1992/023

Institute of Hydrology

Natural Environment Research Council

1990/023

Dr. John H. C. Gash

CEGB RESEARCH FELLOWSHIP ON LAND SURFACE ATMOSPHERE INTERACTIONS

Annual Report 1990

A.J. Dolman

CONTENTS

Appendix A.

Dolman, A.J. (1990a) Using a single column version of a GCM to explore the sensitivity to different land surface descriptions. in : computer modelling in the environmental sciences. D.G. Farmer & M.J. Rycroft (Eds). IMA Conference series. Oxford Univ. Press. pp. 111-123.

Appendix B

Dolman, A.J. (1990c) Outline of a new GCM land surface parameterization, the MITRE model. Unpublished internal note, Institute of Hydrology, Wallingford.

Appendix C

Dolman, A.J., & J.S. Wallace (1991) Lagrangian and K-theory approaches in modelling evaporation from sparse canopies. Quarterly Journal of the Royal Meterological Society : submitted.

Appendix D

Gash, J.H.C., J.S. Wallace, C.R. Lloyd, A.J. Dolman, M.V.K. Sivakumar, & C. Renard (1990) Measurements of evaporation from fallow sahelian savanna at the start of the dry season. Quarterly Journal of the Royal Meteorological Society : Accepted for publication.

Appendix E

Lloyd, C.R., A.D. Culf, A.J. Dolman, & J.H.C. Gash (1991) Estimates of sensible heat flux from observations of temperature fluctuations. Boundary Layer Meterology : submitted.

INTRODUCTION

The land surface provides an important feedback in the global climate system. Improvement in the parameterization of these mechanisms is essential if climate predictions are to become more reliable on a regional scale and if changes in vegetation patterns in response to climatic change are to be modelled with General Circulation Models (GCMs). The existing international research programme in land surface atmosphere interactions seeks to adress these issues by attempting to understand the basic mechanisms of energy exchange of areas the size of a typical GCM grid. Within the programme several large scale international experiments are planned and executed to provide the data required for calibration of descriptions of the spatially averaged interaction of the land surface and atmosphere. The programme of research of the CEGB Fellow at the Institute of Hydrology broadly follows these lines.

In the annual report for 1989 the objectives of the CEGB Fellowship at the Institute of Hydrology were formulated as 1) calibration of existing GCM land surface parameterizations, 2) development of new land surface descriptions, and 3) development of methods of aggregation of heterogeneous land surfaces. The progress made in 1990 in these three subjects areas will be described in this report.

In the calibration of existing parameterizations progress has been made for the tropical rainforest and tropical savannah. A new GCM land surface description was formulated, and theoretical developments in modelling sparse crop energy exchange were used to assess the level of complexity needed in these descriptions. This assessment has lead to a new proposal for a land surface parameterization in the UK Meteorological Office GCM. These developments were regularly discussed within the collaborative project between the Institute of Hydrology and the newly formed Hadley Centre for Climate Prediction and Research (MITRE). In February 1990 an assistant to the Fellow was appointed who started work on aggregation using boundary layer models developed by the Meteorological Office. Experimental work in the Sahel was performed to gain additional insight in regional evaporation through the use of tethersondes.

As the CEGB fellowship forms an integral part of the ongoing Land

Surface Energy Balance research programme at the Institute of Hydrology, developments in this programme will also be described.

CALIBRATION OF EXISTING LAND SURFACE PARAMETERIZATIONS

The tropical rainforest parameterization of the UK Meteorological Office General Circulation Model (UKMO GCM) has been calibrated against data from the Amazon Regional Micrometeorological Experiment in 1982-1985 (Shuttleworth, 1988). In Figure 1 the land surface parameterization of this scheme is shown. Values for the necessary parameters were obtained by calibrating the results of a one dimensional version of the GCM against the observations. A new description of the interception process has been developed and incorporated in the UKMO GCM. Results of the inclusion of this description improve the results dramatically in the 1-D model (Dolman, 1990). Similar improvement has been obtained with the full three dimensional GCM (Lean, pers. communication). The use of this new scheme allows a much more realistic assessment of the consequences of tropical deforestation with GCMs and are likely to improve the GCM surface hydrology in other areas of the world.

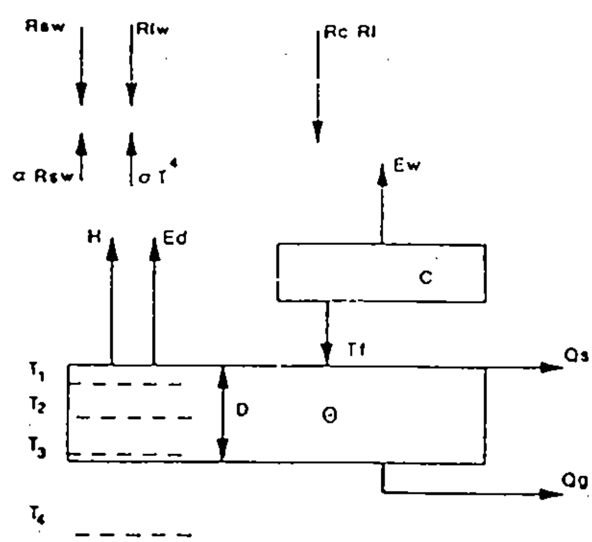

Figure 1. Schematic diagram of the land surface scheme as used in the UKMO GCM and in the 1-D model (Redrawn from Warrilow and Buckley, 1989). Rsw, Rlw are short and longwave radiation, α albedo, Rc and Rl convective and large scale precipitation, Ew and Ed wet and dry canopy evaporation, H sensible heat flux, T_i temperature of the i^{th} soil layer, D rooting depth, θ soil moisture content, Qs and Qg surface runoff and gravitational drainage C canopy water content and T_f throughfall.

The 1-D model experiments also highlighted the control of large scale mid-continental evaporation. The simulations show that evaporation of the mid-continental tropical rainforest area in Brazil is mainly determined by the amount of available energy. In Figure 2 the sensitivity of transpiration (dry canopy evaporation) to the value of the surface resistance is shown. Over quite a large range of resistance, the total transpiration does not change appreciably. Simple stand alone sensitivity experiments however, show a large dependence of transpiration on surface resistance for tropical rainforest (Dolman, et al, 1991, Shuttleworth, 1988). As can be seen from Figure 2 the processes

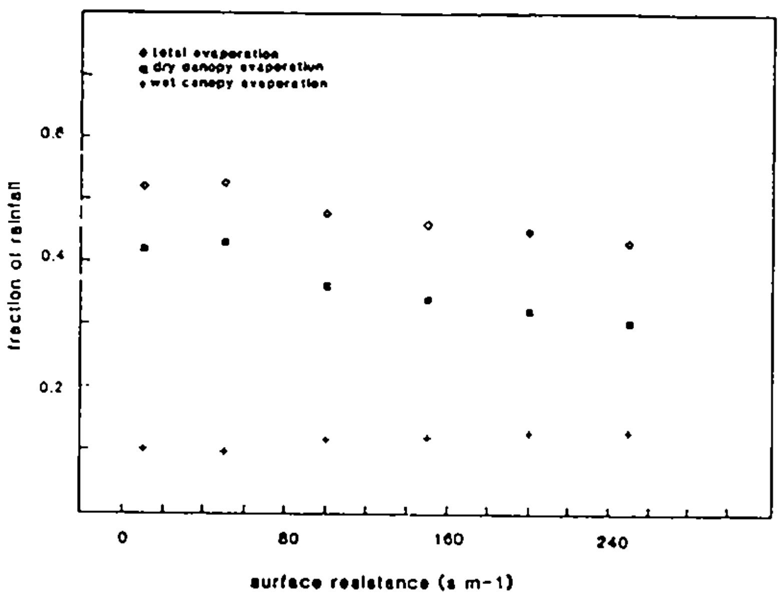

Figure 2.Dry and wet canopy, and total evaporation as a function of surface resistance. Evaporation is expressed as a fraction of annual rainfall. For realistic values of resistance (> 80 sm^{-1}) total evaporation changes hardly as indeed does not transpiration.

of transpiration and interception in the 1-D model experiments balance each other, with the total evaporation remaining roughly at the same amount. This situation holds only for well watered vegetation, but it shows the usefulness of the 1-D model in assessing the sensitivity of the climate to changes in the physics parameterizations of the model. The new Unified Model (UM) 1-D version with an improved vertical resolution near the surface, will be used to further investigate the large scale controls on the surface energy balance.

A start has been made with the calibration of the tropical savannah biome against data obtained by the Sahelian Energy Balance Experiment (SEBEX) (Wallace et al, 1990). Data on sensible and latent heat flux obtained at the end of the wet season and the start of the dry season have been used to obtain simple empirical descriptions of surface conductance of a fallow savannah site in Niger (Gash et al, 1991, appendix).
Figure 3 shows the values of surface conductance, obtained by inverting the Penman-Monteith equation, throughout the observation period. The few points obtained on the days immediately following rainfall on 1-3 October show high conductance values, but by the end of the period the values have fallen to around 2 mm s^{-1}. A simple description of surface conductance based on the number of days after rainfall has been derived and is also shown in Figure 3. This model can be improved

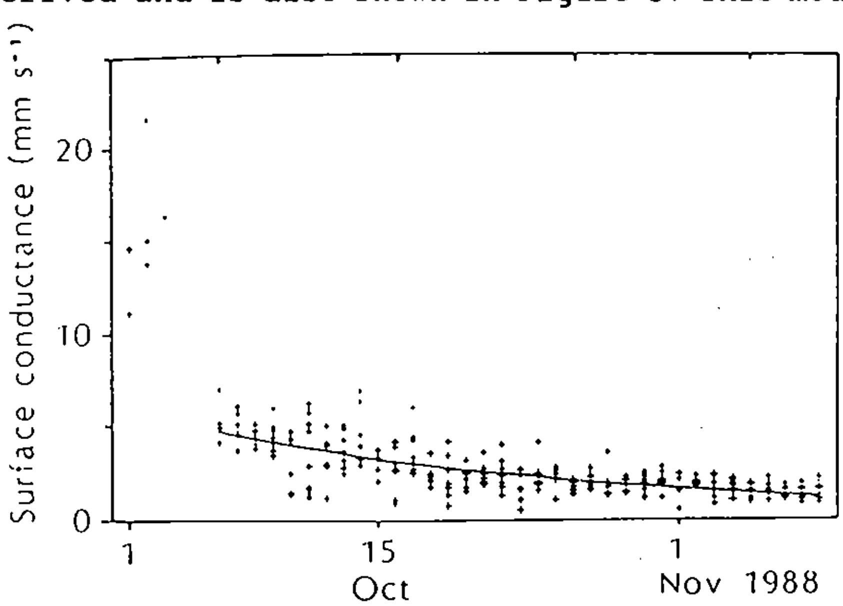

Figure 3.Surface conductance of tropical fallow savannah in Niger during the drying out period. Last rainfall was at 28th September. Values decrease from around 5 to 6 mm s^{-1} to 1 mm s^{-1} when the canopy and grass start dying off as a result of soil moisture stress.

slightly by introducing additional relations of surface conductance with time of day, or soil moisture deficit (Figure 4). Further modelling work is in progress and will attempt to separate the evaporation from the bushes and the grass-soil mixture. Initial results obtained with this model for the tropical savannah are encouraging. The model has been formulated to account for the two components and will form the basis for an extension of the UM modelling scheme and is more fully described in the appendix. Experiments with the UM 1-D model will be initiated to examine the large scale sensitivity of the Sahelian climate to changes in its land surface parameterization.

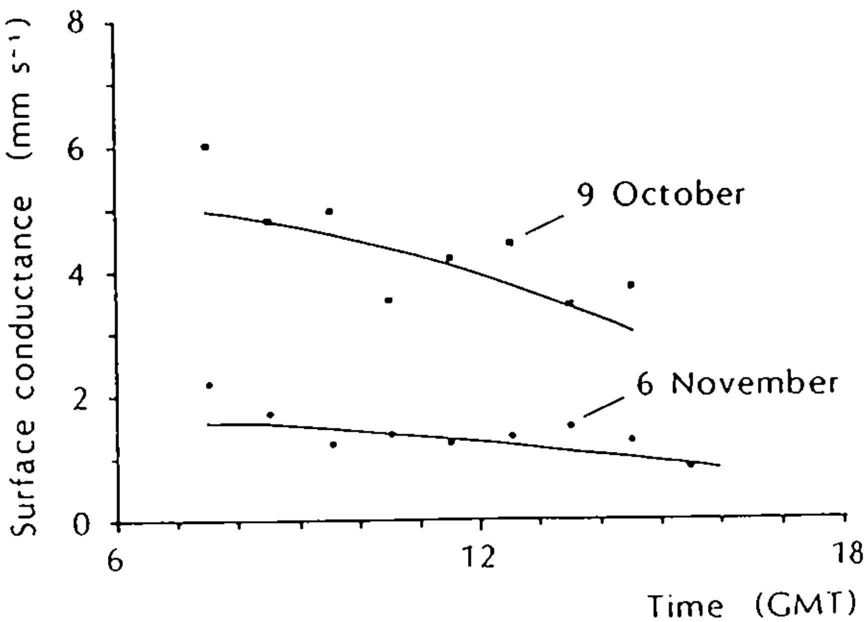

Figure 4.Daily course of surface conductance of fallow savannah in Niger for two days, at the beginning and end of the drying out period. The solid lines represent a simple model description based on time of day and soil moisture content.

DEVELOPMENT OF NEW LAND SURFACE PARAMETERIZATIONS

The basis of the newly proposed land surface scheme is formed by the introduction of a plant canopy with a separate plant canopy surface temperature. Such a model has to describe the in-canopy turbulent transport. It is well known from recent experimental work that the standard description of this transport by gradient-diffusion theory is in some cases too simple a working approximation of the complex in-canopy transport processes. A new evaporation model for sparse canopies, based on Lagrangian turbulent transport processes, was developed to be tested against more simpler approximations of sparse canopy evaporation. This model and the results are more fully described in the appendix (Dolman and Wallace, 1991). Several models were set up to predict evaporation from a sparse millet crop in Niger, West Africa. The Lagrangian model was shown to perform well for a variety of environmental conditions. When the soil was wet a comparison between three models showed relatively few differences.

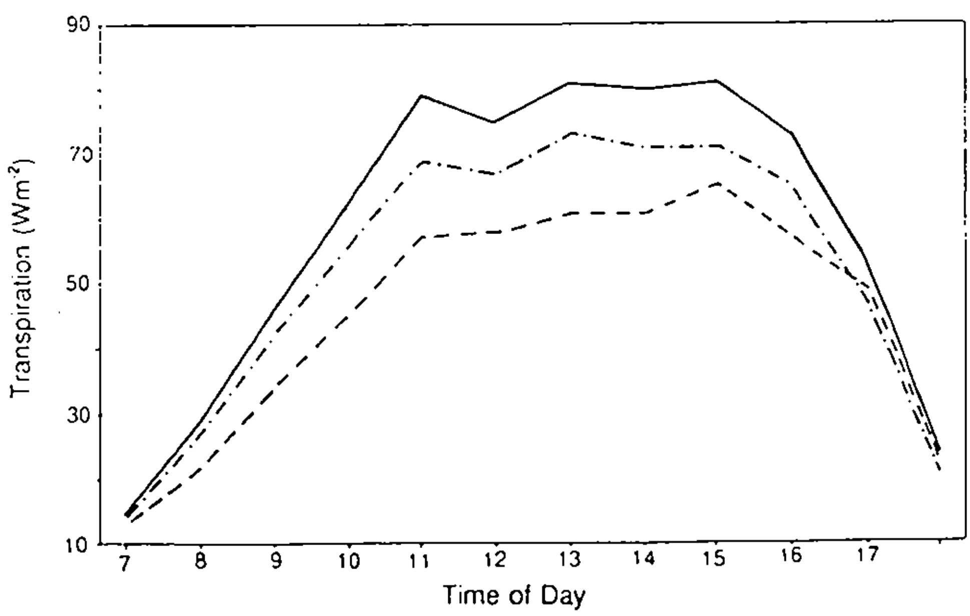

Figure 5.Transpiration from a millet crop in Niger estimated by three models. The solid line represents the transpiration estimated by the new Lagrangian model, the – - line a simple model allowing for soil evaporation, and the – – line a single big leaf description. Taking the interaction between soil and canopy into account clearly increases the transpiration rate from the crop.

