# Surface energy balance: observation and numerical modeling applied to Candiota

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## Abstract

The atmosphere can interact with the terrestrial surface in several ways, therefore any regional-scale dynamical flow model should include an adequate parameterization of the terrestrial surface used as its lower boundary. It is also important to estimate correctly the parameter values used as initial condition of these models. Here, a method of soil-atmosphere interaction, including an one-layer foliage parameterization proposed by Deardorff (1978), is utilized to predict air temperature and specific humidity on the ground and the turbulent fluxes to the atmosphere at Candiota (Brazil) during January, 1994. The purpose is to validate the Deardorff's model in order to adequately describe both the measured radiation and turbulent fluxes. The results indicate good agreement between modeled and observed turbulent fluxes and net radiation. There is a large discrepancy between observed and modeled soil surface heat fluxes. The model is very sensitive to both ground and foliage albedos and emissivities, net leaf area index, retained water on the foliage, concentration of soil moisture and much other parameters which have not been determined during the field campaigns. The discrepancies and uncertainties found in this work indicate that a meteorological experiment must be rethought to observe soil and vegetation parameters with relative precision. The energy flux balances observed in Candiota during other field campaigns are also discussed here.

## 1 Introduction

According to Sellers et al. (1989) the atmosphere can interact with the terrestrial surface in three ways. First, there is an exchange of radiation between the two. Land surface albedo can vary widely, having a direct influence on the radiation absorbed by the surface. A second influence of the land surface is the drag force exerted on the lowest layer of the atmosphere by the roughness elements, mainly vegetation, projecting into the airflow. A third influence of the land surface on the atmosphere is through the availability of moisture for evapotranspiration and the control exerted by vegetation on its release. Shukla and Mintz (1982) have shown that altering the regional soil moisture initialization may have large effects on the continental climates.

The method used here and proposed by Deardorff (1978) involves, basically, solution of an abbreviated energy budget equation to obtain the temperature of a representative foliage element and diagnosis of mean air temperature and humidity within the vegetation layer. The surface temperature and moisture are determined solving a prognostic equation, dependent upon forcing by the sum of the energy fluxes. It contains a mechanism by which a deeper soil layer can influence both surface temperature and humidity. The approximation of double layer was suggested observing that most of temperature and humidity variations in the external layer are associated to the diurnal cycle while the signals in the deeper layer are associated to the seasonal cycle.

Micrometeorological and heat flux measurements obtained in Candiota, during the field campaign of January 1994, are used to validate and calibrate the Deardorff's model. Observed and modeled energy fluxes are compared and theirs results are discussed.

The energy balance obtained in others Candiota field campaigns (July, 1994; February, 1995; May, 1995 and November, 1995) are also shown in this work.

## 2 Data

The experimental site was Candiota, a region of "pampas". Turbulent measurements were made using a micrometeorological tower 12 meter high using the following equipment: Fine Wire Thermometer (temperature); Sonic Anemometer (vertical velocity); Gill Anemometer (horizontal velocities); Kripton Hygrometer (water vapor); Net Radiometer (net radiation); Soil Heat Flux Sensor at 1 cm. The turbulent data were sample with 1-10 Hz frequency and the averaged covariance reported here were evaluated using 5 minutes period data. The turbulent sensors were set up at 9 meters, the radiation sensor at 2 meters.

The data set was interpolated in time using a convergent weighted-averaging interpolation scheme (Barnes, 1964). It is based on the supposition that the distribution of an atmospheric variable at any given time can be represented by the summation of an infinite number of independent waves, i.e., a Fourier integral representation.

## 3 Deardorff's model

A single layer of vegetation which has negligible heat capacity is assumed to be present in the problem. Its density is characterized by the single quantity $\sigma_f$, which is an area average shielding factor associated with the degree to which the foliage prevents shortwave radiation from reaching the ground. The limits of $\sigma_f$ are $0 \le \sigma_f \le 1$; $\sigma_f = 0$ signifying no foliage and $\sigma_f = 1$ signifying complete radiative blocking. The air in close proximity to the foliage is assumed to take on properties intermediate between above-canopy air properties, foliage surface properties, and ground surface properties. A gross energy budget for the foliage layer is established in order to estimate foliage surface temperature. The soil properties are parameterized in function of soil moisture.

