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COMPUTATION OF SOIL WATER BALANCE AND ITS APPLICATIONS IN RAINFED AGRICULTURE

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Soil water balance is an important parameter for numerous applications in agriculture, forestry and hydrology. Soil moisture plays an important role not only for plant growth, development and yields but also for farm operations and practices such as planting, cultivation, harvesting and irrigation. Various climatic, soil and plant factors affect water balance (or components of water balance) in soil. Climatic factors are rainfall (amount, duration and intensity) and potential evapotranspiration of a location. Main soil factors are intake, storage and release characteristics of a soil. Plant factors are crop cover (leaf area index), rooting and water transport characteristics of a crop.

Study of water balance is fundamental to optimise the use of rainfall for crop production especially in the rainfed farming areas. Since rainfall in tropical areas is undependable in terms of its amount, intensity and duration, it is very important to characterize soil water availability (i.e., when, where and how much water is available in the soil profile). This helps match the cropping duration with soil water availability period and

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quantify the risks involved in crop production. Another component of water balance which is rather equally important is runoff or water loss (runoff + drainage). This water could be made available for plant growth or other purposes by water harvesting in farm ponds.

There are several kind of water balance models available in literature which vary in complexity depending upon the intended use of the model. The purpose of this paper is to review various approaches found in literature to quantify or estimate the components of water balance and to give examples on the use of water balance models in agriculture.

SOIL WATER BALANCE MODEL

The basic components of the soil water balance model are (1) the additions of water (precipitation or irrigation, and upward or lateral movement), (2) the losses of water (evapotranspiration, surface runoff, ground water flow and deep percolation), and (3) the change in the plant available water or extractable water content. The components are not independant but are interrelated. For example, amount of runoff depends partially on the rainfall intensity, hydraulic properties of the soil, and surface water content. Quantitatively, the water balance could be represented by the following equation, i.e.,

$$P + I = R + D + E + T + \Delta M$$

Where P = Precipitation

I = Irrigation

R = Surface runoff

D = Deep drainage

E = Soil evaporation

T = Transpiration

ΔM = Change in soil water storage

Precipitation or irrigation are usually measured and other components of water balance are estimated.

Potential evapotranspiration: To estimate the loss of water from soil as evapotranspiration, it is important to know the evaporating power of the environment which is also termed as potential evapotranspiration (PET). According to Penman (1948), PET is defined as the maximum quantity of water which may be evaporated by a uniform cover of dense short grass when water supply in the soil is not limiting. This method has been adopted by FAO (Frère and Popov, 1979) with slight modifications for their water balance studies. Climatological records of temperature, vapour pressure or relative humidity, sunshine duration and wind speed will allow the calculation of PET. The Penman formula reads as follows:

$$PET = \frac{\frac{P_o}{P} \cdot \frac{\Delta}{\gamma}\left[0.75\ R_A\left(a+b\,\frac{n}{N}\right) - \sigma T_k^4\ (0.56 - 0.079\sqrt{e_d})\ \left(0.10 + 0.90\frac{n}{N}\right)\right] + .26\ (e_a - e_d) .\ (1.00 + .54U)}{\frac{P_o}{P} \cdot \frac{\Delta}{\gamma} + 1.00}$$

Where PET = estimation of the potential evapotranspiration for a given period, expressed in mm;

P_o = mean atmospheric pressure expressed in millibars at sea level;

P = mean atmospheric pressure expressed in millibars as a function of altitude for the station where the estimate is calculated;

Δ = rate of change with temperature of the saturation vapour pressure expressed in millibars per degree C;

γ = the psychometric coefficient for the psychometer with forced ventilation = 0.66;

0.75 and 0.95: factors expressing the reduction in the incoming short wave radiation on the evaporating surfaces and corresponding respectively to an albedo of 0.25 and 0.05;

R_A = short wave radiation received at the limit of the atmosphere expressed in mm of evaporable water (1 mm = 59 calories) and taking for the solar constant the value of 2.00 $cal.cm^{-2}.min^{-1}$;

a and b = coefficients for the estimation of total radiation from the sunshine duration (a = .25 and b = .45);

n = sunshine duration for the period considered in hours and tenths;

N = sunshine duration astronomically possible for the given period;

σT_k^4 = Blackbody radiation expressed in mm of evaporable water for the prevailing air temperature;

e_a = saturation vapour pressure expressed in millibars;

e_d = vapour pressure for the period under consideration expressed in millibars;

T_c^o = air temperature measured in the meteorological shelter and expressed in degrees Celsius;

T_k^o = air temperature expressed in degrees Kelvin where

$$T_k^o = T_c^o + 273;$$

$U_{m/s}$ = mean wind speed at an elevation of 2 m for the given period and expressed in m/sec.

Priestley-Taylor Formula

Priestley-Taylor method is used where ET is energy limited occuring from well-watered surface during non-advective conditions. Several investigators (e.g., Jury and Tanner, 1975; Kanemasu et al., 1976) have successfully shown that Priestley-Taylor (1972) formula estimates ET for conditions of adequate water and leaf area index ≥ 2.5, which is given as

$$PET = \alpha\,[S/(S+\gamma)]R_n$$

Where α is proportionality constant for a particular crop and climate and increases with advection; γ is psychrometric constant (mb/°K) at mean temperature, and Rn is the 24-hour radiation (mm water/day).

PET from Class 'A' Pan

To calculate PET, the following formula as suggested by Penman (1948) could also be used

$$PET = a\,E_o$$

Where a is an empirical coefficient that depends upon the crop and stage of plant development and Eo is daily class 'A' pan evaporation.

Estimation of actual evapotranspiration

There are two approaches which are generally followed in the estimation of actual evapotranspiration:

a) Estimation of ET using crop coefficients and soil dryness curves.

b) Estimation of evaporation and transpiration separately since these are two different processes (one physical another physiological).

a) Estimation of ET using crop coefficients:

Penman PET may be said to refer to the ET of a given standard plant under adequate water supply conditions. It is found that actual crop PET is related to PET with a correction factor usually termed "crop coefficient" KC (Jensen, 1968) thus:

$$PET_{crop} = Kc.PET$$

The value of Kc is crop specific (Gommes, 1983) hence varies over the growth cycle (Fig. 1). These values were originally defined by experimental work and details for most crops can be found in Doorenbos and Pruit (1977).

