Modelling the local surface exchange over a grass-field site under stable conditions

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SUMMARY

Observations of near-surface variables are used with the Met Office Surface Exchange Scheme to calculate the exchange of heat and moisture between the surface and the atmosphere for a grass-field site at the Met Office, Cardington, UK. Different methods for modelling the stability near the surface are investigated with the surface temperatures and turbulent heat fluxes compared with data collected from the site. The results show that, whilst using Monin–Obukhov similarity theory can give good results in stable conditions, the form of the stability functions can significantly affect the accuracy in certain stability regimes. The use of log-linear stability functions can cause the turbulent fluxes of heat and moisture to cut off at stabilities which are too close to neutral compared with the observations. The results also show that unrealistically high surface temperatures are obtained in clear-sky nocturnal conditions with the standard surface energy balance. An alternative energy balance which includes a canopy of vegetation above the underlying soil gives better agreement with the observations in these conditions.

KEYWORDS: Surface energy balance Surface exchange

1. INTRODUCTION

The fluxes of momentum, heat and moisture between the surface of the earth and the atmosphere can be considered as the bottom-boundary condition to atmospheric models. For many idealized experiments with, for example, large-eddy simulation or cloud-resolving models that are used to study the physical processes within the atmosphere, these fluxes are considered to be constant. However, for both climate modelling and operational numerical weather prediction (NWP), we are often interested in near-surface variables where the interaction with the surface is crucial and must therefore be parametrized accurately.

Within a grid box over land in an operational NWP model, there are likely to be many different types of land use which will interact with the atmosphere in a complex way. From the surface-exchange schemes in such models we require fluxes of momentum, heat and moisture which are representative of the grid-box average so that the atmosphere evolves in the correct way. However, forecasts are often required for specific locations where the local surface will have a significant influence on near-surface variables and so any observational data consisting of point measurements which are used for verification may not be consistent with the model output. It is therefore difficult to critically assess surface-exchange schemes for operational NWP models.

Soil moisture plays an important role in the exchange of both heat and moisture between the surface and the atmosphere. For example, during periods of low rainfall, the effects of soil moisture become critical as the soil dries out and evapotranspiration becomes limited. In NWP, incorrectly parametrizing the soil moisture in such circumstances can lead to significant errors in forecast variables such as the screen-level temperature. However, during winter in the UK there is sufficient rainfall to prevent soil-moisture control on the transpiration from vegetation. Therefore, to minimize the impacts of these feedbacks, this study has concentrated on a data period during northern hemisphere autumn and winter when there was adequate soil moisture at the field site.

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This study uses a surface-only model in order to remove the interaction and particularly feedbacks between the surface and the atmosphere, therefore concentrating on the performance and behaviour of the surface scheme. This model uses observed radiation and rainfall, along with atmospheric variables of wind, temperature and humidity at a reference height, as forcing data for the surface-exchange parametrization. This means that the response of the surface is constrained to the observed atmospheric variables, but enables us to investigate the surface scheme without the complication of feedbacks between the surface and the atmosphere. Any errors larger than the uncertainty of the observations must be attributed to the surface scheme.

This paper will assess the Met Office Surface Exchange Scheme (MOSES) (Cox et al. 1999) for predicting surface temperature and turbulent heat transfer. The behaviour of the scheme will be compared against detailed surface observations made at a grass-field site at the Met Office, Cardington, near Bedford, UK. Alternative transfer coefficients for turbulent heat exchange which are more appropriate for the exchange from a local surface will also be investigated. Additionally, an alternative surface energy balance and its affect on the surface temperatures predicted by the model will be considered.

2. Model Description

The surface-only model calculates the exchange of momentum, heat and moisture between the land surface and the atmosphere, i.e. it calculates the bottom-boundary condition for NWP over land. The model is based upon MOSES, but has some alterations to the scheme which have been made so that we can assess the performance of the model for predicting the local surface exchange over vegetation. These changes are described later. The surface-exchange scheme uses atmospheric data of air temperature, humidity and wind speed at a level within the surface layer which is specified by the user (for this study the height was set to 10 m). It also uses values of the radiation and precipitation which reach the surface in order to solve the surface energy-balance equation:

\[ R_n = H + \lambda E + G, \]

(1)

where \( R_n \) is the net radiation at the surface, \( H \) is the turbulent heat flux, \( \lambda E \) is the turbulent moisture flux, with \( \lambda \) the latent heat of vaporization of water, and \( G \) is the soil-heat flux. The net radiation is defined to be positive towards the surface and the fluxes are defined to be positive away from the surface.

