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The Relationship Between Incoming Solar Radiation and Land Surface Energy Fluxes

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1. Introduction

Incoming solar radiation (R) is the driver of the land surface energy fluxes: latent heat (E) or soil evaporation (i.e. the natural transfer of water from the topsoil to the atmosphere, although it might include also condensation), sensible heat (H) so-called because it can be "felt" (i.e. it is related to temperature differences between the surface and the atmosphere), and the ground heat flux (G) so-called because it is restricted to the interior of the ground (i.e. it is related to temperature differences between ground layers). All this seems rather obvious during the daytime, when R provides the energy input and apparently the output of E, H and G balance it. However the situation is less clear at night when R is nil but E, H and G may not vanish, while the energy balance must be kept. To understand this simple idea lets consider the following example: under certain conditions, like wet soil in low- and mid-latitudes, E may be considered almost as proportional to R; that is $E \sim (a_1 \times R) + OT$, where a_1 is a proportionality factor (not necessarily a constant) and OT are other terms (in this case: H and G, which are usually smaller than E and $(a_1 \times R)$, but there could be other terms). If we are able to estimate E, R, H and G, or these terms are somehow known, we can tentatively solve for a_1, which could characterize the relationship between R and these fluxes, and the term $(a_1 \times R)$ is called the net radiation (R_n); i.e. the part of R which is actually balanced by E, H and G. The problem, however, is not trivial because even if we restrict ourselves to this simplified case, and we could measure R and E, the smaller terms H and G would have to be assessed as well. Nevertheless we believe that this difficulty may be partially overcome by empirically modeling E. Recall that calculating a_1 with observations of total and net solar radiation: $a_1 = (R_n)_{obs}/R_{obs}$ may not be appropriate for our purposes, because it would not consider E, H and G, and a_1 is but an element of a vector $\mathbf{a}$ yet-unknown. Therefore our goal is to develop a full energy flux model to show that indeed the relationship between R and surface fluxes may be achieved in this empirical way. That is, in this work we will attempt to approximate these surface energy fluxes by simultaneously modeling them based on a simplified energy balance. Although, due to their importance in many environmental issues (from crop-field irrigation (Brisson & Perrier, 1991), (Allen et al., 1998), to the study of the global water cycle (Huntington, 2006)) one usually first looks for R_n in order to evaluate surface fluxes; here we will attempt to model E, H and G in order to estimate $R_n = (a_1 \times R)$. However even if our model is successful, that is even if in general it correlates well with observations, we will examine the situations in which the model fails, the energy balance does not hold, and a simple relationship between R and surface fluxes cannot be established. Thus the limitations of this study should serve as a motivation for future work.

1.1 Modeling approach

The simplest way to determine if a model is appropriate is to compare it with observations. Even though assessing E in general is difficult because it depends not only on the ambient conditions, but also on soil composition and moisture content, here we use observations of soil evaporation (E_{obs}) obtained through micro-lysimetry (Figure 1). That is, we will try to model E from E_{obs} and calculate their correlation coefficient r(E_{mod}, E_{obs}) to evaluate the appropriateness of our model: E_{mod} = $(R_n - H - G)_{mod}$. Although diverse efforts have been devoted to model E for different applications (Penman, 1948), (Priestley & Taylor, 1972), (Twine et al., 2000), (Brutsaert, 2006), (Agam et al., 2010), all these efforts possess different limitations and degrees of difficulty. In other words, there is no general way to model E which is practicable for all situations (Crago & Brutsaert, 1992), and thus we must develop an *ad hoc* E-model to estimate a_1 for our particular case. In this sense we will focus on a relatively simple case, the diurnal variation of bare soil evaporation when water is not a limiting factor (for example wet sand with substantially more than 5% of water (Pavia & Velázquez, 2010)). That is, when the main diurnal surface energy balance is between R_n and E: $R_n \sim E$. Previous works in cases similar to the present one have confirmed that daytime E is highly correlated with R (Pavia, 2008); therefore we should expect our model to reflect daytime better than nighttime conditions. We will perform an experiment with an evaporating tray containing a small amount of wet sand (~35 Kg maximum), so that E_{obs} should be easier to measure throughout the day than R_n. Our hypothesis is that we can obtain E_{mod} from a small number of standard meteorological observations and experimentally-obtained variables, which are chosen by their assumed relationship to energy terms; namely R, air temperature (T_a), surface temperature (T_o), soil temperature (T_s) and observed soil evaporation (E_{obs}). Therefore we will try to fit E_{obs} to a linear combination of terms derived from the above variables. Specifically $E_{mod} = E_{obs} \sim \mathrm{L}(R, \Delta T_a, \Delta T_s)$, where, as it will be explained in the next section, the model E (E_{mod}) is achieved from R, $\Delta T_a = T_o - T_a$, $\Delta T_s = T_o - T_s$ and E_{obs}, through a multiple regression procedure yielding a vector **a** which includes a_1 among other parameters. This approach is physically-motivated by the primary land surface energy balance:

