A Climatonomic Description of the Surface Energy Balance in the Central Sahel. Part II: The Evapoclimatonomy Submodel

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ABSTRACT

In Part I of this article we presented a study of the shortwave radiation budget in the West African Sahel, using Lettau's climatonomy model. In Part II, we apply the second of Lettau's submodels, evapoclimatonomy, to quantifying the surface water balance in the Sahel near Niamey, Niger. The model uses monthly means of ground-absorbed solar radiation and precipitation as input, and it predicts evapotranspiration, runoff, and exchangeable soil moisture. Sensitivity studies show that the model results are most affected by precipitation and by two prescribed model parameters, an evaporivity (e^*) and a residence time (t^*). Model results suggest that in the Sahel, rainfall and not insolation is the limiting factor in determining water balance characteristics.

1. Introduction

The West African Sahel has suffered extensively from drought in recent years. In most regions of the world below-average rainfall usually occurs for only a few consecutive years, but in the Sahel prolonged dry periods with time scales of one or two decades are common. Conditions of above-average rainfall likewise tend to last for a decade or more. The cause of this seemingly unique persistence is as yet unknown, but a number of researchers (see review in Nicholson 1989) have suggested that it may relate to a feedback between the land surface and atmosphere. According to this hypothesis, Sahel drought is self-reinforcing through the changes it evokes on the land surface (e.g., reduced vegetation cover and soil moisture, increased albedo) and their impact on the atmosphere via surface fluxes. Thus, a drought might be triggered by the large-scale general atmospheric circulation but locally maintained by such feedback until the large-scale circulation is sufficiently conducive to rainfall that the feedback effect is overridden.

In order to test this hypothesis, the Sahelian surface energy balance must be quantified. To do this, we are applying Lettau and Lettau's (1975) three-part climatonomy model to describe the energy balance near Niamey, Niger, in the central Sahel (13.29°N, 2.10°E). The three submodels respectively assess the shortwave radiation balance, surface water balance, and the surface thermal balance (Table 1).

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In a companion article (Lare and Nicholson 1990), we present the derivation and results of the shortwave submodel. In the current article, we present the evapoclimatonomy submodel. It predicts evapotranspiration, runoff, and exchangeable soil moisture from two forcing functions: mean monthly precipitation and ground-absorbed radiation. We first describe, in section 2, the general characteristics of Sahel climate, including the two forcing functions. Characteristics of greatest relevance to prescribing model input are emphasized. Section 3 presents the general development of the evapoclimatonomy model and section 4 presents specific input data for our study. Model results and sensitivity studies are described in sections 5 and 6, respectively.

2. Climatic background

The surface water balance is primarily a function of precipitation and solar radiation. In the Sahel, mean annual rainfall ranges from 100 mm in the north to 1200 mm in the south, concentrated in a rainy season ranging from 2 to 5 months (Nicholson 1981). August is the month of peak rainfall. At Niamey mean annual rainfall is 564 mm, most of which falls in the months of May to September (Fig. 1, Table 2); over a third of the annual total falls in August.

The solar energy available to drive the surface hydrologic processes, i.e., the amount of solar radiation absorbed by the ground, can be expressed as $(1 - a_s)G^*$, where a_s is surface albedo and G^* is global radiation. This parameter was calculated with the shortwave submodel (Lare and Nicholson 1990). Here G^* is primarily determined by irradiance (I') at the top of the atmosphere (Fig. 1) and cloudiness (c) (Table 2), both of which peak during the rainy season.

TABLE 1. Summary of climatonomy model characteristics (from Lettau and Lettau 1975).

Submodel	Forcing functions	Response/output
I. Shortwave radiation	Irradiance, I' (top of atmosphere)	Top or planetary albedo, A^* Absorption by submedium, $(1 - a_s)G^*$
II. Evapoclimatonomy	Precipitation, P and Absorption by submedium, $(1 - a_s)G^*$	Soil moisture, m Runoff, N Evapotranspiration, E
III. Thermal radiation	Absorption by submedium minus evapotranspiration, $(1 - a_s)G^* - E$	T_{air}, T_{sfc} Surface fluxes Net radiation, R_{net}

Other factors include attenuation by precipitable water (w), which is also highest during the rainy season, and atmospheric turbidity (β), which attains a maximum from March to May and is relatively high in June and July (Table 2). Surface albedo (a_s) is dictated by vegetation cover; hence, it is lowest from July to October. The net result of these various factors is that ground-absorbed radiation attains a maximum in February, March and April and a minimum from June to August (Fig. 1).

3. The basic model

The evapoclimatonomy submodel is in essence a numerical solution to the integration of the simple hydrologic balance equation:

$$P = E + N + dm/dt \qquad (1)$$

where P is rainfall, E is evapotranspiration, N is runoff and dm/dt is the change in soil moisture storage. The method of solution via integral calculus distinguishes climatonomy from the algebraic accounting of inputs and withdrawals which characterized earlier water balance approaches (e.g., Thornthwaite and Mather 1955; Penman 1956). The model requires input of mass and solar energy (i.e., monthly precipitation and solar radiation absorbed by the ground), referred to by Lettau as "forcing functions." Model output includes predictions of monthly runoff, soil moisture storage, and evapotranspiration.

