EVALUATION OF RECHARGE THROUGH SOILS IN A MOUNTAIN REGION: A CASE STUDY ON THE EMPIRE AND THE SONOITA BASINS

by

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ABSTRACT

A conventional water balance method, employing long-term average values of rainfall, runoff and evapotranspiration yields near-zero recharge values for the Empire and the Sonoita basins. These results, however, are not in agreement with those obtained from an analysis of the local ground water regimes. A different approach for calculating recharge, based on the typical characteristics of these arid basins, is proposed. In particular, both basins are characterized by intense thunderstorms of short duration in the summer which occur usually towards the evening, and shallow, sandy-gravelly soils with a relatively high permeability overlying fractured rocks in the elevated mountain regions. These factors may cause a considerable amount of water to infiltrate through the soil profile, thereby escaping evapotranspiration during the following day. The proposed model deals with separate thunderstorm events using mean values of rainfall intensity and frequency corresponding to elevation. This model was coupled with a numerical solution of the flow equation which was used to solve the one dimensional water flow through a soil profile. The solution includes sink terms and was solved for the simultaneous processes of infiltration, moisture redistribution and evapotranspiration. The results obtained show almost no recharge in the low valleys, but significant recharge in the mountains. The amount of recharge increases with elevation and decreases with the depth of the soil profile.

INTRODUCTION

The present study is a part of a more general study on ground water evaluation and influence of future exploitation on ground water and surface water regime in the Cienega and the Sonoita Basins.

Most of the ground water recharge in the studied area seems to occur in the elevated mountain regions. These regions consist mainly of hard rock formations, partly covered by shallow soils. The exposed formations have a recharge capability of their own which is currently under study. The present study deals with the recharge contribution through the soil covered portion of the mountain regions.

The conventional approach of calculating recharge by subtracting long term averages of runoff and evapotranspiration from the total rainfall results in apparently no recharge. This result does not agree with observations of ground water flows in the corresponding basins.

In order to overcome this discrepancy, the interaction between the constant environmental parameters, namely the soils and rock formations, and the time dependent variables, namely, rainfall and evapotranspiration, is investigated on a short time scale, rather than on the conventional long time scale. More specifically, the objective of the present study is to evaluate the recharge through a soil profile during a single summer thunderstorm event.

BACKGROUND

GEOLOGY

The geology of the studied area was described in detail, among others by Creasey (1967), Drewes (1971, 1972a, 1972b), Hayes and Ramp (1968), Rohrbacher (1964) and Simon (1972, 1974).

In general the area is part of the "Basin and Range" province. The geologic column consists of Precambrian igneous and metamorphic rocks, unconformably overlain by Paleozoic shallow marine carbonates, sandstones and shales. Above these are Mesozoic shallow marine sandstones, shales and carbonates associated with volcanic rocks. The young Cenozoic tectonics starts to differentiate the area into elevated mountains and intermountain basins, which are continuously filled with clastic debris of the rising mountains. The young alluvial formations are thus confined to the "basins". A general schematic section across the mountain and the valley regions is given in Figure 1.

The hydraulic properties of the aquiferous formations, exposed in the mountain regions are in general as follows: the hard carbonates, mainly the Paleozoic limestones and dolomites show an emphasized joint and fractured pattern accompanied by solution channels partly filled with secondary calcite, which suggests a considerable water flow through them. Similar formations elsewhere have hydraulic conductivities (K) within the range of 10-1000 m/day. The igneous formations are partly fractured and weathered. Ground water flow seems to be confined to the fractured zones. The clastic formations, in analogy to other regions, have hydraulic conductivities within the range of 1 m/day for the medium to 3000 m/day for the coarse grained clastics. The K values of coarse gravels can amount to 10,000 m/day.

