A Revised Version of Lettau's Evapoclimatonomy Model

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ABSTRACT

In this paper a revised version of Lettau's evapoclimatonomy model is introduced. Climatonomy is a one-dimensional representation of mean climate, which includes a complete characterization of the surface energy and water balances. The model is fundamentally forced by two input variables: incoming solar radiation and precipitation. All processes related to the ultimate distribution and exchange of the thereby available energy and moisture are parameterized.

The goal of this paper is to describe the model's evolution and demonstrate its utility by sketching results of its application to several diverse climate problems in arid and semiarid Africa. Both the original and the revised evapoclimatonomy models are described, their conceptualization to that of more recent land–atmosphere interaction schemes is compared, their components in terms of the ecosystem are interpreted, and the model improvements incorporated in the current version are proposed. Finally, the use of the model in three applications to studies of climate and ecology in the Sahel and Kalahari regions of Africa is illustrated.

1. Introduction

One of the first climate models developed was the "climatonomy" model of Lettau (e.g., Lettau 1969, 1971; Lettau and Lettau 1969, 1975). The name was derived from the analogy between aerology and aeronomy and was used to signal "quantification" of the hitherto mostly qualitative discipline of climatology. Climatonomy is a one-dimensional representation of mean climate, which includes a complete characterization of the surface energy and water balances. The model is fundamentally forced by two input variables: insolation and precipitation. All processes related to the ultimate distribution and exchange of the thereby available energy and moisture are parameterized. Conceptually, however, the model is quite comparable to current state-of-the-art models of land–atmosphere interaction. The model includes three separate subunits for shortwave radiation, water balance, and total energy balance. This paper deals with the second subunit, "evapoclimatonomy."

Climatonomy has been utilized to model surface climate in diverse regions, including the humid Amazon (Lettau et al. 1979), the tropical Phillipines (Lettau and Baradas 1973), the midlatitude Great Plains of the United States (Corio and Pinker 1987; Pinker and Corio 1987), and arid and semiarid regions of Africa (Nicholson and Lare 1990a; Lare and Nicholson 1990, 1994). It has been applied to a number of climatic problems, such as phenology (Lettau 1973), lake levels (Kutzbach 1980), urban street canyons (Dabberdt and Davis 1978), and deforestation (Lettau et al. 1979). The model has been verified by comparison to observed fields in climatic situations in North America (Lettau 1969; Corio and Pinker 1987) and in the Philippines study, where the correlation between calculated and measured runoff was as high as 0.90. Model results

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have been promising, despite the empiricism in the original model.

The climatonomy model has several advantages. It is relatively simple, requires readily accessible input data, allows for a complete formulation of energy and water balance from a few basic input variables, and provides for feedback loops necessary for the coupling of these balances. An additional advantage is that it is valid at the regional scale. For these reasons, we have selected the evapoclimatonomy submodel for studies of land–atmosphere interactions over Africa, but we have reformulated much of the detail in order to replace empirical relationships with physical parameterization.

The goal of this paper is to describe the model's evolution, as Lettau has published several versions of his model and we have also revised it, and to demonstrate the model's utility by sketching results of its application to several diverse problems. We begin with a description the original evapoclimatonomy model, compare its conceptualization to that of more recent land–atmosphere interaction schemes, and interpret its components in terms of the ecosystem. We then describe the model improvements incorporated in the current version. Finally, we illustrate the use of the model in three applications to studies of climate and ecology in Africa. In two companion articles (Marengo et al. 1996; Nicholson et al. 1996), we verify the model in an application to modeling surface water balance at the sites of the HAPEX-Sahel experiment in West Africa [see Goutorbe et al. (1994) for a description of this experiment].

2. Evapoclimatonomy

a. Model equations and concepts

The evapoclimatonomy submodel is essentially a numerical solution to a simplified form of the hydrologic balance equation:

$$P = E + N + \frac{dm}{dt}, \tag{1}$$

where P is precipitation, E is evapotransipration, N is runoff, and dm/dt is the change in soil moisture storage. Surface water balance is forced primarily by ground-absorbed solar radiation and precipitation, the basic model input. The use of a calculus-based solution rather than algebraic accounting distinguishes this model from prior water balance approaches, such as Penman or Thornthwaite (Thornthwaite and Mather 1955). This also facilitates the incorporation of feedback loops and allows for distinction between immediate and delayed processes of runoff and evapotranspiration (N', N'', E', E'', respectively).

Three assumptions are made in order to maintain simplicity (Lare and Nicholson 1990). The first is that for a continental region with a stable climate there is no net storage of moisture over sufficiently long time periods (e.g., those for which the means of the forcing functions are derived). That is, although soil moisture varies seasonally, the net storage of moisture over sufficiently long time periods is zero:

$$\frac{d\bar{m}}{dt} = 0, \tag{2}$$

where m is the annual average soil moisture. The larger the area and the longer the timescale considered, the better this assumption is. Therefore,

$$\bar{P} = \bar{E} + \bar{N}. \tag{3}$$

The second assumption is that the processes of runoff and evapotranspiration can be further subdivided into immediate ($'$) and delayed ($''$) parts. Thus,

$$E = E' + E'' \tag{4}$$

$$N = N' + N''. \tag{5}$$

Immediate processes are those that occur in the same month as the precipitation, while delayed implies processes that are associated with the rain that fell in previous months. Using this assumption, a quantity termed "reduced precipitation" P' is defined as

$$P' = P - E' - N'. \tag{6}$$

Reduced precipitation is simply the rainfall that does not directly evaporate or run off and is thus available for soil moisture storage. The hydrologic balance equation may then be written as

$$\frac{dm}{dt} = P' - (N'' + E''). \tag{7}$$

The final assumption of the evapoclimatonomy submodel is that the delayed processes vary directly in proportion to soil moisture according to

$$N''(t) = \frac{\bar{N}''m(t)}{\bar{m}}, \tag{8}$$

$$E''(t) = \frac{\bar{E}''m(t)}{\bar{m}}, \tag{9}$$

where $\bar{N}''$ and $\bar{E}''$ are the mean quantities of delayed runoff and evapotranspiration. These quantities are combined in such a way as to yield a process parameter known as "residence time" t^*:

$$N'' + E'' = \frac{m}{t^*}. \tag{10}$$

$$t^* = \frac{\bar{m}}{\bar{N}'' + \bar{E}''}. \tag{11}$$

Residence time is 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. Using (7), (10), and

(11), the hydrologic balance formula is then rewritten as

$$P - E' - N' = \frac{m}{t^*} + \frac{dm}{dt}. \quad (12)$$

In order to calculate the immediate evapotranspiration, another empirical process parameter, "evaporivity" e^*, is defined. Evaporivity is a nondimensional measure of the land surface's ability to use part of the incoming solar radiation to evaporate the rainfall received in a given month (Lettau 1969). It essentially represents the fraction of precipitation immediately evaporated but prorated by the available insolation in the month (a factor analogous to potential evapotranspiration).

