Evaluation of the aerodynamic method of determining fluxes over natural grassland

By B. SAUGIER and E. A. RIPLEY
C.N.R.S.-C.E.P.E., Plant Ecology Department,
P.5051, University of Saskatchewan,
34033 Montpellier, France Saskatoon, Canada

(Received 13 December 1976; revised 19 August 1977)

SUMMARY

Profiles of wind speed, temperature, humidity, and CO₂ content up to a height of 6.5 m were measured over a period of several months above natural grassland in an area with an average fetch of 2 km. Profile shapes have been analysed in relation to atmospheric stability.

In unstable conditions, temperature and CO₂ profiles exhibit similar shapes and depart more from logarithmic than does the wind profile. All profiles are relatively well described by existing empirical stability corrections.

In stable conditions all profiles have approximately the same shape: they depart from the logarithmic profile at moderate stabilities and return to it at strong stabilities. However, the oft-quoted log-linear profile does not provide a very good fit.

Empirical stability corrections have been used to develop a generalized aerodynamic method for flux computation. This method tends to underestimate (by up to 20%) fluxes during unstable conditions and to overestimate (by up to 40%) in moderately stable conditions. These results stem from a comparison of aerodynamic and Bowen ratio estimates of sensible heat flux. They are in essential agreement with the analysis of profile shapes.

The drag coefficient of vegetation may be predicted from the leaf area index. This allows the use of a simplified aerodynamic method, based on measurements of wind speed at one level and of temperature (or H₂O or CO₂) at two levels.

1. INTRODUCTION

The measurement and analysis of water and carbon dioxide fluxes in the atmosphere above vegetated surfaces are pertinent to the understanding of the functioning of plant communities in their natural environment. Although there have been a large number of such measurements in the last two decades (see, for example, Monteith 1976), most studies have been restricted by instrumental difficulties to only a few days of measurements, and have not permitted the following of vegetation responses over a complete growing season. One exception to this is the study of Biscoe et al. (1975).

Direct measurements of H₂O and CO₂ fluxes by eddy correlation techniques, however attractive, have not yet reached the point where they can be used on a routine basis. Ecophysicists still have to rely mainly on the computation of fluxes from corresponding micrometeorological profiles. Two main methods are available: the Bowen ratio method and the aerodynamic method. The first assumes equality of turbulent transfer coefficients for heat, Kₜ, and water vapour, Kₚ – an assumption now well supported by experiments, so that it is at present the most widely used method. The aerodynamic approach relies on similarity between the wind profile and other micrometeorological profiles. It was introduced by Thornthwaite and Holzman (1939) for the computation of evaporation. They assumed logarithmic wind and water vapour profiles, and equality of the turbulent transfer coefficients for momentum (Kₜ) and water vapour (Kₚ). Several variants of their approach have since been devised to take into account the effect of thermal stratification of the atmosphere, and the elevation of the wind profile by vegetation elements through the use of the so-called zero plane displacement height, d (Thom 1971).
There is still no general solution to the problem of relating profiles to fluxes for a thermally stratified atmosphere, although progress is being made in that direction by relating surface fluxes to variables specifying conditions at the upper and lower limits of the planetary boundary layer (e.g. Arya 1975). However, careful simultaneous measurements of profiles and fluxes in recent years have resulted in various empirical formulations (Dyer and Hicks 1970; Webb 1970; Businger et al. 1971; Pruitt et al. 1973) that did not lead to widely different fluxes when applied to measured profiles.

Strong objections have been made to the application of the aerodynamic method above surfaces such as a cornfield (Mukammal et al. 1966) or a pine forest (Thom et al. 1975). In these cases the method gives values of evaporation that are too small when compared with estimates from a lysimeter or from the Bowen ratio.

The present paper seeks to show that the aerodynamic method is still of great interest for routine measurements above large areas of short vegetation such as grassland. Existing empirical stability corrections lead to reasonably good estimates of surface fluxes ($H_2O$ and $CO_2$).

