Recebido em 13 de Março de 1996

Simulation of weather variables

by

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RESUMO

As variáveis meteorológicas constituem entradas essenciais para os modelos de crescimento cultural e outros tipos de modelos agrícolas. Em todas as estações climatológicas e hidrológicas, a temperatura máxima e mínima do ar, humidade medida de manhã e à tarde, e a precipitação diária estão disponíveis. Frequentemente, a insolação é medida e, raramente, a radiação solar global também é medida. Modelos empíricos simples para a estimativa de variáveis de temperatura, humidade, radiação de pequeno e grande comprimento de onda são apresentados. São considerados o passo diário de integração e outros passos mais pequenos. A validação da maioria dos modelos apresentados foi feita com dados horários obtidos em duas localizações no Centro de Portugal.

RÉSUMÉ

Les variables météorologiques constituent des entrées essencielles pour les modèles de croissance végétal e dautres types de modèles agricoles. Dans toutes les stations climatologiques e hydrologiques, la température maximum e minimum de l'air, l'humidité mesurée au matin et dans l'après-midi, et la précipitation journalière sont disponibles. La durée journalière d'insolation est mesurée souvant, au contraire du rayonnement global qui est mesuré rarement. Des modèles empiriques simples pour l'estimative des variables de température, humidité, rayonnement de courtes et grandes longueurs d'ondes son presentés. Le pas d'integration journalier et des pas plus courts sont considerés. La validation de la plus part des modèles présentés est faite avec des données horaires obtenues dans deux endroits du Centre du Portugal.

SYNOPSIS

Weather variables are an essential input for growth models and other types of agricultural models. In all climatological and hydrological stations, maximum and minimum air temperature, humidity measured in the morning and in the afternoon, and daily precipitation are available. Often, sunshine duration is measured and, seldom, global radiation is also measured. Simple empirical models for the estimation of variables of temperature, humidity, short- and long-wave radiation are presented. Both daily step and shorter steps are considered. Validation of most models presented was done with Portuguese hourly data, from two location in Central Portugal.

1. INTRODUCTION

Growth models and other types of agricultural models need weather information. In Portugal, most climatological stations measure maximum and minimum air temperature; dry and wet bulb temperature, measured at least twice a day; sunshine duration; and precipitation. This basic information is also available for the stations of the hydrological network. Unfortunately, from the stand point of the agricultural modeller, humidity is published as monthly mean relative humidity, many times averaged over a number of years. Time-averaged means for derived ratios such as relative humidity are useless (Linacre, 1992). Actual vapour pressure should be published instead. The common practice of using mean relative humidity in conjunction with temperature to calculate mean actual vapour pressure is, therefore, a rough approximation.

Other environmental variables that are not commonly measured modulate the response of organisms to their physical environment (e. g., mean temperature in the light is more important for photosynthesis and to some plants' development). Also, some processes are best modelled with time steps shorter than a day (e. g., detailed canopy absorption of radiation).

In this paper, some weather variables are derived from the simple elements referred to above, which are readily available in nearby stations of the Portuguese climatological or hydrological networks. Daily averaged and instantaneous quantities of temperature, humidity, and solar and terrestrial radiation are analysed from the modeller's perspective. The approaches are as simple as possible, because we feel that this will impinge on its use. Validation of the simple models used is, whenever possible, made with data collected in Portugal. Comparisons of many alternative models are avoid, because of space constraints.

2. MATERIALS AND METHODS

Data used for the validation of the models presented in this paper come from Instituto Geofísico Infante D. Luís (IGIL), Lisbon (latitude: 38° 43' N; longitude: 9° 9' W; 77 m above M. S. L.), and from Quinta da Amoreira (Q. A.), near Cartaxo (latitude: 39° 10' N; longitude: 8° 43' W; 7m above M. S. L.).

In IGIL, global radiation (St), diffuse radiation, and duration of sunshine hours (n) were measured with a Moll-Gorczinsky type of pyranometer (from Kipp and Zonen), another Moll-Gorczinsky with shading ring, and a Campbell-Stokes sunshine recorder, respectively. Three years of hourly data, starting in 1989, were used for the validation of the models of global and diffuse radiation.

Meteorological data collected in Q. A. were: temperature, humidity, precipitation, wind run, global radiation, reflected solar radiation, and total downward and upward radiation fluxes, photosynthetically active radiation (PAR) and reflected PAR. Pyranometers were of the Moll-Gorczinsky type (from Kipp and Zonen); pyrradiometers were from Philipp Schenk (Mod. 8111), and quantum sensors were from Skye Instruments. Hourly data were collected in the first six month of 1992. Calibration details are in Abreu (1994).

3. TEMPERATURE

Temperature is one of the critical variables that drives all biological systems. Air temperature data available for the validation of agricultural models consist, as a rule, of maximum (X) and minimum (N) air temperature. In the Portuguese climatological and hydrological stations, these elements are measured, in a Stevenson screen, at a height of 1.5 m above the ground. Often, however, grass minimum temperature is also available.

3.1. DAILY STEP

Mean temperature during a relevant part of the day is computed as follows:

$$T = kX + (1 - k)N. \tag{1}$$

To compute mean daily air temperature k is 0.5. We used 126 days of data from Q. A. to illustrate the validity of this parameter value. Table 1 shows the statistics of the regression analysis of predicted versus observed values of mean daily air temperature, using Eq. 1 with k equal to 0.5. At least for the conditions of the experimental site, a minor overestimation of the mean daily air temperature occurred. The SE of the estimates was 0.72 °C.

Some physiological processes occur mainly during the light period (e. g., carbon assimilation), and are temperature modulated. In these cases, the relevant temperature is the mean temperature in the light, T_L. This temperature is estimated by Eq. 1, with the fitted value of k. Using 130 days of hourly data from Q. A, this parameter was determined. k was 0.692 ± 0.003; and statistics of the regression analysis of predicted versus observed values of T_L are in Table 1. The values of k found in the literature range from 2/3 (Masle *et al.*, 1989) to 3/4 (Penning de Vries *et al.*, 1989).

3.2. *DIURNAL VARIATION*

Hourly interpolation of temperature between N and X can be achieved by models of different nature and complexity. Parton & Logan (1981) developed a model that divides the day into two periods. The model uses a truncated sine wave to predict day--time temperature changes and an exponential function to predict night-time temperatures. Wann *et al.* (1985) compared this model and two other sine models, and concluded that Parton & Logan's model yielded better results. Reicosky *et al.* (1989), in a similar comparison, that included this model and four others, found that a sine model presented by De Wit *et al.* (1978) had better performance. In this paper this last model is validated.