The next step in the model is to subtract the annual means from each term in the new form of the hydrologic balance equation, resulting in an ordinary differential equation

$$p'(t) = \frac{m - \bar{m}}{t^*} + \frac{d(m - \bar{m})}{dt}, \quad (13)$$

where

$$p'(t) = P - E' - N' - (\overline{P - E' - N'}). \quad (14)$$

This is solved as

$$m - \bar{m} = e^{-t/t^*}\left(\text{const} + \int e^{t/t^*} p' dt\right). \quad (15)$$

Because of the initial assumption of climatic stability, the bracketed term must approach 0, thus determining the integration constant. The equation is solved via stepwise integration until the bracketed term approaches 0.

This entire procedure is completed twice. Initially, the model is run as a diagnostic process for each station using mean monthly values for rainfall and solar radiation to calculate mean annual values for soil moisture storage, total runoff, and evapotranspiration. It is then run a second time as a prognostic tool using the mean values of soil moisture storage and delayed runoff and evapotranspiration calculated during the first run. Observed values of precipitation are used as the forcing function. The final output is the time series of soil moisture storage, runoff, and evapotranspiration.

b. *Calculation of model parameters*

Precipitation is a basic input to the evapoclimatonomy model. All other components of the water balance are calculated by the model. In the original model, parameterization required that either annual runoff or annual evapotranspiration be known and that evaporivity and residence time be reasonably estimated. The model is highly sensitive to these parameters.

In Lettau and Lettau (1969), the requisite annual means were calculated utilizing Budyko's (1986) dryness ratio concept (*the ratio of net radiation to the product of annual precipitation and the latent heat of condensation*). Budyko had empirically determined relationships among the dryness ratio, the runoff ratio (*the ratio of total runoff to precipitation*), and the Bowen ratio (*ratio of sensible to latent heat flux*), so that the runoff ratio could be prescribed from the dryness ratio.

Lettau (1969) suggested that the minimum value of residence time t^* would be on the order of 2–3 months, the probable values in semiarid regions, while a maximum would be 1 year if a stable climate is assumed. An assumption of 3 months for the central plains of North America yielded realistic results, but T^* would be greater for more humid climates. The parameter e^*, evaporivity, was likewise roughly estimated, based on empirical determinations showing that it would generally range from 0.4 and 0.8 (Lettau 1971).

Lettau and Baradas (1973) introduced the concept of "calibrating" the watershed by using four additional process parameters to facilitate the partitioning of delayed and immediate processes. They also introduced a dependency of process parameters on soil moisture. These parameters included two thresholds: a precipitation threshold that must be exceeded for immediate runoff to occur, and a soil moisture threshold that must be exceeded for delayed evaporation to occur. They also utilized ratios to represent the efficiency of soil moisture evaporation and the partitioning of delayed processes into evaporation and runoff. The model was calibrated for the Mabacan River watershed in the Philippines, using available runoff and evapotranspiration data. Unlike prior studies, which dealt only with long-term means, the model was utilized to assess conditions during both dry and wet years.

Although parameters added in the Lettau–Baradas version were empirical, some were quite analogous to physical parameters, such as infiltration capacity and field capacity. This facilitated Lare's (1992) revision of the model.

In a more recent version of evapoclimatonomy, Lettau and Hopkins (1991) replaced the semiempirical watershed calibrations of previous versions with a less subjective diagnostic scheme. The model was verified and used to study interannual variability related to drought and to deforestation.

c. *Soil–plant–atmosphere interactions in the ecosystem*

The precipitation not directly lost by runoff or evaporation of surficially detained water infiltrates through to the soil, where it is stored for longer periods. Soil moisture is depleted in four ways: surface runoff, surface evaporation, extraction via plant roots through transpiration, and gravity drainage from the soil layer to the groundwater (Fig. 1). Surface runoff, also called overland flow, is the lateral movement of water in the porous upper few centimeters of the ground or on the surface itself. The total evapotranspiration is the summation of evaporation of water retained on the soil surface or the surface of plants, evaporation of soil moisture, and transpiration through plants.

The distinction between soil moisture and groundwater is an important one because the utilization of these sources of water is quite different. The soil moisture layer essentially constitutes the root zone of plants; the water contained therein is available to plants. Groundwater is long-term storage and is not directly available to plants.

Although the climatonomy model's partitioning of immediate and delayed processes is a mathematical construct to account for time lags between processes and to facilitate feedbacks in the model, this distinction is consistent with the physical processes involved in soil–plant–atmosphere interactions. It essentially separates temporal variations of runoff and evapotranspiration associated more or less directly with rainfall from those involving subsurface moisture (Nicholson and Lare 1990a).

This being the case, the four physical mechanisms by which soil moisture is depleted are roughly equivalent to the four main evaporation and runoff quantities in the climatonomy model. Immediate evaporation E' is essentially surface evaporation from soil and plants (termed detention and interception water, respectively), plus most of the evaporation that takes place from the upper soil layers. Delayed evaporation E'' would include primarily transpiration of water extracted from the soil moisture layer. Immediate runoff N' is essentially lateral, or surface, runoff; delayed runoff N'' is analogous to gravity drainage.

The partitioning of immediate and delayed processes is also consistent with the current hydrological approach of separately considering storm and interstorm periods. The immediate processes take place mostly during the storm, the delayed during the interstorm periods.

These analogies provide the basis for the physical parameterizations in the revised evapoclimatonomy model, which are discussed in section 2d.

d. The revised evapoclimatonomy model

The revised evapoclimatonomy model incorporates mainly changes in the determination of immediate runoff and evapotranspiration N' and E' and of parameters t^* and e^*. Minor changes are introduced in the calculation of delayed parameters E'' and N'' and soil moisture m. These changes require the introduction of concepts such as potential evapotranspiration, infiltration capacity, gravitational drainage, and residual water content and allow for parameters to vary with soil and vegetation type. The values of e^* and N' also vary according to season.