However, when the soil was dry and a large heat flux emanated from the soil, the in-canopy humidity deficit was shown to be strongly modified by the sensible heat coming from the soil and the Lagrangian model predictions differed from the more simpler descriptions (Figure 5). These results were used to asses the relative importance of the description of in-canopy turbulent transport. From the results it was concluded that the required additional input data and computational complexity do not justify the inclusion of these processes in a GCM land surface parameterization.

On the basis of these results a new land surface description for use in the UM GCM was proposed. The model which includes a separate canopy temperature is based on an extension of the Penman-Monteith "big leaf " description (Figure 6). Rather than solving for the transfer equations of heat and moisture directly, it uses an energy balance approach to estimate the surface evaporation. There is no need for a separate prognostic surface temperature equation in this scheme, so it

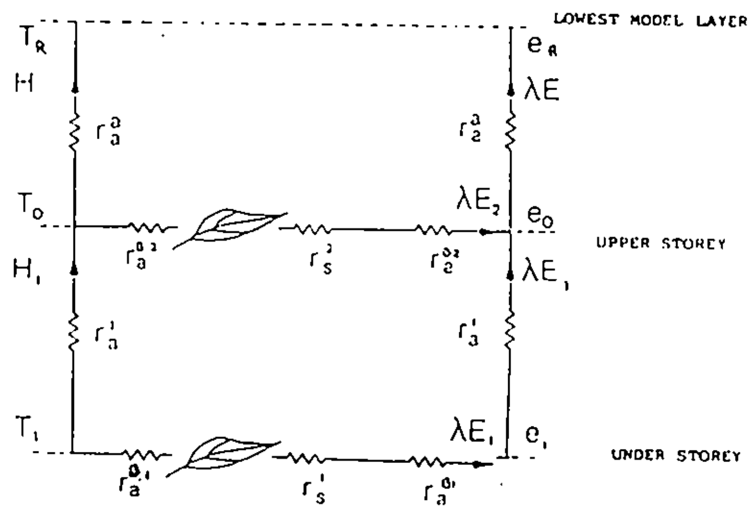

Figure 6.Schematic diagram showing the resistance network used to calculate energy exchange from a vegetation type with an upper and under storey in the MITRE model. Resistances are denoted by the letter R, the subscript s refers to a surface (plant physiological) resistance, the subscript a to an aerodynamic resistance, H is the sensible heat flux and λE the latent heat flux (evaporation).

greatly simplifies computations. The original Penman-Monteith equation was reformulated by Shuttleworth and Wallace (1985) and Choudhury and Monteith (1988) to include the soil as an extra source. In the proposed model (appendix), this approach is further extended by allowing the radiation to reach a canopy and an understorey without being attenuated by canopies at higher layers. The result of this extension is a flexible modelling scheme which can be used for a variety of surfaces. Implementation in the new UM GCM will be performed stepwise in close cooperation with staff of the physical parameterization group at the Hadley centre for climate prediction and research. First the numerical scheme will be adjusted to incorporate a one layer description of the Penman-Monteith equation. In a later stage a canopy will be added. New formulations will be added to describe the aerodynamics of sparse canopies, and the old soil temperature scheme will be reformulated to be compatible to the new energy balance approach. Initial testing of the new model will be performed with the UM 1-D model, whilst SEBEX data will be important in the fine tuning of the parameters. In a further stage the surface resistances of the canopies may be made dependent on environmental variables. At some stage a new formulation of land use types may be introduced to replace the datasets currently used in the GCM. The new model can be seen as a first step towards a fully interactive vegetation, which grows in response to a changing climate.

DEVELOPMENT OF AGGREGATION METHODS

Variations in surface cover have the potential to affect the structure of the boundary layer which provides the link between the surface and the general circulation. Analysis of the HAPEX-MOBILHY experiment in South West France during 1986 has suggested that variations in surface cover may be classified in two types: those which affect the boundary layer and those which do not. The last type of variation appears on a horizontal scale which is small compared to the footprint of the boundary layer i.e. less than 10 km. It is this type of variation which is the subject of present research within the CEGB Fellowship.

To study the effects of advection on area average surface fluxes a 2-D numerical model of the boundary layer is used. This model was developed by the boundary layer branch of the Meteorological Office and is used within a collaborative project to investigate the aggregation of water and heat fluxes. It is currently used

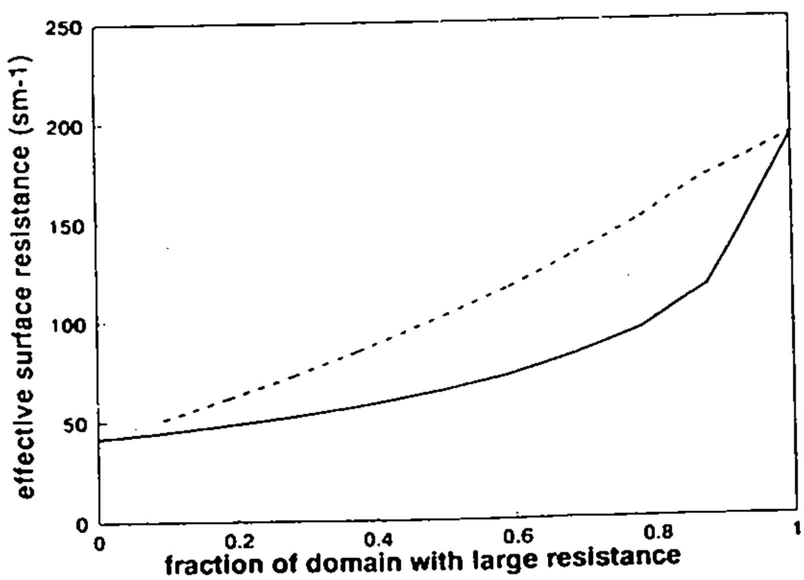

Figure 7. The effective surface resistance as a function of the fraction of the domain covered by the vegetation with the largest resistance. The broken line represents calculations made with a model with no energy balance, the broken line represents a model with a surface energy balance. In both cases the effective resistance is less than mean resistance as a result of non-linearities in the equations.

as a benchmark, against which to test simpler and more approximate descriptions, which are at present based on heuristic arguments. A substantial effort has been put in to reformulating the heuristic models so that they incorporate realistic descriptions of the earth's surface energy balance.
Figure 7 shows an example of the use of two heuristic models, which differ in the fact that model I does have an energy balance and model II does not. The area average resistance as predicted by these models is plotted as a function of the fractional area covered by one type of vegetation. The differences between the approaches suggest that the inclusion of an energy balance in the models is a prime requirement. The 2-D boundary layer model will be extended to include such an energy balance in 1991.

In 1990 a collaborative project was initiated with the French Meteorological Service (CNRM, Toulouse) to study the agregation of surface fluxes with a mesoscale model. The additional funding for travel and subsistence comes from the Alliance programme which is a programme aimed at cooperation between British and French scientific institutions and is run by the British Council and its French counterpart. This mesoscale modelling project will look specifically at situations where part of the domain is wet and will thus complement the previously described work on convective rainfall interception modelling.

EXPERIMENTAL WORK IN THE SAHEL

In addition to the SEBEX programme, and the boundary layer experiment alongside that, in July 1990, tethersondes were used in Niger, West Africa to gain insight into the structure of the surface layer over an area of degraded forest (Tiger bush). The wind profile data were used to determine the height range of this surface layer and the aerodynamic roughness length. Previous experiments with GCMs have shown that changes in the value of roughness length of areas close to deserts may have the same effect as changing the albedo, i.e. reducing the rainfall. Profiles of windspeed and potential temperature were used to determine the value of roughness length for an area which is much larger than those sampled by standard micrometeorological techniques. A value of 0.49 m was found from the wind profiles. The extent to which the logarithmic

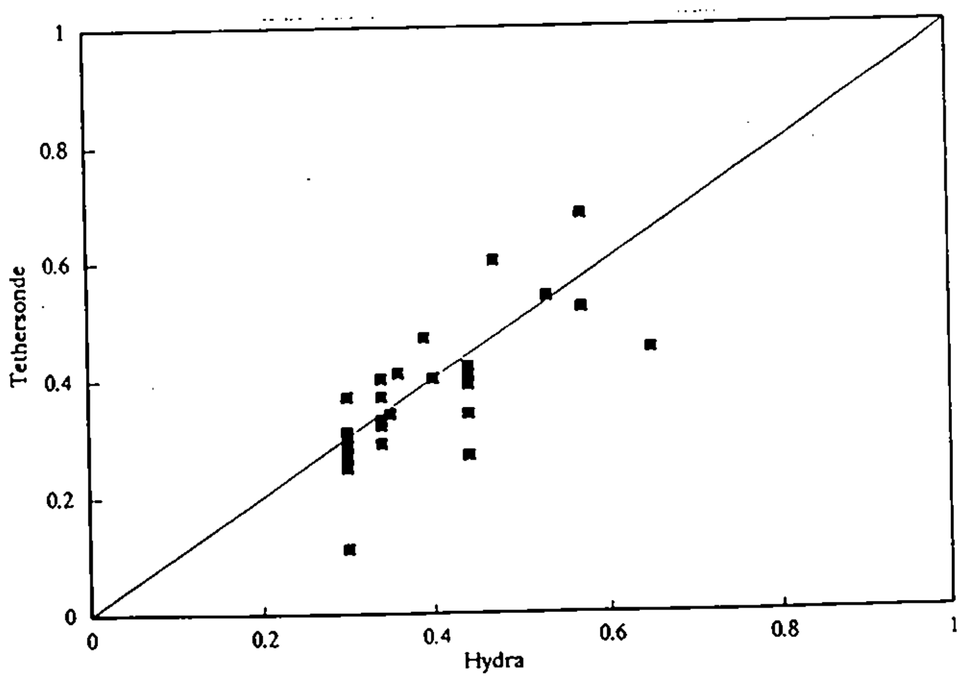

Figure 8. Friction velocities ($\sqrt{\tau}$, where τ is the momentum flux) estimated from the windprofiles measured by a tethered balloon above a degraded forest in the Sahel compared with measurements made by an eddy correlation instrument (Hydra) mounted on a tower. The agreement suggest that the tethered balloon profiles can be used to estimate regional characteristics of momentum and energy fluxes.

layer was valid was determined as 100m. The data on temperature and humidity will be used to determine regional evaporation and heat flux. As an example of this approach to estimate regional fluxes, in Figure 8 the friction velocity as measured by the tethersonde system is compared with that obtained by the Hydra eddy correlation instrument mounted on a tower. The agreement is good, suggesting that the profiles can be used in assessing the regional characteristics of momentum exchange and possibly evaporation.

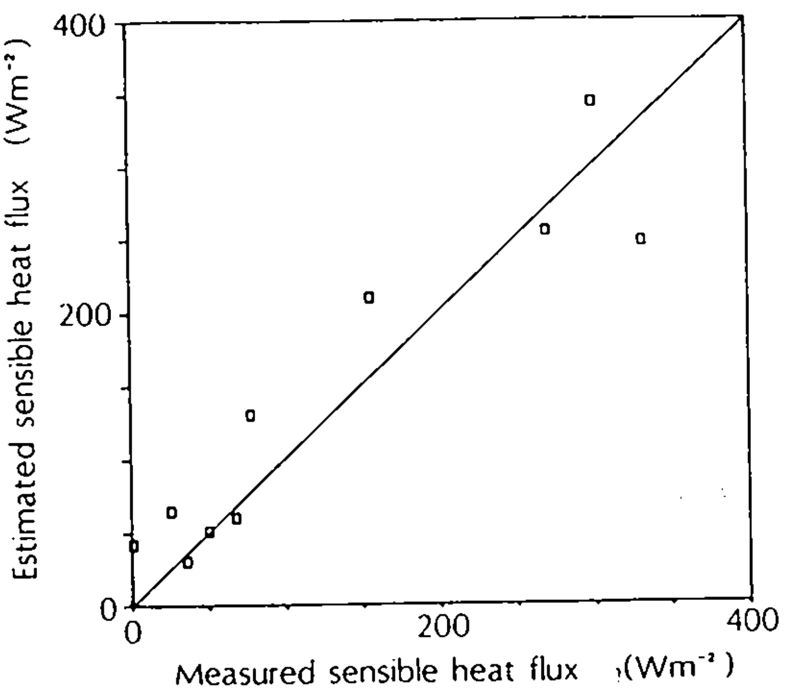

Figure 9. Comparison of sensible heat flux measured by the eddy correlation instrument on a tower above a degraded forest in the Sahel and sensible heat flux estimated by the temperature fluctuation method from a tethered balloon.

As part of the regular programme of instrument development at IH, the tethersonde was equipped with a fast response temperature sensor, to measure the temperature fluctuations in the surface layer. Theory predicts a universal relationship between the fluctuations in potential temperature and sensible heat flux (Lloyd et al, 1991, Appendix). A measurement of these fluctuations may thus be used to obtain estimates of heat flux in an inexpensive way. Especially when mounted on a tethered balloon, this may be a cheap way of obtaining regional estimates of heat flux, which cannot be obtained by standard micrometeorological measurements from height limited towers. Figure 9 shows a comparison of heat flux as estimated by the temperature fluctuation method and the eddy correlation technique at a tower. The results are encouraging and show the technique to have promise in obtaining larger scale estimates of components of the surface energy balance.

PROGRESS IN THE IH LAND SURFACE ENERGY BALANCE PROGRAMME

The Institute has an ongoing programme of research which is specifically designed at improving the land surface representation in General Circulation Models. Tackling the problems associated with this, has led to several new initiatives, both theoretical and practical.

The MITRE link with the new Hadley Centre for Climate Prediction was further strengthened. Regular meetings and discussions over details in the land surface parameterizations in the new UM GCM have taken place in 1990 and will continue. Close collaboration has been initiated to reformulate the land surface description along the lines outlined in the proposal in the appendix.

In 1990 the Anglo-BRAzilian Climate Observational study (ABRACOS) started. This bi-lateral study aims at providing the GCM modelling community with the necessary data on cleared forest to calibrate GCM predictions of tropical deforestation. The Institute of Hydrology is collaborating with several Brazilian institutions (INPA, INPE and CENA). The main field site for detailed study of water and energy balances of cleared areas is a cattle ranch near Manaus. The observation of standard meteorological data is complemented by plant physiological studies and soil moisture measurements. In parallel with these detailed studies climate data will be obtained at two other selected sites west and east of Amazonia (Rondonia and Para respectively) to investigate possible differences in climate between these locations. Initial data have been obtained for the albedo of cleared forest, and the soil moisture measurements at the cleared site have documented the drying out phase quite accurately.

