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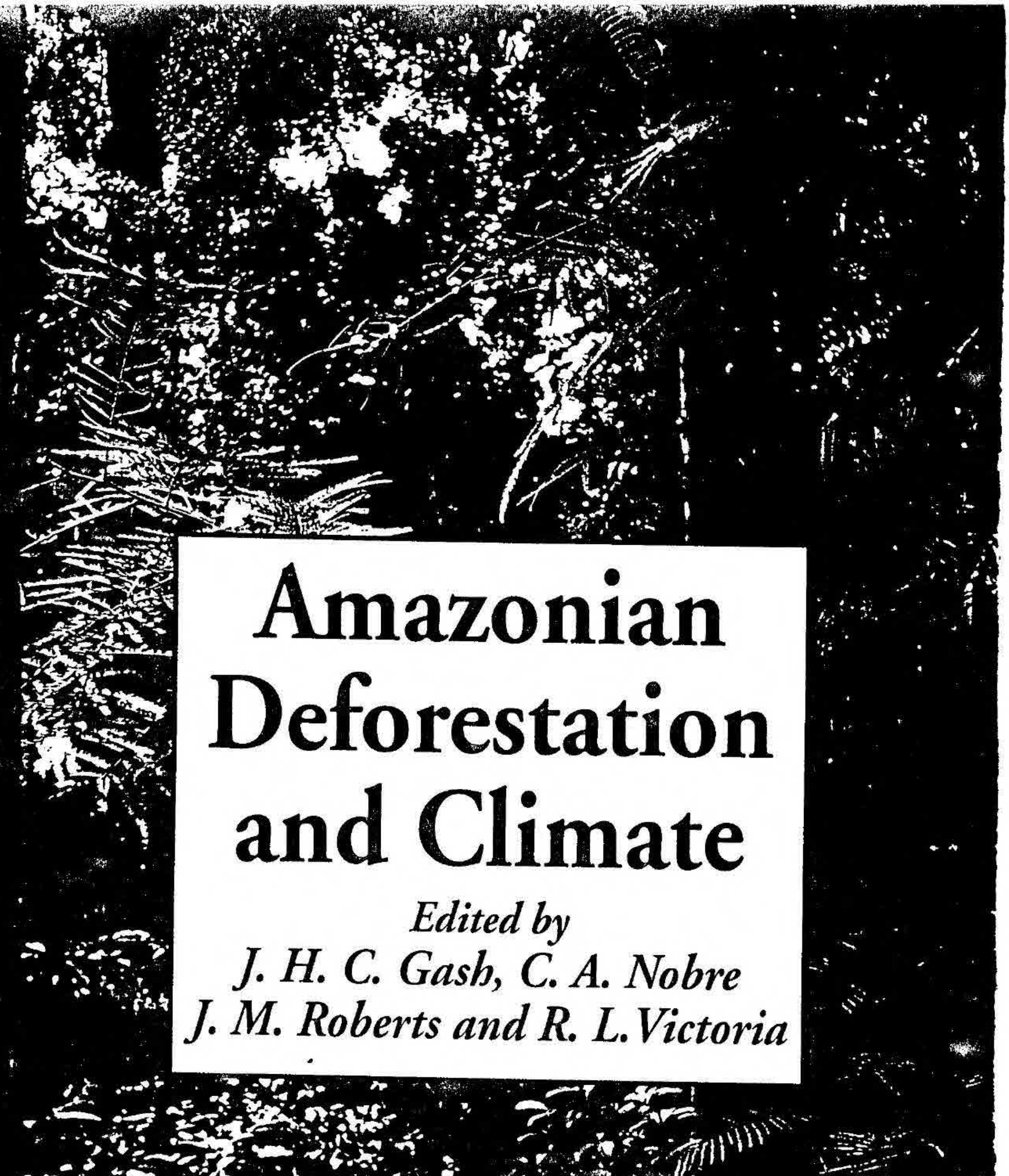

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7 Thermal diffusivity of Amazonian soils

