An Air–Soil Layer Coupled Scheme for Computing Surface Heat Fluxes

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ABSTRACT

An air–soil layer coupled scheme is developed to compute surface fluxes of sensible heat and latent heat from data collected at the Oklahoma Atmospheric Radiation Measurement–Cloud and Radiation Testbed (ARM–CART) stations. This new scheme extends the previous variational method of Xu and Qiu in two aspects: 1) it uses observed standard deviations of wind and temperature together with their similarity laws to estimate the effective roughness length, so the computed fluxes are nonlocal; that is, they contain the contributions of large-eddy motions over a nonlocal area of \( O(100 \text{ km}^2) \); and 2) it couples the atmospheric layer with the soil–vegetation layer and uses soil data together with the atmospheric measurements (even at a single level), so the computed fluxes are much less sensitive to measurement errors than those computed by the previous variational method. Surface skin temperature and effective roughness length are also retrieved as by-products by the new method.

1. Introduction

Surface fluxes of sensible and latent heat are conventionally estimated by the Bowen ratio energy balance (BREB) method (Fritschen and Simpson 1989). The BREB method uses near-surface wind, vertical gradients of temperature and water vapor pressure, and surface energy budget measured by the surface energy and radiation balance system (SERBS), but it does not make full use of the information provided by the similarity law and wind measurements for turbulent flow in the surface layer. The BREB method becomes computationally unstable and produces spurious large values in the computed fluxes when the Bowen ratio is in the vicinity of \(-1\). The profile method uses the equations of the similarity law to compute surface fluxes of sensible and latent heat from the atmospheric measurements in the surface layer (Panofsky and Dutton 1984), but it does not use the information provided by the surface energy budget measurements, and the computed fluxes are sensitive to measurements of the gradients of temperature and humidity across the surface layer. To overcome the drawbacks of these conventional methods, a variational method was developed by Xu and Qiu (1997). This method makes combined use of the information provided by the SERBS measurements, the surface energy balance equation, and the similarity profile equations. The data collected at the Oklahoma Atmospheric Radiation Measurement–Cloud Radiation Testbed (ARM–CART) central station (Stokes and Schwartz 1994) were used to test the variational approach, which was found to be more reliable than the two conventional methods. In addition to the use of an optimal estimation method (Daley 1991; Cohn 1997), the additional data and physical constraints are believed to be the main reasons for the improved performance of the variational method.

When the ARM–CART surface observation data are used to compute the area-averaged fluxes, we see two problems that need to be considered. First, the observation levels in the surface layer are not sufficiently high (2 m for temperature and moisture, 3.4 m for wind speed), so the computed fluxes can only represent the averaged fluxes over a very small area of a few square kilometers. The averaged spacing between the ARM–CART stations (or Oklahoma Mesonet stations) is about...
50 km (or 20 km). To obtain averaged fluxes over a large area covered by these stations, the fluxes computed at each station should be representative of the surrounding area of at least 100 km². Over such an area, contributions to the fluxes come from motions of different scales, ranging from small-scale turbulence to large eddies in the boundary layer (Mahrt 1987). Turbulent motion is mainly responsible for fluxes over a small area, whereas fluxes over a relatively large area are often affected by large eddies. Large-eddy motion shows up as wind gusts, so that standard deviations of wind, as measured at ARM–CART stations, can be used to compute average fluxes for a relatively large area (Beljaars 1987; Beljaars and Holtslag 1991).

The second problem is related to errors in the vertical gradients of temperature and humidity measured over a small vertical distance (1 m) at the ARM–CART stations. Since the two measurement levels are about 1 and 2 m, respectively, above the ground, the true vertical gradient of temperature or moisture is often too small to be accurately detected, especially when the stratification is nearly neutral or slightly unstable. This may cause significant errors in the flux computation. On the other hand, the temperature and humidity differences between the atmosphere and soil are often sufficiently large so that instrument error is negligible. Since soil moisture measurements along with soil type and vegetation index data are available at ARM–CART, these data can be used to improve the accuracy of the flux computation.

Based on the above considerations, we intend to further improve the previous method of Xu and Qiu (1997) by using additional physical constraints: 1) a formula for the observed standard deviation of wind speed or temperature that can be used to compute nonlocal fluxes, including the contributions of multiple scale motions in the planetary boundary layer over an area on the order of 100 km²; and 2) a soil–vegetation layer model coupled with the atmospheric layer that can use soil data together with atmospheric measurements (even at a single level) to make the computed fluxes much less sensitive to measurement errors than those computed by the previous variational method. The goal is to improve the area-averaged representativeness and accuracy of the flux computation and to reduce the sensitivity of the computed fluxes to data errors.

As we will see in this paper, unlike all the previous methods, the new method does not require very accurate measurements of temperature and humidity differences between two atmospheric levels (by the SERBS), so it can be applied to the Oklahoma Mesonet in the near future when soil water content data become available. Thus, in comparison with the previous methods, the new method accomplishes an important step toward using conventional surface data to estimate surface fluxes.

The paper is organized as follows. Data and weather conditions are described in the next section. The new scheme is introduced in section 3 with detailed descriptions of the physical constraints used and the method of solution. The test results with ARM–CART data are presented and analyzed in section 4. Conclusions follow in section 5.

