141

Soil Water Balance in the Sudano-Sahelian Zone (Proceedings of the Niamey Workshop, February 1991). IAHS Publ. no. 199, 1991.

Weather and water in the Sudano-Sahelian zone

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Abstract Evaporation from bare soil surfaces and from vegetation plays a dominant role in the water balance of the Sudano-Sahelian zone. Starting from the formula first derived by Penman, equations are derived for (i) evaporation from bare soil; (ii) potential transpiration from well-watered crop stands; and (iii) actual evaporation from a region. These equations were used to estimate rates of evaporation in southern Niger with emphasis on the contrast between a very dry year (1984) and a wet year (1988). The annual evaporation from bare soil is estimated to be about 230 mm. Potential evaporation from crops in the wet season is close to 6 mm per day, consistent with measurements by Dancette, but is underestimated in the dry season by the original Penman equation and by the Priestley-Taylor equation. Actual regional evaporation estimated between 1984 and 1989 ranged between 300 and 500 mm. Corresponding estimates of runoff are consistent with published measurements ranging from about 30 to 230 mm for areas with little vegetation and from zero to about 110 mm for areas with 50% vegetation.

WORLD WATER

According to several authorities quoted by Baumgartner & Reichel (1975), mean precipitation over the entire global surface is just over 1 m per year, but land surfaces receive only 200 mm per year. The Sudano-Sahelian zone is usually defined in terms of rainfall limits that range from about 400 to over 1000 mm; so in terms of the water it receives, the zone is relatively wet.

Shortage of water in the semiarid tropics is not a consequence of poor annual rainfall. The problem for human settlement and particularly for agriculture, is the seasonal distribution of rainfall (Sivakumar & Wallace, 1991) and the rate at which it is lost by evaporation.

Water vapour held in the earth's atmosphere is equivalent to about 20–30 mm of liquid water, an amount that would be exhausted in less than 10 days if it were not continually replenished by evaporation from the oceans, from soil and from vegetation. In some temperate environments and in the humid tropics, evaporation is a relatively small fraction of annual rainfall; but in semiarid regions evaporation is a major component of the water balance in all years – and it is the only mechanism for loss in very dry years.

This review will be concerned mainly with ways in which evaporation can be quantified in the Sudano-Sahelian zone and with implications for other ways in which water can be lost by percolation or overland flow. The way will then be clear for the following speakers to deal with these processes in more detail.

WATER BALANCE

Within any defined hydrological unit, conservation of mass requires that the input of water over a given time must be precisely balanced by the loss of water and/or by a change in the amount of water stored within the unit. For a simple unit such as the top 2 m of soil within a catchment, the balance can be written as:

$$P = E + R + D + S \tag{1}$$

where P is precipitation, E is evaporation, R and D are the net amounts of water lost by overland flow and deep percolation respectively, and S is the increase in the amount of water stored in the soil. It is both convenient and conventional to express each of these quantities as an equivalent depth of water per unit time, e.g. mm day^{-1}

Throughout this meeting, we shall be concerned with all five variables in equation (1) but will treat them differently in terms of measurement, analysis and management. We will be concerned with the variability and associated unpredictability of precipitation, the one variable that we cannot "manage". Even after 40 years of attempts to "make rain" by seeding clouds, we are still unable to manipulate this input; but models of atmospheric circulation suggest that global warming may eventually be responsible for inadvertent increases or decreases in rainfall. Unlike precipitation, evaporation is a quantity we can modify by management but only to a limited extent.

Runoff is more amenable to management in ways discussed by Lal (1991) and Hoogmoed *et al.* (1991); and an increase in soil water or surface water storage discussed in several papers can be regarded as the prize for success in decreasing evaporation and/or runoff within a catchment.

PRINCIPLES OF EVAPORATION

The rate at which water evaporates from any wet surface is determined (a) by the physical state of the surrounding air as specified by its temperature (T), its vapour pressure (e) and its velocity; (b) by the net amount of heat (H) supplied by processes such as radiative transfer and conduction; and (c) by the wetness of the surface.

The diffusion of heat or vapour from a wet surface into the atmosphere is often treated like a current passing through a resistor in an electrical circuit and many systems from which water evaporates contain two types of resistor. The first (r_a) depends on the thickness of the aerodynamic boundary layer over the surface, and therefore on wind speed and surface geometry. Fluxes of heat, water vapour and momentum all pass through this type of resistor (with a resistance somewhat different for each of these entities). The second (r_s) is a surface resistor which restricts the diffusion of gases such as water vapour and carbon dioxide in their passage through pores in leaves or in soil.