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INTRODUCTION

Soil temperature influences most of the processes of the soil-plant continuum. For example, it affects plant growth rate and the rate of decomposition of organic matter. It is required for the characterisation of the microclimate at the soil surface and is a component of all atmospheric models. Temperature profiles in the soil may be obtained through the integration of the heat conduction equation, which depends on the volumetric heat capacity of the soil, C, and its apparent thermal diffusivity, K, where

$$C = \rho C_b \qquad \mathrm{J\ m^{-3}\ K^{-1}} \tag{1}$$

$$K = k/C \qquad \mathrm{m^2\ s^{-1}} \tag{2}$$

and where k is the apparent heat conductivity in W m^{-1} K^{-1}, ρ is the bulk density of the soil in kg m^{-3} and C_b is the bulk specific heat in J kg^{-1} K^{-1}. However, these parameters vary with the composition, texture and moisture content of the soil. For a soil with a negligible organic fraction, the volumetric heat capacity, as shown by Campbell (1985) and following de Vries (1963), may be obtained from

$$C = C_m \rho/\rho_m + \theta C_w \tag{3}$$

where C_m and C_w are the specific heats of the mineral constituents of the soil and of water respectively, ρ_m is the density of the mineral fraction and θ (kg kg^{-1}) is the volumetric moisture content of the soil.

The apparent thermal diffusivity, which is considered to be constant for a uniform soil, can be determined by several methods using measurements of transient soil temperature. These methods are based on either analytical or numerical solutions of the equation of heat conduction. Horton *et al.* (1983) evaluated six of them: amplitude, phase, arctangent, logarithmic, numerical and harmonic. However, as actual soils are vertically non-uniform, with varying composition, texture and moisture content, more elaborate methods may be used (Lettau, 1954; van Wijk and

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Derksen, 1963). Alternatively, the non-uniform soil may be divided into layers, each considered to be of uniform composition and structure, in which K is assumed constant.

The latter approach is used here to obtain the apparent thermal diffusivity for two deforested areas in Amazonia, each covered with pasture but with contrasting soils: clay and sand.

METHODOLOGY

The one dimensional conduction of heat through the soil is given by

$$C\left(\frac{\partial T}{\partial t}\right) = \left(\frac{\partial}{\partial z}\right)\left(CK\left(\frac{\partial T}{\partial z}\right)\right) \tag{4}$$

where T is soil temperature, t is time and z the depth (see for example Gilman, 1980). Partitioning a non-uniform soil into layers of uniform composition and texture, Equation 4 becomes

$$\frac{\partial T}{\partial t} = K_j\left(\frac{\partial^2 T}{\partial z^2}\right) \tag{5}$$

where $K_j = (k/C)_j$, the apparent thermal diffusivity of layer j, delineated by z_{j-1} and z_{j+1}, j = 1, 2, ... m, the number of layers, not necessarily of equal thickness.

The apparent thermal diffusivity can be obtained by the use of measured temperatures substituted into expressions obtained from analytical solutions of Equation 5. This is the case for the amplitude, phase, arctangent, logarithmic and harmonic methods shown by Horton *et al.* (1983) and referred to here as the 'periodic' methods. Numerical solutions of Equation 5 can also be obtained with a succession of assumed values of K and comparison made with the measured temperature profiles to find the value which minimises the differences between measured and computed profiles. Finally, K can be obtained by making it explicit in the numerical expressions for Equation 4. Five of these methods are compared in this paper, (omitting the harmonic and explicit methods).

Numerical method

Equation 5, put in the form of finite differences by means of a second order weighted scheme (Ralston and Wilf, 1964; Campbell, 1985), gives the solution for layer j, delineated by z_{j-1} and z_{j+1}, and at times t_k and t_{k+1}:

$$\frac{\left(C_{j+\frac{1}{2}}K_{j+\frac{1}{2}}\left[\frac{T_{j+1}-T_j}{z_{j+1}-z_j}\right] - C_{j-\frac{1}{2}}K_{j-\frac{1}{2}}\left[\frac{T_j-T_{j-1}}{z_j-z_{j-1}}\right]\right)}{\left[\frac{z_{j+1}-z_{j-1}}{2}\right]} = C_j\left(\frac{T_j^{k+1}-T_j^k}{\Delta t}\right) \tag{6}$$

