Modelling soil evaporation in an agroforestry system in Kenya

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Abstract

Soil evaporation measurements from bare soil and shaded soil under an agroforestry tree canopy were used to construct a model to predict soil evaporation with and without tree shade. It was found that a simple daily time step model based on the Ritchie (1972) approach was unable to predict daily soil evaporation accurately, but was capable of providing good estimates of cumulative soil evaporation over hydrologically significant periods (weeks–months). This model was used to show how trees could reduce annual soil evaporation directly beneath their canopy by an average of 35% (compared to completely bare soil), equivalent to 21% of rainfall. In sparse agroforestry tree canopies the area average saving is smaller, depending on tree leaf area index (LAI). The model also demonstrated how annual saving in soil evaporation due to a tree canopy might vary with rainfall, with a maximum of around 180 mm being achieved once rainfall exceeded \(1000\) mm year\(^{-1}\). This saving in soil water is very significant and will help offset the enhanced evaporative losses associated with tree canopies due to interception and re-evaporation of rainfall or as tree transpiration.

Keywords: Agroforestry; Grevillea robusta; Soil evaporation; Modelling; Shade

1. Introduction:

In many agricultural systems in semi-arid regions crops often use a small fraction of the rainfall input since there can be substantial losses of water via soil evaporation, runoff and drainage (e.g. see Wallace and Batchelor, 1997). Of these, soil evaporation is often the largest component. For example, in the semi-arid regions of the Middle East and West Africa, direct soil evaporation from sparse barley or millet crops can account for between 30% and 60% of rainfall (Cooper et al., 1983; Allen, 1990; Wallace, 1991). If some of this unproductive loss of water could be retained in the soil and used as transpiration, yields could be increased without increased rainfall or the use of supplemental irrigation.

Agroforestry is a mixed cropping system where the introduction of trees into a crop may lead to an overall increase in the proportion of rainfall that is used as transpiration (Wallace, 1996). This could be achieved if the tree/crop mixture reduced soil evaporation, runoff or drainage and the water saved retained in the soil and used as tree or crop transpiration. In tropical climates, where soil evaporative losses are high, the extra shade provided by a tree canopy may reduce soil evaporation. It is this aspect which is the
focus of this paper and the effects of agroforestry on runoff, drainage and transpiration are dealt with elsewhere (Jackson et al., 1998).

Soil evaporation theory and several field studies of soil evaporation suggest that total soil evaporation is determined (at least in part) by the radiant energy reaching the soil surface (e.g. see Hillel, 1980). This ‘first phase’ of soil evaporation (described more fully below) could, therefore, be reduced by canopy shade. However, after the first phase is over, soil evaporation rates are determined by the soil hydraulic properties and should, therefore, be independent of shade. Total evaporative loss from the soil is the sum of losses in the energy limited first phase and the hydraulically limited second phase. The net effect of shade on cumulative soil evaporation over periods of several weeks or more will, therefore, depend on the total amount of time the soil spends in first and second stage drying. This will be a function of soil type and the frequency with which the surface is re-wetted by rainfall.

This paper assesses the effects of canopy shade on soil evaporation over hydrologically significant periods (weeks–years) by presenting an analysis of soil evaporation data from an agroforestry trial at Machakos, Kenya. Details of the techniques used and data obtained are given in a companion paper by Jackson and Wallace (1999). These data are used to determine the parameters required for a modified version of the Ritchie (1972) model. The model is then used to calculate evaporation from bare and shaded soil for a series of years with a range of rainfalls.

2. Theory

The model used to predict direct evaporation of water from bare soil is based on the commonly used Ritchie (1972) approach, which considers evaporation to occur in two distinct phases. Initially evaporation from the soil proceeds at the potential rate \( E_{0b} \) (mm day \(^{-1}\)) during the ‘first phase’ immediately following re-wetting of the surface by rain. In this phase soil evaporation is determined by the surface energy balance. Phase I lasts for a number of days \( t_1 \) until the total amount of water evaporated is \( U \) (mm), after which the ‘second phase’ begins, when the rate of soil evaporation is less than potential and is determined by the hydraulic properties of the soil. Theoretical considerations and laboratory and field observations have shown that cumulative Phase II soil evaporation is proportional to the square root of time (e.g. see Ritchie, 1972; Hillel, 1980; Monteith, 1981). Soil evaporation in Phases I and II can, therefore, be expressed mathematically as,

\[
\sum E_{s1} = \sum_{i=0}^{t_1} E_{s0} = U \quad t < t_1 \quad (1)
\]

\[
\sum E_{s2} = \alpha \sqrt{(t - t_1)} \quad t > t_1 \quad (2)
\]

where \( \sum E_{s1} \) and \( \sum E_{s2} \) are the cumulative amounts of soil evaporation (mm) in the first and second drying phases, respectively. \( \alpha \) (mm day \(^{-0.5}\)) is assumed constant for any particular soil and is a function of soil diffusivity (Black et al., 1969). As a result of its simple and practical nature the Ritchie approach is widely used in practice (e.g. see, for example, Dierckx et al., 1986; Daamen et al., 1995; Wallace and Holwill, 1997).