The equations describing how a wet surface exchanges heat and water vapour with the air passing over it include surface temperature as a variable. Although this quantity is continuously monitored over the whole earth by radiometers mounted on satellites, routine measurements close to the ground are rare. Fortunately, Penman (1948) discovered a simple way of eliminating surface temperature from the equations in which it appears. Then if the evaporation rate E is expressed in mass per unit area and per unit time and L is the latent heat of evaporation, the flux of latent heat (LE) becomes a function of five quantities, viz.

$$LE = LE\ (H,\ T,\ e,\ r_a,\ r_s) \tag{2}$$

The nature of this function in the Penman-Monteith (PM) equation (Monteith, 1981) is given in Appendix 1. Only one further detail is needed here: the numerator of the equation contains a term proportional to the saturation deficit of the air which is:

$$D = e_s(T) - e \tag{3}$$

where e_s is the saturation vapour pressure of air at temperature T.

The main obstacle to the operational use of the PM equation is that r_s, like surface temperature, is seldom measured in a routine way. The equation has therefore been used for diagnosis rather than prognosis in the analysis of experiments where evaporation rate was measured along with H, T, e and r_a in order to determine r_s. From a substantial body of knowledge accumulated over 25 years, it is clear that whereas the value of r_s for water and for thoroughly wetted soil is zero, it is about 60 s m^{-1} for most types of well-watered vegetation.

In the material that follows, equation (2) is used initially in a conventional way to specify the rate of evaporation from a bare soil surface or from a crop with foliage whose "wetness" can be prescribed by the value of r_s. The equation is then manipulated to obtain a new formula which does the reverse: it estimates r_s from the state of the atmosphere and therefore provides a way of measuring the *actual* evaporation from natural surfaces on a regional scale.

EVAPORATION FROM BARE SOIL

Because vegetation is sparse throughout the Sudano-Sahelian region, evaporation directly from the soil surface represents a major component of the water balance. Even on land that is cropped, soil is likely to contribute at least 30% of the water lost by evaporation during the growing season (Wallace, 1991). Despite the importance of this component, attempts to estimate soil evaporation regionally are very rare.

When, after rain, water begins to evaporate from bare soil, it is replenished by the upward diffusion of liquid water from wetter soil below. Water also diffuses as vapour in a direction determined by gradients of vapour pressure; so, at night, the surface of the soil can be re-wetted by distillation from warmer, deeper soil to a cooler dry surface.

Formal equations for water, vapour, and heat transfer have been used in models that simulate evaporation from soil (e.g. Van Bavel & Hillel, 1976) but relevant soil parameters are often lacking and models become complex when they are extended to take account of processes in the atmosphere as well as in the soil. However, the evaporation from drying soil can be predicted from the PM equation if it is assumed that the resistance to the diffusion of water vapour upwards to the soil surface is proportional to the amount of water previously lost by evaporation (see Appendix 1). In effect, this is a two-layer model in which the top layer of soil, through which vapour diffuses, is assumed to be completely dry whereas the lower layer, from which water evaporates, is assumed to be at field capacity. Despite this gross simplification, the model predicts that an initial maximum rate of evaporation determined by weather (E_{s0}) gives way to a rate proportional to the square root of time as well as to E_{s0}. Dependence on the square root of time is consistent both with theory for an isothermal column of soil and with field observations on bare soil in the field where vertical gradients of temperature are always present.

Figure 1 shows how well the simple theory fits measurements of evaporation from a Vertisol and an Alfisol at ICRISAT Center, India (Vollebergh, 1984) and from a sandy Entisol at ICRISAT Sahelian Center (ISC), Niger (Wallace *et al.*, 1989). An even better fit to the observations could be obtained by accepting the conventional assumption that the evaporation rate remains constant until the surface of the soil becomes dry. It appears that this constant-rate phase lasted for about three days in the Vertisol, two in the Alfisol and one in the sandy Entisol. However, for the purposes of calculating total evaporation over periods of a week or longer, no significant error is generated by ignoring the constant-rate phase.

The constant (A) defining the relation between soil resistance and accumulated water loss (see Appendix 1) depends, *inter alia*, on maximum volumetric soil water content. This quantity appears to be around 10% for the most common soils in the Sudano-Sahelian zone, implying that the formula may need little adjustment between sites. Moreover, when annual evaporation from the sandy Entisol was estimated from rainfall recorded at ISC, doubling the value of A from 2.5 to 5 mm^2 day^{-1} increased evaporation only by about 10%. This is equivalent to increasing A/E_{s0} from 0.5 to 1 mm (in the belief that the value obtained at ISC is probably minimal), in which case total evaporation ranged from 217 mm in 1984 and 1987 to 257 mm in 1986. The range is small, possibly because the maximum rate of evaporation from a wet soil surface is less in wet, cloudy years than in drier and more sunny years.