where

$$T = pT^{k+1} + (1-p)T^{k} \tag{7}$$

and

$$C_{j+\frac{1}{2}} = \frac{1}{2}(C_j + C_{j+1}), etc \tag{8}$$

and p is a weighting factor which may vary from 0 to 1. A factor of p = 0.5 corresponds to the classic scheme of Crank-Nicholson. According to Campbell (1985), p equal to 0.6 minimises the errors in the energy balance and this value is therefore used in this study. The stability, convergence and consistency of the numerical integration (Horton *et al.*, 1983) is assured if

$$\Delta T \leq \frac{(\Delta z)^2}{2K} \tag{9}$$

To determine K corresponding to the layer j, a sequence of solutions of Equation 6 was derived with a measured initial temperature profile and with measured temperature boundary conditions. Successive constant trial values of K are used and the sum of the squared differences between the computed temperatures and their corresponding measured values is determined. The value of K which minimises this difference is considered to be the one corresponding to the layer considered, under the conditions imposed.

Amplitude

Following the procedures given by Horton *et al.* (1983) the temperatures measured at the two depths, z_1 and z_2, delimit a layer assumed to be uniform, so that

$$K = \frac{\pi}{P}\left[\frac{z_2 - z_1}{\ln(A_1/A_2)}\right]^2 \tag{10}$$

where A_1 and A_2 are the amplitudes at depths z_1 and z_2, respectively, of the corresponding transient temperature records, and P is the cyclic time period, in this case the 24 hours from 0 to 24 h local time. It should be noted that this and subsequent periodic methods result from an idealised analytical solution of the soil heat conduction equation.

Phase

$$K = \frac{P(z_2 - z_1)^2}{4\pi L^2} \tag{11}$$

where L is the time interval between measured maxima, or minima, of the soil temperature at z_1 and z_2, during the period P.

Arctangent

$$K = \frac{\pi}{P}(z_2-z_1)^2 \left\{ \arctan \left[\frac{(T_1-T_3)(T'_2-T'_4)-(T_2-T_4)(T'_1-T'_3)}{(T_1-T_3)(T'_1-T'_3)+(T_2-T_4)(T'_2-T'_4)} \right] \right\}^{-2} \tag{12}$$

where temperatures T_i and T'_i, are at depths z_1 and z_2 respectively, and are taken from four successive times, i = 1 to 4, each six hours apart during the 24 hour period P.

Logarithmic

$$K = \left[\frac{0.0121\,(z_2-z_1)}{\ln \left[\frac{(T_1-T_3)^2+(T_2-T_4)^2}{(T'_1-T'_3)^2+(T'_2-T'_4)^2} \right]} \right]^2 \tag{13}$$

where the notation is the same as for Equation 12.

To obtain reliable daily values of K with these four simple idealised periodic methods, it is essential that the daily temperature records are regular and have only one maximum and minimum during the 24-hour period considered, and that their derivatives vary slowly. In addition there should be no rain, and consequently the soil heat flux should have a slow and regular variation, especially during the daylight hours. This occurs particularly on clear days or at least when there are no strong gradients of solar radiation caused by rapidly moving convective clouds. Although some of the short period 'noise' can be accommodated by Fourier analysis and filtering (Gilman, 1977), the occurrence of rain eliminates the possibility of a constant K value throughout the day because of the change in heat capacity and conductivity with changing soil moisture (Wierenga *et al.*, 1969; Campbell, 1977).

The numerical integration used to obtain values for K with the numerical method, and also to calculate temperature profiles in any situation, are very robust. The predicted temperature profiles within a layer are always smooth, whatever the temperatures at its boundaries. However, as the objective is to obtain constant values of K for a layer during the whole 24-hour period, and the dependence on soil moisture, data from rainy days are also unacceptable for this method.