To calculate actual evapotranspiration (ET), PET_{crop} is adjusted according to soil water availability. For this purpose plant available water is considered to be the total amount of water from field capacity to permanent wilting point. Contradictory view points exist on the availability of soil moisture (Fig. 2). Using the appropriate curve in the Figure 2, actual evapotranspiration (ET) is calculated as follows.

$$ET = \frac{\text{Available water}}{\text{Available water capacity}} Z.Kc.PET$$

Where Z is the adjustment factor for different types of soil dryness curves.

b) Separate estimations of soil evaporation and transpiration

Soil evaporation: Evaporation (E) from the soil surface is separated into two stages - constant rate and the falling rate. The constant rate stage of soil evaporation depends upon the amount of energy the soil surface receives because the surface is wet and evaporation is limited by energy (Ritchie, 1972). The fraction of energy (τ) supplied to the soil surface depends on crop cover or leaf area index (LAI) and is given by

$$\tau = Rns/Rn = \exp(-.398\ LAI)$$

Where Rns and Rn is 24-hour net radiation at soil surface and above the crop canopy.

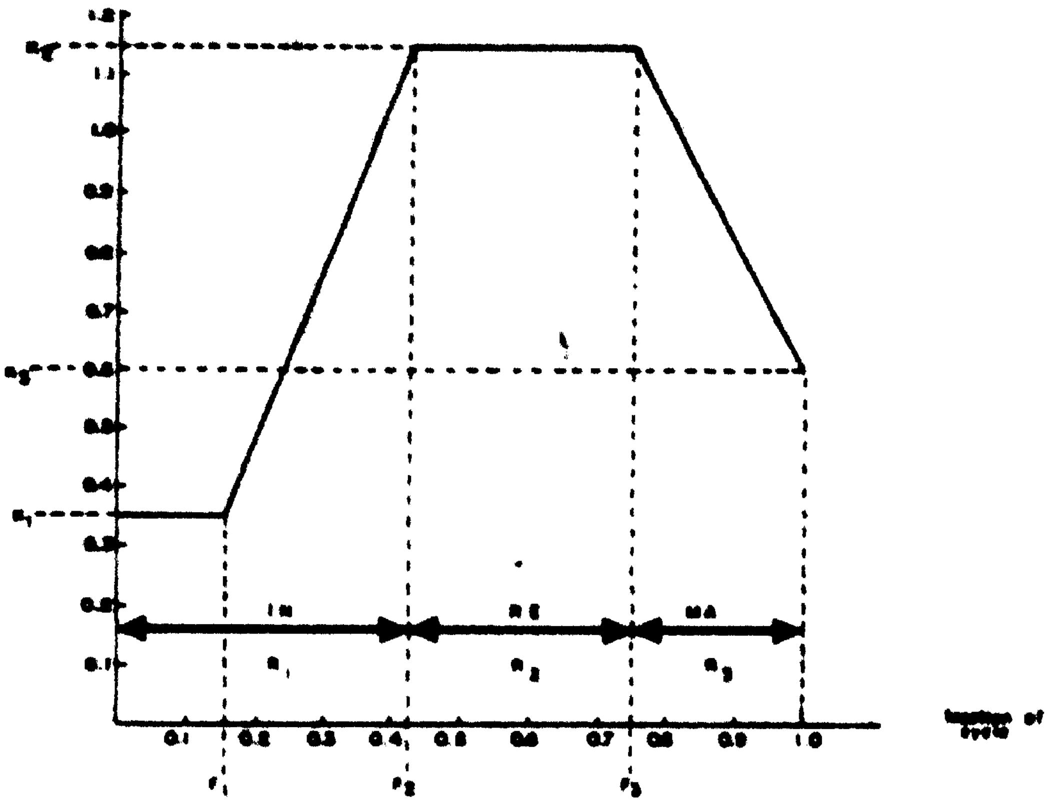

Fig. 1. Variation of the crop coefficient (K_c) over the different phenological stages: from initial (IN) to reproductive (RE) and maturing or yield formation (MA). F1, F2 and F3 are characteristic points of the cycle which, together with K1, K2 and K3, allow the interpolation of the intermediate values of K_c. The yield reduction factors are R1 through R3. In general R2 is highest (Gommes 1983).

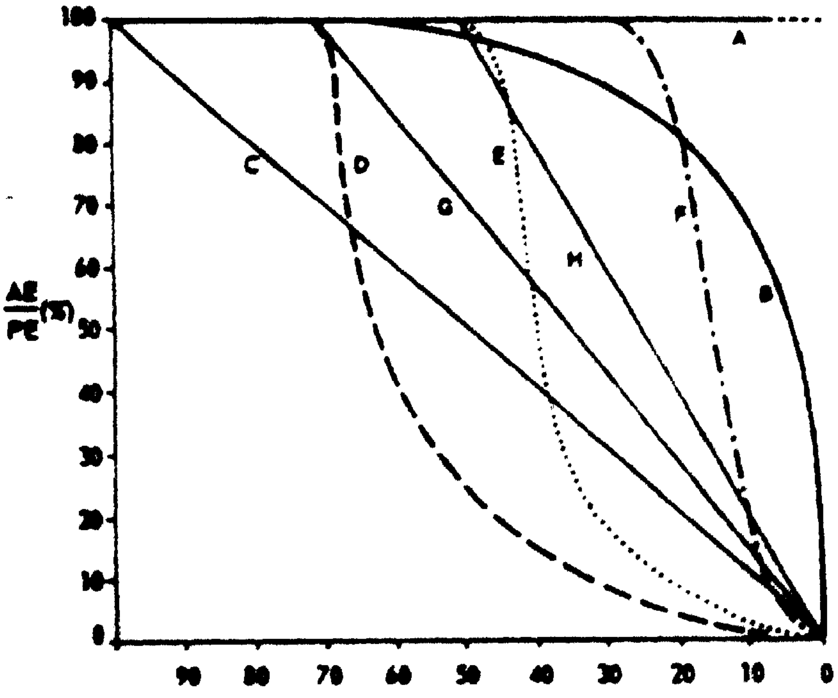

Fig. 2. Various proposals for the relationships between AE:PE ratio and available soil moisture (Baier and Robertson 1966).