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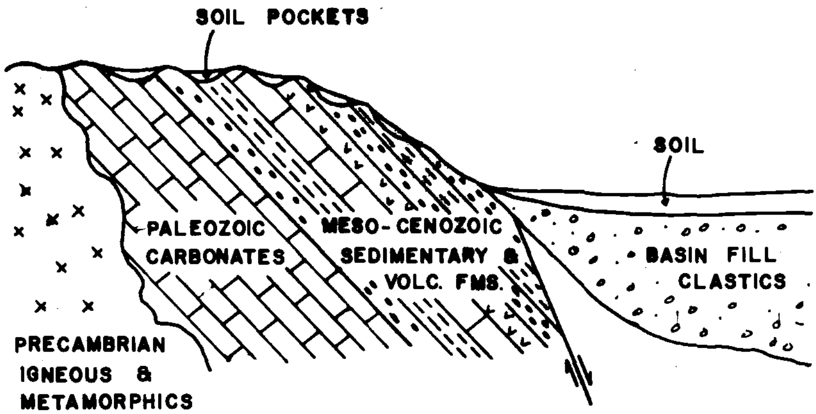

Figure 1. A schematic geological cross-section.

The described formations are partly exposed and partly overlain by a relatively shallow soil cover.

SOIL COVER

The type and distribution of the soils in the study area are discussed in detail by Richardson (1971) and Richardson and Miller (1974). In general, the various soil types can be grouped into two main groups: a) The low valley and lowland deep soils, namely the Sonoita - Bernardino - White House - Hathaway - Canelo - Caralampi Associations, and b) the mountain shallow soils, namely the Faraway - Barkerville - Tortugas - Lampshire Associations.

The schematic section (Fig. 1) across the mountain and the valley region illustrates the distribution and the thickness changes of the soil cover.

In the valleys and lowlands the soils are continuous, and relatively thick, over 150 cm. In the mountain regions the soils are 10-50 cm thick and they appear in a rather patchy distribution according to the local relief, forming lenticular pockets and covering some 50% of the surface area.

Permeability according to Richardson (1971) and Richardson and Miller (1974) is low, between 0.06 and 0.2 inch/hour for the deep valley and lowland associations. It is moderately high, 2.0 - 20 inch/hour, for the shallow mountain associations.

RAINFALL

Rainfall variation and distribution in the area are discussed in detail in Sellers and Hill (1974). The annual rainfall variation shows two distinct rainy seasons which differ from each other in origin and intensity. The winter rains in southeast Arizona are responsible for some 35% of the annual rainfall and are mainly originated by Pacific fronts. The summer rains, which are related to moisture penetration from the Gulf of Mexico, amount to some 65% of the annual rainfall. The later are orographic, convection type thunderstorms of high intensity and short duration, starting in the late afternoon and early evening.

Duckstein et al. (1973) show for the summer rains in this region that there exists a linear relationship between elevation, and both mean number of thunderstorm events per season and mean amount of rainfall per event. Using the equations derived by them the results for the relevant elevations in the studied area are summarized in Table 1. It is evident from this table that variations of rainfall per event are very small within the given elevations, and the number of events accounts mainly for the total summer rainfall differences for the various elevations.

Table 1. Number of events, rainfall per event and total summer rainfall vs. elevation.

ELEVATION h	NUMBER OF EVENTS E(N)	RAINFALL PER EVENT E(R)		TOTAL SUMMER RAINFALL E(Z) = E(N)·E(R)	
feet		inches	mm	inches	mm
10,000	43.6	.30	7.7	13.22	335.8
9,000	40.5	.29	7.4	10.79	274.1
8,000	37.4	.28	7.1	9.83	249.7
7,000	34.2	.26	6.8	8.87	225.3
6,000	31.1	.25	6.5	7.91	200.9
5,000	28.0	.24	6.2	6.95	176.5

THE MODEL

OBJECTIVE

The purpose of the model, as stated before, is to simulate the particular situation which prevails in the studied area. It combines the existance of shallow and permeable soils, overlying fractured and permeable rocks, intense and short rainfall events and a time-span devoid of evapotranspiration. The model analyzes a single typical thunderstorm event assuming that the amount of water which drains below the soil profile and the root zone escapes evapotranspiration and may, thus, be considered as net recharge to the ground water system.

DESCRIPTION

The basic element represented by the model is a single soil pocket, a series of which overlie the bedrock relief in the mountain region. The geometry of such an element, 10 to 50 cm deep, is schematically illustrated in Fig. 2. The different pockets in the area are assumed to be completely separated in the sense that there is no flow from one element to another. The soil in the element is assumed to be a stable, isotropic and homogeneous porous medium. The initial water content is taken to be uniform all over the element, sufficiently low to prevent any considerable initial water flow in the system. The hydraulic properties of the soils in the study area are given in Figs. 3 and 4.