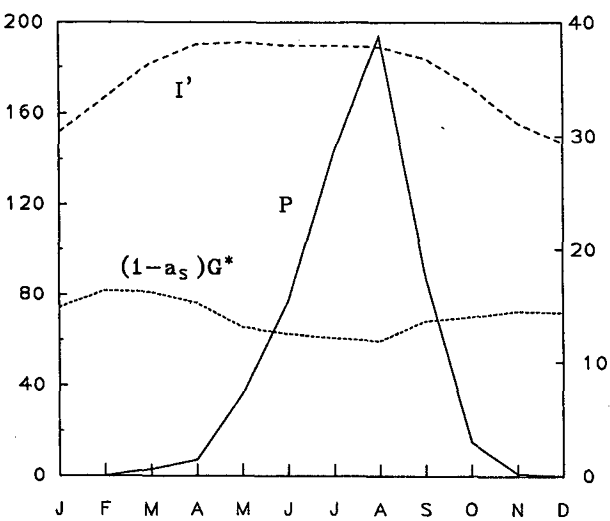


FIG. 1. Forcing functions for evapoclimatonomy submodel: ground-absorbed solar radiation $(1 - a_s)G^*$ (MJ m^{-2} day^{-1}) (as calculated by the shortwave radiation submodel) and precipitation P (mm month^{-1}). The forcing function for the shortwave submodel, irradiance I' (MJ m^{-2} day^{-1}), is included for comparison.

The basic premise of the model is as follows:

1) For a stable climate the long-term mean of dm/dt is zero, i.e., there is no net change in soil moisture storage. Thus:

$$\bar{P} = \bar{E} + \bar{N}. \qquad (2)$$

The larger the region evaluated and, up to a limit, the longer the time period considered, the more valid this assumption.

2) Evapotranspiration and runoff can both be partitioned into immediate and delayed processes:

$$E = E' + E'' \qquad (3)$$

$$N = N' + N'' \qquad (4)$$

where the single primes denote "immediate" processes occurring in the same month as the rainfall and the double primes denote runoff and evaporation of rain which fell in previous months.

The physical rationale for assumption 2) is the need to distinguish between the time variations of E and N coupled with concurrent precipitation and those supplied by subsurface moisture independent of rainfall.

The model requires that two of the three annual means ($\bar{P}$, $\bar{E}$, $\bar{N}$) be known. Precipitation is a readily available quantity and the others can be calculated with the aid of characteristic climatic ratios defined as:

$$B = \bar{Q}/L \cdot \bar{E} \qquad (5)$$

$$C = \bar{N}/\bar{P} \qquad (6)$$

$$D = \bar{R}_{net}/L \cdot \bar{P} \qquad (7)$$

where Q is sensible heat transfer, L is the latent heat of vaporization, R_{net} is net radiation at the surface (Budyko 1986), B is the Bowen ratio, and C and D are

TABLE 2. Forcing and input to the evapoclimatonomy submodel. $(1 - a_s)G^*$ = ground absorption of solar radiation (MJ m^{-2} day^{-1}) as calculated by the shortwave radiation submodel (Lare and Nicholson 1990); c = cloudiness (percent); w = precipitable water (cm); β = turbidity coefficient; a_s = surface albedo; P = precipitation (mm month^{-1}); e^* = evaporivity; t^* = residence time (month).

Variable	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Ann
$(1 - a_s) G^*$	14.9	16.4	16.3	15.3	13.1	12.5	12.2	11.9	13.7	14.0	14.5	14.4	14.5
c	45	49	61	68	73	70	76	78	71	63	50	51	63
w	1.4	1.3	1.6	2.3	3.6	4.2	4.5	4.7	4.7	3.9	2.2	1.6	2.9
β	0.37	0.31	0.63	0.69	0.69	0.51	0.51	0.44	0.36	0.45	0.40	0.33	0.47
a_s	0.30	0.30	0.31	0.31	0.32	0.30	0.23	0.20	0.20	0.22	0.26	0.28	0.27
P	0	0	3	7	36	77	143	194	88	15	1	0	47
e^*	0.80	0.80	0.80	0.70	0.65	0.60	0.50	0.50	0.70	0.75	0.80	0.80	0.50
t^*	1.5	1.5	1.5	1.5	1.5	1.5	1.5	1.5	1.5	1.5	1.5	1.5	1.5

termed the runoff and dryness ratios, respectively. Using these definitions, the basic surface energy balance equation

$$L\cdot\bar{E} = \bar{R}_{\text{net}} - \bar{Q} \tag{8}$$

can be rewritten as

$$L\cdot\bar{E} = \bar{R}_{\text{net}}/(1 + B) \tag{9}$$

where

$$\bar{E} = \bar{P} - \bar{N} = (1 - C)\bar{P}. \tag{10}$$

This produces the relationship

$$(1 + B)\cdot(1 - C) = D. \tag{11}$$

Examining a large number of watersheds, Budyko determined a semiempirical relationship between the runoff and dryness ratios, such that

$$C = 1 - \tanh D. \tag{12}$$

If either ratio is known, the other two can be calculated. We estimated D, using the method described by Lettau (1978) and Kutzbach (1980) to approximate $\bar{R}_{\text{net}}$. This allowed the calculation of C and, thus, $\bar{N}$ [from Eq. (6)] and $\bar{E}$ [from Eq. (2)].

Equations (3) and (4) allow for the definition of a quantity called "reduced precipitation" (P'), which represents the amount of the mass input that is not lost by surficial processes of immediate runoff and evapotranspiration and thus is available for soil moisture storage. Therefore,

$$P' = P - N' - E'. \tag{13}$$

In view of the preceding discussion, the soil moisture storage term is reduced to

$$dm/dt = P' - (N'' + E''). \tag{14}$$

The above equation (14) is solved using two empirical concepts, "evaporivity" e^* and a "residence time" t^*, which are subsequently defined and used to partition the immediate and delayed processes.