The force-restore method is used to estimate the ground surface temperature ($T_g$). It includes a restoring term that contains the deep soil temperature $T_2$:

$$\partial T_g / \partial t = -c_1 H_A / (\rho_s c_s d_1) - c_2 (T_g - T_2) / \tau_1 \qquad (1)$$

$$\partial T_2 / \partial t = -H_A / (\rho_s c_s d_2) \qquad (2)$$

where, $T_2$ is the mean soil temperature over layer of depth $d_2$; $H_A$ is the sum of fluxes to atmosphere; $c_1, c_2$ are dimensionless constants; $\rho_s$ is the density of soil; $c_s$ is the specific heat of the soil; $d_1, d_2$ are the soil depths influenced by the diurnal and annual temperature cycles, respectively and $\tau_1$ is the diurnal period.

A method, analogous to the force-restore method of predicting surface temperature, is used to predict the ground surface moisture. The specific humidity at the surface is then related to the ground surface moisture content. This permits evaporation to dry out the ground surface and so reduce the evaporation rate from bare soil in comparison with evapotranspiration.

Assuming that most of the vertical movement of the volumetric concentration of ground soil moisture (w) within the soil can be described by a diffusion process, the equation for the volumetric concentration of soil moisture can be written as:

$$\partial w_g / \partial t = -C_1 (E_g + 0.1 E_{tr} - P_g) / (\rho_w d_1') - C_2 (w_g - w_2) / \tau_1 \qquad (3)$$

$$\partial w_2 / \partial t = -(E_g + E_{tr} - P_g) / (\rho_w d_2') \qquad (4)$$

where, $w_g$ is the ground surface value of w; $\rho_w$ is the density of liquid water; $C_1$ and $C_2$ are constants analogous to $c_1$ and $c_2$; $d_1$' is a depth to which the diurnal soil moisture cycle extends; $w_2$ is the vertically averaged value of w over a ticker layer $d_2$' below which the moisture flux is negligible; $E_g$ is the evaporation rate at the ground surface; $E_{tr}$ is the foliage transpiration rate and $P_g$ is precipitation rate felt at the ground surface.

Next section discusses the results obtained using the model of Deardorff to simulate Candiota conditions.

## 4 Results

An application of the Deardorff model is made by simulating 48-hours period of the conditions observed in Candiota during days 19 and 20 of January, 1994.

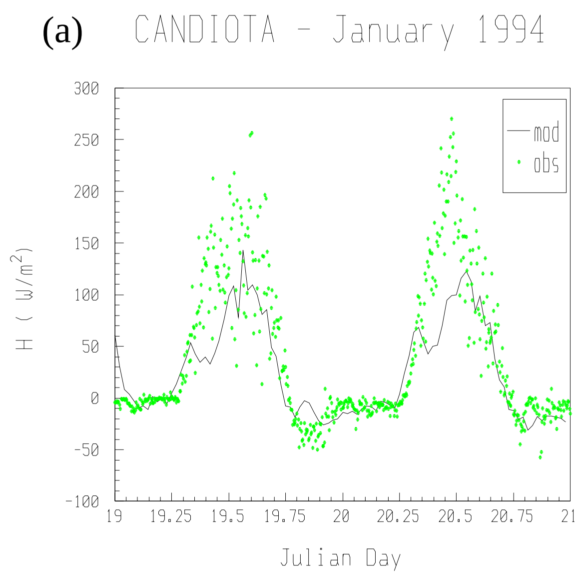
Observed air temperature, specific humidity and wind speed at 2 m height were used as imput of the model. During this period there was little or no cloudiness and the mean wind speed varied between 1.5 m $s^{-1}$ and 5 m $s^{-1}$.

| **Parameter** | **Value** |
|---|---|
| $T_2$ (mean soil temperature of the deeper layer ) | 301.6 K |
| $T_g$ (ground surface temperature) | 298.2 K |
| $w_2$ (volumetric concentration of soil moisture of the deeper layer) | 0.4 |
| $w_g$ (volumetric concentration of soil moisture) | 0.1 |
| $\sigma_f$ (foliage shielding factor) | 0.25 |
| $K_s$ (soil thermal diffusivity) | 1.22 x $10^{-6}$ $m^2$ $s^{-1}$ |
| $z_0$ (bare surface roughness) | 0.01 m |
| $z_{0f}$ (vegetation roughness) | 0.03 m |
| $\varepsilon_g$ (ground surface emissivity) | 0.95 |
| $\varepsilon_f$ (foliage emissivity) | 0.95 |
| $\alpha_f$ (foliage albedo) | 0.2 |
| $\alpha_g$ (ground surface albedo) | 0.15 |
| $d_1$'(soil deep influenced by the diurnal soil moisture cycle) | 0.10 m |
| $d_2$' (soil deep influenced by seasonal soil moisture variations) | 0.50 m |
| d (zero displacement length) | 0.23 m |

**Table 1:** Parameters used in the Deardorff model.

Soil parameters and other relevant values are stipulated as indicated in Table 1.