The radiation balance at the surface can be considered in two components, the solar radiation and the long-wave radiation. The present model can be integrated in two modes, either using net solar and net long-wave radiation, or using the downward components of both the solar and long-wave radiation. For this study, the downward components of the radiation have been used and the upward components have been calculated. The upward solar radiation is diagnosed using a constant surface albedo which is specified at the beginning of the model run. The upward long-wave radiation is parametrized using

\[ L_{w\uparrow} = \epsilon_s \sigma T_s^4 + (1 - \epsilon_s) L_{w\downarrow}, \]

(2)

where \( L_{w\uparrow} \) is the upward long-wave radiation, \( L_{w\downarrow} \) is the downward long-wave radiation, \( T_s \) is the surface temperature, \( \epsilon_s \) is the surface emissivity and \( \sigma \) is Planck’s constant.
The turbulent fluxes of heat and moisture are calculated using the equations

\[ H = \rho c_p C_H U \Delta \theta = \frac{\rho c_p \Delta \theta}{r_A}, \]  
\[ \lambda E = \rho \lambda C_E U \Delta q = \frac{\rho \lambda \Delta q_s}{r_A + r_S}, \]  

where \( \rho \) is density, \( c_p \) is specific heat capacity at constant pressure, \( C_H \) and \( C_E \) are the transfer coefficients for heat and moisture, respectively, \( U \) is the wind speed at the reference level, \( \Delta \theta \) is the potential-temperature difference between the surface and the reference level, \( \Delta q \) is the specific-humidity difference between the surface and the reference level, \( \Delta q_s \) is the difference between the saturated specific humidity at the surface and the specific humidity at the reference level, \( r_A \) is the aerodynamic resistance and \( r_S \) is the surface resistance.

In the Unified Model of the Met Office (Cullen 1993) the transfer coefficient is parametrized using explicit functions of the bulk Richardson number

\[ C_H = \frac{1}{U r_A} = \frac{C_{Hn}}{1 + Ri_b / Pr}, \]  

where \( C_{Hn} \) is the transfer coefficient in neutral conditions and \( Ri_b \) is the bulk Richardson number; \( Pr \) is a form of the turbulent Prandtl number and is calculated using

\[ Pr = \frac{\log \left( \frac{(z_1 + z_{0m})/z_{0h}}{(z_1 + z_{0m})/z_{0m}} \right)}{\log \left( \frac{(z_1 + z_{0m})/z_{0h}}{(z_1 + z_{0m})/z_{0m}} \right)}, \]

where \( z_1 \) is the height of the first model level, \( z_{0m} \) is the roughness length for momentum and \( z_{0h} \) is the roughness length for temperature. These atmospheric functions are chosen so that turbulence is sustained in very stable conditions. The justification for maintaining the turbulence in very stable conditions in a NWP model is that it is unlikely that the turbulence will cease over all of the types of land use within a grid box, which even for mesoscale modelling is of the order of \( 10 \times 10 \) km, i.e. the transfer coefficients notationally take account of subgrid heterogeneity.

In contrast, this study considers the surface exchange over homogeneous vegetation cover and so a maintained turbulent flux produced by the heterogeneous nature of a grid box will not be suitable for our purposes. For this reason the present model will compare the turbulent fluxes using the scheme from the Unified Model of the Met Office for calculating the aerodynamic resistance, along with standard Monin–Obukhov similarity theory (e.g. see Garratt 1992). Two stable stability functions are considered: log-linear functions

\[ \Psi_M(z/L) = \Psi_H(z/L) = 5z/L, \]  

and those taken from Beljaars and Holtslag (1991),

\[ \Psi_M(z/L) = z/L + 0.67(z/L - 14.3) \exp(-0.35z/L) + 9.5, \]  
\[ \Psi_H(z/L) = (1 + 2z/3L)^{3/2} + 0.67(z/L - 14.3) \exp(-0.35z/L) + 8.5, \]

where \( \Psi_M \) is the stability function for momentum, \( \Psi_H \) is the stability function for heat, \( z \) is height and \( L \) is the Obukhov length. The stability functions for moisture are taken to be the same as those for heat since they are both advected and diffused by the same flow. The roughness length for temperature and moisture are both taken to be a tenth of
the roughness length for momentum which agrees with, for instance, Garratt (1992) or Hopwood (1995).

A detailed description of MOSES can be found in Cox et al. (1999) so only a brief description will be given here. In MOSES the surface exchange is a Penman–Monteith surface energy balance with a diagnosed ‘skin’ temperature. It incorporates a four-layer model for soil moisture and soil temperature in order to introduce moisture dependent soil thermal properties and soil-water phase change. The soil hydraulic properties are described by ‘effective’ soil hydraulic characteristics derived from the mean fractions of sand, silt and clay within an NWP model grid box using the regression relations of Cosby et al. (1984). Soil-water phase changes are included via a parametrization based upon the ‘effective’ capacity approach of Williams and Smith (1989), where the maximum unfrozen soil water at a given temperature is derived from the soil-water suction curve (Black and Tice 1988). The stomatal resistance for vegetation is dependent on temperature, humidity deficit, photosynthetically active radiation, soil moisture content and CO₂ concentration. Since we are only considering night-time, when there is no photosynthetically active radiation, the stomatal resistance becomes large and hence the evapotranspiration is negligible.