SITES, DESCRIPTION AND DATA

Data were recorded in the ABRACOS project at Fazenda Dimona, near Manaus, Amazonas and Fazenda Nossa Senhora, near Ji-Paraná, Rondônia (Gash *et al.*, 1996). Soil structure at each site (in terms of percentage clay: silt sand) is 75%: 10%: 15% at Manaus and 15%: 10%: 75% at Ji-Paraná (Wright *et al.*, 1996; Hodnett *et al.*, 1995). The data used were soil temperatures, energy fluxes and moisture profiles, measured during two dry season data missions: Mission 1 at Fazenda Dimona, 25 September to 25 October 1990 and Mission 4/5 at Fazenda Nossa Senhora, 1 April to 6 August 1993. Soil temperatures were monitored using thermistors (Campbell

Scientific, Shepshed, UK) and recorded as ten-minute averages, together with soil heat flux, using a programmable logger (Campbell Scientific, Shepshed, UK); see Wright *et al.* (1992). Soil temperatures at 10 cm are the mean of three sensors; 20 and 40 cm temperatures were from a single sensor at each depth. Soil moisture content was measured at approximately seven day intervals using a Neutron Probe (Didcot Instruments, Abingdon, UK) with each datum being the mean soil moisture estimated from eight permanently installed access tubes (Hodnett *et al.*, 1995). Radiation balance data are from the automatic weather station at each site (Culf *et al.*, 1996).

RESULTS AND DISCUSSION

Examples of typical data from wet and dry periods during 1993 at Fazenda Nossa Senhora are shown in Figures 1 and 2 which illustrate the influence of radiation

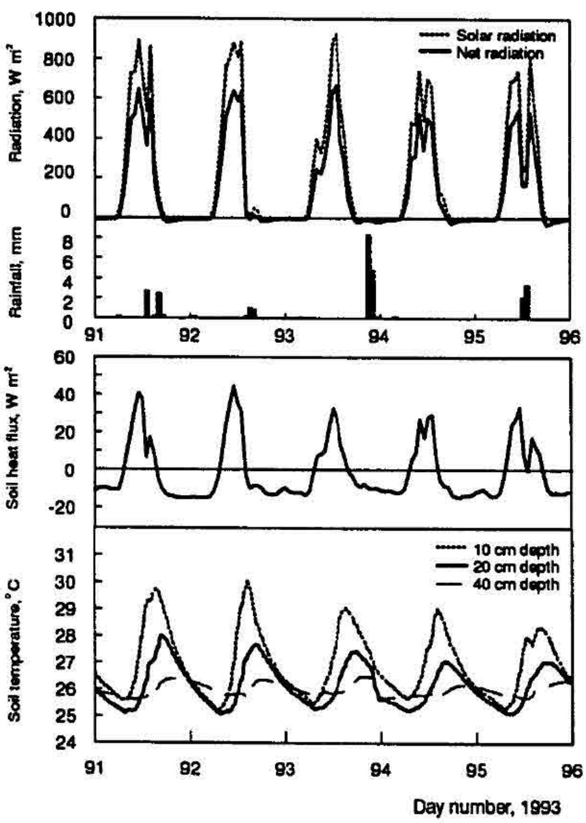

Figure 1 Wet season energy fluxes, rainfall and soil temperature measured at Fazenda Nossa Senhora (Days 91-95, 1993)