The SEBEX experiment had its last field season in 1990. Data have been obtained for three different land use types in the Sahel and the micrometeorological work was complemented by boundary layer soundings. Initial results have shown that there are large differences in the micrometeorological behaviour of the degraded forest area and the fallow savannah. Analysis and subsequent modelling of these differences will be performed in 1991. The data will be used to calibrate the new land surface scheme in the UM GCM and will probably be used to calibrate the Simple Biosphere model of the NASA GCM. The data were furthermore used to test the temperature fluctuation method of sensible heat flux estimation over four different surfaces (Lloyd et al, 1991, Appendix)

The boundary layer links the surface to the general atmospheric circulation and is of immediate importance in land surface process studies for GCMs. The radio soundings in the Sahel have been used to test simple models of boundary layer growth. The results show that these models can describe the growth of the boundary layer quite accurately, both for low boundary layer heights as occur in the wet season as well as for the very high boundary layer heights observed in the dry season, when most of the energy received at the surface is returned as sensible heat. In the future the models may be inverted to obtain estimates of GCM parameters on the appropriate scale. The data also provide possibilities of validating the GCM boundary layer height predictions.

Analysis continued on the data obtained in the FIFE experiment in Kansa. A most important result is the finding that sites with a different leaf area index gave similar evaporation rates and have similar surface conductances. Modelling surface conductance of these sites continues.

For 1992 a large scale international field experiment is planned in Niger. This experiment in the WMO coordinated HAPEX series will investigate the energy and water fluxes in an area of 100 by 200 km near Niamey. Three main sites will be installed along the rainfall gradient in the Sahel. Several groups from the UK, the USA, Holland, France and Germany will participate in this project. Within the project IH has a leading role in coordinating micrometeorological flux and vegetation studies.

In May and June 1991 the EFEDA project will take place in an desertification threatened area in Spain. This project which is partly funded by the European Community will again involve participation of several research groups from different countries. Site selection and planning is in an advanced state of progress.

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APPENDIX A

USING A SINGLE COLUMN VERSION OF A GCM TO EXPLORE THE SENSITIVITY TO DIFFERENT LAND SURFACE DESCRIPTIONS

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1.1 INTRODUCTION

General Circulation Models (GCM's) attempt to model the three dimensional structure of atmospheric motions. On a global basis 50% of the radiation at the top of the atmosphere reaches the surface, where about two thirds of that amount is being converted into sensible and latent heat (water vapour flux), the remainder reflected and used to warm up the surface (GARP, 1975). The land surface plays an important role in this process, partly through the positioning of the continents relative to the oceans, and partly through the fact that the surface characteristics of the land surface determine the way in which the exchange processes take place. Early experiments with GCM's have shown a great sensitivity of the global circulation patterns to changes in the land surface parameterization. Shukla and Mintz (1982) have shown that altering soil moisture initialization in a GCM has a pronounced effect on the rainfall pattern over continental land masses. Sud et al. (1985) and Sud et al. (1988) showed that changing the surface roughness in a GCM has a large influence on the horizontal convergence of moisture in the atmospheric boundary layer and on the distribution of convective rainfall. Charney et al. (1977) showed how an increased albedo in the Sahel region could lead to further desertification through a feedback process whereby a reduced evaporation rate would result in lower cloudiness and lower precipitation rates. Reviews of the influence of the land surface on global circulation patterns can be found in Mintz (1984) and Rowntree (1986).

Most of these experiments are sensitivity tests of the model climate to the alteration of a single parameter such as albedo or surface roughness. Being physically not very realistic, these experiments served mainly to express the need for more accurate descriptions of the interaction of the land surface with the lowest layers of the atmosphere. Recently a number of new land surface parameterization schemes have been developed by Dickinson (1984), Sellers et al. (1986) and Noilhan and Planton (1989). They incorporate a considerable degree of complexity in their description of vegetation-atmosphere interaction. In developing these models, they are usually calibrated against micrometeorological data (e.g. Sellers et al. (1989), Pinty et al., (1989)). Very few experiments exist where complete global climate simulations are made with these schemes (Sato et al., 1989).

rainfall through a relation of the standard deviation of the dewpoint temperature and air temperature. Gaussian distributions are assumed for all the random variables and a random number generator is used to derive actual values once a day. The avoid large step changes in the atmospheric variables from day to day, the actual value used on a single day is interpolated from the values of two days.
Four subroutines deal with the physical processes such as radiation, dry and moist convection and large scale (frontal) precipitation and surface and boundary layer turbulent exchange. The model iswritten in Fortran. Timesteps in the model are typically 300 to 900 seconds and integrations are usually performed over a full annual cycle. For the experiments described in this paper, the model was run on an IBM PS/2 machine with 25 Mhz processor. A complete annual cycle takes several hours CPU time.

1.3 LAND SURFACE SCHEME

The land surface scheme presently used in the UKMO GCM is shown in figure 1. Surface temperature is a prognostic variable and is calculated using a four layer surface–soil temperature

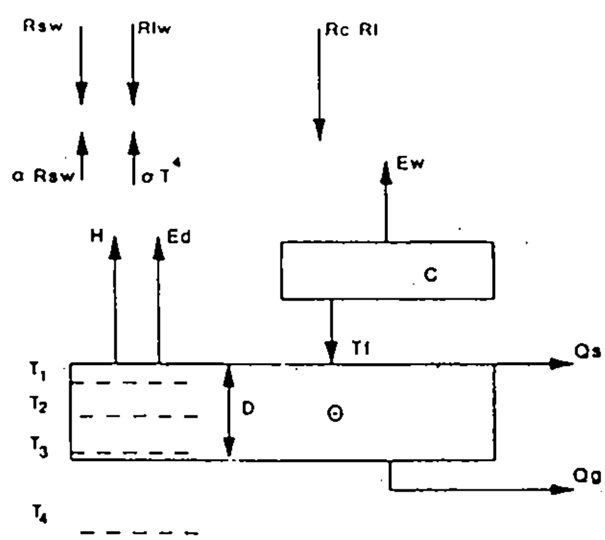

Figure 1. *Schematic diagram of the land surface scheme as used in UKMO GCM and in the original version of the SCM (Redrawn from Warrilow and Buckley, 1989). Rsw, Rlw short and longwave radiation, α albedo, Rc and Rl convective and large scale precipitation, Ew and Ed wet and dry canopy evaporation, H sensible heat flux, T_i temperature of the i^{th} soil layer, D rooting depth, θ soil moisture content, Qs and Qg surface runoff and gravitational drainage, C canopy water content, T_f throughfall.*

TABLE 1.

Results of GCM deforestation experiments. LW refers to the results of Lean and Warrilow (1989), DHS to Dickinson and Henderson Sellers (1988) and SNS to Shukla et al. (1989). Values refer to the difference between deforested and control climate.

	LW	*DHS*	*SNS*
evaporation (mm day^{-1})	−0.61	−0.52	−1.36
rainfall (mm day^{-1})	−1.34	0	−1.76
surface temperature (°C)	1.98	2−5	2.5

2.1 SINGLE COLUMN MODEL EXPERIMENTS: WET CANOPY EVAPORATION

Overestimation of interception loss is likely to be caused by the very low average rainfall intensities in a GCM grid. Typical average rainfall intensities are 0.25 mm hr^{-1} in the SCM. Average rainfall intensities as measured in the Reserva Ducke are 5.15 mm hr^{-1} (Lloyd et al., 1988). This difference in rainfall intensities causes problems in the modelling of interception loss. Typical canopy storage capacities for forest are about 1 mm, for the tropical rainforest, Lloyd et al. (1988) estimate a value of 0.7 mm. In GCM's which take account of interception loss, the rainfall is essentially assumed to cover the complete grid. As the grid rainfall is a composite of several storms of variable intensity, duration and surface coverage, this is probably not very realistic.

Shuttleworth (1988b) devised a scheme whereby the grid average rainfall is assumed to cover only a small part of the grid if the rainfall is of convective origin. The local rainfall rate in the rain covered area is assumed to follow a negative exponential probability distribution

$$f(P_l) = \frac{\mu}{P} \exp\left[\frac{-\mu P_l}{P}\right] \qquad (2.1)$$

where P and P_l are the grid average and local rainfall rate respectively and μ the proportion of the grid which is wet. The average grid throughfall rate follows as

$$T = P \exp\left[\frac{-\mu C_m}{P}\right] \qquad (2.2)$$

where C_m is defined as the maximum canopy infiltration rate per timestep.

In figure 2 the sensitivity of grid average throughfall to changes in the fraction of the grid wetted before full canopy saturation is reached. Although the scheme appears to perform better than WJS1 it does so because it effectively allows rainfall to run off from unsaturated canopies. Shuttleworth (1988a) reports a substantial sensitivity of modelled interception loss to the canopy storage capacity. The main reason for this is that, contrary to temperate forest interception where evaporation during the storm is important, most of the interception loss in rainforests is made up by evaporation after the rainfall has ceased. Changing the value of S, the canopy capacity from 2.5 to 0.7 mm hardly affected the modelled interception loss in the original scheme.

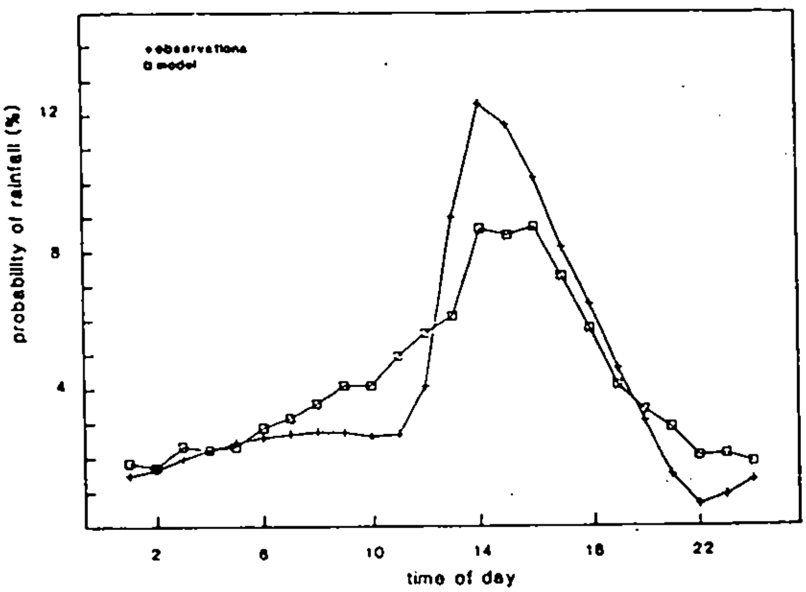

Figure 3. *Comparison of observed and predicted frequency distribution of rainfall.*

Figure 3 shows the predicted frequency distribution of rainfall with time of day at Reserva Ducke Manaus, compared with observations reported by Lloyd (1990). The dominant convective nature of Amazonian rainfall, with showers occurring predominantly in the early afternoon, is reproduced quite well by the SCM. This suggests that the grid average rainfall is correct. Given the sensitivity of modelled interception loss to the exact value of μ (figure 2), and the variation in rainfall amount with time of day, it is quite likely that assuming a constant value of μ=0.3 is not realistic. The Shuttleworth scheme can be extended by incorporating an explicit dependence of μ on the grid average rainfall rate according to

$$\mu = P / P_r \qquad (2.3)$$

where P_r is an observed average rainfall rate for convective storms. From observations in the Reserva Ducke (Lloyd et al., 1988), P_r is found to be 5.15 mm hr^{-1}. The results of this scheme (WJSII) are also given in table 2. The result compare well with the observations. The calculated

which shows that the rate of change in specific humidity, q depends on the evaporation rate E and the height of the boundary layer. Any increase in surface resistance and hence potential decrease in evaporation will result in a decrease in humidity and increase in surface temperature (and corresponding non-linear increase in saturated specific humidity at the surface). The resulting gradient will then be a balance between the decrease in lowest layer humidity and increase in surface saturated humidity through an increase in surface temperature.
The present boundary layer model has a fixed boundary layer height and thus tends to present a strong feedback of the boundary layer to surface evaporation. Neglecting the sink of humidity caused by moist convection, it is essentially a closed box. Such a model tends to produce evaporation rates mainly dependent on the available energy. The relative insensitivity to changes in surface resistance suggest a similar phenomenon may be operating in the SCM.

3.1 DISCUSSION

It has been shown that the incorporation of a subgrid parameterization to allow for spatially varying surface coverage of convective rainfall improves the partitioning of evaporation between wet and dry canopy evaporation. Nevertheless, the original models produced the correct total evaporation. The primary cause for this that on a larger scale the evaporation will always be constrained by the available energy. Average net radiation in the SCM varies around 105 W m^{2}, Lean and Warrilow report a figure of 147 W m^{-2}. Observed net radiation is about 120 W m^{2}. As observations show that 90% of the available energy in the Amazon Basin is used to recycle rainfall through wet and dry canopy evaporation (Shuttleworth, 1988a), a correct prediction of net radiation is an obvious first requisite to be able to predict the total evaporation.
The SCM showed a remarkable low sensitivity to changes in the surface resistance. Boundary layer modelling experiments with more complex models show such a moderation to be likely, although not on the scale as observed in the SCM. It remains to be explored in further experiments whether an increased resolution and a varying boundary layer height will show a more pronounced response to changes in surface resistance. Furthermore, the neglect of entrainment at the top of the boundary layer may give a much lower sensitivity to the specification of surface resistance.
The SCM has however proved to be a very useful model in the development of new land surface parameterizations for use in GCM's. It reproduces clearly the large scale phenomena very well and the fact that it interacts with the surface in a way that micrometeorological models cannot, makes it very suitable for the purpose of developing new land surface parameterizations in GCM's.

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APPENDIX B

OUTLINE OF A NEW GCM LAND SURFACE PARAMETERIZATION, THE MITRE MODEL
Draft, not to be quoted

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Rationale

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The current generation of land surface models in GCM's (BATS and Sib, Sellers et al, 1985 and Dickinson et al, 1986) rely on the straightforward solution of the transfer equations of heat and moisture. They include a canopy which distinguishes them from the Met Office scheme (Warrilow et al, 1986). The inclusion of a canopy is especially important in the case of sparse canopies, where the soil can be an important additional source of heat and water vapour. The equivalent holds for vegetation consisting of an upper and understorey.

It is proposed to take the current GCM land surface models one step further and introduce the combination equation of Penman (1948) with the resistance formulation of Monteith (1965) to solve directly for the fluxes of heat and water vapour. Extensions of the single source, big leaf model of Monteith (1965) to include the soil component, have been given by Shuttleworth and Wallace (1985) and Choudhury and Monteith (1988) and form an integral part of the new model. This approach is here extended to incorporate the case of vegetation consisting of two separate layers.