An indirect check on the validity of these estimates of evaporation is provided by the relation between precipitation and runoff in the Sudano-Sahelian zone, plotted for three types of surface by Davy *et al.* (1976) from measurements by Dubreuil. Some of the measurements relevant to bare soil surfaces are plotted in Fig. 2(a) (open squares) along with part of a

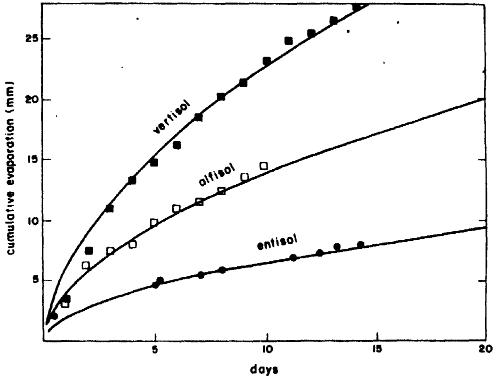

Fig. 1 Cumulative evaporation from three types of soil as a function of time after last complete wetting. The points are measurements and the three curves were fitted by using equation (A5) with an initial evaporation rate of E_{s0} = *5.0 mm day*$^{-1}$ *and with the following values for the soil-dependent parameter* A/E_{s0}*: Vertisol – 10 mm; Alfisol – 3 mm; Entisol – 0.5 mm.*

curve which Davy *et al.* appear to have drawn by eye through the complete set of points for each of the three surfaces. The curve in Fig. 2(a) is for "steppe and thorny steppe with less than 50% crop fields" and applies to basins up to 2000 km^2.

The full squares in Fig. 2(a) represent a fraction of rainfall minus estimated soil evaporation at ISC for each year from 1984 to 1989. When runoff was assumed to be half of this net loss, the six ISC points were found to be congruent with the much larger data set used by Davy *et al.* (1976). The relatively small inter-annual differences in E_s are demonstrated by the proximity of points to the straight line representing an annual soil evaporation of 230 mm. The significance of Fig. 2(b) is discussed later.

EVAPORATION FROM VEGETATION

Potential transpiration

The term "potential transpiration" is often used to describe the evaporation from a plant stand completely covering the ground and freely supplied with water – a specification drawn up by Penman (1948). A further restriction

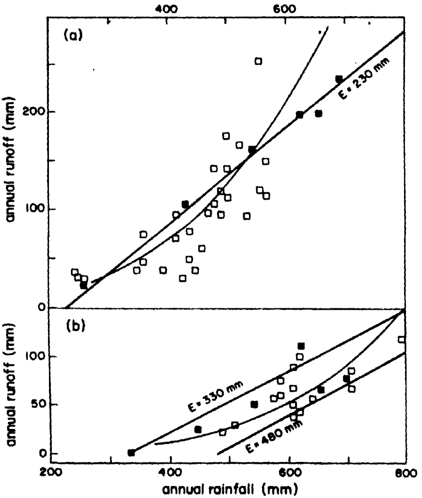

Fig. 2 Runoff as function of rainfall for catchments with little vegetation as plotted by Davy et al. (1976) from the work of Dubreuil (open points fitted by thin curve). Full points obtained as a fraction f of the difference between regional rainfall and evaporation for southern Niger, 1984–1989. In (a) f = ½ and in (b) f = ¹/₃ (see text).

sometimes imposed is that the vegetation should be "short" because turbulence is more vigorous over a tall crop than over a short one. However, because both latent and sensible heat transfer depend on turbulence and because their sum is constant, one cannot increase without the other decreasing. Commonly the ratio of the two heat fluxes is such that neither depends strongly either on wind speed or on the aerodynamic roughness of the surface and the turbulence that it generates.

Potential evaporation (E_T), as thus defined, can be determined theoretically or experimentally in a number of ways:

(a) From the original Penman (1948) formula for evaporation from open water. Penman multiplied this quantity by an empirical season-dependent factor to obtain E_T for vegetation.

(b) From the Penman-Monteith equation assuming an arbitrary value for the surface resistance r_s (see Appendix 1).

(c) From other physically-based formulae, notably that derived by Priestley & Taylor (PT) (1972) (see Appendix 1) in which radiation and temperature are the only weather variables and there is no term equivalent to a surface resistance.
(d) By direct measurements using lysimeters, by monitoring changes in soil water content, or by measuring fluxes of water vapour in the atmosphere (see Wallace, 1991).

The last method is the most reliable but there are few environments in which a uniform stand of vegetation can be readily maintained throughout the year and this problem is acute in the semiarid tropics. By skillful management, Dancette (1976) was able to obtain consistent measurements of E_T for well-watered grass grown on lysimeters at three sites in Senegal from 1968 to 1970 (Fig. 3). At the most northerly site (Richard Toll) annual rainfall is about 300 mm, at the most southerly (Sefa) it is 1300 mm and at the intermediate station of Bambey it is 650 mm. The seasonal variation in potential evaporation rate is therefore much smaller than the variation in rainfall. It is also in the opposite direction because high rainfall is associated both with high humidity and with a loss of radiant energy intercepted by cloud.