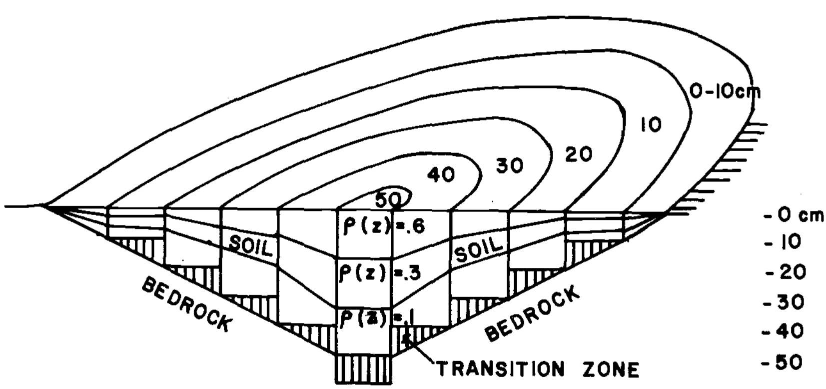

Figure 2. A schematic geometry of a single soil element.

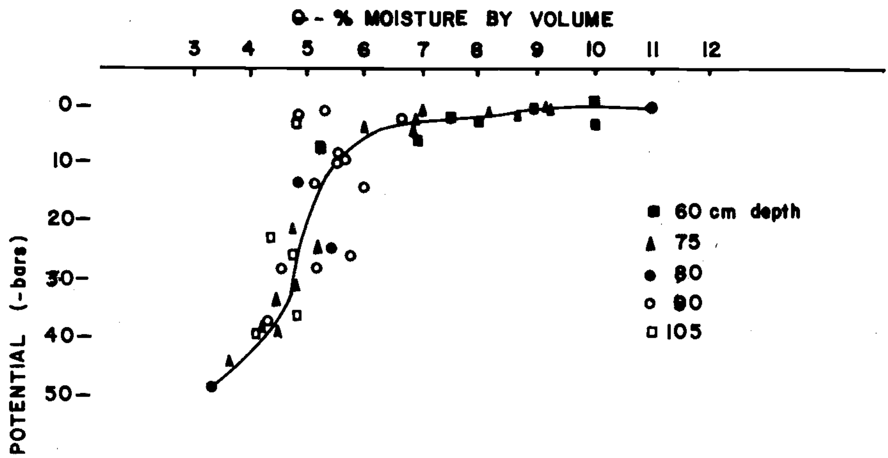

Figure 3a. Moisture release curve for the soil (after Sammis, 1974).

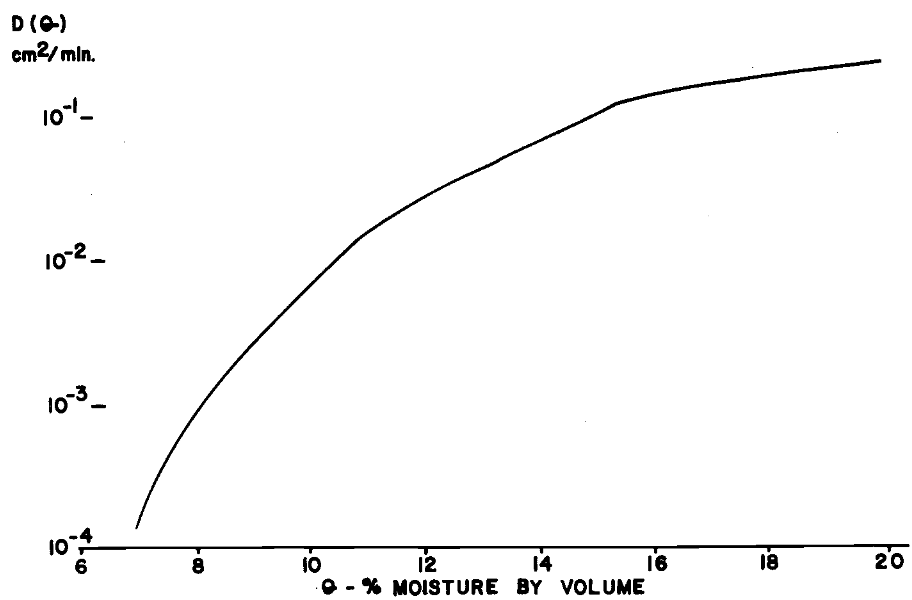

Figure 3b. Diffusivity as a function of moisture content.