TABLE 1

Statistics from linear regression analysis of predicted versus observed values. n is the number of daily observations, $\overline{p}$ and $\overline{o}$ are average predicted and observed values, S_p and S_o are their standard deviations, a and m are the intercept and slope of the line, and SE *is the standard error. S_t = global radiation, S_d = diffuse radiation, L_d = atmospheric long-wave radiation*

Model	n (days)	$\overline{p}$	$\overline{o}$	S_p	S_o	a, m	SE (r^2 in %)
Daily mean temperature, in °C (Eq. 1, with $k = 0.5$)	126	14.71	15.10	3.59	3.80	a=-0.20 m=1.04	0.72 (96)
Day-time mean temperature, in °C (Eq. 1, with $k = 0.692$)	130	17.94	17.28	4.34	4.03	a=-0.26 m=1.05	0.79 (96)
S_t, in MJ $m^{-2}d^{-1}$ (Eq. 5, with a=0.236 and b=0.575)	1081	17.57	17.59	8.73	8.77	a=0.28 m=0.98	1.33 (98)
S_d, in MJ $m^{-2}d^{-1}$ (Eq. 10, with a'=0.965 and b'=-0.834)	1081	6.44	6.36	2.78	2.93	a= 1.03 m=0.85	1.23 (81)
S_d, in MJ $m^{-2}d^{-1}$ (Eq. 11, with A=1.234 and B=0.894)	1081	6.46	6.36	2.73	2.93	a= 1.15 m=0.84	1.21 (81)
L_d, in MJ $m^{-2}d^{-1}$ (Using Eq. 16, with the apparent emissivity given by Eq. 19, where C=0.043 and D=0.807)	128	318.21	316.21	27.95	27.97	a=39.03 m=0.88	13.14 (78)

The model assumes that minimum temperature in day i (N_i) occurs at sunrise (t_r, in hours), and maximum temperature (X_i) at 1400 hours (True Solar Time). Instantaneous air temperature is given by

$$T = (X_{i-1} + N_i)/2 + [(X_{i-1} - N_i)/2]\cos[\pi(t+10)/(t_r+10)]$$

with $0 \leq t < t_r$

$$T = (X_i + N_i)/2 - [(X_i - N_i)/2]\cos[\pi(t-t_r)/(14-t_r)] \qquad (2)$$

with $t_r \leq t < 14$

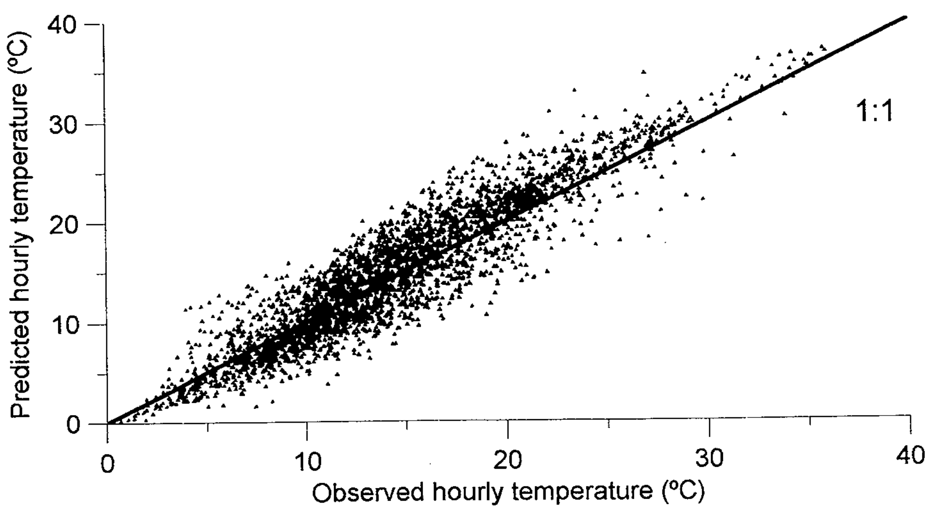
$$T = (X_i + N_{i+1})/2 + [(X_i - N_{i+1})/2]\cos[\pi(t-14)/(t_r+10)]$$

with $14 \leq t < 24$

For the validation of the model in Eq. 2 we used hourly values of temperature from 130 days, measured in Q. A., in 1992. The plot of predicted versus observed values (Fig. 1) shows that the

FIGURE 1

Predicted versus observed hourly air temperature, using the model in Eq. 2



overall performance of the model is good throughout all the range of values that occurred (i. e., from about 0°C to 38 °C). The statistics of the regression analysis of predicted versus observed values of hourly temperature are presented in Table 2.

TABLE 2

Statistics from linear regression analysis of predicted versus observed values. n is the number of hourly observations, $\overline{p}$ *and* $\overline{o}$ *and are average predicted and observed values,* S_p *and* S_o *are their standard deviations, a and m are the intercept and slope of the line, and* SE *is the standard error.* $S_t(t)$ = *instantaneous global radiation,* $S_d(t)$ = *instantaneous diffuse radiation*

Model	n (hours)	$\overline{p}$	$\overline{o}$	S_p	S_o	a, m	SE (r^2 in %)
Air temperature, in °C (Eq. 2)	3001	9.05	8.63	4.43	3.81	a=-0.49 m=1.06	2.58 (86)
$S_t(t)$, in Wm^{-2} (Eq. 9, clear sky)	1040	600.80	600.80	221.38	250.86	a = 78.05 m = 0.87	36.99 (97)
$S_t(t)$, in Wm^{-2} (Eq. 9, all days of 1990)	3431	525.45	525.45	232.35	262.39	a = 66.12 m = 0.84	72.19 (90)
$S_d(t)$, in Wm^{-2} (Eq. 11, all days of 1990 and 1991) A=1.1464 B=0.9360	7386	167.22	167.51	75.36	95.00	a = 55.35 m = 0.67	40.56 (71)
$S_d(t)$, in Wm^{-2} (Eq. 14, all days of 1990 and 1991)	7386	167.78	167.51	89.17	95.00	a = 29.16 m = 0.84	40.00 (80)

4. HUMIDITY

Atmospheric moisture is a determinant of many biological processes and is a variable that should be used by most agricultural models. Transpiration and carbon dioxide assimilation, for example, are intertwined and dependent on atmospheric humidity.

