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Note on computing screen temperatures, humidities and anemometer-height winds in large-scale models

G.D. Hess and B.J. McAvaney
Bureau of Meteorology Research Centre, Australia
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The predictor-corrector method of computing screen temperatures, humidities and anemometer-height winds proposed by Hess et al. (1995) has been reconsidered in the light of a range of different methods of determining the fluxes and stresses in models of the land surface. We propose methods of computing effective roughness lengths and mean 'surface' values of temperature and moisture that are compatible with the computed stresses and fluxes at the scale of the grid box. These mean surface values can be used to generate appropriate grid-box values for screen temperatures, humidities and winds. Our method applies to the subgrid 'tiles' of individual surface types within a grid box (we include the possibility that a tile can be on bare or vegetated land, over the ocean or on sea-ice).

Introduction

Recently Hess et al. (1995) presented a simple, efficient predictor-corrector method to obtain winds, temperatures and humidities at screen and anemometer heights. This method was originally designed for application in large-scale models of the atmosphere which employ a Richardson number-based bulk transfer scheme such as proposed by Louis (1979; 1983). Implicit in the original discussion was an assumption of a homogeneous surface. Since more advanced large-scale models now commonly use sophisticated land-surface schemes which implicitly recognise more than one surface type in a model grid cell, some updating of the original scheme is required. However, the general principles of the predictor-corrector system still apply. We now consider the extension of the technique to a range of land-surface schemes which in turn may use a range of bulk transfer schemes.

Corresponding author address: Dr G.D. Hess, Bureau of Meteorology Research Centre, GPO Box 1289K, Melbourne, Vic. 3001, Australia.

The purpose of this note is to recommend algorithms to calculate suitable estimates of the terms used by the predictor-corrector system of Hess et al. (1995) and to present sample results from our large-scale atmospheric model coupled to a relatively complex land-surface scheme.

Extensions to the methodology

Because of the interest that has been expressed in the use of the Hess et al. (1995) method (especially in the context of the Atmospheric Model Intercomparison Project (AMIP) Phase II, Gleckler (1996)), we would like to make several additional comments:

- the method is applicable to Louis schemes where the roughness length for momentum and heat are the same, and to those where they have different values (see for example, Garratt (1994), pp. 243-244, for a specification of roughness lengths with different values for momentum and heat);

- the method is applicable to those models (e.g. Koster and Suarez 1992) which use a Penman-Monteith (Monteith 1965) formulation of the sensible and latent heat fluxes:

$$\lambda E_0 = \frac{\frac{de_{0sat}}{dT}(R_n - G_0) + \rho c_p\left(\frac{e_{0sat} - e_1}{r_a}\right)}{\frac{de_{0sat}}{dT} + \gamma\left(\frac{r_a - r_s}{r_a}\right)} \qquad ...1$$

$$H_0 = \frac{\gamma\left(\frac{r_a + r_s}{r_a}\right)(R_n - G_0) - \rho c_p\left(\frac{e_{0sat} - e_1}{r_a}\right)}{\frac{de_{0sat}}{dT} + \gamma\left(\frac{r_a + r_s}{r_a}\right)} \qquad ...2$$

where λ is the latent heat of vaporisation, E_0 the evaporative flux, H_0 the sensible heat flux, R_n the net radiation, G_0 the heat flux going into the soil, e the vapour pressure of water vapour, r_a the aerodynamic resistance, r_s the stomatal resistance, ρ the air density, c_p the specific heat of air at constant pressure, and $\gamma \equiv \rho c_p/(0.622\lambda)$. The subscript 0 indicates a value at the surface, the subscript 1 indicates a value at the lowest model level and the subscript *sat* indicates the saturated value.

To apply the procedure given by Hess et al. (1995) to more general formulations such as the above, the only modification necessary to the original scheme is to replace the Louis form of the drag law with the actual drag law used in the land-surface scheme at the lowest model level and evaluate the friction velocity u^*, and subsequently to use Eqns 1 and 2 to evaluate the scaling parameters Θ_* ($= -H_0/\rho c_p u_*$) and Q_* ($\equiv -E_0/\rho u_*$). The parameter r_a is given by $1/(C_H[Ri_b,(z_1-d)/z_0,(z_1-d)/z_H]U_1)$ where C_H is the bulk transfer coefficient for heat, Ri_b the bulk Richardson number, z_1 the height of the lowest model level, z_0 the momentum roughness length, z_H the heat roughness length (which in some Louis schemes is the same as the momentum roughness length), d the zero-plane displacement height, and U_1 the magnitude of the horizontal wind at the lowest model level.