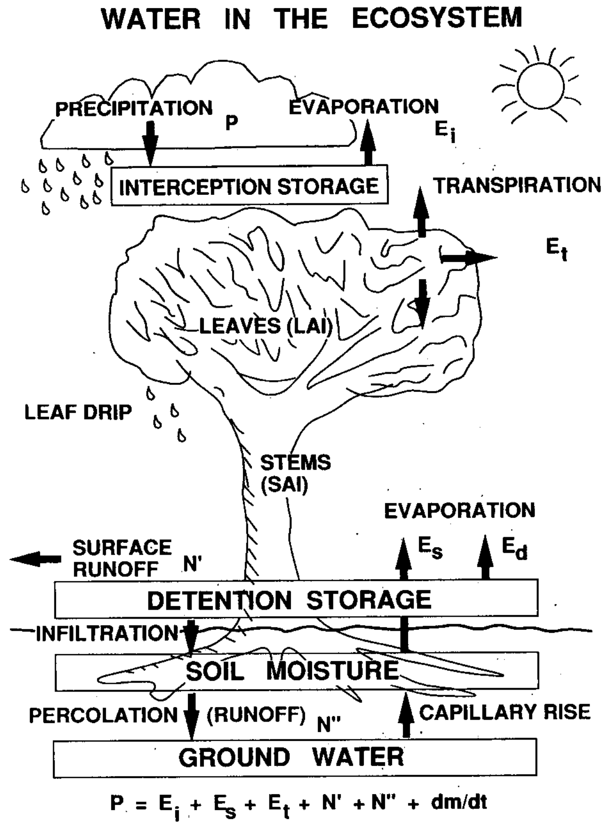


FIG. 1. Relationship between evapoclimatonomy and biophysical processes involved in surface hydrology.

This revised version of the model has been applied in studies of surface water balance in the domain of the HAPEX-Sahel regional experiment, as detailed in Marengo et al. (1996) and Nicholson et al. (1996). Here the main changes are summarized. One is referred to Lare (1992) and Lare and Nicholson (1994) for more complete details.

The original model assumed an arbitrary value of residence time (Lettau 1969). Residence time t^* is now calculated after Serafini and Sud (1987) as a function of potential evapotranspiration, wilting point, and field capacity, with an adjustment of values for various vegetation types. Wilting point and field capacity are determined according to Saxton et al. (1986) as a function of soil type (i.e., percent sand and clay content). Thus, t^* is now parameterized as

$$t^* = \frac{\gamma}{\lambda \text{PET}} \ln \left(\frac{e^{\text{Wfc}} - 1}{e^{\text{Wp}} - 1} \right) \left(\frac{\text{Wfc} - \text{rwc}}{\text{Wfc} - \text{Wp}} \right), \quad (16)$$

where

$$\lambda = \frac{\alpha_v}{\text{Wfc} - \text{Wp}}$$

and

$$\gamma = 1 - e^{\alpha v}.$$

Parameters Wp and Wfc are the wilting point and field capacity, respectivelly, rwc is the residual water content, and PET is potential evapotranspiration in millimeters per day or millimeters per month, and accordingly, t^* is in days or months. The factor α_v accounts for vegetation type (Serafini and Sud 1987). Mintz and Serafini (1984) suggested using $\alpha_v = 6.81$ to cover most vegetation types.

In the earlier versions of the model (Lare and Nicholson 1990, 1994; Nicholson and Lare 1990a; Farrar et al. 1994), the soil moisture estimate only refers to exchangeable soil moisture m, that is, moisture available for the processes of runoff and evapotranspiration. Although it is this quantity that is calculated by the model equation (15), the residual water content (rwc) must be added to the calculated exchangeable soil moisture to obtain the true soil moisture contents. Residual water content is determined after Rawls and Brakensiek (1989):

$$\mathrm{rwc} = 10[0.2 + 0.1\,(\%\ \mathrm{om}) + 0.25\,(\%\ \mathrm{clay})(\mathrm{cec})^{0.45}]\ \delta, \quad (17)$$

where rwc is the residual water content in millimeters, (% om) is percent organic matter content, (% clay) is percent of clay content, (cec) is cation exchange capacity, and δ is the bulk density of the entire soil layer (0–100 cm), as estimated by Rawls and Brakensiek (1989).

The parameter e^*, evaporivity, is defined as a nonlinear measure of the capacity of the land surface to use a portion of monthly solar radiation to evaporate precipitation received in the same month. It serves the same function as potential evaporation in traditional water balance models. Previous versions of the model (Lettau 1971; Pinker and Corio 1987; Nicholson and Lare 1990a; Farrar et al. 1994) have prescribed arbitrary values of e^* based on vegetation, soil, and topography. In the revised version (Lare 1992; Marengo et al. 1996; Nicholson et al. 1995), e^* is analytically derived by using the normalized difference vegetation index, or NDVI.

This index (Tucker et al. 1981, 1983, 1985; Tucker and Sellers 1986) is based on reflectance differences in the visible and near-infrared portions of the solar spectrum. In semiarid regions it correlates strongly with such parameters as greenleaf biomass and leaf area index (LAI) but is essentially a measure of photosynthetic activity and hence a measure primary production (Prince 1991). Lare (1992) reasoned that e^* may be thought of as the ratio between soil evaporation and transpiration from plant surfaces and noted that because of the intimate link between photosynthesis and transpiration (Tucker and Sellers 1986), NDVI is an indicator of transpiration. On the basis of relationships given in Lowry and Lowry (1989), Lare derived the semiempirical formula

$$e^* = (1.25 + 6.25\ \mathrm{NDVI})^{-1} \quad (18)$$

for the calculation of e^*.

In our initial work (Nicholson and Lare 1990a), immediate runoff N' was assumed to be 0, based on a literature review suggesting that it is minimal in arid and semiarid regions. In the revised model, it is calculated using the approach developed by Milly and Eagleson (1982) and adapted by Warrilow (1986). In essence, immediate runoff is a function of mean monthly rainfall rate, mean monthly effective hydraulic conductivity, and a proportionality constant based on the average fractional area of a grid or region in which precipitation occurs. Infiltration rate is a function of hydraulic conductivity, which is determined by soil moisture content and soil texture, following Saxton et al. (1986). The rate is modified by vegetation density and structure and ground cover, as in Rawls and Savabi (1989), Rawls and Brakensiek (1989), and Wilcox et al. (1990).

Thus, N' is be expressed as

$$N' = P \exp\left(-\frac{\alpha F}{P}\right), \quad (19)$$

where P is precipitation rate, F is the infiltration rate, and α is the proportionality constant (Warrilow 1986). The infiltration rate F is determined following Saxton et al. (1986), Rawls and Savabi (1989), and Rawls and Brakensiek (1989) and depends on percent sand and clay contents for the surface layer (0–10 cm) for a particular soil type, the presence of vegetation canopy, ground cover, surface rocks, leaf litter, and crusting of the soil. This is somewhat different from the previous version of the model used by Nicholson and Lare (1990a, 1994), in which the hydraulic conductivity is assumed as the infiltration rate for bare soil only. The proportionality constant α assigns a particular fraction of a grid box, such that within the prescribed area it is assumed that precipitation is occurring. Since rainfall over Africa is primarily convective in nature, α is set to 0.66 (Eagleson et al. 1987).