2.1. Phase I evaporation

First phase evaporation is determined by the potential evaporation rate from bare soil, \( E_{0b} \), which Ritchie calculated using the Penman (1963) formula. However, Ritchie assumed that for soil below a crop the aerodynamic term in this formula could be ignored so that \( E_{0b} \) was given by

\[
\lambda E_{0b} = \frac{\Delta + \gamma}{\Delta} R_{ns} \quad (3)
\]

where \( R_{ns} \) is the net radiation above the bare soil (W m\(^{-2}\)), \( \Delta \) is the rate of change of saturated vapour pressure with temperature (kPa K\(^{-1}\)), \( \gamma \) the psychrometric constant (kPa K\(^{-1}\)) and \( \lambda \) the latent heat of vaporisation of water (J kg\(^{-1}\)). Several authors have reported situations where the aerodynamic term in the Penman equation is not negligible, but can be up to 40% of the total potential evaporation (e.g. Daamen et al., 1993; Wallace and Holwill, 1997) and in these cases potential soil evaporation was calculated using the Penman–Monteith equation (Monteith, 1965) with a surface resistance of zero, i.e.,

\[
\lambda E_{0b} = \frac{\Delta(R_{ns} - G_s) + \rho c_p D / r_{sa}}{\Delta + \gamma} \quad (4)
\]
where $G_s$ is the soil heat flux ($W \, m^{-2}$), $D$ is the vapour pressure deficit of the air (kPa) and $r_a$ the aerodynamic resistance to evaporation ($s \, m^{-1}$), $\rho$ the density of air (kg m$^{-3}$) and $c_p$ the specific heat of air at constant pressure (J kg$^{-1}$ K$^{-1}$). In the present study in Kenya daily total values of $E_{0,0}$ were calculated as the sum of hourly $\lambda E_{0,0}$ values calculated using Eq. (4) and hourly measurements of temperature (needed to calculate $\Delta$), vapour pressure deficit and wind speed ($u$, m s$^{-1}$). As $G_s$ was not measured it was taken as a fixed fraction of $R_{ns}$ (i.e. $G_s = 0.2 \, R_{ns}$, see Daamen et al., 1995) and $r_a$ was estimated using the formula for neutral atmospheric conditions, i.e.

$$r_a = \frac{\ln^2\{(z-d)/z_0\}}{k^2u} \tag{5}$$

where $z$ is the reference level height (0.75m), $d$ the zero plane displacement (=0 for bare soil), $z_0$ is the roughness length (= 0.01 m for bare soil), $k$ is von Karman’s constant (0.41) and $u$ is the wind speed at height $z$. Further details of the micro-climate measurements are given by Wallace et al. (1995). Eq. (5) may not be directly applicable under a canopy, however, $E_{0,0}$ is not highly sensitive to $r_a$. An analysis of $E_{0,0}$ calculations for 161 days in 1995 showed that a 25% change in $r_a$ only produced, on average, a 6% change in $E_{0,0}$. Eqs. (4) and (5) should, therefore, produce a reasonable estimate of bare soil potential evaporation both in the open and under a sparse tree canopy.

2.2. Phase II evaporation

In Ritchie’s model soil evaporation switches from Phase I to Phase II when the cumulative evaporation exceeds a value, $U$. The value of $U$ is soil dependent and is as low as 3 mm in very sandy soils (Daamen et al., 1993; Wallace and Holwill, 1997). A value of 6 mm was used by Ritchie for a Plainfield sand, 9 mm for a loam and 12 mm for a clay. The soil in the present study is a sandy loam and would be expected to have a value between 6 and 9 mm. This is supported by inspection of the soil evaporation data (Jackson and Wallace, 1999) which showed that soil evaporation on the day after wetting by rain could reach 6 mm, but on the subsequent day total evaporation was usually well below potential rate and in the range 3–4 mm. Assuming that soil evaporation could proceed at the potential rate for several hours on the second day after rain (e.g. also see Daamen and Simmonds, 1996), the value of $U$ used here in the modelling of soil evaporation was taken as 7.5 mm. This value is midway between that for a sand and a loam (Ritchie, 1972). The other parameter required to calculate Phase II evaporation is $\alpha$ and the value of this for the soil at the site in Kenya was determined from observations of soil evaporation after rainfall (see later).

