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Modelling Soil Physical Processes and Crop Growth in the Semi-Arid Tropics*

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Abstract

Process-based models help scientists to understand large systems and to extrapolate limited experimental information from one location to another. Unfortunately, most models of soil physical processes neglect the effects of spatial and temporal variability. For accurate representation of this field-scale variability in soil physical processes, models of these processes should use scaling techniques.

The soil water budget of the root zone and submodels of its components are discussed. Most of the submodels for infiltration are appropriate to homogeneous soil whose surface is ponded instantaneously. It is suggested that the Mein and Larson submodel of infiltration under rain, or a modification of it to accommodate variable rainfall intensity, would be appropriate for water budget models in the semi-arid environment. The transpiration component of the water budget is very well described by the Penman-Monteith model, which fails to describe the evaporation from a drying soil. And though the four-layer model of Choudhry and Monteith describes evaporation from both the leaves and the soil adequately, it has not been tested experimentally.

Two examples of crop growth models, CERES and RESCAP, are presented to illustrate how submodels of soil physical processes (e.g., a submodel of the water budget of the root zone) are used in conjunction with crop development and growth submodels to model the whole crop production system.

Introduction

When a scientist conducts a field experiment, he knows that his measurements and his conclusions may be specific to both site and season. He also knows that extrapolating site- and season-specific conclusions to other locations and years is a gamble. Yet he can never hope to work in every agroecological niche with every possible distribution of rainfall.

The scientist, therefore, has no choice but to extrapolate from very limited evidence. To do this with minimum risk of error, he must rely on the guidance of principles which enable him to understand the processes, help to bridge the gap in our knowledge between different levels of organization, and they can be used to set our ideas in the perspective of large

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systems. For example, crop growth and yield, as determined by weather and soil type, may be understood better by studying water budgets and nutrient dynamics, along with the development and growth of a crop. Several submodels may be used to describe each of these processes. Thus, a model of the water budget of the root zone of a crop will normally have rainfall, infiltration, runoff, evaporation, and deep drainage as component submodels.

Generally, before a scientist uses process models to predict the behavior of a whole system, he first identifies the problem, assesses the resources and constraints in the system, tries to understand the mechanisms governing various processes operating within the system, and then formulates and tests hypotheses. This procedure may lead to a prescription for solving a particular problem (Figure 1). Sometimes the measurement and analysis of parameters may lead to empirical relations whose coefficients are site- or season-specific. These empirical submodels are also often used in simulation models for large systems. In the Resource Management Program (RMP) at the International Crops Research Institute for the Semi-Arid Tropics (ICRISAT), we try to develop process models that are applicable in the semi-arid tropics (SAT) so that the benefits of crop simulation can be passed on eventually to farmers throughout the region.

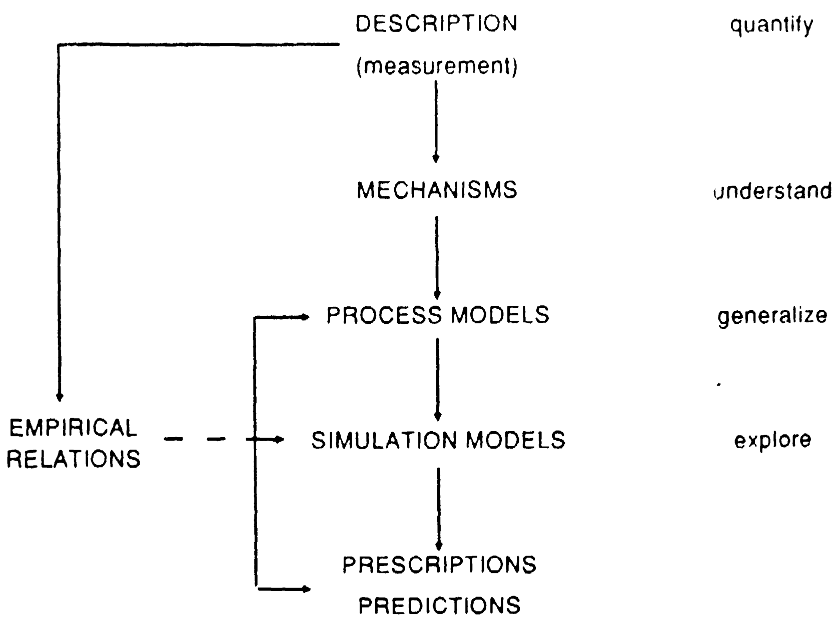

Figure 1. Steps in process-oriented research in agriculture.

In many crop models, the supply of water and nutrients is assumed not to limit growth. In the SAT, however, erratic weather, particularly variability in the rainfall distribution within and between seasons, infertile soils, and lack of capital for improvement characterize agricultural production systems. To model crop productivity in such situations, it is necessary to consider (i) processes occurring in the soil, which determine the availability of water and nutrients to the roots; and (ii) the processes that determine the rate at which roots grow into wetter zones of the soil. Also because temperatures in the SAT are high and often exceed a mean of 18°C throughout the year (Landsberg *et al.*, 1963; and Thompson, 1965), soil temperature models to predict seedling emergence and root growth are necessary. The literature on crop modelling is voluminous. In this paper, we review some of the models relevant to soil physical processes and crop growth in the SAT.

The water budget of the root zone

The difference between the amount of water added W(in) and the amount of water withdrawn W(out) from a given volume of soil during a specified period is equal to the increase in water content W during that period.

$$W = W(in) - W(out) \qquad (1)$$

As shown in Figure 2a, the value of W(in) includes precipitation P, irrigation Ir, upward capillary flow into the root zone U, and runon Ro, from areas higher in elevation than the field under consideration. The value of W(out) is the sum of runoff Rf, deep drainage D, evaporation E, and transpiration T. The evaporation term includes the amount of water that evaporates directly from the soil, part of the water that evaporates from temporary storage of water in surface depressions, and evaporation of water intercepted by leaves during rainfall. The water budget of the root zone becomes:

$$W = (P + Ir + U + Ro) - (Rf + D + E + T)$$
$$\text{(Gains)} - \text{(Losses)} \qquad (2)$$

Figure 2b presents a flow chart of the main processes in the model described by eq. (2) and illustrates how the process submodels described in the following sections fit into the overall water budget of the root zone.

If we define cumulative infiltration I as:

$$I = (P + Ir + Ro) - Rf \qquad (3)$$

and combine the evaporation term E with the transpiration term T, as ET, the model represented by eq. (2) then becomes,

$$W = (I + U) - (D + ET) \qquad (4)$$

which can further be simplified to

$$W = I - ET \qquad (5)$$

if it is assumed (as with most water budget models) that U and D are negligible compared to the other two terms. If U and D are large, as when roots approach a water table (in which case U would be substantial) or for sandy soils (where D may be relatively large in wet years), then eq. (4) has to be used.

Infiltration models

Many mathematical models, some empirical and others theoretically based, have been used over the years to describe the infiltration term in eqs. (3), (4), and (5).