The delayed processes E'' and N'' are functions of soil moisture. In the revised version of the model, E'' and N'' depend on the exchangeable soil moisture content m, rather than the total soil moisture. Immediate evapotranspiration E' then can be estimated as

$$E' = e^*(P)\left(\frac{\mathrm{Rs_{abs}}}{\overline{\mathrm{Rs_{abs}}}}\right), \quad (20)$$

where $\mathrm{Rs_{abs}}$ and $\overline{\mathrm{Rs_{abs}}}$ are the monthly and annual ground-absorbed global solar radiation, respectively.

As indicated in Lare (1992), delayed runoff represents gravitational drainage. Therefore,

$$N'' = K(\theta), \quad (21)$$

where K is the hydraulic conductivity (usually unsaturated), which is a function of the soil moisture content

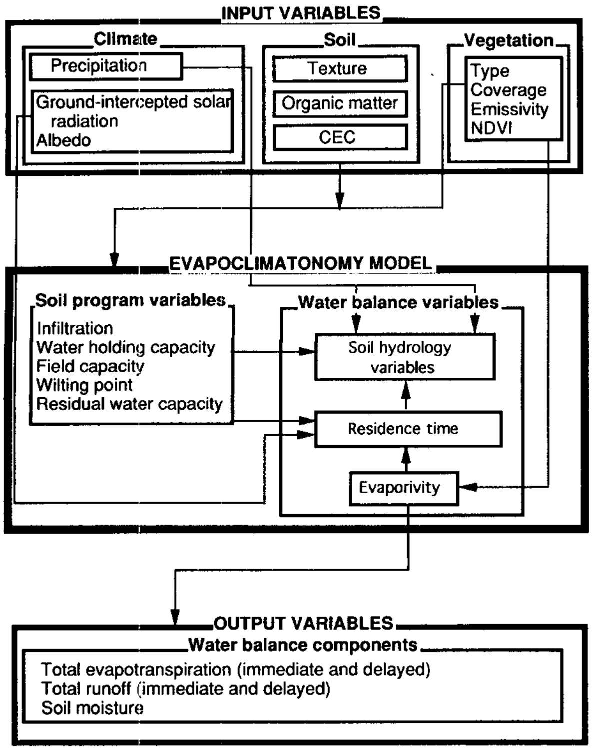


FIG. 2. Schematic overview of the revised evapoclimatonomy model.

θ in the soil layer from 1 to 100 cm (Warrilow 1986). In the model, by using calculated soil moisture contents from the evapoclimatonomy submodel, mean annual delayed runoff is iterated until convergence is obtained between mean annual delayed runoff determined by the evapoclimatonomy submodel (8) and the mean annual delayed runoff calculated in (21). Delayed evaporation E'' is calculated as a residual.

3. Recent model applications

This section sketches the results of several diverse applications of the revised model and includes a comparison between calculations by the original and revised model. Details of the applications are provided in Lare and Nicholson (1990, 1994) for the original version and in Marengo et al. (1996) and Nicholson et al. (1996) for the revised version.

Figure 2 gives a schematic overview and summary of the revised version of evapoclimatonomy. The model requires monthly values of rainfall, surface albedo, solar radiation, NDVI, and potential evapotranspiration, as well as cation exchange capacity, sand, clay, and organic matter content for two soil layers. Solar radiation and potential evapotranspiration values were taken from FAO (1984). Surface albedo data were taken from global monthly fields derived by Dorman and Sellers (1989) using the Simple Biosphere Model (SiB) of Sellers et al. (1986). Other vegetation parameters such as the upper- and lower-story vegetation cover and the emissivity were taken from Dorman and Sellers (1989) and Stull (1988). NDVI data, calculated from AVHRR data from the NOAA satellites (Justice et al. 1986), were obtained from C. J. Tucker of NASA/Goddard Space Flight Center. Information on soil texture, organic matter, and cation exchange capacity were obtained from Zobler (1986) and Webb et al. (1991).

a. NDVI, soil moisture, and ET in West Africa: Original versus revised evapoclimatonomy models

Nicholson and Lare (1990a) applied Lettau's original model to assessing the surface water balance in the central Sahel, utilizing input data for Niamey, Niger. The model had assumed $N' = 0$, which according to Eagleson and Segarra (1985) was generally true for semiarid regions. Figure 3 shows model calculations of evapotranspiration, runoff, and soil moisture for Niamey (Nicholson and Lare 1990a, 1991). Evapotranspiration and rainfall, the forcing functions of the model, peak in August; soil moisture lags by 1 month. Monthly runoff never exceeds 2 mm and is near 0 most of the year. The calculated soil moisture ranges from near 0 at the end of the dry season to about 25–50 mm at the beginning and end of the rainy season and about 80—90 mm in the peak months of August and September. This compares favorably with observations from a similar location in Mali (de Ridder et al. 1982).

The model also gives reasonable values of evapotranspiration, with actual equalling the potential only in the wettest month (Sivakumar et al. 1984; FAO 1984). The validity of the results is further supported by the favorable comparison between calculated soil moisture and NDVI (Fig. 3). The agreement with observed soil moisture is good and the phenology of NDVI generally matches the season cycle of model-calculated soil moisture. The exception is the apparent minimum in NDVI in May and June, a minimum that reflects reduction of NDVI by aerosols and atmospheric moisture, rather than a decline in growth. Both NDVI

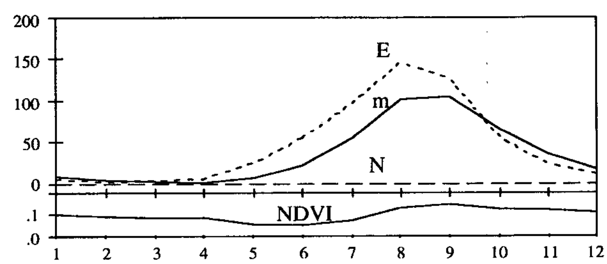


FIG. 3. Water balance at Niamey, calculated with the previous version of evapoclimatonomy. Units are evapotranspiration in millimeters per month; soil moisture in millimeters; and runoff in millimeters per month (from Nicholson and Lare 1990).