2.3. Modelling the effects of a canopy on soil evaporation

The presence of vegetation can affect soil evaporation in several ways. The first is via the shading of the ground which alters the net radiation, temperature, humidity and wind speed above the soil. These alterations in micro-climate combine to reduce the potential bare soil evaporation rate beneath the canopy. In the agroforestry trials in Kenya all of the above variables were measured beneath the canopy and used to calculate $E_{0,0}$ under shade (Jackson and Wallace, 1999). When these daily $E_{0,0}$ values were available, e.g., in 1995, they were used in the soil evaporation model. When daily $E_{0,0}$ data were not available (i.e. 1984–1988), daily mean values of $E_{0,0}$ beneath a canopy, calculated from the data in periods when data were available (222 days during the 1995 short rains and 1996 long rains), were used in model simulations.

The second way in which the canopy affects soil evaporation is via the enhanced drying of the surface soil layer due to the abstraction of water by roots. Most vegetation types will have active roots in the top soil layers and as these extract water the soil will dry at a quicker rate than it would were there are no roots. Daamen et al. (1995) showed that accounting for root water abstraction in millet was important for the accurate prediction of soil evaporation. They proposed a simple modification of the Ritchie approach which effectively speeds up soil drying in Phase II when there is vegetation present. This modification was used here to calculate the rates of shaded soil evaporation in Phase II. Following Daamen et al. (1995) root water abstraction from the upper soil layers is assumed to be a fixed proportion ($V$) of soil evaporation. Daamen et al. (1995) justify this assumption by arguing that soil evaporation and root abstraction will be well correlated since they both depend on the matric potential in the soil. At the end of each day, therefore,
the cumulative loss of water from the uppermost soil layer is $\sum (1 + V)E_{s2}$. This sum is used to calculate an equivalent time into second stage drying, $t_{eq}$, i.e.,

$$t_{eq} = \left(\sum (1 + V)E_{s2}/\alpha\right)^2$$

(6)

Soil evaporation on the following day is then calculated as

$$E_{s2} = \alpha \sqrt{t_{eq} + 1} - \alpha \sqrt{t_{eq}}$$

(7)

This calculation of $E_{s2}$ is used when a tree canopy is present and replaces that given for bare soil (i.e. Eq. (2)).

In the current paper $V$ was assumed to be 1.0, the same as the value used by Daamen et al. (1995). They have pointed out that the value of $V$ will depend on a number of factors, including leaf area index (LAI), root density and soil moisture distributions in the surface soil layers. However, a primary factor will be LAI and the tree LAI during the period studied in this paper, i.e., between 1 and 2.5 (Jackson and Wallace, 1999) is similar to the LAI of the millet crops studied by Daamen et al. (1995). The net effect of introducing root water abstraction is to reduce the cumulative amount of soil evaporation in Phase II. Root water abstraction is not included in Phase I drying as the soil layer which supplies the water for this evaporation is only the top few centimetres where active roots are unlikely to be present due to disturbance caused by surface tillage and high surface temperatures. Daamen et al. (1995) also concluded that root abstraction is only likely to be important in second stage drying, since any significant Phase I root abstraction in the surface soil layer may be compensated by reduced drainage immediately following rain.

Another way in which a canopy can affect soil evaporation is via a reduction in the rainfall input to the soil surface due to canopy interception of rainfall. This affects the surface water storage, which is compared to the value of $U$ to define when soil evaporation switches between Phase I and Phase II. In this paper we can use a very simple way of dealing with interception since the sensitivity of the modelled values of $E_s$ to the value of the net reduction in the rainfall is low (see below). It is, therefore, sufficient to assume that all rainstorms are reduced by a fixed amount which we take as twice the canopy storage capacity. This figure is deduced as follows. For a tree canopy with full cover, canopy storage is ~1 mm (e.g. see Teklehaimanot and Jarvis, 1991). This means that after every rainstorm 1 mm of water would be held and evaporated from the canopy. However, a similar amount of water is also evaporated from the canopy during rainfall (Gash et al., 1995), giving a total loss of ~2 mm per storm. Modeled values of $E_s$ presented later in this paper are for the soil directly beneath the agroforestry tree canopy. Hence the ‘full cover’ value of canopy storage (2 mm) was used in these calculations. Modeled values of annual total $E_s$ only changed by 4% for a 50% change in canopy storage.