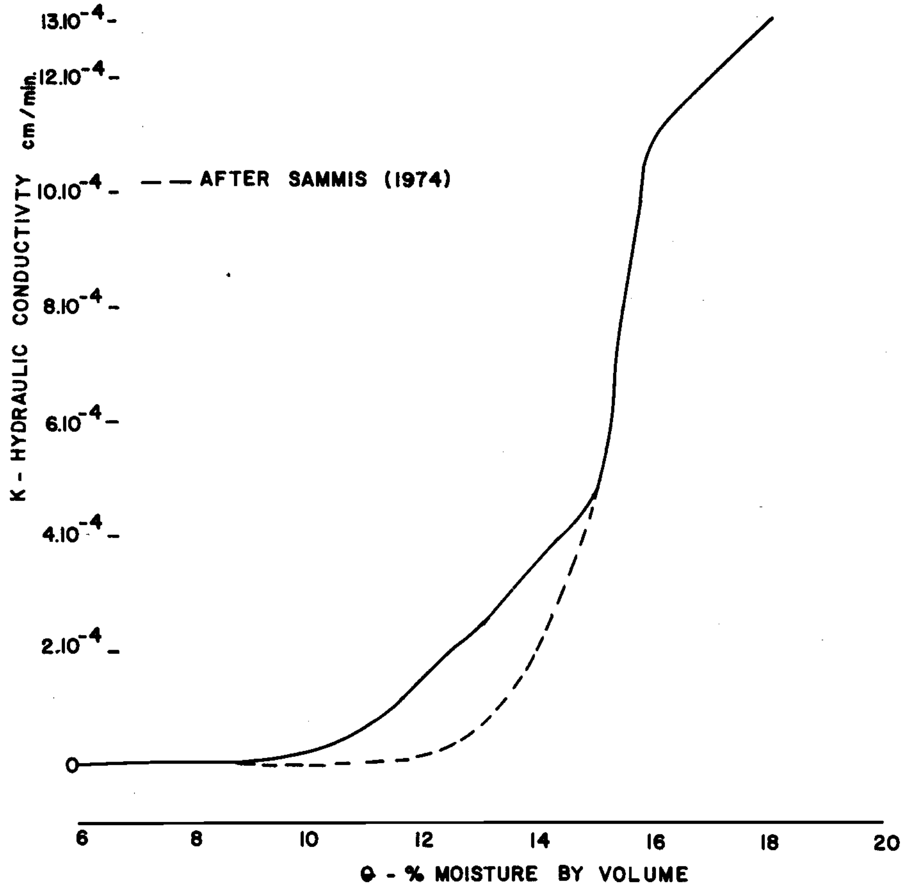

Figure 4. Hydraulic conductivity as a function of soil moisture.

The differential equation that governs the one dimensional vertical flow in the system can be expressed in terms of the diffusivity form as follows:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(D(\theta)\frac{\partial \theta}{\partial z}\right) - \frac{\partial K(\theta)}{\partial z} - S \qquad (1)$$

where θ is the soil water content (cm^3/cm^3), t is a given time (minutes), $D(\theta)$ being the soil water diffusivity (cm^2/min.), $K(\theta)$ being the soil hydraulic conductivity (cm/min.), z is the vertical coordinate, regarded positive downwards and S is a sink function (t^{-1}). Since hysteresis is not taken into account in the present study, both $D(\theta)$ and $K(\theta)$ are assumed to be a function of θ.

The above equation was solved by Diner (1970), excluding the sink term for the following initial and boundary conditions:

$$\theta_{(z,0)} = \psi_{(z)} \quad ; \; 0 \leq z \leq Z$$

LOWER BOUNDARY CONDITIONS

Since the model assumes a homogeneous and isotropic medium its lower boundary, namely, the soil-bedrock interface is expressed in two different ways, representing two different boundary conditions. The two cases, both found in the studied area are:

a) $K_r \leq K_s$ and b) $K_r \geq K_s$, where K_r and K_s are the hydraulic conductivities of the bedrock and the soil, respectively.