2. Data and weather conditions

a. Data

The observational data used in this study were collected by the surface Bowen ratio energy balance system installed at 10 ARM–CART stations, including the central facility (CF) and nine extended stations. These stations are located in Kansas and Oklahoma. The collected surface meteorological and soil data include wind speed, direction, and fluctuation (standard deviation) at one level; temperature, vapor pressure, and their fluctuations at two levels; air pressure near the ground; net radiation flux; and soil heat flux, soil temperature, and soil water content in the top 5-cm layer. Latent and sensible heat fluxes, estimated from application of the SERBS method to Bowen ratio measurements, are also recorded. These data are sampled continuously and averaged over 30-min intervals. The detailed descriptions of the SERBS and related instruments at the ARM–CART stations can be found in Xu and Qiu (1997) and Stokes and Schwartz (1994).

The new method developed in this paper is tested with SERBS data collected during 5–19 July 1994 at the 10 ARM–CART stations. Since the test results are similar for all the stations, in this paper we present the results for only the CF station in this paper. The area around CF consists of open fields, no outstanding obstacles, and covered by about 90% grass (0.4 m on average) and 10% weeds (0.6 m on average). The soil type is silty clay loam.

b. Weather conditions

Cloud amounts were recorded by field observers at the CF station for all cloud types. Formation and dissipation of each type, as well as their amount, and their relations to the weather types and precipitation, were also recorded. Cloud amounts are plotted in Fig. 1 for three basic types of clouds: cumulus, stratocumulus, and mid/high clouds. The soil water content (moisture fraction) and rainfall amounts are shown in Fig. 2. During the aforementioned 14-day period, there were 8 dry and 6 wet days. They were all typical summertime convective days, in which five convective storms passed over the CF station during the wet days. Other events included light and dense fog and light rain, as illustrated in Fig. 1. There was no significant rain before 13 July 1994, but then five rainfall events caused sharp increases in soil water content. There were also three very light rainfall events recorded by field weather observers at the CF station, but they were not detectable by the rain gauge and had no significant impact on soil moisture.
During the first eight days, the soil water content was depleted due to dry winds, steady solar radiative heating in the daytime, and lack of rainfall, as shown in Fig. 2. The soil water content was kept at a very low level, close to the wilting point in the last few dry days, even with the two light rainfalls during the morning of 7 July and the afternoon of 12 July. For the discussion that follows, the first eight days are considered the dry period, whereas the next six are considered the wet period.

During the dry period, there were always fractional mid to high clouds. Cumulus developed locally in the early morning, then evolved into stratocumulus in the afternoon, and mostly dissipated during the nighttime. This type of daily cycle of cloud variation was maintained during the entire dry period until heavy precipitation occurred on the evening of 13 July. During the wet period, the sky was mostly cloudy without diurnal variation. The variations of soil water content in the wet period were also very different from those in the dry period. These variations, as will be seen in the later sections, account for the differences in the computed fluxes between these two periods.

3. Physical constraints and method of solution

a. Similarity laws for wind and temperature profiles in the surface layer

Near the surface, profiles of horizontal wind speed and temperature can be described by the following similarity laws:

\[
U(z) = \frac{U_0}{\kappa} \left[ \ln \left( \frac{z}{z_{om}} \right) - \Psi_M \left( \frac{z}{L} \right) + \Psi_M \left( \frac{z_{om}}{L} \right) \right],
\]

(1)

\[
T(z) = T_s + \frac{\theta_v}{\kappa} \left[ \ln \left( \frac{z}{z_{om}} \right) - \Psi_H \left( \frac{z}{L} \right) + \Psi_H \left( \frac{z_{om}}{L} \right) \right],
\]

(2)
where $U_a$ is the friction velocity defined by $U_a^2 = \tau/\rho$ in association with the wind stress $\tau$ and air density $\rho$; $\kappa = 0.4$ is the von Kármán constant; $\theta_a$ is the flux temperature scale associated with the sensible heat flux $H = -c_p\rho U_a \theta_a$, where $c_p$ is the specific heat at constant pressure; $L = \left[U_{zs}(z)/kg\theta_a\right]$ is the Obukhov length; $T_s$ is the surface skin temperature; $\Psi_m$ and $\Psi_H$ are the stability functions; and $z_{om}$ and $z_{oh}$ are the roughness lengths for momentum transport and heat transport, respectively. For the unstable case ($\theta_a < 0$ or $L < 0$), the stability functions $\Psi_m$ and $\Psi_H$ are the same as those in (6) and (7) of Xu and Qiu (1997). For the stable case ($\theta_a > 0$ or $L > 0$), the stability functions are the same as those used in (8) and (9) of Xu and Qiu (1997).

Högström (1996) compared various published formulas for nondimensional wind and temperature profiles. He showed that the various formulas agree to within 10%–20% for $|z/L| \ll 0.5$. For unstable stratification, the various formulas for nondimensional temperature profiles continue to agree within this accuracy up to at least $-z/L = 2$. The various formulas for nondimensional wind profiles continue to agree for moderately strong unstable stratification up to $-z/L = 1$, but diverge after $-z/L > 1$ for very unstable stratification (see his Fig. 3). Since the measurement levels were low ($\leq 3.4$ m) and $z/L$ was rarely much below $-1$ for the ARM data used in this study, the formulas used in (1) and (2) are not affected by the profile divergence problem in the free-convection limit.

b. Effective roughness length

When area-averaged surface stress (or momentum flux) is computed by using the similarity law, the meaning of roughness length, $z_{om}$, used in the formulation should be redefined to consider the nonlocal effect. In this case, the traditional local roughness length should be replaced by a so-called effective roughness length. The local roughness length represents a geometric feature of a small area, and the effective roughness length represents the total effect of roughness and heterogeneous terrain averaged over a large area (Fiedler and Panofsky 1972; Mason 1988). According to Beljaars and Holtslag (1991), the effective roughness length accounts for both the averaged surface obstacles' geometry over the concerned area and the dynamic influence of surface characteristics on air parcels flowing over that area. The effective roughness length is about one order of magnitude larger than the local roughness length over grassland. The dynamic influence of surface characteristics on air parcels is closely related to large-eddy motions. During the nighttime, large-eddy motions are suppressed by stable stratification in the boundary layer, so the effective roughness length is small. During the daytime, as convective large-eddy motions develop vigorously, the effective roughness length is greatly enhanced.