2.4. Evaporation on rain days

First phase evaporation on days after rainfall is described by the above Phase I evaporation model. However, evaporation may also occur on days when it rains and evaporation during these days contributes significantly to the total soil evaporation (i.e. about one third in an average year at Machakos). Normally the only data available are daily rainfall totals, with no information on when the rain fell within a day or what the duration and intensity of the rainfall was. In this case if rainfall can occur at any time of the day, the simplest assumption that can be made is that all the rain falls instantaneously at midday. Soil evaporation on this day can then be estimated as the mean of the daily rates before, $E_s^p$, and after, $E_s^{eq}$, the rain storm. The rate before the storm is that which would occur if it had not rained on the day in question and hence will depend on the time since the previous rain storm. The rate after the storm will be the appropriate potential soil evaporation rate. One further adjustment to $E_s$ on rain days was made to account for the duration of rain storms, during which soil evaporation was assumed to be zero. Rainfall duration was estimated from the daily rainfall total divided by the mean rainfall intensity for the site. In the current study the mean rainfall intensity of 2.2 mm h$^{-1}$ was calculated from 754 rain events recorded at Machakos between 1992 and 1996. In practice, when comparing modelled with measured soil evaporation on rain days it was found that the model set soil evaporation too low ($E_s = 0$) on days with large storms (>26 mm day$^{-1}$). Although this only occurred on 5% of rain days, we chose to avoid this by limiting the duration of rainfall to a maximum of 6 h during the day.
3. Materials and methods

The soil and weather data used in this paper were collected at the Machakos field station of the International Centre for Research in Agroforestry (ICRAF), located ~80 km south-east of Nairobi, Kenya at 1°33'S, 37°8'E. The climate of the site is sub-humid with a bi-modal rainfall distribution consisting of a short rainy season of 265 mm usually lasting from late October to late December, and a longer rainy season of 345 mm running from late March to the end of May (Jackson and Wallace, 1999). The mean annual rainfall is 782 mm, which is about half of the annual average potential evaporation of 1450 mm (Huxley et al., 1989).

The site has a south-west facing slope of ~22% with a reddish-brown sandy clay loam soil of variable depth (0.2–2 m). The soil has a number of distinct horizons and is underlain by layers of first weathered and then coherent rock (gneiss) at varying depths. Throughout the entire 6 years of the experiment no permanent water table was observed within the soil profile or weathered rock. Further details of the site and soils are given by Wallace et al. (1995) and Jackson and Wallace (1999).

The agroforestry trial within which measurements were made was established in 1991 using Grevillea robusta (A. Cunn. ex R. Br.) seedlings. A total of 25 plots measuring 20 m × 20 m were laid out in a randomised block design (Wallace et al., 1995). In this paper we use soil evaporation measurements made in either completely bare soil or in plots with trees planted on a 3 m × 4 m grid. Measurements of soil evaporation were made when the trees were between 3 and 5 years old, but because of regular pruning their canopies only covered between 20% and 50% of the ground. Further details of the layout and management of the agroforestry trial are given by Jackson and Wallace (1999).

Direct evaporation of water from the soil was measured using 24 small soil lysimeters (150 mm diameter × 100 mm deep). Approximately 200 lysimeters were installed within the bare and tree covered plots at the start of each season. They were pushed into the soil using as little force as possible, and left for several weeks to allow the soil in them to settle after any soil disturbance during their installation. When the lysimeters were required for measurement, they were carefully extracted and the perforated sides and base sealed with polythene and a PVC base plate. To complete the installation, the lysimeters were lowered into lined holes located either in a bare soil plot or beneath the tree canopy in another undisturbed part of the trial. Following rainfall events the lysimeters were weighed every morning and afternoon on a balance with a resolution of 0.1 g, equivalent to 0.01 mm of water. Further details of the procedure used in the micro-lysimetry technique and the uncertainties associated with it are given by Jackson and Wallace (1999).

The weather data needed to calculate the potential soil evaporation rates from both bare soil and shaded soil were measured at appropriate locations. Within the tree plot hourly measurements of temperature, humidity and wind speed were made using an aspirated wet and dry bulb thermometer system and cup anemometer placed beneath the tree canopy, 0.75 m above the ground. Net radiation was measured beneath the tree canopy using two radiometers (Model Q6, Radiation Energy Balance Systems, Seattle, USA) located 0.75 m above the soil. The aerodynamic resistance needed to estimate $E_{s0}$ was calculated using Eq. (5) using a reference height of 0.75 m. Separate net radiation measurements were made over a bare soil plot, again using a radiometer 0.75 m above the surface. Reference level measurements of temperature, humidity and wind speed used to calculate bare soil potential evaporation were recorded ~2 m above the tree canopy and assumed to be the same over the adjacent bare soil plot (only ~40 m away). However, in this case $r_0$ was calculated using Eq. (5) with a reference level of 7.8 m.