The Green and Ampt (1911) and Philip (1957) models are based on one-dimensional downward infiltration into an

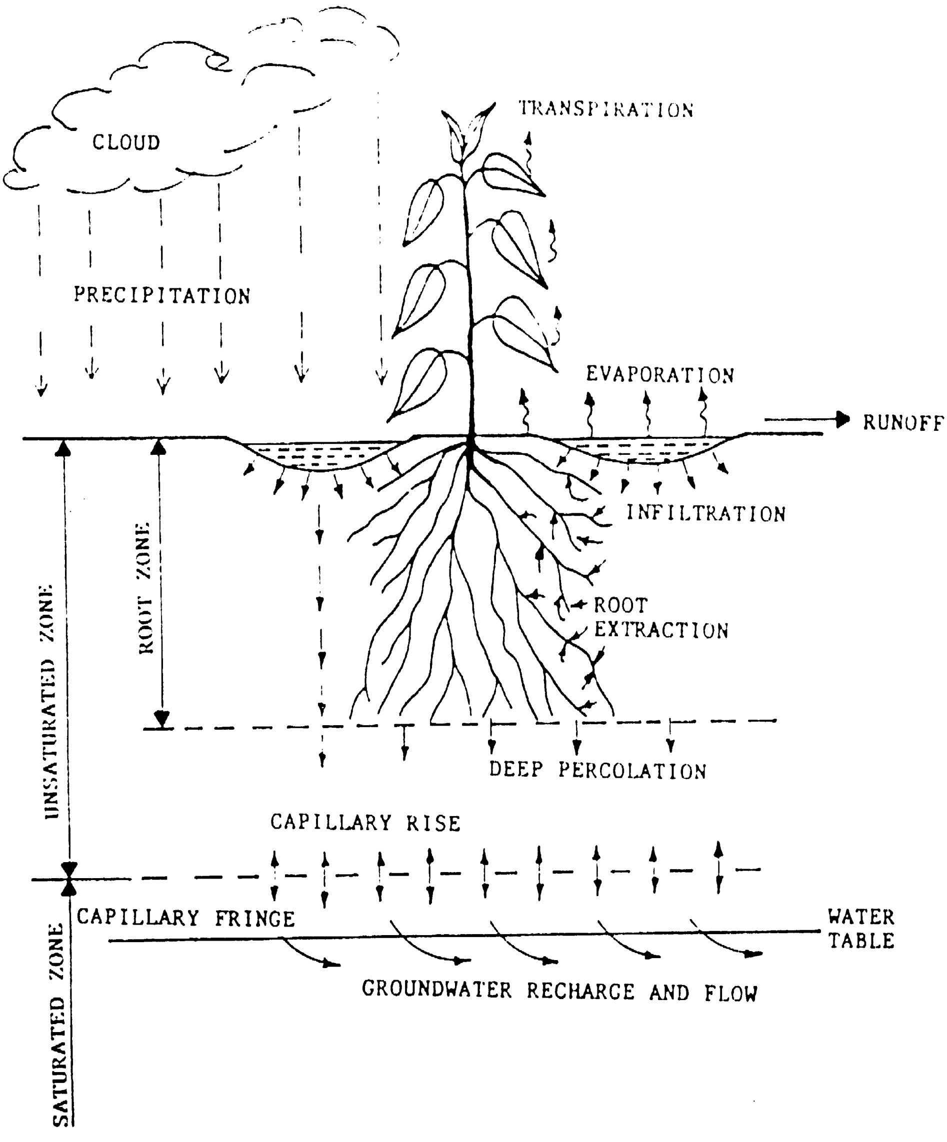

Figure 2a. The water budget of a root zone (schematic).

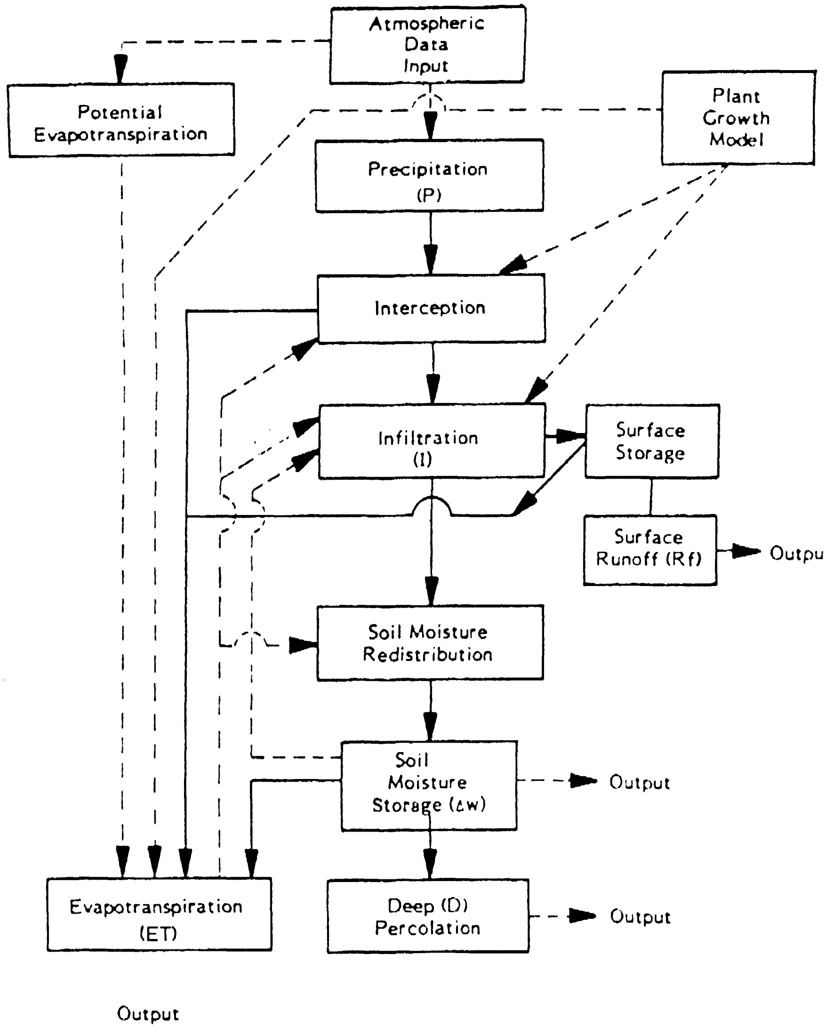

Figure 2b. Basic flow chart of the water budget model. Solid lines represent actual movement of moisture. Broken lines indicate flow of influence between model components.

infinitely long column of a nonswelling soil, with uniform initial water content w_i ($m^3\ m^{-3}$), whose surface (z=0) is ponded instantaneously and maintained at a saturated volumetric moisture content w_o ($m^3\ m^{-3}$). A further assumption in the derivation of the Green and Ampt model is that a sharp and definable wetting front exists between soil at w_i and soil at w_o and that the saturated hydraulic conductivity K_o (ms^{-1}) of the wet region is constant. This boundary condition of instantaneous surface ponding is perhaps appropriate to border irrigation of a field and modelling of field drainage problems, but it is less appropriate for infiltration under rain.

At the start of most rainfall events, all rain infiltrates but the capacity of the soil to absorb water declines until the infiltration rate is less than the rain intensity. At this point (often refered to as "time to ponding"), free water first appears at the surface of the soil and this marks the time beyond which both runoff and erosion may be initiated. The models of Green and Ampt, and Philip do not apply to the initial phase of rainfall when there is no ponding but the Mein and Larson (1973) model extends the Green and Ampt equation to describe constant rainfall infiltration.