(a)

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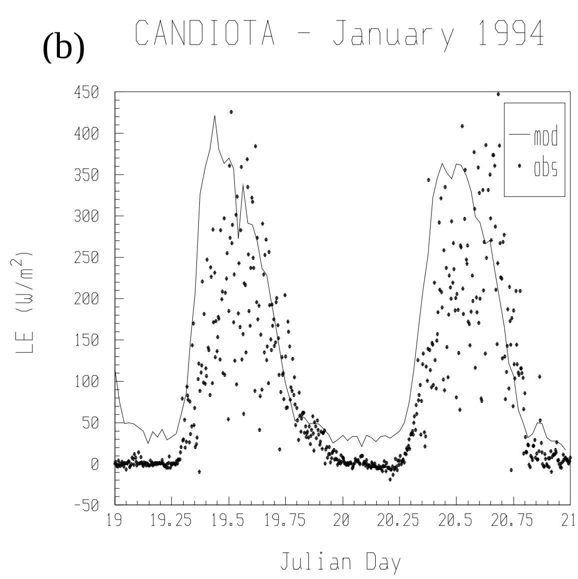
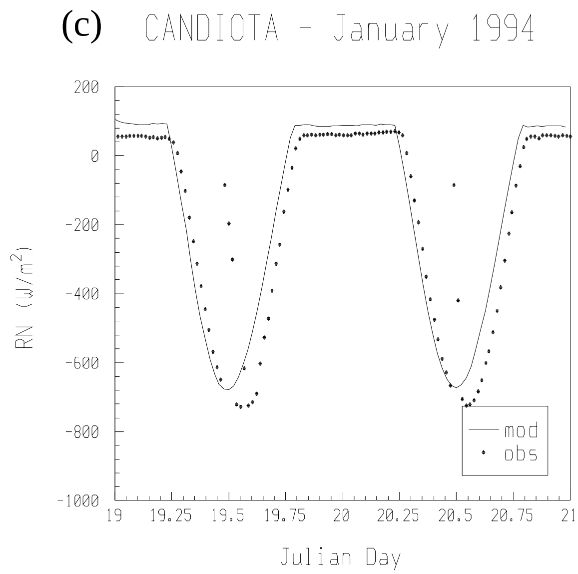


**Figure 1:** Observed and modeled (a) sensible heat flux (H); (b) latent heat flux (LE); (c) net radiation (Rn) and (d) soil energy flux at the surface (G). Positive when directed upward.



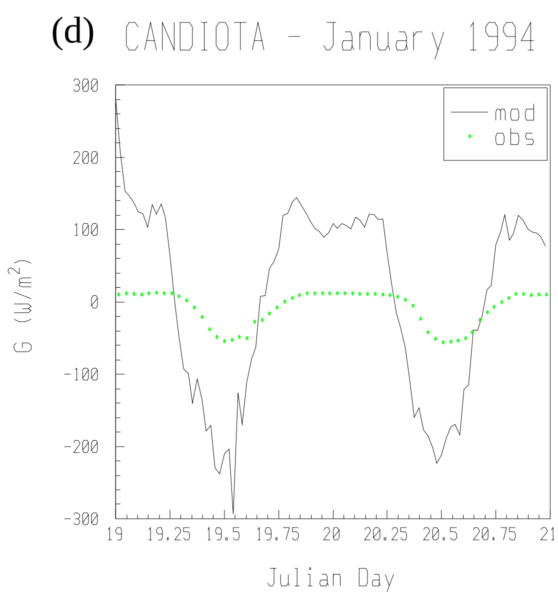




**Figure 1**: Continuation.

The model was able to reproduce the observed patterns of sensible heat flux (Fig. 1a), latent heat flux (Fig. 1b) and the net radiation (Fig. 1c).

A large discrepancy, however, was found between the modeled and observed soil heat flux at the surface (Fig. 1d). Some reasons can account for the observed discrepancy. First, the lack of measurements of soil and vegetation parameters used as initial conditions of the model. The model has been shown to be very sensitive to variations of foliage shielding factor ($\sigma_f$). The arbitrary value of $\sigma_f$ used here was selected to yield satisfactory values of net radiation. Second, the observed energy balance for this experiment, shown in Fig. 2, indicates that the balance between the energy terms is not exact, having a large residue. Imprecise measurements of the soil energy flux at the surface could be responsible for an important part of this residue.