Thus the soil evaporation during the constant rate stage (E1) is calculated by

$$E1 = \tau \, (PET) = mm/day \quad \ldots\ldots\ldots\ldots\ldots \quad (1)$$

During the falling rate stage, soil evaporation (E2) depends upon the water transmitting properties (c) of soil and is expressed as

$$E2 = [ct^{½} - c\,(t-1)^{½}] = mm/day \quad \ldots\ldots\ldots\ldots\ldots \quad (2)$$

where t is the days into the falling rate stage.

The program is initialized using equation 2 to compute E2 and continues to use E2 until a rain exceeds or equals 6 mm, it then switches to equation 1 to compute E1. E1 is computed each day until its sum (Σ E1) reaches a threshold value U (which depends primarily on soil texture) then it uses equation 2 again. An additional limitation is that there must be energy to evaporate the water, therefore if E2 is greater than E1, than E2 is considered equal to E1.

Potential transpiration (Tp):

Ritchie and Burnett (1971) have shown that when water movement to the plant roots is not limited, the evaporation rates from developing cotton and grain sorghum canopies is given by

$$Tp = PET\,(-0.21 + 0.70\,LAI^{½}) \text{ when } 0.1 \leq LAI \leq 2.7$$

The nonlinearity of the relation between Tp and LAI is the result of at least two interacting factors: (1) less competition for radiation per unit of leaf area during initial stages of plant growth and (2) the partitioning of a large fraction of net radiation at the dry soil surface between plant rows to sensible heat flux causing increased canopy temperature and consequently increased Tp (Ritchie and Burnett, 1971). Upper limit of 2.7 of LAI represents minimum LAI necessary for full cover of canopy. For crop canopies with LAI > 2.7, Tp = PET. When LAI $\leq$ 0.1, TP is considered negligible. Kanemasu et al. (1976) using Preistley-Taylor (1972) formula calculated Tp as follows:

$Tp = \alpha_v (1-\tau)\ [s/(s + \gamma)]\ Rn$ when crop cover < 50%

$Tp = (\alpha - \tau)[s/(s + \gamma)]\ Rn$ when crop cover > 50%

Where $\alpha_v = (\alpha - 0.5)/0.5$

Influence of soil water content on transpiration:

Ritchie (1973) reported that transpiration from sorghum or corn is not affected by soil water deficit until the available water in the root zone is less than 0.3 of the maximum available moisture content (θmax). Thus when available water content in the root zone is between 1 and 0.3 of the maximum, actual transpiration (T) is considered equal to Tp. When available water content is less than 0.3 of the maximum then

$T = TP\ \theta a/0.3\ \theta max$

Where θa is the actual available water content in the root zone.

Advective contribution:

Since Priestley-Taylor (1972) method is not good for accounting water loss from the crop by advective energy, Kanemasu et al. (1976) approximated advection (A) as

$A = 0.1T$ Tmax > 33°C for sorghum

$A = 0.25T$ Tmax > 31°C for soybean.

Then $ET = E + T + A$

Surface Runoff:

When rainfall occurs some water often runs off and becomes unavailable to the crop. Therefore to estimate the recharge of soil profile it is important to estimate amount of runoff that occurs with each rainstorm. Various approaches have been used in water balance modeling. Some models do not consider runoff and deep drainage separately. Any amount of water input into the soil after the soil profile is full to its maximum storage capacity (field capacity) is considered as water loss (runoff + deep drainage). However, in some models the runoff or infiltration is estimated to calculate profile recharge. Kanemasu et al. (1978) calculates effective precipitations (Pe) and runoff as follows:

$$Pe = R^{.75} \qquad \text{when } R > 1.0 \text{ inch}$$

$$Pe = R \qquad \text{when } R \leqslant 1.0 \text{ inch}$$

$$\text{Therefore runoff} = R - Pe$$

Baier et al. (1978) uses a simplified relationship between moisture content in the top soil zone and daily precipitation total to estimate infiltration into the soil. On days with rainfall $\leqslant$ 1.0 inch the total amount of rainfall is considered to infiltrate into the soil. On days with rainfall $>$ 1.00 inch, infiltration (Infl) into the soil is less than the daily rainfall, because it is limited by runoff as a function of rainfall and the moisture already in the top zone of soil, and is computed as:

$$Infli = 0.9177 + 1.811 \log RRi - 0.97 \,[sj(i-1)/cj]\, \log RRi$$

Where RRi = rainfall (inches) on day i

$Sj(i-1)$ = soil moisture in the Jth zone on day i-1

Cj = available water capacity of the Jth zone

$J = 1$

The remainder of the daily rainfall is assumed to be lost as runoff.

Williams et al. (1984) uses SCS curve number (USDA, Soil Conservation Service, 1972) equation to calculate runoff in their EPIC (Erosion-Productivity Impact Calculator) model which is briefly described as:

$$Q = \frac{(R-0.2S)^2}{R+0.8S}$$

Where Q is the daily runoff, R is the daily rainfall, and S is a retention parameter. The retention parameter is related to soil water content with the equation.

$$S = Smx \left(1 - \frac{SW}{SW + \exp\left[W1 - W2\ (SW)\right]}\right)$$

Where SW is the soil water content in the root zone minus the wilting point (-15 bar). Smx is the maximum value of S, W1 and W2 are shape parameters. Values for W1 and W2 are obtained from the simultaneous solution of the above equation assuming that S = Smx at wilting point and S = S3 at field capacity. At saturation S is allowed to approach its lower limit of zero.

Profile water recharge and drainage:

In the water balance models soil profile is either considered as a single layer or divided in to discrete layers of either uniform or variable thickness to represent the soil below the total rooting depth. The infiltration and redistribution of water throughout the layered soil profile is often treated in two rather different procedures.

In the simple method the infiltrated water is freely transmitted to lower layers by gravity or out of the profile if it was the lower most layer. The upper limit of water for each layer is set at field capacity. When the antecedent water content plus inflow of water exceeds field capacity of that layer then the excess water is allocated to next lower layer. This process is repeated for all layers and excess water from the lowest layer is considered as deep drainage. In this method upward movement or redistribution of water is not allowed unless it is an added feature.