2. SITE AND INSTRUMENTATION

Measurements were taken at the Matador Field Station (50°42'N 107°43'W; 682 m), Saskatchewan, as part of the International Biological Program (Coupland et al. 1974). The land, protected from grazing, had a mean slope of 0·6%. In spite of a sharp drop to Diefenbaker Lake at the southern boundary, the fetch exceeded 2 km for the prevailing wind directions (Figs. 1(a), (b)). The soil surface was covered with short (2·5 cm) grass consisting largely of dead leaves with a lesser, varying, amount of green leaves.

Table 1 summarizes the profile instrumentation. The sensors, described more fully in Ripley and Saugier (1972, 1974), were positioned on four masts 50–80 m to the southwest of the recording trailer. Anemometers were calibrated in a wind tunnel before and after each measurement period, and intercomparisons on the site showed their readings to agree within 2%. A single calibration curve was therefore used for all anemometers.

Thermocouples were used for air temperature. They were lagged to give them a response time of about 90 s when aspirated at 5 m s$^{-1}$ in their radiation shields. They were referenced to a brass block buried at a depth of 1 m in the soil. Block temperature was measured with a resistance thermometer.

The psychrometers used platinum resistance thermometers that were calibrated in-
Figure 1. (a) Location of the micrometeorological site (black dot) in relation to the overall study area (8 km²) and adjacent lands. Study area contour elevations are in metres. (b) Average wind rose for May to September 1970 and 1971. Centre value (1-8%) is the frequency of calm winds.

dividually against a standard resistance thermometer. These sensors were sheathed with 9 mm diameter stainless steel tubes to provide time constants of 2 min for the wet bulbs and 3 min for the dry bulbs. At the aspiration rate of 7 m s⁻¹ the radiation error compensated almost exactly for the difference between heat and water vapour transfer coefficients. This allowed the psychrometric constant to be taken equal to its thermodynamic value: 61 Pa K⁻¹ (1 Pa K⁻¹ = 0.01 mb K⁻¹) at the Matador elevation. Psychrometers were mounted on a rotating mast that brought them to the same level for 30 minutes every 6 hours. Typical accuracy of both thermocouple and resistance thermometers was about 0.05K.

The CO₂ sampling system has been described fully in Saugier and Ripley (1974) and in Ripley et al. (1973). Air was pumped from the sampling nozzles through a desiccant to a sensitive differential infrared gas analyser that was frequently calibrated with standard gas mixtures. The accuracy for measured differences was about 0.1 vpm after applying a volume correction for the removal of water vapour (Parkinson 1971).

Net radiation was measured with a Swissteco allwave radiometer, calibrated against an Ångström pyrhiometer. The heat flux at the surface of the soil was calculated from its value at 5 cm, measured with flux plates, corrected for the change in heat storage in the 0–5 cm layer.
All measurements were recorded every three minutes on magnetic tape, using a digital data logger with a resolution of 1 μV for DC voltage and 0.01 ohm for resistance. Bad data, such as caused by inadequate water supply to the psychrometer wicks or by desiccant restriction in the CO₂ sampling lines, were deleted after visual inspection of profile plots. Raw data were converted to physical units and averaged for half-hour periods. Measurements were taken almost continuously for 2 ½ months in 1970 and for 5 months in 1971.

Both the raw and the summarized data are available on magnetic tape, on punched cards and in printed form, from the authors. An abbreviated hourly summary for 1971 is contained in Ripley and Saugier (1973). The climatology of the area has been reviewed by Ripley (1973).

3. Computation of Fluxes

The Bowen ratio and aerodynamic methods of flux computation have been described many times (e.g. Webb 1965; Saugier 1974), and only features that are relevant to the present study will be reviewed here.

(a) Bowen ratio method

The ratio between fluxes of sensible and latent heat may be written as

$$\beta = H/\lambda E = \gamma \Delta T/\Delta e$$

where $\lambda$ is the latent heat of vaporization, $\gamma$ the thermodynamic value of the psychrometric constant, and $\Delta T$ and $\Delta e$ the differences in temperature and in water vapour pressure between two heights of measurements. Eq. (1) assumes equality of turbulent transfer coefficients of heat and water vapour, which assumption is supported by measurements over a wide range of atmospheric stabilities (for a discussion see Monin and Yaglom 1971 and Saugier 1974).

