept, indicates that the inclusion of the effect of water vapour expansion/compression can lead to a more negative daytime CO_2 flux and to a lower nighttime CO_2 flux compared with those after the WPL correction. Therefore, the difference in magnitudes of CO_2 flux using the WPL correction and our proposed correction depends on the amount of the water vapour in the total moist air.

5. Conclusions

Under an unstable atmosphere, thermal plumes usually develop. As the plume temperature varies, the expansion/compression of the plume occurs. Consequently, this expansion/compression of the total moist air (not the dry air only that is used in Webb et al. (1980)) leads to the variation in the density of constituents (e.g., water vapour and CO_2). From this physical point of view, we derived two equations for correcting water vapour and CO_2 fluxes that differ from those proposed by WPL. Our results indicate that latent heat fluxes after correction using two methods are quite similar. However, CO_2 fluxes after correction using two methods were different, with the CO_2 flux during the day obtained from our proposed correction being approximately 0.9 μmol m^{-2} s^{-1} more negative than that obtained from the WPL correction. During the night, our proposed correction yielded a CO_2 flux that was smaller (less positive by 0.3 μmol m^{-2} s^{-1}) than those obtained from the WPL correction. Integrated over the growing season, when our proposed correction was applied, we obtained a larger carbon sink (i.e., there was more CO_2 uptake by vegetation) as compared with the WPL correction. Our data show that the NEE from April 23 to October 21, 2002 was approximately 49 g C m^{-2} more negative when our proposed correction was applied than the WPL correction, indicating an approximately 26% larger carbon sink during the growing season in this high-latitude boreal forest. Larger differences may be possible at other forests, depending on the characteristics of surface vegetation and in particular on the magnitude of latent and sensible heat fluxes.

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Appendix A: A Three-Dimensional Continuity Equation for CO_2

Massman and Lee (2002) derived a fundamental equation of continuity for the measurement of CO_2 flux, which combines the WPL correction, non-ideal adevctive effects, and the pressure covariance. Starting from the continuity equation and following the procedures in Massman and Lee (2002), we re-derive the fundamental three-dimensional continuity equation for CO_2, and then compare it with the equation we derived in the text. The difference from Massman and Lee (2002) is that, instead of using the equation of continuity for the dry air, we use the continuity equation for the total moist air.

The continuity equations for CO_2 and the total moist air can be rewritten:

$$\frac{\partial \rho_c}{\partial t} + \nabla \cdot (\vec{V}\rho_c) = S_c, \tag{A1}$$

$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\vec{V}\rho) = S, \tag{A2}$$

where $\vec{V}$ denotes velocity, S_c and S are source/sink terms for CO_2 and the total moist air, respectively.

We define the mixing ratio, ω, as the ratio of CO_2 to the total moist air (or CO_2 mass fraction) that is different from that used in Massman and Lee (2002) and Paw U et al. (2000) in which CO_2 mass fraction is the ratio of CO_2 to the dry air, so that we obtain

$$\omega = \frac{\rho_c}{\rho}. \tag{A3}$$

After applying a Reynolds decomposition and Reynolds averaging to Equations (A1)–(A3), we obtain

$$\frac{\partial \overline{\rho_c}}{\partial t} + \nabla \cdot (\overline{\vec{V}}\,\overline{\rho_c} + \overline{\vec{V}'\rho_c'}) = \overline{S_c}, \tag{A4}$$

$$\frac{\partial \bar{\rho}}{\partial t} + \nabla \cdot (\overline{\vec{V}}\,\bar{\rho} + \overline{\vec{V}'\rho'}) = \overline{S}, \tag{A5}$$

$$\overline{\omega} = \frac{\overline{\rho_c}}{\overline{\rho}}. \tag{A6}$$

Using the same procedure as Massman and Lee (2002), multiplying Equation (A5) by $\overline{\varpi}$, subtracting Equation (A4) and then using the incompressibility assumption for the mean flow($\nabla \cdot \vec{V} = 0$), I obtain

$$\overline{\rho}\frac{\partial \overline{\varpi}}{\partial t} + (\overline{\vec{V}}\overline{\rho_c} + \overline{\vec{V}'\rho'} \cdot \nabla \overline{\varpi} - \overline{\vec{V}}\overline{\varpi} \cdot \nabla \overline{\rho}) + \nabla \cdot (\overline{\vec{V}}\overline{\rho_c} + \overline{\vec{V}'\rho_c'} - \overline{\varpi}\overline{\vec{V}'\rho'}) = \overline{S}_c - \overline{\omega S}. \tag{A7}$$