The effective roughness length can be estimated in several ways (Garratt 1992). In this paper, the simple deviation method of Beljaars (1987) and Beljaars and Holtslag (1991) is used. Since air parcels in the surface layer always tend to follow the upstream terrain, standard deviations of winds and/or temperature contain information that records the effects of upstream terrain on air motion. According to the Monin–Obukhov similarity theory and the scaling laws for turbulent motion within the surface layer (Panofsky and Dutton 1984; Garratt 1992; DeBruin et al. 1993), the profiles of the normalized standard deviations of wind speed and temperature can be described by

$$\frac{\sigma_u}{U_a} = f_u(z/L)$$  \hspace{1cm} (3)

and

$$\frac{\sigma_T}{\theta_a} = f_T(z/L),$$  \hspace{1cm} (4)

where $f_u$ and $f_T$ are the universal functions of the stability parameter $z/L$.

In this paper, (3) is used for the case of a nonconvective surface layer and $f_u$ is given by (DeBruin et al. 1993)

$$f_u = 2.2(1 - 3z/L)^{0.3}, \quad \text{for } 0 \leq z/L \geq -1,$$

$$f_u = 2.5, \quad \text{for } z/L > 0.$$  \hspace{1cm} (5)

Here, (3) and (5) are applicable only up to moderately unstable conditions ($z/L \geq -1$). When $z/L < -1$, turbulence motions are driven primarily by thermal buoyancy, so $\sigma_u$ changes little with height within the depth of the local free convection, and reliable formulations are hardly available for $f_u$. In this case, (4) can be used in place of (3) with $f_u$ given by (Tennekes 1972)

$$f_T \approx -0.4(\kappa/(\xi B^{-1}) - 1), \quad \text{for } z/L < -1.$$  \hspace{1cm} (6)

In addition to the physical constraints described in the previous subsection, constraints (3) and (4) allow use of the observed standard deviations of wind speed or temperature (averaged every 30 min at the Oklahoma ARM–CART stations) to determine the effective momentum roughness length $z_{om}$. The effective roughness length $z_{oh}$ for heat flux can be related to $z_{om}$ by the following empirical formula (Thom 1972):

$$z_{oh} = z_{om} \exp(-\kappa B^{-1}),$$  \hspace{1cm} (7)

where $B^{-1} = (T_s - T_0)/\theta_a$ is the Stanton number, $T_s$ is the surface skin temperature as in (2), and $T_0$ is aerodynamic temperature at the height of the momentum roughness length $z_{om}$. In the limit approaching a smooth surface, (7) gives $z_{om} \rightarrow 0$, $T_s \rightarrow T_0$, and $z_{oh} \rightarrow z_{om}$. When the surface is not very smooth, $T_s$ can be significantly different from $T_0$, and thus $z_{oh}$ can be significantly different from $z_{om}$. Huband and Monteith (1986a,b) measured $T_s$ and $T_0$ under nearly neutral conditions and found that $T_s - T_0$ is about $1^\circ$C. According to Garratt
and Hicks (1973) and Brutsaert (1982), when the atmospheric stratification changes from very unstable to stable over flat grassland, $\theta_a$ can decrease from 0.5°C to $-0.5^\circ$C and $T_a - T_v$ can decrease from 3°C to $-3^\circ$C. This yields $B^{-1} = 6$, or $z_{0h} \approx 0.1 \ z_{hm}$ using (7), and this value is used in this paper for flat grassland.

c. Surface energy balance equation

As in Eq. (10) of Xu and Qiu (1997), the surface energy balance equation has the following form:

$$H + \lambda E = R + G,$$

(8)

where $H = -c_p \rho U_a \theta_a$ is the sensible heat flux; $\lambda E = -\lambda \rho U_a q_a$ is the latent heat flux, where $\lambda$ is the latent heat of evaporation; $G$ is soil heat flux; and $R$ is the net radiative flux. In principle, $\lambda E$ can be computed together with the moisture scale $q_a$ by using the similarity laws and profile method [see (1)–(3) of Xu and Qiu (1997)] with humidity measured at two or more vertical levels in the surface layer. In practice, as explained in Xu and Qiu (1997), the profile method requires very accurate measurements of humidity differences between any two measurement levels, especially when they are not far apart from each other. Humidity differences (between 0.96 and 1.96 m) can be accurately measured by SERBS at the ARM–CART stations, but the area-averaged moisture flux (or latent heat flux) may still not be well estimated by using the similarity law alone, especially when the land surface is covered by vegetation. Thus, unlike the situation for sensible heat flux, the biological processes in the soil–vegetation layer need to be considered in the moisture flux formulation, and this consideration will be critical if the atmospheric measurements are not sufficiently accurate or available only at a single level (such as the conventional measurements from the Oklahoma Mesonet).