(i) The Green and Ampt model

Considering water movement into soil under ponded conditions, Green and Ampt (1911) applied Darcy's law to a

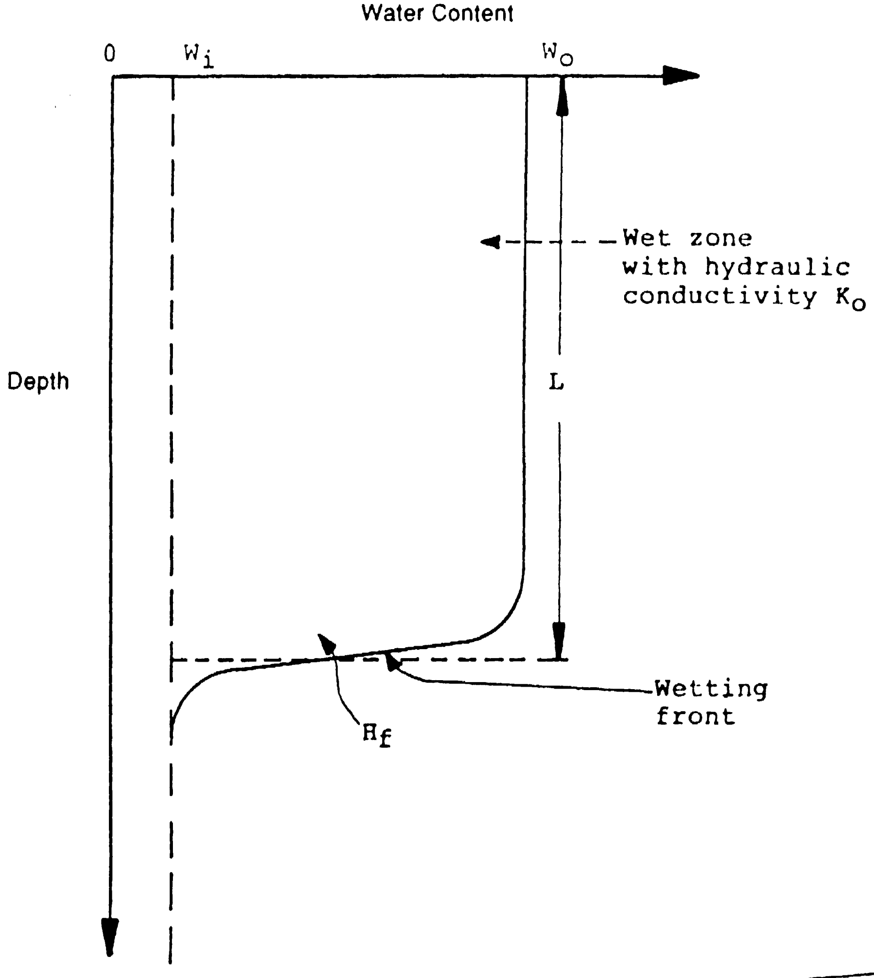

Figure 3. Schematic diagram of the moisture content profile for the derivation of Green and Ampt model.

vertical soil column (Figure 3) to yield and infiltration flux f (ms^{-1}).

$$f = K_o(L+H_f)/L = K_o(1+H_f/L) \quad (6)$$

where H_f is the average matric potential at the wetting front (m), which is the division between the soil at w_i ($m^3 m^{-3}$) and the wet region assumed saturated at water content W_o ($m^3 m^{-3}$) and L (m) is the distance from the soil surface to the wetting front. The cumulative depth of water infiltrated, I (m), is then given by

$$I = (w_o - w_i)L = nL \quad (7)$$

which when substituted into eq. (6) for L yields,

$$f = K_o[1 + nH_f/I] \quad (8)$$

If I = 0 at time t =0, then the integral form of eq. (8) may be written as,

$$I = K_ot + \lambda \, ln(1+I/\lambda) \quad (9)$$

which is convenient for modelling because it relates the infiltration rate to depth infiltrated from the start of infiltration. In eq. (9), $\lambda = (w_o - w_i)(H_o - H_f)$, and H_o is the pressure head at the soil surface.

The average matric potential H_f is usually calculated by integration of the soil water potential (ψ) versus relative hydraulic conductivity (K_r) curve so that

$$H_f = \int_{\psi_o}^{\psi_i} K_r d_\psi \quad (10)$$

In eq. (10), ψ_o is the saturated value of the soil water potential (m), and ψ_i the initial value of the soil water potential (m).

Green and Ampt's model has been validated for infiltration into initially dry sandy soils, which exhibit a sharp wetting front.

(ii) The Philip model

Philip's model was the first general solution of the differential equation governing one-dimensional downward infiltration as a function of time [I(t)]. His quasi-analytical solution is a power series with the form

$$I(t) = s_1t^{1/2} + s_2t + s_3t^{3/2} + ... + s_nt^{n/2} + K_it \quad (11)$$

in which the coefficient $s_2, s_3, ... s_n$ are calculated from the hydraulic conductivity K (ms^{-1}) and soil water diffusivity D (m^2s^{-1}) as functions of water content, and K_i is the unsaturated hydraulic conductivity at w_i. The first term in eq. (11) describes exactly the horizontal entry of water into soil under the same initial and boundary conditions (i.e., infinitely long column at w_i, whose surface is ponded instantaneously and maintained at w_o) but without gravity as a factor. The other terms are a consequence of the gravitational field. Terms beyond the fourth in the infinite series are usually very small at short time periods, but they become significant as t increases. In practice, it is generally sufficient to describe ponded infiltration for short time periods by the two-parameter form of eq. (11) which is:

$$I(t) = st^{1/2} + At \quad (12)$$

In eqs. (11) and (12), s_1 or s is the sorptivity, which describes the initial absorption of water by soil as a result of the matric potential gradients alone; A is a constant. At large times, the differential of eqs. (11) and (12) do not converge to the expected linear asymptote. In an attempt to extend the validity of the Philip two-parameter model to longer times, a number of authors (e.g., Morel-Seytoux, 1981; Youngs, 1982) have taken $A=K_o$ but have had only partial success. In a recent rigorous examination, Kutilek *et al.* (1988) have shown that the parameters K_o and λ in the Green and Ampt model (eq. 8) and the S and A in Philip's model (eq. 12) are all time dependent (Figure 4). Consequently, they may give erroneous results if used to extrapolate information to other places or for time periods beyond which they are valid.

(iii) The Mein and Larson model

Mein and Larson (1973) extended the Green and Ampt model to constant rainfall. Their two-stage infiltration model utilizes field values of saturated hydraulic conductivity and water content and is described by two equations. Stage 1, up to the time to surface ponding t_p, is of the form:

$$I_p = [H_f (W_o - W_i)] / [(f/K_o) - 1] \qquad (13)$$

where I_p is the cumulative infiltration up to the time of surface ponding. Equation (13) describes the infiltration process for a constant rainfall rate prior to the initiation of runoff. The second stage of postponding infiltration is described by an equation identical to the Green and Ampt model. Many other models varying in mathematical complexity have been published since Mein and Larson's derivation to describe infiltration during constant and also variable intensity rainfall (e.g., Morel-Seytoux, 1976, 1978, 1982; Hachum and Alfaro, 1977; and White *et al.*, 1982).

In many hydrologic models, the infiltration parameters (K_o and H_f of Green and Ampt or S and A of Philip) are used for curve fitting. This does not necessarily give an accurate representation of the process. Ideally, infiltration curves should be generated using independently measured soil properties, which can be used as input parameters for submodels of the infiltration process in the larger hydrologic model.