$$R_n = E + H + G, \tag{1}$$

Where R_n would be approximated by the R term (Gay, 1971) and the sensible heat flux H and the ground heat flux G would be similarly approximated by the ΔT_a and ΔT_s terms, respectively. Therefore it is anticipated that the multiple-regression parameter-vector **a** resulting from our model may give a preliminary assessments of the relative importance of R on each of these surface energy flux densities.

2. Methods

In this section we describe the original technique to find the relationship between R and the surface energy fluxes. This includes the experimental evaluation of E, the approximation made of $H \sim \Delta T_a$ and $G \sim \Delta T_s$ from the observed temperatures, and the multiple regression method to optimize these approximations.

2.1 The experiment

A 27-d experiment was performed from 12 February to 11 March 2011, in Ensenada, Mexico (31° 52′ 09′′ N, 116° 39′ 52′′ W) at 66 m above mean sea level. It consisted of a bird-guarded wet-sand evaporating tray (equipped with temperature sensors at depths z_o = 0.02 m, for T_o, and z_1 = 0.07 m for T_s) set on an electronic scale to register the varying weight (w) next to a meteorological station recording R and T_a among other variables (see Figure 1). All variables are registered at Δt = 300 s intervals, and the total number of samples is N = 7776. See (Pavia & Velázquez, 2010) for more details on similar experiments.

Fig. 1. The experimental setup: the meteorological station, the evaporative tray and the weighing scale used in the study.

2.2 The empirical approach

We begin by calculating a time series of weight-change time-rates $\Delta w_i = (w_{i-1/2} - w_{i+1/2}) / \Delta t$ [Kg s^{-1}], where $w_{i-1/2}$ and $w_{i+1/2}$ represent smooth averaged weight values (e.g. precipitation has been filtered out), which is used to obtain a time series of observed evaporation, $(E_{obs})_i = \lambda \times \Delta w_i / A$ [W m^{-2}], where $\lambda = 2.45 \times 10^6$ [J Kg^{-1}] is the latent heat of water vaporization, and $A = 0.23$ m^2 is the evaporating surface area. Then we fit $(E_{obs})_i$ to the corresponding series of R_i, $(\Delta T_a)_i$, and $(\Delta T_s)_i$, that is:

$$(E_{mod})_i = a_1 R_i + a_2 (\Delta T_a)_i + a_3 (\Delta T_s)_i; \; i = 1,2,\ldots,7776. \tag{2}$$

And the problem is now reduced to finding the values of a_1, a_2, and a_3.

2.3 The statistical method

A simple technique to try to solve the above problem is a least-square multiple regression procedure, which in this case is formulated as follows. First we construct the vector:

$$\mathbf{y} = \left[(E_{obs})_1 \; (E_{obs})_2 \; \ldots \; (E_{obs})_{7776} \right], \tag{3}$$

and the matrix:

$$\mathbf{X} = \begin{bmatrix} R_1 & R_2 & \ldots & R_{7776} \\ (\Delta T_a)_1 & (\Delta T_a)_2 & \ldots & (\Delta T_a)_{7776} \\ (\Delta T_s)_1 & (\Delta T_s)_2 & \ldots & (\Delta T_s)_{7776} \end{bmatrix}. \tag{4}$$

Then we posit that $\mathbf{y}_{mod} = \mathbf{aX}$, where $\mathbf{a} = [a_1 \; a_2 \; a_3]$ is the coefficients-vector to be found. Using (3) and (4) this is done by minimizing $Z \equiv (\mathbf{y} - \mathbf{aX})(\mathbf{y} - \mathbf{aX})^T$; that is $\partial Z / \partial \mathbf{a} = 0$, which finally yields $\mathbf{a} = \mathbf{y}\mathbf{X}^T (\mathbf{X}\mathbf{X}^T)^{-1}$ and consequently $\mathbf{y}_{mod} = [\, (E_{mod})_1 \; (E_{mod})_2 \ldots (E_{mod})_{7776} \,]$.