In the first case ($K_r \leq K_s$) the boundary conditions at Z represent a surface under which the vertical drainage is controlled only by gravity. The two phase hydraulic conductivity function acts as two different layers. The first phase represents the lower hydraulic conductivity of the bedrock which prevents water drainage from the overlying soil profile until the moisture content in the lowermost soil profile reaches the breakthrough point and a considerable flow through the boundary is initiated at time t_1. The vertical drainage is, therefore, negligible at z=Z for $0 \leq t \leq t_1$. Being controlled only by gravity, and following the breakthrough time t_1 to the end of the process at time T, the boundary conditions at Z are given by:

$$\frac{\partial\theta}{\partial z}(Z,t) = 0 \ ; \ t_1 \leq t \leq T$$

To simulate the first boundary condition where $K_s \geq K_r$ the model was applied separately for each particular soil thickness examined, i.e., 10, 15 and 20 cm, and the vertical drainage was obtained at the bottom.

In the second case it is assumed that the soil-bedrock interface does not limit the downward flow, the bedrock and the overlying weathered zone which is transitional to the soil, are having the same hydraulic characteristics as those of the soil. The interface represents herein only a lithological transition and the lower boundary of the root zone. The driving forces of the vertical flow at the lower boundary are controlled by moisture content gradient and the gravity. This case was simulated by defining the boundary conditions at z=Z, deep enough to avoid any considerable effect on the region of interest. The model was applied to a depth of 50 cm and the flow was calculated for the relevant 10, 15 and 20 cm, depths. The flow was considered as the volumetric net change of soil moisture in the profile after the deduction of evapotranspiration for the corresponding time interval.

RAINFALL INTENSITY

The downward flow velocity at the surface (z=0) is given by the rainfall instantaneous intensity expressed as:

$$V_{(z=0)} = -D(\theta)\frac{\partial\theta}{\partial z} + K(\theta) + E(\theta) \text{ at } z=0. \qquad (2)$$

where $E(\theta)$ is the surface evaporation function. This function can be neglected since it was shown (Diner, 1970) that the effect of $E(\theta)$ compared to transpiration is negligible. The variations in rainfall per event for the different elevations (Table 1) are small and they are insignificant when expressed as rainfall intensity. An average rainfall was chosen and intensities were calculated for the durations of 30, 60, and 120 minutes being 0.0266, 0.0113 and 0.0056 (cm/min) respectively. Although the first duration is most frequent (Sellers and Hill, 1974) the two others were also examined.

RUNOFF

Runoff is not taken into account in the model assuming that there is no lateral flow from one soil element to another. The infiltration capability of the soil in the study area (0.2 cm/min.) is higher than the given rainfall intensity (0.0226 cm/min.) and therefore it is assumed that runoff is not initiated from the soil elements which is in agreement with the relative low runoff/rainfall ratio in the study area. Any lateral runoff from adjacent bedrock exposures to the bottom of the soil element is additional to the rainfall which falls directly on the element and might therefore increase the calculated recharge.

EVAPOTRANSPIRATION

The rate of water uptake by roots from a volume unit of soil per time unit, $S(z,\theta,t)$ (cm^3 water/cm^3 soil per min) is a function of root density at each depth $\rho(z)$ (cm^3 roots/cm^3 soil), the hydraulic characteristics of the soil $f(\theta)$ (dimensionless) and evaporation capability of the atmosphere throughout daytime, expressed as a time dependent function $\alpha(t)$ having the same units of S(min^{-1}). For the present study it was assumed that

$$S(z,\theta,t) = f(\theta)\cdot\rho(z)\cdot\alpha(t) \qquad (3)$$

The integration of $S(z,\theta,t)$ over the entire depth results in the actual evapotranspiration given as:

$$Et_{(t)} = \int_0^Z \alpha(t)\cdot f(\theta)\cdot\rho(z)\,dz \qquad (4)$$

The root density function assumed that the water uptake is proportional to the root density in a given depth. According to Gardner (1964) and to field observations the root density was assumed for different types of vegetation to be some 60, 30 and 10% for the upper, middle and lower thirds of the soil profile. The bedrock is considered to be devoid of roots.

$f(\theta)$ is assumed to be a step function of the form:

$$f(\theta) \begin{cases} =1 & \theta^* \leq \theta \leq \theta_{max} \\ =0 & \theta^* \geq \theta \end{cases} \qquad (5)$$

where θ^* is the volumetric water content at the wilting point.