Application to land-surface models

Physically based land-surface models ('SVAT' models, or surface-vegetation-atmosphere-transfer models) relate evaporation, sensible heat transfer and the momentum stress to the vegetation and the soil. Typically such schemes have a canopy-interception reservoir and multiple soil reservoirs for water, and stomatal resistances affected by environmental stresses. Subgrid-scale variability is sometimes accounted for by a 'mosaic' approach where a heterogeneous large-scale grid box is divided into relatively homogeneous subregions ('tiles' of the mosaic) and separate energy and water balances are calculated for each tile. Usually each tile maintains its own prognostic equations for soil moisture contents and soil temperatures, but sometimes the soil moisture is kept uniform over the grid box and only the vegetation is tiled. Often this land-based tiling approach is extended to include the case where a tile may be a lake, sea or sea-ice. Sometimes the geographic location of each tile is accounted for, otherwise the subgrid-scale heterogeneity is represented more simply by using the fractional cover of each surface type (Pitman 1991).

Once the physical characteristics and the energy and water balances of the tiles are known, the next step is to calculate the transfer of momentum, heat and moisture from the entire grid box to the atmosphere of the 'host' model. Three approaches for calculating this transfer have been suggested: (a) 'flux aggregation' (where the fluxes are averaged over the grid box, using a weighted average with the weights determined by the area covered by each tile); (b) 'parameter aggregation' (where parameters such as roughness length, albedo, leaf-area index, stomatal resistance, soil conductivity, etc., are derived in a manner which attempts to best incorporate the combined nonlinear effects of the different tiles in the grid box; this technique looks at the grid box as a whole and does not need the water and energy balance calculations over the separate tiles); and (c) a combination of the flux aggregation and parameter aggregation methods (see for example, Claussen (1995b)).

The method described by Hess et al. (1995) is applicable to land-surface schemes involving both subgrid-scale tiling and fractional cover. To apply this method we aggregate the surface fluxes corresponding to different subgrid areas to determine the mean sensible heat flux $\langle H_0 \rangle$ and latent heat flux $\lambda\langle E_0 \rangle$ over the grid box:

$$\langle H_0 \rangle = \rho c_p\, U_1 \sum_i f_i C_H [Ri_{b\,i}, (z_1-d_i)/z_{0\,i}, (z_1-d_i)/z_{H\,i}]\,(\Theta_{0i} - \Theta_1) \qquad ...3$$

$$\langle E_0 \rangle = \lambda\,\rho\, U_1 \sum_i f_i C_H [Ri_{b\,i}, (z_1-d_i)/z_{0\,i}, (z_1-d_i)/z_{H\,i}]\,(Q_{0i} - Q_1) \qquad ...4$$

where f_i is the fractional cover and d_i the zero-plane displacement height for surface type (tile) i,

We then need to calculate values for $\langle \Theta_0 \rangle$ and $\langle Q_0 \rangle$. In most cases simple linear averaging of the surface potential temperature and specific humidity over the grid box does not provide values of $\langle \Theta_0 \rangle$ and $\langle Q_0 \rangle$ that are consistent with the fluxes from the atmospheric model and hence can be considered 'incorrect'. The processes governing the interaction between the atmosphere and the land surface are, in general, inherently nonlinear so that simple linear averaging techniques are not appropriate. There might be a strong temptation to try to verify our grid-box model results with point measurements of temperature and humidity averaged averaged over different tiles. This approach should be resisted because of the difficulty in correctly performing the averaging process. Remotely sensed measurements averaged over areas corresponding to the grid-scale of the model are the most appropriate for verification data.

The mean surface values can, however, be computed by combining the flux aggregation and the parameter aggregation techniques described above (Eqns 3 and 4) with a parameter aggregation technique based on the following equations:

$$\langle H_0\rangle = \rho c_p C_H[\langle Ri_b\rangle, (z_1 - \langle d\rangle)/\langle z_0\rangle, (z_1 - \langle d\rangle)/\langle z_H\rangle] \quad U_1(\langle\Theta_0\rangle - \Theta_1) \qquad \ldots 5$$

$$\lambda\langle E_0\rangle = \lambda\,\rho\, C_H[\langle Ri_b\rangle, (z_1 - \langle d\rangle)/\langle z_0\rangle, (z_1 - \langle d\rangle)/\langle z_H\rangle] \quad U_1(\langle Q_0\rangle - Q_1) \qquad \ldots 6$$