One of the major and as yet unmet challenges in hydrologic modelling is spatial and temporal variability; all the models discussed in the preceding sections and most of the others not considered here use an infiltration behavior determined or estimated at one site in the watershed. The spatial variability of basic hydraulic properties in agricultural lands is well recognized (e.g., Nielsen *et al.*,1973; Carvallo *et al.*, 1976; Warrick *et al.*, 1977; and Keisling *et al.*, 1977). Recently, Sharma *et al.*, (1980) have used a similar media concept and eq. (12) to scale ponded infiltration characteristics from 26 sites in a 9.6 ha watershed. The scaling technique thus provides a basis for representing soil spatial variability in terms of a single stochastic variable, the scaling factor which enables scattered data sets to be coalesced so that a mathematical relationship can be arrived for modelling. We, therefore, suggest that a scaled form of Mein and Larson infiltration submodel be used in large simulation models of the water budget of the root zone in order to overcome the problem of the variability caused by soil heterogeneity.

Evaporation models

Evaporation (ET) is a complicated process, which depends on many atmospheric, soil, and plant factors. A common approach is to estimate or measure the potential rate of ET from prevailing weather and then to compute the fraction of that potential achieved, given the current status of plants and soil. This procedure involves variables for determining (a) potential ET, (b) plant-water related characteristics, and (c) soil-water related characteristics. Many models of potential evaporation have been developed based on energy budgets, aerodynamic profiles, or a combination of the energy budget and equations

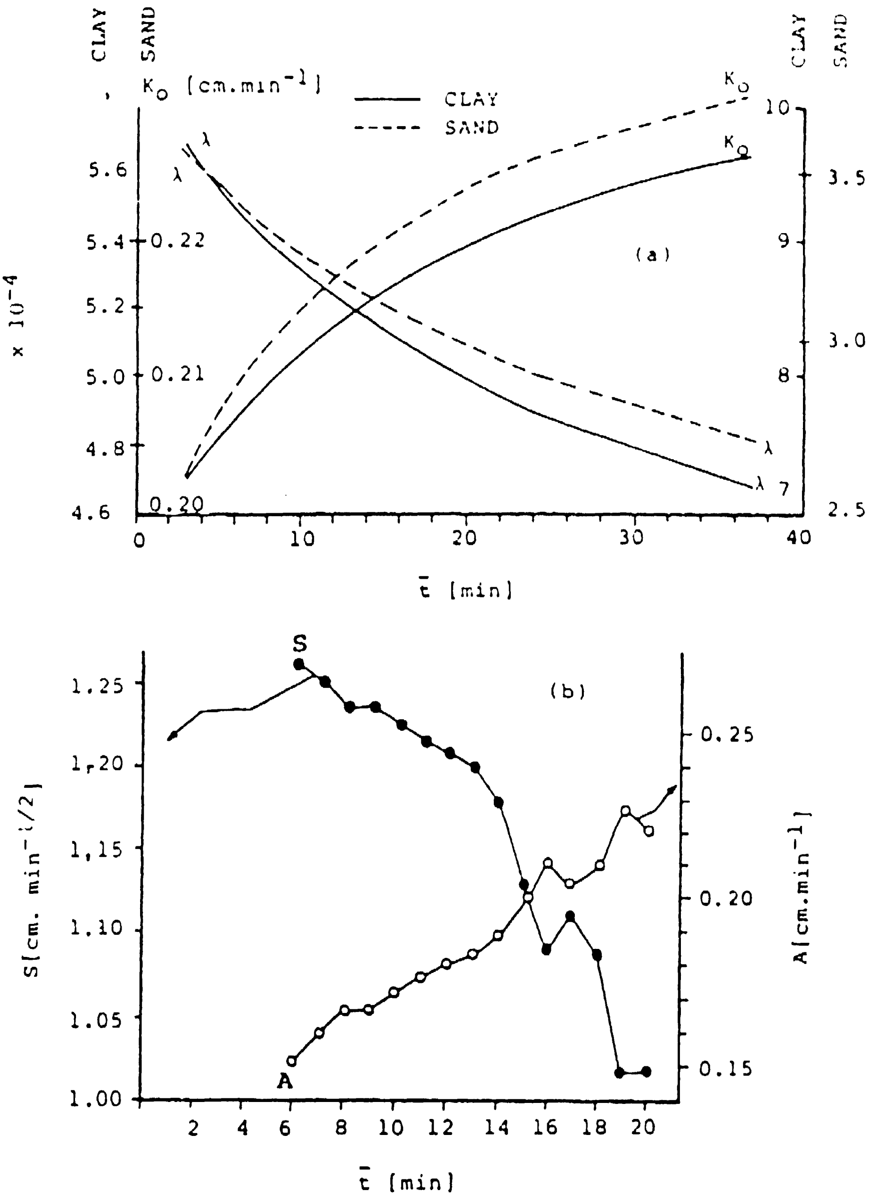

Figure 4. (a) *The dependence of the coefficients K_o and $\lambda = (\theta_s - \theta_i)(H_o - H_f)$ in the Green and Ampt equation (6) upon average time t.*

(b) *The dependence of the coefficients S and A in the Philip's equation (12) upon the averaged time t for the experimental infiltration data.*

Source: Kutilek *et al.* (1988).

governing the transport of heat, water vapor, and momentum in the lowest few meters of the atmosphere. Of the several models, we will briefly discuss the Penman (1948) equation, which represents one of the more reliable models for potential evaporation (PE), the Penman-Monteith (Monteith, 1981) equation, which is an extension of Penman's model to transpiration T, and the Ritchie (1972) model, which is also a modification of the Priestley and Taylor (1972) equation for soil evaporation.

(i) Penman's model

The classic PE model of Penman (1948) was derived by eliminating surface temperature from the energy budget and aerodynamic transport equations. Originally, this model was formulated to describe evaporation from a water surface. Without going into its derivation, Penman's equation contains basically two components relating to (i) adiabatic heat and water vapor exchange and (ii) diabatic exchange (Monteith, 1981). In symbols,

$$LE = LE_a + LE_d \tag{14}$$

where LE_a is the amount of latent heat transferred adiabatically (i.e., all the heat from evaporation is provided by the transfer of heat from air with temperature T_a and vapor pressure e_a to a wet surface; and LE_d is additional heat supplied to saturated air, e.g., by the absorption of radiation (Monteith, 1981)

The components LE_a and LE_d are given by

$$LE_a = (\rho c/r_a)\{[e_s(T_a) - e_a] / [\Delta + \gamma]\} \tag{15}$$

and,

$$LE_d = (\Delta H)/(\Delta + \gamma) \tag{16}$$

where ρc is the volumetric specific heat of air at constant pressure, r_a is the resistance for heat and mass transfer, γ is the psychrometer constant, Δ is the rate at which saturation vapor pressure increases with temperature, $e_s(T_a)$ is the saturation vapor pressure at air temperature, e_a is the prevailing vapor pressure of air, H is the difference between the net radiation (R_n) of the surface and the rate G at which heat is conducted away from the surface.