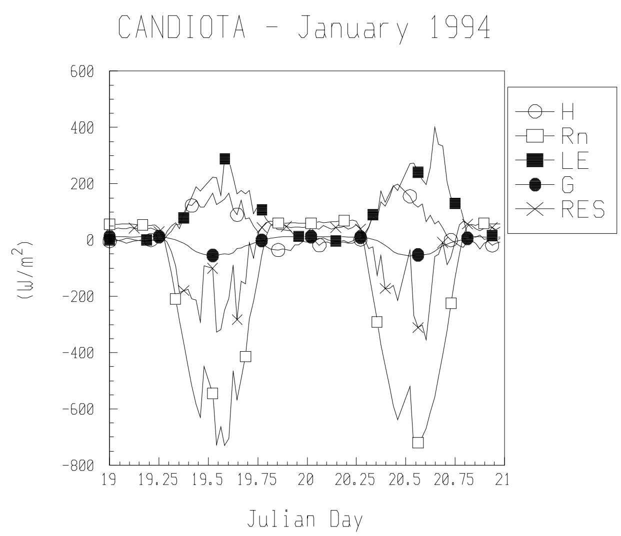
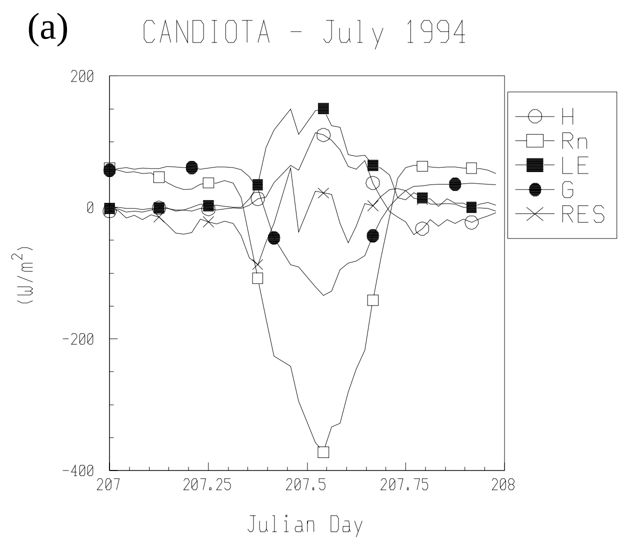



**Figure 2:** Energy flux balance. RES is the energy balance residue. Positive when directed upward.

Trying to understand the discrepancy found between the modeled and observed soil heat flux at the surface it is worth to look at the energy balances obtained in Candiota during other field experiments.

As can be seen in Figures 2 and 3, the residues of the energy term balances are significant for the 5 field campaigns. Those residues are proportional to the amplitude of the fluxes suggesting a problem of representativity of the measurements of the soil energy fluxes.



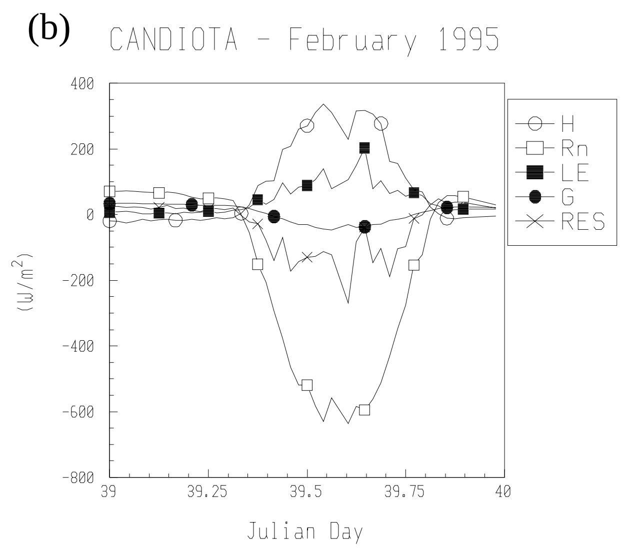
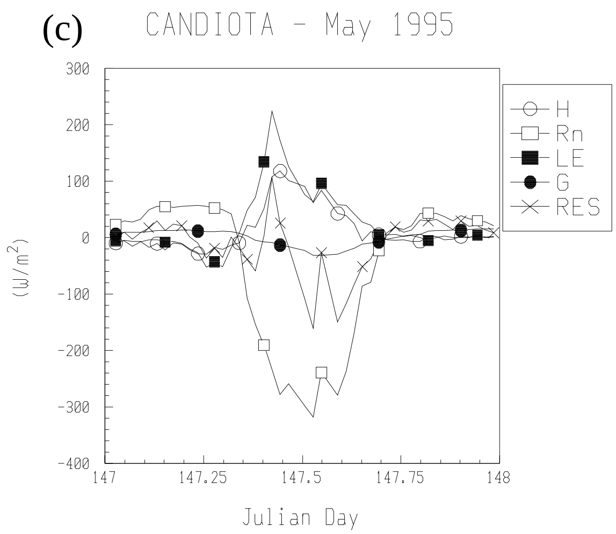