Conservation of energy at the vegetation level gives another relationship involving $H$ and $\lambda E$:

$$H + \lambda E = R_n - S$$

where $R_n$ is net radiation above the vegetation and $S$ is heat flux into the soil surface. (Terms involving plant metabolism and heat stored in biomass are small and have been omitted.)

In this work, temperature and humidity were measured at 5 levels using dry and wet bulb thermometers. If $T$ and $T'$ are dry and wet bulb temperatures, the psychrometric equation may be written in differential form as

$$\Delta e = (s + \gamma)\Delta T' - \gamma \Delta T$$

where $s = de_s(T')/dT'$; and $e_s(T')$ is the saturation vapour pressure at temperature $T'$, the psychrometric constant for the instruments used in this study being taken equal to its thermodynamic value, $\gamma$ (see section 2). Elimination of $\Delta e$ between equations (1) and (3) then gives

$$1/\beta = \lambda E/H = [(s + \gamma)/\gamma] \Delta T'/\Delta T - 1$$

Eqs. (2) and (4) allow easy computation of $H$ and $\lambda E$. Since measurement errors in $T'$ and $T$ are similar, the ratio $\Delta T'/\Delta T$ has been computed as the slope of the orthogonal regression line fitted to the $T'$ v. $T$ measurements (5 points).

(b) Aerodynamic method

In this method, theoretical profiles of wind speed ($u$), temperature ($T$), water vapour pressure ($e$), and CO₂ concentration ($c$) have been fitted to the measurements to derive
scale parameters $u_*, T_*, e_*$ and $c_*$ for each profile. Following Panofsky (1963), one may define stability functions, $\psi$, that correct the shape of the usual logarithmic profiles, according to the following relations:

$$u = \left( u_*/k \right) \ln \left( \frac{(z-d)/z_0}{\psi_\infty(z/d)/z_0} \right)$$

$$T = -(T_*/k) \ln(z-d) - \psi_H(z/d)/L) + \text{constant}$$

and relations similar to (5b) for $e$ and $c$. $k$ is the von Kármán constant, taken as 0.40; $d$ the zero plane displacement height; $z_0$ the roughness length; and $L$ the Obukhov length $-\rho c_p T_*/k q H$, where $\rho$ is air density and $T$ air temperature). The same stability function, $\psi_H$, was adopted for water vapour and for CO$_2$ profiles.

Functional forms of $\psi_\infty$ and $\psi_H$ have been determined only empirically. The stability corrections of Dyer and Hicks (1970) - first suggested by Businger (1966) - for unstable conditions, and of Webb (1970) for stable conditions, have been used as the best fits of recent measurements. The resulting expressions for $\psi_\infty$ and $\psi_H$ are given in Paulson (1970).

Since $L$ is a function of surface fluxes ($u_*$ and $H$), it cannot be computed directly from the observed profiles. Thus another stability parameter, the gradient Richardson number $Ri$, is computed first from gradients of virtual potential temperature and wind speed, obtained by fitting second-order polynomials in $\ln z$ to the observed profiles and taking the derivative at a reference height $z_1$. $L$ is then computed using the following expressions, which form the basis of the stability corrections of Dyer and Hicks (1970) and of Webb (1970):

$$\begin{align*}
\text{unstable} (Ri \leq 0) & \quad L = z_1/Ri \\
\text{stable} (0 < Ri \leq 0.16) & \quad L = z_1(1/Ri - 4.7)
\end{align*}$$

Then $u_*/k$ is computed as the slope of the regression line of measured wind speed $v$. $\ln (z/d)/z_0 - \psi_\infty$ (Eq. (5a) and Fig. 2). Other flux parameters, $T_*$, $e_*$ and $c_*$, are obtained in a similar way. This procedure has the advantage of using all the measurement levels. Finally, the fluxes are computed as $\rho c_p u_* T_*$ (sensible heat), $(\rho c_p/\gamma)u_* e_*$ (latent heat), and $(44/29)\rho u_* c_*$ (carbon dioxide).