To facilitate the solution of (14), a number of assumptions are made in order to parameterize the immediate and delayed processes of runoff and evapotranspiration. First, delayed runoff N'' and delayed evapotranspiration E'' are assumed to vary in direct proportion to soil moisture m (Lettau 1969) so that

$$N''(t) = \overline{N''}m(t)/\bar{m} \tag{15}$$

$$E''(t) = \overline{E''}m(t)/\bar{m}. \tag{16}$$

The sum of the above equations yields an equation that defines a new characteristic dimensional parameter:

$$N'' + E'' = (\overline{N''} + \overline{E''})m/\bar{m} = m/t^* \tag{17}$$

where $t^* = \bar{m}/(\overline{N''} + \overline{E''})$ denotes a time interval most conveniently expressed in months. Physically, t^* can be interpreted as a "residence time" or "turnover period" that is characteristic of a basin and signifies the time required for a volume of water equal to the annual mean of exchangeable soil moisture to be depleted by the "delayed" processes of runoff and evapotranspiration. Thus, t^* could be expected to be 2 or 3 months or more.

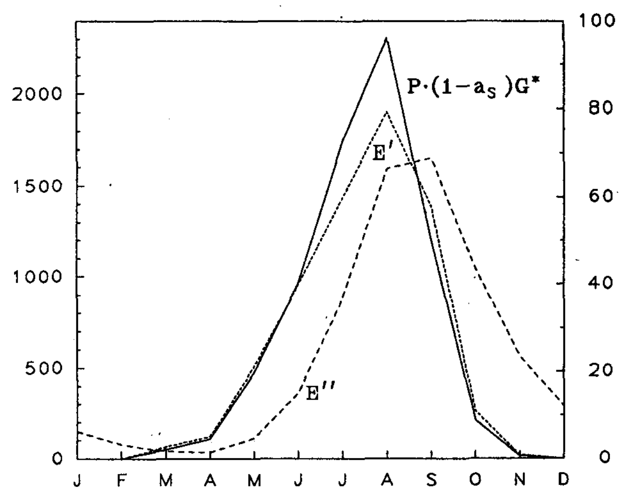


FIG. 2. Comparison between the product of the two evapoclimatonomy forcing functions $P\cdot(1 - a_s)G^*$ and immediate and delayed evapotranspiration, E' and E'', respectively. Units are mm month^{-1} for evapotranspiration and W m^{-1} for the forcing function.

TABLE 3. Monthly mean evapoclimatonomy results for Niamey, Niger. The values of soil moisture m, and the immediate, delayed and total processes of runoff (N', N'' and N) and evapotranspiration (E', E'' and E) have units of mm month^{-1}. Annual values are expressed as a monthly mean.

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Ann
m	9.4	4.8	2.6	2.1	7.2	22.7	55.6	100.8	104.6	66.2	35.6	18.3	35.8
N'	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
N''	0.1	0.0	0.0	0.0	0.1	0.2	0.4	0.8	0.8	0.5	0.3	0.1	0.3
N	0.1	0.0	0.0	0.0	0.1	0.2	0.4	0.8	0.8	0.5	0.3	0.1	0.3
E'	0.0	0.0	2.7	5.1	21.1	39.7	59.8	79.3	57.9	10.9	0.8	0.0	23.1
E''	6.2	3.2	1.7	1.4	4.7	15.0	36.7	66.4	68.9	43.6	23.5	12.1	23.6
E	6.2	3.2	4.4	6.5	25.9	54.7	96.4	145.8	126.8	54.5	24.3	12.1	46.7

The parameterization of immediate processes is likewise achieved using proportionalities based on the annual balance equations. We have simplified the calculations by assuming $N' = 0$, which is generally the case in semi-arid regions such as the Sahel (Lettau 1969; Eagleson and Segarra 1985). For immediate evapotranspiration (E'), Lettau (1969) defined an additional characteristic parameter (e^*), termed the "evaporivity," which is applied in the following equations:

$$E'(t) = e^* \cdot P(1 - a_s)G^* / \overline{(1 - a_s)G^*} \quad (18)$$

$$\overline{E'} = \overline{e^* \cdot P}. \quad (19)$$

In the above, $(1 - a_s)G^*$ and $\overline{(1 - a_s)G^*}$ are the monthly and annual "forcing functions," i.e., energy input via absorbed solar radiation. Thus, e^* is a nondimensional measure of the capacity of the land surface to use a portion of the monthly solar radiation to evaporate precipitation received during the same month. Tentative evaluations by Lettau (1971) and others have suggested that e^* will normally be between 0.4 and 0.8. Lettau chose a value of 0.7 for New Delhi, a semiarid subtropical climate with summer rainfall, comparable to Niamey.