The effective roughness for momentum (based on skin friction only) $\langle z_0\rangle$ can be defined as (Mason 1988):

$$\left(\ln\left(\frac{l_b}{\langle z_0\rangle}\right)\right)^{-2} = \sum_i f_i\left(\ln\left(\frac{l_b}{\langle z_{0i}\rangle}\right)\right)^{-2} \qquad \ldots 7$$

where l_b is the 'blending height', that is, the height at which the air flow senses the blended influence from the whole grid box. Different formulae for the blending height, which include a scale of horizontal variation (e.g. Mason 1988; Claussen 1991), have been proposed. However, for simplicity and because of the approximate nature of the theory, we take the blending height to be the height of the lowest model level (in our case $l_b \approx 75$ m; Wieringa (1986) suggests using 60 m). If a height other than the lowest model level is chosen, then corrections must be applied to determine the fluxes at the blending height (see Claussen 1991). The present theory assumes that the blending height is located within the surface boundary layer.

The value of the effective roughness length for momentum $\langle z_0\rangle$ is strongly influenced by isolated obstacles or small rough patches of surface cover. The effective roughness length for heat $\langle z_H\rangle$ behaves differently. It responds primarily to the dominant surface cover. Typically, the effective roughness length for heat is much smaller than that for momentum, but the ratio has been found experimentally to vary over an enormous range of values. Betts and Beljaars (1993) found $\langle z_0\rangle/\langle z_H\rangle = 18$ for data of the First ISLSCP Field Experiment (FIFE), whereas Beljaars (1995) found $\langle z_0\rangle/\langle z_H\rangle = 10^7$ for the data of the Oklahoma Boundary Layer Experiment (BLX83)). A formula for the effective roughness for heat, based on the assumption of near-neutral thermal stratification, and a constant heat flux vertical profile but a variable momentum flux vertical profile, is given by Beljaars and Holtslag (1991):

$$\ln(l_b/\langle z_H\rangle) = \frac{\ln(l_b/z_0)\ln(l_b/z_H)}{\ln(l_b/\langle z_0\rangle)} \qquad \ldots 8$$

where z_0 and z_H are the local values of the dominant surface cover. There is a great deal of uncertainty, however, still remaining in the choice of an optimal formulation for $\langle z_H\rangle$ and hence we do not yet feel confident in recommending a specific choice.

In the above discussion we were only concerned with determining the effective exchange of scalars like heat and moisture. When computing the momentum flux, the effective roughness length should also include the effect of topographical roughness (form drag) (see, for example, Wood and Mason (1993) and Claussen (1995b)). The recent emergence and use of more physically based subgrid-scale orographic drag parametrisation schemes (which include the generation and dissipation of gravity waves) such as Lott and Miller (1997) is providing a framework for the inclusion of form drag that is separate from details of a land-surface scheme.

In applications over tall vegetation, the height z in Eqns 1 to 6 and Eqns 8 to 13 in Hess et al. (1995) should be replaced by $(z\text{-}d)$ where d is the zero-plane displacement height. There are many practical difficulties in determining the value of d. (At this time the uncertainties in d are such that there is no justification for introducing separate displacement heights for momentum and heat.) The usual procedure of obtaining d involves careful fitting of neutral wind profiles to a logarithmic function. However, in the context of large-scale modelling, this procedure is impractical. In principle d depends on the height of the canopy h and the spacing of the roughness elements, but, for simplicity we suggest d=0.7h (see Wieringa 1993; Garratt 1994, p. 290) and we recommend that the grid-area average be computed as a linear average weighted by the fractional cover. The introduction of d complicates direct comparisons with anemometer and screen-height point observations, because sections of the grid box refer to values of wind speed, temperature and humidity above a canopy, while other sections may refer to values above the ground. On the other hand, adopting a common methodology when including d enables consistent inter-model comparisons to be performed.

When the fluxes averaged over the grid area are used in Eqns 5 and 6 to determine the mean surface temperature and moisture, certain anomalies or inconsistencies can arise (see the discussion of the 'Schmidt paradox' by Lettau (1979) and Claussen (1991) for example). The mean variables and the fluxes are related in a highly nonlinear way. Small regions of strong turbulence in unstable conditions can dominate the grid-area averaged fluxes, but have less effect on the vertical mean value deficit between the surface and the lowest model level, leading to a situation of apparent counter-gradient transport. When exchange processes over rough surfaces involving significant canopies are averaged with those over smooth surfaces, a similar inconsistency can occur. Even more extreme cases can occur in a grid box that has subgrid tiles consisting of land, sea-ice and ocean (see Stössel and Claussen 1993). Even for these extreme cases the general principles of flux aggregation still apply. The fluxes in Eqns 3 and 4 have been determined by aggregating the fluxes for the individual tiles, but the bulk transfer coefficient C_H is determined by aggregating the roughness parameter to form an effective roughness parameter.