Atmospheric humidity is measured in all climate stations, with a psychrometer. In Portugal, the bulbs of the pair of thermometers are placed at a height of 1.5 m in a standard screen. Often, dirty wicks, poor ventilation, and impure or insufficient water, reduce evaporative cooling of the wet thermometer, which introduces errors in this measurement.

Cloudless days were analysed for the diurnal changes of humidity, described in many ways. As Fig. 2 shows for eight cloudless days (DOY: 114, 118, 120, 121, 123, 124, 131, and 132), it is very important to choose the right measure of humidity. Vapour pressure (e) was almost constant throughout the day and vapour deficit (D) at the occurrence of minimum temperature was very small and reached a peak around 1400 hours, when maximum temperature is reached. Diaz (1989) as a result of similar observations developed a simple model to compute the maximum saturation deficit (D_x):

$$D_x = \frac{s(X-N)}{1-a\ s(X-N)} \quad , \tag{3}$$

where s is the change of saturation vapour pressure with temperature at mean air temperature, and a is a constant with a value of 0 for humid areas and rainy days, and around 0.08 kPa^{-1} for dry and arid areas. In 130 days of 1992 that were analysed only 14 days had maximum relative humidity lower than 90 %. Since the weather in 1992 was drier than average, Eq. 3 may be used successfully in all Portugal, even with a equal to 0.

The average vapour deficit for intire 24 hour days (D_a) is sometimes computed as the difference of the saturation vapour pressure at mean daily temperature, and the saturation vapour pressure at minimum temperature (Campbell & Stokle, 1992):

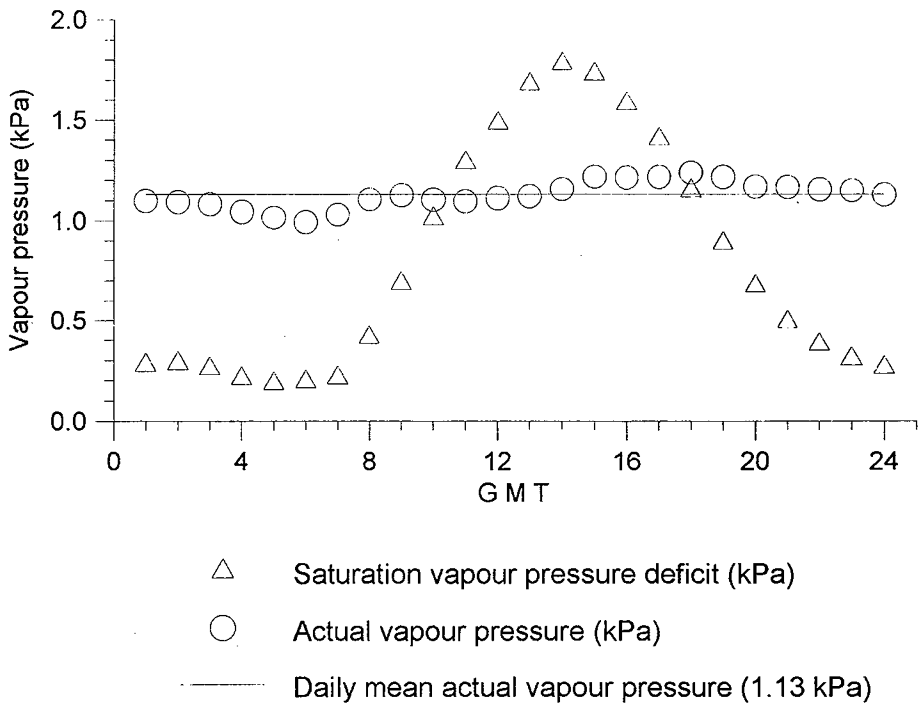
$$D_a = e_s[(X+N)/2] - e_s(N).$$

Since most crop processes occur during daytime, saturation deficit for the simulation of those processes should be averaged for this period. Campbell & Stokle (1992) propose that $D_d = 0.7D_x$.

Approximations of the saturation vapour pressure function, $e_s(T)$, and of the slope of this function on temperature, $s(T)$, are given by Campbell (1977), Brutsaert (1982), Monteith & Unsworth (1990) and Linacre (1992).

FIGURE 2

Diurnal variations of actual vapour pressure and saturation vapour pressure deficit (in kPa), on eight cloudless days



5. SOLAR RADIATION

Solar radiation is short-wave radiation ($0.15\mu m \leq \lambda \leq 3.0\mu m$) that originates on the Sun's photosphere, and is the radiation that reaches the top of the Earth's atmosphere. It is strongly depleted until it reaches the surface of the globe. The radiant flux density on a horizontal surface placed at this level is referred to as 'solar global radiation', S_t ($0.3\mu m \leq \lambda \leq 3.0\mu m$). Part of global radiation comes directly from the Sun as 'beam radiation', S_b; the rest has little, if any, directional nature, it is 'diffuse radiation', S_d. The radiation that reaches a horizontal surface is partly reflected by the surface, according to the reflection coefficient (ρ) of the surface. Thus, the short-wave radiation balance (S_n) is

$$S_n = (1 - \rho)S_t \; . \tag{4}$$

The reflection coefficient is a surface property and has been compiled for many surfaces (e. g., Gates, 1980; Abreu, 1985; Monteith & Unsworth, 1990). Here simple models for the daily step and instantaneous estimation of global and diffuse radiation, and photosynthetically active radiation (PAR) are presented and validated.

5.1. *GLOBAL RADIATION*

5.1.1. DAILY STEP

Global radiation (S_t) is measured routinely in Bragança, Oporto, Penhas Douradas, Coimbra, Castelo Branco, Lisbon, Évora, Faro, Angra do Heroísmo, Ponta Delgada, Porto Santo, and Funchal. Some of these cities have more than one station measuring radiation. Moreover, the majority of the climatological stations in Portugal measure the duration of sunshine hours (INMG, 1990 a, b, 1991 a, b, c, d). However, in the region of Madeira only five out of fourteen climatological stations measured sunshine. The instrument used for these measurements is the Campbell-Stokes sunshine recorder. This instrument was adopted by the World Meteorological Organisation (WMO) in 1962 as a standard of reference (WMO, 1965).