The land surfaces at the ARM–CART stations are mostly covered by grasses. With dense vegetation coverage, moisture flux is governed not only by the atmospheric process in the surface layer but also by the biological processes in the soil–vegetation layer. The bulk effect of these coupled processes can be described by the Pemmán–Monteith evaporation equation (Monteith 1973; Hillel 1980):

$$
\lambda E = \rho C_p [q(T) - q_h \gamma/r_s - s(R + G)]/s + (1 + r_s/r_u) \gamma,
$$

(9)

where $s = \partial q/h\partial T$; $q_a$ is the saturation specific humidity at the measurement level of $q$; $\gamma = c_p / \lambda$; $r_s$ is the aerodynamic resistance to sensible and latent heat fluxes; and $r_u$ is the bulk canopy resistance to moisture flux—a function of solar radiation, leaf area index, and soil water potential [see (13)–(14)]. In (9), the moisture flux (or surface evaporation) is related to net radiative flux, soil heat flux, and moisture deficit near the ground.

d. Aerodynamic resistance and canopy resistance

The sensible heat flux can be related to the temperature difference between the air and ground in the following bulk form:

$$
-H/\rho C_p = (T(z) - T_a)/r_u,
$$

(10)

where $r_u$ is the aerodynamic resistance. Substituting $H = -c_p \rho U_a \theta_a$ with (2) into (10) gives

$$
r_u = \frac{1}{k \rho U_a} \left[ \ln \left( \frac{z}{z_{0h}} \right) - \Psi_h \frac{z}{L} \right] + \Psi_t \left( \frac{z}{L} \right).
$$

(11)

This formulation will be used to compute the aerodynamic resistance $r_u$. The aerodynamic resistance to latent heat flux is assumed to be the same as the aerodynamic resistance to sensible heat flux computed by (11).

The canopy resistance $r_c$ characterizes water transportation from the total canopy, which is a bulk measure of moisture flux resistance for transpiration from canopy leaves’ stomata to the air around them. This canopy resistance estimation grossly simplifies the real situation, but it provides the best available simple description of moisture transport near the surface (Garratt 1992). The value of $r_c$ is controlled by the following factors: 1) visible solar radiation (0.4–0.7 $\mu$m) received by leaves, 2) air humidity and temperature around leaves, and 3) the soil water content (Deardorff 1978; Sellers et al. 1986; Noilhan and Planton 1989).

Following Dickinson (1984) and Noilhan and Planton (1989), the above factors 1 and 2 can be combined into a single parameter, the unstressed canopy resistance,

$$
r_c^* = \frac{r_c}{\text{LAI}} F_1 F_2 F_3 F_4^{1/1},
$$

(12)

where $r_{min}$ is the minimum resistance (50 s m$^{-1}$ for grass), LAI is the leaf area index, and an estimated value of LAI = 2 is used in this paper (for the ARM CF station covered by about 90% grass and 10% weeds). Here, $F_1$, $F_2$, $F_3$, and $F_4$ are factors related to solar radiation, temperature, and moisture, respectively. The solar radiation factor is given by (Dickinson 1984)

$$
F_1 = \frac{1 + f}{1 + r_{min} r_{max}},
$$

where $r_{max}$ is the maximum resistance (800 s m$^{-1}$ for grass), $f = 1.1 (S_f / S_u) \text{LAI}^{-1}$, $S_f$ is the incident solar radiative flux absorbed by canopy, $S_u$ is the upper-limit value of $S_f$, and $S_u$ equals 120 W m$^{-2}$ is used for grass in this paper. According to Monteith (1976), $S_f$ can be estimated by $S_f = \beta \chi e^{-\beta \chi}$, where $\chi$ is the extinction coefficient of canopy to shortwave radiation [$\chi = 0.6$ is used for grassland in the growing season (Ripley and Redmann 1976)] and $\beta = 48\%$ is used for the percentage of the visible part (photosynthetically active part over wavelengths 0.4–0.7 $\mu$m) of the downward solar radi-
ative flux $S_o$ on top of the grass cover. In this paper, the downward solar radiative flux is computed by $S_o = S_n(1 - \alpha_s)^{-1}$, where $\alpha_s = 0.2$ is the albedo of the grass cover and $S_n$ is the net solar radiative flux. Since $S_n$ is not directly measured at the ARM CF station, the empirical formula of Paltridge and Platt (1976) is used to compute the net longwave radiative flux $L_o$ from the observed air temperature $T_o$, computed surface skin temperature $T_s$, and estimated cloud-base temperature (the details are omitted here). Then the net solar radiative flux is obtained by $S_n = R - L_o$. The moisture and temperature factors are estimated by (Sellers et al. 1986 and Dickinson 1984)

$$F_z = 1 - 0.0025[e_s(T_s) - e(z)],$$

and

$$F_t = 1 - 0.0016[298.0 - T(z)]^2,$$

where $e_s(T_s)$ is the saturation water vapor pressure at the surface skin temperature $T_s$, $e(z)$ is the water vapor pressure, and $T(z)$ is the temperature at the measurement level $z$.