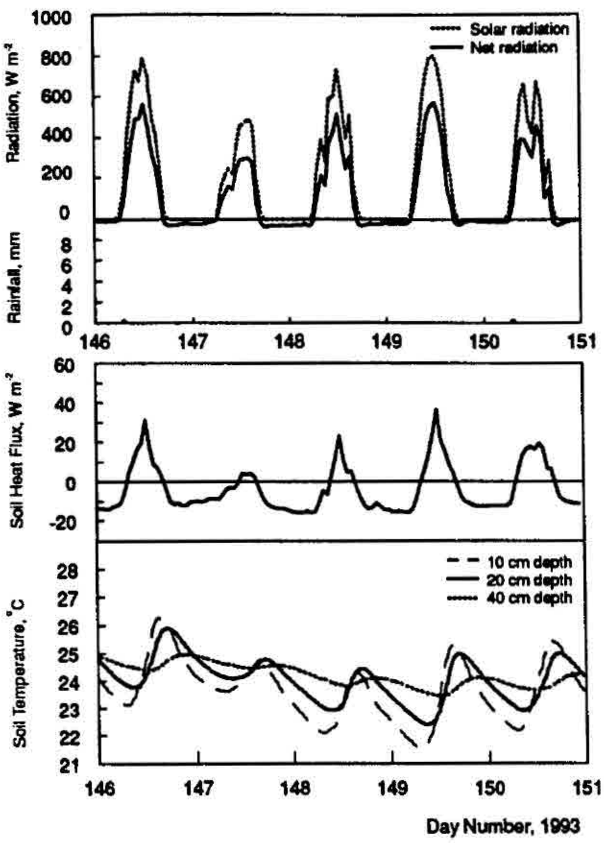

Figure 2 Dry season energy fluxes, rainfall and soil temperature measured at Fazenda Nossa Senhora (Days 146-150, 1993)

and rainfall on the soil heat flux and soil temperatures. The cloudy day (Day 150, 1993) has some insignificant rainfall (0.13 mm) at 08.00h and moderate fluctuations in cloudiness, yet presents a relatively smooth soil heat flux, and is typical of those selected to compute the apparent soil thermal diffusivity. The interaction between rainfall and soil moisture content at 10, 20 and 40 cm depths, for both sites, is shown in Figure 3. The soil water drainage after rainfall, in these periods with unsaturated soils, is rapid for the sandy soil of Fazenda Nossa Senhora, while the decrease of moisture is moderately slow during the rain-free periods. The high values of soil moisture for 10 and 20 cm depth on Day 115, 1993 are a consequence of moisture measurement very soon after the 15 mm storm. At Fazenda Dimona there was very little rain during the period shown, and although the moisture content is high, the decrease with time is slow.

Equations 6 and 10-13 were applied to estimate thermal diffusivity on nine selected days from Fazenda Nossa Senhora for which the above conditions were satisfied and

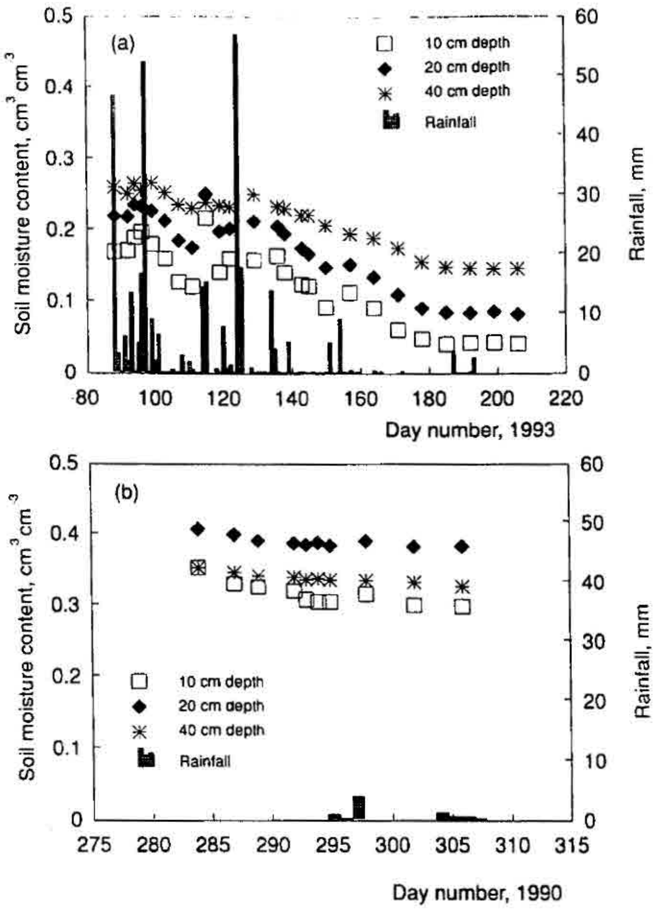

Figure 3 Soil moisture content and rainfall at (a) Fazenda Nossa Senhora and (b) Fazenda Dimona