An empirical formula that has long been used to estimate global radiation (Prescott, 1940), with good results, is

$$T_t = S_t/S_0 = a + b\frac{n}{N} \; , \tag{5}$$

where T_t is the daily total transmittance of the atmosphere (i. e., the ratio between global radiation and the extra-terrestrial solar radiation on a horizontal surface, S_0), n/N is the relative sunshine duration (where n is the duration of actual sunshine hours, and N is the duration of possible sunshine hours). S_0 and N are functions of latitude and time of the year, and can be obtained from appropriate tables (e. g., Smithonian Institution, 1966; Garg, 1982) or computed (e. g., France & Thornley, 1984; Spitters *et* *al.*, 1986; Monteith & Unsworth, 1990). The parameters a and b in Eq. 5 were calculated monthly for Lisbon by Peixoto & Marques (1980) (Table 3). Gonçalves (1985) reports those parameters for Bragança (Table 3). In the following analysis, we use the parameter values obtained by linear regression of daily total transmittance (S_t/S_0) on relative sunshine duration (n/N) on three years of daily data for Lisbon (IGIL): $a = 0.236 \pm 3.5 \times 10^{-3}$ and $b = 0.575 \pm 5.2 \times 10^{-3}$, with a $r^2 = 0.92$.

TABLE 3

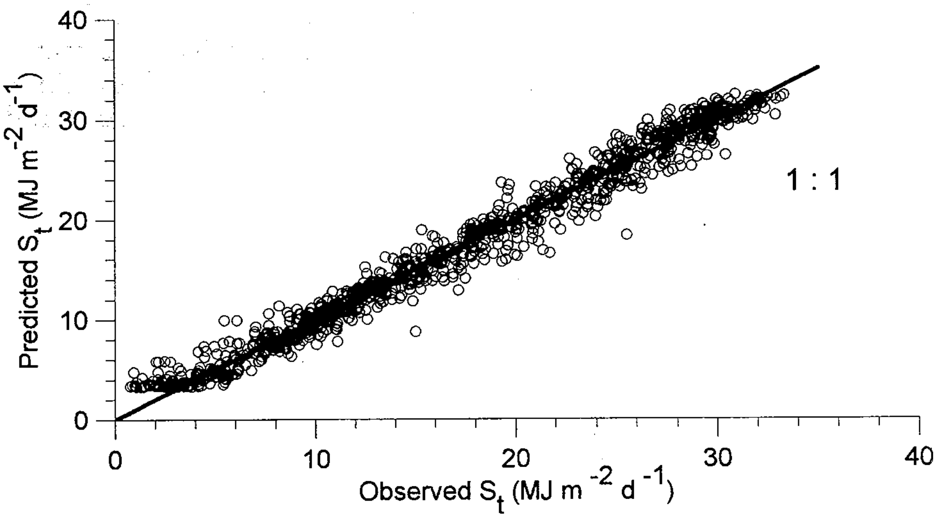
Regression analysis of the daily total transmittance (T_t) and relative sunshine duration (n/N) as reported by Peixoto & Marques (1980), for Lisbon (IGIL), and by Gonçalves (1985), for Bragança. The intercept of the regression line a and the slope b, and related coefficients of determination (r^2) *are shown*

	Peixoto & Marques (1980) (data from 1957 - 1976)			**Gonçalves (1985) (data from 1956 - 1983)**		
Month	a	b	r^2 (%)	a	b	r^2 (%)
Jan	0.227	0.509	89	0.21	0.58	72
Feb	0.224	0.522	91	0.19	0.64	88
Mar	0.230	0.533	90	0.27	0.51	67
Apr	0.226	0.546	90	0.28	0.51	67
May	0.237	0.529	87	0.28	0.53	72
Jun	0.255	0.509	81	0.27	0.53	52
Jul	0.262	0.495	82	0.16	0.66	66
Aug	0.246	0.493	75	0.23	0.56	34
Set	0.245	0.498	76	0.37	0.36	34
Oct	0.230	0.491	90	0.21	0.57	85
Nov	0.223	0.491	90	0.18	0.64	81
Dec	0.222	0.484	91	0.22	0.54	72
Average	0.236	0.508		0.24	0.55	

Fig. 3 shows the scatter diagram of predicted S_t from measurement of n/N versus observed values of S_t. Table 1 presents selected statistics of the regression of predicted on observed S_t values. These statistics confirm the visual evaluation of the model.

FIGURE 3

Predicted versus observed daily global radiation (T_t), using the model in Eq. 5



The linear regression shows a highly significant correlation coefficient (0.98), and a SE of the estimates of 1.34 MJ m^{-2} d^{-1}, which represents only 7.6 % of the data mean. Peixoto & Marques (1980) and Gonçalves (1985) made a more detailed analysis,since they used a longer series of data and made the regressions for decade or monthly sub-sets of data (Table 3). Nevertheless, their parameters are in generally good agreement with ours, but the fraction of the variation accounted for by the model is generally lower in their reports. It is probably safe in any location and environment to use average values reported by Doorenbos & Pruitt (1975), Doorenbos & Kassan (1979), Brutsaert (1982), and Linacre (1992): $a = 0.25$ and $b = 0.50$.

Often, there are no measurements of sunshine duration, but there are observations of the degree of cloudiness. The simple evidence that a portion of the day is either clear or overcast leads to the assumption that

$$\frac{C}{10} + \frac{n}{N} = 1 , \qquad (6)$$

where C is the cloudiness in tenths. This is probably satisfactory for most purposes, when no other information is available. Nevertheless, the actual relation between the degree of cloudiness and relative sunshine duration may be different than in the previous equation. For example, De Vries (1955), in the Netherlands, found

$$0.88\frac{C}{10} + 1.12\frac{n}{N} = 1 \; . \tag{7}$$

5.1.2. DIURNAL VARIATION

To compute instantaneous total solar radiation we set the instantaneous total transmission coefficient ($T_t(t)$) equal to the daily mean value (T_t):

$$T_t(t) = T_t \; ,$$

or

$$S_t(t)/S_0(t) = S_t/S_0 \; . \tag{8}$$

This approximation is likely fairly good for clear and overcast days but bad on partly cloudy days. The justification is that it makes use of all available data and does sum to the correct daily radiation. The instantaneous irradiance on a horizontal surface on the top of the atmosphere, $S_0(t)$, is given by the solar constant, S_{0c}, corrected for azimuth solar angle, Θ, and radius vector of the Earth, r (ratio of sun-earth distance to mean sun-earth distance). The resulting equation is

$$S_t(t) = (S_{0c}\cos\theta/r^2)S_t/S_0 \; . \tag{9}$$

The solar constant is 1370 $\mathrm{Wm^{-2}}$, and cos Θ and r are easily calculated (e. g., Garg, 1982, and France & Thornley, 1984).