The canopy resistance is stressed (with an increased value) when the soil is not saturated with water. In this case, the above factor 3, that is, the effect of the soil water, should be taken into account. This factor can be parameterized in several ways, such as by using soil water content (Feddes et al. 1976; Noilhan and Planton 1989), leaf water potential (Pinty et al. 1989), or soil water potential (Feddes et al. 1976; Sellers et al. 1986), Sceicz and Long (1969) and Turner (1974) measured the response of canopy resistance to soil water potential or leaf water potential. They found that depending on the species of vegetation and environmental conditions (mainly environmental light intensity), the canopy resistance could increase gradually as leaf water potential decreases, remain unchanged within a rather broad range of water potential, or remain constant within a range and then increase abruptly as water potential falls below a critical value. According to Sceicz and Long (1969), the canopy resistance for short grasses and crops, which is better labeled ground-cover resistance, maintains its unstressed (minimum) value until the soil water potential decreases to $-4$ bars; then the resistance increases almost linearly as soil water potential decreases to the wilting point at $-12$ bars, although there is no experimental data beyond the wilting point. The permanent wilting soil water potential varies with species, ranging from $-10$ to $-20$ bars. According to Turner (1974), the permanent wilting soil water potential for the short vegetation is lower than that for tall vegetation. Therefore, the wilting soil water potential for grasses should be close to the lower limit ($-20$ bars), and the ground-cover resistance can be parameterized in terms of soil water content as follows:

$$r_s = r_s^\varphi,$$

for $\varphi > \varphi_c$,

$$r_s = r_s^\varphi\left[1 + \left(\frac{r_{\text{max}}}{r_s^\text{LAI}} - 1\right) \times (\varphi - \varphi_{\text{wilting}})/(\varphi_{\text{wilting}} - \varphi_c)\right],$$

for $\varphi_{\text{wilting}} < \varphi < \varphi_c$,

$$r_s = r_{\text{max}} / \text{LAI},$$

for $\varphi \leq \varphi_{\text{wilting}}$, (13)

where $\varphi_{\text{wilting}}$ is the wilting soil water potential and $\varphi_c$ ($=-4$) is the critical soil water potential. By definition, the soil water potential is related to the soil water content $w$ (directly measured moisture fraction) by

$$\varphi = \varphi_c \left(\frac{w}{w_s}\right)^{-b},$$

where $\varphi_c$ is the saturated soil water potential, $w_s$ is the saturated soil water content, and $b$ is a constant. These constant parameters depend on soil texture. For the ARM CF station, the soil texture is characterized by silty clay loam, so $\varphi_c = 0.356$ m, $w_s = 0.477$, and $b = 7.75$ are used in this paper according to Clapp and Hornberger (1978, see their Table 2).

The above resistance formulations neglect the dew-related processes. Dew is a common phenomenon in the early morning of summer season but is difficult to handle. When dew forms on the leaves, the leaf surface should be treated as water, and the evaporation process should be similar to that from a body of water. If the processes within the canopy (short canopy) can be ignored, then the total resistance should be the aerodynamic resistance only. Dew usually forms under a very stable condition during nighttime. In this case, the aerodynamic resistance is very large, implying that both sensitive and latent heat fluxes are very small. Thus, the changes of these fluxes due to the influence of dew should also be very small, especially in comparison with the daytime strong surface fluxes. Based on this consideration, the dew-related complication is ignored in this study.

### e. Method of solution

The input observational data include wind speed $U(z)$ and its standard deviation $\sigma_U(z)$ at $z = 3.4$ m, air temperature $T(z)$ and its standard deviation $\sigma_T(z)$, vapor pressure $e(z)$, and atmospheric pressure averaged at $z = 1.5$ m; net radiative flux $R$; soil heat flux $G$; soil water content $w$; and cloud coverage. With these input data, the six unknown parameters, $U_\vartheta^*$, $\theta_\vartheta^*$, $z_{\text{con}}$, $z_{\text{oh}}$, $T_o$, and $\Lambda E$, can be solved iteratively from (1)–(3) or (4) and (7)–(9). The iterative procedure starts with first guesses of $U_\vartheta^* = 0.1$ m s$^{-1}$, $z_{\text{con}} = 0.01$ m, $\theta_\vartheta^* = 0.1$ K, and $T_o = 300$ K. The iteration cycle consists of the following steps:

1) Compute the Obukhov length $L = [U_\vartheta^* T(z)]/\kappa g \theta_\vartheta^*$. 

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2) Solve for \( U_h \) and \( z_{om} \) from (1) and (3) [or (4)] under the constraint of \( z_{om} \geq 0.01 \) m.
3) Compute \( z_{om} \) from (7) and \( \Delta E \) from (9).
4) Compute \( H \) from (8) and \( \theta_a \) from \( H = -c_p \rho U_h \theta_a \).
5) Compute \( T_r \), from (2) and check the convergence by
\[
J = \alpha_{\theta} |\theta_a^{(n)} - \theta_a^{(n-1)}| + \alpha_{u} |U_h^{(n)} - U_h^{(n-1)}| + \alpha_{T} |T_r^{(n)} - T_r^{(n-1)}| \leq \varepsilon, \tag{15}
\]
where \( \alpha_{\theta} = 10 \text{ m}^{-1} \text{s} \), \( \alpha_{u} = 1.0 \text{ (K)}^{-1} \), \( \alpha_{u} = 10.0 \text{ m}^{-1} \), and \( \alpha_{T} = 0.1 \text{ (K)}^{-1} \) are chosen to be inversely proportional to the maximally tolerated errors (multiplied by \( \varepsilon \)) for their respectively associated variables, the superscript \( n \) denotes the value at the \( n \)th cycle of iteration, and \( \varepsilon = 10^{-4} \) is the convergence criterion used in this paper. In step 2, the constraint of \( z_{om} \geq 0.01 \) is used to ensure that the effective roughness length is not smaller than the local roughness length (that is, 0.01 m for flat grassland). Although a rigorous convergence proof seems difficult for (15), all the solutions obtained with the above iterative procedures converge within the accuracy of \( J = 10^{-4} \) as in (15).