3. Results and discussion

The above procedure gave $a_1 = 0.48$, $a_2 = -3.77$ [W m^{-2} K^{-1}], $a_3 = -14.25$ [W m^{-2} K^{-1}], which are used in (2) to evaluate E_{mod} [W m^{-2}]. The comparison of the evolution of E_{mod} and E_{obs} is presented in Figure 2. These two series have a correlation coefficient $r(E_{mod}, E_{obs}) = 0.90$, which indicates that our method has been rather successful to model E. In addition we will try to relate each term of E_{mod} to surface energy fluxes using (1); that is $a_1 R = E_{obs} - a_2 \Delta T_a - a_3 \Delta T_s$, or $0.48 R = E_{obs} + 3.77 \Delta T_a + 14.25 \Delta T_s$. The most important term of the model is $0.48 \times R$, because most of E occurs during the daytime. This means that here the net radiation is principally proportional to the absorbed radiation: $R_n \sim b_1 (1 - \alpha) R + b_2 T_a + b_3 T_o$, where $b_1 = a_1 / (1 - \alpha)$, $b_2 = 0$, $b_3 = 0$, and α is the wet-sand albedo. Moreover if $0.10 < \alpha < 0.25$ we obtain reasonable values for b_1: $0.53 < b_1 < 0.64$ (Gay, 1971), (Stathers et al., 1988). Obviously this assumption is not valid during the nighttime, when $E \sim 0$ but not nil. Likewise we may consider the sensible heat flux to be approximated by $H = -a_2 \times \Delta T_a$.

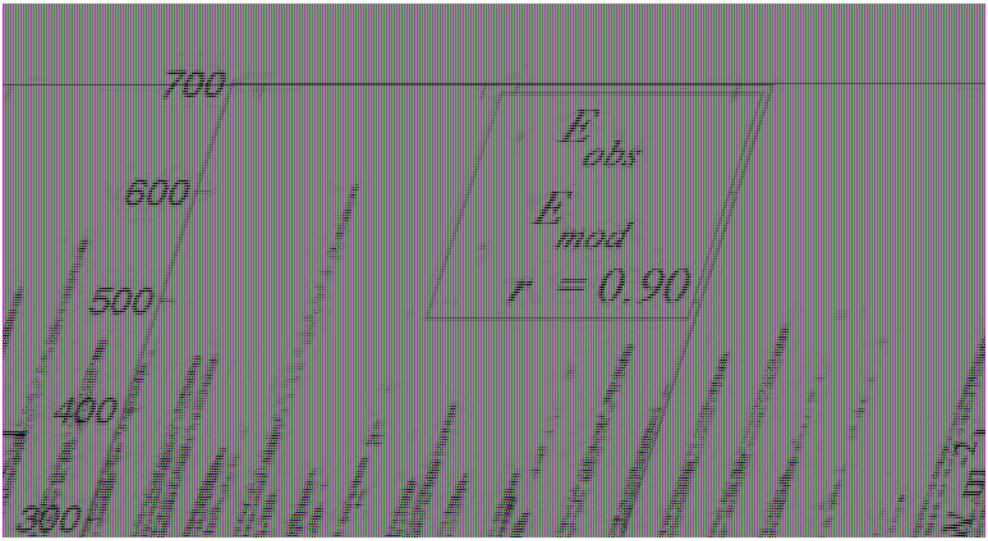

Fig. 2. Comparison between modeled soil evaporation E_{mod} and observed soil evaporation E_{obs}.

Comparing this approximation with its theoretical expression (Stathers et al., 1988) $H = (\rho\ c_p\ / r_{aH}) \times \Delta T_a$, where $\rho \sim 1.0$ [Kg m^{-3}] is the density of air, $c_p \sim 1000.0$ [J Kg^{-1} K^{-1}] is the specific heat capacity of air at constant pressure, and r_{aH} is the aerodynamic resistance to heat transfer between the surface and z ~ 2.0 m (the height at which T_a is measured), we straightforwardly get $r_{aH} = -(\rho\ c_p\ /\ a_2) \sim 265$ [m^{-1} s]. This value is a very good approximation to the one obtained from its simplest theoretical form, i. e. for stable atmospheric conditions (Webb, 1970):

$$r_{aH} = \frac{[\ln\ (z/z_T) + 4.7 \times (z\ /\ L)] \times [\ln\ (z/z_M) + 4.7 \times (z\ /\ L)]}{k^2 u} \sim 293\ [\mathrm{m^{-1}s}], \tag{5}$$