Spitters *et al.* (1986) use an empirical relation between $T_t(t)$ and Θ to correct for a decrease in atmospheric transmission on the extremes of the day. On clear days, it may be useful to do this correction, with empirical parameters determined in the region under study. Our inspection of the data from IGIL and Q. A. show that there are important differences between stations in the clear--sky transmissivity at the extremes of the day. For this reason, we validated the simpler approach (Eq. 9) on Lisbon (IGIL) data.

The statistics from the predicted versus observed linear regression analysis performed on clear days and all days of 1990 are shown in Table 2. Clearly, the magnitude of the SE associated with the model shows that it gives reliable values only for clear days. If it is important to know instantaneous radiation, then measurements on an appropriate time scale would be needed.

5.2. *DIFFUSE RADIATION*

5.2.1. DAILY STEP

Diffuse radiation (S_d) is measured regularly in Bragança, Oporto, Coimbra, Castelo Branco, Lisbon, Évora and Faro.

A linear equation relating the ratio of diffuse radiation to global radiation (F_d) and relative sunshine duration (n/N) has been widely used (e. g., Peixoto, 1981):

$$F_d = S_d/S_t = a' + b'\frac{n}{N} \,. \tag{10}$$

Bristow, Campbell & Saxton (1985) proposed a model that estimates S_d with only one measured weather variable: S_t. Their equation is

$$T_d = T_t(1 - \exp(A - A\ B/T_t)) \ \text{ with } \ T_t > 0 \,, \tag{11}$$

where $T_t = S_t/S_o, T_d = S_d/S_o$, and A and B are parameters. These authors assume that

$$A = 0.6/(B - 0.4). \tag{12}$$

Substitution of Eq. 12 in Eq. 11 yields

$$T_d = T_t\,\{1 - \exp\,[0.6(1 - B/T_t)/(B - 0.4)]\,\} \,, \tag{13}$$

which leaves only one coefficient, B, the maximum clear-sky transmissivity. Evaluation of maximum clear-sky transmissivity is possible with limited measurements of transmissivity (i. e., T_t) in clear days.

Working with three years of daily values of global and diffuse radiation, and sunshine duration from Lisbon (IGIL) the coefficients in Eq. 11 and 13 were computed.

The linear regression of F_d on n/N (see Eq. 10) yielded $a' = 0.965 \pm 4.7 \times 10^{-3}$ and $b' = -0.834 \pm 7.0 \times 10^{-3}$, with a $r^2 = 0.93$. The regression line and the relationship F_d versus n/N are shown in Fig. 4 A. The plot of predicted versus observed values of S_d using Eq. 10 are shown in Fig. 4 B. Statistics of the linear regression of predicted versus observed values of S_d are shown in Table 1. Table 4 shows the monthly values of the coefficients a' and b' in Eq. 10 calculated by Peixoto & Marques (1980) for Lisbon (IGIL).

TABLE 4

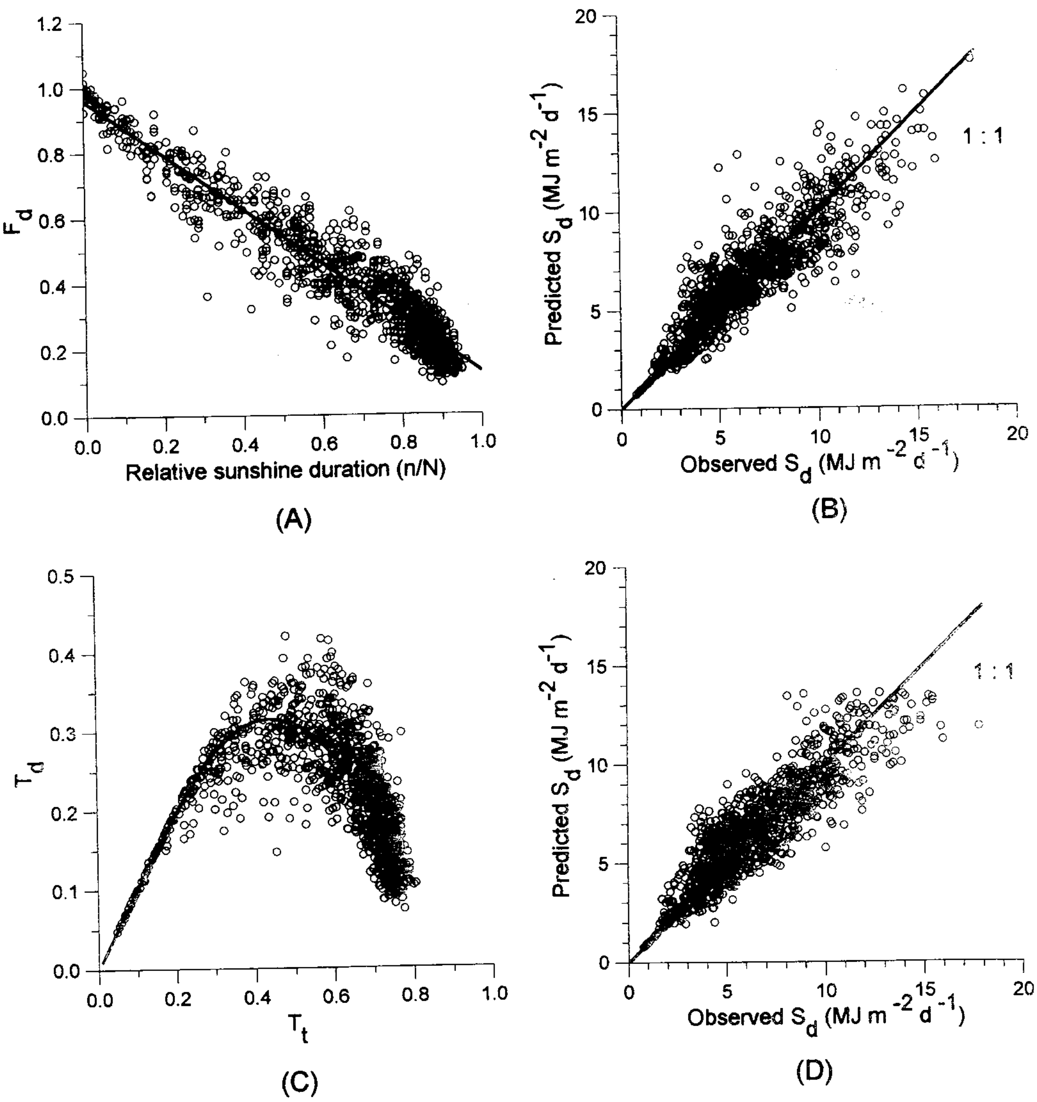
Regression analysis of the daily diffuse radiation transmittance (T_d) and relative sunshine duration (n/N) as reported by Peixoto & Marques (1980) for Lisbon (IGIL). The intercept of the regression line a' and the slope b', and related coefficients of determination (r^2) are shown. Data are from 1 January 1957 until 31 December 1976