The transpiration is a function of atmospheric conditions as long as θ is above the critical θ^* value. In this case and for these soils θ^* was found to be 0.06 (Sammis, 1974). The step function seems to be realistic being also in agreement with the water retention curve obtained by Sammis (1974) and Mehuys et al. (1975) which approaches a step function (Fig. 3a). The sink intensity, $\alpha(t)$, is time dependent. It is considered zero during the thunderstorm when relative humidity exceeds 100% and during the following night. During the successive 12 hours of day time $\alpha(t)$ was determined from equation (4) replacing the instantaneous actual, by the potential evapotranspiration.

The total typical summer day potential evapotranspiration was chosen to be 6mm/day. The actual evapotranspiration obtained from the material balance of the model was 4.5 mm/day.

The model was applied for 1440 minutes the first half simulating the night hours and the second half of the day time when evapotranspiration is effective.

FIELD MEASUREMENTS

In order to compare the model's soil moisture content with field data, soil samples were taken in the study area from various depths and their moisture content was determined. The sampling was carried out 36 hours following a rainy winter evening of some 10 mm rainfall and a winter day of low evapotranspiration. The corresponding higher rainfall and lower evapotranspiration, about one fourth of the summer amount, were imposed on the model. The moisture content of both field samples and model results was compared.

RESULTS

DRAINAGE THROUGH A SOIL PROFILE, $K_s \geq K_r$

The results of this case where the soil hydraulic conductivity is higher than that of the bedrock show that there was no drainage from the 20 cm thick soil profile. The results obtained from the 10 and 15 cm thicknesses are represented as drainage rate curves accompanied by an evapotranspiration histogram in Fig. 5. The cumulative drainage in mm for the different thicknesses and rainfall intensities is given in Table 2. The results can be summarized as follows:

(a) Regarding the 10 cm profile most of the drainage occurs during the night and the early morning when evapotranspiration is still small.

(b) Drainage starts in the early evening, earlier for the two higher rainfall intensities. After a while the drainage rate becomes approximately the same for the different intensities.

(c) During the day the drainage rate decreases sharply due to increase of evapotranspiration.

(d) The cumulative drainage amounts to 2.10 mm for the most frequent half hour rainfall duration, with a somewhat higher value for the hour duration and somewhat lower for the two hours duration.

(e) Regarding the 15 cm profile, there is a longer time lag to the start of drainage which occurs only in the morning. Most of the water in the soil profile does not escape evapotranspiration and drainage accordingly is drastically reduced to relatively small amounts which increase with rainfall intensity.

DRAINAGE THROUGH A SOIL PROFILE $K_r \geq K_s$

This represents the case of higher bedrock hydraulic conductivities than those of the soil. A cumulative drainage curve calculated for a depth of 10 cm is compared to a cumulative curve for the same depth where $K_s \geq K_r$, accompanied by an evapotranspiration curve (Fig. 6). It is evident from this comparison that although the drainage rate is different for both cases in different time interval the final cumulative result is similar. The change of moisture content with time at the bottom of the 10 cm profile for both cases is illustrated in Fig. 7. Both cases differ from each other by their build up of moisture content. Where $K_s \geq K_r$, a higher moisture content is gained above the interface due to its flow limiting nature, compared to the other case where $K_r \geq K_s$. The temporary decrease in flow due to the limiting nature of the boundary is compensated by the higher moisture content which increases the flow as a result of the higher hydraulic conductivity. The final, cumulative result is thus similar in both cases.

SOIL SAMPLES MOISTURE CONTENT

The results of the soil moisture content of the field samples and the results obtained from the model are given in Table 3. Three out of four results are similar to that obtained in the model.

