APPLICATION OF INVERSE ANALYSIS FOR SURFACE TEMPERATURE PREDICTION OF SEMI-ARID REGION OF N-E BRAZIL

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Abstract. *For this study the data were collected from the micrometeorological tower with 10 meters of height at the experimental site located in the semi-arid region of the Institute-basin of Federal University of Campina Grande in the city of São João do Cariri-PB N-E Brazil. The data of velocity, temperature, specific humidity, net radiation, surface temperature and pressure were collected in the interval of 20 minutes from 28-02-2001 to 09-03-2001. In this study the inverse theory is applied for the estimation of the surface temperature by solving the surface energy balance equation using the forms used in the limited area regional model for obtaining turbulent heat and momentum fluxes in the surface layer. The Levenberg-Marquardt algorithm is used to retrieve value of the wind (considered as a parameter for estimation of bulk Richardson number associated with the turbulent diffusion). It is observed that there are significant differences between the calculated (predicted) and measured values of the surface temperature. Also the measurements errors of the data acquisition system are considered according to the standard deviation furnished by manufacture. Then it is found that the predicted and the measured values of the surface temperature are quite close to each other.*

Key Words: thermophisic properties, ground temperature, inverse problems.

1. INTRODUCTION

The earth's surface temperature is an important parameter in the atmospheric phenomena and processes occurring in the boundary layer. It involves in the surface energy balance, evaporation, evapotranspiration and desertification processes and is also considered as an indicator of environmental degradation and climate change. Further, the determination of diurnal variation of ground surface temperature plays a critical role in the modeling of atmospheric planetary boundary layer and, by extension, in mesoscale numerical weather prediction (NWP) models. At ground surface, the various flux terms of the energy balance equation unfortunately are not a linear function of the surface temperature (T_g). These terms include: short wave flux due to solar radiation, infrared flux from the atmosphere, thermal emission at the ground, sensible heat transfer between the air and the ground, latent heat flux due to evaporation/condensation and rainfall, and soil heat flux due to the vertical thermal gradient within the soil. Of these, only the first has well defined, externally forced diurnal and seasonal periods and then only under cloudless skies. Therefore, the difficulty in obtaining an analytical solution for the heat equation lies not in the differential equation itself but rather in accurately prescribing the external forcing and in the nonlinear feedback nature of the upper boundary condition. For this reason, the value of the surface temperature (T_g), in NWP-type models are usually determined by numerical methods. Deardorff (1978) found the so-called force-restore method to be the most satisfactory. This method is used in many NWP-type models (Zhang and Anthes, 1982; Tuccillo and Phillips, 1986). Bhumralkar (1975) has performed numerical experiments on the computation of ground surface temperature in an Atmospheric General Circulation Model. Recently Best (1998) described a model to predict the surface temperature of variety of surfaces, by solving the surface energy balance equation iteratively, using the standard meteorological data. Figueiredo (2000) has used a statistical model to adjust the prognosticated maximum and minimum temperature for 24 hours using the limited area atmospheric model for the city of Botucatu-SP, Brazil. Antonio and França (2000) have estimated the earth's surface temperature using the NOAA-AVHRR satellite data. Peres and Câmara (2002) have studied the estimation of earth's surface temperature using inverse theory and METEOROSAT Second Generation (MSG) data, based on the on the method of Watson (1992) and Faysash (1999). In this study the inverse theory is applied for the prediction of the surface temperature of semi-arid region of N-E Brazil, by solving the surface energy balance equation.

Recently the inverse theory has received much attention in micrometeorological studies and several successful applications have been presented within the last decade (Siquira et al., 2003, 2000; Raupach, 1988; 1989a,b; Denmead and Roupach 1993; Denmead, 2000, Denmead et al., 2000; Katul et al., 1997, 2001; Massman and Weil 1999; Leuning 2000; Leuning et al., Simon, et al., 2002; Hsich, et al., 2003; Paz, 2002; Paz et al., 2004; among others). In short the inverse modeling technique uses a measured data set of an atmospheric quantity and an assumed model relationship that describes the physical processes of the quantity to produce the measured data set as a set of parameters (Wolff and Bange, 2000). In other words the technique uses appropriate model assumptions that are based on theoretical assumptions to fit measured data (Zittel, et al., 2002). The technique is based on the assumption of a relationship (operator **D**) between the model parameter **M** and the measured parameter **C,** i.e., $\mathbf{C}_{obs} = \mathbf{D}\,\mathbf{M}_{mod}$, where, **D** is also called as dispersion matrix (ixj terms), $\mathbf{C}_{obs}$ is a vector (with i rows) and $\mathbf{M}_{\mathbf{obs}}$ is a vector with j rows.