Month	a'	b'	r^2 (%)	Month	a	b	r^2 (%)
Jan	0.95	-0.80	88	Jul	0.90	-0.82	81
Feb	0.95	-0.81	92	Aug	0.97	-0.88	83
Mar	0.95	-0.83	86	Sep	0.93	-0.84	88
Apr	0.93	-0.81	88	Oct	0.92	-0.80	88
May	0.91	-0.81	86	Nov	0.95	-0.83	90
Jun	0.92	-0.83	88	Dec	0.95	-0.81	94

To determine the coefficients in Eq. 11 a non-linear regression algorithm was used (Marquardt, 1963). Fitted coefficients are: $A = 1.234 \pm 30.7 \times 10^{-3}$ and $B = 0.894 \pm 5.4 \times 10^{3}$. The relationship between T_d and T_t is shown in Fig. 4 C, where the fitted curve is also shown. The plot of predicted versus observed values is shown in Fig. 4 D. Statistics of the linear regression of predicted on observed values are presented in Table 1. Bristow *et al.* (1985) found, by inspection of the actual data on clear-sky transmissivity, that the value of B in Eq. 13 was 0.76 for Pullman (winter rainfall climate in Washington State, USA) and 0.74 for Townsville (summer rainfall climate in Queensland, Australia), and used these values in their simulations.

FIGURE 4

(A) Ratio of diffuse radiation to global radiation (F_d) as a function of relative sunshine duration (n/N). The regression line is also shown. (B) Predicted versus observed daily diffuse radiation (Sd), using the model in Eq. 10. (C) Relationship between daily diffuse (T_d) and total transmittance of the atmosphere (T_t). The fitted curve is also shown. (D) Predicted versus observed daily diffuse radiation (S_d), using the model in Eq. 11



Both models performed well in predicting diffuse radiation. Nevertheless, the model of Bristow *et al.* (1985) was marginally better, as can be seen by the lower SE (see Table 1). The SE of the estimates of diffuse radiation are, in relative terms, about three times greater than those found for global radiation (see Table 1). Fortunately, the sensitivity of growth models to errors in separating diffuse from direct radiation is not high, because the errors in the estimate of one quantity are associated with compensating errors in the other quantity.

5.2.2. DIURNAL VARIATION

When hourly values of global radiation are available, a model similar to the model used by Bristow *et al.* (1985) for daily total of global radiation, and fitted parameters, seems to be ideal (Eq. 11). An alternative model, that doesn't need hourly values of global radiation has long been used (Collares-Pereira & Rabl, 1979):

$$S_d(t) = S_0(t)\frac{S_d}{S_0} \qquad \text{with} \quad S_d(t) \leq S_t \ , \tag{14}$$

where $S_d(t)$ and $S_0(t)$ are the instantaneous diffuse and global radiation, respectively. Two years (1990 and 1991) of hourly values of global radiation from Lisbon (IGIL) were used to validate these models. The fitted parameters in Eq. 11 are: $A = 1.1464$ and $B = 0.9360$. The statistics of the regression analysis of predicted versus observed values for both models are shown in Table 2. Although the SE are similar, Eq. 14 leads to better estimates over the full range of values used, because the slope of the line is closer to 1 and the y-intercept is closer to 0. This indicates that the instantaneous diffuse transmittance of the atmosphere $(S_d(t)/S_0(t))$ is quite insensitive to cloud conditions and solar height (Collares-Pereira & Rabl, 1979).

5.3. *PHOTOSYNTHETICALLY ACTIVE RADIATION*

Short wave radiation in the visible waveband ($0.4\mu m \leq \lambda \leq 0.7\mu m$), because of its role in the photochemical conversion of the energy of light in the cloroplasts of the plants, is often called the 'photosynthetically active radiation', PAR. This quantity can be measured as energy flux density (W m^{-2}) or a quantum flux density (mol m^{-2} s^{-1}).

The fraction of PAR to total energy in the extraterrestrial solar spectrum is 0.45, but at the surface of the globe it increases to 0.50 (Monteith & Unsworth, 1990). The relation between quantum content and energy is a conservative quantity. Monteith & Unsworth (1990) report values of 2.1, 2.3 and 2.9 μmol J^{-1} (total radiation), in California, the English Midlands, and Texas, respectively. Using data from 141 days, we obtained 2.105±0.034 μmol J^{-1} (total radiation), which is the same relation that was found in California (above).

Rayleigh scattering is wavelength selective and predominant in clear skies. Hence, for a clear sky, the scattered diffuse component in PAR is relatively larger than in total radiation. The experiments of Burtin *et al.* (1981) show that, under clear skies, the fraction diffuse in the PAR band is 1.4 times the fraction diffuse of total global radiation. Nevertheless, the correction to account for this higher diffuse fraction in the PAR waveband is largely outweighed by another correction that could be introduced to account for circumsolar radiation under clear skies (Spitters *et al.*, 1986).

Although, near sunrise and sunset, the ratio of quantum content and energy is slightly lower than the daily average it is almost constant. We used linear regression analysis on 1168 values of hourly data from Q. A. to determine this ratio: 2.110±0.002 μmol J^{-1} (total radiation). The SE was 34.84 μmol m^{-2} s^{-1}, with $r^2 = 1.00$.