4. Test results and analysis

a. Fluxes computed by the new method and BREB method

The method described in the previous section is used to compute the nonlocal fluxes from the SERBS data collected during 5–19 July 1994 at the CF station (as well as other stations). Since direct eddy correlation measurements of sensible or latent heat fluxes are not available yet at the ARM–CART stations, the computed fluxes can be compared only with the fluxes obtained by the conventional BREB method. Smith et al. (1992) compared the fluxes obtained by the BREB method with the direct eddy correlation measurements collected during the First International Satellite Land Surface Climatology Project Field Experiment, and they found no serious problem for the consistency of the fluxes obtained by the two methods (except for the well-known fact that the BREB method becomes computationally unstable and produces spurious spikes in the computed fluxes when the Bowen ratio is in the vicinity of \(-1\)). In this paper, the fluxes computed by the BREB method are denoted by BR, while the nonlocal fluxes computed using our new method are denoted by NLC. As shown in Figs. 3a,b, the overall agreements between the BR fluxes and the NLC fluxes are quite good, and spurious spikes in the BR fluxes are eliminated by the new method. Note also that the new method uses only a single level of atmospheric data, and thus the results are not affected by the errors in temperature and humidity gradients determined from measurements across small vertical distances (1 m in the SERBS).

By comparing Fig. 3a with Fig. 3b, we can see that the daytime value of the Bowen ratio changed dramatically as the wet period started with heavy precipitation in the evening of 13 July. The daytime Bowen ratio was in the range of 3–4 during the dry period but reduced to about 0.5 during the wet period. This implies that the incoming solar radiative forcing is balanced mainly by sensible heat fluxes on a dry day but is balanced largely by latent heat flux on a wet day. During dry days or over dry land, the surface evaporation is severely limited by lower soil water content. As the latent heat flux is reduced, the sensible flux is enhanced according to the surface energy balance relation (8). The increase of sensible heat flux is caused mainly by high surface temperature. During wet days or over wet land, surface evaporation is enhanced by high soil water content. The enhanced surface evaporation causes the surface temperature to decrease and thus reduces the sensible heat flux. Furthermore, wet soil has a greater soil heat capacity than dry soil and, hence, does not tend to warm as rapidly. Consequently, turbulence is weaker on wet days (or over wet land) than for a dry day (or over dry land). This feature was also observed during the field experiment at Boardman, Oregon, (Doran 1992). In particular, turbulence was found to be stronger over semiarid rangeland than over an irrigated wheat field, and the Bowen ratio \( H/AE \) was close to 0.5 over wet land and as large as 5 over dry land. Clearly, soil water content has strong impact on the surface fluxes and related turbulent processes, although the quantitative aspect may be complicated by other factors remaining to be studied.

b. Effective roughness length and nonlocal fluxes

The effective roughness length is obtained as a byproduct of the flux computation. The temporal variation of the effective roughness length is plotted together with the measured standard deviation for wind speed, \( \sigma_u \), in Fig. 4. As shown, the effective roughness length generally follows the variation of \( \sigma_u \) with high values in the daytime and lower values in the nighttime. The maximum of the effective roughness length can reach a half meter, a value much larger than the local roughness length (about 1.0 cm) over grassland. The daytime- and nighttime-averaged effective roughness lengths are shown in Fig. 5 for 14 days. Since the terrain is flat with no obvious obstacles surrounding the ARM–CART station, wind and temperature fluctuations (measured by standard deviations) detected in the daytime are caused mainly by unstable convective motions in the boundary layer. During the nighttime, the boundary layer stratification is stable, large-eddy motions are suppressed, and the effective roughness length reduces to nearly the same value as the local roughness length. This causes the daily cycle in the variation of the effective roughness length shown in Figs. 4 and 5. The daily cycle is especially strong for the dry period (Fig. 5). During the dry period, a hot and dry day is usually followed by a
very stable night, and the effective roughness length is greatly enhanced by vigorous convective motions developed in the daytime. During the wet period, however, severe storms occurred in the nighttime (see Figs. 1 and 2), which caused strong wind fluctuations (measured by $\sigma_u$) and enhanced the effective roughness length (Figs. 4 and 5). For the total 14 days, the half-daily averaged effective roughness length ranges from 1–3 cm (nighttime) to 5–12 cm (daytime). The averaged daytime effective roughness length is close to the value (10 cm over grassland) obtained by Beljaars and Holtslag (1991) at Cabauw, Netherlands, during the MESOGERS-84 field experiments.