The second method is to treat the layered soil profile by a solution of the Darcy's unsaturated flow equation in which each layer is assumed to be uniform in moisture content, capillary pressure, and unsaturated conductivity. Mathematical solutions vary from the simple finite difference with large time steps to finite element with near analytical results. This treatment of water flow can be used to represent nearly all situations including upward or downward flow between layers, widely varying characteristics within the profile, time distribution of infiltration and redistribution among layers, water tables, and plant water withdrawl. The choice of which soil water movement calculation to employ depends upon the accuracy required. For readily drained soils where withdrawl of water by the plant dominates the water profile development and casual accuracy is required, the free flow procedure would be adequate.

Root extraction:

Extraction of water from various layers is generally computed based upon root distribution and relative water content, which is given by

$$Tl = Tp(Sl/Cl)\ RCl$$

Where Tl = actual transpiration from layer l

Tp = potential transpiration

Sl = actual available water in layer l

Cl = available water capacity of layer l

RCl = active root concentration expressed as ability of roots to extract available soil water and is a function of time and varies from 0 to 1.

SOME EXAMPLES ON THE APPLICATION OF WATER BALANCE MODELS IN AGRICULTURE

Water balance models have been used by various workers for several purposes. Some examples of its use in rainfed agriculture are as follows:

1) Length of the growing season.

Virmani (1975) used WATBAL (Keig and McAlpine, 1974) to derive estimates of week-to-week changes in available soil moisture in relation to potential evaporative demand. The duration of the crop growing period in the Hyderabad region as determined by total available soil moisture was estimated (Table 1). The data show that on shallow Alfisols (AWC = 50 mm) the length of the growing season fluctuates between 12 and 21 weeks, on deep Vertisols (AWC = 300 mm) the duration varies from 20 to 31 weeks. Thus, soil type plays a dominant role in defining the growing period in a given rainfall situation.

Table 1. The length of the growing season in three soils having a different available water storage capacity (AWC), Virmani (1975).

Probability	Available root profile water storage capacity					
	Low (50 mm)		Medium (150 mm)		High (300 mm)	
	Weeks	Period	Weeks	Period	Weeks	Period
Mean	Jun 25 - Oct 28	17	Jun 25 - Nov 25	21	Jun 25 - Nov 25	25
90%	Jun 25 - Sep 30	12	Jun 25 - Oct 21	16	Jun 25 - Nov 18	20
75%	Jun 25 - Oct 7	13	Jun 25 - Nov 4	18	Jun 25 - Dec 2	22
Medium	Jun 25 - Oct 21	16	Jun 25 - Nov 18	20	Jun 25 - Dec 23	25
25%	Jun 25 - Nov 11	19	Jun 25 - Dec 9	23	Jun 25 - Jan 14	28
10%	Jun 25 - Nov 25	21	Jun 25 - Dec 23	25	Jun 25 - Feb 4	31

From the sowing rains up to the week when the availability of profile moisture reduces ET/PET to 0.5 (ET and PET stand for actual and potential evapotranspiration).

The shallow Alfisols exemplify a low AWC situation; deep Alfisols and medium deep Vertisols, a medium AWC situation and the deep Vertisols a high AWC situation.

2) Characterising soil water availability.

A reliable estimate of intra-seasonal probabilities of water deficits is provided by estimates of soil moisture variations occuring over the growing season. The medium amounts of available water present in the root profile of three soil types of low, medium and high "available" water capacity (AWC) are shown in Figure 3. The amount of available moisture in the low AWC soils does not exceed 60-70 per cent of the total available water capacity; there is a marked decrease in the amount of available water in the first half of August (to less than 25 mm). Since the evaporation demand during dry periods in the rainy season often exceeds 25 mm per week, a break in the continuity of rains exceeding one week would be quite hazardous to crops on the low AWC soils. These results emphasize that to increase and

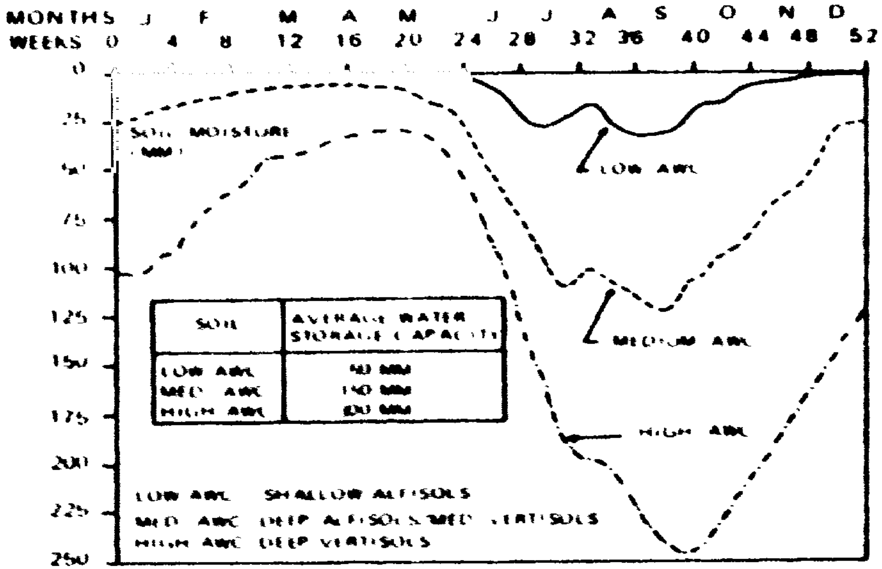

Fig. 3. Weekly soil moisture storage in three soils (Hyderabad 1901 - 1970 data) (Virmani 1975).

