J. v. Bakel

NOTE 1437 July 1983

INSTITUTE FOR LAND AND WATER MANAGEMENT RESEARCH
Wageningen, the Netherlands

THE GENERATION OF RIVER ALIMENTATION IN RESPONSE TO PRECIPITATION; A SOIL PHYSICAL APPROACH

dr. G.P. Wind and drs. A. Vandenberg

(Chapter of PAO-course
Real-time River Flow)

CONTENTS

10. THE GENERATION OF RIVER ALIMENTATION IN RESPONSE TO TO PRECIPITATION; A SOIL PHYSICAL APPROACH

G.P. WIND
Institute for Land and
Water Management Research
Wageningen, The Netherlands

A. VANDENBERG
National Hydrology Research
Institute,
Ottawa, Canada

10.1. ABSTRACT

River alimentation can be simulated with the aid of models developed by agricultural hydrologists. One-dimensional models with a Fourrier boundary condition are the most appropriate. They are simulating subsurface drain outflow and surface run-off from an input of precipitation and potential evaporation. The generation of long time series of several decades is economically possible.

A number of simulation-runs have been made over two very wet autumns and winters. A sensitivity analysis has been made of the effects of properties of soil and drainage on peak discharge rates. Properties determining storage capacity had strong influence. Properties governing flow processes were less important. The general conclusion is: the coarser the soil and the better the drainage the more smooth is the discharge pattern. Another conclusion: reclamation of swampy land to poorly drained agricultural land sharply increases peak discharges; these will decrease after improvement of the drainage.

10.2. INTRODUCTION

The ultimate cause of river flow is rainfall in its drainage basin. Rainfall is transformed however, in the soil system so that river alimentation rates seldomly equal precipitation rates. Mostly precipitation rates are reduced and retarded by storage- and flow processes in the soil and drainage system.

Water engineers are needing river alimentation rates in order to compute river flow data which are needed for the design of e.g. river profiles and levees.

Agricultural hydrologists are exploiting models of transient flow in saturated and unsaturated soil, which produce water discharges from the soil as an output. Although these models have been developed to simulate moisture conditions for crop growth and soil tillage they need a correct simulation of sub-surface drain outflow and surface run-off. These output-data might be important for use as input-data for river flow forecasting.

This paper gives some information on models which can be used and on the significance of soil- and drainage characteristics for the transformation of rainfall to water discharge rates.

10.3. MODELS OF TRANSIENT FLOW OF WATER IN SOILS

Soil scientists are using models of storage and flow of water in soils. Their aim is to predict workability, actual evapotranspiration and crop yield from an input of precipitation and other wheather data. Reason for these simulations is testing the effects of hydro-meliorative measures as drainage and soil improvement.

The models contain a saturated and an unsaturated past; because groundwater tables can rise and fall the level of the transition between saturated and unsaturated soil is generated by the model itself. Three types of models can be distinguished:

1. Approximation models;
2. One-dimensional models;
3. More-dimensional models.

Each of them is describing the processes of flow and storage in the unsaturated and saturated zone. In the given order the degree of correctness is increasing and with that the reliability but also the computing time and costs.

10.3.1. Approximating models

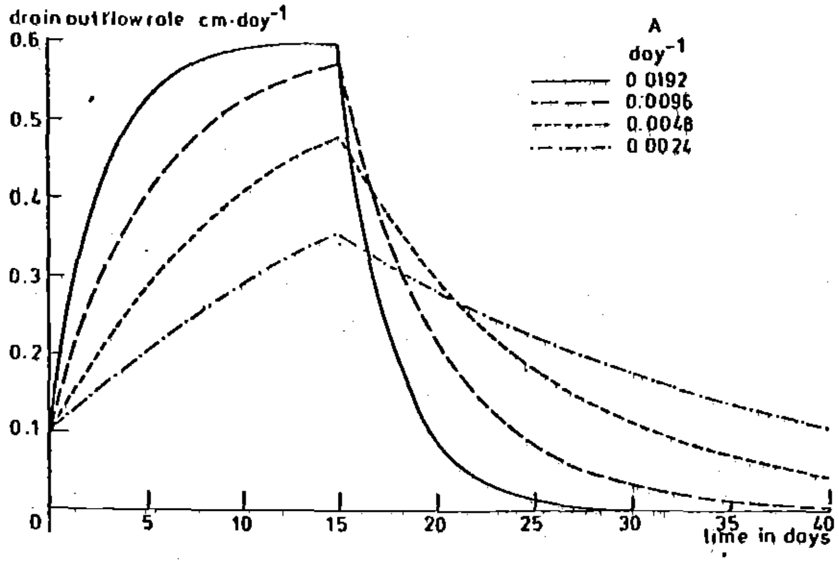

Fig. 1. Drain outflow rate computed by the 'De Zeeuw Hellinga' approximation model with 4 drainage intensities (A). Porosity 5%. Initial condition 0,1 cm.day^{-1}. Input 15 days rainfall of 0.6 cm.day^{-1}, after that zero rainfall.