2. SITE AND INSTRUMENTATION

The experimental site is located in the semi-arid region of the institute-basin of Federal University of Campina Grande-UFCG, in the city of São João do Cariri (latitude $07^0$22'44"S and longitude $36^0$32'00"W) and altitude of approximately 465 m, covered with sparse vegetation of the type caatinga and pasture. The region is considered to be semi-arid with less abundant rainfall from the month of February to April. The soil is rocky and the vegetation is sparse and gives the appearance of strips alternatively between the rocky soil and the sparse vegetation.

A micrometeorological tower of ten meters is located in the middle of the sparsely vegetated surface of the experimental site described above. The net radiometer and radiometer (both Campbell Scientific Inc., Q-7) were installed at the canopy top at the height of 6 meters. The soil temperature were measured by the temperature sensor (Campbell Scientific Inc., model 108) of precision of $0{,}001^0$ C, installed one at the surface (to measure surface temperature) and the other three in the depths of 2, 5 and 15 cms. The air temperatures were measured by, two Cromel Constantin thermocouple of 25 μm and 74 μm diameters, one located at the surface of the vegetation and the other at 10 m of height. The temperature and specific humidity were also measured by a HMP 45C (Campbell Scientific Inc.) thermo hygrometer located at the top of the bushes. The wind velocity, temperature and specific humidity were measured at 1,5 and 10 m heights. The precipitation and the humidity of the soil were also measured. The observations for this work were made during the period from 28-02-2001 to 09-03-2001. The measurements were taken during the interval of 20 minutes by the data acquisition system CR23X of Campbell Scientific Inc., continuously connected to the battery of 12 volts and 55 AH, accompanied to a solar panel of potential of 50 W. The datalogger is programmed to control all the experiment and to unload the data to the microcomputer.

3. METHODOLOGY AND DISCUSSION

The prognostic equation for the surface temperature, T_g is obtained in Limited Area Model (ALM) at grid points from the surface energy balance equation suggested by Bhumralkar (1975) as:

$$G(0)=S - R_L - LE - H \tag{1}$$

Where,
G(0) represents soil heat flux at the surface
S is a solar energy reaching the earth's surface
R_L is net long wave radiation
LE represents latent heat flux and
H represents sensible heat flux
The soil heat flux is defined by

$$G(z,t) = \lambda \frac{\partial T_s(z,t)}{\partial z} \tag{2}$$

where λ is a soil thermal conductivity and.T_s is soil temperature.

T_s may be obtained by solving the soil heat transfer equation for (simplification) homogeneous soil

$$\frac{\partial T_s}{\partial t} = \frac{\lambda}{c_g} \frac{\partial^2 T_s}{\partial z^2} \tag{3}$$

with boundary condition

$$T_s(z,t) = \overline{T} + \Delta T_0 \text{ sen } (\varpi t) \tag{4}$$

as,

$$T_s(z,t) = \overline{T} + \Delta T_s \, e^{-z/D} \text{ sen}(\varpi t - z/D) \tag{5}$$

where

c_g is soil specific heat (J kg^{-1} K^{-1})
$\overline{T}$ is daily mean temperature

$\overline{T}$ ΔT_0 is surface thermal amplitude
ϖ is frequency of oscillation and

$$D = \left(\frac{2\lambda}{c_g \varpi} \right)^{1/2} \text{ is damping depth.}$$

Now substituting Eq. (5) in Eq. (2) one may have

$$G(z,t) = \left(\frac{\varpi c_g \lambda}{2} \right)^{1/2} \left[\frac{1}{\varpi} \frac{\partial T_s(z,t)}{\partial t} + T_s(z,t) - \overline{T} \right] \tag{6}$$

A rate of temporal variation of a layer at depth Z may be written in the form

$$c_g \frac{\partial T_s}{\partial t} = -\left[\frac{G(Z,t) - G(0,t)}{Z} \right] \tag{7}$$

As, the aim of this work is to calculate the surface temperature of the soil, rather than the mean temperature of soil at finite depth, so Eq. (7) may be applied for a layer of soil of 1 cm depth, and considering that the mean temperature of this layer is approximately equal to the surface temperature, T_g; so one may have

$$T_s(1,t) \approx T_g\ (t) \tag{8}$$

Eq. (8) represents a reasonable approximation for the present work, seeing the uncertainty of the definition of the surface itself specially for the regional scale. Now from Eqs. (2), (3) and (4) for the layer of thickness Z=0,01 one may have

$$c_g \frac{\partial T_g}{\partial t} = G(0) - \left(\frac{\varpi\, c_g \lambda}{2} \right) \left[\frac{1}{\varpi} \frac{\partial T_g}{\partial t} + T_g - T \right]$$

which can be solved for G(0) following Bhumralkar (1975) as:

$$T_g^{\,t+\Delta t} = T_g + \frac{f_1}{f_2 + f_3} \tag{9}$$

where,

$$f_1 = R_n - LE - H - \left(\frac{\lambda\, c_g \varpi}{2} \right)^{1/2} \left(T_g - \overline{T} \right) \tag{10}$$

$$f_2 = \left(\frac{c_1}{\Delta t} \right) + 4\sigma_B\, T_g^{\,3} \tag{11}$$

$$f_3 = c_H \left(1 + \frac{L}{c_p} w_g \frac{dq_s}{dT_g} \right) + \left(\frac{\lambda\, c_g \varpi}{2} \right)^{1/2} \tag{12}$$

where other notations are defined in (Bhumhalkar, 1975 and Paz, 2002).