**Figure 3:** Energy flux balances. (a) July 1994; (b) February 1995; (c) May 1995 and (d) November 1995. Positive when directed upward.

(c)




(d)

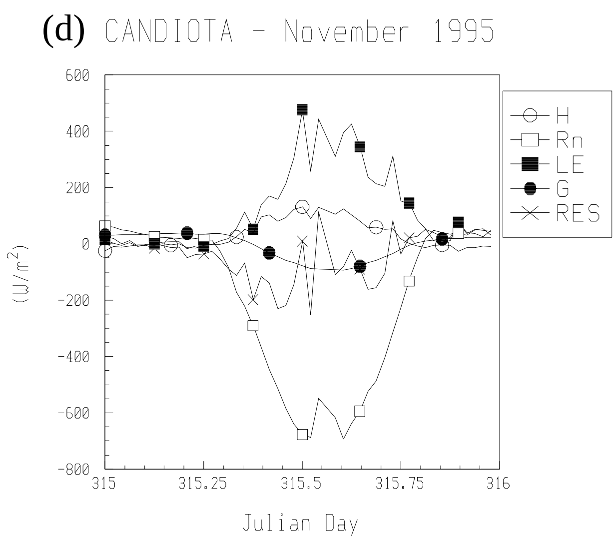



**Figure 3:** Continuation.

During the field campaigns of July and May the amplitude of the fluxes were smaller than those obtained during the warmer period, January, February and November (Figures 2 and 3). The fluxes do not present a very well defined seasonal pattern indicating that the soil moisture is a determinant factor of the energy balance.

## 5 Discussion

Comparison between simulated turbulent fluxes and observations are very seldom found in literature and when found the agreement between modeled and observed results are usually quite poor. For instance, Manzi et al. (1994) have not obtained a good comparison using a model of boundary condition for Amazon Rain Forest. Deardorff (1978) has not compare simulated turbulent fluxes with observations.

Here, despite the uncertainties in the experimental conditions, the Deardorff's model was able to reproduce the observed patterns of sensible heat flux, latent heat flux and the net radiation obtained in Candiota during the experiment of January 1994. The large discrepancy found between the modeled and observed soil heat flux can be due to the lack of measurements of soil and vegetation parameters used as initial conditions of the model and/or due to the representativity of the measurements of the soil energy fluxes.

In order to use observational data as imput of model predictions, care needs to be taken in the selection of experimental data sets and careful assessments need to be made of the effects of uncertainties in the experimental conditions. For example, uncertainty in the soil composition and surface properties can produce significant changes in model outputs. It is unlikely that any atmospheric meso scale model will perform well without previous calibration of the lower boundary conditions.

Next field campaigns must determine both soil heat flux and the parameters listed in Table I with the maximum possible precision.

## 6 References

Barnes, S.L., 1964: A technique for maximizing details in numerical weather map analysis. *J. Appl. Meteor.,* **3**, 396-409.

Deardorff, J.W., 1978: Efficient prediction of ground surface temperature and moisture, with inclusion of a layer of vegetation. *J. Geophys. Res.,* **83**, 1889-1903.

Manzi, A.O. and S. Planton, 1993: Implementation of the ISBA parameterization scheme fo land surface processes in a GCM - an annual cycle experiment. *Journal of Hydrology,* **155**, 353-387.

Sellers, P.J., W.J. Shuttleworth and J.L. Dorman, 1989: Calibrating the simple biosphere model for Amazonian Tropical Forest using field and remote sensing data. Part I: Average calibration with field data. *J. Appl. Meteor.*, **28**, 727-759.

Shukla, J. and Y. Mintz, 1982: Influence of land-surface evapotranspiration on the Earth's climate. *Science*, **215**, 1498-1501.