If there is a rectilinear relation between drain discharge rate, V_D, and hydraulic head, h, of the groundwater above the drainage base the following differential equation holds:

$$V_D dt + p dh = i dt \tag{1}$$

Herein stands t for time, p for apparent porosity of the soil and i for precipitation rate. The linear relationship between drain discharge and hydraulic head is described by

$$A = \frac{V_D}{h} \tag{2}$$

in which A is called the drainage-intensity.

Integrated eq. (1) becomes:

$$V_D(t) = Ah(t) = V_0 e^{-\frac{A}{p}t} + i(1-e^{-\frac{A}{p}t}) \tag{3}$$

Herein is V_0 the drain discharge rate at the previous day. Such a model is developed by De Zeeuw and Hellinga (1958). When for every day the actual precipitation (i) is introduced, this simple model calculates draindischarge taking into account the amount of stored water.

Fig. 1 gives an example of such a calculation. After an initial condition of 0.1 cm rain per day, it is raining 15 days with a rate of 0.6 cm.day^{-1}. Thereafter precipitation rate drops to zero. The apparent porosity p = 0.05, and there are 4 drainage intensities. Discharge rates apparently are dependent on the drainage intensity. So it seems possible to calculate drain discharge rates, which equals river alimentation rate with this simple model, provided that A and p are known.

Drainintensity A might be a rough approximation for the relation between discharge rate and hydraulic head, which is not totally correct. Apparent porosity however, is a factor which is essentially not constant. Its value changes with depth of the groundwatertable and with the flux in the unsaturated zone. Moreover, transport of water through the unsaturated zone needs time; time-lags varying between zero and several months.

Therefore such models can serve only as a very rough approximation of river alimentation.

10.3.2. One dimensional models

A fairly correct treatment of what happens in the unsaturated zone is applied in one-dimensional models. These are calculating both fluxes and storage in a number of soil layers.

The models are based on a combination of the vertical unsaturated flux equation.

$$V = -K(\psi)\left(\frac{d\psi}{dz} + 1\right) \tag{4}$$

and the continuity equation

$$\frac{\partial\theta}{\partial t} = -\frac{\partial v}{\partial z} \tag{5}$$

Herein is V the vertical flux in cm.day^{-1} positive in upward direction
K the hydraulic conductivity in cm.day^{-1}
ψ soil moisture pressure head in cm, negative above groundwater

z vertical distance below surface in cm (negative)

θ volumetric moisture content

t time in day.

Two relations should be known; $\psi(\theta)$ the moisture characteristic and $k(\psi)$ the conductivity curve. The latter can be expressed by a number of functions; often an exponential function is used:

$$k(\psi) = k_o e^{\alpha\psi} \quad (6)$$

Herein k_o is the conductivity at zero moisture pressure head, which is often considerably lower than saturated conductivity, and α is a coefficient (cm^{-1}) determining the rate of decrease of k with decreasing ψ.

For the models the different equations (4) and (5) are discreted to finite difference or finite element models. Mostly layers of $\Delta Z = 10$ cm are used. The time step size is chosen according to a stability criterion; mostly $\Delta t < 0.01$ day. This causes that the models require much computertime. Nevertheless some models can be used for long time series, e.g. 30 years, at reasonable costs. Examples of these models are SWATR by Feddes et.al.(1978) and FLOW by Wind and Van Doorne (1975). SWATR fits well for summer conditions; FLOW for winter conditions.

Every one-dimensional model requires an initial condition and two boundary conditions. The upper boundary condition is given by the input of daily rain and evaporation data. The lower boundary condition can be the Fourrier condition that a relation is given between flux and potential (pressure head) at the bottom of the model. This relation can be the same as in eq. (2) or it can be made more complicated.

10.3.3. More dimensional models

If the relation between hydraulic head midway between the drains and drain discharge is not univocal the one-dimensional models can not be used. This is the case if the discharges relation is not the same for all places between two drains. If so a physically correct description of the saturated zone must be brought into the model. That requires a two- or three- or quasi-three-dimensional model. Such models are described a.o. by Zaradny and Feddes (1979) and Neuman, Feddes and Bresler (1975).

Because such models need very much computer-time they cannot be used for long time series up to now. The computing costs will be prohibitive.

In the following only a one-dimensional model will be used to show its possibilities for calculation of river alimentation.