measurements of soil moisture were available (see Table 1). Within the Fazenda Dimona data only two such days occurred and the soil moisture profiles were interpolated for the relevant period. This was possible because of the small changes in soil moisture between measurements and the absence of rain. Volumetric heat capacity was determined using Equation 3 with C_m = 2.4 MJ m^{-3} K^{-1} and ρ_m = 2.65 Mg m^{-3} for both clay and quartz, and C_w = 4.18 MJ m^{-3} K^{-1}. Soil bulk density ρ was taken as 1.1 for the clay and 1.3 Mg m^{-3} for the sandy soils (Wright *et al.* 1996).

The apparent thermal diffusivity computed by the amplitude method consistently showed the least scatter when compared with the other three periodic methods, and only the results from the amplitude method are presented here for comparison with the numerical method. The 'arctangent' and 'logarithmic' methods performed particularly badly. The scatter in all of the periodic methods is due to their strong sensitivity to small errors in the data. Diffusivities from the 'amplitude' method for

Table 1 Apparent thermal diffusivities and measured data for selected days

Day	Rain	Measured soil moisture Volume fraction			Storage	Thermal diffusivity ($\times 10^6$) Amplitude		Numerical	SDELQ*	rms**
	mm	$m^3 m^{-3}$			mm	$m^2 s^{-1}$			°C	°C
		10 cm	*20 cm*	*40 cm*	*0-40*	*10-20*	*20-40*	*10-40 (cm)*		
Fazenda Nossa Senhora, 1993										
11	0.51	0.120	0.174	0.230	64.4	3.092	0.862	1.410	22.160	0.395
129	0.0	0.157	0.212	0.250	77.6	2.049	1.205	1.330	1.094	0.089
145	0.0	0.121	0.166	0.22	162.9	2.212	0.862	1.530	4.165	0.170
150	0.13	0.091	0.147	0.206	53.5	1.880	0.927	1.450	2.198	0.122
171	0.0	0.059	0.109	0.174	40.1	2.084	0.983	1.670	6.649	0.217
178	0.0	0.046	0.090	0.155	33.7	1.700	0.898	1.410	1.304	0.095
192	0.13	0.041	0.083	0.146	31.1	1.722	0.958	1.410	2.539	0.134
199	0.0	0.041	0.085	0.145	31.2	2.074	1.393	1.250	3.873	0.164
206	0.0	0.039	0.081	0.145	30.4	1.828	1.062	1.373	5.129	0.190
						2.07 (0.42)	1.02 (0.18)	1.43 (0.12)		
Fazenda Dimona, 1990										
282	0.0	0.350	0.408	0.357	146.4	0.169	0.453	0.063	10.960	0.277
288	0.0	0.320	0.393	0.342	139.0	0.146	0.453	0.067	13.586	0.315
						0.158	0.453	0.065		

* $\Sigma(T_{measured} - T_{numerical})^2$ for depths equal to 20 cm (142 values) ** rms = $(SDELQ/142)^{0.5}$

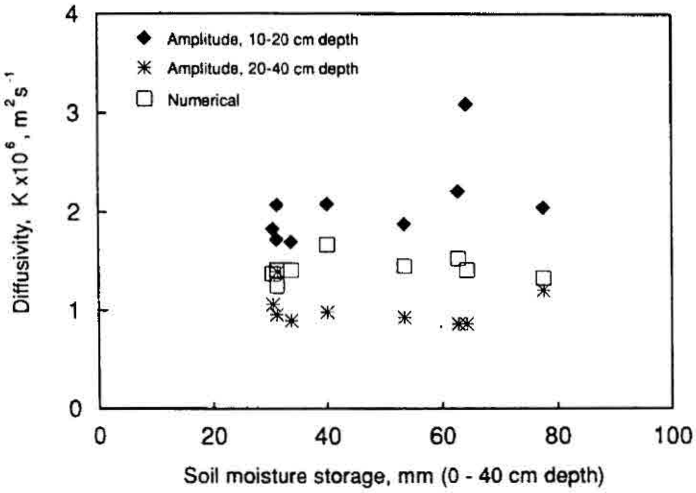

Figure 4 Calculated apparent soil thermal diffusivity for Fazenda Nossa Senhora.