AJD 31/10/90

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Theory

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Describing the fluxes of moisture and heat as

$$\lambda E = (\varrho c_p/\gamma)\ (e_r-e_s)\ /\ (r_a+r_s) \qquad (1)$$

$$H = \varrho c_p\ (T_r-T_s)\ /\ r_a \qquad (2)$$

where ϱ is the density of dry air, c_p the specific heat of air, e_s and e_r the vapour pressure at the surface and reference height respectively, T_s, T_r, the respective temperatures, r_a the aerodynamic resistance to vapour transport from the surface to the reference height, and r_s the surface resistance to water vapour transport (the sum of the stomatal resistances), and introducing the slope of the saturated vapour pressure versus temperature curve

$$\Delta = (e_s-e_r)/(T_s-T_r) \qquad (3)$$

and the energy balance at the surface

$$A=\lambda E + H - G \qquad (4)$$

where, G, is the ground heat flux, it is possible (Monteith, 1965) to eliminate the surface values of temperature and humidity and obtain a combination equation for evaporation:

$$\lambda E = \{\Delta A + \varrho c_p \delta e/r_a\}\ /\ \{\Delta+\gamma(1+r_s/r_a)\} \qquad (5)$$

where δe is the vapour pressure deficit at reference height. This equation has been extensively used for both diagnostic and predictive purposes in the last decade. The sensible heat flux is given by

$$\lambda E = \alpha\, C_1\, PM_1 + (1-\alpha)\, C_2\, PM_2 \tag{6}$$

where

$$PM_1 = [\, \Delta A + \{\varrho c_p \delta e - \Delta r_a^1 A_1\}/(r_a^a + r_a^1)\,]$$
$$[\, \Delta + \gamma\, \{1 + (r_s^1 + r_a^{b1})/(r_a^a + r_a^1)\}]^{-1} \tag{7a}$$

$$PM_2 = [\, \Delta A + \{\varrho c_p \delta e - \Delta r_a^2 A_2\}/(r_a^a + r_a^2)\,]$$
$$[\, \Delta + \gamma\, \{1 + r_s^2/(r_a^a + r_a^2)\}]^{-1} \tag{7b}$$

and the coefficients C are given as functions of the resistances:

$$C_1 = R_2\, (R_1 + R_a)/\{R_1 R_2 + \alpha R_2 R_a + (1-\alpha)\, R_1 R_a\} \tag{8a}$$

$$C_2 = R_1\, (R_2 + R_a)/\{R_1 R_2 + \alpha R_2 R_a + (1-\alpha)\, R_1 R_a\} \tag{8b}$$

and

$$R_1 = (\Delta + \gamma) r_a^1 + \gamma (r_s^1 + r_a^{b1}) \tag{9a}$$

$$R_2 = (\Delta + \gamma) r_a^{b2} + \gamma r_s^2 \tag{9b}$$

$$R_a = (\Delta + \gamma) r_a^a \tag{9c}$$

The total available energy is given by

$$A = \alpha\, A_1 + (1-\alpha)\, A_2 \tag{10}$$

This model is different from both the Shuttleworth and Wallace (1985) and Choudhury and Monteith (1987) models in that the sparseness of the vegetation is such that both the canopies or the soil

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Comparison with the "unified model" scheme

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The incorporation of this scheme makes a detailed soil temperature scheme in a way superfluous, as the surface fluxes are no longer calculated from the gradient in temperature and humidity over the lowest model layer but from their values at the lowest model layer. The scheme can however still be used for diagnostic purposes for instance in a two-layer mode. Furthermore the soil scheme in the case of ice and snow melt the scheme would still be necessary. For radiation calculations the surface values of temperature can be obtained by solving the transfer equations, assuming similarity of aerodynamic and radiative surface temperature. Stability corrections can be incorporated through an iterative loop which calculates *L*, the Monin-Obhukov length scale. These corrections affect only the aerodynamic resistances. For example:

$$r_a^a = u/u_*^2 + \{ \ln (z_0/z_{0h}\} + \Psi_m - \Psi_h \} \{ku_*\}^{-1} \qquad (12)$$

where Ψ_m, and Ψ_h are the integral diabatic correction factors for momentum and water vapour. It is proposed to calculate the friction velocity from a knowledge of the roughness length and displacement height of the vegetation. The model uses leaf area index as an additional variable, which is not available in the Wilson dataset. Inclusion of leaf area allows interactive calculation of d and z_0 as a function of leaf area, and above all, leaf area can be used to prescribe seasonal variation in canopy parameters.

The fraction of the available energy leaving the system as ground heat flux has to be specified. In the unified model it is calculated as a rest term after the fluxes of heat and moisture have been determined.

The surface resistance of the vegetation can be made dependent on environmental variables. Ultimately, the proposed model could be extended to incorporate carbondioxde uptake.

In the new model the treatment of snow and ice can be updated to the level as currently used in the IH Distributed Model (IHDM). This would include drip of water through the snowpack and heatconduction through the pack. It could also be extended to include an age dependent albedo.

AJD 31/10/90

APPENDIX C

LAGRANGIAN AND K-THEORY APPROACHES IN MODELLING EVAPORATION FROM SPARSE CANOPIES

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Submitted To The Quarterly Journal of The Royal Meteorological Society
27 November 1990

AJD/27/11/90

1. INTRODUCTION

The partitioning of energy into latent and sensible heat at the earth's surface is of great importance in the general circulation of the atmosphere and the regional climate. Previous developments in modelling the land surface atmosphere interaction have shown that for vegetation with near complete canopy cover, the canopy may be represented as a single "big-leaf" with the sources of sensible and latent heat located at the same level $d+z_0$, where d is the zero plane displacement height and z_0 the roughness length (see Monteith, 1981). When the canopy cover is incomplete, the big leaf approach tends to be less appropriate as the exposed soil becomes a significant additional source of heat and water vapour.

Recently a number of models have been described (Shuttleworth and Wallace, 1985, Choudhury and Monteith, 1988) which attempt to model the partitioning of energy for soil-vegetation systems with more than one effective source. Such models have to describe the transport of heat and water vapour from the soil, through the canopy, to the reference level above the canopy. Traditionally the transfer of a scalar quantity, c within a canopy has been parameterized with local gradient diffusion (K-theory):

$$F(z) = - K(z)\ \delta c/\delta z \qquad (1)$$

where K(z) is the diffusion coefficient and F(z) the scalar flux.

A number of studies have emerged during the last decade questioning the validity of K-theory for within canopy transfer (e.g. Finnigan and Raupach, 1987) . The observation of countergradient fluxes (Denmead and Bradley, 1985), implying negative K-values is the most dramatic example of the failure of equation 1. Generally, K-theory requires the characteristic length scale of the dominant eddies to be small compared with the distance over which the gradient changes appreciably. This is violated within most plant canopies where the length scale of the turbulence is of the same order as the canopy height. Although these objections have been known for some time, until recently the lack of a practical alternative has led most

2. THEORY

(a) Nomenclature

a	Constant used in the calculation of boundary layer conductance (a=0.01)
$a_{i,j}$	I, jth element of coeficient matrix
A	Coefficient matrix
A, A_s	Total available energy flux leaving the complete crop, the soil substrate, as sensible and latent heat per unit ground area (W m^{-2})
b_i	Ith element of column vector
B	Column vector
C_d	Mean drag coefficient of the leaves (dimensionless)
C_r, C_i	Concentration of scalar quantity at reference height and in the ith layer
C_{ni}, C_{fi}	Near field and far field concentration of scalar quantity in the ith layer
c_p	Specific heat of air at constant pressure (J kg^{-1} K^{-1})
d	Zero plane displacement (m)
D_r, D_i	Vapour pressure deficit at reference height and in the ith layer (mbar)
e_r, e_i	Vapour pressure at reference height and in the ith layer (mbar)
f_c	Fraction of net radiation absorbed by the canopy (dimensionless)
F_i, F_s	Scalar flux emanating from the ith layer and soil substrate
G	Soil heat flux (W m^{-2})
g_b	Leaf boundary layer conductance (m s^{-1})
g_s	Leaf stomatal conductance (m s^{-1})
h	Crop height (m)
H_i	Sensible heat flux from the ith layer (W m^{-2})
J	Flux variable H- (Δ/γ) λE (W m^{-1})
k	Von Kármán's constant (dimensionless)
K	Eddy diffusion coefficient (m^2 s^{-1})

z_0'	Roughness length of the soil substrate (m)
α, α'	Extinction coefficient for net radiation and windspeed (dimensionless)
β	$\gamma/(\Delta+\gamma)$ (dimensionless)
Δ	Rate of change of saturated vapour pressure with temperature (mbar $°K^{-1}$)
γ	Psychrometric constant (mbar $°K^{-1}$)
λE_c, λE_s, λE_i	Latent heat flux from the complete crop, soil, and the i^{th} layer (W m^{-1})
ρ	Density of air (kg m^{-3})
σ_w	Variance of the vertical wind velocity (m^2 s^{-2})

the variance of the vertical wind velocity. C_{nr} is the near field concentration at the reference height. The scalar flux is given as

$$F = \sum_{j=1}^{m} S_j \; z \qquad (4)$$

where S_j is the source density at layer j. The appearance of the near field concentration in equation (3) is a result of the boundary condition at the reference height

$$C_r = C_{nr} + C_{fr} \qquad (5)$$

In the calculation of the near field contribution it is assumed that the turbulence is locally homogeneous. It can then be shown (Raupach, 1989a) that the near field concentration can be described as the convolution of the source density distribution with a 'near field kernel' function:

$$C_{ni} = \sum_{j=1}^{m} (S_j/\sigma_{wj}) \left| Kn \; \{(z-z')/(\sigma_{wj} \; t_{lj})\} + Kn \; \{(z+z')/(\sigma_{wj} t_{lj})\} \right| \; z' \qquad (6)$$

where z' is the source height and $Kn(\varepsilon)$ the near field kernel function derived by Raupach (1989a) for $\varepsilon > 0$, as

$$Kn(\varepsilon) = Kn(-\varepsilon) = -0.3984 \; \ln \; \{1-e^{-\varepsilon}\} - 0.15623 \; e^{-\varepsilon} \qquad (7)$$

It can be seen from equation (6) that the near field concentration at any level involves contributions from all sources in the canopy. However, due to the highly peaked nature of the near field kernel function only those contributions for which $|z-z'| \gg \sigma_w t_l$ contribute substantially to the near field effect.

where ϱ is the density of air, c_p the specific heat of air, γ the psychrometric constant, T the temperature, e the vapour pressure, and r_b and r_s the boundary layer and stomatal resistance of layer i. At each level the energy balance equation applies

$$R_{ni} = H_i + \lambda E_i \tag{11}$$

where R_{ni} is the available energy in the form of net radiation (at the lowest level the available energy is given by R_{n0} - G, where G is the soil heat flux). Introducing Δ , the rate of change of the saturated vapour pressure with temperature yields two equations, similair to the Penman-Monteith equation (Monteith, 1981, Monteith, 1973), for latent and sensible heat with the new variable D, the vapour pressure deficit:

$$H_i = \{ \gamma\, r_v\, R_{ni} - \varrho c_p\, D_i \} / \{ \Delta r_b + \gamma r_v \} \tag{12a}$$

$$\lambda E_i = \{ \Delta\, r_b\, R_{ni} + \varrho c_p\, D_i \} / \{ \Delta r_b + \gamma r_v \} \tag{12b}$$

where $r_v = r_b + r_s$. From the definition of Δ it follows that

$$D_i - D_r = \Delta \{ T_i - T_r \} - \{ e_i - e_r \} \tag{13}$$

Referring to equation 8 it can be shown that

$$D_i - D_r = \{\Delta/(\varrho c_p)\} \sum_{j=1}^{m} M_{ij}\, J_j\, z \tag{14}$$

where the new flux variable J can be obtained from substitution of equation (14) into (12) and follows as

$$J_i = H_i - (\Delta/\gamma)\, \lambda E_i \tag{15}$$

This is the same flux variable as described by Chen (1984) and McNaughton (1976). Substitution of

This system can be solved with standard matrix procedures for systems of linear equations. Substitution of D_i into equation (12) yields the source density distribution. Temperature and vapour pressure distributions can be found from the equivalent versions of equation (14) for temperature and sensible heat source distributions and vapour pressure and latent heat source distributions respectively. The general theory of the multi-layer evaporation model with a Lagrangian within canopy transfer theory is now complete.

(d) Auxiliary equations

The linear system as specified in equations (17) and (18a,b) needs information about the distribution of available energy in the canopy, boundary layer and stomatal resistances and, above all the turbulence profiles are needed in the Lagrangian submodel σ_w^2 and t_l.

Analysis of turbulence profiles for several plant canopies led Raupach (1989b) to suggest that in the absence of measurements, the profiles of the vertical velocity variance and the Lagrangian time scale may be approximated from

$$\sigma_w(z) = u_* \, 1.25 \qquad z/h > 1 \qquad (19a)$$

$$\sigma_w(z) = u_* \{ 0.25 + z/h \} \qquad 0 < z/h < 1 \qquad (19b)$$

$$t_l(z) = (u_* / h) \{ \max [0.3, k (z - d) / 1.25 h] \} \qquad (20)$$

where u_* is the friction velocity. The boundary layer resistances are calculated according to Choudhury and Monteith (1988). First the windspeed has to be extrapolated downwards from a reference height, z_r to the top of the canopy:

$$u(h) = u_r \{ \ln (h - d) / z_0 \} / \{ \ln (z_r - d) / z_0 \} \qquad (21)$$

where R_n is net radiation measured above the canopy, α the net radiation attenuation coefficient for millet measured as α=0.41 (Wallace et al., 1990), and L′ the leaf area between the top of the canopy and z.

(e) Single and dual source models based on K-theory diffusion

The concept of K-theory has been extensively used in physically based evaporation models (Penman, 1948, Monteith, 1981). The utility of these models is demonstrated by their increasingly wide adoption in the fields of meteorology, hydrology and agriculture. As the most simple model for evaporation from sparse canopies a modified Penman-Monteith equation suggested by Wallace et al (1990) is considered

$$\lambda E_c = \{ \Delta f_c R_n + \varrho c_p D_r / r_a\} / \{ \Delta + \gamma(1+r_c^s/r_a\} \qquad (26)$$

where D_r is the vapour pressure deficit at reference height and r_c^s a bulk stomatal resistance of the canopy. The modification consists of the use of the net radiation absorbed by the canopy, given as a fraction, f_c of the net radiation at reference height. From equation (25), f_c can be expressed as

$$f_c = 1 - \exp(\alpha L) \qquad (27)$$

The aerodynamic resistance, r_a is calculated from crop height, h, and windspeed at the reference height, u_r according to

$$r_a = [\ln^2\{(z_r-d)/z_0\} + \ln(z_0/z_m)\ln\{(z_r-d)/z_0\}] \{k^2 u_r\}^{-1} \qquad (28)$$

The second term takes into account the difference in transfer of momentum and heat and water vapour. The natural logarithm of the ratio of the roughness lengths for water vapour and momentum is taken as typical for field crops. The values of aerodynamic resistances between the substrate and canopy source height and canopy source height to reference level are calculated following Shuttleworth and Wallace (1985). They assumed that the values of z_0 and d for a fully developed canopy are given by the Monteith (1973) relationships. then, implicit in the way r_a^s and r_a^a vary between a closed canopy and bare soil substrate is a corresponding variation in the values of d and z_0.

The third K-theory model used in this paper is the updated version of the Shuttleworth and Wallace model as described by Shuttleworth and Gurney (1990). This follows Choudhury and Monteith's (1988) description of aerodynamic transfer by replacing the simple crop-height relationships of d and z_0 with more sophisticated relationships based on second order closure modelling (e.g. Shaw and Pereira, 1982). These relationships are given in this paper as equations 22, 22a and 22b. They show how they effective sink height for momentum varies with leaf area. It is important to note that in the both the Shuttleworth-Gurney approach, as in the Shuttleworth-Wallace approach, the source height of water vapour and temperature remains fixed at $d+z_0$, calculated according to the relations of Monteith (1973) as previously given. They only vary with height of the crop. The resulting values of the resistances are calculated using equations 45 and 46 of Shuttleworth and Gurney (1990). Also, rather than using a constant value of boundary layer resistance, as in the Shuttleworth-Wallace model, a boundary layer resistance which varies with in canopy windspeed is used (see equations 23 and 24, this paper).

is compared with a simpler K-theory model.