where we have chosen in (5) the following values, $z_T = 0.0002$ m for the surface roughness length for sensible heat transfer, $L = 10$ m for the Monin-Obukhov length, $z_M = 0.0005$ m for the surface roughness length for momentum, $u = 2.0$ [m s^{-1}] for the mean wind speed at z = 2 m height, and $k = 0.4$ is the von Kármán constant (Stathers et al., 1988). Similarly if we approximate the soil heat flux obtained by integrating the heat conduction equation (Peters-Lidard et al., 1998): $G = (\kappa\ \partial T/\partial z)_o \sim \kappa\ (\Delta T_s\ /\Delta z)$, where κ is the soil thermal conductivity and $\Delta z = (z_1 - z_o) = 0.05$ m, and compare it with our estimate of $G = -a_3\ \Delta T_s$ we straightforwardly obtain $\kappa = (14.25 \times 0.05) = 0.7125$ [W m^{-1} K^{-1}], which is a reasonable value, although somewhat low since for water $\kappa = 0.6$ [W m^{-1} K^{-1}] and for soil minerals $\kappa = 2.9$ [W m^{-1} K^{-1}] (see Table I of Peters-Lidard et al. (1998)). Therefore we may consider that as $E_{obs} \sim E$, $a_1\ R \sim R_n$, $a_2\ \Delta T_a \sim H$, and $a_3\ \Delta T_s \sim G$, the surface energy balance is approximately satisfied ($R_n \sim E + H + G$). And if we calculate the mean diurnal variations during our 27-d observation period (defined positive toward the surface):

$$< F >_i = \sum_{j=1}^{27} (F_i)_j, \quad i = 1,2,\ldots,288, \tag{6}$$

where F is any of the approximated energy fluxes considered here, we observe that, as expected, most of the time the magnitudes of $< H >$ and $< G >$ are smaller than those of $< E >$ and $< R_n >$ (up to about one order of magnitude during the daytime). However their progresses during the day are more telling (see Figure 3); that is the $< R_n >$ maximum around noon, the $< G >$ minimum at mid-morning and the $< H >$ minimum in the afternoon; all suggest that our empirical approach is appropriate. Perhaps it can be improved with better observations, but these results are definitely encouraging.

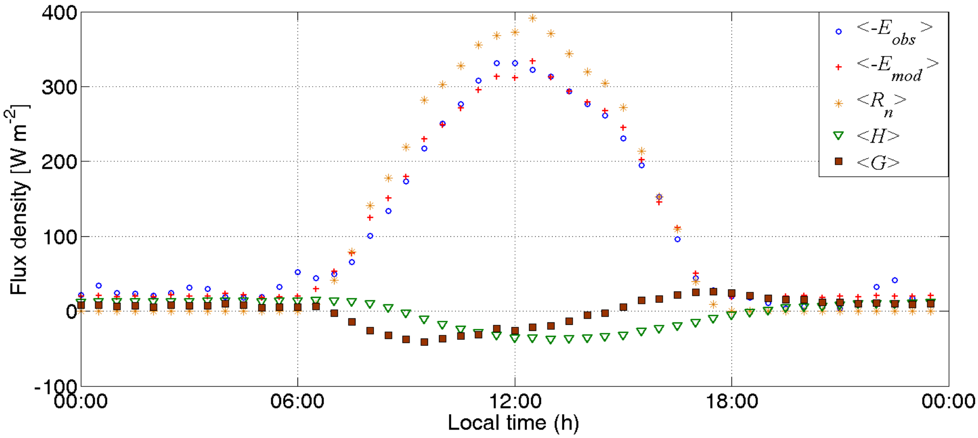

Fig. 3. The diurnal variation of the different approximate energy fluxes (6) defined positive towards the surface. Note that evaporations are plotted with a negative sign. For clarity fluxes are plotted at 5Δt intervals.

Nevertheless we must acknowledge the limitations of our empirical model. Since in this case E so strongly depends on R and only marginally on ΔT_a and ΔT_s, we cannot include more terms, for example terms related to wind speed and relative humidity (which are also related to R), because that could render the model unstable as these terms do not significantly contribute to explain variance. For example if we focus on 5 March 2011 (see Figure 4), a particularly windy and dry day apparently resulting from a brief Santa Ana event (see Raphael (2003) for a description of this kind of events), we observe that E_{mod} underestimates E_{obs}, especially during the nighttime early hours. In this situation E_{mod} can only be appropriately modeled if we could include in our algorithm wind and humidity observations, which as mentioned before is not possible. Yet the model clearly indicated that in this case other evaporative causes, besides the ones related to energy fluxes, were also related to R and playing a role in E. And, on the other hand, when we tested our model with independent data (Figure 5); that is using the current values of the vector **a** with new observations R_i, $(\Delta T_a)_i$, and $(\Delta T_s)_i$ for the period 17-29 May 2011, we found that now the model overestimates the observations: $(\Sigma\, E_{obs})\ /\ (\Sigma\, E_{mod}) \sim 0.7$.