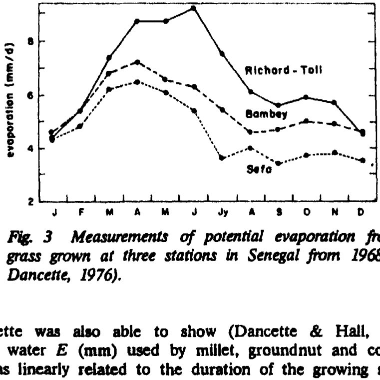

Fig. 3 Measurements of potential evaporation from well-watered grass grown at three stations in Senegal from 1968 to 1970 (from Dancette, 1976).

Dancette was also able to show (Dancette & Hall, 1979) that the amount of water E (mm) used by millet, groundnut and cowpea grown in Bambey was linearly related to the duration of the growing season d (days). Measurements for all three species fitted the relation $E = 5.7\ (d - 16)$, equivalent to a constant evaporation rate of 5.7 mm day^{-1} after an induction period of 16 days when evaporation was effectively nil.

The measurements for Bambey can be compared with potential evaporation estimates for ISC in Niger (somewhat drier) using standard climatic records in the equations associated with methods (a) to (c) (Figs 4(a) and (b)). Comparing Figs 3 and 4, it appears that the closest agreement between Bambey measurements and ISC estimates is given by the PM equation with

r_s = 60 s m^{-1}. Agreement is closer in the first half of the season than in the second, possibly because of senescence at the end of the rainy season.

Both in a very dry year (1984) and in a wet one (1988), the original Penman formula and the PM equation agree well during the rainy season. During the dry season, the Penman formula systematically predicts less evaporation because the aerodynamic term in the equation is substantially

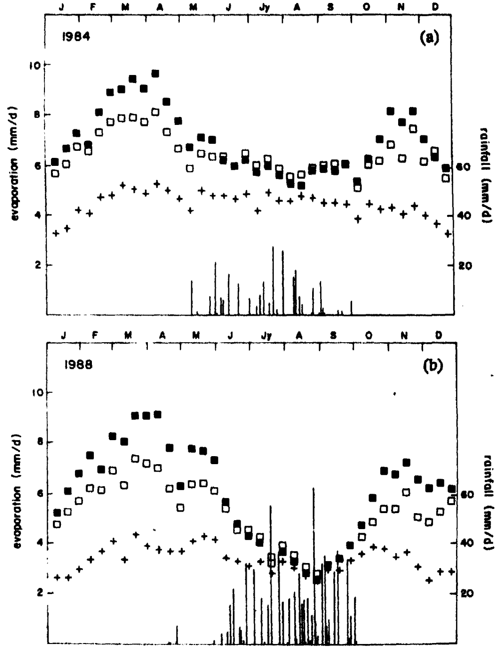

Fig. 4 Estimates of potential evaporation and measurements of rainfall for ICRISAT Sahelian Center in (a) a dry year, 1984 and (b) a wet year, 1988.
■: Penman-Monteith equation with r_s = 60 s m^{-1}; □: Penman formula; +: Priestley-Taylor equation.

underestimated as shown by Thom & Oliver (1977).

During the whole of the dry season, the PT formula grossly underestimates E_T, a conclusion already reached by Sivakumar *et al.* (1991) on the basis of West African data and by Gunston & Batchelor (1983) for 30 stations in the tropics. Even in the wet season, the PT formula systematically underestimates E_T compared with the Penman and PM equations when rainfall is deficient (1984), but the three formulae agree closely when it is abundant (1988).

Advection In principle, the PT type of equation (in which there is no saturation deficit term) cannot provide a correct estimate of evaporation unless the saturation deficit of air to which foliage is exposed is the same as the deficit in air passing over the foliage. This condition appears to be satisfied rapidly when air moves over a surface when r_s is 60 s m^{-1} or less, more slowly as r_s increases, and not at all when there is virtually no evaporation. Similarly, all versions of the Penman equation fail in conditions where there are significant horizontal gradients of temperature and humidity, as when air in equilibrium with a dry surface passes over a crop or *vice versa*.

In practice, the most important case is when dry air passes over an irrigated area. Near the upwind boundary of the area, evaporation is usually substantially faster than predicted by Penman-type equations because the flux of water vapour leaving the surface is larger than the flux at the height where temperature and vapour pressure are measured.

Several attempts have been made to explore this problem theoretically (e.g. Philip, 1987) but none includes the full complexity of changes in surface roughness or in canopy resistance which responds to spatial changes of saturation deficit.