The concepts described above transform the basic budget equation (1) into

$$\begin{aligned} P - E' - N' &= E'' + N'' + dm/dt \\ &= m/t^* + dm/dt. \quad (20) \end{aligned}$$

Subtracting the annual average of all terms from the above equation (20) yields

$$p'(t) = (m - \bar{m})/t^* + d(m - \bar{m})/dt \quad (21)$$

where p' stands for the time series $P - E' - N' - (\overline{P - E' - N'})$. The ordinary differential equation is solved by

$$m - \bar{m} = e^{-t/t^*}\left[\text{const} + \int e^{t/t^*} p' dt\right] \quad (22)$$

where "const" denotes an integration constant that is determined by the requirement that the bracketed value (i.e., the annual mean) of the right side must vanish for a stable climate. This equation is solved using stepwise integration, starting with an assumed initial value (m_1) of soil moisture. In a stable climate, the value for the thirteenth month (m_{13}) must equal m_1. The procedure is iterated until reasonable agreement occurs, i.e., until $|m_{13} - m_1| < 0.005$ mm. Usually only two or three iterations are necessary.

4. Input parameters for the evapoclimatonomy submodel

Several variants of the evapoclimatonomy submodel have been used (Lettau 1969; Lettau and Baradas 1973; Lettau and Lettau 1975); each require somewhat different initial input. The choice depends on the nature of the climate being simulated and on data availability. In our case, e^* and t^* were initially prescribed and N'

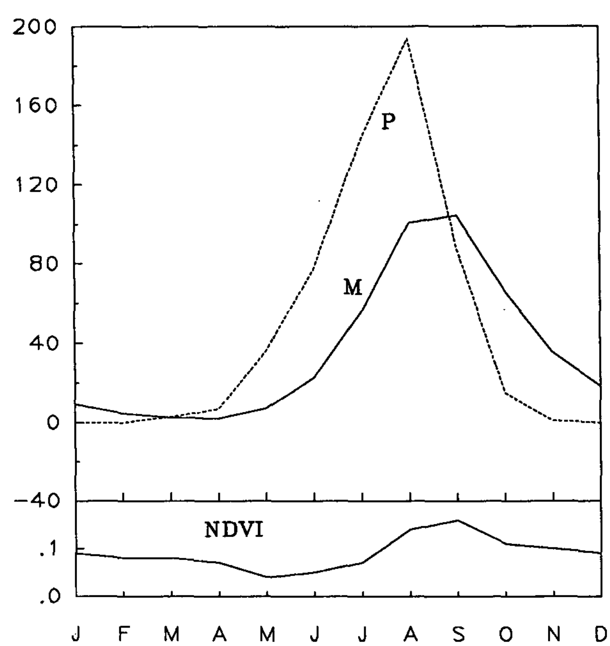


FIG. 3. Comparison of calculated soil moisture m, precipitation P, and the normalized difference vegetation index (NDVI) from NOAA satellites. The last parameter represents the ratio of radiation absorption in the red and near-infrared bands of the solar spectrum and provides an estimate of the vigor of vegetation growth. P has units of mm month^{-1} and m has units of mm.

was assumed to be zero. An estimate of N' can be based on climatic considerations or, for a region where the basin hydrologic balance is well known, on ancillary data for the region.

As with Pinker and Corio (1987) and the various studies by Lettau and coworkers, we have based our estimates of e^* and t^* on soils, topography, results of previous applications of climatonomy and other water balance studies for the region. We have chosen 1.5 months for t^* (Table 2). As pointed out by Pinker and Corio (1987), monthly variations of e^* are to be expected. We have selected appropriate values based on the typical ranges given by Lettau (1971), values used by Pinker and Corio (1987) and the seasonal variation of global radiation with respect to the seasonality of rainfall at Niamey. Our monthly values of e^* range from 0.5 to 0.8, with low values corresponding to the rainy season (Table 2) when the available radiant energy is at a minimum.

The forcing functions of the evapotranspiration submodel are rainfall and ground-absorbed solar radiation (Fig. 1, Table 2), the latter having been calculated by the shortwave submodel. The two forcing functions are out of phase with each other. Rainfall has a sharp peak in July and August (Fig. 1) and remains near zero from January through May. Although insolation also peaks during and just prior to the rains, the ground absorption peaks in February through April. It is lowest during the rainiest months because of high cloud cover and atmospheric water vapor. The total model forcing (Fig. 2) is roughly the product of rainfall and ground-absorbed solar radiation.

5. Model results

The model output consists of soil moisture m; immediate, delayed and total evapotranspiration E', E'', and E, respectively; and the equivalent runoff parameters N', N'' and N (Table 3 and Figs. 2 and 3). The seasonal variation of immediate evapotranspiration E' closely parallels that of the total forcing function (Fig. 2); both have a sharp maximum during July and August, the two rainiest months. Delayed evapotranspiration peaks in August and September, lagging rainfall by about 1 month. Its seasonal variation reflects that of soil moisture (Fig. 3). Delayed exceeds immediate at the end of the rainy season, (i.e., September) and throughout the dry season to February, but immediate evapotranspiration is the dominant process late in the dry season and during most of the rainy season. Both attain maxima between 65 and 80 mm month^{-1}.

The model predicts very little runoff; a maximum of 0.8 mm month^{-1} occurs in August and September and is all delayed runoff since N' is set to zero. Total runoff is zero from February through April. This is consistent with two characteristics of Niger River discharge: its relatively constant volume as it traverses the Sahel from the Niger Bend to Niamey and its December/January maximum at Niamey, after which time subsurface runoff nearly ceases (Billon 1986).