In Figure 1a the predicted source density profiles of latent heat are shown for three different times on 13 August 1986. In the morning the soil latent heat flux is a major contribution in the system, however during the day the canopy contribution increases. Figure 1b shows the vapour pressure deficit profile which exhibits little variation with height. The vapour pressure deficit is low in the morning as a consequence of the relatively low temperature (Figure 1c). During the course of the day the temperature and vapour pressure deficit increase. The shape of the two concentration profiles agrees qualitatively with published measurements, although the variation with height is much less than previously observed in a similar crop in Australia (Begg et al, 1964). However, in the present case the leaf area index was only 1.1 and in the Australian study a value of 8.0 is quoted. The low leaf area apparently enhances the turbulent diffusion in the canopy, resulting in quite small gradients of the scalar quantities in the canopy. The Lagrangian model however, seems to be capable of realistically simulating source density profiles and profiles of temperature and vapour pressure deficit.

It is of important to assess the sensitivity of the Lagrangian model to changes in its basic input parameters. In Table 2 the results of a sensitivity analysis are shown for a number of key parameters. The sensitivity of the model to a parameter is expressed as the percentage change in total evaporation over a day as a result of a change of ±25% in that parameter. Generally the sensitivity is low exept for the specification of stomatal conductance, where a 25% change propagates as a 19% change into the predicted transpiration. Sensitivity to the turbulence profiles as given in equations (19) and (20) is low. The sensitivity to a change of 25% in the σ_w profile is 6.2 % and is the highest sensitivity of all the other parameters. As this parameter helps to specify the turbulent diffusion in the canopy it suggests that improvement in the aerodynamic submodel may come from a better understanding of σ_w in this type of sparse canopies.

difference in predicted transpiration was greatest. The humidity deficit ratio is the space averaged vapour pressure deficit within the canopy, divided by the value at reference height. It may be noted that the main cause of the difference in the output between the P-M and S-W models is also due to the differences in the treatment of the within canopy deficit in these models (Wallace et al, 1990).

The prediction of a higher in-canopy deficit in the Lagrangian model, however, is not a result of the inclusion of the near field effect, but depends primarily on the specification of the far field diffusion coefficients . For instance, setting the near field contribution to zero and using the model with only a far field description of turbulent diffusion, did not change the predicted transpiration appreciably. The relatively small contribution of the near field effect in these conditions is caused by the low source density profile of crops with low leaf area indices and low (net) radiative absorption.

5. CONCLUDING REMARKS

The Lagrangian model developed in the current paper appeared to be capable of predicting realistic profiles of temperature, vapour pressure deficit and latent and sensible heat flux for a sparse crop of millet in Niger, West Africa. Of the three K-theory based models, the Penman-Monteith "big-leaf" model, predicted lower transpiration than the two models which explicitily take the soil contribution into account (the Shuttleworth-Wallace and Shuttleworth-Gurney models). When a large evaporative flux emanates from the soil, the difference in transpiration between the three models is small. This is because the large soil evaporation flux dominates the flow of water vapour through the canopy and the resulting in-canopy water vapour deficits do not differ markedly between the models (and the value at reference height). Conversely, when there is a large heat flux from the soil, the in-canopy deficits differ significantly between the models and hence the models predict different tranpiration rates.

The Shuttleworth-Gurney model which incorporates a leaf area and height dependent location of the canopy sink of momentum (the values of d and z_0), generally predicted up to 10% more transpiration than the Shuttleworth-Wallace model in which the d and z_0 vary implicitly with the calculated values of with an improved understanding of the aerodynamics of this type of canopy. In developping sparse crop evaporation models further, the measurement of an within-canopy deficit and temperature may prove to be a prime requirement for future experiments with this type of sparse crop.

DATE	LAI	CROP HEIGHT	SOIL EVAPORATION
17 September 1985	0.26	2.3 m	2.5 mm
2 July 1986	0.23	0.6 m	0.3 mm
13 August 1986	1.10	2.8 m	0.6 mm

Table 1. Characteristics of the soil and crop

AJD/27/11/90

DATE	*P-M*	*S-W*	*S-G*	*Lag*
17 Sept 1985	0.49	0.47	0.49	0.45
2 July 1986	0.80	0.91	1.02	1.16
13 August 1986	1.84	2.06	2.16	2.47

Table 3. Predicted daily total transpiration (mm) by the Penman-Monteith (P-M), Shuttleworth-Wallace (S-W), Shuttleworth-Gurney (S-G) and Lagrangian model (Lag)