To avoid situations of inconsistency we impose several constraints based on observational considerations, but determined through our experiences using one particular tiled land-surface scheme (the BASE scheme, Desborough (1997)): when the computed flux is non-zero, the mean wind speed must be equal to or greater than 0.25 m s^{-1}; the sensible and latent heat fluxes cannot be resolved to values less than 1 W m^{-2} (when the grid-area averaged fluxes are less than 1 W m^{-2}, the mean potential temperature or specific humidity deficit between the surface and the lowest model level is set to zero). The chosen value of 1 W m^{-2} is much less than the observational error, typically 15 W m^{-2}, in observational flux estimates. We would suggest that the setting of these parameters should be checked for different land-surface schemes; we have found the above criteria to be robust during long runs with our climate model with BASE.

Equations 5 and 6 are solved for the surface potential temperature $\langle\Theta_0\rangle$ and specific humidity $\langle Q_0\rangle$ averaged over the grid box.

Once $\langle Q_0\rangle$ is known, an effective surface wetness parameter (β) can be obtained from (see Kondo et al. (1990) and Mahfouf and Noilhan (1991)):

$$\lambda\langle E_0\rangle = \rho C_H U_1 \beta (Q(\langle T_0\rangle)_{sat} - Q_1) \qquad \ldots 9$$

where $\langle T_0\rangle$ is the grid-box mean surface temperature corresponding to the potential temperature $\langle\Theta_0\rangle$, and β is defined as $(\langle Q_0\rangle) - Q_1)/(Q\ \langle T_0\rangle)_{sat} - Q_1)$. This relation defines the actual grid-box mean evaporation in terms of the potential evaporation.

In the procedure that has been developed, no allowance is made for deviations from the Monin-Obukhov stability relationships in the transition layer located immediately above the roughness elements (see Garratt 1994, pp. 58-60). The procedure calculates Louis scheme profiles and these approximate Monin-Obukhov profiles.

Calculation procedure involving a land-surface scheme

We now present the details of how we have proceeded for a case using a tiled land-surface scheme (BASE; Desborough 1997). The BASE land-surface scheme uses a bulk formulation for each subgrid-scale surface component. It then uses flux aggregation to provide estimates of the sensible $\langle H\rangle$ and latent heat $\lambda\langle E\rangle$ fluxes over the grid box, and computes the effective roughness length (using Eqn 7) for the grid box. The scheme also provides an estimate of the effective radiative potential temperature $\langle\Theta_*^r\rangle$ for the grid box. This temperature is determined from the upward long wave radiative flux in the surface energy balance calculation over the grid box.

In addition to the standard information passed between BASE and the host model we requested that the BASE scheme also provide to the large-scale model an estimate of $\langle u_*^2\rangle$ for the box by aggregation of the friction velocity for each surface type. We calculate the grid-box mean momentum flux from

$$\langle u_*^2\rangle = \sum_i f_i \left(\frac{k}{\ln\frac{(z - d_i)}{z_{0i}}}\right)^2 F_m\,(Ri_{bi}, (z_1 - d_i)/z_{0i})\,U_1^2 \qquad \ldots 10$$

where F_m $[Ri_{bi}, (z_1 - d_i)/z_{0i}]$ is the Louis stability function for momentum for tile i.

We then proceed as follows:

Case A

1. Check for cases where $\Delta\Theta = \langle\Theta_*^r\rangle - \Theta_1$ has a sign inconsistent with $\langle H\rangle$, or $\langle H\rangle < 1$ W m^{-2}. These cases will be considered to be neutrally stratified.
2. For these cases, calculate a neutral-stability bulk transfer coefficient using the effective roughness length provided by BASE (using Eqn 7), and solve for $\langle\Theta_0\rangle$ from Eqn 5 and for $\langle Q_0\rangle$ from Eqn 6. The neutral bulk transfer coefficient is given by:

$$C_H = \left(\frac{k}{\ln\left(\frac{(z_1 - \langle d\rangle)}{\langle z_0\rangle}\right)}\right)\left(\frac{k}{\ln\left(\frac{(z_1 - \langle d\rangle)}{\langle z_H\rangle}\right)}\right) \qquad \ldots 11$$