The first term on the left-hand side (LHS) is the storage term, and the second term on the LHS is the advective term; see Baldocchi et al. (2000), Massman and Lee (2002) and Paw U et al. (2000) for detailed discussions about the influence of advection on flux measurements. Here we are interested in the third term that explains CO_2 flux divergence.

The CO_2 flux can be written from the third term on the LHS of Equation (A7) as

$$F_c = \overline{\vec{V}}\overline{\rho_c} + \overline{\vec{V}'\rho_c'} - \overline{\varpi}\overline{\vec{V}'\rho'}. \tag{A8}$$

The first term on the right-hand side (RHS) of Equation (A8) denotes the influence of the mean velocity on flux (e.g., the mean velocity caused by mesoscale effects, heterogeneous terrain, etc.). The second term on the RHS of Equation (A8) is the flux measured by eddy covariance systems. The third term on the RHS of Equation (A8) is equivalent to the correction term caused by air-parcel expansion/compression of the total moist air. Substitution of Equation (8) in the text into the third term on the RHS of Equation (A8) yields

$$\begin{aligned} F_c &= \overline{\vec{V}}\overline{\rho_c} + \overline{\vec{V}'\rho_c'} - \frac{\overline{\rho_c}}{\overline{\rho}}\left\{(1 - \frac{m_a}{m_v})\overline{\vec{V}'\rho_v'} - (\overline{\rho}_a + \frac{m_a}{m_v}\overline{\rho}_v)\frac{\overline{\vec{V}'T'}}{\overline{T}}\right\} \\ &= \overline{\vec{V}}\overline{\rho_c} + \overline{\vec{V}'\rho_c'} + \left\{\frac{\overline{\rho_c}}{\overline{\rho}}(\frac{m_a}{m_v} - 1)\overline{\vec{V}'\rho_v'} + \frac{\overline{\rho_a}}{\overline{\rho}}\overline{\rho}_c(1 + \frac{m_a}{m_v}\frac{\overline{\rho}_v}{\overline{\rho}_a})\frac{\overline{\vec{V}'T'}}{\overline{T}}\right\} \\ &= \overline{\vec{V}}\overline{\rho_c} + \overline{\vec{V}'\rho_c'} + \left\{\frac{\overline{\rho_c}}{\overline{\rho}}(\mu - 1)\overline{\vec{V}'\rho_v'} + \frac{\overline{\rho_a}}{\overline{\rho}}\overline{\rho}_c(1 + \mu\sigma)\frac{\overline{\vec{V}'T'}}{\overline{T}}\right\}, \end{aligned} \tag{A9}$$

where $\mu = m_a/m_v$, and $\sigma = \overline{\rho}_v/\overline{\rho}_a$.

Comprehensive discussion about the influence of the pressure perturbation, which is not included here, can be found in Massman and Lee (2002). The one-dimensional form of Equation (A9) is exactly the same as Equation (19) in the text when the influence of the mean vertical velocity on flux (the first term on the RHS of Equation (A9) is neglected in the case of a single point measurement). The identity of Equation (A9) and Equation (19), which are based on two different physical principles, one from expansion/compression concept and the other from the continuity equation, provides confidence in our approach. As mentioned in Massman and Lee (2002), such a derivation of three-dimensional continuity equation for CO_2 not only avoids the effects resulting from different initial assumptions, but also provides generalized information that is not included in the results of Webb et al. (1980).

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