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AJD/27/11/90

FIGURE 1A

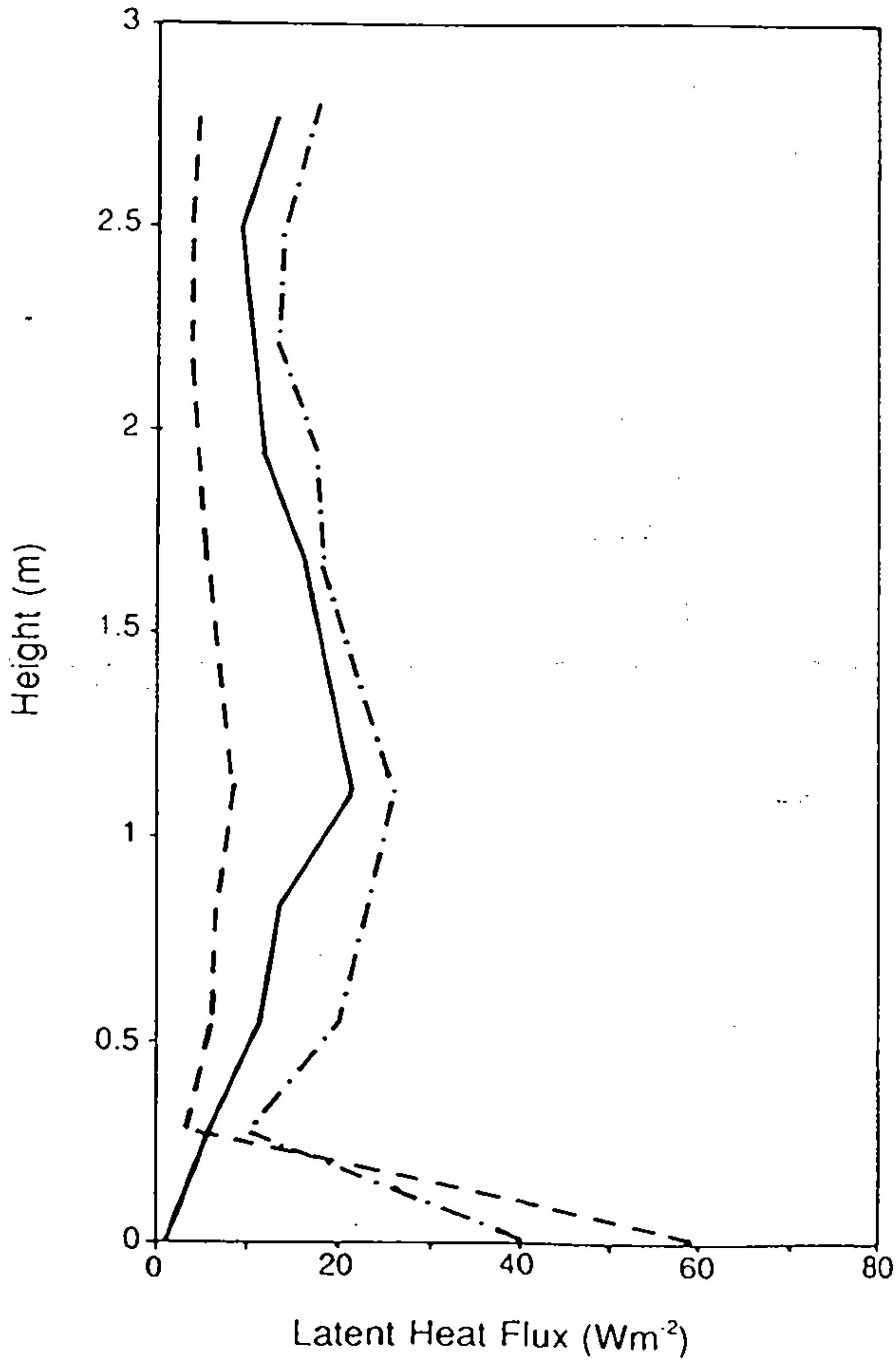


FIGURE 1C

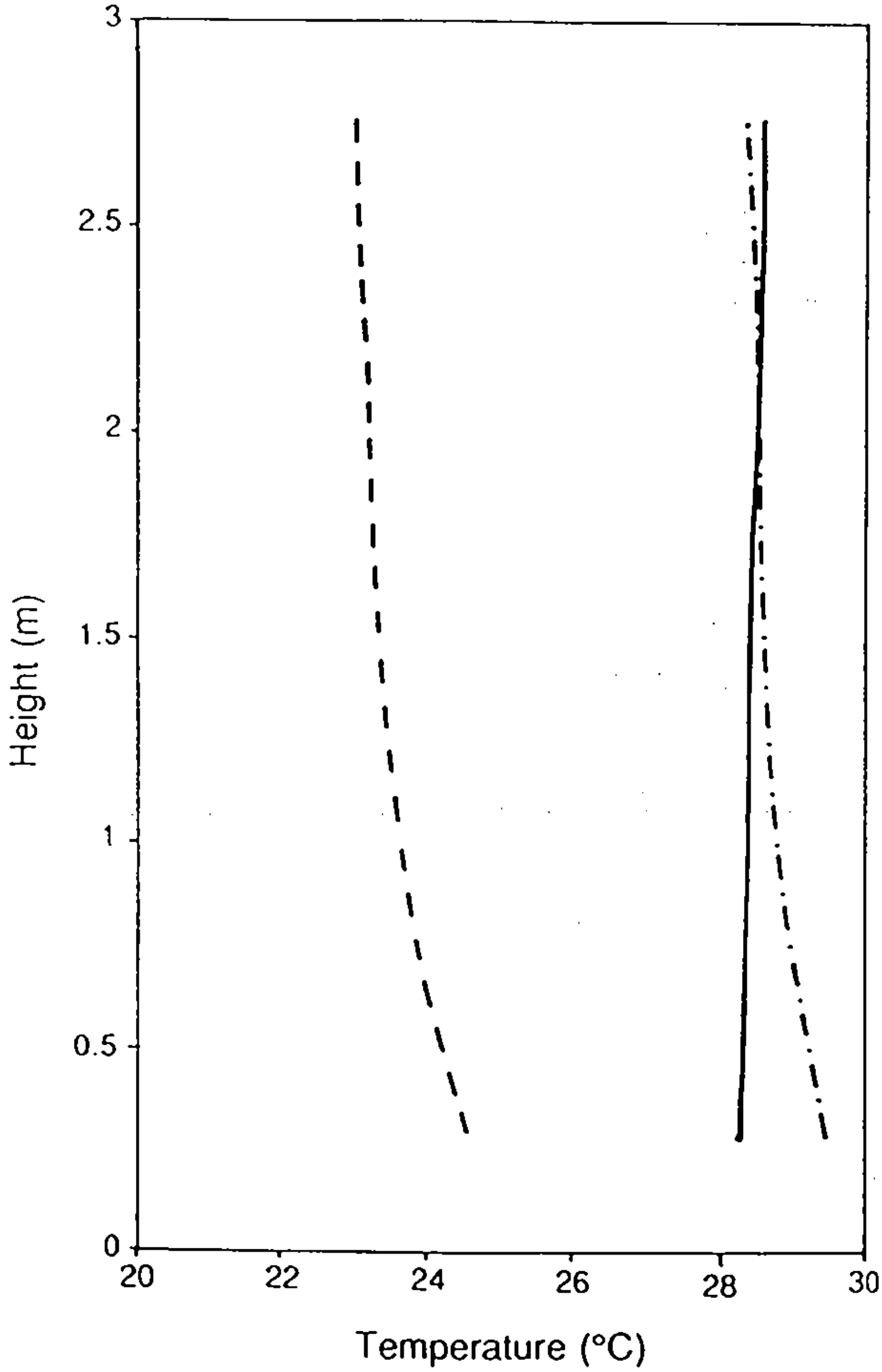


FIGURE 2B

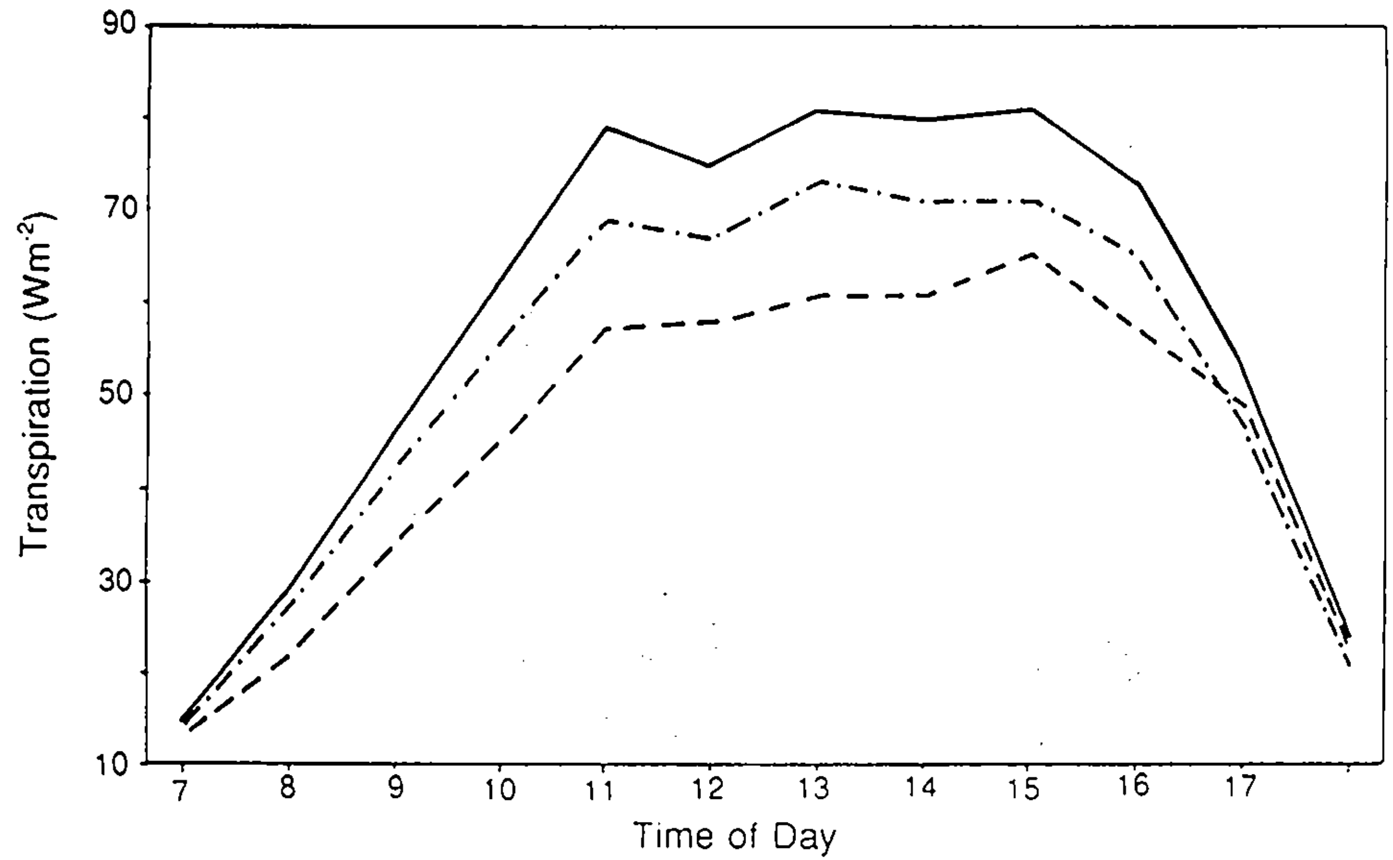

Transpiration (Wm-2)
90
70
50
30
10
7
8
9
10
11
12
13
14
15
16
17
Time of Day

FIGURE 3A

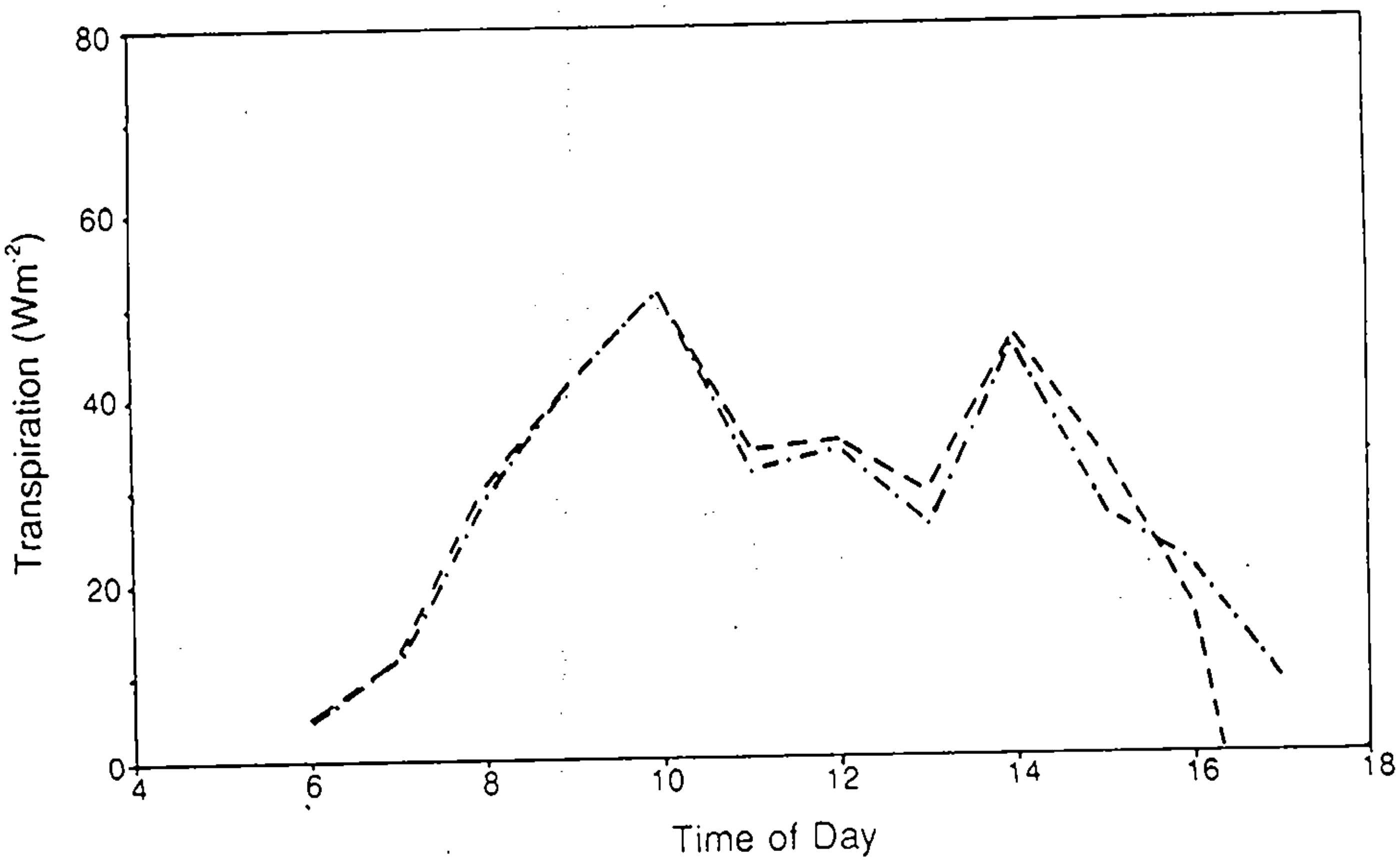
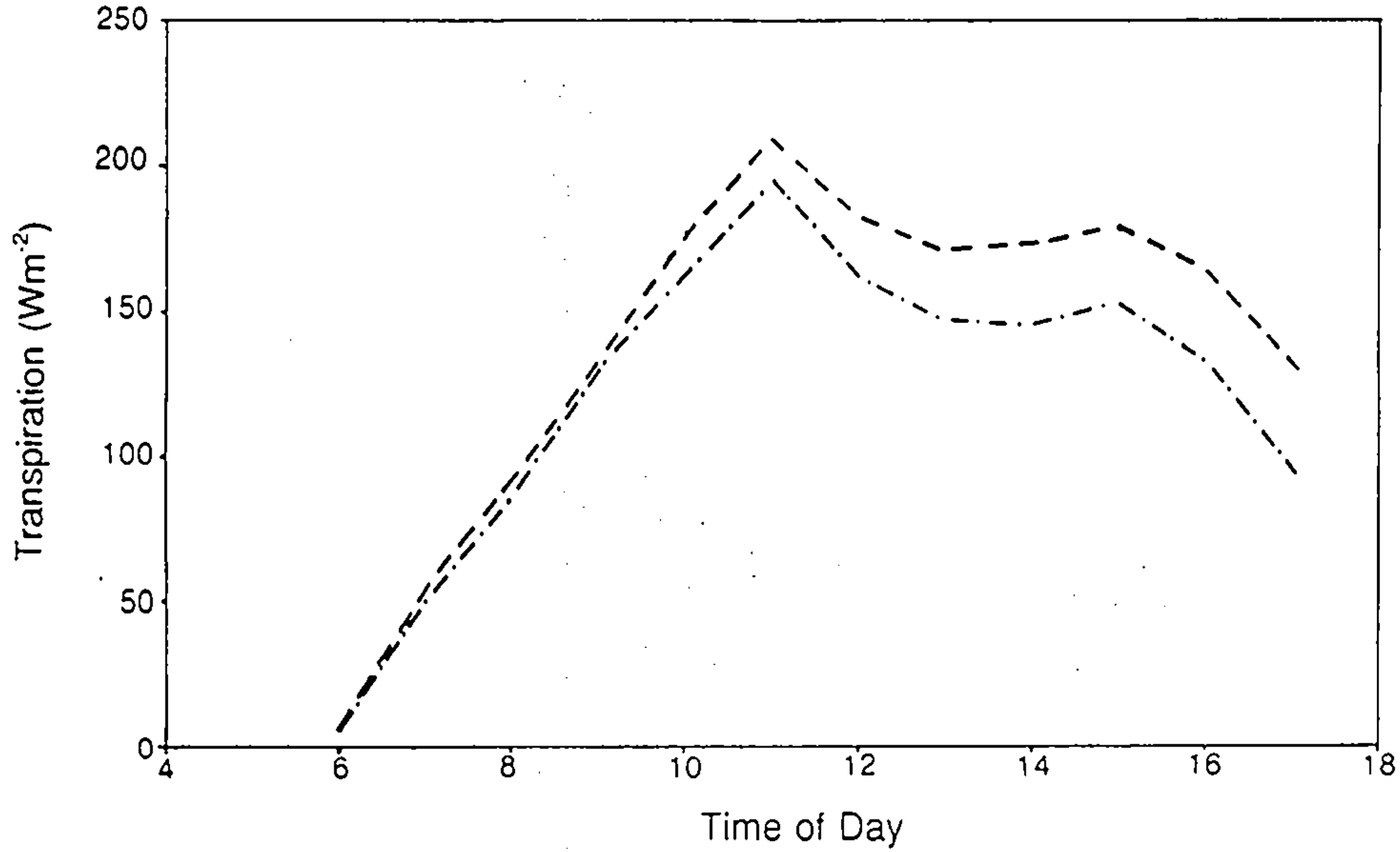


FIGURE 3C



250
200
150
100
50
0
4
6
8
10
12
14
16
18
Transpiration (Wm^{-2})
Time of Day

APPENDIX D

SBX-EVAP.QJ4 28-6-90

MEASUREMENTS OF EVAPORATION FROM FALLOW SAHELIAN SAVANNAH AT THE START OF THE DRY SEASON

by

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erosion. Charney (1975) proposed a mechanism whereby the higher albedo of bare soil could produce an atmospheric feedback process which resulted in lower rainfall. General Circulation Models (GCMs) have modelled this effect (see Rowntree, 1988). Less attention has been paid to modelling the effects of changes in the other surface energy fluxes which might occur as desertification proceeds. This is partly because it is only recently that the land surface sub-models of GCMs have become sufficiently realistic to allow the incorporation of the evaporative characteristics of different vegetation types, but also because there are little, if any, data against which such models could be calibrated. This paper describes data which have been collected as part of a study to meet this need. The measurements described here were made in a fenced area of fallow bushland with a good cover of vegetation. In any investigation into the climatic effects of desertification they could be used as the limiting case of vegetation in good condition.

2. SITE

The measurements were made between 28 September and 10 November 1988 in an area of fallow bushland on the International Crops Research Institute for the Semi-Arid Tropics (ICRISAT) experimental farm at Sadoré, 45 km south of Niamey, Niger (Lat. 13^{o} 15′N, Long. 2^{o} 17′E). Some preliminary measurements at this site have been described by Wallace *et al.* (1990). The area is flat and the site was situated so that there was a fetch of at least 300 to 500 m of bushland in all directions. Further details of micrometeorological aspects of the site are given by Lloyd *et al.* (1990). The observations were made at the start of the dry season in 1988 when the surrounding area was mainly millet stubble or grazing land. There was a windbreak of Neem (*Azadirachta indica*) trees approximately 10 m high on the eastern boundary of on top of a 10 m high tower.

A second Hydra was mounted at the top of a variable height, pneumatic mast. This Hydra was operated at 3 m, 6 m and also at 9 m. There was also a short period when the two instruments were intercompared on the tower at 12.3 m.

An automatic weather station was mounted at the top of the same tower as the eddy correlation mast. This measured solar radiation (Kipp and Zonen, Delft, The Netherlands), aspirated wet and dry bulb temperature (Campbell Scientific Instruments, Sutton Bonington, UK), and wind speed and direction (Didcot Instrument Co., Abingdon, UK). These instruments were logged on a Campbell Scientific CR10 solid state data logger. For the purposes of the analyses presented in this paper all the measurements were processed to provide hourly average values. Other measurements not directly relevant to this paper included net radiation, reflected solar radiation, soil temperature and surface temperature.

Soil moisture content was derived from 5 gravimetric samples taken along a 100 m transect some 25 m to the south of the observation tower. The samples were taken to a depth of 500 mm. Measurements were made once after the only substantial rainstorm during the observation period, and otherwise at weekly intervals. The transect was moved 1 m to the south for each new measurement to avoid any areas where the vegetation had been damaged by the previous sampling.

The climatological data were taken at the standard agrometeorological station situated about 1 km from the experimental site. Readings of temperature,

kg^{-1} at the end.

Rainfall during the measurement period is shown in Figure (3); the only substantial rainfall occurred on 2 September. The decline in the soil moisture content of the top 500 mm of soil, expressed as percentage of the maximum value observed during the study, is also shown in Figure (3).

Figures (2) and (3) show that the evaporation remained close to potential for the first ten days following the end of the wet season, but then declined. By the end of the measurement period, ie after six weeks without rain, the evaporation was less than 30 per cent of potential. While this was partly a result of the increase in potential evaporation in response to the greater humidity deficits encountered as the dry season progressed, there was also a decline in evaporation in absolute terms from approximately 4.5 to 1.5 mm per day. This is illustrated in Figure (4) which shows the daily trend in evaporation and sensible heat flux for two example days, one near the start of the measurement period and one near the end. The soil moisture in the top 500 mm declined for the first two weeks of the dry season but then remained virtually constant, implying that during the last four weeks of measurements the evaporated water must have been extracted from below 500 mm. Assuming an evaporation rate of 5 mm per day for the three days of missing data starting on 29 September, over the whole period of observations 135 mm of water was observed to be evaporated, of which 78 mm came from the soil below 500 mm during the last four weeks of measurements.

Hours with negative u_*^2, which can occur at low windspeeds when the cup anemometer is within its stalling region, were also omitted. Between 3 and 5 October the fast-response thermometer did not work, there were no heat flux measurements, and therefore the surface conductance could not be calculated. This gave a total of 266 hours available for analysis.

Figure (5) shows the individual surface conductance data points for each day throughout the observation period. The few points obtained on the days immediately following the rainfall at the end of September, ie 1 to 3 October, show conductances in the range 11 to 22 mm s^{-1}, but by 6 October (day number 280) the conductances have fallen rapidly to between 4 and 7 mm s^{-1}. Figure (6) shows the trends in surface conductance for the two days at the start and end of the measurement period shown in Figure (4). On both days after declining for the first few hours the conductance changes relatively little for the remainder of the day. A similar trend in surface conductance was observed by Kim and Verma (1990) for drought stressed prairie grass. The decline in surface conductance as the dry season progresses is the dominant feature of the data apparent in Figures (5) and (6); by the end of the measurement period conductances had fallen from values of about 4 mm s^{-1} to between 1 and 2 mm s^{-1}.

(d) Surface conductance: modelling

The Penman-Monteith equation assumes a one-dimensional transfer of water vapour from a hypothetical single source. Savannah bushland is a natural vegetation with complex three-dimensional structure. It has at least two gives r^2 = 0.65. The parameters for this equation are again given in Table 1. The result of evaluating Equation (4) is also shown in Figure (6), for the two example days at the start and end of the measurement period.

From a plant physiological point of view a more realistic model of surface conductance should also take account of its variation in response to environmental conditions. As well as green leaf area index and soil moisture, solar radiation and atmospheric humidity deficit are normally seen to be correlated with plants' stomatal control. However in the present case, with the exception of the long term decline in surface conductance with elapsed time (and thus with green leaf area and soil moisture), the range of variation is narrow. At this latitude the solar radiation rises and falls rapidly at the start and end of each day, and there are almost no data for solar radiation less than 300 W m^{-2}, the range normally observed for a response to radiation. As a consequence there is no obvious response of the surface conductance to solar radiation, and the stomata appear to be light-saturated for all the data collected. Dividing the data into three periods to reduce the influence of the long term decline in surface conductance, it is possible to derive a relationship between surface conductance and humidity deficit, but the correlation is poor. For example a regression between surface conductance, in mm s^{-1}, and specific humidity deficit, in g kg^{-1}, for the middle period between 16 and 25 October inclusive, gave the relationship

$$g_s = 3.86\ (+/-\ 1.11) - 0.054\ (+/-\ 0.019)\ D' \qquad (5)$$

with r^2 = 0.10.

shown in Figure (2). The roughness length, z_0, was taken as 0.14 m, the value derived for this fallow bushland site by Lloyd *et al.* (1990). The values of $g_{s,d}$ necessary to give the observed daily evaporation are shown in Figure (7). The values are less than the average of those derived for each hour, as in this model the evaporation is spread over 24 hours. The variation in $g_{s,d}$ can also be described well by a function equivalent to Equation (3)

$$g_{s,d} = a_l e^{b(N - 280)} \qquad (8)$$

This is also shown on Figure (7), with the values of the parameters being given in Table (1).

5. CONCLUDING REMARKS

The results presented here are intended to provide some preliminary data on the rates of evaporation and the consequent surface conductances which might be expected from this type of Sahelian savannah vegetation at the start of the dry season. The data demonstrate that the evaporation remains close to potential for ten days after the last rainfall. After that when some 50 mm of water have been evaporated, the rate of evaporation falls, but even after five weeks, when some 135 mm have been evaporated there is still appreciable evaporation with rates of 1 to 2 mm per day.