In the first instance, $d$ was computed for each half-hour's data to give the line of best fit to the observed wind profile. A seasonal pattern of variation of $d$ was thus established: 10 cm in early May and late September, 12 cm in July, in phase with the amount of green

![Figure 2](image-url)  
Figure 2. Illustration of the method used to derive $u_*$ from the measured wind profile.
leaves. However, this procedure often produced values of \( d \) that varied considerably from the trend line and occasionally were even negative. Since part of this variability is thought to be real (due to changes in canopy structure with wind speed and to the influence of buoyancy within the canopy), in the final computations \( d \) was allowed to vary by only \( \pm 2.5 \text{ cm} \) around the seasonal trend line.

4. Evaluation of the methods

Accuracy in the determination of fluxes by the methods described above is limited by: (a) errors in the measurements; and (b) hypotheses concerning turbulent transfer mechanisms. As regards (a), it is possible to assess the likely accuracy in the measurements from a knowledge of each sensor type, its capacity to retain its calibration, and the accuracy of the measuring system. Such analyses give only minimum estimates of real errors occurring in field measurements. More realistic evaluations may be made by simply looking at departures of measurements from theoretical profiles. Observed departures in the present study lead to errors in flux parameters, \( u^*, T^*, T'_a, c_a \), that are about 3\% for wind speed, 5\% for dry bulb temperature, 10\% for wet bulb temperature, and 5 to 10\% for CO\(_2\) concentration.

In regard to (b), the assumed equality of eddy diffusivities for heat and water vapour is not likely to introduce serious errors, but equality of these diffusivities to the one for CO\(_2\) might be questioned since it has never been verified by direct simultaneous measurements of fluxes and profiles. The validity of the stability corrections used in the aerodynamic method may be tested by comparing the shapes of measured and theoretical profiles, or by comparing the fluxes computed by the two methods – aerodynamic and Bowen ratio.

To compare profile shapes, the simple, but sensitive, method of Swinbank and Dyer (1967) was used. The profile of any atmospheric property, \( X \), may be characterized by a shape function defined as \( S_X = (X_2 - X_1)/(X_3 - X_1) \), where \( X_1, X_2 \) and \( X_3 \) are measurements of the property at three (usually logarithmically-spaced) levels.

Figures 3(a), (b), (c) show the shape functions for wind speed, temperature and CO\(_2\) profiles plotted against the gradient Richardson number at a height of 1 m, \( R_i_{1m} \). Solid lines represent theoretical shape functions computed using a constant value of 14 cm for \( d \). Experimental data have been separated into 12 stability classes, on the basis of \( R_i_{1m} \), and mean values of the shape functions have been plotted with their standard deviations against the geometric mean of each \( R_i \) range. The number of runs is indicated for each class. Only the best CO\(_2\) profiles have been used to compute \( S_x \) and in the neutral case for temperature, extravagant values of \( S_T \) caused by very small temperature gradients have been deleted. The measurement heights chosen were those closest to the levels 40, 160, and 640 cm above the zero plane, and are indicated on each figure. Consequently, all three functions show values near 0.5 at neutrality. Increasing instability causes the vertical gradients to decrease more at the higher levels, resulting in an increase in \( S_x \), while increased stability tends to decrease \( S_x \). Several conclusions may be drawn from Figs. 3(a) to (c):

1. Data points for temperature and CO\(_2\) are in good agreement (bars overlap) over most of the range, the only exceptions occurring for \( R_i = 0 \) and 0.005. It is thus justified to assume proportionality, and likely equality, of turbulent diffusivities for heat and CO\(_2\). This assumption, necessary for computing CO\(_2\) flux by any profile-based method, has had little experimental support so far.

2. Both sets of data are relatively well accounted for by the theoretical line in unstable conditions, giving confirmation of the Dyer and Hicks (1970) corrections for both heat and mass transfer.

3. Wind speed data are fitted better by the KEVPS representation (using again
**Figure 3.** Variation with stability of the profile shape functions \( S_x = (X_2 - X_1)/(X_3 - X_1) \); A. for wind speed; B. for temperature; C. for CO\(_2\) concentration. The solid lines represent theoretical expressions, and dots with bars the measured data with standard errors of the means. The number of data values is indicated for each stability class.

\( z/L = Ri \), with \( \gamma = 18 \); see Panofsky (1963) and Paulson (1970) for definitions) than by that of Dyer and Hicks (1970). The implication is that the latter stability corrections do not fully account for the observed departure of the wind profiles from the logarithmic shape.