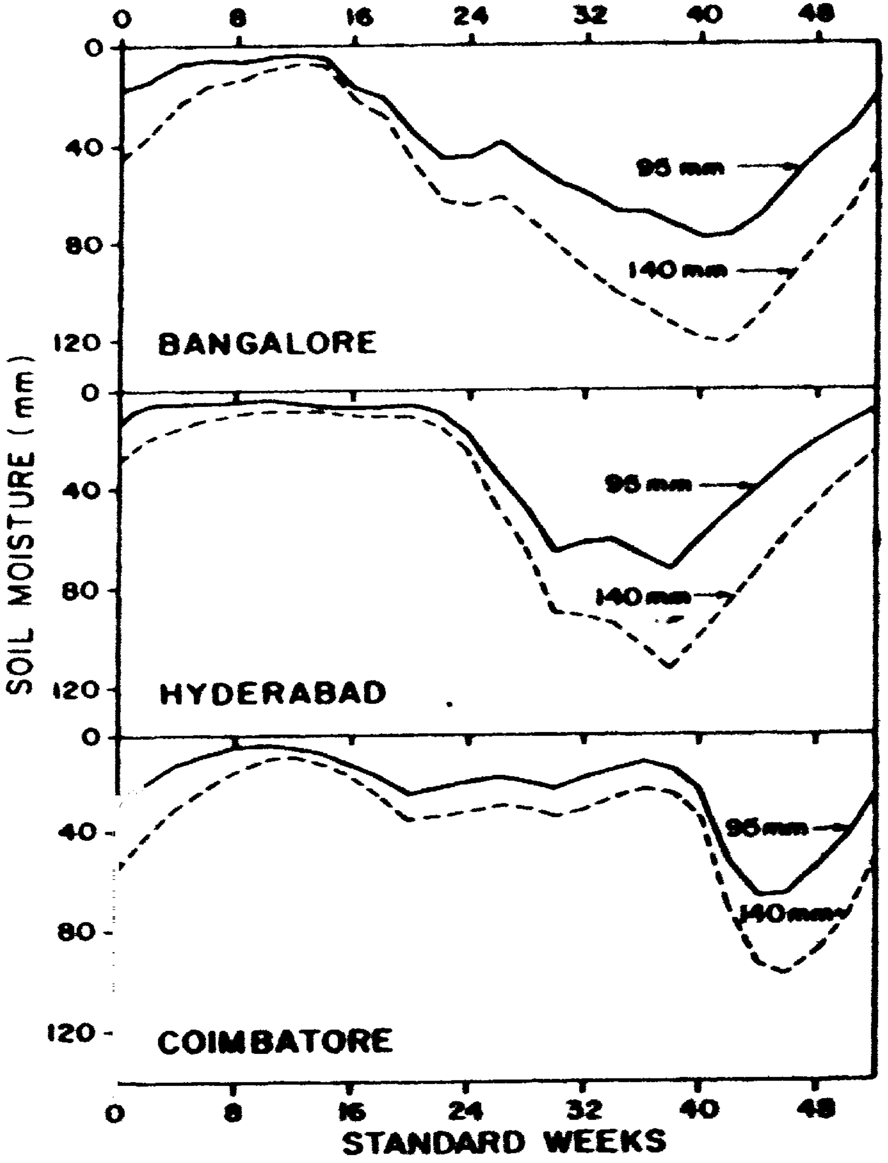

Fig. 4. Simulated profile moisture changes under two assumed soil water storage capacities for three locations in India.

stabilize crop production in such soil regions there is a need for developing alternative water resources to break intra-seasonal droughts. One alternative is to collect runoff during periods of excess rainfall and reuse the collected water through a farm water storage and application facility.

Sivakumar et al. (1983) compared the soil water availability using WATBAL in the Bangalore, Hyderabad and Coimbatore region (Fig. 4). Data shows that Bangalore has longer soil water availability period (about 36 weeks) than the Hyderabad region (about 28 weeks). At Coimbatore the soil water availability period is as long as that at Bangalore but the amount of water available is much less especially in the initial part of the season Such a variations in soil moisture availability has strong implications for the type of crops and cropping systems to be grown. Coimbatore region can support only drought hardy and short duration crops while Hyderabad and Bangalore provide more favourable moisture environment for crops.

Soil water balance models could also be used to characterise soil water availability at the beginning of the postrainy season i.e., after the harvest of rainy season crop. Huda and Virmani (1986) compared the soil water availability at the beginning of postrainy season for Indore, Dharwar, Patancheru and Anantapur after the harvest of a rainy season sorghum (Fig. 5). Data shows that at Patancheru and Indore about 100 mm of water is available in 8 out of 10 years, while at Dharwar it is only 50 mm at same probability. Anantapur will not have any available water at the beginning of the postrainy season. These data show that a