Given the net radiation and the turbulent fluxes of heat and moisture, the surface energy balance can be used to diagnose the soil-heat flux at the surface. This soil-heat flux can then be used as the top-boundary condition for the solution to the diffusion equations,

\[
\frac{\partial T}{\partial t} = \frac{1}{C} \left[ \frac{\partial}{\partial z} \left( k \frac{\partial T}{\partial z} \right) \right], \tag{10}
\]

\[
G = -k \frac{\partial T}{\partial z}, \tag{11}
\]

that can then be solved for the surface and soil temperatures, where \( T \) (K) is temperature, \( C \) (J m\(^{-3}\)K\(^{-1}\)) is the volumetric heat capacity of the soil and \( k \) (J s\(^{-1}\)m\(^{-1}\)K\(^{-1}\)) is its thermal conductivity. The bottom-boundary condition required in order to solve these equations is provided by a zero heat-flux condition.

To solve the diffusion equations in MOSES, four soil levels have been chosen as a compromise between accuracy and computational expense, with the depths of these four levels being partly determined by the hydrology scheme to ensure that there is a model level at a depth of 1 m. However, when compared to the analytical solution for a sinusoidal surface soil-heat flux with a diurnal period, the four soil levels used in MOSES give inaccuracies in both the amplitude and phase of the solution. Therefore, since this study does not have the same computational constraint on computer usage as an operational NWP model, a more accurate nine-level soil scheme is used. The nine levels are staggered in such a way as to optimally capture the form of the exponential decay of the temperature variations with depth in the soil, whilst retaining a level at a depth of 1 m. Details of the nine-level scheme are given in the appendix.

Best (1998) showed that when using the standard surface energy balance (1) to represent vegetation in clear-sky, low-wind-speed conditions, the observed night-time difference between the surface temperature of the vegetation and the screen temperature of up to 10 K could not be achieved. The addition of a vegetation canopy to the surface energy balance allowed these temperature differences to develop (Best 1998). Other modelling studies (e.g. Viterbo and Beljaars 1995) have also shown the importance of a canopy surface temperature for improved model performance. Therefore, results from adding the vegetation canopy of Best (1998) to the surface energy balance of MOSES
will also be considered. The vegetation canopy, which represents 100% vegetation cover, is a simplified version of the canopy suggested by Deardorff (1978) and is modelled as a layer of vegetation over the underlying soil. This layer of vegetation exchanges radiation and turbulent fluxes with the atmosphere, but only exchanges radiative fluxes with the underlying soil. The vegetation canopy represents a closed canopy. It is possible to extend the scheme to represent partly open bare soil, but this will not be discussed here.

A conceptual diagram of this vegetation canopy is shown in Fig. 1. The surface energy balance (1) is now replaced with two energy-balance equations, one for the vegetation canopy and one for the underlying soil. These two equations are, respectively:

\[
H_v C \frac{dT_s}{dt} = (R_n - H - \lambda E) - (\varepsilon_s \sigma T_s^4 - \varepsilon_g \sigma T_g^4),
\]

(12)

\[
G = \varepsilon_s \sigma T_s^4 - \varepsilon_g \sigma T_g^4,
\]

(13)

where \(H_v\) is the height of the canopy, \(C\) is the volumetric thermal capacity of the canopy, \(T_s\) is the temperature of the canopy, \(T_g\) is the temperature at the surface of the underlying soil, \(\varepsilon_s\) is the emissivity of the canopy and \(\varepsilon_g\) is the emissivity of the underlying soil.

3. INSTRUMENTATION AND DATA PROCESSING

The observations to be described below were collected over a three-month period during the autumn of 1997 at the Met Office field site at Cardington, UK (52.06°N,
00.25°W). They were part of a long-term surface-layer-based study of a slightly inhomogeneous site.

(a) The measurement site

The site itself is in a river valley that is approximately 60 m deep and 8 km wide. The measurements were made about 3 km from the southern edge of the valley; Figs. 2 and 6 show the location of the field site relative to the local topography and land use, respectively. About 2.5 km south-east of the site there is a ridge oriented approximately north-east to south-west, although the ridge is not straight and its orientation varies along its length. The ridge rises 50 m above the level of the site over a horizontal distance of about 1 km. To the south-west, the direction corresponding to the prevailing wind, the fetch is flat, consisting of arable fields planted with winter wheat forming a 5 cm high canopy with low wire fences and isolated low bushes and trees surrounding them. A distance of 0.5 km north-east of the observation site are two large hangars. The town of Bedford lies about 3 km to the north-west of the site.