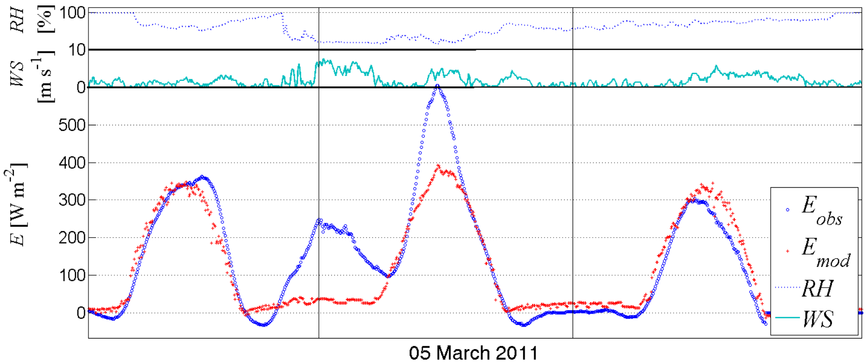

Fig. 4. Close up centered on 05 March 2011. Wind speed *WS* and relative humidity *RH* are also shown schematically.

Although correlation were still high, $r(E_{mod}, E_{obs})$= 0.83, in this situation E_{obs} were limited by the lack of moisture in the wet sand. Here again the model clearly indicated that in this case the sand was drier than when we first calculated **a**, as the average weight of the evaporating tray in this case was 21.0 Kg, compared to 25.5 Kg in the original experiment.

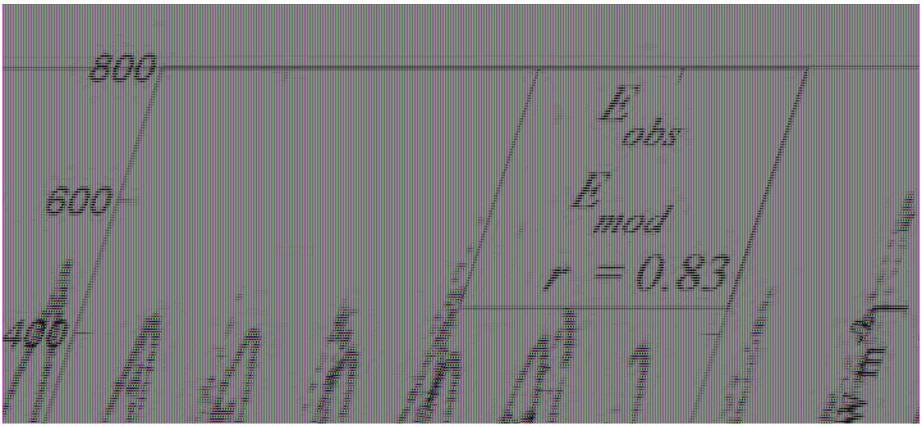

Fig. 5. Same as Figure 2, but for the test period with independent data. Sand was drier after 26 May.

Finally we compared our method (Figure 6) with a previous technique developed for modeling 7-h (08:30 to 15:30 h, local time) total soil evaporation (Pavia, 2008). In this case evaporation, in mm, is given by:

$$E^{(1)}_{mod} = 0.8 \times \left[0.1525 \times (\overline{T}_a - 18) + 0.0053 \times (\overline{R} - 404) \right] + 2.2 \text{ [mm]}; \quad (7)$$

where the overbar indicates dimensionless mean values during the 7-h observing period, and the corresponding values for our model are computed by:

$$E^{(2)}_{mod} = \sum_{k=1}^{84} (E_{mod})_k \times \Delta t / \lambda \ [\text{mm}], \quad (8)$$

where k = 1 corresponds to 08:30 h local time. The higher correlation given by the second model: $r(E^{(2)}_{mod}, E_{obs}) = 0.9$ versus $r(E^{(1)}_{mod}, E_{obs}) = 0.7$ of the first model, furthermore suggests that the new model improves the predictions.