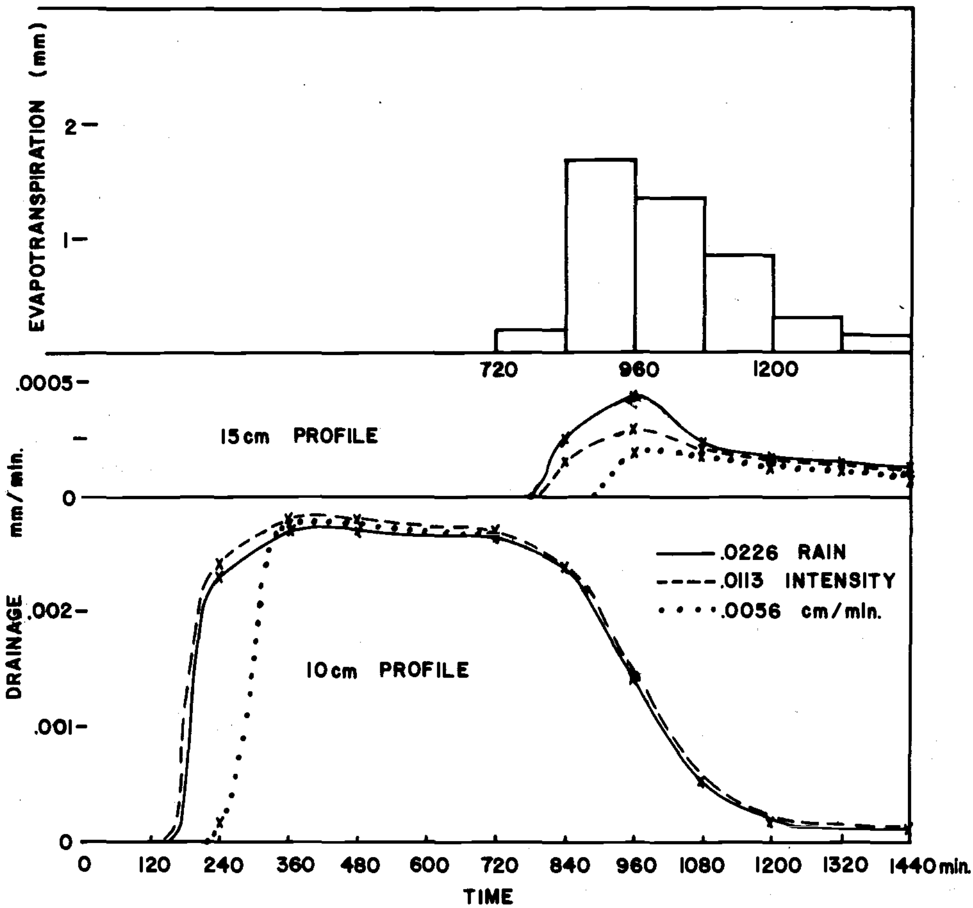

Figure 5. Drainage and evapotranspiration from a soil profile following a rainfall event.

Table 2. Cumulative drainage per event from different thickness of soil profile and different rainfall intensities (mm).

RAINFALL INTENSITY cm/min	DRAINAGE THROUGH		
	10 cm thickness	15 cm thickness	20 cm thickness
.0056	1.92	.08	0
.0113	2.17	.11	0
.0226	2.10	.14	0