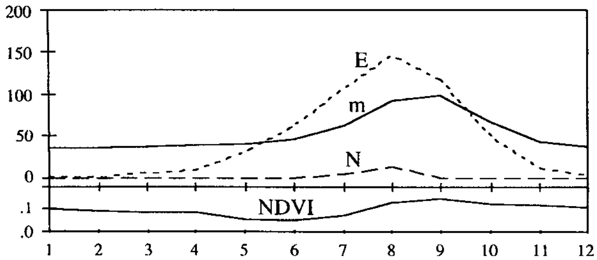


FIG. 4. As in Fig. 3 but with calculations from the revised evapoclimatonomy model. Note that the soil moisture estimates includes model-calculated exchangeable soil moisture plus the residual soil water content.

and soil moisture appear to lag rainfall by one month, a realistic result.

Calculations with the revised model for Niamey are shown in Fig. 4. The major differences are in the soil moisture estimates. The original version calculates only exchangeable soil moisture, while the revised version estimates the total soil moisture content by adding the residual water content to calculated exchangeable soil moisture. Peak values of soil moisture so-calculated are slightly lower than the original estimates. Since the residual water content is approximately 30 mm, this means significantly lower exchangeable soil moisture. This is a result of a greater proportion of evaporation resulting from delayed processes (Lare 1992). Other notable differences between calculations with the original and revised models include increased runoff during the rainy season and lower evapotranspiration in September. Evapotranspiration is otherwise similar to that from the original model in magnitude and seasonal cycle, peaking in August one month prior to peak soil moisture.

The calculated soil moisture varies from around 40 mm (approximately the wilting point value) at the beginning and end of the rainy season, reaching maximum values of about 70–90 mm in August and September. It is important to indicate that the soil moisture estimates obtained with the revised version of the evapoclimatonomy model compare most favorably with neutron probe measurements at the nearby region that constitutes the domain of the recent HAPEX-Sahel experiment, particularly for the August–October season. As in the case of the previous version of the model, the agreement with observed soil moisture is good and the phenology of NDVI matches the season cycle of model-calculated soil moisture, except for the May–June NDVI minimum induced by atmospheric effects.

The revised model was applied to stations along a north–south transect following the rainfall gradient in the Sahel from the semidesert grassland to the forest (Fig. 5a). Except for Niamey, the most arid station, correlations between NDVI and both soil moisture and ET are excellent, but the NDVI–ET relationship gets progressively better, while the NDVI–soil moisture re-

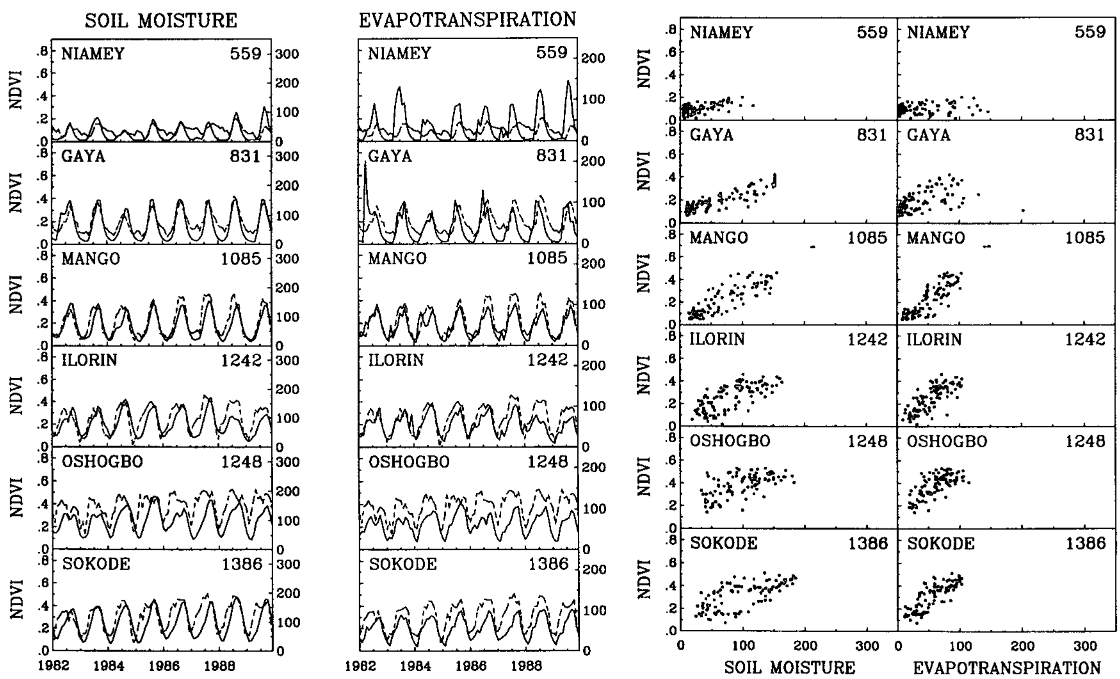


FIG. 5. NDVI (dashed lines) versus soil moisture (mm; solid lines) and NDVI versus evapotranspiration (mm month^{-1}; solid lines) for six stations along a north–south transect in West Africa. Mean annual rainfall in millimeters is given in the upper right of each diagram. (a) Monthly time series 1982–89. (b) Scatter diagrams of data in (a).

lationship gets progressively poorer with increasing rainfall. At relatively dry stations there is a linear relationship between NDVI and soil moisture with little scatter; the scatter increases and the correlation decreases at wetter stations, as NDVI levels off with increasing soil moisture (Fig. 5b). Except for the driest station, where the NDVI–ET correlation is extremely weak, ET generally increases with NDVI. There is much scatter at the drier stations but at the wetter stations, a clear linear relationship with comparatively little scatter is apparent. Although the absolute values of the correlations do not systematically change along the wet–dry gradient, the ratio of the NDVI-soil moisture correlation to the NDVI-ET correlation generally decreases southward with increasing rainfall. This is evident not only for the six stations in the transect but for a group of about 32 stations in West Africa for which the model was run.

These results are physically realistic because in drier regions vegetation growth is limited by soil moisture, but in wetter regions (with nearly 100% vegetation cover) transpiration is the dominant contributor to ET and photosynthesis (which NDVI represents) is linearly related to ET. The point at which the NDVI–soil moisture and NDVI–ET correlations are roughly equal, represented by the station Mango (mean annual rainfall of 1085 mm), marks the transition from water-limited to radiation-limited environments.

b. *Continental fields of water balance parameters*

Lare and Nicholson (1994) adapted the evapoclimatonomy model for the derivation of spatial fields of water balance parameters over West and Southern Africa. For a small number of base stations, solar radiation and ancillary data (e.g., atmospheric constituents) were obtained and their values assigned to other nearby stations in the networks. Local information on rainfall, vegetation, and soil type was used for each station in the network.