(ii) The Penman-Monteith model

Penman (1953) considered a single leaf as providing an additional resistance r_s to heat and vapor exchange between the leaf surface and the atmosphere, and he treated such a system as isothermal because the wet walls surrounding the substomatal cavity (source of vapor) are effectively at the same temperature as the epidermis (source or sink of heat). He modified γ to

$$\gamma^* = \gamma(1 + r_s/r_a) \tag{17}$$

Equation (17) was extended by Monteith (1965, 1981) for a crop canopy by assuming that the effective sources (or sinks) for heat, mass, and momentum are all at the same level. The resistance r_s can then be interpreted as the effective stomatal resistance of a set of n parallel resistors, each representing a layer of canopy. The resistance r_s of the whole canopy is given by

$$r_s = 1 / \sum_{i=1}^{n} (L_i/r_i) \tag{18}$$

where r_i is the stomatal resistance and L_i is the leaf area index of layer i. In principle, by obtaining representative measurements of ri with a porometer and of L_i by sampling, r_s can be estimated, and eqs. (15), (16), and (17) can then be used to obtain an estimate of transpiration (Azam-Ali *et al.*, 1984). This is not a routine method, however, because of the labor of sampling.

Equation (14) has been found not to be valid for evaporation from a drying soil because the pores from which water is evaporating have usually a different temperature from the surface where sensible heat exchange with the atmosphere occurs (Fuchs and Tanner, 1967). To apply eq. (14) to the rate of evaporation E_s from a drying soil, Monteith (1981) assumed that evaporation of water occurs from wet soil below an isothermal layer of increasing thickness and described E_s with a model,

$$E_s = (2t/A^* + 1/E^2)^{-1/2} \qquad (19)$$

where E_o is the evaporation rate at time t=0 and is determined by the state of the atmosphere. In eq. (19), A^* is defined as

$$A^* = r_{sa} (\Delta + \gamma) E_o / m\gamma \qquad (20)$$

where r_{sa} is the aerodynamic boundary layer resistance for heat transfer between the soil surface and air at a reference height (e.g., screen height), and m is a function of the liquid and gaseous diffusivities of the soil.

In a recent analysis, Choudhry and Monteith (1988) presented a model for heat budget of homogeneous land surfaces, which assumes four layers defined by the following boundaries:

(i) from a reference height in the atmosphere to an effective sink for momentum within the foliage or the soil surface if foliage is absent,

(ii) from the virtual sink to the soil surface,

(iii) an upper completely dry layer of soil extending down from the surface, and

(iv) a lower wet layer of soil.

Differences in heat and vapor flux across these layers were specified by resistances, e.g., surface resistance to vapor loss from the foliage, and soil resistance to vapor loss assumed proportional to the depth of the dry layer. Using these assumptions, Choudhry and Monteith (1988) derived a model for transpiration incorporating eqs. (17) and (18) and a new soil evaporation equation, allowing to some extent for temperature gradients in the soil, viz.

$$LE_s = \{x\Delta' R_s + \rho c [\Delta' (T_m - T_b)/ r_l + n D_b/r_2]\}/(\Delta' + \gamma^*) \qquad (21)$$

where Δ' is the rate at which saturation vapor pressure at the interface between the dry and the wet soil changes with temperature, R_s is the net absorption of radiation by the soil surface, T_m is the temperature at the bottom of the wet soil layer, T_b is the temperature of air in the canopy, r_l is the resistance proportional to the depth of the wet layer, D_b is the saturation deficit in the canopy, and r_2 is the resistance between the soil surface and the canopy. In eq. (21), x, n and γ^* are functions of resistances defined by

$$x = (1 + r_u/r_2)^{-1} \qquad (22)$$

$$n = [1 + (r_u + r_2)/r_l]\, x \qquad (23)$$

$$\gamma^* = r(1 + r_d/r_2)/n \qquad (24)$$

where r_u is the thermal resistance proportional to the depth of the dry soil layer and r_d is the corresponding resistance to the diffusion of water vapor. Equation (21) and the associated definitions in eqs. (22), (23), and (24) are yet to be validated experimentally but conclusions from the model about the interaction of water vapor fluxes from soil and from foliage, are consistent with field observations.

An example of the use of the Penman-Monteith submodel to estimate evaporation is the Prosper model of soil-plant-atmosphere water flow (Goldstein and Mankin, 1972), which has been coupled to the Terrestrial Ecosystem Hydrology Model (TEHM), (Huff *et al.*, 1977) to provide a mechanistic watershed model (Swift *et al.*, 1975; Luxmoore *et al.*, 1977; Peck *et al.*, 1977; and Sharma and Luxmoore, 1979).

(iii) Ritchie's model

When a crop is sown, the field is bare (except in cases where mulch is applied) until the seeds germinate. Then follows a vegetative growth period, during which complete canopy cover may be established. During the period when the field is bare and also at full canopy, eqs. (14), (15), (16), and (17) may be used successfully. Between germination and the establishment of full canopy cover or if there is poor seedling establishment, canopy cover is incomplete and eqs. (14), (15), (16), and (17) will not apply. Ritchie (1972) assumed that the term containing the vapor pressure deficit $[e_s(T_a) - e_a]$ and the aerodynamic resistance could usually be neglected so that substitution of eqs. (15) and (16) into eq. (14) gives,

$$PE_s = \Delta H_s/(\Delta + \gamma)L \qquad (25)$$

where PE_s is the potential evaporation from soil, and H_s is the difference between the net radiation R_{ns} at the soil surface and the soil heat flux. Net radiation at the soil surface was calculated by applying Beer's law to the interception of the downward net flux within foliage to give

$$R_{ns} = R_n \exp(-0.4\,L_l) \qquad (26)$$

All the CERES (Crop Estimation through Resource and Environment Synthesis) crop growth models use Ritchie's submodel to estimate the evaporation term in the water budget subroutine.

Crop growth models

The main processes involved in crop growth are related to fields of knowledge that have developed independently of each other (Figure 5). Models that attempt to simulate crop growth usually contain elements from these fields which need to be coordinated. Consequently, even though the subject of this paper deals primarily with soil physical processes and crop growth models, parts of the discussion that follow touch on processes in other disciplines that are important in crop growth.

Apart from possible nutrient deficiencies in soils, crop growth in the SAT is limited by water shortage during part of the season. The availability of water determines the duration of the growth period. This is particularly true of the dry regions of the SAT, like the Sahelian zone. However, when water is freely available, crop growth rate increases up to a maximum determined by weather and by the fertility level of the soil. Water shortage results in stomatal closure, which in turn reduces the assimilation of carbon dioxide. The water budget aspects of modelling have been considered extensively by many researchers in the past, but it is only recently that some of the seemingly intractable problems associated with modelling the nutrient budget have been tackled. The water and the nutrient budgets interact in their influence on crop growth, and modelling of the interaction is