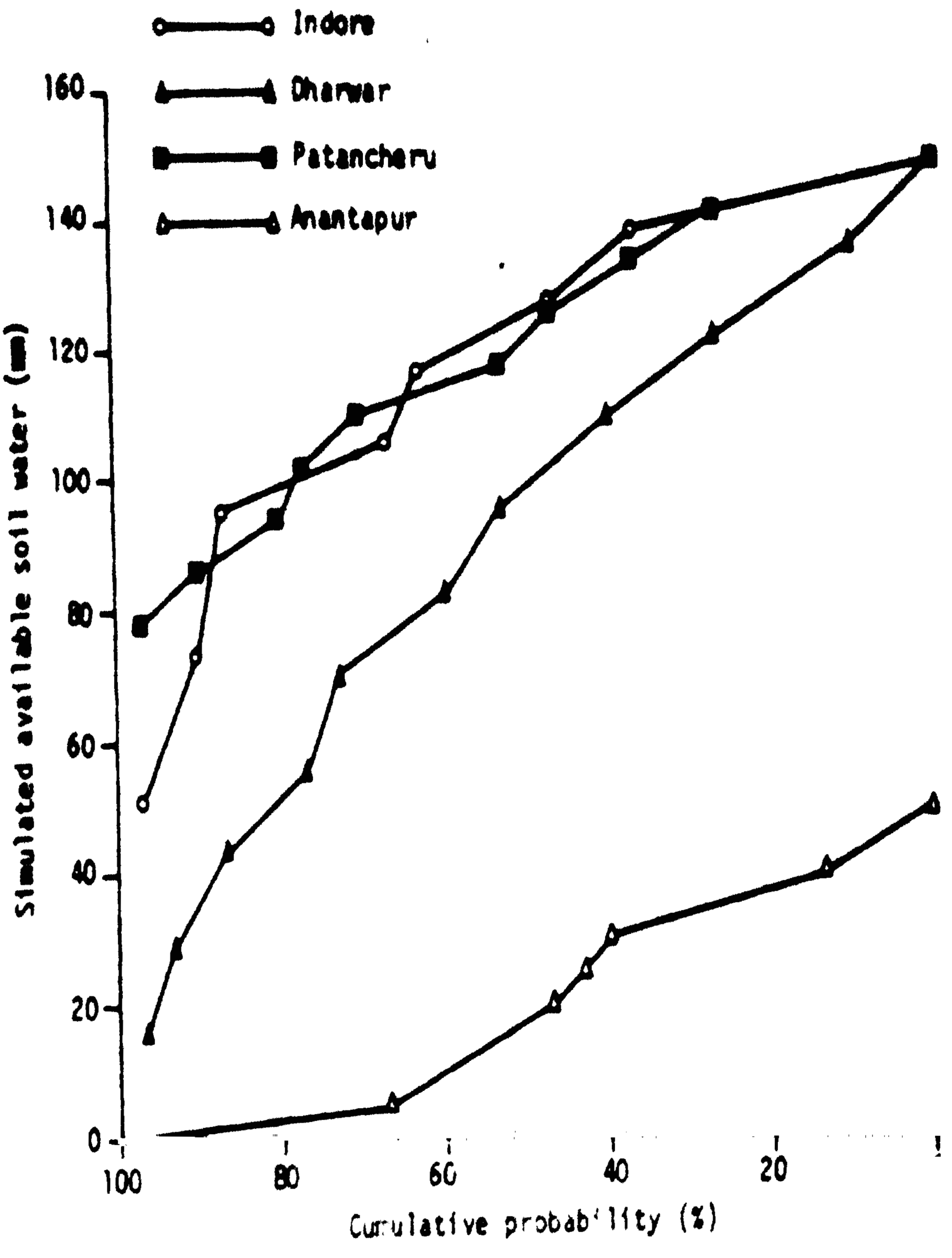

Fig. 5. Cumulative probability (%) of simulated available soil water at the beginning of the postrainy season after the harvest of rainy season sorghum for four selected locations. (Data base: 1941 to 1970) (Huda and Virmani 1986).

postrainy season crop could be successfully grown at Patancheru and Indore without any significant supplemental irrigation, while crops at Dharwar will need supplemental irrigation in most years. Crops at Anantapur will need full irrigation if the crops are to be grown.

3) Moisture availability in relation to crop demand.

It is generally accepted that evapotranspiration requirements of most crops expressed as a fraction of the potential evaporative demand is about 0.4 at the seedling stage, about 0.6 at the vegetative stage, 0.8 at the flowering and the reproductive stage, and about 0.4 at maturation and seed filling stage. Virmani (1975) used WATBAL method to estimate weekly changes in the available moisture in the root profile for 1901-70 data. The probability of meeting the estimated evapotranspiration from a given sowing date (last week of June for Vertisols and first week of July for Alfisols) were computed for a 65-70 day pearl millet and two sorghums of different growth durations (Table 2). The data show that deep Vertisols are able to meet ET in most years for millet, as well as medium and long duration sorghum. Deep Alfisols and medium deep Vertisol are comparatively better suited for millet and 90-100 day sorghum than for the long duration sorghum. The shallow Alfisols are suited to short duration crops like millet. The incidence of drought at various stages of crop growth increases with medium and long duration sorghum.

Table 2. Probabilities of moisture availability to meet crop demand in three soil types at specified growth stages of crops of varying durations (based on 1901-1970 data for Hyderabad), Virmani (1975).

Growth stage	Period	ET/PET required	Shallow Alfisols	Deep Alfisols Medium Vertisols	Deep Vertisols
		Sorghum (130-150 day)			
Seedling	4 week	0.30	0.41	0.66	0.70
Vegetative	6 week	0.60	0.38	0.74	0.73
Reproductive	6 week	0.80	0.17	0.65	0.80
Maturity	4 week	0.60	0.06	0.48	0.67
		Sorghum (90-100			
Seedling	2 week	0.30	0.59	0.77	0.79
Vegetative	4 week	0.60	0.49	0.75	0.72
Reproductive	4 week	0.80	0.42	0.72	0.66
Maturity	4 week	0.60	0.41	0.86	0.87
		Pearl Millet (65-			
Seedling	2 week	0.30	0.59	0.77	0.79
Vegetative	2 week	0.60	0.75	0.86	0.88
Reproductive	4 week	0.80	0.47	0.77	0.79
Maturity	2 week	0.60	0.72	0.72	0.98

4. Selection of suitable crops and cropping patterns.

Using water balance procedure, Virmani (1975) estimated the water availability (ratio of actual to potential evapotranspiration) in different soil types and compared with the water demand of crops to assess the suitability of crops in a given soil-climate system. An example of fitting of long (130-150 day), and medium (100-110 day) and short (65-70 day) duration crops has been shown in Figures 6 and 7. It is clear from the data that traditional varieties (130-150 days) will flower in late September and reach physiological maturity in late October or early November. The crop will be caught in a water deficit situation at the reproductive stage in most years on low AWC soils. The medium

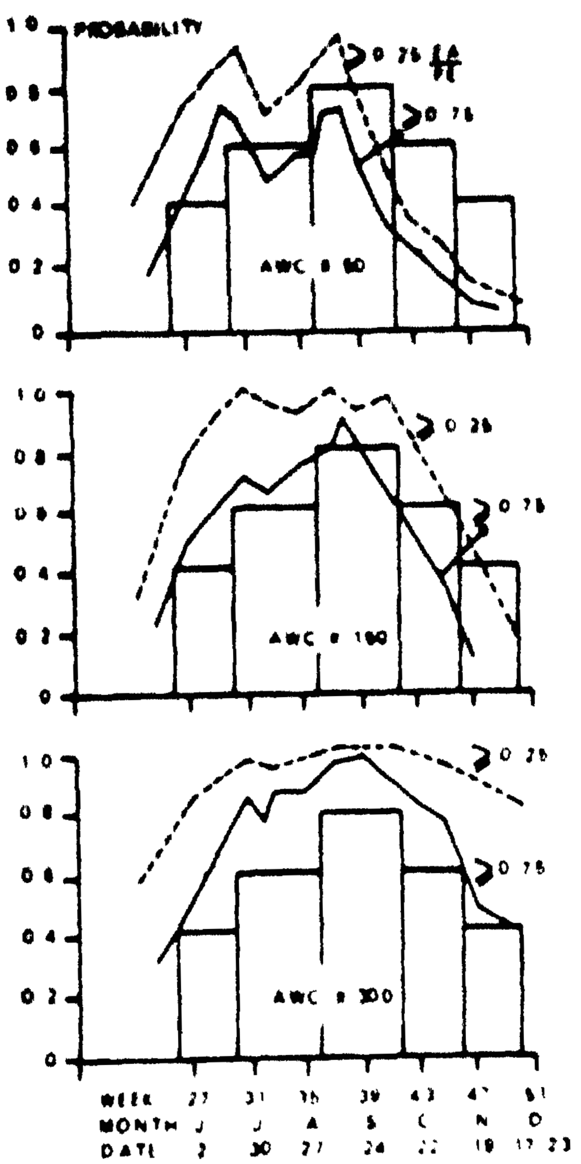

FIG. 6. FITTING OF A LONG DURATION (130-150) CROP IN THREE SOILS

(BARS = CROP WATER REQUIREMENT AND CURVES = WATER AVAILABILITY AT TWO PROBABILITY LEVELS)