The model was utilized to calculate soil moisture, evapotranspiration, and runoff during the ten wettest and ten driest years in each region. Wet-minus-dry calculations for the wettest months (August and January, respectively) are shown in Fig. 6. A strong contrast is evident between the two regions in the consistency, size, and magnitude of the wet-minus-dry anomalies. Changes are much larger and spatially coherent over the Sahel.

One of the most interesting results is the change of moisture and ET gradients in the Sahel in wet and dry years, a consequence of increased precipitation resulting in higher runoff in wetter regions to the south but higher soil moisture and ET in drier regions to the north. Using the thermal wind relationship and a simple convective model, it was demonstrated that the change in gradients could decrease the speed of the African easterly jet by 3–8 m s^{-1}, depending on latitude (Lare and Nicholson 1994). This is consistent with observed contrasts in the jet in wet and dry years (Newell and Kidson 1984). This strongly suggests the possibility of hydrologic feedback in the drought since the jet provides the instability and energy for the development and maintenance of rain-bearing disturbances in the region.

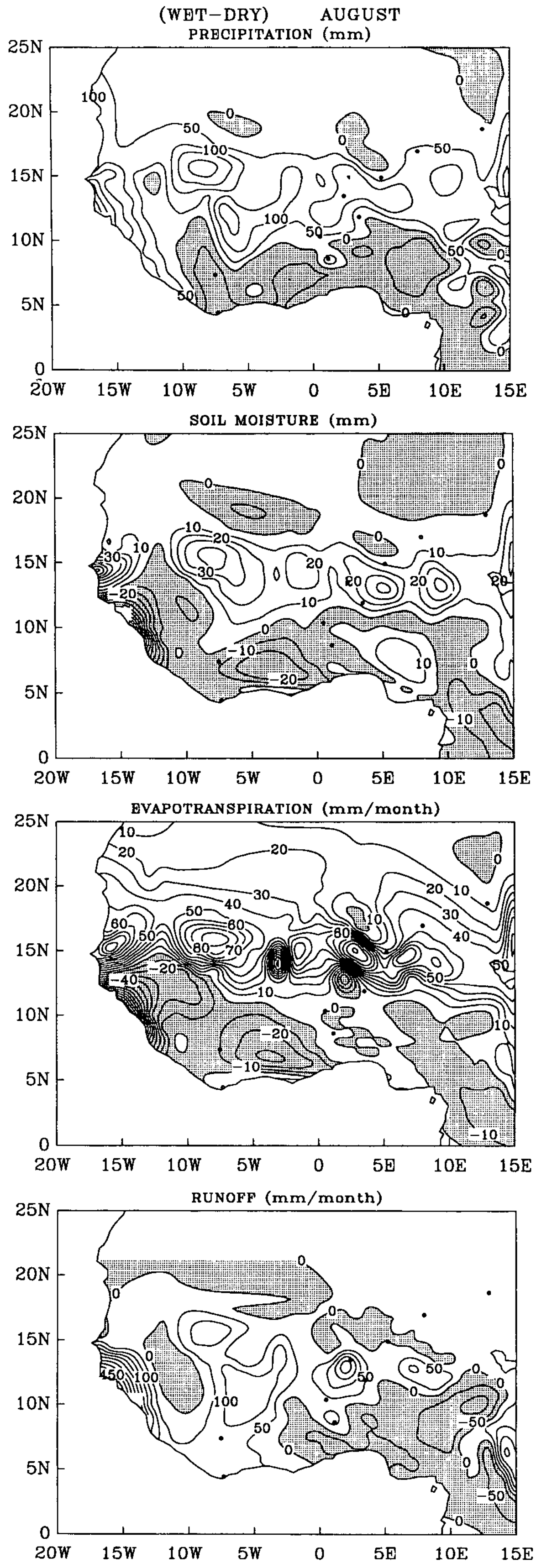
c. *Studies of the relationship between soil moisture and vegetation growth*

Evapoclimatonomy was recently applied to assessing soil moisture in Botswana in order to better understand relationships that had been identified between vegetation growth (as assessed via NDVI) and rainfall elsewhere in Africa (Nicholson et al. 1990). Two interesting aspects of the relationships warranted further examination: the differences in vegetation growth in East and West Africa under similar conditions of rainfall and the log-linear form of the NDVI–rainfall relationship, with NDVI leveling off to a constant value as rainfall increases (Fig. 7). Specific goals of the Botswana study were to determine the reasons for different growth rates under similar conditions of climate, to determine the influence of soil type on the NDVI–rainfall relationships, and to evaluate the saturation of the index at higher levels of rainfall. Details of the study are given in Farrar et al. (1994).

The study exploited NDVI–rainfall ratios and NDVI–soil moisture ratios as proxy indicators of the biophysical functions of rain-use efficiency (LeHouerou and Hoste 1977) and water-use efficiency (Tucker and Sellers 1986). These functions are described by the ratios of productivity (of which NDVI is an approximate measure) to rainfall and productivity to transpiration (roughly a function of soil moisture in semiarid regions). Various authors (e.g., LeHouerou and Hoste 1977; Noy-Meir 1985) have suggested using these to characterize ecosystems and to monitor land surface changes.

The ratios of NDVI to rainfall are markedly different in East, West, and Southern Africa (Table 1); the values suggest that rain-use efficiency is highest in Southern Africa and lowest in West Africa. To determine whether or not this was due to climatic contrasts reflected in the rate of generation of soil moisture, climatonomy was applied to three stations western and Southern Africa. The ratio of soil moisture to rainfall was considerably greater in West Africa than Southern Africa, thus demonstrating that this factor could not explain the more efficient growth in Southern Africa (Table 2).