Reliable experiments on the effects of advection on crop evaporation are also rare. In one trial on irrigated rice in southern Australia, Lang *et al.* (1974) found that the rate of evaporation 20 m from the leading edge of the field was about 120% of the rate 500 m downwind. Estimates of advection were obtained by subtracting the rate of evaporation predicted by the PT equation from the rate measured by lysimeters. The scale of advection estimated in this way was anomalously large, probably because the PT equation substantially underestimated the true rate of potential evaporation. When the PM equation was used to calculate reference evaporation, the enhancement of evaporation by advection was about 40–50% near the leading edge of the field, decreasing to about 10% at 500 m. It follows that even in small irrigation schemes with dimensions of a few hundred metres, the underestimation of evaporation using the PM equation is unlikely to exceed 20%.

Actual evaporation

The rate of evaporation from a crop stand cannot reach the potential rate, as defined above, if:

(a) The maximum rate of water uptake by roots is less than the demand for water set by the microclimate of the foliage. In this case, leaf cells lose part of their water and stomata close, bringing the rate of transpiration closer to the point at which it matches uptake.

(b) Ground cover is incomplete so that the amount of radiant energy intercepted by foliage is less than the amount received on a horizontal surface.

During a dry spell, both these constraints may operate together: reduction of leaf growth and the curling of leaves are mechanisms used by plants to minimize stress during drought. Here, processes (a) and (b) will be treated separately.

Restricted uptake Over the past 30 years, many field studies have examined the response of crops to drought in terms of the diurnal and seasonal behaviour of stomata. However, the main function of stomata during drought is to act as a valve that matches water supply and demand as closely as diurnal changes in both these quantities will allow. Supply is determined not by stomatal behaviour but by the movement and activity of root systems and the distribution of water in the layers of soil which they penetrate.

Unfortunately, much less is known about the ability of roots to capture water than about the ability of stomata to control its subsequent loss. The size and rate of growth of roots have been determined for a range of species but there are few measurements of the resistances and potentials in the soil/root system that determine the rate of water uptake.

In a simple empirical model bypassing the kind of detail that is rarely available, water uptake by an extending root system can be expressed as a function of two parameters only, both of which can be estimated from changes in soil water content throughout the growing season and at a range of depths (Monteith, 1986, 1988). The parameters are: the velocity (u) with which the maximum depth of extraction moves downward as a root system extends; and the time constant for the process of extraction at a given depth,which can usually be expressed as an exponential function of time.

Based on evidence obtained on a Vertisol and on an Alfisol at ICRISAT Center in India, u is about 3 to 4 cm per day for both sorghum and millet and time constants are of the order of 40–80 days for healthy stands of these species. Theory gives the maximum rate of extraction as $u\theta$ where θ is the maximum water available to plants per unit soil volume.

For $\theta = 0.1$, a representative value for soils in the Sudano-Sahelian zone, the maximum rate of supply is about 3–4 mm per day. This range is consistent with measurements by Azam-Ali *et al.* (1984) who grew millet on deep sand at the AGRHYMET Centre in Niamey. The crop was sown at the end of the rainy season and so depended entirely on water stored in the profile which was extracted to a depth of about 2.5 m. A transpiration rate of 3–4 mm day^{-1} is somewhat less than the potential rate of demand during the rainy season as given in Figs 3 and 4 and is much less than the demand during the dry season. It follows that both the transpiration and the growth of cereals in the Sudano-Sahelian zone will proceed at sub-potential rates unless they receive water regularly as rain or irrigation.

Incomplete ground cover When there is not enough foliage to shade the ground completely, the value of net radiant energy as used in the Penman, PM and PT equations is smaller and the value of r_s is larger than for complete cover. However, the soil surface receives more energy and is better ventilated so that evaporation from the soil surface proceeds faster than when the canopy is complete.

This type of system is difficult to explore experimentally but several theoretical schemes have been developed. The original Penman equation treats a canopy as a single wet layer exposed to the atmosphere (one layer model) whereas the PM equation also takes account of surface resistances to vapour diffusion (two-layer model: canopy and atmosphere). Shuttleworth & Wallace (1985) described a three-layer model in which the canopy was divided into an upper component (foliage) and a lower component (below foliage) which received heat and water vapour from the surface of the soil where conditions were prescribed. Choudhury & Monteith (1988) then added the soil as a fourth layer, defining its effective resistance by equation (A2) and also taking account of temperature gradients in the soil (which the treatment in the Appendix does not).

Common to three- and four-layer models is the conclusion that, when cover is incomplete, the wetness of the soil surface plays an important part in determining the saturation deficit of air surrounding foliage and the relative amount of total evaporation contributed by transpiration (Ritchie, 1983). Water that evaporates from soil beneath a crop should not be regarded as entirely wasted because the humidification of air within the canopy reduces the demand for water imposed on leaves by their microclimate and therefore allows stomata to open more widely, with the consequence of a faster photosynthetic rate.