Calculated soil moisture ranges from 2 mm in April to 101 mm in August and 105 mm in September (Fig. 3). Although the lack of soil moisture measurements at Niamey precludes direct comparison with observed values, this result agrees favorably with measurements from Niono Ranch, a location to the west in Mali with a similar climatic regime (de Ridder et al. 1982). There m ranged from about 40–45 mm at the beginning and end of the rainy season (July and October) to 80–100 mm in August and September, compared to ~56–66 mm and ~101–105 mm, respectively, in our calculations for Niamey. The overall seasonal cycle of calculated soil moisture, with a minimum from January to May and an abrupt rise and maximum in August and September, is similar to that shown by Sivakumar et al. (1984) and by Sivakumar and Gnoumou (1987) for comparable stations in Mali and Burkina Faso. The delayed processes of evapotranspiration E'' and runoff N'' in Table 3, as assumed by the model, show the same seasonal variations as soil moisture, peaking somewhat later than rainfall.

It is interesting to note that the vigor of vegetation growth around Niamey, as assessed remotely by the normalized difference vegetation index (NDVI) from NOAA satellites, shows seasonal variations which strongly parallel those of soil moisture and both tend to lag Sahel rainfall by about 1 month (Fig. 3). The delayed evapotranspiration, which is largely controlled by vegetation, peaks sharply in the months of greatest vigor of growth. Soil moisture is rapidly depleted as the rains diminish in October, and vegetation growth also drops markedly.

Various climatological water and energy balance parameters provide an additional check on model results. These include the runoff and evaporation ratios (N/P and E/P), the dryness ratio $R_{net}/L \cdot P$ (where L = latent heat of vaporization and R_{net} = net radiation), the Bowen ratio B of sensible to latent heating ($Q/L \cdot E$) and the ratio of actual to potential evapotranspiration E/E_p. All but the Bowen ratio have been calculated on a monthly basis (Table 4) by assuming L

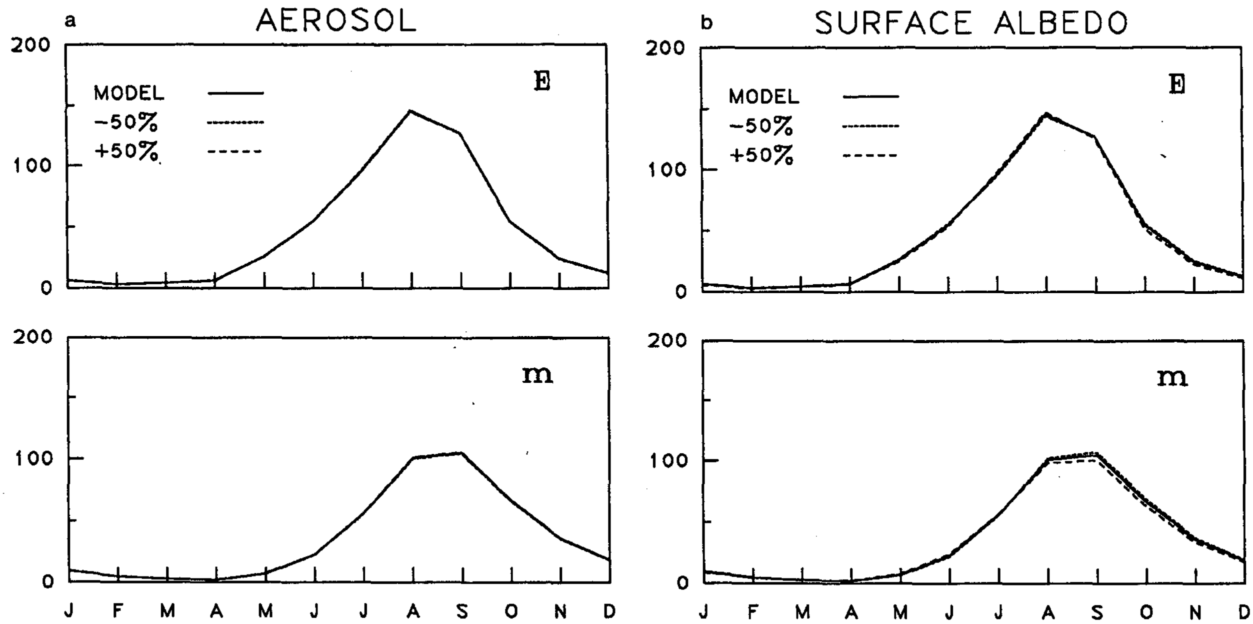
TABLE 4. Climatological water and energy balance parameters as calculated by model.

Parameter	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Ann
N/P	—	—	0	0	0.003	0.003	0.003	0.004	0.009	0.03	0.30	—	0.01
E/P	—	—	1.47	0.93	0.72	0.71	0.67	0.75	1.44	3.63	24.3	—	0.99
$R_{net}/L \cdot P$	—	—	49.5	21.4	3.7	1.6	0.9	0.6	1.6	9.1	119.4	—	2.9
E/E_p	0.03	0.01	0.02	0.03	0.11	0.28	0.56	1.00	0.82	0.29	0.12	0.06	0.23

$= 2.43 \times 10^6$ J kg^{-1} and approximating R_{net} in the manner described by Kutzbach (1980) and Lettau (1978). Potential evapotranspiration is taken from *Agroclimatological Data* (FAO 1984). The runoff ratio remains near zero most of the year but attains a value of 0.3 in November. The dryness ratio is exceedingly high during the dry season but less than 1 during the months of July and August, implying that the net radiant energy input is insufficient to evaporate all of the monthly rainfall. Actual evapotranspiration is as little as 1% of the potential during the dry season but reaches 100% during August. The annual means of N/P, E/P, $R_{net}/L \cdot P$, B, and E/E_p are 0.01, 0.99, 2.9, 1.9 and 0.23, respectively. According to Sellers (1965) and Budyko (1986), these ratios are more typical of deserts than semiarid subtropical savannas, but these values are appropriate for areas with a dryness ratio of 2.9, as is the case with Niamey.