Case B

3. For other cases, we relate the aggregated momentum flux $\langle u_*^2\rangle$ found from Eqn 10 to the value found from parameter aggregation:

$$\langle u_*^2\rangle = \left(\frac{k}{\ln\left(\frac{(z - \langle d\rangle)}{\langle z_0\rangle}\right)}\right)^2 F_m\,(\langle Ri_b\rangle, (z_1 - \langle d\rangle)/\langle z_0\rangle)\,U_1^2 \qquad \ldots 12$$

where F_M is the Louis stability function for momentum for the grid box. Since $\langle z_0\rangle$ is known from Eqn 7, we then solve Eqn 12 for the value of the effective Richardson number $\langle Ri_b\rangle$.

4. Compute the bulk transfer coefficient C_H:

$$C_H = \left(\frac{k}{\ln\left(\frac{(z_1 - \langle d \rangle)}{\langle z_0 \rangle}\right)}\right)\left(\frac{k}{\ln\left(\frac{(z_1 - \langle d \rangle)}{\langle z_H \rangle}\right)}\right) \quad \ldots 13$$

$$\times F_H \left[\langle Ri_b \rangle, (z_1 - \langle d \rangle)/\langle z_0 \rangle, (z_1 - \langle d \rangle)/\langle z_H \rangle\right]$$

5. Calculate $\langle \Theta_0 \rangle$ from Eqn 5 and $\langle Q_0 \rangle$ from Eqn 6.

The procedure followed in case B above was prompted by a diagnostic methodology suggested by Stössel and Claussen (1993). The values for $\langle \Theta_0 \rangle$ and $\langle Q_0 \rangle$ obtained are then used in calculating the screen temperatures and humidities using the procedure of Hess et al. (1995).

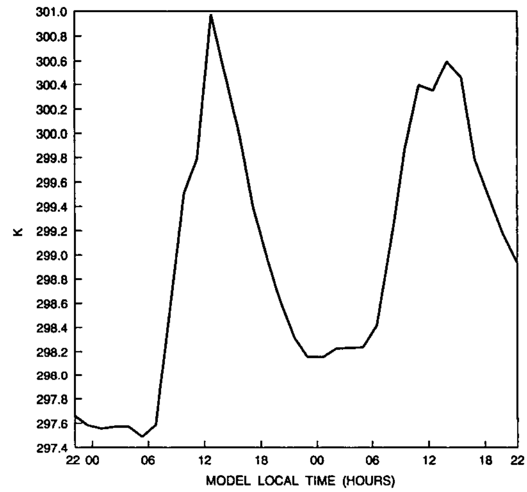
Illustrative results for the BMRC model

We illustrate the results of this technique using the output of the BMRC AGCM. The experiments are similar to those described in Hess et al. (1995), except that the land-surface scheme (BASE) was used with a radiation calculation every 90 minutes and model data were archived at the time of each radiation calculation. In the current model configuration $z_H = z_0$ and $d = 0$.

Figure 1 illustrates the diurnal variation for $\langle \Theta_0 \rangle$ for two grid boxes. One grid box (see Fig. 1(a)) is typical of the tropics and includes both vegetated land and sea (14.5°S and 142.5°E – 11 per cent of the grid box is covered by sea); the other grid box (Fig. 1(b)) is typical of a subtropical semi-arid region (34.7°S and 138.7°E). The data points are 90 minutes apart, commencing at 2200 local time (LT). For the tropical point in Fig. 1(a) the diurnal cycle in $\langle \Theta_0 \rangle$ follows the general pattern in behaviour of the (small) sensible heat flux (not shown); the effective Richardson number over the grid box (not shown) remained stable except for small unstable excursions near local midday. The effective radiative potential temperature showed a very similar behaviour. For the point in the semi-arid, low-vegetation area, evaporation was severely limited by the availability of soil moisture and hence the large diurnal range in sensible heat flux associated with the unstable conditions indicated by the effective Richardson number (not shown) produces a strong diurnal variation in $\langle \Theta_0 \rangle$; there is also a substantial warming trend over the period. As expected, the effective radiative potential temperature behaves in a very similar manner. Model results such as these should, in principle, be checked using satellite radiometric data over a grid-box area. However until such data become readily available we must rely on the consistency of the method, the physical reasonableness of the predicted behaviour and the agreement of the method with direct calculations using smaller-scale models (e.g. Claussen 1995a,b). Since the behaviour of $\langle \Theta_0 \rangle$ is physically reasonable for these (and other points), we consider that our recommended formulation is appropriate.