4. Results and discussion

Fig. 1 shows soil evaporation data for three drying events in 1996. Both measured $E_s$ and calculated $E_{s0}$ are lower under the canopy than in open bare soil. These data are used to define the parameter $\alpha$ needed to model Phase II soil evaporation. On the first day following rainfall in each event soil evaporation is close to the potential rate and so it is assumed that Phase I evaporation dominated. Note that the rainfall event on the 8 April 1996 occurred before dawn, so the $E_s$ data shown follow this event. On the second day
after rain evaporation is generally significantly below the potential rate, especially in open bare soil, so it was assumed that this was the first day of Phase II drying. The switch from Phase I to Phase II is not always easy to identify and some of the problems in defining the two phases of soil evaporation are discussed later.

Fig. 2 shows the variation in the cumulative evaporation from the soil under three different degrees of exposure, for the three drying events shown in Fig. 1. The plots show that there is some variation in slope for different drying events. In contrast, for any given event there is little difference in the slope of the lines for different exposure, as expected if Phase II evaporation is independent of evaporative demand. However, the differences in slope between different drying events indicates that Phase II evaporation may not always proceed at the same rate. This could be a consequence of different degrees of soil wetness after the rain event which preceded the drying cycle. Indeed the highest rates of soil evaporation shown in Fig. 2 (Event b) came after the greatest antecedent rainfall (76 mm in the previous 3 days, Table 1). Although similar amounts of soil evaporation were lost on the first day of Phase II drying in each of the three events (Table 1), the highest cumulative soil evaporation over 4 days occurred under the greatest potential evaporation. This indicates that although Phase II evaporation is very much less than potential, it may not be entirely independent of either soil wetness or atmospheric demand.

Experimental results and modelling of soil evaporation reported by Daamen and Simmonds (1996) and Wallace and Holwill (1997) show that in practice soils do not simply move from Phase I to Phase II drying after a given number of days drying. Firstly, Phase I drying does not always last for an integral number of days and can even be much less than 1 day in very sandy soils. This means that when using a daily time step model, it can be difficult to define the beginning of Phase II drying accurately. This will account for some of the differences in evaporation observed on the first day of Phase II drying shown in Fig. 2. Another complication arises because of the observation by Daamen and Simmonds (1996) and Wallace and
Holwill (1997) that soil evaporation during Phase II can return to the potential rate in the early part of a day due to re-hydration of the soil surface layer overnight. This will mean that Phase II evaporation may have some dependence on evaporative demand, as suggested above. This kind of behaviour can be predicted using an hourly time step model with a multi-layer soil moisture description and complete surface energy balance (e.g. see Daamen and Simmonds, 1994). Simple daily time step models which assume Phase II evaporation independent of evaporative demand will only be able to give an approximate description of soil evaporation under these conditions.

Despite the approximate nature of the Ritchie model in Phase II, we will see later that it is still possible to use a simple daily time step model to obtain reasonable estimates of cumulative soil evaporation over hydrologically significant periods (weeks–months). To do this the data shown in Fig. 2 (a,b,c) were averaged for each time after the start of Phase II drying and the mean and standard deviations are shown in Fig. 2(d). The straight line fitted to these averaged data provides the following equation for calculating Phase II cumulative evaporation

$$\sum E_{s2} = 4.95\sqrt{(t - t_1)} - 1.43 \quad (8)$$

A measure of the uncertainty in any calculation of Phase II evaporation is given by the ratio of the standard deviation/mean for each point in Fig. 2(d), and this varies from 0.13 to 0.21 as time increases from 1 to 4 days after the start of Phase II drying. The fitted line in Fig. 2(d) does not go through the origin. This could arise if some Phase II (i.e. sub-potential) evaporation had already occurred on the day before the summation is started. The soil evaporation model in this paper uses Eq. (8) instead of Eq. (2) to calculate $E_{s2}$.