6. Sensitivity studies

We have tested the sensitivity of the evapoclimatonomy submodel both to changes initiated in the shortwave submodel, which provides input, and to changes of evapoclimatonomy submodel parameters. In the first case, atmospheric aerosols, cloudiness, surface albedo,



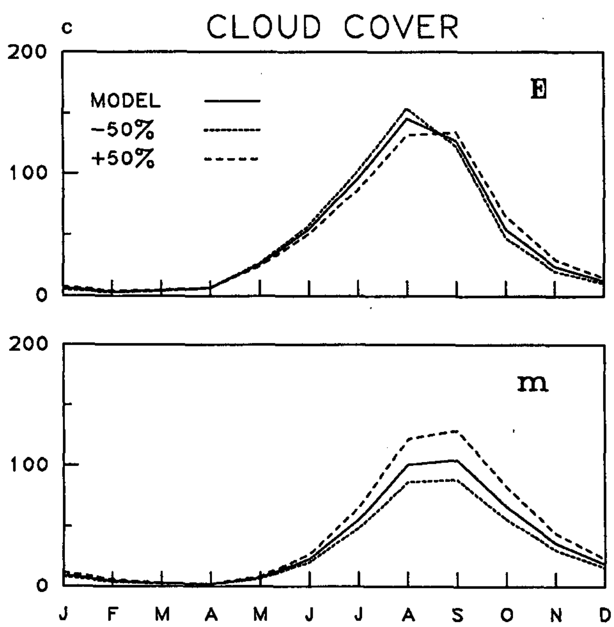


FIG. 4. (a) Effect of varying aerosol concentration by ±50% on evapotranspiration E (mm month^{-1}) and soil moisture m (mm), as calculated by evapoclimatonomy submodel. (b) As in panel (a), but varying surface albedo by ±50%. (c) As in panel (a), but varying cloud cover by ±50%.

rainfall, and ground-absorbed solar radiation were each varied by ±50%, holding other parameters constant. In the second case, e^* was varied between 0.5 and 0.9 and t^* was varied between 2 months and 4 months, in order to compare our results with those of Pinker and Corio (1987).

The sensitivity to aerosols is exceedingly low because, as tests of the shortwave submodel indicated, these have little impact on ground-absorbed solar radiation (Lare and Nicholson 1990). Changes of ±50% in aerosol concentration altered model calculated evapotranspiration and soil moisture by less than 1% (Fig. 4a).

The impact of surface albedo changes is only somewhat higher (Fig. 4b). A 20% increase/decrease was shown previously to decrease/increase ground-absorbed radiation by 4%–8% (Lare and Nicholson 1990). A 50% change in surface albedo changes the calculated soil moisture and evapotranspiration by less than 9% in all months.

The sensitivity of the evapoclimatonomy submodel to cloudiness is higher than to surface albedo (Fig. 4c). A 50% change in cloudiness alters global radiation between 26% and 47% during the wet season. The corresponding effect on soil moisture and evapotranspiration calculated by the evapoclimatonomy submodel is a change of 1%–25%, with greater effects in the wetter months.

We have also tested the model's sensitivity to the two primary forcing functions, precipitation and ground-absorbed solar radiation. A 50% change in rainfall alters both evapotranspiration and soil moisture by ≈50% (Fig. 5a), with greater rainfall increasing both E and m. An increase in absorbed radiation has no effect on either evapotranspiration or soil moisture (Fig. 5b); this suggests that under mean climatic conditions in the Sahel rainfall is the limiting factor in evapotranspiration. A 50% decrease in ground-absorbed solar radiation, however, decreases evapotranspiration by about 20%, but it has no effect on soil moisture. This asymmetric response suggests that at lower levels of insolation, both available solar energy and precipitation control the rate of evapotranspiration.

The evapoclimatonomy submodel shows considerably greater sensitivity to evaporivity e^* and residence