REGIONAL EVAPORATION

In principle, evaporation from a region encompassing many types of land surface can be estimated in two distinct ways: by assessing the loss of water from each component separately, using one of the methods already outlined; or by measuring both spatial and temporal changes in the amount of water vapour held in the atmosphere.

In practice, however, working with individual components is rarely feasible because surface characteristics, especially wetness, are unknown. Moreover, accurate figures for the fraction of land under different types of management are not usually available in unpopulated regions though some can be derived from satellite images.

As an example of the second and less common method, Guowei & Yi (1989) estimated the water balance of mainland China from radiosonde ascents made twice a day at 53 stations. However, this type of estimate has not been used operationally because of the large number of measurements needed to assess spatial and temporal changes of water content.

Bouchet (1963) suggested that regional evaporation could be estimated by making the intuitive assumption that actual and potential rates of evaporation are "complementary", i.e. that their sum is constant. Although Morton (1983) has achieved some success with this method by a series of empirical modifications to the Penman equation, it is not supported by the behaviour of the Convective Boundary Layer (the lowest 1–2 km of the atmosphere) as analysed by McNaughton & Spriggs (1989).

In a new method for estimating regional evaporation described in the Appendix, it is suggested that the so-called "constant" of the Priestley Taylor equation can be treated as a simple function of mean daily saturation deficit divided by Δ, the rate at which saturation vapour pressure changes with temperature (at mean air temperature). Using this assumption to write equation (A5), daily rates of evaporation at ISC for two contrasting years were estimated: 1984 which was exceptionally dry (335 mm) and 1988 which was wet (563 mm).

In preliminary calculations, the evaporation for 1984 substantially exceeded rainfall, probably because the rain recorded at ISC was less than the regional average (whereas temperature and saturation deficit change relatively little from station to station in the same region). The ISC temperature and vapour pressure record was used along with the average rainfall for three stations close to the River Niger: Say (20 km from ISC), Tillabery (155 km from ISC) and Gaya (335 km from ISC).

The bottom half of Figs 5(a) and (b) gives daily rainfall and 5-day means of estimated daily evaporation for the region. The upper half of each figure contains the predicted daily change of soil water deficit for a layer of soil with a specified maximum deficit. This value and the value of D_m in equation (A7) were obtained as follows.

Because the regional mean rainfall for 1984 was only 335 mm and because daily totals exceeding 20 mm were rare at the four stations used, runoff was assumed to be negligible in this year. The value of D_m/Δ used in equation (A7) and the value of the maximum deficit were then adjusted until two conditions were satisfied:

(a) The annual total of evaporation should be close to the annual rainfall. A value of $D_m/\Delta = 12.3$°C gave an evaporation of 341 mm against rainfall of 335 mm.

(b) The amount of soil water carried over from 1984 to 1985 should be zero. This gave the maximum deficit as about 30 mm, appropriate for bare soil but less than would be expected for natural vegetation and an order of magnitude too small for a crop with a vigorous root system. However, the value may well be representative of a region in which vegetation occupies a relatively small fraction of the total land surface.

In Fig. 5(a), the (actual) evaporation rate barely reaches 3 mm day^{-1} and follows rainfall closely so that cumulative rainfall and cumulative evaporation are highly correlated. Wallace *et al.* (1990a) reported somewhat higher values of evaporation from bare soil and from fallow bushland: between 3 and 4 mm day^{-1} over a period of five days in September of 1986. For most of the dry 1984 season, the amount of water held in the soil profile (i.e. 30 mm minus the soil water deficit) is less than 20 mm, implying a major shortage of water for agricultural production.

The same values of D_m/Δ and maximum deficit were used to estimate

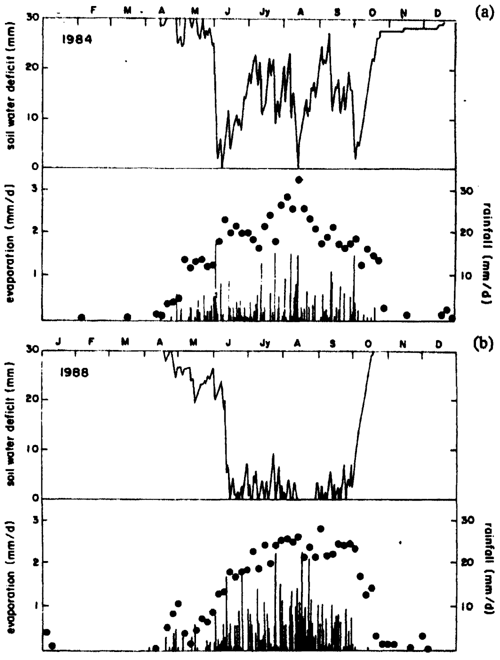

Fig. 5 Lower part of each graph: rainfall (bars) and estimated actual evaporation for southern Niger (●). Upper part: moisture deficit in soil layer containing 30 mm water at field capacity. Results are shown for both (a) a dry year, 1984 and (b) a wet year, 1988.

regional evaporation for 1988 when the mean rainfall was 563 mm (Fig. 5(b)). Again, the seasonal pattern of evaporation matches rainfall; but in contrast to 1984, available soil water exceeds 30 mm for most of the season and the difference of 238 mm between rainfall and evaporation implies that a substantial amount of water was available for runoff and recharging groundwater.