Table 2 Soil water storage (0-40 cm) versus soil moisture at specific levels — linear regression parameters

Depth (cm)	Intercept	Slope	r^2	Soil moisture range
Fazenda Nossa Senhora				
15*	8.55	371.86	0.998	0.060 - 0.233
30**	-18.58	424.50	0.988	0.113 - 0.250
Fazenda Dimona				
15*	0.46	383.95	0.998	0.340 - 0.424
30**	-0.77	586.69	0.986	0.354 - 0.409

* Average between 10 and 20 cm measured values of θ
** Average between 20 and 40 cm measured values of θ

the 10 to 20 cm and 20 to 40 cm soil layers are shown in Table 1 and Figure 4. The numerical method — in principle the most appropriate if the selection conditions are satisfied — provided the most consistent results for the 10 to 40 cm soil layer (also given in Table 1 and Figure 4).

The overall water storage in the top 40 cm of soil was found to correlate linearly with the values of the soil moisture at individual depths, with correlation coefficients (r^2) between 0.986 and 0.998. Notwithstanding the underestimation of soil moisture near the surface using the neutron probe, this is a useful result for interpolating soil moisture during rain-free periods to correspond to specific K values. The parameters of these regressions are shown in Table 2.

To satisfy the stability criterion given by Equation 9, the maximum integration time interval Δt was 600 s at Fazenda Dimona and between 150 s and 200 s at Fazenda Nossa Senhora. The measurements were recorded at 600 s intervals and when the time integration increments were shorter than 600 s, the soil temperatures at the boundaries were interpolated using cubic splines between the measured values. The same method was used to interpolate vertically for the initial temperature profile. Eight depths, unequally spaced, were used for the integrations between 10 and 40 cm: 12.5, 15.0, 17.5, 20.0, 22.5, 25.0, 30.0, and 35.0 cm. At the Fazenda Nossa Senhora site (Table 1) the sums of the squared differences between 142 measured and computed soil temperatures at the 20 cm depth for each of the nine selected days vary between 1.09 and 22.16, with seven of them under 5.2. Individual root mean square (rms) differences range from 0.089 to 0.395°C. Figure 5 shows the measured and calculated soil temperatures for the selected day (Day 150, 1993).

For the sandy soil of Fazenda Nossa Senhora (Figure 4) the apparent thermal diffusivities obtained using the numerical method vary from 1.25×10^{-6} to 1.67×10^{-6} m^2 s^{-1}, while the range of soil water storages (0-40 cm) is 30.4 to 77.7 mm, corresponding to soil moisture contents at 15 cm of 0.06 to 0.18 m^3 m^{-3}, for a porosity

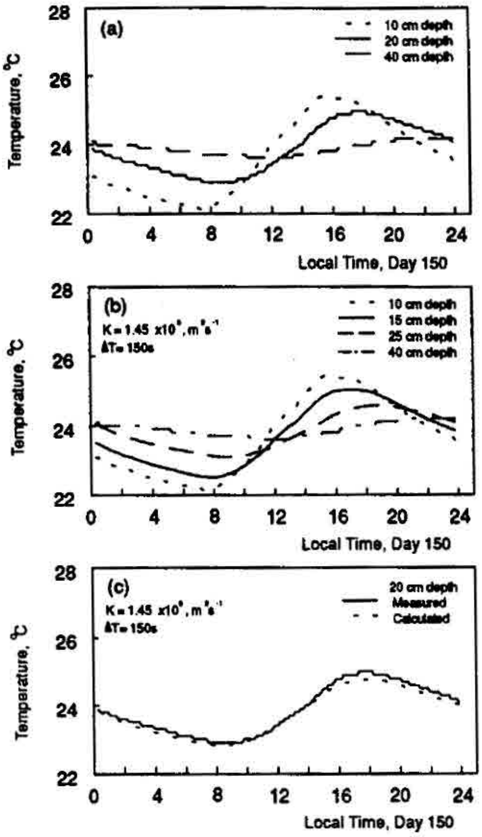

Figure 5 Soil temperatures at Fazenda Nossa Senhora for Day 150, 1993; (a) measured, (b) calculated and (c) a comparison between measured and calculated at 20 cm depth