The method described in the previous section is also used to compute the local fluxes with a fixed local roughness length ($\zeta_{om} = 1$ cm). In this case, the standard
deviation of wind or temperature and the similarity law in (3) or (4) are not used, and the computed fluxes are denoted by LC. The correlation diagram between the LC fluxes and the BR fluxes is shown in Fig. 6, whereas the correlation diagram between the NLC fluxes (computed in the previous subsection) and the BR fluxes is shown in Fig. 7. Note that the BR fluxes are computed from the local measurements of SERBS without using the observed standard deviations of wind speed and temperature, so they are local fluxes. This explains why the BR fluxes are more closely correlated to the LC fluxes in Fig. 6 than the NLC fluxes in Fig. 7. The correlation diagram between the NLC fluxes and the LC fluxes is plotted in Fig. 8. Most of the points are distributed closely along the diagonal line in the vicinity of the origin. The points become scattered and deviate from the diagonal line as the flux values increase during daytime. Note that the computed daytime NLC fluxes of sensible and latent heat are both positive, so they are both constrained by the local value of $R + G$ measured by the SERBS and used in (8). This explains why the computed NLC fluxes are not increased as much by daytime large-eddy activities, although the effective roughness is enhanced.

c. Surface skin temperature

Surface skin temperature, that is, the surface radiative temperature, is the temperature at which the surface emits infrared radiation outward. Aerodynamically, this temperature is sometimes referred to as the temperature at the heat losing level $z_{ok}$ [see (2)]. As an important parameter in a surface–atmosphere system, the surface skin temperature determines loss or gain of sensible heat from the surface and relates the surface characteristics to fluxes through energy balance near the surface. This temperature, however, is hard to measure in the presence of vegetation, although it can be either estimated by an infrared radiometer or derived from wind and temperature profiles near the surface (Huban and Monteith 1986a,b). In this paper, the surface skin temperature is a computational parameter.

Figure 9 shows that the retrieved surface skin temperature is about 4°–5°C higher than the air temperature (measured at $z = 1$ m) during most of the day, 4°C lower than the air temperature on noncloudy nights, and only slightly lower than the air temperature on cloudy or rainy nights. For the night of 15 July, however, the retrieved surface skin temperature is higher than the air temperature by 2.5°C, and this is rather unusual. We checked the energy balance for that night and found that the soil heat flux was larger than the upward net longwave radiative flux by about 10–20 W m$^{-2}$. This implies that the soil heat flux overly compensated the heat loss caused by infrared radiation at the ground surface, so the surface skin temperature could be higher than the
air temperature. This unusual situation might be related to changing weather conditions. In the early night, a thunderstorm passed the ARM–CART site with precipitation, then the sky cleared somewhat followed by dense fog during the early morning of 16 July. The dense fog reduced the net infrared radiation from the surface, allowing the surface skin temperature to become higher than the air temperature.

d. Sensitivity test

Although the method developed in this paper has been tested with the data collected at the ARM–CART stations, it is not clear whether the method can still give reliable estimates of fluxes when the air temperature and humidity data (such as those obtained at the Oklahoma mesonet stations) are not as accurate as those measured by the SERBS at the ARM–CART stations. To examine the sensitivities of the computed fluxes to errors in temperature and humidity measurements, errors of ±0.3°C in air temperature and ±10% in relative humidity are added to the data. Fluxes computed from the original data and error-contaminated data are denoted by \( \hat{F} \) and \( \hat{\bar{F}} \), respectively. The relative rms error is defined by

\[
\text{Relative rms error} = \frac{\sqrt{\sum (\hat{F} - \hat{\bar{F}})^2}}{\sqrt{\sum \hat{F}^2}}
\]
where the summation is made for all the 14 × 48 time levels (30 min apart for the 14 days from 5 to 19 July 1994). As listed in Table 1, when the air temperature data has a positive error of 0.3°C, the latent heat flux is overestimated by 3.8% in the dry period and by 2.4% in the wet period; the sensible heat flux is underestimated by 1.6% in dry period and by 4.5% in the wet period. The results reverse when the air temperature data has a negative error of −0.3°C. When the humidity data has a positive error of 10%, the sensible heat fluxes are overestimated and the latent heat fluxes are underestimated in both the dry and wet periods. These results also reverse when the moisture data error changes sign. Clearly, the relative rms errors in Table 1 are all within ±10%. Note that the relative rms errors for the fluxes (and especially sensible heat fluxes) computed in the wet period are larger than those in the dry period. This suggests that the method is more accurate on dry days than on rainy days.

When the original variational method was developed by Xu and Qiu (1997), it was tested with the same ARM data as used in this paper except for the period of 10–19 July (instead of the entire 14 days from 5 to 19 July). The sensitivities of the computed fluxes to data errors were also examined. As shown in Table 1 of Xu and Qiu (1997), adding a small error of 0.05°C to the air temperature data caused an rms error of 8.5 W m⁻² in the computed sensible heat fluxes and an rms error of 4.7 W m⁻² in the computed latent heat fluxes. Adding a very small error of 0.001% to the air humidity data caused an rms error of 4.4 W m⁻² in the computed sensible heat fluxes and an rms error of 8.2 W m⁻² in the computed latent heat fluxes. The rms amplitudes of the computed sensible heat fluxes and latent heat fluxes were about 100 W m⁻² (see Figs. 2 and 3 of Xu and Qiu 1997), so the relative rms errors caused by adding a small error of 0.05°C to the air temperature data can be estimated by \( \text{RE} \approx 8.5\% \) for the computed sensible heat fluxes and \( \text{RE} \approx 4.7\% \) for the computed latent heat fluxes. The relative rms errors caused by adding a small error of 0.001% to the air humidity data can be estimated by \( \text{RE} \approx 4.4\% \) for the computed sensible heat fluxes and \( \text{RE} \approx 8.2\% \) for the computed latent heat fluxes. These relative rms errors are larger than those in Table 1, even though they were caused by much smaller errors in the air temperature and humidity data. Thus, the fluxes computed by the new method are much less sensitive to measurement errors than the fluxes computed by the previous variational method.