As a further, indirect check on the validity of these estimates of regional evaporation, a fraction of the difference between rainfall and evaporation was plotted against rainfall (Fig. 2(b)). In this case, assuming that the fraction was one third (compared with a half for bare soil) gave a set of annual values (full squares) consistent with the observations reported by Davy *et al.* (1976) for "thorny steppe, arboraceous steppe and savanna with at least 50% crop fields" (open squares and fitted curve) Evaporation is not independent of rainfall as in Fig. 2(a) but ranges from about 330 to 480 mm.

CONCLUSION

In this review, several ways have been outlined of establishing how the rain that falls in the Sudano-Sahelian zone is subsequently returned to the atmosphere by evaporation at the point where it falls. Progress in measuring and estimating evaporation, can be summarized in terms of three spatial scales. First, on the scale of 0.01 to 1 km^2, evaporation from bare soil or from vegetation can be determined from continuous measurements of vapour flux over uniform surfaces; from intermittent measurements of soil water content (where uniformity is again important); or by appeal to the PM equation provided a surface resistance can be chosen with confidence. Wallace (1991) gives relevant technical details.

Second, on the much larger scale of 10^4 to 10^7 km^2 which includes the entire Sudano-Sahelian zone, a net input or output of water vapour could, in principle, be obtained from atmospheric soundings but the number of radiosonde stations currently operating in the zone is too small to hold errors within acceptable limits. An alternative is to use satellite images to estimate rainfall (from cloud top temperatures (Milford & Dugdale, 1990)), evaporation (from surface temperatures (Seguin *et al.*, 1989)) and water near the soil surface (from microwave emissivity). A vast data base already exists but cannot be properly exploited until more attention is paid to collecting "ground truth" and to developing the mechanistic models that are needed to minimize empiricism.

The gap between these two scales – roughly from 1 to 10^4 km^2, will shortly be filled by an international energy balance experiment (HAPEX-II-Sahel), to be mounted in Niger in 1992. HAPEX-II-Sahel should provide much of the information needed to interpret and exploit records from satellites. It should also generate new insights into the behaviour of the Convective Boundary Layer in the Sudano-Sahelian zone as a basis for developing a better method of estimating actual evaporation from climatic records than the one I have proposed here.

This meeting provides a timely opportunity to review what we now know or do not know about the water balance of the Sudano-Sahelian zone. I hope that material from HAPEX-II-Sahel will substantially advance our knowledge for the benefit of all those who live and work in this taxing but precious environment.

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APPENDIX 1:

EXTENSIONS OF THE PENMAN-MONTEITH (PM) EQUATION

General form

A general form of the PM equation for the loss of latent heat from a surface of specified resistance is:

$$LE = \frac{\Delta H + \rho c_p D/r_a}{\Delta + \gamma\,(1 + r_s/r_a)} \tag{A1}$$

Parameters not defined previously are:

Δ = increase of saturation vapour pressure with temperature (Pa°C^{-1}),

γ = psychrometer constant (66 Pa°C^{-1}),

ρc_p = volumetric specific heat of air (J m^{-3} °C^{-1}).

Equation (A1) is obtained by eliminating surface temperature from a set of primary equations describing the heat balance of a surface such as soil or a vegetation canopy and the exchange of sensible and latent heat between that surface and the air passing over it (Monteith, 1981).

Allen *et al.* (1989) recently compared predictions of evaporation using five versions of the Penman formula (including the original) with measurements of evaporation from grass or lucerne grown on well-maintained lysimeters in different climates.

The PM equation gave the best agreement between prediction and measurement both in humid and in dry climates and for months when the evaporation rate reached a maximum as well as for annual totals. On the basis of this study, a group of experts meeting at FAO in May, 1990 recommended that the formula should be adopted as a basis for calculating crop water reqirements in place of the formula referred to as FAO-24 (Doorenbos & Pruitt, 1977). Crop coefficients will continue to be used, as by Abdulmumin & Misari (1990) for northern Nigeria, until they can be replaced by seasonally-changing values of r_s.