of 51.1%. The overall mean diffusivity from this method is 1.45×10^{-6} m^2 s^{-1}. These diffusivities are consistent with the values published by Viswanadham and Mohana Rao(1972) and somewhat higher than the values of 1.0 and 0.7 $\times10^{-6}$ m^2 s^{-1} given for wet sand by Rosenberg *et al.* (1983) and Arya (1988) respectively. However, these published values are for wetter sand of lower porosity. There is a slight tendency for the diffusivities from this method to be lower when the soil moisture storage is at its lowest and highest (see Figure 4). Although this trend is not strongly evident in these results, it is consistent with a maximum diffusivity occurring at intermediate soil moisture contents (Wierenga *et al.*, 1969; Campbell, 1977).

The two estimates for diffusivity for the clay soil at Fazenda Dimona using the numerical method (Table 1) are 0.063×10^{-6} and 0.067×10^{-6} m^2 s^{-1}, corresponding to soil moisture contents of 0.36 and 0.38 m^3 m^{-3} for the 15 and 30 cm depths respectively, with a porosity of 59%. The individual r.m.s. differences for these K values are 0.277 and 0.315°C. These diffusivities seem particularly low and are not consistent with the estimates from the 'amplitude' method. This suggests heterogeneities in the 10 to 40 cm layer allowing the thinner layers of the 'amplitude' method to give more plausible values. Wright *et al.* (1996) show that porosity changes from 1.06 to 1.15 in the top 30 cm of this soil. Values for clay soil of 0.2 and 0.18 - 0.51×10^{-6} m^2 s^{-1} are given by Rosenberg *et al.* (1983) and Arya (1988) respectively, although diffusivities for the clay soil of Manaus might be expected not to conform to published values because of its unusually light texture and high porosity (Tomasella and Hodnett, 1996).

These results indicate the need for further work, especially on the moisture content dependence of apparent thermal diffusivity for these and other Amazonian soil types. For successful prediction of the soil surface heat flux boundary condition, more attention must be given to the performance of theory in the shallowest layers of the soil: not only over 24-periods when total soil heat flux is small, but also for individual hours to give accurate estimation of the daytime surface energy budget.

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RESUMO

A difusividade térmica do solo (K) – igual à razão entre a condutividade térmica (k) e a capacidade térmica "bulk" (C) – é um importante parâmetro para a determinação da distribuição da temperatura e do fluxo de calor no solo. Este parâmetro, que contribui para a caracterização do microclima do solo, depende do conteúdo de umidade, bem como da composição e da textura do solo. Medidas da temperatura do solo às profundidades de 10, 20 e 40 cm, sob diferentes condições de umidade, foram utilizadas para determinar K em duas áreas desmatadas, com solos argiloso e arenoso, da Fazenda Dimona (Manaus) e da Fazenda Nossa Senhora (Ji-Paraná), respectivamente. Compararam-se vários métodos para a obtenção de K, incluindo uma solução numérica da Equação de Condução de Calor. Utilizaram-se as temperaturas do solo nas profundidades de 10 a 40 cm como condições de contorno, e um perfil de temperatura inicial medido, para obter valores otimizados de K que ajustem a temperatura em 20 cm. Os valores de K são apresentados como função do armazenamento de água no solo até a profundidade de 40 cm. Encontraram-se também relações lineares entre este armazenamento e a umidade em diferentes profundidades. Obteve-se um valor médio de K, para nove dias selecionados e utilizando o método numérico, igual a $1,45 \times 10^{6}$ m^2 s^{-1} para o solo arenoso, valor este consistente com aqueles obtidos pelos métodos analíticos. Com uma amostra de dois dias para o solo argiloso, obtiveram-se valores menos consistentes de K, variando de 0,07 a $0,45 \times 10^{6}$ m^2 s^{-1}.