(b) Instruments

The observations were made at a number of heights using an array of mast-based instrumentation. The observations used in this study are a subset of the full instrumented array and come from the following:

(i) Placed at 10 m were: a Gill Solent asymmetric ultrasonic anemometer, the sensor head of which contained three non-orthogonal transducer paths at angles of 120° and
path lengths of 0.2 m; an Ophir IR-2000 dual-wavelength infrared hygrometer separated laterally from the sonic by a distance of 1 m (since both the sonic head and the Ophir are relatively bulky with maximum dimensions of about 0.5 m, the 1 m separation is a compromise involving the mechanical strength of the mount, minimizing the mutual effects of flow distortion around the instruments and maintaining correlation between the two sets of measurements for the determination of latent-heat flux); a fast response (0.01 s response time) platinum resistance thermometer (PRT) consisting of a helical coil of 25 μm diameter and nominal resistance 150 ohms attached to the frame of the anemometer so that it is laterally separated from the measuring volume by 0.05 m for the determination of absolute temperature and sensible-heat flux. The sonic temperature, although corrected for wind errors as per Schotanus et al. (1983) and Kaimal and Gaynor (1991), was not used to generate the sensible-heat flux as the moisture correction could not be specified. However, it was used for the quality control of the sonic data as documented later; an Automated Systems Laboratories (ASL F250) ventilated and shielded slow-response temperature sensor was used to quality control the PRT as documented later. Errors due to interference by the frame and ultrasonic transducers are small, within about 30° of the main axis of the sonic head (Wyngaard and Zhang 1985; Grant and Watkins 1989), and, to minimize these errors plus those induced by mutual Ophir flow distortion, the whole mount was placed upon a manually operated rotator so that the whole instrument package pointed into the wind for as much of the time as possible. The direction of the rotator is estimated to be accurate to 5 degrees.

(ii) Placed at a height of 4 m were: two Schulze-Dake dual-wavelength (SD) radiometers, one switched to operate in net mode and one to output upper and lower domes separately; a Kipp and Zonen (KZ CM21) solarimeter; and a Heimann KT15 radiation thermometer with a viewing angle of 30° allowing it to view a ground area of approximately 4 m².

(iii) Placed below the grass surface were four sub-soil PRTs at depths of 0.01, 0.04, 0.07 and 0.1 m.

Additionally there was a Setra pressure transducer and an ASL F250 ventilated and shielded slow-response temperature sensor at 1.2 m, a tipping bucket raingauge and a grass-surface PRT housed in an aspirated radiation shield.

(c) Data collection and quality control

The analogue outputs from the above sensors were digitized at 20 Hz by a 12 bit analogue/digital converter, encoded and transmitted by a low-power 400 MHz radio transmitter to a ground station. At the ground station, the signals were separated, decoded and passed to a MicroVax II computer for subsequent storage and processing. All the instrumentation used in this study, except the PRTs, used their generic manufacturers calibrations. The PRTs were calibrated at the Met Office, Cardington against an ASL F250 over the temperature range -10 to 40 °C. These calibrations were then applied during the data processing stage.

For the data presented in this paper only one in five records was saved giving an effective sampling rate of 4 Hz. Grant (1991) and Hopwood (1993) showed that this sampling rate was adequate for resolving turbulent motions responsible for the transport of momentum at the heights at which the measurements were taken. The data are automatically recorded for continuous two-hour periods in 'runs' of 68 min duration. This data period was chosen so that the number of data points was a power of two for the generation of spectra.
The analysis was carried out by splitting each run into segments each of 17 min in length, an adequate averaging period given the 4 Hz sampling rate for the calculation of fluxes under stable conditions to better than the 10% level of confidence (Kaimal and Finnigan 1994; Howell and Sun 1999). The wind components from the various sensors were rotated into a frame where the x-axis was aligned along the mean wind direction over the segment, the z-axis being defined by making the average vertical velocity equal to zero. The contribution to the variances and fluxes due to linear trends, if present, over the 17 min averaging period were also removed. Finally the turbulence statistics were calculated. The heat fluxes were not corrected for the effects of sonic velocity contamination, but in stable conditions this error is small (Hignett 1992). The sonic anemometer itself is an absolute instrument and few corrections to its observations are required.

However, large errors can be generated if the instrument is not correctly levelled. Care was taken to achieve this and the levelling of the sonic anemometers was brought in situ to within an error of about 2°. Further improvement was, where necessary, achieved by determining the sonic tilt from the variation of the apparent wind inclination with wind direction (e.g. Grant and Watkins 1989). Any estimated tilt was then used to rotate the Reynolds stress tensor into a co-ordinate frame with the x-axis in the direction of the mean wind and the z-axis vertically upwards. It must be noted that the vertical velocity over individual 17 min periods will not necessarily be zero and that in weak wind cases the computed fluxes will become sensitive to this correction (Mahrt 1999). The temperature sensors, prior to the experimental period, were intercompared at the same height to generate a set of zero offsets for application during data processing. In addition to the aforementioned calibration checks the following general quality control checks were carried out on the data during data processing:

(i) Occasions when the surface wind directions were between 000 and 070° were excluded because of the presence of the Cardington airship hangars.