(4) In stable conditions the three sets of data show a similar trend with a minimum at about \( Ri_{\text{trim}} = 0.05 \). This supports the equality of the eddy diffusivities for momentum, heat and CO\(_2\), again with the exception of the CO\(_2\) points at \( Ri = 0 \) and 0.005. This is essentially in agreement with the findings of Webb (1970). However, the theoretical line fails to fit the data even at moderate stabilities. This line represents the log–linear profile with
$\alpha = 4.7$, the value found by Businger et al. (1971), and close to that, 5.2, reported by Webb (1970).

The disagreements observed between 'theoretical' and measured shape functions emphasize the difficulties experienced when using empirical stability corrections that have no real theoretical justification. Corresponding errors in the fluxes have been estimated by plotting observed and predicted values of $\psi_M$ and $\psi_H$ vs. $(z-d)/L$. They lead to a maximum underestimation of 7% in both $u_*$ and $T_*$ (i.e. 14% in sensible heat flux) for unstable conditions, and to an overestimation of 20 to 50% in aerodynamic fluxes for moderate ($0 < R_i < 0.05$) stabilities.

To check the validity of these tentative conclusions, the values of sensible heat fluxes,
computed by the aerodynamic and energy balance methods, have been compared in relation to atmospheric stability. The computed half-hour fluxes have been grouped into three classes: unstable, $R_i < -0.03$ (Fig. 4(a)); stable $R_i > +0.01$ (Fig. 4(c)); and near neutral (Fig. 4(b)), the remaining cases. Although there is a considerable amount of scatter, the near-neutral class shows quite good agreement between the two methods. The unstable class, too, shows good agreement for smaller fluxes, but a tendency is revealed for aerodynamic estimates to be lower than energy balance values as the magnitude of the fluxes increases.

There is obviously strong correlation between instability and flux magnitude. For the stable class, aerodynamic estimates are considerably more negative than energy balance estimates. The data of Fig. 4 have been summarized (Table 2) using six stability classes. It may be observed that the agreement between aerodynamic and energy balance estimates is best for the slightly unstable case. The lower aerodynamic estimates for the greater instability cases, as well as the higher estimates for the stable cases, are consistent with the analysis of the shape function given above. Averaged over the entire 5370 half-hour cases, shown in Figs. 3 and 4, the aerodynamic estimate is 45 W m$^{-2}$, which is 19% less than the energy balance estimate (55 W m$^{-2}$). If only positive values of $H$ are considered, the difference reduces to 8%.*

5. A SIMPLIFIED AERODYNAMIC METHOD

In computing flux parameters such as $u_*$ and $T_*$ from the corresponding profiles, even if all (i.e. 7 or 8) measurement levels are used, more weight is given to the extreme levels, because of the nature of the computation of a regression line. One may thus ask whether accurate results could not be obtained using only two levels, one close to the vegetation and the other as high as permitted by the fetch. For even greater simplification, one may use only one level for wind speed since the wind must drop to zero near the ground (drag coefficient concept).

Such an approach would use relation (5b) for computing $T_*$, and (5a) would be replaced by

$$u_* = ku(z)/\left[\ln\left((z-d)/z_0\right) - \psi_M\right] \quad (7)$$

Prior knowledge of $d$, $z_0$, and $L$, the stability parameter, are required for use of Eq. (7). The values of $d$ and $z_0$ are determined from the vegetation structure (height and density), and $L$ may be computed from $u(z)$ and the difference in temperature, $T_2 - T_1$, between two levels.

To test the validity of this approach, several regressions of $u_*$ (computed using all levels of wind measurements) against wind speed at the top measurement level have been computed, one of which is shown in Fig. 5. In spite of the expected effect of stability, the correlation between $u_*$ and $u$ is quite good and permits an accurate determination of the slope. This relation can be used to derive $u_*$ from $u$. Even if the stability correction is ignored, the

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* And to 24% if 0-41, rather than 0-40, is used for $k$, as pointed out by a referee.
resulting error in $u_*$ does not exceed 15%, provided $u(6.35\,\text{m})$ is greater than 5 m s$^{-1}$, which was usually the case at Matador in the daytime. If the stability correction of Dyer and Hicks is applied, the error is reduced to less than 5%.