6. TERRESTRIAL RADIATION

Terrestrial radiation is long-wave radiation emitted by the natural surfaces of the earth. The energy flux density emitted by a full radiator (B, in W m^{-2}), or black body, is given by Stefan--Boltzmann's Law

$$B = \sigma T^4 , \tag{15}$$

where σ is the Stefan-Boltzmann constant (5.67×10^{-8} W m^{-2}K^{-4}), and T is the temperature in kelvins (K).

6.1. *UPWARD LONG-WAVE RADIATION*

The radiant emittance of natural surfaces (φ) can be estimated by

$$\varphi = \varepsilon B = \varepsilon \sigma T^4 , \tag{16}$$

where ε is the emissivity, considered independent of wavelength. Gates (1980), Abreu (1985), and Monteith & Unsworth (1990) provide long-wave emissivities of many natural surfaces.

The upward long-wave flux density (L_u), received on a horizontal surface facing down, is given by Eq. 16 plus a minute contribution from the part of atmospheric radiation (L_d) that is reflected on the ground. The corresponding equation is

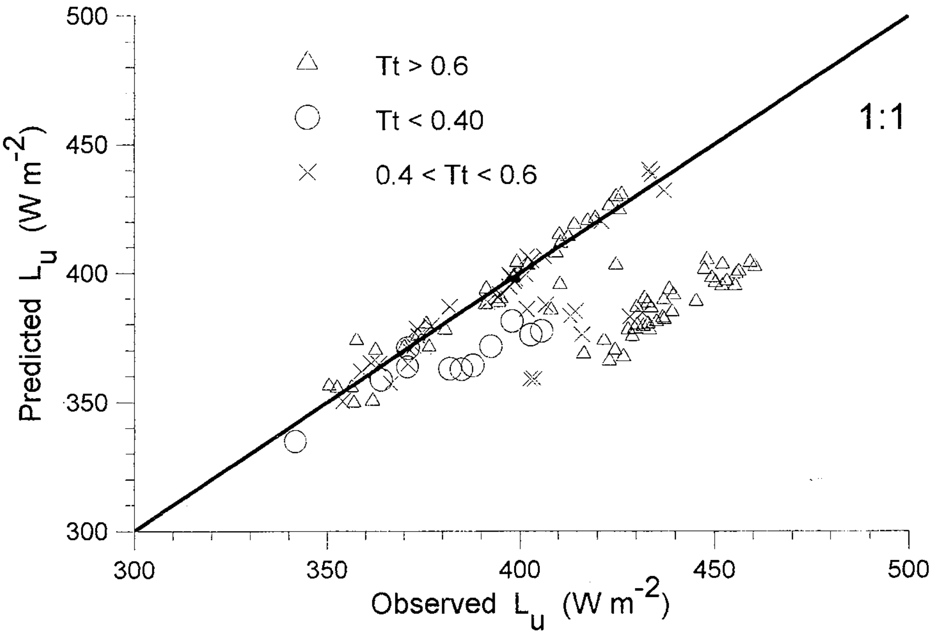
$$L_u = \varepsilon \sigma T^4 + (1 - \varepsilon) L_d . \tag{17}$$

Notice that, according to Kirchhoff principle, emissivity (ε) equals absorptivity.

Thus, L_u can be calculated accurately by Eq. 17, if 1) the appropriate emissivity is used; and 2) the superficial temperature of the ground is known. Although the first condition is easily satisfied for most surfaces, superficial temperature is rarely simulated by growth models. For a cropped soil with a high water status, it is close enough for must proposes to use mean air temperature to calculate mean daily values of L_u. Nevertheless, this practice in rainfed crops in warm environments leads to overestimation of L_u in clear days. Fig. 5 shows that Eq. 17, with $\varepsilon = 0.97$, usually underestimates L_u measured over a wheat crop, under Mediterranean conditions. Clear-sky measurements are usually responsible for this underestimation, since in overcast days predicted values tend to be similar to the measured.

FIGURE 5

Predicted versus observed daily values of upward long-wave radiation (L_u), using the model in Eq. 17, with $\varepsilon = 0.97$



6.2. *ATMOSPHERIC LONG-WAVE RADIATION*

6.2.1. DAILY STEP

The main emitters (and absorbers) of long-wave radiation in a clear atmosphere are water vapour, carbon dioxide and ozone. Whenever present, water droplets (or ice crystals) in the clouds are the only effective emitters (and absorbers) in wave-lengths between $8 - 13\mu m$, the so called 'atmospheric window'. All emitters present in the atmosphere are 'grey bodies' emitting proportionally to temperature as described by Eq. 16. The downward long-wave flux density from the atmosphere (L_d), received on a horizontal surface, can therefore be derived from the profiles of water and temperature in the atmosphere, and cloudiness. Unfortunately, where estimates of long wave are needed only simple climate elements are likely available.

Under clear sky conditions, the simplest formulae estimate L_d using screen temperature as a proxy. Since more than half of L_d comes from the lower 100 m of the atmosphere (Geiger, 1961), a strong relation exists between screen temperature and the temperature of the centre of emission. Monteith & Unsworth (1975) calculate the apparent emissivity of the atmosphere on clear days (ε_a) and use it in an equation similar to Eq. 16. Their empirical relation is

$$\varepsilon_a = 1.06 - \frac{119}{\sigma T_a^4} \, , \tag{18}$$

where T_a is standard screen temperature, in kelvins.