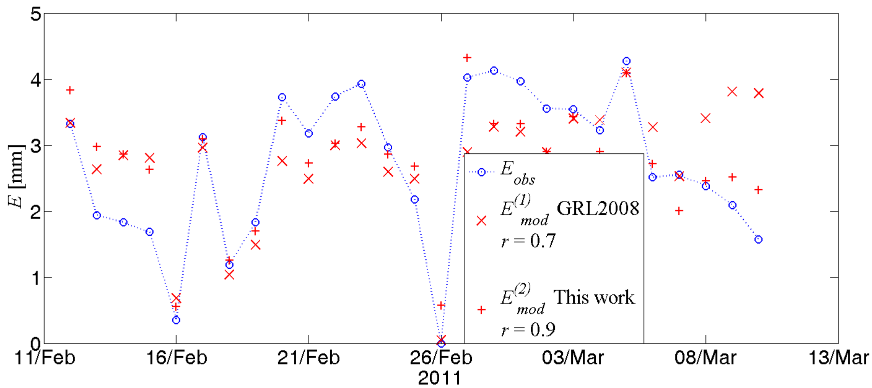

Fig. 6. Comparison of the 7-h total E obtained with present model $E^{(2)}_{mod}$ and that obtained with the model GRL2008 of Pavia (2008) $E^{(1)}_{mod}$.

4. Conclusions

The main objective of this work, which is the optimal estimation of **a** by the empirical modeling of soil evaporation, has been achieved (see Figures 2 and 6). This vector represents the relationship between solar radiation and surface energy fluxes. Nevertheless it has a drawback, since a_1 is proportional to R it is pointless when the incoming solar radiation is nil. However this empirical approach, physically motivated by the surface energy balance, yields promising results by still suggesting an energy balance at night; i.e. when $R = 0$. For example, we conclude that in this case the net radiation $R_n = a_1 R \sim b_1 (1 - \alpha) R$ is largely a function of the absorbed solar radiation, because here we are dealing with substantially wet sand and most of the evaporation occurs during the day (see Figure 7); but we also conclude that the sensible heat flux $H = a_2 \times \Delta T_a \sim (\rho\, c_p / r_{aH}) \times \Delta T_a$, since the value obtained here for the aerodynamic resistance to heat transfer $r_{aH} = 265$ m^{-1} s is very close to its theoretical estimation $r_{aH} = 293$ m^{-1} s obtained with (5) (see Figure 8). And, similarly, we conclude that the ground heat flux $G = a_3\, \Delta T_s \sim \kappa\, (\Delta T_s / \Delta z)$, since the value obtained here for the thermal conductivity $\kappa = 0.7125$ [W m^{-1} K^{-1}] is within the expected range (Peters-Lidard et al., 1998) of values: 0.6 to 2.9 [W m^{-1} K^{-1}] (see Figure 9).

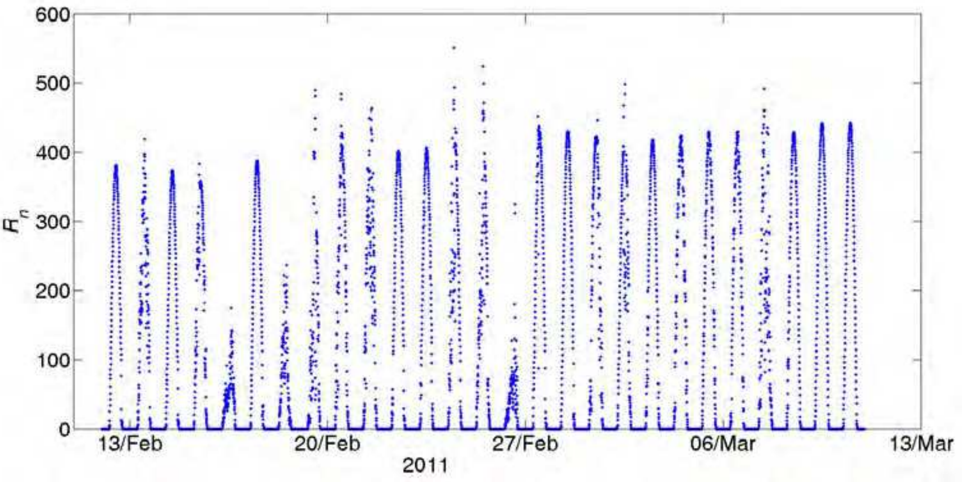

Fig. 7. Time series of the modeled net radiation.

The shapes of the progresses (see Figure 3) of their mean diurnal values (6) furthermore support these conclusions.