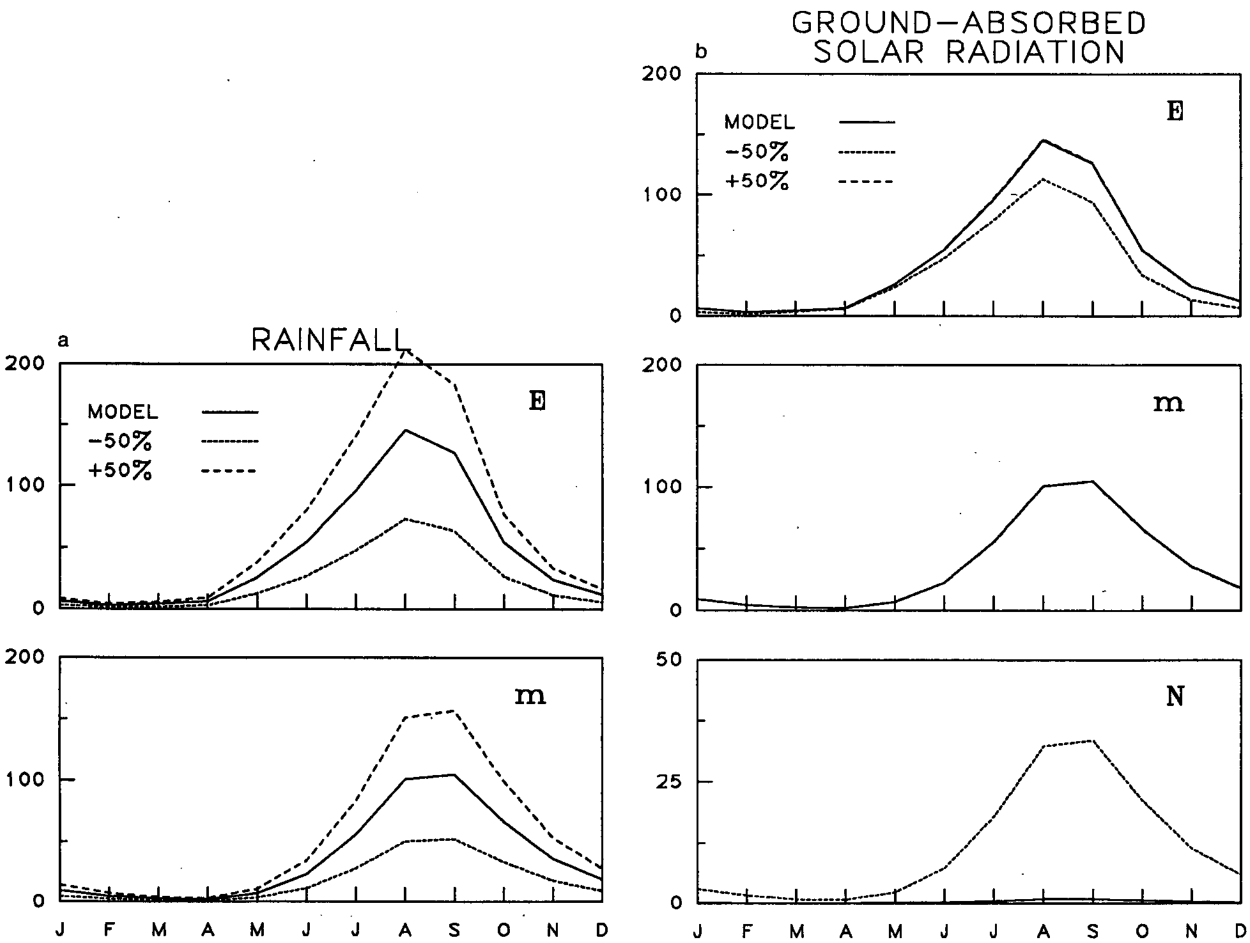


FIG. 5. (a) Effect of varying rainfall by ±50% on evapotranspiration E (mm month^{-1}) and soil moisture m (mm), as calculated by the evapoclimatonomy submodel. (b) As in panel (a), but varying ground-absorbed solar radiation by ±50%.

time t^* (Fig. 6). During the wet season a one month change in t^* (holding e^* constant) alters evapotranspiration by up to ~10% and soil moisture by 10%–30%; increasing t^* decreases E but increases m. The dry season effect is different: higher residence time increases both E and m since the delayed process E'' is enhanced by higher soil moisture content. The dry season changes in E and m are, however, only a few mm. Varying e^* (with constant t^*) modifies only the amplitude of the seasonal cycle of evapotranspiration and not the annual mean. The effect on soil moisture is large, however, and an increase of e^* from 0.5 to 0.7 decreases soil moisture during the wettest months by ~30% or ~30 mm. Our results are comparable to those of Pinker and Corio (1987). In their climatonomy model for Kansas, a one-month change in t^* resulted in a 33% change in soil moisture. As in our study, e^* affected only the seasonal cycle of evapotranspiration; for example, increasing e^* from 0.5 to 0.9 increased wet season evaporation by up to ~20 mm month^{-1} but reduced it by up to 20 mm month^{-1} during the dry season.

7. Summary and conclusions

The forcing functions of the evapoclimatonomy model are monthly precipitation and ground-absorbed solar radiation, as calculated by the shortwave climatonomy submodel (Lare and Nicholson 1990). The model results show that precipitation is almost completely accounted for by evapotranspiration and soil moisture storage, with little loss as runoff. The evapotranspiration is almost equally partitioned into immediate and delayed processes, and the delayed component peaks in September, approximately one month after rainfall has achieved its maximum. Soil moisture likewise peaks in September. Total evaporation ranges from near zero in February through April to nearly 150 mm month^{-1} in August. Soil moisture is close to 100 mm in August and September and exceeds 20 mm from June through November.

The model results show little sensitivity to shortwave radiation or factors influencing it, such as cloudiness, aerosols and surface albedo. This suggests that rainfall is the limiting factor in evapotranspiration and other