Fig. 3 shows a comparison of daily measured and modelled soil evaporation for the period 9 February–9 June 1995. In general, modelled soil evaporation follows the pattern of measured evaporation quite well, but there are clearly occasions when the modelled values are quite different from measurements. This is highlighted in Fig. 4, where a plot of daily modelled against measured soil evaporation for the same period shows a large scatter about the 1 : 1 line. This shows that there is fairly high uncertainty in individual daily values of soil evaporation estimated with the current model. The root mean square deviation of the modelled values from measurements shown in Fig. 4 is 1.0 mm for the bare soil and 0.8 mm for the soil below the tree canopy. However, it is the cumulative soil evaporation over periods much longer than a day which is of hydrological significance. Fig. 5, therefore, shows a time series plot of the cumulative measured and cumulative modelled soil evaporation. The model and measurements remain very close throughout most of the period compared. In the bare soil, the model cumulative total for the entire period shown is 134 mm compared to a measured total of 127 mm. Below the tree canopy the modelled total evaporation for the same period is the same as the measured total at 101 mm. The largest differences between measured and modelled $E_s$ occurred in the bare soil when small rainstorms generated higher modelled $E_s$ than were measured at that time. For example, this happened on 30 April and 16 May when rainfall was 5.5 and 2 mm, respectively (see, Figs. 3 and 5).

Fig. 5 also clearly shows that evaporation losses under the tree canopy are lower than in bare soil. The measured change from bare to shaded soil for the entire period shown is $-27$ mm or a decrease of 21%. The equivalent modelled change is slightly higher at $-33$ mm or 25%. Within the model, reduced potential
evaporation accounts for $-9.5$ mm, interception for $-9.5$ mm and root abstraction for $-14.4$ mm. In the model, therefore, interception has as large an effect as reduced potential evaporation rate and the largest single effect is due to root abstraction of water in the upper soil layers.

The good agreement of the cumulative data and poor agreement of the daily data can be explained as follows. The total amount of water which the model predicts to be evaporated over the first few days after rain is constrained by the amount of water in the surface soil layers, i.e., the value of $U$. Underestimates of $E_s$ early in a drying cycle, therefore, tend to be compensated by overestimates later in the cycle and vice versa. For example, Fig. 3(a) shows that on 27 March modelled bare soil $E_s$ was well below that measured, but the reverse was true on the following 2 days. The net result was that cumulative measured and modelled $E_s$ over the 3 days (27–29 March) were very similar at 10.5 and 11 mm, respectively. The cumulative totals predicted by the model are therefore much more reliable than individual daily values. Our overall conclusion is that the model described here gives reliable cumulative soil evaporation predictions over periods around a week or longer, both for completely bare soil and for soil beneath a tree canopy.

4.1. Extrapolation to longer time periods

The long term effect of a tree canopy on soil evaporation is illustrated in Fig. 6. Here the cumulative seasonal amount of soil evaporation from both bare and shaded soil at Machakos was calculated over an 18 month period using the model described earlier. The mean potential bare soil evaporation rate ($E_{s0}$) used for this calculation was 4.4 mm day$^{-1}$, calculated from measurements made during the 1995 short...
and 1996 long rains. Fig. 6(a) shows that total bare soil evaporation is \( \approx 59\% \) of rainfall over the three wet seasons in 1994 and 1995. Just under half of the evaporation occurs in Phase I (\( \approx 47\% \)) when evaporative demand determines the loss rates. Similar high losses of water due to soil evaporation have been reported by Cooper et al. (1983), Wallace (1991) and Wallace and Holwill (1997).

Fig. 4. Daily modelled versus measured soil evaporation for (a) bare soil and (b) soil directly beneath a tree canopy. Solid circles [●] show Phase I evaporation and open circles [○] show Phase II evaporation.

Fig. 5. (a) Cumulative modelled [●] and measured [○] bare soil evaporation, and modelled [□] and measured [■] soil evaporation below the tree canopy, over the 1995 long rainy season at Machakos, Kenya. (b) Daily rainfall over the same period.
The degree to which the presence of a tree canopy can affect soil evaporation is shown in Fig. 6(b). For these calculations $E_s$ was reduced to 3.7 mm day$^{-1}$ (again calculated from measurements made beneath the tree canopy in the 1995/1996 period used above) and the effects of root water abstraction and canopy interception were introduced. Under shade $E_s$ decreases to $\sim$41% of rainfall, still a very significant proportion of the soil water balance. The net effect of the canopy over the 18 month period shown was to reduce soil evaporation by 220 mm, or 18% of rainfall.

Table 2 shows the results of a series of calculations of soil evaporation at Machakos for the years 1984–1988. Part (a) shows annual total values of soil evaporation in both Phase I and Phase II. Apart from the driest year (1987), more bare soil evaporation occurs in Phase I than Phase II. Total soil evaporation varies from 50% to 89% of rainfall, the higher proportions being lost in dry years.