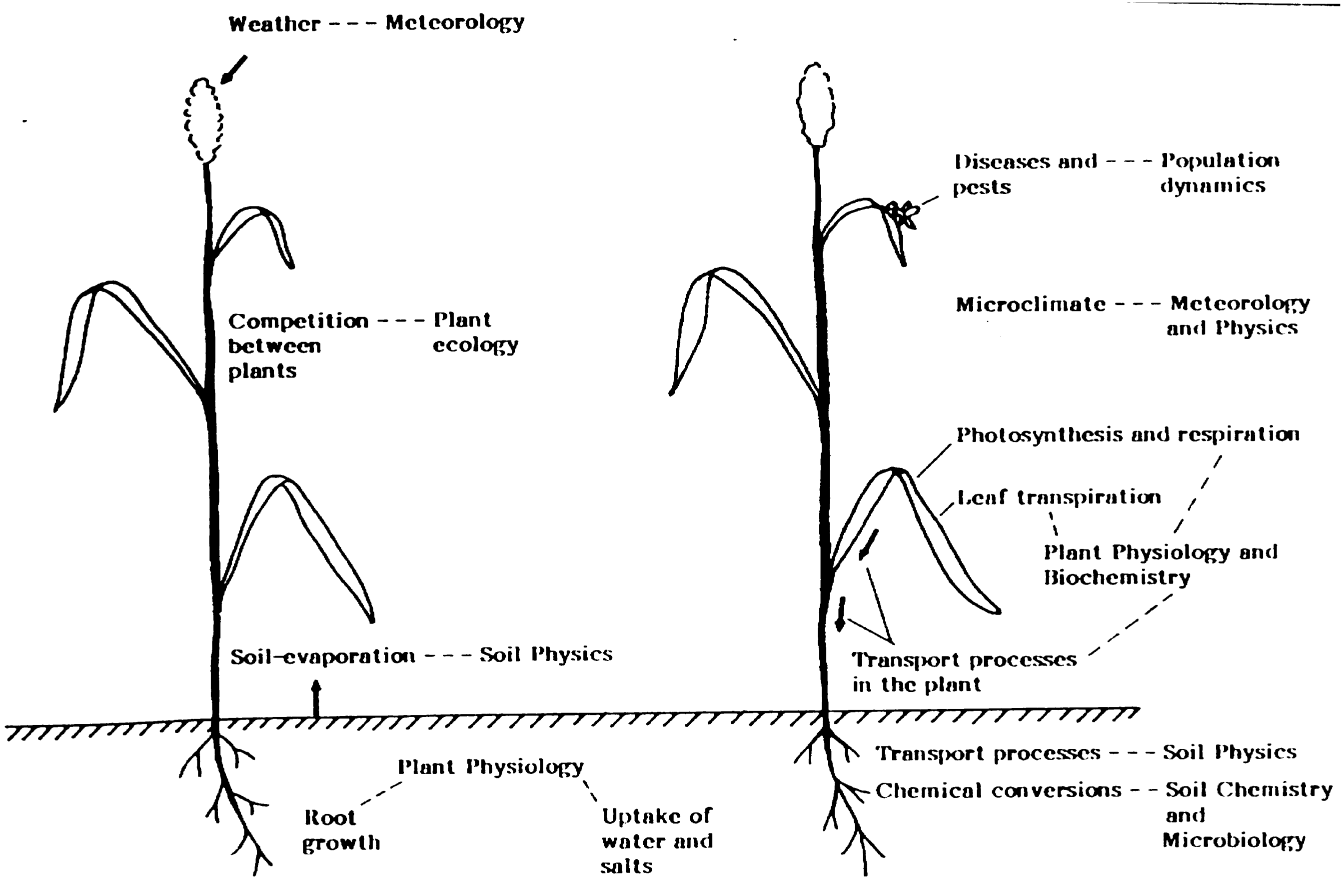

Figure 5. Fields of knowledge that need consideration in a study of plant growth.

essential for successful crop growth models.

We will discuss two crop models, the CERES model (Ritchie, 1985) and the Resource Capture (RESCAP) model (Monteith *et al.*, 1988). The first model considers the relationship between water use by plants and the associated dry-matter production, while the second emphasizes the role of shoots and roots in relation to the control of dry-matter production by the supply of radiant energy *or* water.

(i) The CERES crop models

There are currently CERES crop models for wheat, sorghum, millet, maize, and rice. In this discussion, only the main common features will be outlined, leaving out detail to be found in Jones and Kiniry 1986; Ritchie and Otter, 1985; and Ritchie and Alagarswamy, 1988. The CERES models deal with the main factors that determine the final yield of the crop, i.e., (i) the development and duration of growth phases related to plant genetics and the environment, (ii) morphogenesis of the vegetative and the reproductive structures, (iii) growth of leaves, stems, and senescence of leaves, (iv) biomass accumulation and partitioning between leaves, stem, and roots, (v) effect of soil water on growth and development, and (vi) effect of nitrogen on growth and development (Ritchie, 1985).

The models evaluate the soil water budget of a crop or fallow land, using a submodel similar to eq. (2). The soil profile is divided into layers, and the limit to which water can increase in the layer, i.e., drained upper limit (DUL), and saturated upper limit (SUL) together with the lower limit (LL) of plant water availability are inputs for each soil layer. Values for these limits must be obtained in the field and not from conventional laboratory measurements of wilting point and field capacity, which have been found inaccurate when used in these models. Daily rainfall and irrigation (if applied) are used as inputs. Infiltration of water into the soil is calculated as the difference between rainfall or irrigation and runoff (see eq. 3). Runoff is calculated using the USDA-Soil Conservation Service (1972) curve number technique, which specifies runoff curves by numbers varying from 0 (no runoff) to 100 (all runoff). The Soil Conservation Service (USDA, 1972) handbook provides runoff curve numbers for various hydrological and soil-cover complexes.

The drainage rates are calculated using an empirical relation of the form,

$$w_t = (w_o - w_u) \exp(-K_d\, t) + w_u \tag{27}$$

where w_o is the saturated volumetric water content, w_u is the drained upper limit water content, w_t is the water content for any time t after field saturation, and K_d is a conductance parameter.

Evaporation is calculated using a modified version of eq. (25) (Priestley and Taylor, 1972), and eq. (26) is used to separate transpiration from soil evaporation on the basis of radiant energy reaching the soil, the time after the surface layer was wet, and the leaf area index.

The CERES model calculates absorption of water by roots using the "law of the limiting" approach, in which the larger of the soil or root resistances determines the flow rate of water into roots. The radial resistance to water flow into a single cylindrical root is assumed to vary with soil-limited water absorption rate q, such that

$$q = \{4\pi\, K(w)\, [h_r - h_s]\} \,/\, \ln\, (c^2/r^2) \tag{28}$$

where $K(w)$ is the hydraulic conductivity of the soil, h_r is the water potential at the root surface, h_s is the water potential of the bulk soil, r is the root radius, and c is the radius of the cylinder of soil through which water is moving. In the CERES models, $K(w)$ (cm day^{-1}) is calculated using an empirical relation:

$$K(w) = 10^{-5} \exp [62(w-LL)] \qquad (29)$$

The plant-limited flow rate is taken to be approximately 0.03 cm day^{-1}. The smaller value of the maximum soil-limited absorption and the plant-limited flow rate is converted into an uptake rate for a layer of soil, using the root length density and the depth of the soil layer. The sum of the maximum root absorption from each soil layer gives the maximum uptake of water from the profile. If this uptake rate is less than the maximum transpiration rate, the transpiration rate is set equal to the maximum absorption rate. On the other hand, if the maximum absorption rate is greater than the transpiration rate, the maximum absorption calculated for each layer is reduced proportionally so that the uptake rate from the profile equals the transpiration rate.

Potential dry-matter production is considered in the CERES crop models to be a linear function of intercepted photosynthetically active radiation (PAR) (i.e., wavelength bands from 0.4 to 0.7 microns). For wheat it is assumed that 3.05 gram of total biomass are produced for every MJ of PAR intercepted, whereas the constant for conversion for sorghum dry-matter production is 4.0 gram per MJ of intercepted PAR. The proportion of incoming PAR transmitted by a canopy is assumed to be an exponential function of leaf area index L_i.