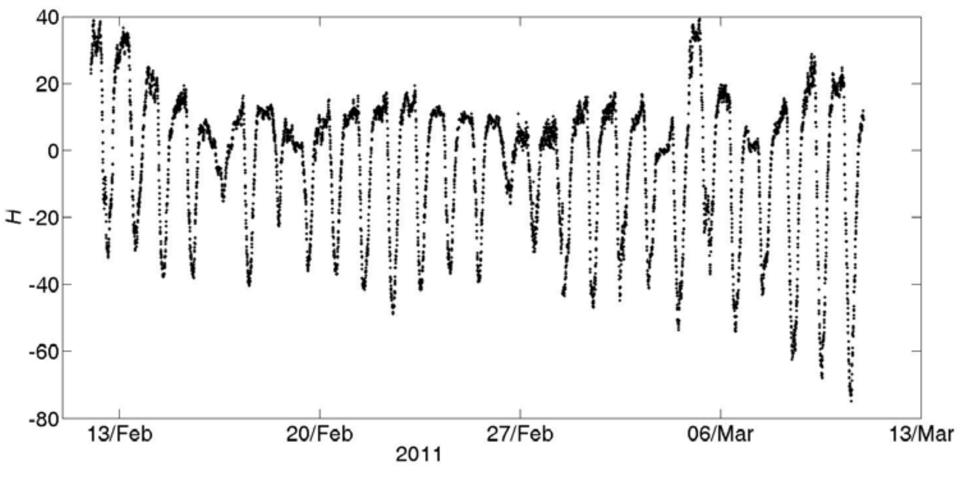

Fig. 8. Time series of the modeled sensible heat.

However our empirical model is limited because statistically it is not possible to have more than a few terms. Considering wind speed and relative humidity terms in our algorithm may result in better predictions during Santa Ana events. Considering single temperature terms may improve the net radiation term R_n = b$_1$ (1 – α) R + b$_2$ T_a + b$_3$ T_o, as b$_2$ and b$_3$ become non-zero. This in turn may improve the estimations of the H and G terms, which may result in better predictions when the wet sand becomes drier, for example. Efforts to overcome these limitations are in progress, i.e. trying to model the difference between evaporation and net radiation (E_{mod} - R_n) = L(T_a, ΔT_a, ΔT_s) or L(T_s, ΔT_a, ΔT_s), since T_a and T_s are correlated. Nevertheless the present empirical approach provides an interesting alternative to more sophisticated methods.

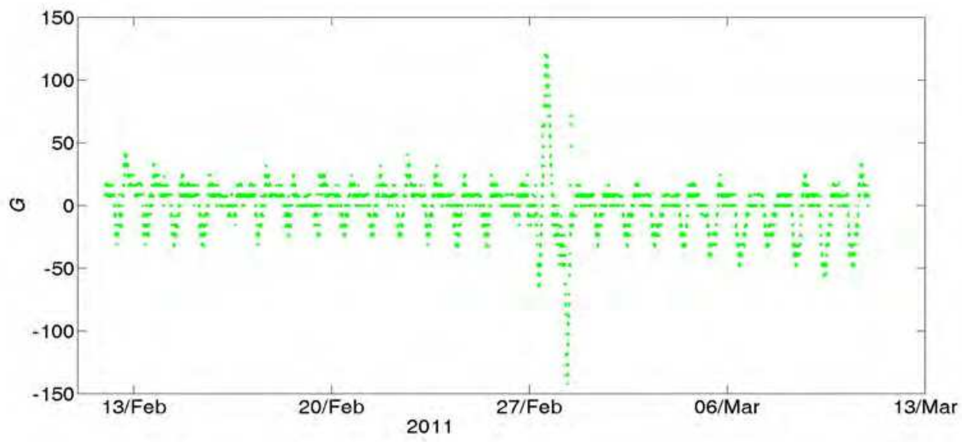

Fig. 9. Time series of the modeled ground heat flux.