Table 2(b) shows similar figures for soil evaporation under a canopy. Both Phase I and Phase II evaporative losses are reduced by the presence of the tree canopy, the reduction being greatest in Phase II. The amount of rainfall lost as soil evaporation goes down by between 108 and 196 mm, depending on rainfall. On average, the presence of a tree canopy can be regarded as ‘saving’ around 157 mm water per annum, or 21% of rainfall. The canopy has a proportionally greater effect in dry years, when the percentage of rainfall saved rises to 33%.

4.2. Prediction of evaporation in different climates

The effect of a tree canopy on soil evaporation in drier and wetter rainfall climates than Machakos was estimated using the model described here with rainfall data from Sadoré, Niger and Kimakia, Kenya. Identical soil parameters were used for each site but the potential bare soil evaporation rates were altered in each location. Potential soil evaporation for Sadoré, Niger was taken as 4.0 mm day$^{-1}$ (Wallace and Holwill, 1997). No potential soil evaporation data were available for Kimakia so conventional Penman potential evaporation data were used for this site (Blackie and Edwards, 1979). The mean daily Penman potential evaporation rate for the 16 years period 1958–1973 was 3.0 mm day$^{-1}$. Canopy shade was simulated at all the three sites by assuming that potential soil evaporation was reduced by the same amount as measured at Machakos (i.e. 15%) and the simple treatments of canopy interception of rainfall and root abstraction of soil water are included in all calculations.

Fig. 7 shows the results from the above calculations. The reduction in $E_s$ due to the presence of a canopy increases as rainfall increases up to $\sim$1000 mm year$^{-1}$, Fig. 7(a), after which it remains roughly constant at $\sim$180 mm. Fig. 7(b) shows the reduction in soil evaporation due to the presence of a canopy as a percentage of rainfall. When rainfall is less than 500 mm year$^{-1}$ the reduction in $E_s$ is equivalent to $\sim$30% of rainfall at both the Kenyan and Nigerian sites. At the Kenyan sites this percentage
declines steadily as rainfall increases to less that 5% when rainfall exceeds ~2500 mm year⁻¹. In contrast, the reduction in $E_s$ due to a canopy declines much more rapidly with increasing rainfall in Niger than in Kenya. This is mainly due to the different rainfall climate in Niger where much more of the rain falls as large, high intensity storms, leading to fewer soil drying events. This is reflected in the hourly mean rainfall rate for Sadoré, Niger of 5.7 mm h⁻¹, compared to 2.2 mm h⁻¹ at Machakos in Kenya.

Table 2
Calculated amounts of annual evaporation from (a) bare and (b) shaded soil for Machakos, Kenya

<table>
<thead>
<tr>
<th>Year</th>
<th>Phase I $E_{s1}$ (mm)</th>
<th>Phase II $E_{s2}$ (mm)</th>
<th>Total $E_s$ (mm)</th>
<th>Rainfall $P$ (mm)</th>
<th>$E_s/P$ (%)</th>
<th>Bare-shade $E_s$ (mm)</th>
<th>$E_s/P$ (%)</th>
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<tr>
<td>Mean 1984–1988</td>
<td>243</td>
<td>209</td>
<td>452</td>
<td>761</td>
<td>59</td>
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</table>

(b) Soil under canopy

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<th>Year</th>
<th>Phase I $E_{s1}$ (mm)</th>
<th>Phase II $E_{s2}$ (mm)</th>
<th>Total $E_s$ (mm)</th>
<th>Rainfall $P$ (mm)</th>
<th>$E_s/P$ (%)</th>
<th>Bare-shade $E_s$ (mm)</th>
<th>$E_s/P$ (%)</th>
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<td>75</td>
<td>231</td>
<td>597</td>
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</table>

Fig. 7. Annual saving in soil evaporation due to a tree canopy for the same soil (sandy loam) at three different sites; Machakos (Kenya) [●], Kimakia (Kenya) [▲] and Sadoré (Niger) [○]. Data shown as (a) absolute amount in mm and (b) as a percentage of rainfall.
5. Conclusions