CERES crop model performance

Figure 6 presents an example of a comparison of simulated wheat yield with measured yield from about 25 sites, ranging in latitude from 36° S in Australia to 50° N in England. The mean of the absolute difference between estimated and measured yields for the 168 data sets was 1070 kg/ha, with a standard error of ± 60 kg/ha (Ritchie and Otter, 1985). Figure 7 presents an example of a comparison of the simulated and measured grain yield of sorghum at three different sites. The observed and predicted values are scattered close to the 1:1 line.

(ii) The RESCAP model

The RESCAP model was developed by Monteith *et al.* (1988) primarily to predict the growth and yield of sorghum and pearl millet, given appropriate environmental conditions and genetic coefficients. All the parameters used in the model can be measured in the field. It is general enough to be adapted to any seed-producing crop. The model assumes that (i) at all growth stages, the rate of production of dry matter C (kg m^{-2} day^{-1}) per unit of intercepted solar radiation S (MJ m^{-2} day^{-1}) is effectively constant and has value e when water is not limiting; (ii) the amount of dry matter produced per unit of water transpired is inversely proportional to mean saturation deficit whether water is limiting or not. Growth is also assumed to be light-limited if the demand for water to transpire is less than the amount of water that the roots can supply and is calculated as,

$$C = f_i \, e \, S \qquad (30)$$

where f_i is the fraction of radiation intercepted by foliage and estimated by a relationship similar to eq. (26),

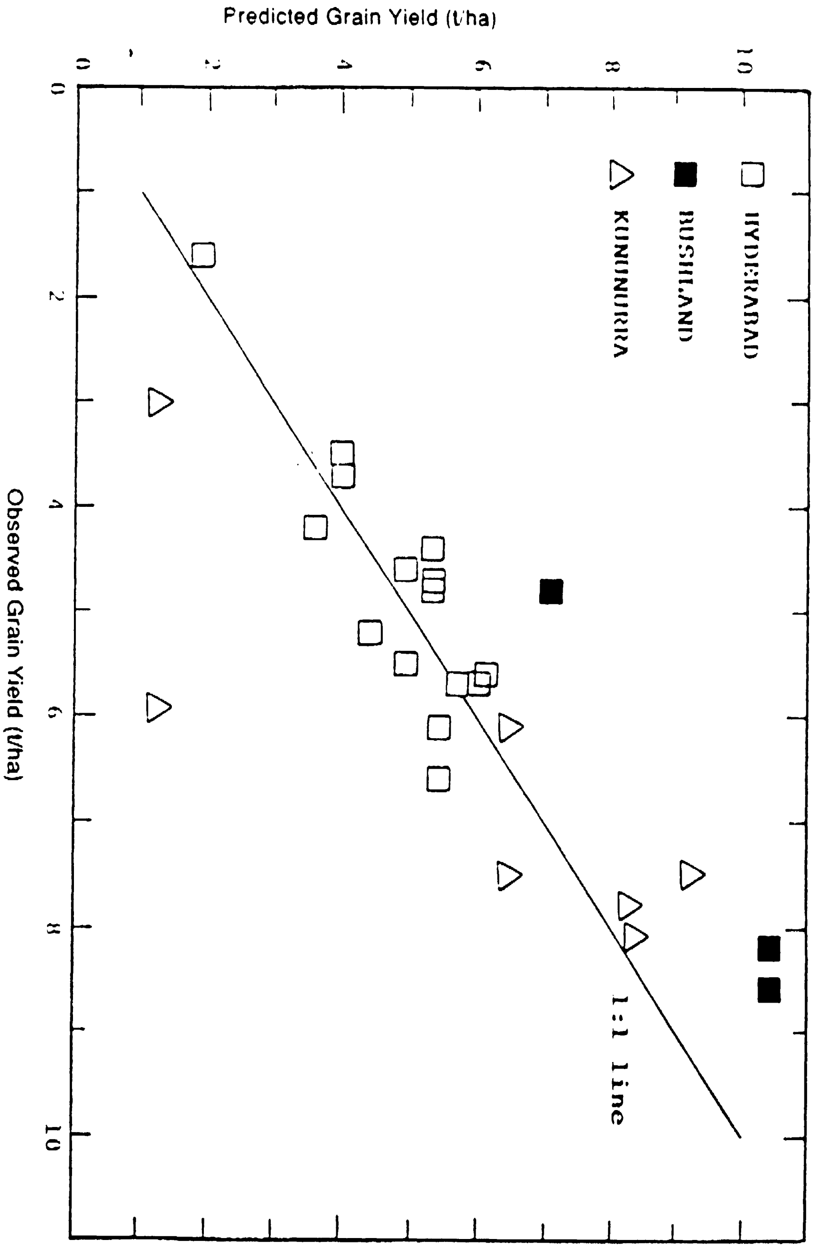

Figure 7. *Relation between predicted and observed grain yield of sorghum, using the CERES model.*

Source: Ritchine and Alagarswamy (1988).

$$f_i = 1 - \exp(-kL_i) \quad (31)$$

where k is a radiation extinction coefficient and L_i is the leaf area index which is increased daily by dL_i (= increase in dry weight multiplied by leaf area ratio D). The leaf area ratio (m^2 leaf per g plant) is calculated as

$$D = A_l (1 - x_r - x_s) \quad (32)$$

with A_l as the specific leaf area (m^2 leaf per g leaf), x_r as the fraction of dry matter allocated to roots, and x_s as the fraction of dry matter allocated to stems.

Extraction of water

The size and the distribution of the root system is specified in the RESCAP model by

(i) the downward velocity of the root "front" (u_r) to reflect the thickness of the layer of soil traversed by the root in a day. For sorghum the velocity of the root front measured by using a neutron probe to follow the downward movement of the drying front during a period when the crop received no water was found to reach a maximum value of 0.035 m day^{-1} about 20-30 days after emergence (Monteith, 1986).

(ii) the root length per unit volume l_v which is a function of depth and therefore of time. The values of l_v is assumed to be inversely proportional to the square root of rooting depth d_r, so that l_v at d_r is

$$l_v(d_r) = l_v(1)\,(d^* / d_r)^{0.5} \quad (33)$$

where d^* is the minimum root depth and $lv(1)$ is the root length per unit volume for layer 1. Since x_r is the fraction of dry matter allocated to roots

$$x_r C = \rho l_v u_r \quad (34)$$

In eq. (34) ρ is the root weight per unit length (kg m^{-1}) and C and u_r are both functions of time. During the initial stages of growth, x_r is set arbitrarily at a value of 0.3 and eqs. (33) and (34) are used to compute u_r. When u_r reaches its maximum value of u^*, u_r is then set at u^* and x_r is allowed to decrease with increasing depth.

The available water content of a soil layer permeated by roots from a time t=0 is assumed to decrease exponentially with time, following Passioura (1983), and is calculated from the expression,

$$AW(z,t) = w_e(z) \exp\{-[t/k l_v]\} \quad (35)$$

where AW(z,t) is the available water, which is a function of depth and time t, $w_e(z)$ is the total amount of water at depth z extractable by roots between t=0 [w (z,0) = $w_e(z)$] and t infinite (w = 0), and k is an arbitrary constant with the dimensions of a diffusion coefficient. Differentiation of eq. (35) multiplied by u_r gives the amount of water extracted from a layer of soil at time t.

Evaporation from the soil surface

The rate of evaporation E_s from wet soil without cover was assumed to be 0.9 times the rate of evaporation from a class A pan EP. Ground cover reduced the rate by a factor of $(1-f_i)$. E_s is then given by

$$E_s = 0.9\,(1-f_i)\,(w - w_a)/(w_{fc} - w_a)\,EP \quad (36)$$

where w is the actual water content of the layer, w_a is the air-dry water content, and w_{fc} is the water content at field capacity. The air-dry water content is assumed to be 1/3 of the value at 1.5 MPa.