(ii) The PRT temperature, used in the flux calculations, had to be within 0.5 K of the ASL temperature. This provided a consistency check for contamination by rain or device failure.

(iii) The standard deviation of the sonic temperature over each 17 min average had to be less than 0.4 K. This test, combined with the output of the tipping-bucket raingauge, provides a method of deducing periods when the sonic data were seriously affected by rain and has been shown by Grant and Watkins (1989) to be an adequate constraint.

(iv) The longitudinal (or x) component of the shear stress, averaged over 17 min, had to be less than zero.

(v) The standard deviation of the vertical wind component $\sigma_w$ had to satisfy

$$0.7 < 1.7\sigma_w[\ln(20/0.01) - \psi(20/L)]/U < 1.5,$$

where $U$ is the wind speed at the measurement height and $\psi$ is the integrated non-dimensional wind profile. This assumes that $\sigma_w/u_\ast = 1.4$, where $u_\ast$ is the friction velocity and is independent of stability under weakly stable conditions (Nieuwstadt 1984; Smedman 1988). This, together with check (iv), ensures that only conditions where turbulence is maintained are used in the analysis.

(d) Data analysis

The analysis of observational data in statically stable flows requires issues of representativity or variability to be addressed. The most important are instrument losses due to limited response and impact of terrain-generated effects such as land-use heterogeneity or small-scale orography.
In stable conditions, where the turbulent intensity is likely to be small in magnitude, the ability of the instrumentation to capture the turbulence accurately becomes particularly important. Finite instrument response time can eliminate some of the small-scale flux (Moore 1986; Horst 1997). The effect of instrumental response time on the derived fluxes may be assessed by considering spectra and cospectra. From Kaimal (1973), in the stable surface layer the spectral peak wavelength for vertical velocity \( \lambda_m \approx L \). However, typically \( l_m = kR_i L \approx 0.1L \) where \( l_m \) is the mixing length, \( k \) is the von Kármán constant and \( R_i \) is the flux Richardson number. The scaling of the spectral peaks on mixing is consistent with local scaling. A brief comparison with present spectra, shown in Fig. 3, showed that the use of surface-layer formulae was at least adequate to estimate response requirements and hence quality control also. The cospectra as shown in Figs. 4 and 5 fall off faster at high wave numbers than the variance spectra and have a larger spectral-peak wavelength. The overall loss due to a response on a length-scale \( l_r \) will vary as \( (\lambda_m/l_r)^{4/3} \) for covariances and \( (\lambda_m/L_r)^{2/3} \) for variances. In the observations presented in this paper, the mixing length is 4 m and \( \lambda_m \approx 50 \) m. The main response-limiting factor is the sonic anemometer length constant \( \approx 0.05 \) m. This suggests that, in general, velocity and temperature instrument losses should be confined to just a few percent.
However, with respect to moisture fluxes, the separation of 1 m between the sonic anemometer and the Ophir hygrometer may introduce errors into the latent-heat flux calculation due to decorrelation between the velocity and moisture measurements. Since the separation is in the lateral or cross-wind direction it is not necessary to apply a lag when calculating the $\overline{w'q'}$ covariance. However, the spatial separation will cause some decorrelation of the two perturbations. Moore (1986) calculated the transfer functions for lateral separations in isotropic turbulence and found that the instrument separation should be less than 10% of the measurement height in neutral or unstable conditions, or less than 0.7% of the Monin-Obukhov length under stable conditions, to avoid significant flux measurement errors. In this study the first condition is just fulfilled but the second was always violated. The heat-flux covariance could be corrected via convolution of the measured $\overline{w'q'}$ cospectra with a transfer function based upon $\overline{w'T'}$ cospectra at high frequencies, assuming the cospectral shapes are of the same form in this frequency regime. However, this has not been done in this study as the cospectral loss is of the order of 30% in the calculated latent-heat flux which is approximately 10% in overall magnitude when compared to the calculated sensible-heat flux. Hence, as latent-heat fluxes are not being compared here, the correction was not attempted.
Figure 5. Normalized heat-flux cospectra (diamonds). The cospectral estimates are normalized by the total heat flux and the wave number by height above the local surface. The dashed line represents the Kansas neutral form (Raimali et al. 1972).