One limitation of using Eq. (7) is the oft-quoted variation of roughness height with wind speed. Values of $z_0$ computed from wind speed profiles using stability corrections have been plotted against wind speed at the lowest measurement level (Fig. 6), using the same runs as in Fig. 5. There is a significant variation of $z_0$ at low wind speed, but, for the quasi-totality of the daytime runs, $z_0$ stays between 0.9 and 1.3 cm, implying variation in $u_*^2/u$ from 0.061 to 0.065, an excursion of only 3% on each side of the average value 0.063. Variation in $z_0$ with wind speed is somewhat greater when the vegetation becomes denser (from 2 to 3.4 cm in early July 1971), but even in this worse case the error made in using an average $u_*^2/u$ ratio does not exceed 5%.

Density of the vegetation is usually characterized by the leaf area index: i.e. the total area of leaves (counting one side only) per unit area of soil. The Matador vegetation consists...
of a relatively constant large amount of dead leaves and a varying, but smaller, amount of green leaves. Because green leaves tend to be located higher in the canopy than dead ones, they experience a higher wind speed and contribute more to the resistance offered to the wind (i.e. to the momentum flux). The resistance offered by a canopy of effective leaf area index \(L_a\) may be written as \(\rho L_a Cu^2\) where \(\rho\) is air density and \(C\) the drag coefficient of a single leaf, assumed here to be constant. For a canopy composed of both green and dead leaves, therefore,

\[
L_a = \int_0^h u^2 (dL_a + dL_d) / \int_0^h u^2 dz
\]

(8)

where \(dL_a, dL_d\) represent the leaf area indices of green leaves and of dead leaves, respectively, in a canopy layer of thickness \(dz\), and height \(h\). Using average profiles of canopy leaf area indices and wind speed (Ripley and Redmann 1976) leads to the expression

\[
L_a = 2.16L_a + 0.76L_d
\]

(9)

The momentum flux, \(\tau\), may now be written as

\[
\tau = \rho u_a^2 = \rho L_a Cu^2
\]

(10)

or

\[
u_u/u = L_a^2 C^4
\]

(11)

Eq. (11) may be tested by computing \(u_u/u\) and \(C^4\) at different times during the growing season. Table 3 gives the result of such computations for two periods in 1970 and three in 1971. Corresponding values of \(z_0\) are also given, computed as \((z-d)\exp(-ku/u_0)\), with \(z = 6.35\) m, \(d = 0.14\) m and \(k = 0.40\). The resulting values of \(C^4\) are quite constant with a

<table>
<thead>
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<th>Date</th>
<th>(L_a)</th>
<th>(L_d)</th>
<th>(L_a)</th>
<th>(u_u/u(6.35m))</th>
<th>(z_0(\text{cm}))</th>
<th>(C^4)</th>
</tr>
</thead>
<tbody>
<tr>
<td>16-21 July 70</td>
<td>1.5</td>
<td>3.05</td>
<td>5.56</td>
<td>0.0735</td>
<td>2.7</td>
<td>0.0312</td>
</tr>
<tr>
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<td>4.62</td>
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<td>0.0327</td>
</tr>
<tr>
<td>6-16 May 71</td>
<td>0.35</td>
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<td>3.75</td>
<td>0.0630</td>
<td>1.1</td>
<td>0.0326</td>
</tr>
<tr>
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<td>3.9</td>
<td>4.36</td>
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<td>0.0322</td>
</tr>
<tr>
<td>3–13 July 71</td>
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<td>4.2</td>
<td>5.25</td>
<td>0.0732</td>
<td>2.7</td>
<td>0.0320</td>
</tr>
</tbody>
</table>

Table 3. Influence of leaf area index (LAI) on the ratio \(u_u/u(6.35m)\) and the associated mean roughness length \(z_a\). \(L_a, L_d\) are the respective LAIs of green leaves and dead leaves; \(L_a\) is the effective aerodynamic LAI (= 2.16L_a + 0.76L_d); and \(C^4\) the square root of the drag coefficient \((= u_u/u)^{-1} L_a^2\).