Phenology

Phenology is divided into the usual

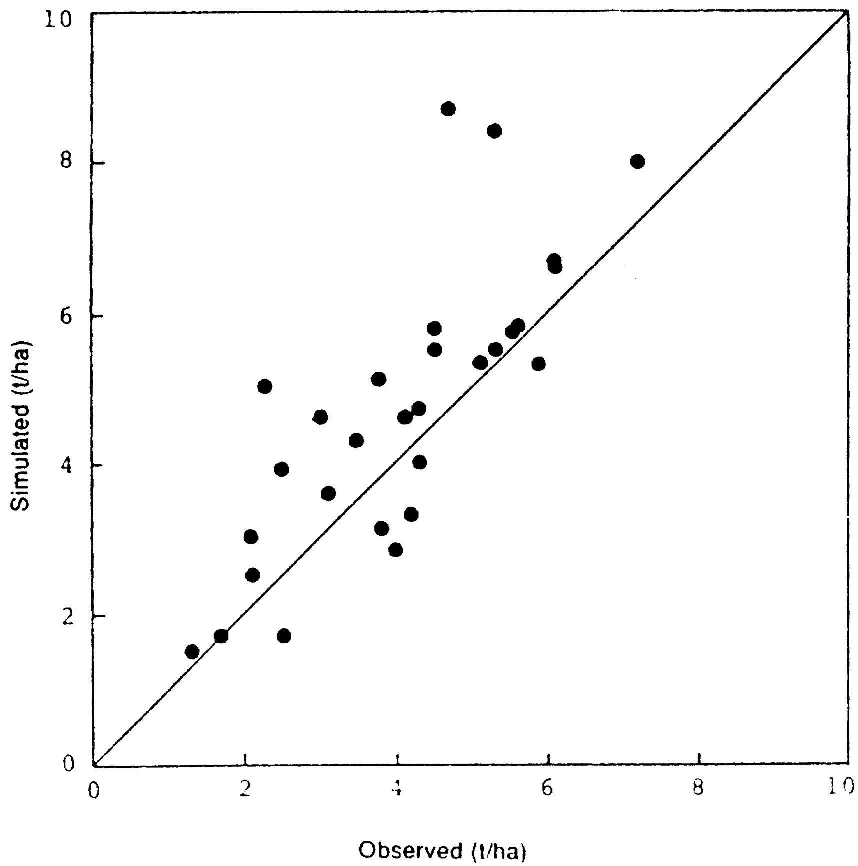

Figure 8. *Simulated versus measured sorghum grain yields for 29 crop-year data sets.*

Source. Monteith *et al.* (1988).

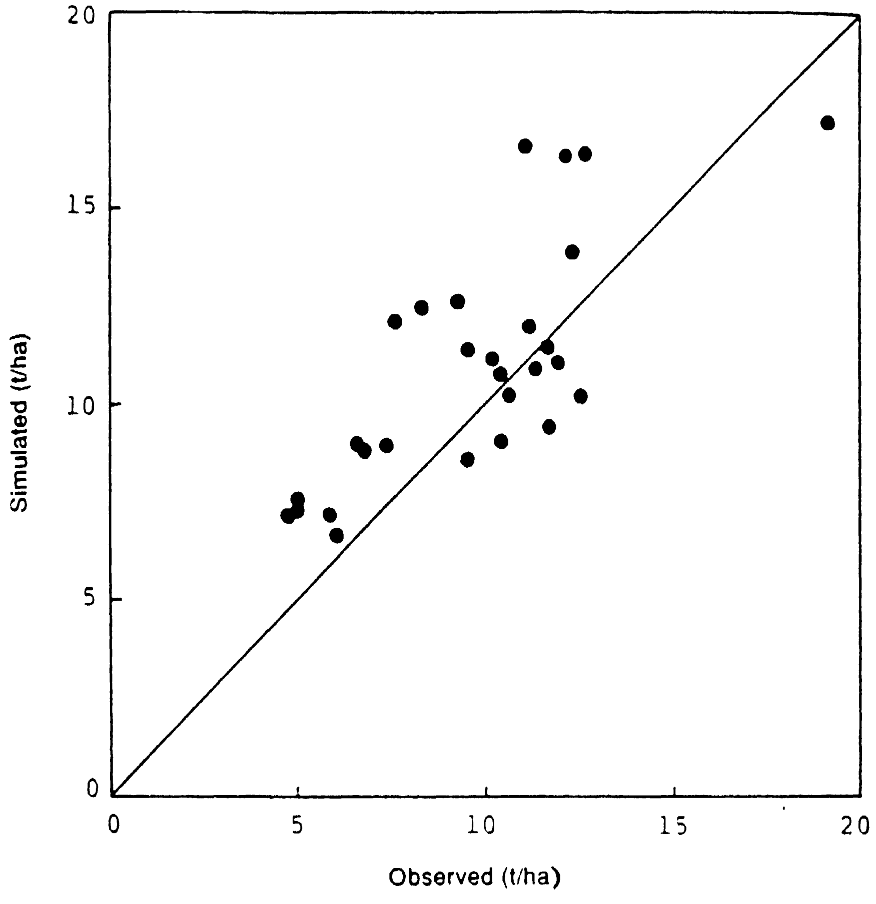

Figure 9. *Simulated versus measured total dry-matter production of sorghum for 29 crop-year data sets.*

Source: Monteith *et al.* (1988).

three stages : GS1 from emergence to panicle initiation; GS2 from panicle initiation to anthesis; and GS3 from anthesis to maturity. The length of each stage is specified in terms of thermal time above a base of 7°C, and daylength. The daily mean temperatures were assumed to be the average of the reported maximum and minimum temperatures, except when the maximum temperature exceeds 38°C in which case a value of 38°C was assigned.

RESCAP crop model performance

An example of a comparison between simulated sorghum grain yield and measured grain yield for 29 experiments conducted at ICRISAT is presented in Figure 8. Figure 9 also presents a comparison between the simulated and measured total dry matter for the same 29 experiments. Except for three points in Figure 8, the agreement between measured and simulated grain yield and also that between measured and simulated dry-matter production is very good.

Conclusion

We have discussed in this paper some of the infiltration and evaporation submodels often used in water budget models. Of the many models for infiltration, the Green and Ampt and the Philip two-term models have been used extensively by many researchers. Both models are for infiltration processes where the soil surface is ponded instantaneously, e.g., flood or border irrigation. Therefore, they do not adequately describe infiltration under rainfall conditions. The Mein and Larson equation, however, models infiltration during rainfall events and is an appropriate component for a model of crop growth under rainfed conditions. All or most of the infiltration models have a serious drawback in that they disregard spatial and temporal heterogeneity of soils.

On closer examination the Choudhry-Monteith four-layer evaporation model is appealing because it also includes evaporation from drying soil. However, it has not been tested experimentally and it also involves intricate mathematical expressions, which may not be user-oriented.

Two examples of crop growth models have been presented to illustrate how different forms of water budget submodels are used together with crop development and growth submodels to model the whole crop production system. The RESCAP model is still being refined and it offers a new approach to crop modelling. There is an urgent need to incorporate nitrogen and phosphorus dynamics and effects of pests and diseases in crop growth models.

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