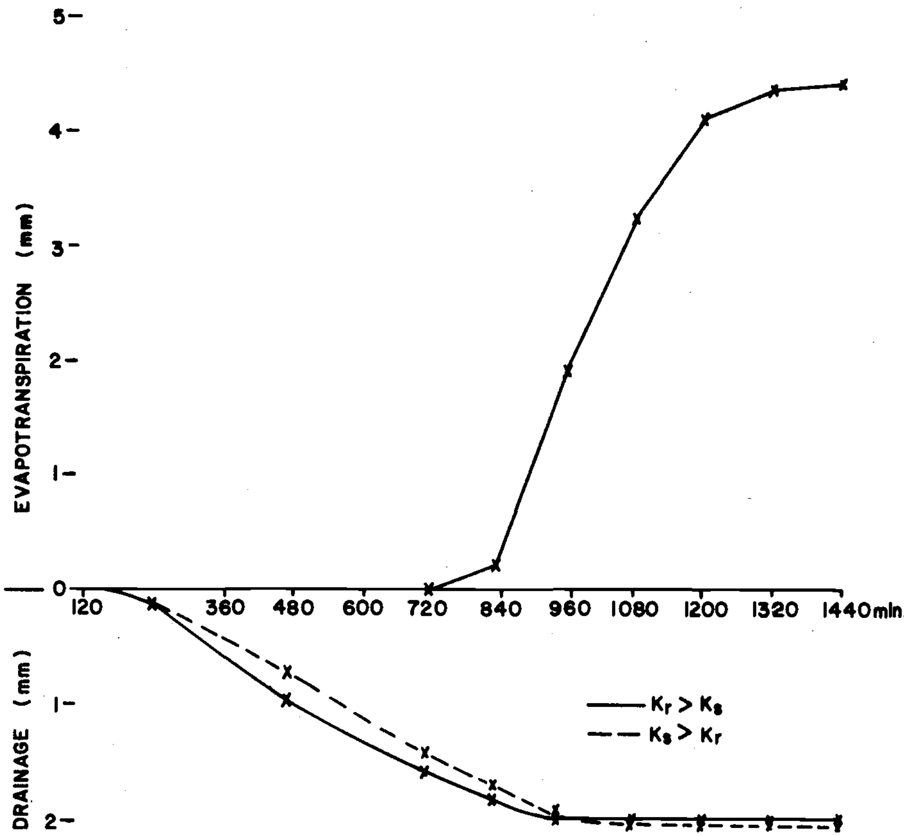

Figure 6. Cumulative evapotranspiration and drainage from a 10cm soil profile.

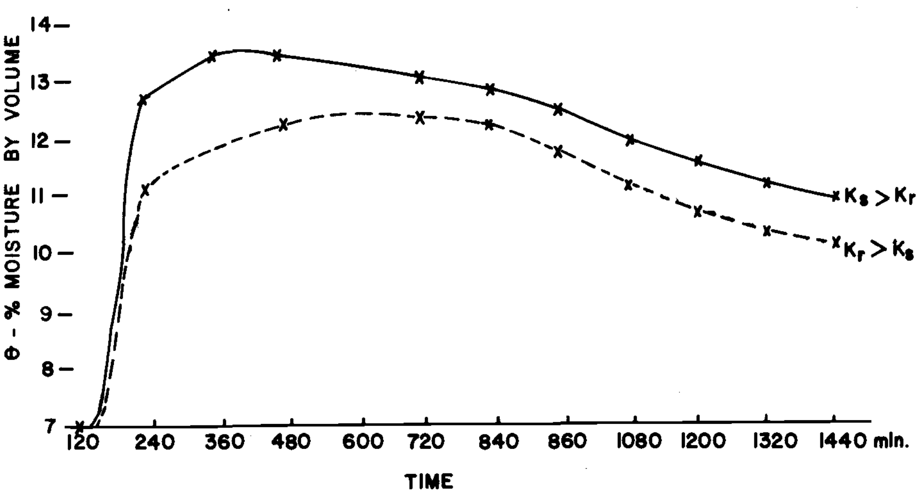

Figure 7. Moisture content at the bottom of a 10cm soil profile as a function of time.

Table 3. Measured moisture content from some field samples.

NO.	DEPTH (cm)	LOCATION	% MOISTURE CONTENT	% Moisture Content Computed by Model
1	10	Empire Mts.	11.4	
2	10	Empire Mts.	18.7	11.7
3	10	Empire Mts.	9.8	
4	10	Empire Mts.	10.1	

CONCLUSIONS

(a) There is no recharge from the summer type rainfall through soils exceeding a depth of 15 cm, which includes all of the lowland and part of the mountain soil cover.

(b) One can calculate the total recharge through a soil profile during the summer by multiplying the net recharge throughout one event by the number of events for a given elevation (Table 1). The results show a small recharge of 4-6 mm through the 15 cm profile and a considerable one of 59-92 mm for the 10 cm profile.

(c) The percentage of recharge through a 10 cm soil profile out of an average rainfall per event is calculated to be some 31%. In the mountain region the soils, 10-50 cm thick, cover some 50% of the total surface area. Assuming a radial symmetry of a soil element (Fig. 2) it is evident that the ratio between the 10 cm surface area and the total area of the element is approximately $2\Delta r/R$, where Δr is the width of the 10 cm deep soil segment and R is the radius of the element. Using these figures one can arrive at a rough estimate of the overall recharge contribution, out of the total rainfall, through these soils as follows: $0.31 \times 0.50 \times 0.40 \times 100 = 6.2\%$

This summer contribution of 6.2% to the water balance is considerable, having the same magnitude of the runoff values obtained in the study area from rainfall/runoff relationship.

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