Campbell & Stokle (1992), fitted the following equation to L_d data under cloudy skies:

$$\varepsilon_{ac} = \varepsilon_a + \frac{1 - \varepsilon_a}{1 + C\exp(D\ T_t)} \, , \tag{19}$$

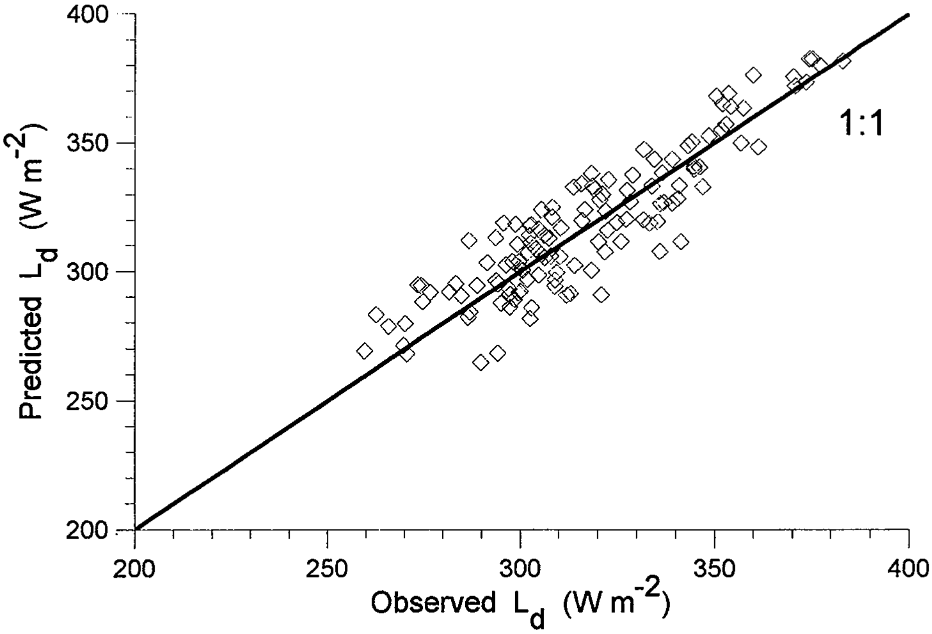
where the parameters C and D where equal to 0.048 and 7.1, respectively.

Using Eq. 18 in the Stefan-Boltzmann's Law, clear-sky atmospheric radiation was estimated closely (Table 1). SE was 8.86 W m^{-2}, which is 3 % of the average of the measured quantity. Notice that the coefficients in Eq. 18 were determined for the very different conditions of the English Midlands, which is a positive indication about the generality of this equation.

The atmospheric radiation on cloudy days, using Eq. 19 for apparent emissivity, was estimated using fitted parameters determined on 128 days of data: $C = 0.043$ and $D = 8.07$. The plot of predicted versus observed values is shown in Fig. 6. The SE from the regression line is 13.1 W m^{-2} , and 78 % of the variation in L_d (i. e., $r^2 = 78$ %) has been accounted for by the empirical model (Table 1). Notice that when the original parameters were used r^2 was 76 %.

FIGURE 6

Predicted versus observed daily values of atmospheric long-wave radiation (L_d), using Eq. 16, with the apparent emissivity of the atmosphere given by Eq. 19



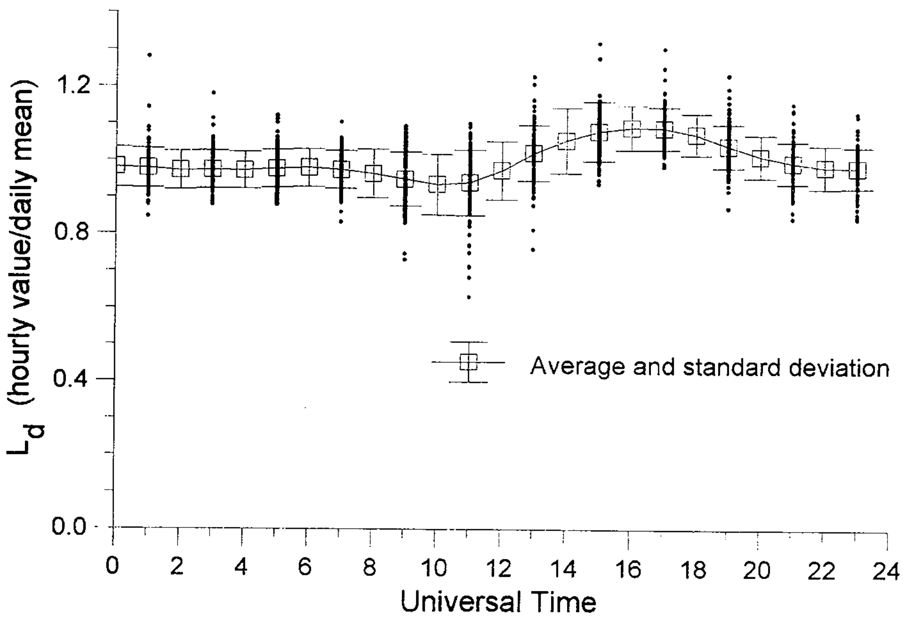
6.2.2. DIURNAL VARIATION

Under clear skies, diurnal changes of atmospheric radiation occur due to an increase of temperature and, on a smaller scale, precipitable water, in the lower portion of the atmosphere. Clouds distort this underlying pattern in a random fashion. Six month of hourly data from Q. A. (all days from January to June of 1992) were used to illustrate those diurnal changes (Fig. 7). Hourly values were normalized by mean L_d of that day, to allow for the simultaneous representation of all data points. L_d is below the mean until about mid-day, showing a small increase in the afternoon, until it reaches a peak value around 16 hours (Universal Time). Later, there is a small decline; near mid-night mean L_d is reached again. Idso (1974) reports diurnal changes on clear days, in a very hot and arid environment, that roughly resembles ours. Paltridge (1970) reports, under overcast skies, smaller diurnal changes of L_d that are unrelated to screen temperature.

It is well known (Paltridge, 1970; Idso, 1974) that seasonal variations are important in simulating the diurnal time-course of L_d, what further complicates the problem.

FIGURE 7

Normalised diurnal changes of atmospheric long-wave radiation (L_d), using six months of hourly data from Q. A. (all days from January to June of 1992). Only odd-hour values are plotted for the sake of clarity. Standard error bars are also shown



6.3. *NET LONG-WAVE RADIATION*

Estimation of net long-wave radiation (L_n) of a surface is possible when its temperature is known, which is seldom the case. The net long-wave is also given by the difference $L_d - L_u$, calculated as indicated above. Ordinarily, however, L_n is replaced by its isothermal counterpart (L_{ni}) where the surface temperature is replaced by screen temperature:

$$L_{ni} = (\varepsilon_{ac} - \varepsilon_s)\sigma T_a^4 \quad , \tag{20}$$

where ε_{ac} is given by Eq. 19, and ε_s is the surface emissivity.

ACKNOWLEDGMENTS

We thank Junta Nacional de Investigação Científica e Tecnológica (JNICT) for financial support, and Professor J. Pinto Peixoto for supplying the data from Instituto Geofísico Infante D. Luís.

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