Fig. 1 The mean 'surface' potential temperature $\langle \Theta_0 \rangle$ (full lines)for points selected over: (a) a point in the Australian tropics; and (b) a subtropical desert point in south central Australia. The data are for two 'typical' days selected from a 45-day run, using an initial condition set at 1 January. The abscissa shows the local time in hours.

(a)



(b)

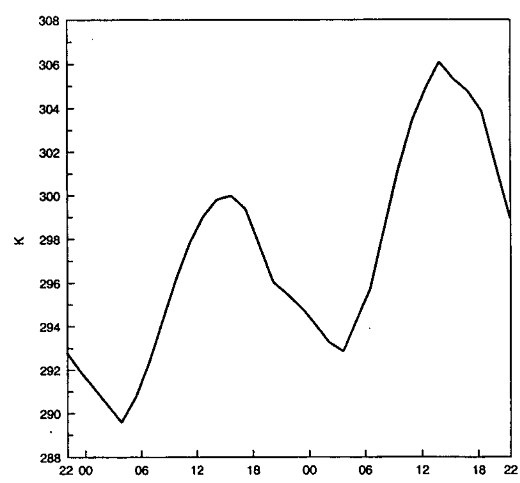
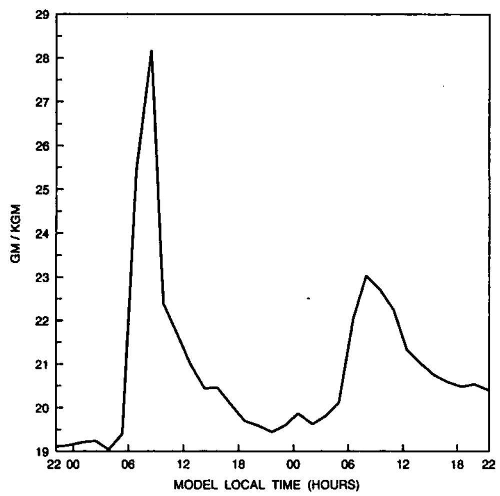


Fig. 2 The same as Fig. 1, except for mixing ratio $\langle Q_0 \rangle$.

(a)



(b)

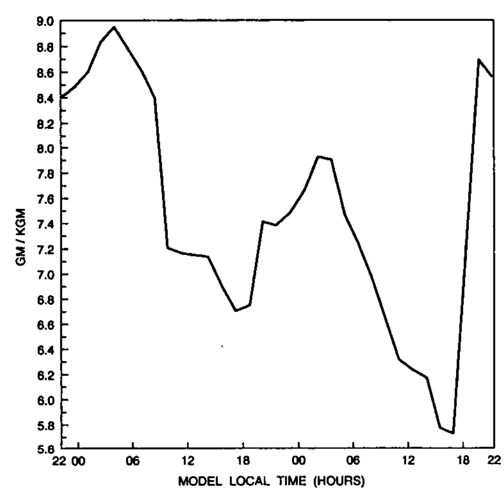


Figure 2 illustrates the variation in $\langle Q_0 \rangle$ for the same grid locations shown in Fig. 1. For the tropical point shown in Fig. 2(a) there is a strong diurnal variation in the latent heat flux (not shown). $\langle Q_0 \rangle$ for this point also shows a large amplitude diurnal signal which is strongly modulated by the near-neutral stability that exists during most of the daylight hours on the second day. For the semi-arid grid-point in Fig. 2(b) there is, as expected, very little latent heat flux (< 15 W m^{-2}) and the behaviour of $\langle Q_0 \rangle$ follows the behaviour of the specific humidity at the first model level (not shown) and is relatively dry. Again, the behaviour of $\langle Q_0 \rangle$ over the day is physically reasonable.

Summary

We have proposed methods of computing effective roughness lengths and mean surface values of temperature and moisture on the scale of a grid box that are compatible with the computed stresses and fluxes at the grid-box scale and which can be used to generate appropriate grid-box values for screen temperature, humidities and winds. Our method applies to the subgrid 'tiles' of a 'mosaic' grid box for a land-surface model as well as to the more simple differentiation between a vegetated surface and bare land. Results using a land-surface scheme illustrate physically consistent behaviour for the 'grid-box average' variables. We recommend that our method be used in conjunction with our previous method of calculating surface air temperatures, humidities and winds so that grid-box measures for these variables can be obtained that are consistent with the stresses and latent and sensible heat fluxes output by a numerical model.

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