Now, the surface temperature can be calculated from Eq. (9) where H and LE are obtained from the similarity theory of Monin-Obukhov, using forms of Bussinger et al., (1971) or other formulation for the similarity empirical functions of Monin-Obukhov.

The method of Levenberg-Marquardt (Press et al., 1992) is used to solve the problem of identification. The adjust of the merit function S^2 can be written for the temperature in the form

$$S^2(\beta, t) = \sum_{i=1}^{N} \left[\frac{T^i{}_{g(obs)} - T^i{}_{g(mod)}}{d_i} \right] \qquad (13)$$

The soil's prognostic temperature $T_g{}^i$ presents the following functional dependency:

$$T_g{}^i = F(R_n, T_g, LE, H, t, z, z_0, \lambda, \varpi, c_g, K_g, \rho, c, c_p, q, q_s, u) \qquad (14)$$

The data obtained were submitted to consistency analysis based on the conventional criteria within the practical norms of the operational meteorology.

The process of the identification requires a preliminary analysis of the sensibility of the variable obtained by model in terms of the parameters and objectives of the estimation. The coefficient of reduced sensibility is used and represented graphically to evaluate the possibility of the satisfactory application of the method of identification. The physical properties of air, soil and wind velocity are evaluated. Although the analysis of wind velocity is a variable furnished by the micrometeorological station, justifies the analysis because it evaluates indirectly the sensibility of the model to the Richardson number to include the effect of the stability of the boundary layer. So, finally, one may have

$$T_g{}^{t+\Delta t} = F(R_n, T_g, t, z_0, \lambda, \varpi, c_g, c, c_p, u, t) \qquad (15)$$

A numerical code is developed in FORTRAN language to perform an analysis for the prognostic model for the surface temperature and to calculate the coefficient of sensibility. In this case the sensibility of the temperature surface in relation to wind velocity (u), considered as a parameter in calculating the Richardson number. The surface temperature calculated from the model and the corresponding measured values is shown in Figure (1).

Usual values were considered for roughness (z_0), the frequency of the surface temperature (ϖ), humidity parameter (w_g), specific heat of soil (c_g), thermal conductivity of the soil (c),and specific heat of air (c).

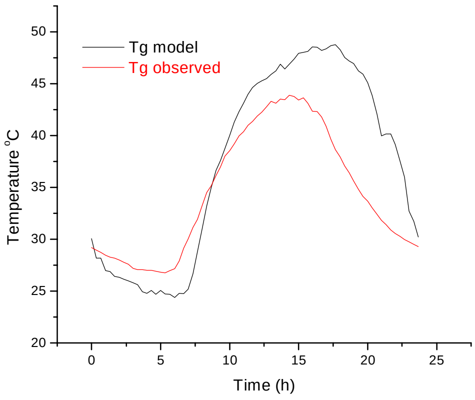

Figura 1. Observed and modeled values of Surface temperature before identification.

The performance of the model as seen in the figure are satisfactory, although there are mean relative errors of 12.7%.

Figure (2) represents the modeled and measured values of the surface temperature after the identification of the parameters. The relative mean errors after identification, reduces to 1.4%. It can be seen from Figure (2) that the dispersion of the data has reduced significantly, and there is a quite good agreement between the modeled and measured values after the identification. This shows the efficiency of the inverse modeling for surface temperature prediction. Further details of this study will be reported in a future paper.

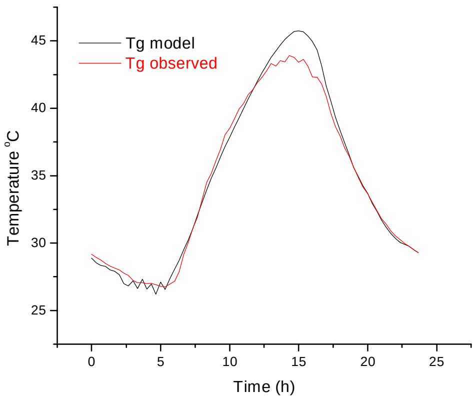

Figura 2. Observed and modeled values of surface temperature after identification.

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