Evaporation from soil

To apply equation (A1) to soil, the resistance to the upward diffusion of water vapour within the soil profile is assumed to increase in proportion to the amount of water lost by evaporation (E_s) since the last complete wetting, i.e.

$$\int E_s \, dt \qquad \text{(A2)}$$

where m is a constant which is a function of the soil diffusivity. Substituting for r_s in equation (A1) and rearranging terms (see Monteith, 1981) gives:

$$E_s = \{2t/A + 1/E_{s0}^2\}^{-0.5} \qquad \text{(A3)}$$

where

$$A = r_a(\Delta + \gamma)\, E_{s0}/\gamma m \qquad \text{(A4)}$$

and E_{s0} is the initial maximum rate of soil evaporation when $r_s = 0$. Integrating equation (A3) gives the cumulative evaporation as:

$$\int E_s \, dt = \{2At + A^2/E_{s0}^2\}^{0.5} - A/E_{s0} \qquad \text{(A5)}$$

Evaporation from vegetation

When equation (A1) is applied to vegetation, r_s is identified as the resistance of a leaf canopy, treated, in effect, as one large leaf. Despite this sweeping simplification, the equation has performed consistently when applied to many types of crop stand and even to forests, provided the ground is well covered by foliage. Much more complex schemes are needed to describe and explore how transpiration is distributed layer by layer within a canopy (Raupach & Finnigan, 1988).

When vegetation is sparse and the soil surface is wet, so that both leaves and soil make a significant contribution to the total loss of water from the system, the simplest assumption is that the individual components depend on the relative amounts of radiant energy they intercept. This method, originally suggested by Ritchie (1972), has been widely used in models of crop growth and water use, but is subject to error which can be large for sparse crops growing on dry soil (Wallace *et al.*, 1990b).

Priestley-Taylor (PT) equation

McNaughton (1976) showed that when air passes over an extensive and uniform land surface, the vertical gradient of saturation deficit decreases. When it vanishes, equation (A1) reduces to $LE = [\Delta/(\Delta + \gamma)]H$. For well-watered vegetation, LE often exceeds this estimate by a factor α between 1.2 and 1.3. The PT equation (Priestley & Taylor, 1972) is:

$$LE = \alpha[\Delta/(\Delta + \gamma)]H \qquad \text{(A6)}$$

Regional evaporation

Equation (A1) has rarely been used to estimate regional evaporation because of the difficulty of specifying an appropriate value of r_s, especially when evaporation is limited by the supply of water to different types of surface. A new paradigm is needed to make progress. Instead of regarding daily mean values of temperature and vapour pressure as quantities that determine evaporation rate, can they be treated as consequences of that rate?

Provided daily mean values of radiation and windspeed do not change much with time, a decrease in the supply of water for evaporation would be expected to increase mean air temperature and to decrease vapour pressure so that the saturation vapour pressure deficit would increase. In the extreme, found only in arid and semiarid regions, the evaporation rate approaches zero at the height of the dry season at which time the value of D would be expected to reach a maximum.

Exploring this line of reasoning with the help of daily weather records from ICRISAT Sahelian Center, the quantity D/Δ was found to be a more consistent index of atmospheric dryness than D alone (Fig. A1). For reasons to be discussed elsewhere, the so-called constant (α) of the PT equation was assumed to decrease

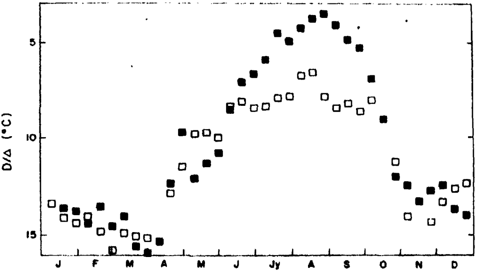

Fig. A1 Value of D/Δ estimated from climatic records at ICRISAT Sahelian Center in 1984 (□) and 1988 (■). Note similarity of records in dry season but not in rainy season. The vertical scale is inverted to emphasize the inverse relation between D/Δ and water supply or evaporation rate.

linearly with D/Δ from a maximum value of α_m when D is zero to zero when D has a maximum value D_m. The relation can be written:

$$\alpha/\alpha_m = [1 - (D/\Delta)/(D_m/\Delta)] \qquad \text{(A7)}$$

It was then possible to eliminate D from equations (A1) and (A5) to establish the dependence of E on r_s. This relation was found to be consistent with field observations of E and r_s from diverse sites (Monteith, 1965), provided α_m was set at about 1.6. Reassuringly, McNaughton (1989) was able to demonstrate that the same observations were consistent with de Bruin's (1983) model of the Convective Boundary Layer. (This is the layer, about 1–2 km deep, within which heat and water vapour are exchanged with the earth's surface.) Further evidence for the validity of equation (A5) at least in the Sudano-Sahelian zone, is given in the main text.