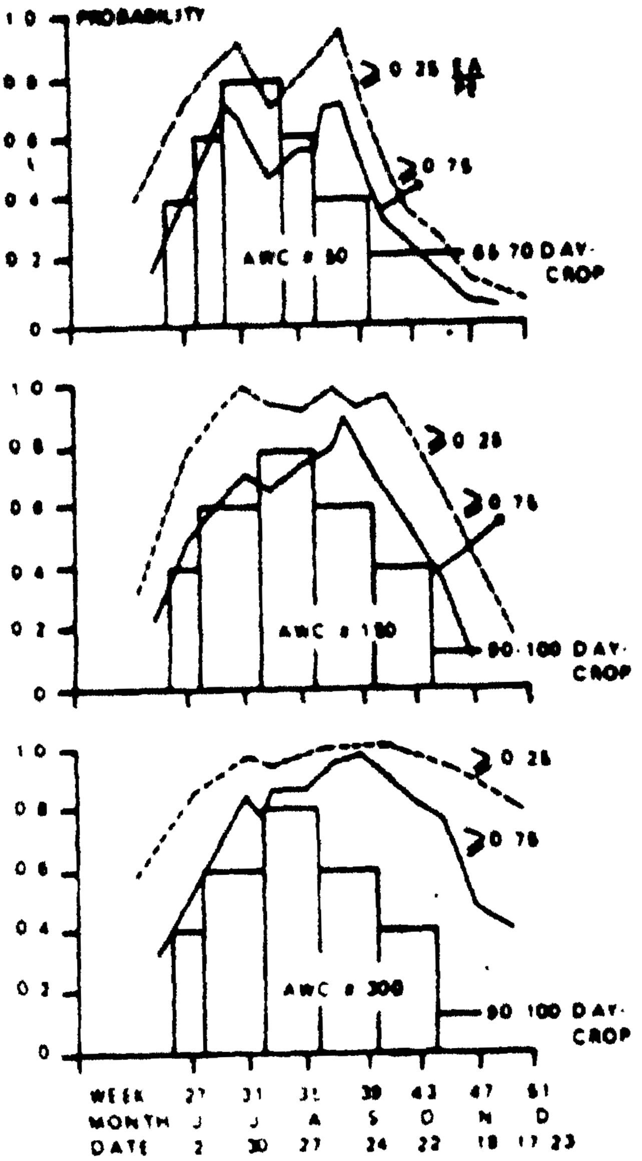

FIG. 7. FITTING OF SHORT (65-70) & MEDIUM (90-100) DURATION CROPS IN THREE SOILS

(BARS = CROP WATER REQUIREMENT AND CURVES = WATER AVAILABILITY AT TWO PROBABILITY LEVELS) and shorter duration varieties (90 - 110 days) will flower in the last half of August and reach physiological maturity in late September. Such varieties may frequently be subject to moisture stress due to drought in August. Such a analysis on water balance of a location in conjuction with rainfall analysis helps fit a variety or cropping system in an environment.

5. Characterising runoff potentials.

In rainfed farming areas it is important to know the surface runoff potentials or excess water losses from the agricultural land in order to develop appropriate land and water management practices to conserve soil and water and to develop water harvesting and storing facilities. Water balance models can generate information on amount and frequency of water loss from various soil types and locations. Such a study has been done for the Hyderabad region for soils varying in water holding capacity from 50 mm to 230 mm (Table 3). It is clear from the data that probability of water loss from low water holding capacity soils (AWC 50 mm) is almost double of that from soils having greater available water holding capacity (AWC = 230 mm). At least 10 of rainfall is lost in 96% of years from the low AWC soils (AWC = 50 mm) compared to 50% of years from the high AWC soils (AWC = 230 mm). At least 100 mm of rainfall is lost in 24% of years from the 230 mm AWC soils, whereas similar losses occur in 59% of years from the 50-mm AWC soils. Such an information has strong implications for developing land and water management practices for different agroclimatic environments.

Table 3. Probabilities of water loss (runoff + drainage) in soils of different available water capacities in the Hyderabad region.

Available water capacity (mm)	Water loss (mm)						
	> 10	> 25	> 50	> 75	> 100	> 150	> 200
	Probability (%)						
230	51	44	37	29	24	17	14
200	63	51	44	36	29	17	16
160	69	64	51	46	37	24	17
140	77	69	59	47	40	26	19
95	87	83	71	60	47	34	23
50	96	91	80	74	59	44	27

Huda and Virmani (1986) estimated the probability of water loss from soils of Indore, Dharwar, Patancheru and Anantapur (Fig. 8). Data shows that soils of Indore have greater potential for water loss (about 200 mm at 80% probability) in most years compared to Patancheru, Anantapur and Dharwar (30 to 75 mm at 80% probability). In 2 out of 10 years Indore could have 500 mm of rain water lost from the soil profile. Such a data indicate stronger need for developing soil and water conservation practices for the Indore region compared to other three locations.