Thus, \(u_u\) may be estimated within a few percent at neutrality, using relations (9) and (11) and measurements of leaf area indices and wind speed. Combining equations (7) and (11) for neutral conditions \((\psi_M = 0)\), gives

\[
L_a^2 C^4 = k/\ln((z-d)/z_0)
\]

(12)

This expression permits computation of \(z_0\) at any time, and thus use of relation (7) to compute \(u_u\) even when a stability correction has to be applied. The Obukhov length, \(L\) (= constant \(x u_u^2/T_u\)), may be obtained by an iterative method, or from \(Ri\) as previously.

This simplified aerodynamic method requires the measurement of wind speed at only one level, and temperatures at two levels, in order to compute sensible heat flux with an accuracy of about 20% in the daytime and somewhat less (up to 40%) at night, using the stability corrections described in section 3. Fluxes of water vapour and of CO₂ may also be computed in the same way, from differences in humidity and CO₂ concentration at two levels.
The method that has just been described is similar to the aerodynamic method of Thornthwaite and Holzman (1939), but has two significant advantages:

(a) The drag coefficient approach introduces a relationship between the aerodynamic properties of the surface and a measurable canopy parameter, the leaf area index. This permits the accurate determination of \( u_0 \) from a single wind measurement level.

(b) The use of relatively accurate stability corrections decreases the errors in the computation of the daytime fluxes to within about 20%.

6. CONCLUSION

Profiles of wind speed, temperature, humidity, and CO\(_2\) content of the air have been measured above natural grassland over periods of several months, with the aim of following the diurnal and seasonal evolution of H\(_2\)O and CO\(_2\) fluxes over an entire growth period.

The data were of sufficient quality to permit an analysis of the influence of stability on the shape of the profiles, with the exception of humidity measurements that were used only in the Bowen ratio method.

In the unstable range, temperature and CO\(_2\) profiles were found to have an identical shape that is well described by the Dyer and Hicks (1970) stability correction for temperature. The wind speed profile is affected less by instability than the temperature (or CO\(_2\)) profile, but more than is predicted by the Dyer and Hicks correction for wind speed. The KEYPs profile, with \( \gamma = 18 \), would give a better fit.

In the stable range, all profiles have an identical shape (with the possible exception of CO\(_2\) near neutrality, although the accuracy of the CO\(_2\) data does not permit of a definite conclusion): as stability increases, the profiles at first depart from logarithmic, and gradually return to it beyond a critical stability (in this case for \( R_{1m} \sim 0\-05 \)) as has been noted by Webb (1970). The shape of the profiles is not described very well by the log–linear expression even at moderate stabilities.

Errors in flux computations caused by using the preceding (unstable – Dyer and Hicks 1970; stable – Webb 1970) stability corrections have been assessed by a simple error evaluation technique and by a comparison of aerodynamic and Bowen ratio estimates of the sensible heat flux. The fluxes are slightly underestimated (up to 20%) during unstable conditions and overestimated (up to 100%) during stable conditions. If stable cases are restricted to \( R_{1m} < 0\-05 \), which seems reasonable when looking at the shape functions (Fig. 3), the overestimation decreases to 40%, and possibly less if provision is made for systematic errors in the Bowen ratio method (see regression line in Table 2). The aerodynamic method may thus be recommended for use above relatively smooth surfaces in the daytime, or at night when the stability is weak to moderate. Better stability corrections, as may be developed in the future, would of course result in even more accurate calculations of the fluxes.

A simplified aerodynamic method, based on measurements of wind at one level, temperature (and H\(_2\)O and CO\(_2\) concentrations, etc.) at two levels, and a drag coefficient, calculated from the leaf area indices of both green and dead leaves, has been developed and shown to give reasonably accurate flux estimates.

An analysis of computed H\(_2\)O fluxes is presented in Ripley and Saugier (1978).

ACKNOWLEDGMENTS

The authors are grateful to the National Research Council of Canada for the funding provided through the Canadian Committee for the International Biological Programme. They would also like to thank D. Spittlehouse and D. Couturier for their assistance in the collection and analysis of the field data.
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