On an annual basis it has been shown that, on average, a tree canopy can reduce soil evaporation by 35% (compared to completely bare soil), equivalent to 21% of rainfall (Table 2). This figure applies to $E_s$ immediately below a tree canopy, i.e., over the fraction of the land area which is considered as ‘covered’ by the trees. Jackson and Wallace (1999) also present data for this location and found a similar reduction in $E_s$ of 30% when the tree LAI was 2.5. However, they also showed that $E_s$ was reduced by a smaller amount (20%, when LAI = 2.5) in the ‘uncovered’ fraction of ground which is not fully shaded. The area average reduction in $E_s$ due to a sparsely planted agroforestry system can be estimated from the mean of the values for the covered and uncovered areas given by Jackson and Wallace (1999) (Fig. 4) and this can be expressed as a fraction of rainfall using the mean value of bare soil $E_s/P = 0.59$, (Table 2). This gives the area average reduction in $E_s$ due to a tree canopy with and LAI of ~2.5 of 25%, equivalent to 15% of rainfall. Similar calculations for a LAI of 1 give the area average reduction in $E_s$ as 10%, or 6% of rainfall. This means that in sparse tree canopies the area average saving in water due to reduced soil evaporation is smaller than the figure given above for complete cover (i.e. 21% of rainfall, Table 2), but it is still significant and will help offset the enhanced evaporation losses associated with canopy interception or tree transpiration.

Daamen et al. (1995) reported that sparse millet crops growing on sandy soil in Niger reduced seasonal soil evaporation by between 12% and 16%. The canopy LAIs associated with these crops were between 1 and 2.7 respectively. In the present study, area average reductions in $E_s$ due to tree shade are similar at the lower LAI (i.e. ~1), but larger at the higher LAI (i.e. ~2.5). Higher soil evaporation and greater reductions due to shade would be expected in the present study since the soil has a higher water holding capacity. This is reflected in the value of $U$ which has been estimated as only 3 mm for the sandy soil in Niger (Daamen et al., 1993; Wallace and Holwill, 1997), but 7.5 mm for the sandy loam in Kenya. As a consequence the Kenyan soil stays in Phase I for longer and soil evaporation is, therefore, under a greater degree of control by the energy available at the soil surface which is affected by shade. This effect will be most pronounced at high LAIs. Differences in $E_s$ between the present study and the work of Daamen et al. (1995) would also be expected due to differences in rainfall climatology. Less water is saved due to canopy shade in Niger where the rain falls in relatively few intense storms separated by comparatively long dry spells. This effect can be seen in Fig. 7(b) where the saving in $E_s$ in Niger is generally less than that in Kenya at the same rainfall. Conversely, more saving in $E_s$ occurs in Kenya, where the rain falls at lower intensities separated by comparatively short periods.

The model used in this study has demonstrated that the presence of a canopy could affect $E_s$ in three ways. By reducing the potential evaporation rate at the soil surface, by reducing the input of rain to the soil due to canopy interception and by reducing the soil surface water content due to water abstraction by roots. The latter two of these effects have been fairly crudely modelled and yet appear to have important influences on $E_s$ below a canopy. There is, therefore, scope for improving the modelling of rainfall interception and surface root water abstraction. The former could be achieved by using the complete Gash et al. (1995) sparse forest canopy model rather than the simplistic approach used here. Improved specification of the value of $V$ used in the $E_s$ model, which was used to represent the enhanced drying of the soil surface due to root abstraction, will require a more complete water balance analysis in which changes in surface soil water content can be separated into soil evaporation, root abstraction and drainage.

Rainfall interception in the agroforestry system with an LAI of 2.5 (i.e. ~50% cover) has been estimated by Wallace et al. (1995) to be ~11% of rainfall. The same density tree canopy has been shown to produce a saving in soil evaporation of 15%, when annual rainfall is close to average (780 mm). In this type of rainfall climate the effects of a tree canopy on rainfall interception and soil evaporation should, therefore, tend to cancel each other, with small gains being made in the order of 4% of rainfall. A similar picture should hold for canopies with lower (or higher) LAIs, since both $E_s$ and interception will vary in a similar way with cover. However, where rainfall is very low (<500 mm year$^{-1}$), the reduction in $E_s$ due to a tree canopy may be greater than the interception loss,
so the net benefit to the soil water balance may be greater than 4%. Conversely, in rainfall climates where either annual rainfall is high (e.g. \( \gg 1000 \text{ mm} \)) and/or rainfall intensities are high (e.g. \( \gg 3-4 \text{ mm h}^{-1} \)), losses due to interception may be greater than savings due to reduced soil evaporation. The net balance between reduced soil evaporation and increased rainfall interception could, therefore, be positive, neutral or negative, depending on the rainfall climate and soil type.

Clearly the introduction of trees into a crop may also affect the soil water balance due to the transpiration by the trees and potential effects on runoff and drainage. A detailed analysis and discussion of these factors are outside the scope of the current paper. However, the savings in soil evaporation identified here need to be taken into account in calculating the net effect of introducing trees into crops on the soil water balance.

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