It should be recognised that the surface conductances calculated in this paper are bulk values which represent the restriction to evaporation from the surface as a whole. Indeed Shuttleworth and Gurney (1990) have shown that for canopies with low leaf areas, surface conductances such as these will be a poor representation of the physiological behaviour of the canopy. To

TABLE 1

Equation	a_1	a_2	a_3	b	r^2
3	4.75			−.041	.57
4	4.94	−.30	−.023	−.041	.65
8	4.57			−.045	.85

The coefficients derived for Equations (3), (4) and (8), together with their correlation coefficients.

with the Thom and Oliver aerodynamic function and the data from the Sadoré agrometeorological station. The curve is the result of fitting Equation (8) to the data from 6 October 1988 (day 280) onwards.

Department/Tropsoils, Texas A and M University.

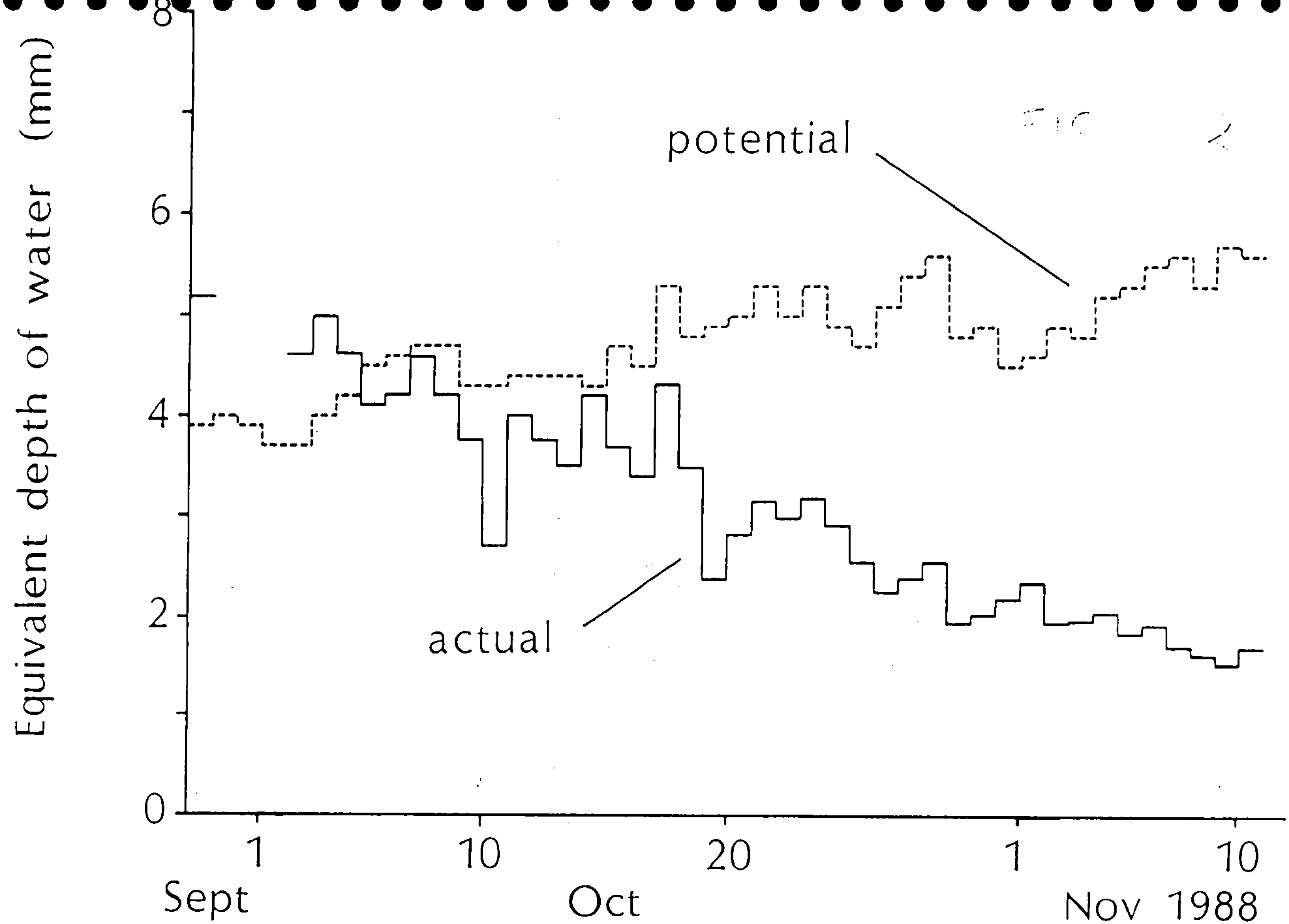
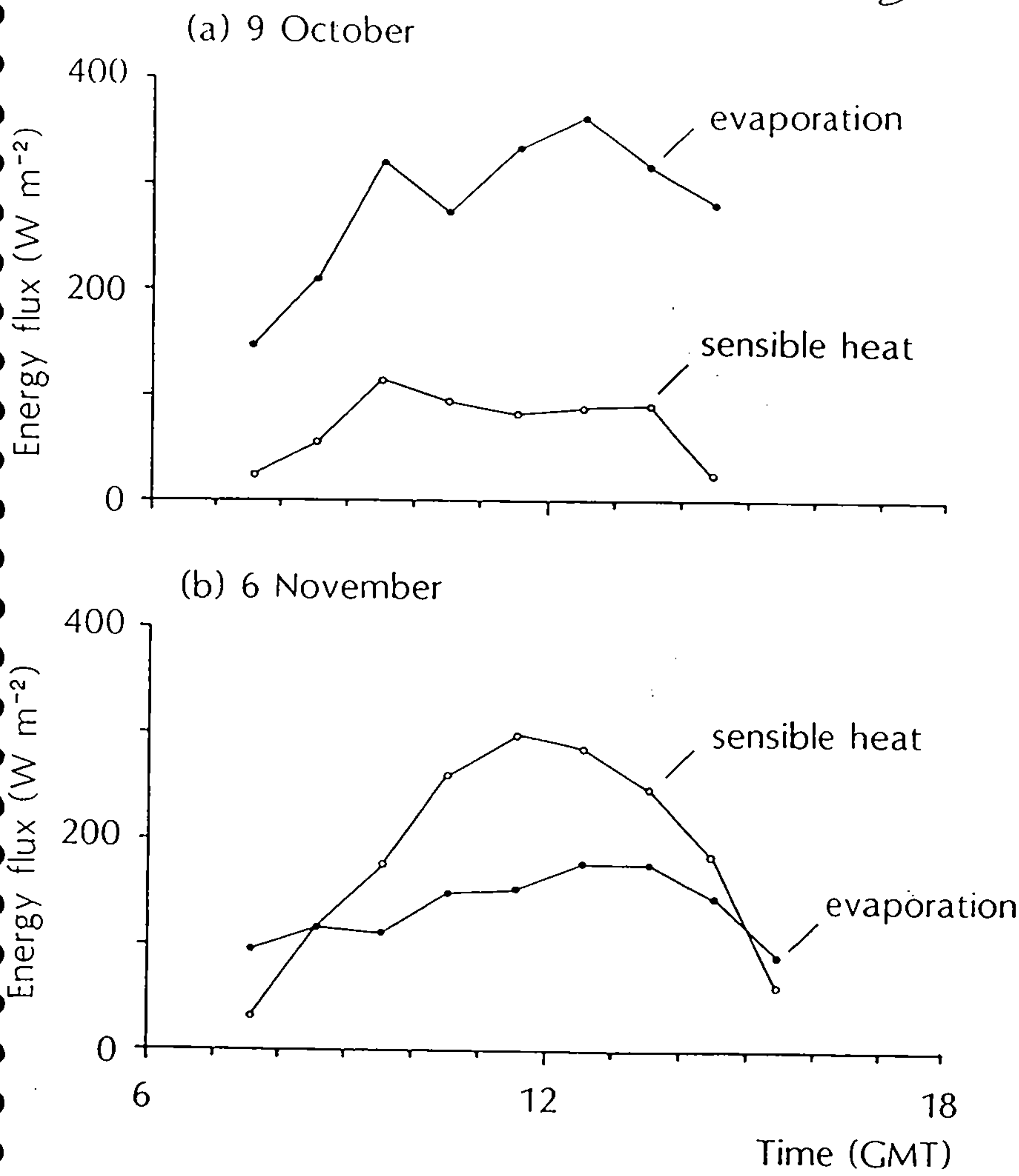

Equivalent depth of water (mm)
0
2
4
6
8
potential
actual
1
10
20
1
10
Sept
Oct
Nov 1988

Fig 4


(a) 9 October
400
200
0
Energy flux (W m^{-2})
evaporation
sensible heat
(b) 6 November
400
200
0
Energy flux (W m^{-2})
sensible heat
evaporation
6
12
18
Time (GMT)

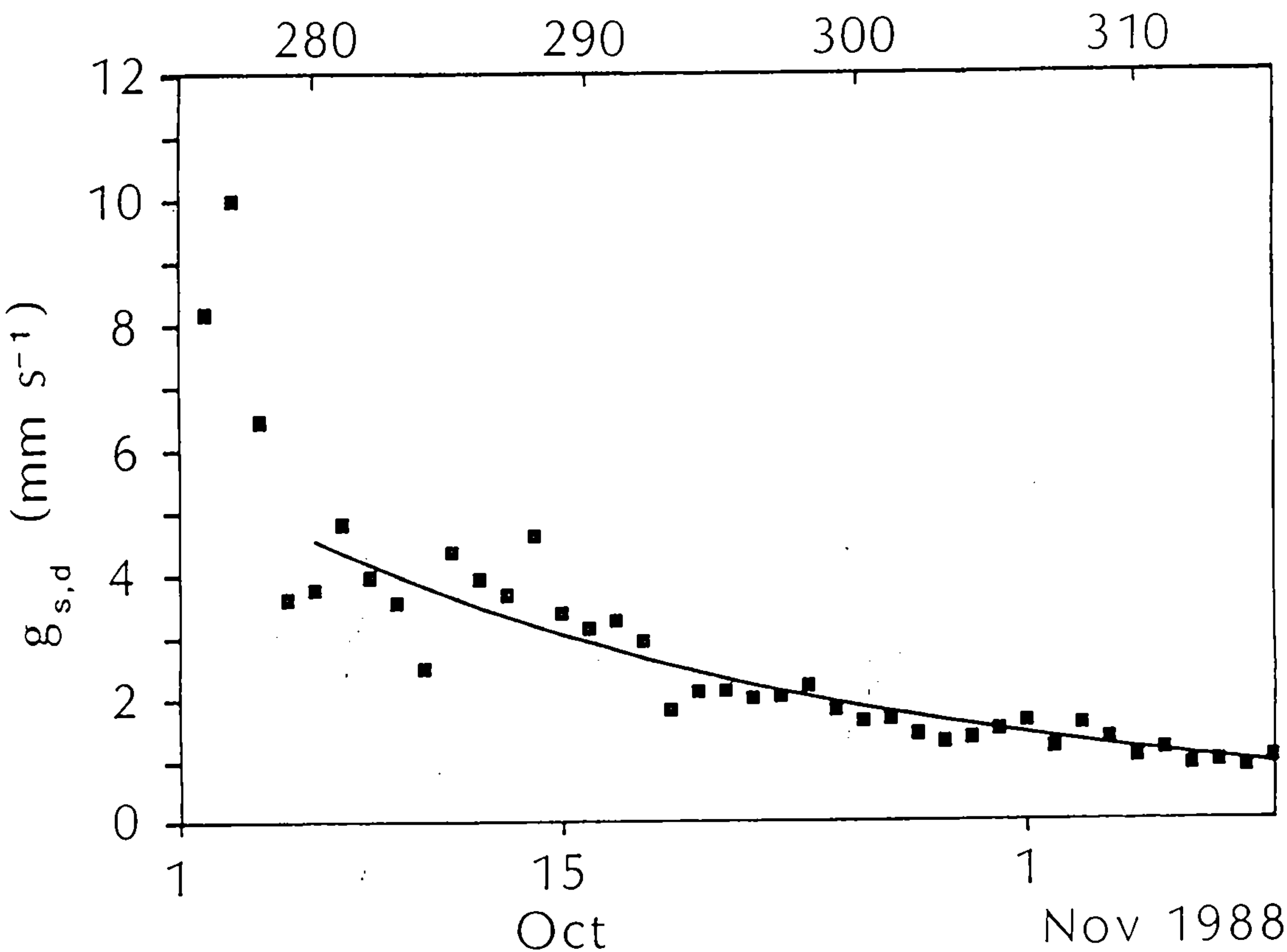


Fig 7

APPENDIX E

ESTIMATES OF SENSIBLE HEAT FLUX FROM OBSERVATIONS OF TEMPERATURE FLUCTUATIONS

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2nd Draft, to be submitted to

Boundary Layer Meteorology

18-12-1990

1. INTRODUCTION

Progress in the understanding of land surface-atmosphere interactions over the last decade has greatly benefited from an increased ability to measure the fluxes of sensible and latent heat over a variety of land surfaces. Among the techniques employed, the eddy correlation technique is increasingly favoured by experimenters as it constitutes a direct measurement of the fluxes. Eddy correlation measurements are, however, prone to a number of errors (see Moore, 1986) and routine application of the eddy correlation technique still requires a considerable investment in equipment and manpower.

In recent years emphasis has shifted from small scale micrometeorological studies over homogeneous vegetation to the study of large scale, area-average fluxes over inhomogeneous vegetation (eg André *et al.*, 1988). These studies require the installation of a substantial number of field sites over a large area, and are usually complemented with a measurement of fluxes from aircraft to study their aggregation. A measurement technique which is simple and inexpensive, and can be used both for surface and airborne measurements, could prove to be very valuable in these experiments. One promising technique is the temperature fluctuation method. This method allows the sensible heat flux to be estimated from only the standard deviation of the temperature fluctuations. Previous work has highlighted the potential of this technique and good results were obtained for a number of land surface types (Tillman, 1972; de Bruin, 1982, Weaver, 1990). In this paper we compare the sensible heat flux estimated from the standard deviation of temperature fluctuations with measurements of the sensible heat flux made by eddy correlation instrumentation. Results will be shown for four surfaces in Niger, West Africa: bare soil, millet, fallow savannah and tiger bush (degraded forest).

A further advantage of the temperature fluctuation method is that the only measurement required is a scalar. Accurate levelling of the instrumentation is therefore unnecessary, making the method particularly suitable for application on a platform replace z by z_m = z-d, where d is the zero plane displacement height.

Following Tillman (1972) it is possible by rearranging the above equations to obtain the following explicit relations for the sensible heat flux

$$\overline{w'\theta'} = \{(\sigma_\theta / C_1)^3 \quad (kgz_m / \theta) \quad (C_2 - z_m/L)/(-z_m/L)\}^{1/2} \tag{5}$$

$$\overline{w'\theta'} = \{(\sigma_\theta / C_1)^3 \quad (kgz_m / \theta)\}^{1/2} \tag{6}$$

where k is von Kármán's constant and g is the acceleration due to gravity. Equation (5) covers the whole of the unstable region and Equation (6) applies in the limiting case of free convection. Tillman, basing his analysis on work by Wyngaard *et al.* (1971), obtained the values C_1=0.95, C_3=2.5 and $C_2=-(C_1/C_3)^3$=0.05 for the constants in Equations (5) and (6). In the analysis in this paper we use Equation (6), the free convection approximation of Equation (5), because it does not need information on u_* and stability. Furthermore, the data presented by Tillman suggest that it is a very good approximation to Equation (5).