The evapoclimatonomy model was applied to assessing monthly soil moisture at a total of 16 Botswanan stations representing five texturally diverse soil types, and the results were compared to monthly composited NDVI (Fig. 7, Table 3). Differences in the NDVI–rainfall and NDVI–soil moisture relationships



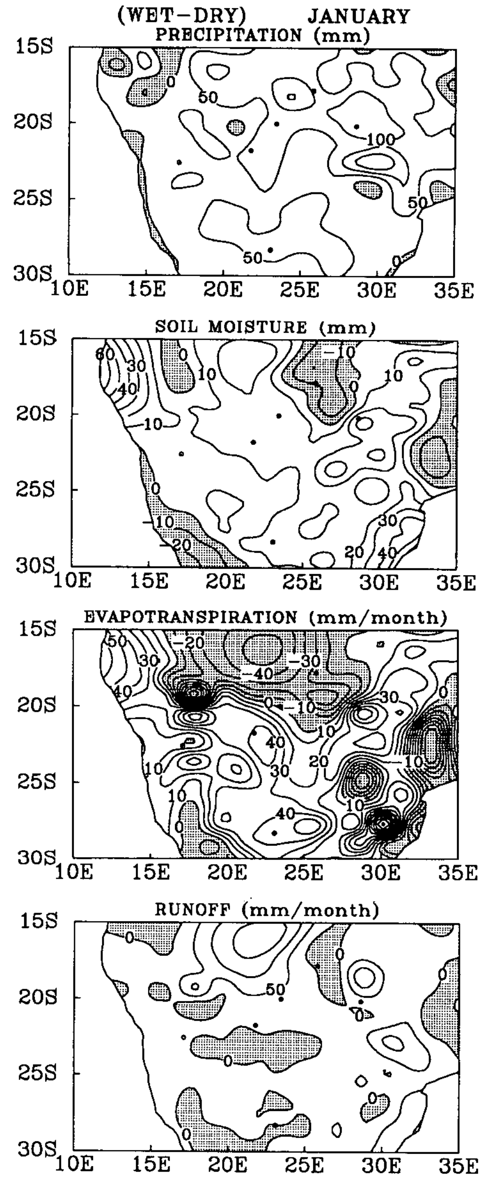


FIG. 6. Wet-minus-dry calculations of soil moisture, evapotranspiration and runoff over West and Southern Africa (from Lare and Nicholson 1994). These are from the ten wettest and ten driest years in each region for August and January.

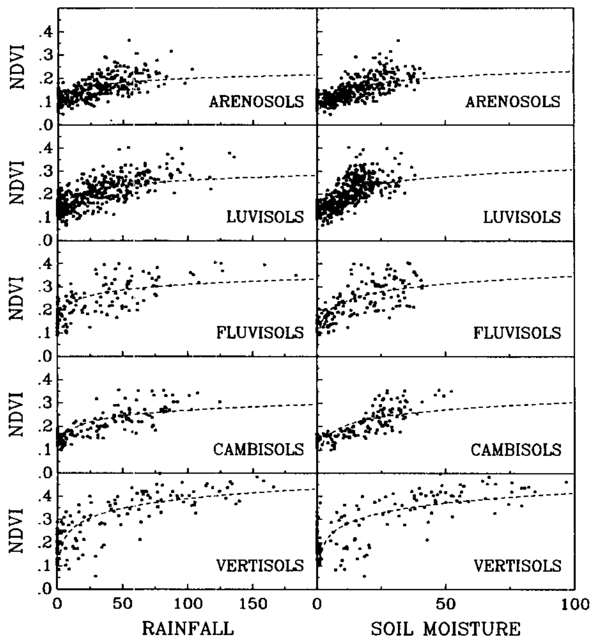


FIG. 7. NDVI–rainfall and NDVI–soil moisture relations for sixteen stations in Botswana representing five soil types (from Farrar et al. 1994). Rainfall is in millimeters per month; soil moisture is in millimeters.

are clearly apparent. The NDVI–rainfall slope generally decreases with sand content, but with soil moisture the relationship is more complex. The lowest ratios correspond to the two extremes, the arenosols, which are over 90% sand, and the vertisols, which are nearly 80% clay. The reason is that initial growth in the clay-rich soil is so rapid that the index readily "saturates" (NDVI levels off), the canopy density becomes so dense that maximum photosynthetic activity is attained and no further increase occurs even with more available moisture. The environment becomes effectively "radiation-limited."

These results can also help to interpret the saturation of NDVI noted earlier. At least two explanations are plausible. One is that although rainfall continues to increase, the soil becomes saturated, and, despite increasing rainfall, moisture availability to plants ceases to increase. The other is that the index saturates because the environment becomes radiation-limited rather than water-limited as the photosynthetic capacity is reached. Figure 7 shows that soil moisture continues to increase after NDVI has leveled off, demonstrating that water-availability is no longer the limiting factor in growth. Thus, the "saturation" of the index probably marks the transition between water-limited and radiation-limited environments, a result that can be exploited in energy balance and climate models.

The above studies were carried out with the original evapoclimatonomy model. The revised model was also applied to stations in Botswana and neighboring southern African countries in order to more explicity determine the rainfall thresholds at which the index saturates. Some of the results are shown in Fig. 8, which illustrates the correlation between NDVI and rainfall at

TABLE 1. Annual integrated NDVI (4-yr mean), annual rainfall (4-yr mean in millimeters), and rain-greenness ratio (RGR) for vegetation formations in East, West, and Southern Africa (from Nicholson et al. 1990; Farrar et al. 1994).

Vegetation zone	Annual NDVI	Annual rainfall	RGR
East Africa			
Lowland forest	4.3	1320	3.2
Coastal mosaic	4.0	1014	3.9
Coastal forest	5.2	1962	2.7
Upland forest	4.3	1157	3.9
Wet miombo	3.7	1102	3.5
Dry miombo	3.8	948	4.3
Itigi thicket	3.9	700	5.6
Shrub-thicket	2.5	546	5.1
Shrub-thicket mosaic	3.3	803	4.6
Semidesert shrub	1.1	297	3.5
Southern Africa			
Southern woodland	2.6	951	2.8
Northern woodland	1.5	460	3.5
Grassland–woodland transition	0.9	249	3.9
Grassland	0.6	110	5.7
West Africa			
Mopane woodland	2.0	313	6.4
Transition zone	2.7	339	8.1
Kalahari Thornveld	2.4	356	6.8

TABLE 2. Comparison of soil moisture generation rates at comparable stations in the Sahel and the Kalahari, as indicated by the ratio of mean monthly soil moisture m to mean monthly rainfall R.

Kalahari station	Mean annual rainfall	m/R	Sahel station	Mean annual rainfall	m/R
Kasane	673	0.76	Gaya	831	0.90
Gaborone	531	0.35	Niamey	559	0.83
Dibete	381	0.52	Tahoua	381	0.58

TABLE 3. NDVI–soil moisture M slope, NDVI–rainfall regression R slope, and NDVI–rainfall ratios, using the best correlated rainfall interval [slopes multiplied by 1000; note that R slope values differ slightly from those in Marengo et al. (1996) because they are based only on the subset of 17 stations used in the soil moisture analysis]. Approximate sand content (%) given in parentheses.