For an observation site where the terrain surrounding it is not classically homogeneous, the turbulence characteristics at a measurement height may not be in equilibrium with the local surface but will reflect the conditions some distance upwind or over a source area. In particular, nocturnal boundary layers (Derbyshire 1995a) are slow to adjust to surface variations and the impact of small-scale obstacles. This is because the mixing is at a smaller scale and significantly weaker. Hence, small-scale features are more likely to individually influence the surface layer compared to surface layers with near-neutral or unstable stratification. Also, even a small amount of heterogeneity can extend the turbulence to larger Richardson numbers compared to that at a homogeneous site (Derbyshire 1995b). Therefore, identification of the flux footprint area is important in deducing the representativity of the observations. The ideas of Schmid and Oke (1990) allow the calculation of the dimensions of the area that effectively contributes to the exchange processes at the measurement height and location for a given stability and terrain type. For the observations presented in this paper the qualitative area that contains 50% of the sources that contribute to the total flux at the 10 m measurement height is shown in Fig. 6 superposed on the land-use classifications surrounding the Cardington field site. This corresponds to a $z/L = 0.1$, where $z$ is the measurement height, and a roughness length of 0.01 m. This shows that the main contribution to the observed fluxes
Figure 6. Land use of a 10 x 10 km square centred upon the field site at Cardington based upon the 25 m resolution Institute of Terrestrial Ecology land-use dataset for the UK. Also shown are the qualitative area that contains 50% of the sources that contribute to the total flux at the 10 m measurement height for three wind directions, 200°, 250° and 300°, within the prevailing wind sector. Areas in black represent unclassified data.
is from the arable fields beyond the Cardington site. These arable fields, as discussed in the site description earlier, contained winter wheat at the time of the study forming a 5 cm high canopy.

For the purposes of this study, observations of downward short- and long-wave radiation from the KZ and SD radiometers, 10 m mean wind-speed, temperature and specific humidity from the 10 m sonic, PRT and Ophir, were used to provide the forcing data for the model. In the canopy model validation the observed fluxes were provided by the sonic anemometer, PRT and Ophir. Also the grass surface temperature and the 1.2 m ASL temperature were used in the validation. The remaining instrumentation was used to provide quality control and consistency checks.

4. RESULTS

One of the biggest uncertainties in modelling the nocturnal surface energy balance is parametrizing the transfer coefficient for the turbulent fluxes. Re-arranging (3) gives \(C_E \propto H/\Delta \theta\), but in nocturnal conditions \(H\) has typically large relative errors and \(\Delta \theta\) can increase to large values, and so it is not sensible to directly compare the observed transfer coefficient with the modelled value. However, a good solution to the surface energy balance should give good agreement of both \(H\) and \(\Delta \theta\) with the observed values. Therefore the results are split into two sections: one on the temperature differences between the surface and screen level for the surface-exchange scheme with and without a vegetative canopy included, and one on the turbulent transfer of heat from the vegetation to the atmosphere. Hence, assessment can be made on the impacts of including a decoupled vegetative canopy and/or the form of stable boundary-layer stability functions on parameters important to local NWP forecast accuracy (surface-screen temperature differences) and parameters important to the accuracy of grid-box NWP forecast evolution (surface fluxes).

This comparison was achieved by using the observed downward short- and long-wave radiation, 10 m wind speed, temperature and specific humidity to force the model with and without a decoupled canopy. The observed data were interpolated to 15 min intervals to coincide with the time step employed in the model. The model was then integrated for 34 days within the two-month period of October and November 1996 producing simulated surface temperatures and heat fluxes at each model time step. These were then compared to the observed temperatures and heat fluxes over the period.

For the simulations presented in this study the vegetation properties required by the model were set to be appropriate for grass: i.e. leaf area index = 3, roughness length for momentum \(z_{0m} = 0.01\) m, canopy water capacity = 0.63 kg m\(^{-2}\), albedo \(\alpha = 0.194\) and root depth \(D_{root} = 0.57\) m.

(a) Temperature difference

Figures 7(a) and (b) were obtained using Monin–Obukhov similarity with the stable
stability functions of Beljaars and Holtslag (1991). The results are plotted against a bulk Richardson number which is defined by \(Ri_b = g z (T - T_s)/T U^2\), where \(T_s\) is the surface temperature, \(T\) is the observed 10 m temperature, \(U\) is the observed 10 m wind speed, \(z\) is the height of these observations (i.e. 10 m) and \(g\) is the acceleration due to gravity. Results without the vegetation canopy are presented in Fig. 7(a), which shows that, after a bulk Richardson number of around 0.15, there is no change in the difference between the surface and screen-level temperatures, with a maximum value obtained at around \(-2.5\) K. This does not agree with the observations where the temperature difference
continues to increase with stability to a value of around 5 K at a bulk Richardson number of 0.35.

Results with the vegetation canopy are shown in Fig. 7(b). The modelled temperature difference decreases with stability in a similar way to the observations, but to larger values. The temperature difference also remains approximately constant with larger values of stability, which again agrees with the observations.