Equation (6) shows that the estimate of sensible heat flux is sensitive to the value of d. Figure 1 shows the sensitivity of the estimated flux to the error in height above the zero plane displacement. The sensitivity of the esimated flux to d is considerable. Clearly the method is best employed at heights which are large compared to d, and in the case of a balloon mounted sensor care must been taken to ensure that the height of the sensor is accurately established.

Details of the measurement period , the height of the vegetation, and instrumentation, and the measured or estimated displacement height and roughness length are given in Table 1.

4. Instrumentation

Measurements of heat flux were made with the Institute of Hydrology Mk 2 Hydra eddy correlation device (Shuttleworth *et al.*, 1988). This includes a sonic anemometer for measuring vertical windspeed and a fast response thermocouple for measuring temperature fluctuations. The sensible heat flux and the variance of the temperature fluctuations are calculated in real time.

These instruments were deployed over each of the sites: over the bare soil at a height of 3.3 m, over the millet at 4.15 m, over the fallow savannah at 12.3 m and over the tiger bush at 15.9 m.

In addition, on two days a balloon (AIR Inc, Boulder, USA) was tethered at a height of approximately 25 m, above the tiger bush site. Suspended below this balloon was a fast response thermocouple of the type used in the Hydra, with only those parts of the Hydra electronic circuitry necessary to record the standard deviation of the temperature fluctuations.

5. Results

In Figure 2 positive values of sensible heat flux for the bare soil site, as estimated from Equation (6), are compared with the fluxes measured by the Hydra. The agreement is good. Table 2 shows the result of a regression of the estimated flux on the measured flux, the regression has been forced through the origin. The correlation in d, together with the proximity of the Hydra to the surface of the crop, is probably the major cause of the relatively poor results for the millet crop.

Recently Weaver (1990) suggested that the dimensionless stability function may be different over different surfaces and may not be a universal function. Our experimental results do not support this view, as good agreement between observed and estimated heat flux was found for a range of very different surfaces. Furthermore, the shape of the dimensionless function $f(z/L)$ is in close agreement with our measurements. Figure 7 shows the function compared with our measurements made over the fallow savannah. The results substantiate Tillman's (1972) use of the relationship (Equation 6), which since it is based on the universal functions given by Monin and Yaglom (1971), should be generally applicable.

The good results obtained with Equation (6) - the free convection limit - also agree with the conclusion of Tillman (1972) and de Bruin (1982) that this equation is a good working approximation for all unstable conditions. This is useful since if the full form of Equation (5) is used, values of u_* and H are first required to calculate the stability. While it is possible to obtain these values in the same way as de Bruin (1982) and Weaver (1990) with a measurement of windspeed and a process of iteration, the simplicity of Equation (6) has obvious attractions.

The results presented in this paper show that the temperature fluctuation method can be used to produce estimates of sensible heat to a high degree of accuracy over a variety of surface types. In particular, the excellent results of the airborne measurements suggest that the technique has the potential to produce estimates of area-averaged fluxes over inhomogeneous vegetation.

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Moore,C.J., 1986. 'Frequency response corrections for eddy correlation systems'. *Boundary-Layer Meteorol.*, 37: 17-45.

Site	Period	Vegetation height (m)	Instrument height (m)	Displacement height (m)	Roughness length (m)
Bare soil	25 MAY 87 - 03 JUL 87	---	3.3	0.0	0.0005
Millet	01 SEP 87 - 02 OCT 87	2.4	4.15	1.13	0.21
Fallow savannah	01 OCT 88 - 11 NOV 88	2.3	12.3	1.4	0.14
Tiger bush	29 JUN 90 - 23 JUL 90	3 - 4	15.9	2.0	0.4
Tiger Bush (Balloon)	11 JUL 90 & 13 JUL 90	3 - 4	25	2.0	0.4

Table 1. Characteristics of measurement period, site and crop. The values of d and z_0 for bare soil is based on Oke (1987), for millet on the relationship between leaf area index and height (Choudhury and Monteith (1988), for fallow savannah on Lloyd et al. (1991), and for tiger bush on Dolman et al. (1991).

Figure captions

Figure 1 Percentage error in the sensible heat flux estimated by the temperature fluctuation method as a function of the percentage error in the height above the zero-plane displacement. Calculations assume θ=303.2 K, σ_θ=0.5 K.

Figure 2 A comparison of sensible heat flux measured by the eddy correlation method with that estimated from the temperature fluctuation method over bare soil. The 1:1 line is also shown.

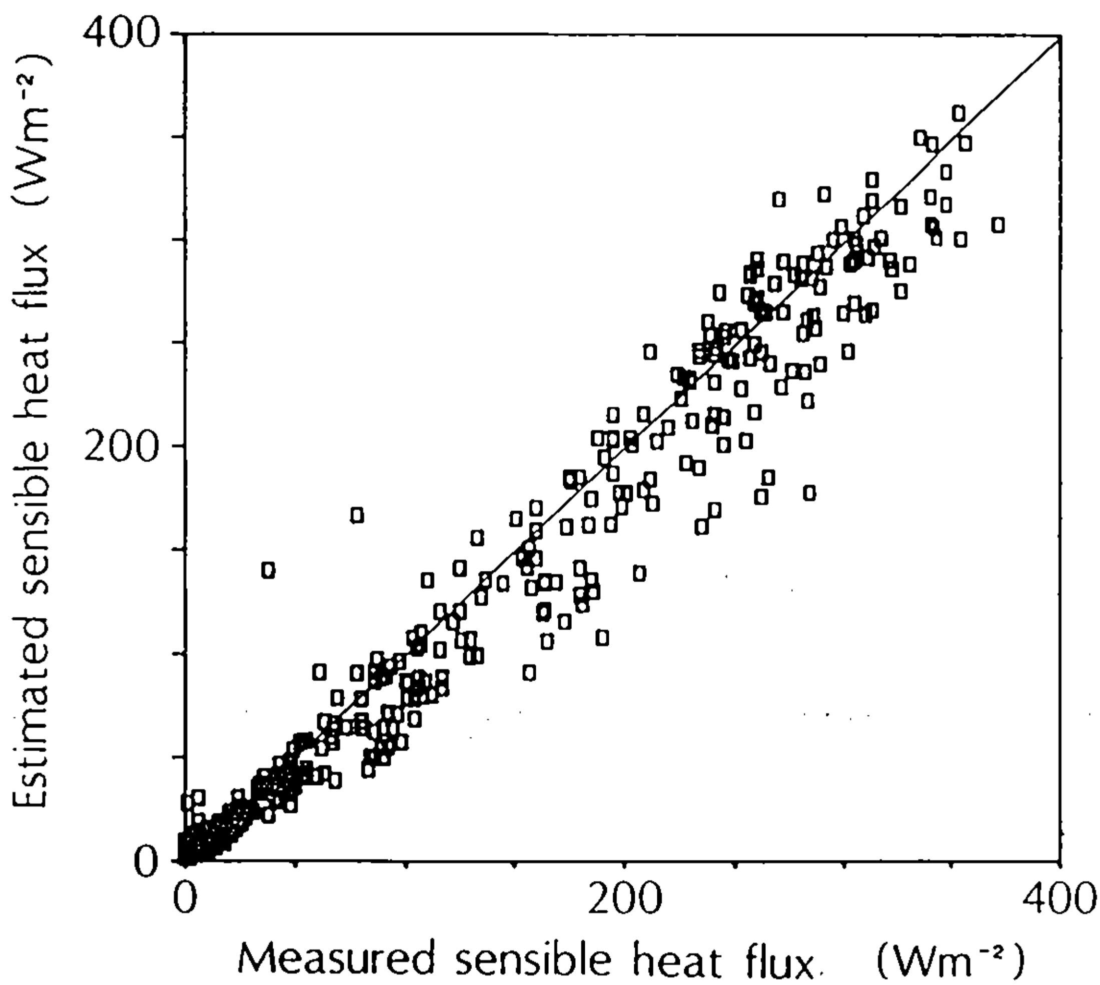
Figure 3 As for Figure 1 but over mature millet.

Figure 4 As for Figure 1 but over fallow savannah.

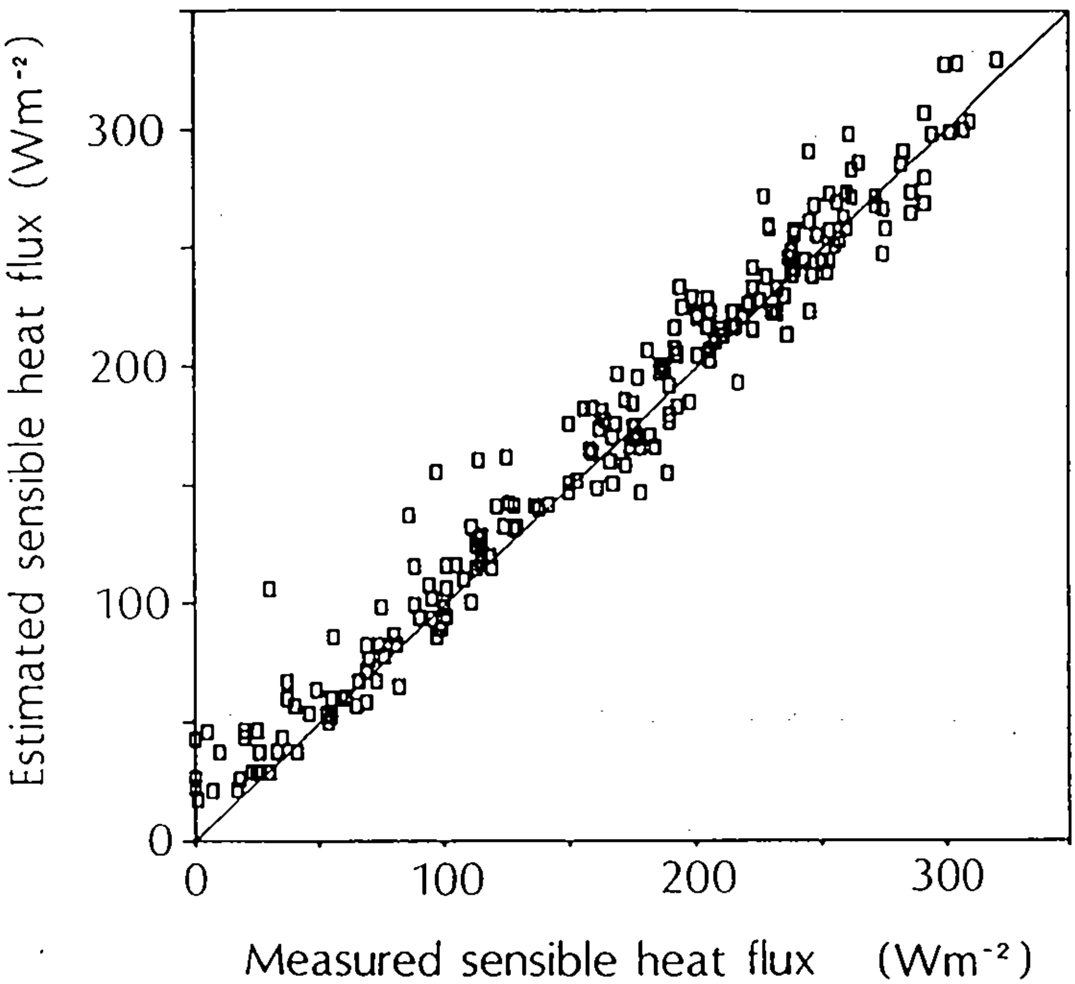
Figure 5 As for Figure 1 but over tiger bush.

Figure 6 A comparison of sensible heat flux measured by an eddy correlation instrument mounted on a tower with that estimated from measurements of temperature fluctuations made with a balloon borne sensor.

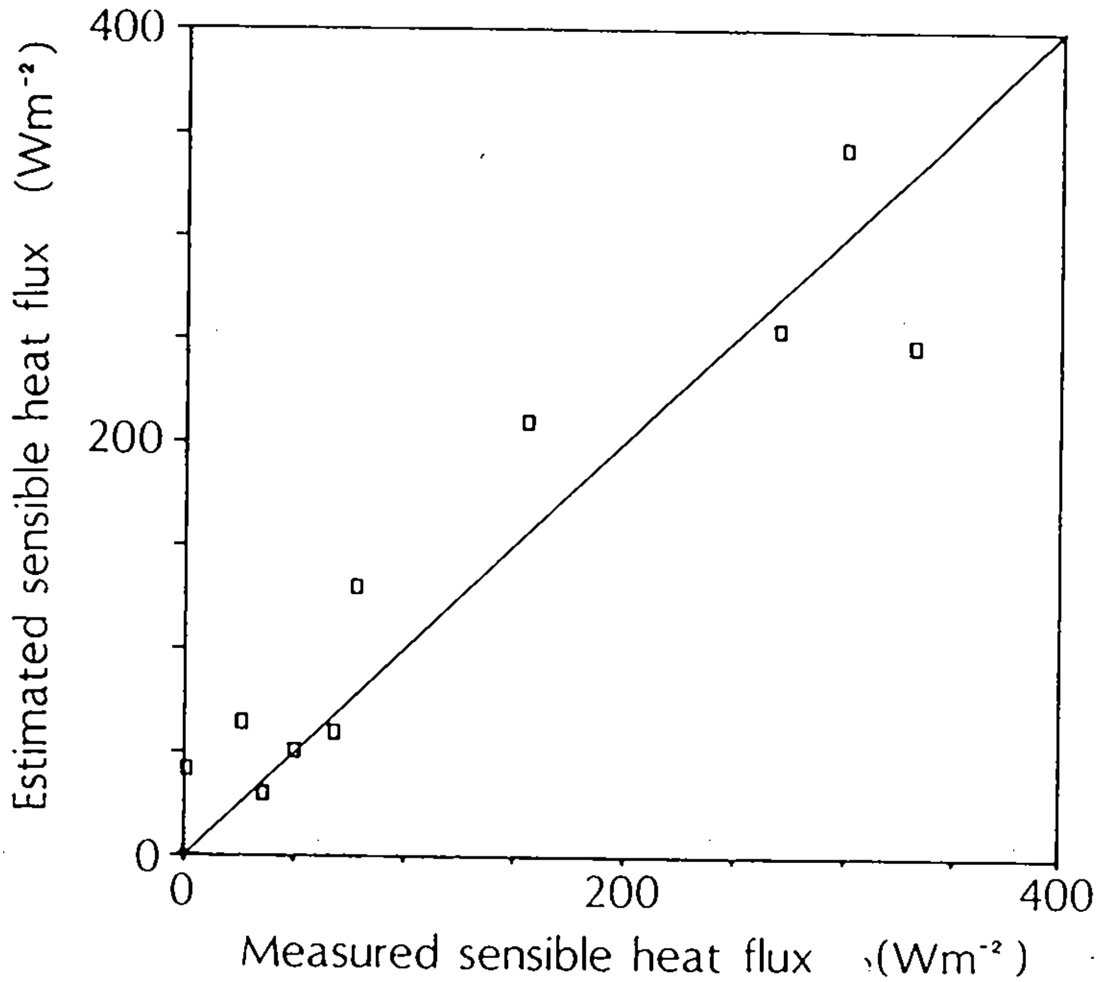
Figure 7 Normalised standard deviation of temperature as a function of stability. Open squares: observations made over fallow bush. Solid line: the universal theoretical relationship as given by Equation 3.



400
200
0
0
200
400
Estimated sensible heat flux (Wm⁻²)
Measured sensible heat flux (Wm⁻²)



Estimated sensible heat flux (Wm⁻²)
Measured sensible heat flux (Wm⁻²)
0
100
200
300



400
200
0
0
200
400
Estimated sensible heat flux (Wm⁻²)
Measured sensible heat flux (Wm⁻²)

Natural
Environment
Research
Council