5. Summary

The relationship between incoming solar radiation and the surface energy fluxes E, H and G has been investigated by empirically modeling E through a multiple regression method. We propose this new empirical model of wet sand evaporation, which gives excellent results when moisture is not a limiting factor and wind and air humidity are not extreme (see Figure 10), as a means to establish this relationship (represented here by **a**). The algorithm was physically motivated by the surface energy balance $R_n = E + H + G$; i.e. we do not consider other terms (i.e. relative humidity or wind speed). In this sense we measured R, T_a, T_o, T_s, and E_{obs}, in order to model E from R, $\Delta T_a = T_o - T_a$, and $\Delta T_s = T_o - T_s$. Namely E_{mod} = a$_1$ R + a$_2$ ΔT_a + a$_3$ ΔT_s; where E_{mod} is the model E, and the coefficients a$_1$, a$_2$, and a$_3$ are determined through multiple regression. Therefore the model provides also a preliminary assessment of the relative importance of energy fluxes. That is, making $E = E_{obs}$, R_n = a$_1$ R, H = a$_2$ ΔT_a, and G = a$_3$ ΔT_s, we get a$_1$ R = E_{obs} - a$_2$ ΔT_a - a$_3$ ΔT_s. Comparison of model results with observations may serve to identify the active role of other variables (wind speed or air humidity) on evaporation, when the model underestimates observations; or the departure from saturation of the evaporating media, when the model overestimates observations. These two cases represent extreme situations when the relationship between solar radiation and surface energy fluxes can not be established by this simple model.

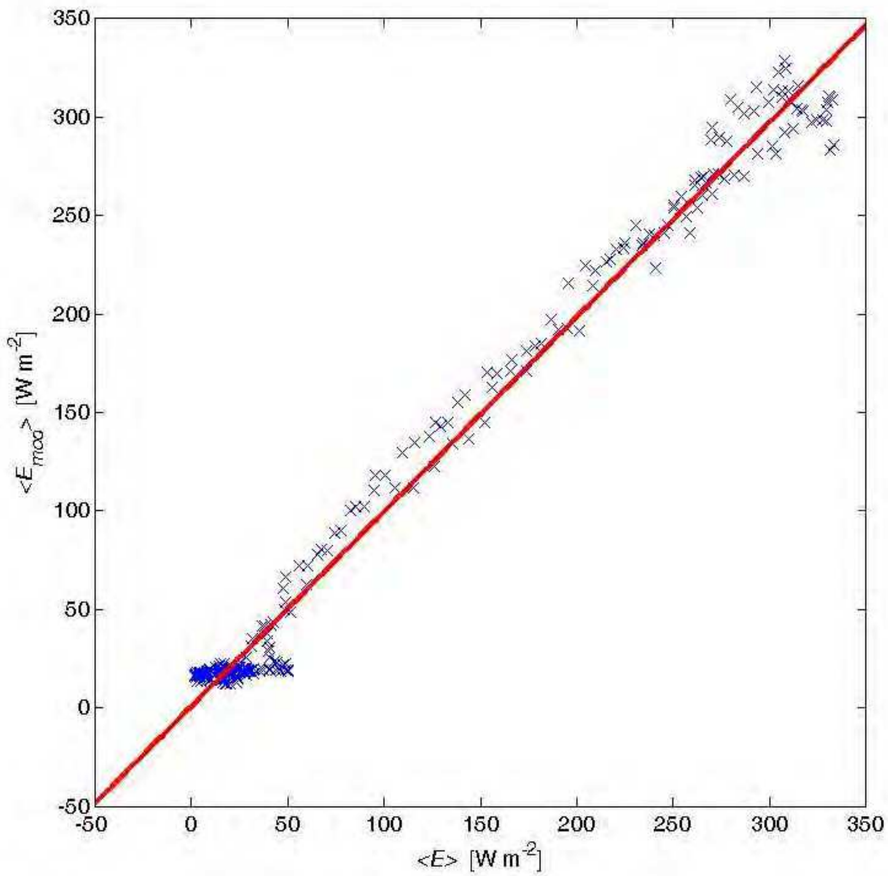

Fig. 10. Mean observed evaporation versus mean modeled evaporation calculated with equation (6). The slope of the linear fit is ~1.0 (red line).

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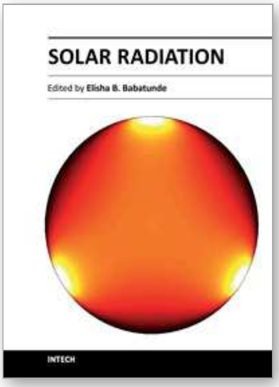

Solar Radiation

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The book contains fundamentals of solar radiation, its ecological impacts, applications, especially in agriculture, architecture, thermal and electric energy. Chapters are written by numerous experienced scientists in the field from various parts of the world. Apart from chapter one which is the introductory chapter of the book, that gives a general topic insight of the book, there are 24 more chapters that cover various fields of solar radiation. These fields include: Measurements and Analysis of Solar Radiation, Agricultural Application / Bio-effect, Architectural Application, Electricity Generation Application and Thermal Energy Application. This book aims to provide a clear scientific insight on Solar Radiation to scientist and students.