Figures 8(a) and (b) show a direct comparison between the modelled and observed temperature differences, with and without the vegetation canopy, respectively. The error bars represent the standard errors in the mean, whilst 2 K error lines are shown to help the reader determine the magnitude of the model errors. The lowest values of the temperature differences occur on calm clear nights which have large values of $Ri_b$. This figure confirms that the simulation of the temperature difference without the canopy is poor for the lowest values of the temperatures. When the canopy is introduced into the energy balance, the agreement of the temperature difference lies within the 2 K error lines for these low temperatures. So the canopy is representing a process which is missing in the standard energy-balance equation (1).

A time series of surface temperature from the 4–8 November 1996 is shown in Fig. 9. Plotted on the graph along with the observed values are model results with and without the vegetation canopy. It is worth noting at this point that, although the surface temperature goes below the freezing point of water during this period, the soil temperatures remain above freezing (not shown), and so the energy released from the
Figure 8. Binned scatter plot of observed against modelled temperature difference between surface and screen level, with standard error. Dashed lines represent ±2 K error lines. (a) Without vegetation canopy. (b) With vegetation canopy.

Phase change of the water within the soil does not have any affect upon these results. The agreement between the model results and the observations are good except for the period of the night between the 7–8 November. This time series demonstrates that the missing process in the standard energy balance only becomes important on radiation nights, i.e. when there are clear skies and calm winds. On such radiation nights, the energy balance for vegetation without a canopy is effectively just a balance between the net long-wave radiation and the flux of heat from the soil. It is this flux of energy from the soil, stored up during the daytime, which provides the energy to the surface that can maintain a higher temperature of the vegetation than is observed. The balance of energy for the vegetation canopy on a radiation night, however, is between the different long-wave radiation components, and so the vegetation is capable of cooling to more realistic temperatures.

The turbulent heat fluxes for all of the data obtained from the model without the vegetation canopy are shown in Fig. 10(a). As the stability increases, the turbulent fluxes are underestimated when compared to the observations and are close to zero. This is caused by the gradient in temperature being too small, due to the incorrect temperature differences between the surface and screen level. Including the vegetation canopy gives more accurate fluxes with increasing stability (Fig. 10(b)). This confirms that the errors in the surface temperatures obtained without a vegetation canopy are caused by the way
the energy balance is represented, and not due to error compensation in the turbulent-flux formulation.

(b) Turbulent fluxes of heat

The results shown in this section were all obtained using the vegetation canopy in the surface energy balance. The bulk Richardson number formulation (Fig. 10(d)) agrees well with the observation over the range of stabilities, apart from near-neutral conditions where the modelled fluxes are overestimated compared to the observations. Although there is agreement to within 20% of the observed values, the bulk Richardson number formulation does overestimate the flux, as expected since this scheme has been designed to maintain turbulence with increasing stability.

Using the stability functions of Beljaars and Holtslag (1991) with Monin-Obukhov similarity theory (Figs. 10(a) and (b)) also gives good agreement with the observed fluxes, except in near-neutral conditions where the modelled flux is also overestimated. With these stability functions, the modelled fluxes tend to be slightly underestimated with increasing stability compared to the observed values. The log-linear stability functions (Fig. 10(c)) prevent turbulence for bulk Richardson numbers greater than 0.2, so surface exchange is limited to that allowed by the radiative flux. By contrast the observed fluxes have non-zero values up to at least bulk Richardson numbers of 1.0.

These results are clearly demonstrated in the time series of the turbulent heat flux for the night of 7–8 November 1996 (Fig. 11). This is the clear calm night during the time series of surface temperature (Fig. 9) discussed in the previous section. Using the log-linear stability functions gives a zero flux which is not seen in the observations. The bulk Richardson number formulation overestimates the observed flux, whilst the best agreement is obtained from the Beljaars and Holtslag (1991) stability functions.

The temperature differences between the surface and screen level for the log-linear functions are shown in Fig. 7(c). By preventing turbulence with increasing
Figure 10. Binned scatter plot of night-time sensible-heat fluxes against bulk Richardson number. Triangles—mean and standard error for observations, asterisks and dashed lines—mean and standard error for model. (a) Without vegetation canopy using Monin–Obukhov similarity with Beljaars and Holtslag functions. (b) With vegetation canopy using Monin–Obukhov similarity with Beljaars and Holtslag functions. (c) With vegetation canopy using Monin–Obukhov similarity with log-linear functions. (d) With vegetation canopy using bulk Richardson number formulation.

Figure 11. Time series of sensible-heat flux for the night of 7–8 November 1996. Solid line—observations, dashed line—model with Monin–Obukhov and Beljaars and Holtslag functions, dashed-dot line—model with Monin–Obukhov and log-linear functions, dashed-treble-dot line—model with bulk Richardson number formulation.
stability, there is insufficient warmer air mixed down to the surface, and so the surface temperature cools too far below the screen temperature compared to the observations.