### Table 1. Sensitivities of the computed fluxes (H and \( \lambda E \)) to errors in the temperature and moisture data, where RE is the relative rms error defined in (22).

<table>
<thead>
<tr>
<th>Data error</th>
<th>Dry period</th>
<th>Wet period</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>RE for H</td>
<td>RE for ( \lambda E )</td>
</tr>
<tr>
<td>+0.3°C in ( T )</td>
<td>−1.6</td>
<td>3.8</td>
</tr>
<tr>
<td>0.3°C in ( T )</td>
<td>1.6</td>
<td>−3.7</td>
</tr>
<tr>
<td>+10% in ( q )</td>
<td>2.8</td>
<td>−6.6</td>
</tr>
<tr>
<td>−10% in ( q )</td>
<td>−2.9</td>
<td>6.8</td>
</tr>
</tbody>
</table>

5. **Conclusions**

An air–soil layer coupled method is developed to compute nonlocal surface fluxes of sensible and latent heat from data collected by SERBS at the Oklahoma ARM–CART stations. In comparison with the method of Xu and Qiu (1997), this coupled method uses two types of additional data and physical constraints: 1) the observed standard deviations of wind and temperature together with the similarity laws for these standard deviations and 2) the soil water content together with a soil–vegetation layer model. With the first type of additional data and physical constraints, the method can estimate the effective roughness length that accounts for
both the averaged surface obstacles’ geometry and the dynamic influence of surface characteristics on air parcels flowing over a nonlocal area of $O(100 \text{ km}^2)$. The computed fluxes are nonlocal as they contain the contributions of large-eddy motions. With the second type of additional data and physical constraints, the method becomes much less sensitive to measurement errors than the previous variational method and can be used even when the atmospheric temperature and humidity data are available only at a single level, as shown in this paper.

The method is tested with data collected by SERBS at the Oklahoma ARM–CART stations during 5–19 July 1994. The computed nonlocal fluxes are compared with the fluxes computed by the BREB method. The overall agreements are quite good, and spurious spikes in the fluxes computed by the BREB method are eliminated by the new method. The results show that the computed surface sensible heat flux is larger than the latent heat flux for the dry days, and the Bowen ratio can reach about 3–4 in the daytime. For the rainy days, however, the Bowen ratio is about 0.5. This suggests that the incoming solar energy near the surface was largely balanced by the latent heat flux during the wet days but largely balanced by the sensible heat flux during the dry days.

As a byproduct, the surface skin temperature is retrieved by the present method. The averaged skin temperature is about 3–5°C higher than the air temperature near the ground (1 m high) in the daytime, but is 1–4°C lower than the air temperature in the nighttime with clear sky. On raining or cloudy days, the difference between the surface skin temperature and air temperature is small. The effective roughness length is also computed as a byproduct. The computed effective roughness length shows strong daily variations ranging from 1 to 3 cm in the nighttime and 5 to 14 cm in the daytime. Since the terrain is flat and covered by short vegetation, the local roughness length can be estimated approximately by the fixed value of 1 cm. Since differences between the effective roughness length and the local one can be quite large, it is necessary to consider the effect of large eddies in order to compute the nonlocal fluxes, though the method developed in this paper is still limited by the local nature of the soil heat flux data and net radiative flux data (measured by the SERBS).

The observed standard deviations of wind and temperature contain large-eddy information on the scale of about ten or several tens of kilometers, but virtually no information on the mesoscale flows. The numerical simulation of Zhong and Doran (1995) showed that the contributions of mesoscale flows to the vertical fluxes of sensible and latent heat are very small in the surface layer. The contributions of large-eddy motions of different scales to surface fluxes vary with atmospheric conditions. It is clearly demonstrated here that important information concerning the effective roughness length is contained within the observed variances of wind and temperature [Eqs. (3)–(4)]. Numerical weather prediction models could significantly benefit from determination of effective roughness lengths; however, they generally use only specified local roughness length scales. Models carrying predictive expressions for turbulence variances (e.g., second-order closure models) should, similar to the manner demonstrated in this paper, be capable of extracting effective roughness lengths from their turbulent moment information.

The data used in this paper include soil heat flux and net radiative flux directly measured by the SERBS. These data are routinely available at the Oklahoma ARM–CART stations but not at the Oklahoma Mesonet stations and other conventional surface stations. The Oklahoma Mesonet has much higher spatial resolution than the ARM–CART stations, and most of the surface stations over the U.S. continent and around the world are conventional. It is thus desirable to extend the current method and make it applicable to data collected at conventional surface stations. As we have seen in this paper, the current method does not actually use the temperature difference and humidity difference (though accurately measured by the SERBS) between two atmospheric levels, so the method should be applicable to the Oklahoma Mesonet in the near future when soil water content data become available. In particular, since the net solar radiative flux is directly measured by the Oklahoma Mesonet and the net longwave radiative flux can be estimated by using the empirical formula of Patridge and Platt (1976) as in this paper, the current method will require only one additional formula to estimate soil heat flux and, in this case, the formula of Al Nakshaband and Kohnke (1965) should be suitable. Furthermore, it is also possible to extend the current method to use conventional surface station data without soil water content measurement. The related technique is under our investigation.

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