6. Use of water balance models in yield predictions.

Most studies have shown that yield and water use by crops are closely related (de Wit, 1958 and Hanks, 1974). In general, studies indicate that dry matter production and ET are more closely related than grain yield and ET. Improvements in these correlations are observed when transpiration (T) is used instead of ET and when the effects of water stress at critical growth stages are incorporated in the models relating ET with crop

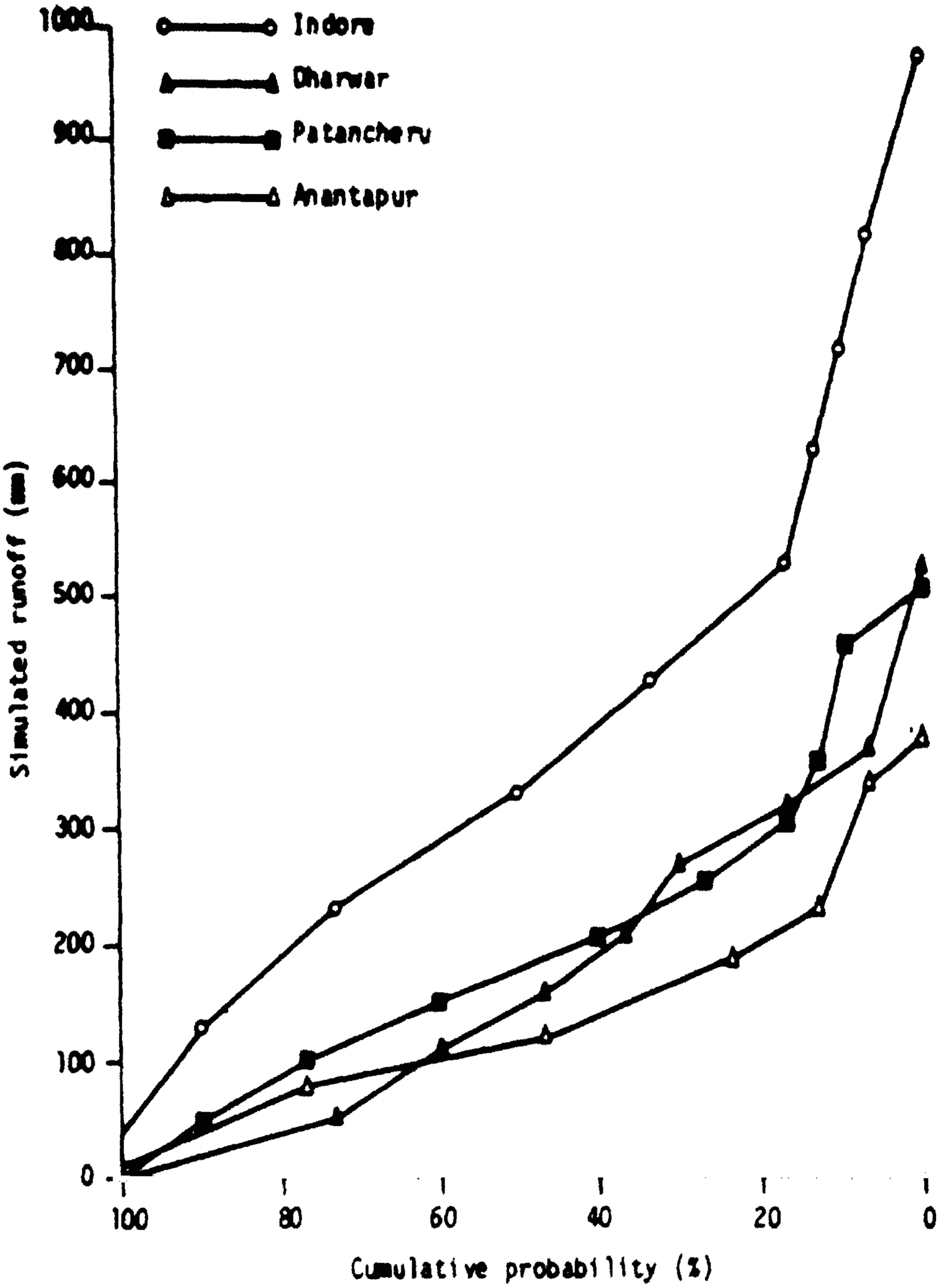

Fig. 8. Cumulative probability (%) of simulated runoff for four selected locations. (Data base: 1941 to 1970) (Huda and Virmani 1986).

yield. Water balance models could be used to estimate transpiration and evapotranspiration to predict yields through the ET-based yield models. Stewart et al. (1977) reported the following relationship for dry matter:

$$Y/Ym = 1-\beta o\ ETd = 1 - \beta o + \beta o\ ET/ETm$$

Where βo is the slope of the relative yield (Y/Ym) versus ETd ; and ETm is defined as the ET required for Ym; ETd is given by 1 - ET/ETm. The value of βo is obtained from field measurements.

In some models the effect of growth stage on yield response to water stress have been included. Rasmussen (1979) developed the following grain yield water use equation for winter wheat:

$$Y(kg/ha) = 1.92 \left(\Sigma (T/ETo)\right)_1^{0.172} \left(\Sigma (T/ETo)\right)_2^{0.104} \left(\Sigma (T/ETo)\right)_3^{0.646}$$

Where subscripts 1, 2 and 3 refer to growth stages emergence to jointing, jointing to heading, and heading to soft dough, respectively. Daily T/ETo ratios are summed during each of the stages. ETo is estimated by Priestley-Taylor (1972) equation. Rasmussen (1979) reported a correlation coefficient (r) of 0.68 between predicted and observed yields. Better results were obtained with a data subset in which water was the primary yield-limiting factor ($r^2 = 0.91$).

We at ICRISAT have also studied the relationship of sorghum grain yield with evapotranspiration. The relationship is:

$$Y/Ym = (ET/ETM)_1^{0.19} \ .\ (ET/ETM)_2^{0.55} \ .\ (ET/ETM)_3^{0.33}$$

Where ET and ETM refer to actual and maximum evapotranspiration at a given growth stage. Subscripts 1, 2 and 3 are growth stages emergence to panicle initiation, panicle initiation to 50% flowering, and 50% flowering to physiological maturity, respectively. This model accounts for 84% variation in yield ratio (Y/Ym).

Water balance models are also integral part of the more elaborate process based dynamic models such as the ones developed for sorghum, wheat, soybean, corn etc. Water balance models have been included in these crop models to develop water stress coefficient based upon soil water prediction in the root zone.

From the foregoing it is clear that there are several approaches in modeling soil water balance which depend upon the intended use of the water balance model and the accuracy required in predicting various components of water balance. Water balance models have numerous application and only a few have been discussed here in context of rainfed agriculture and many more could be envisaged or found in the literature on soil water balance.

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