The modelled temperature differences between the surface and screen level with the bulk Richardson number formulation (Fig. 7(d)) agree well with the observed temperature differences. The agreement is also better than when using the Beljaars and Holtslag function with Monin–Obukhov similarity theory.

5. Conclusions

This study has investigated the exchange of heat between the surface and prescribed near-surface atmospheric conditions. We have used observations from a field site at Cardington, UK, to compare with the performance of a surface-exchange parametrization with varying stability formulations and two surface energy-balance conditions. In most atmospheric conditions the model of the surface exchange gives reasonably accurate values of both surface temperature and surface sensible-heat fluxes.

The choice of stability functions at night can have significant impact on both the surface temperature and the sensible-heat flux. The bulk Richardson number formulation has been designed to maintain turbulence in increasingly stable conditions and overestimates the sensible-heat fluxes at this (relatively homogeneous) site. However, there is good agreement for the surface temperatures.

Using Monin–Obukhov similarity with the stability functions of Beljaars and Holtslag (1991) gives better agreement with the observations over all of the range of stabilities for the sensible-heat fluxes. The agreement for the surface temperatures is not so good as the bulk Richardson number formulation. Using Monin–Obukhov similarity with log-linear stable stability functions cuts off the flux of heat with increasing stability too quickly, compared to the observations. This leads to incorrect lower surface temperatures as the warmer atmospheric air is no longer mixed down to the surface.

It is unclear whether the bulk Richardson number formulation or the Monin–Obukhov similarity with the stability functions of Beljaars and Holtslag (1991) give the best overall agreement with the observations. However, it is clear from the results that using Monin–Obukhov similarity with the log-linear stability functions gives poor agreement with the observations. So it is important to ensure that if Monin–Obukhov similarity is used, then care is taken in choosing the stability functions.

Whilst using the standard surface energy-balance equation (1) gives good agreement between modelled and observed surface temperatures in most stability conditions, the surface temperatures are too high on clear-sky radiation nights with strong surface inversions. In these situations, the coupling of the surface to the underlying soil is too strong, supplying a source of energy which maintains higher surface temperatures. An alternative surface energy balance including a simple vegetation canopy has weaker coupling to the underlying soil and thus reduces this supply of energy to the surface. Hence, the surface temperatures are reduced and are in better agreement with the observations.

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APPENDIX A

The soil temperatures within a surface-exchange scheme are calculated by solving the diffusion equation (10). The bottom-boundary condition is provided either by a fixed temperature, or more commonly by a zero-flux condition. The top-boundary condition comes from the surface soil-heat flux which is determined from the surface energy balance. For a sinusoidally varying surface soil-heat flux $A \exp(i\omega t)$ as a top-boundary condition, the analytical solution for the soil temperatures is

$$ T = T_0 + \frac{A}{\sqrt{C\omega k}} \exp \left( -\sqrt{\frac{\omega C}{2k}} z \right) \exp \left( i \left( -\sqrt{\frac{\omega C}{2k}} z + \omega t - \frac{\pi}{4} \right) \right), $$

(A.1)

where $T \rightarrow T_0$ as $z \rightarrow \infty$, $A$ is the amplitude of the surface soil-heat flux and $\omega$ is the frequency. This solution is an exponentially decaying sinusoidal wave with depth.

To avoid discretization problems with a limited number of soil levels, it is desirable to stagger the levels so that they are consistent with the exponential decay, i.e. choose $z_j$ such that

$$ \exp \left( -\sqrt{\frac{\omega C}{2k}} z_j \right) = \frac{j}{n} $$

(A.2)

for $j = 0, 2, \ldots, n - 1$, where $n$ is the number of soil levels. So,

$$ z_j = \sqrt{\frac{2k}{\omega C}} \ln \left( \frac{n}{j} \right). $$

(A.3)

The two main frequencies for numerical weather forecasting are the diurnal period and the annual period. Therefore, depths are calculated with $\omega$ set for both the diurnal and annual frequencies. The two sets of depths are then merged to provide one continuous grid which should resolve both diurnal and annual variations accurately. Since the bottom soil level should be at an infinite depth for this method ($j = 0$), the bottom soil layer thickness has been set to twice the thickness of the layer above.

To be consistent with the hydrology scheme within MOSES, values of $k$ and $C$ were chosen so that there are an integer number of layers within the top 1 m of soil. Hence, a nine-soil-level scheme has been devised, where the soil temperatures are averages over layers which have boundaries at 0.0228, 0.0523, 0.938, 0.1648, 0.4364, 1.0000, 1.7926, 3.1488 and 5.8611 m, and the soil-heat fluxes are averages over layers which have boundaries at 0.0258, 0.0602, 0.1125, 0.2251, 0.6878, 1.3453, 2.3453 and 4.496 m, with the bottom soil-heat flux set to zero as the lower-boundary condition.

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