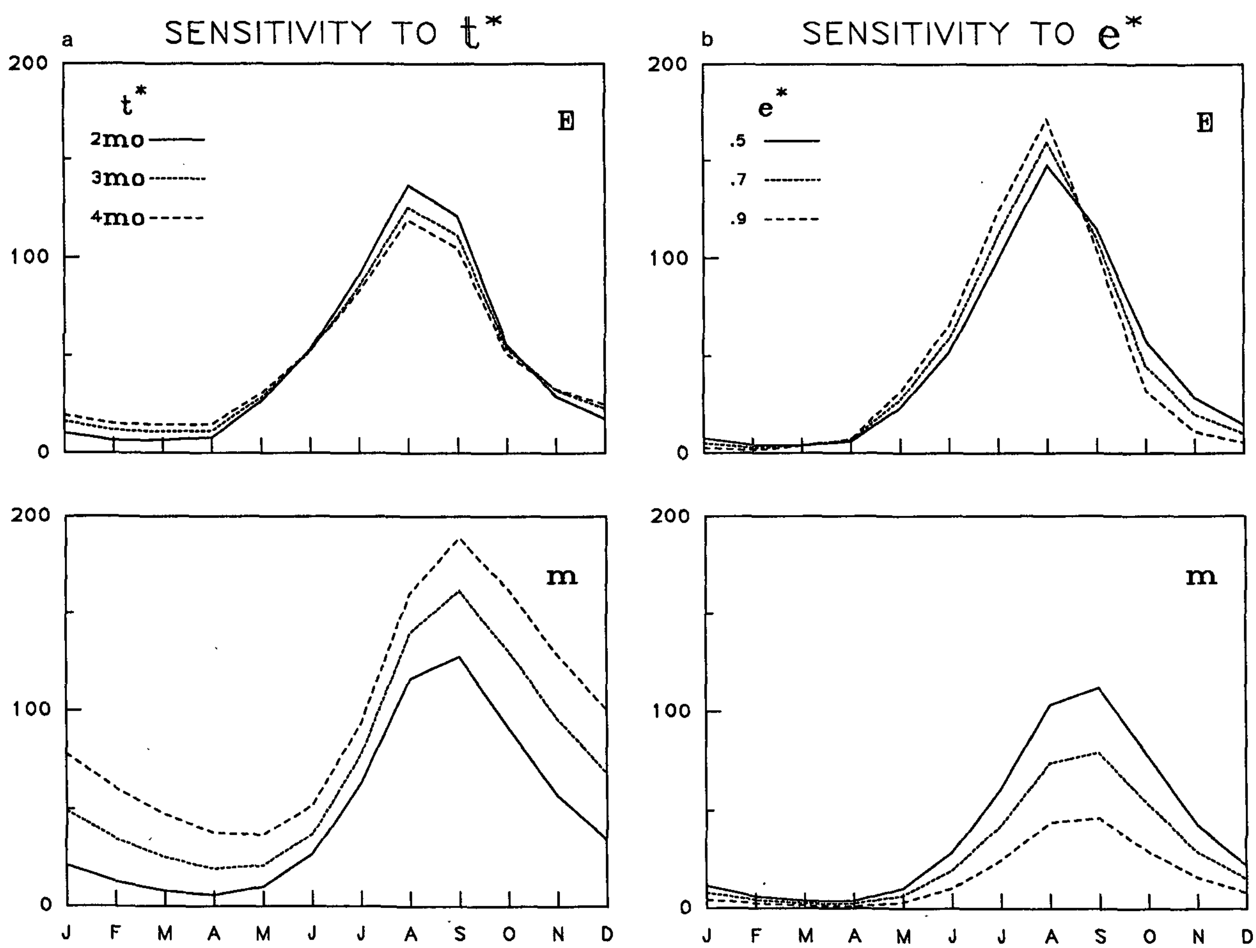


FIG. 6. Calculated values of evapotranspiration E (mm month^{-1}) and soil moisture m (mm) for (a) residence times t^* of 2, 3 and 4 months and (b) evaporivity ratios e^* of 0.5, 0.7 and 0.9.

hydrologic processes in the Sahel. A 50% change in rainfall evokes approximately the same degree of change in evapotranspiration and soil moisture.

The model is highly sensitive to e^* and t^*, the evaporivity and residence time. These values have little influence on total evaporation, but they markedly affect the amplitude of the seasonal cycle. Their greatest impact is on soil moisture. Changing e^* from 0.5 to 0.7 decreases peak soil moisture by ~30%. A one month increase in t^* increases peak soil moisture by ~20% and modifies soil moisture in the drier months by a considerably larger proportion.

The values of soil moisture storage calculated by the model are in close agreement with observed values at a comparable location in Mali (de Ridder et al. 1982). The model also produces reasonable values of evapotranspiration (Sivakumar et al. 1984), with actual equaling potential only in the wettest month (FAO 1984). This gives us confidence in the model's ability to reasonably replicate the most basic hydrologic exchange processes at the regional scale.

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APPENDIX

List of Symbols

Symbol	*Description*
a_s	surface albedo
B	Bowen ratio
C	runoff ratio
c	cloudiness (percent)
D	dryness ratio
dm/dt	change in soil moisture storage (mm month^{-1})
E	total evapotranspiration (mm month^{-1})
E_p	potential evapotranspiration (mm month^{-1})
E'	immediate evapotranspiration (mm month^{-1})
E''	delayed evapotranspiration (mm month^{-1})
e^*	evaporivity
G^*	global radiation
I'	irradiance at the top of the atmosphere (MJ m^{-2} day^{-1})
L	latent heat of vaporization (2.43×10^6 J kg^{-1})
m	exchangeable soil moisture (mm)
N	total runoff (mm month^{-1})
N'	immediate runoff (mm month^{-1})
N''	delayed runoff (mm month^{-1})
P	precipitation (mm month^{-1})
P'	reduced precipitation (mm month^{-1})
Q	sensible heat flux into the atmosphere (MJ m^{-2} day^{-1})
R_{net}	surface net radiation (MJ m^{-2} day^{-1})
t^*	residence time (month)
w	precipitable water (cm)
β	turbidity coefficient

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