Soil–vegetation		M slope	R slope
Arenosols	(90–95)	3.5	1.5
Luvisols	(50–60)	5.6	1.7
Fluvisols	(40–55)	4.7	1.6
Cambisols	(30–35)	4.4	1.7
Vertisols	(~10)	3.2	2.0

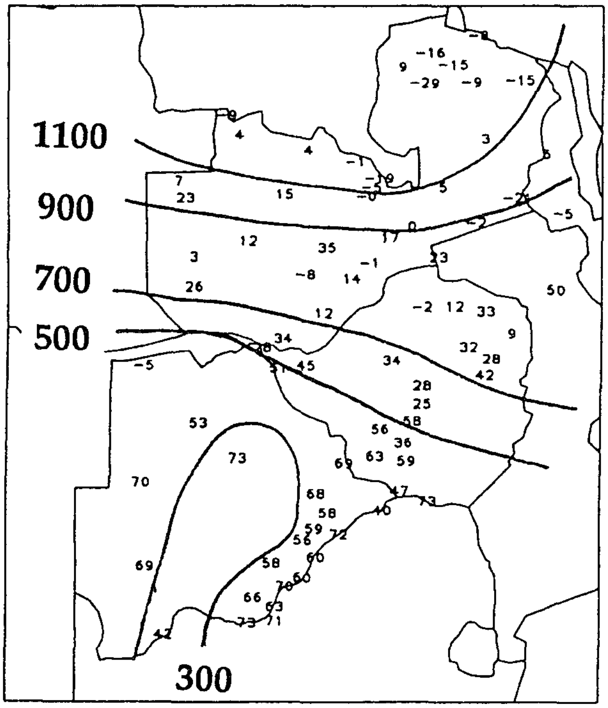


FIG. 8. The correlation (× 100) between monthly composed NDVI anomalies and monthly rainfall anomalies over Southern Africa. Country boundaries and rainfall isohyets (mean annual rainfall) are indicated.

73 stations in the region. For raw monthly data, the correlations (not shown) are excellent everywhere because of the similar seasonal cycles of the two variables. When the annual cycle is removed, the NDVI–rainfall correlation drops off sharply in the area where mean annual rainfall increases from 500 to 700 mm per year. This suggests that the index saturates at these levels of rainfall. A regression analysis of the scatter of the NDVI–rainfall relationship (Fig. 9), using a technique described by Draper and Smith (1981), fixes the saturation point at approximately 700 mm per year. Similar analyses for East and West Africa give about 750 and 1300 mm, respectively.

We have stratified the Southern African data into two groups representing stations with mean annual rainfall above and below this saturation point. Figure 10 shows scatterdiagrams of NDVI versus model-calculated soil moisture. For the "dry" stations there is a reasonably good relationship between NDVI and soil moisture (although some saturation occurs during the wettest years), but for the "wet" stations there is little relationship between the two variables. Similarly, the correlation between NDVI and rainfall is 0.29 for wet stations and 0.72 for dry stations. This supports the suggestion that the saturation point represents the transition from water-limited to radiation-limited environments. It also suggests that the transition occurs at a much lower threshold over Southern Africa than in West Africa. This is consistent with the higher NDVI–rainfall ratios over Southern Africa, as noted earlier.

4. Conclusions

The conceptualization of the evapotranspiration climatonomy model is consistent with current more physically based models. This made conversion to physical parameterization feasible and relatively simple. The results of the original and revised model are not vastly different so that the model is reasonably robust. However, the revised model has the advantages that it is readily transferable to other geographic regions and that soil types can be explicitly distinguished in model calculations.

The model has many potential applications. In numerous parts of Africa it has produced reasonable results when applied to diverse climatic problems. Although a one-dimensional model, it is readily adapted to the calculation of spatial fields. It has the advantage of producing water balance parameters representative of regional averages. It also has potential application to numerous ecological and climate sensitivity studies.

The excellent comparison with NDVI and the physically realistic relationships between soil type and soil moisture generation suggest that the model produces reliable results. The model was, however, explicitly verified using data for the HAPEX-Sahel experiment

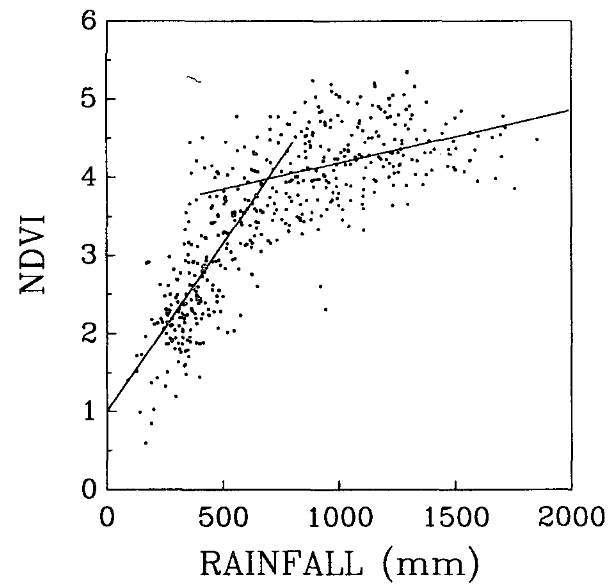


FIG. 9. Scatter diagram of annually integrated NDVI (sum of twelve monthly values) versus annual rainfall at 73 stations in Southern Africa. The solid lines indicate the two linear regression lines producing the best fit of the data (Draper and Smith 1981); the intersection of the lines is used to indicate the saturation of NDVI with respect to rainfall and delineate a dry and a wet regime.

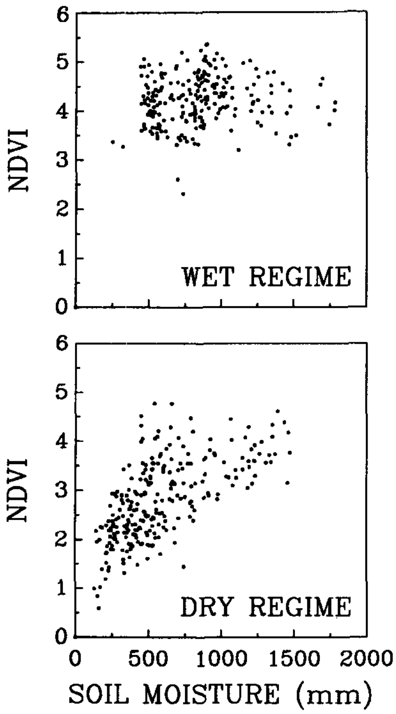


FIG. 10. Scatter diagrams of annually integrated NDVI and model-calculated soil moisture (mean of twelve monthly values) for wet and dry regimes, as defined in Fig. 9.

in Niger, West Africa. This is the subject of the second part of this article (Marengo et al. 1996). A second application of the model to HAPEX-Sahel data on daily timescales is described in Nicholson et al. (1996).

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