10.4. APPLICATION OF ONE-DIMENSIONAL MODELS TO FORECASTING OF RIVER ALIMENTATION

10.4.1. Models used

For the purpose of simulation of river alimentation the models FLOW by Wind and Van Doorne (1975) and ELAN by Wind and Mazee (1979) are used. Both models have in common that eq. (4) is applied in integrated shape using eq. (6) as $k(\psi)$ relation.

The flux equation then becomes

$$V = \frac{k_2/a - k_1}{1 - 1/a} \qquad (7)$$

Herein V is vertical flux in $cm.day^{-1}$, positive in upward direction, k is the hydraulic conductivity in $cm.day^{-1}$, index 1 refers to the upper layer and index 2 to the lower.
The factor a is

$$a = e^{\alpha \Delta z} \qquad (8)$$

Herein α is the soil constant from eq. (6) and Δz is the distance in cm between the layers 1 and 2. Mostly $\Delta z = 10$ cm is used as an intermediate between accuracy and computing costs. Computing costs are reversely proportional to the third degree of Δz.

FLOW is a numerical model written in FORTRAN IV; ELAN is an electronic analog based on the resemblance of eq. (7) with Ohms law. Both models are based on the same principles and are giving the same output results. Therefore details of ELAN shall not be treated in this paper.

10.4.2. Storage of moisture in the unsaturated soil

Difference in vertical flux above and below a soil layer causes an increase or decrease of moisture in that layer according to the continuity equation (5). An increase of moisture content also causes an increase of soil moisture pressure head and of conductivity.

Relations between pressure head and moisture content are given in fig. 2 for the soils used in this paper: clay, sandy loam and sand.

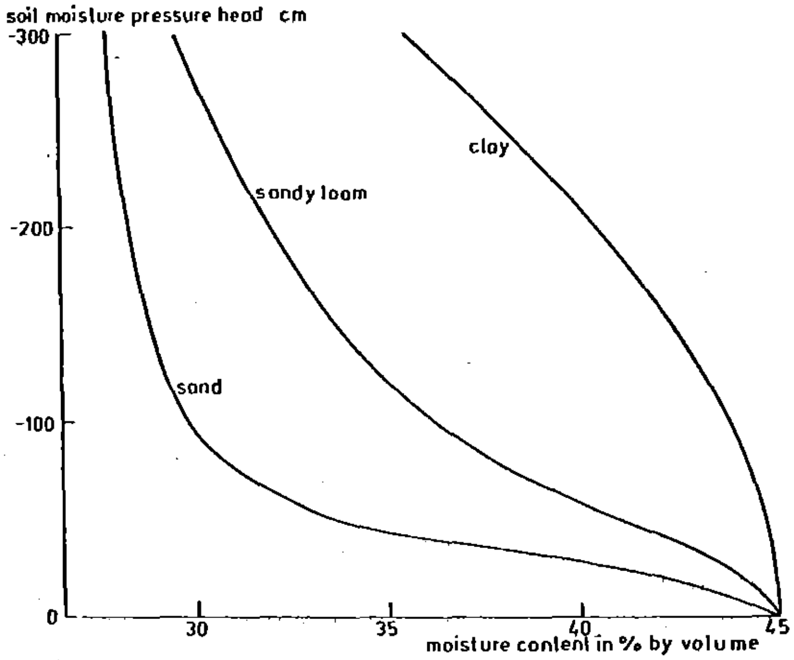

Fig. 2. Soil moisture characteristics, $\psi(\theta)$ relation, of the soils used in the simulations

10.4.3. Infiltration and run off

Expression (7) is also applied for computing infiltration, assuming k_1 is zero during rainfall. When infiltration rate is lower than precipitation rate water is stored upon the soil surface. The amount of this is called pool depth, symbol p (cm).

Surface run off is thought to be dependent on pool depth. In the following expression (9) is applied

$$V_s = c.p^2 \qquad (9)$$

Herein V_s is surface run-off rate and c is a proportionality factor, dimension: $cm^{-1}.day^{-1}$. Low values of c are representing a high resistance for over-land flow. Different expressions might also be used instead of (9). In this paper always c = 1 is used, except in the 'swamp' of fig. 7 where c = 0.01.

Drainage

The drainage function is serving as lower boundary condition for the model. According to Hooghoudt (1947) there is a relation between drain outflow rate (V_0) and hydraulic head of the groundwater.

In its most simple form this relation is rectilinear for the hydraulic head midway between two parallel drains.

$$V_D = -A\,\psi_D \tag{10}$$

Herein V_D is drain outflow rate and ψ_D is the soil moisture pressure head at draindepth. A is the drainage intensity in day^{-1}.

Application of this (10) boundary condition gives rise to 3 equations with 3 unknown factors: V_D, ψ_D and the height of groundwater table. These can only be solved by an iteration procedure. Nevertheless a more complex expression than (10) can be applied.

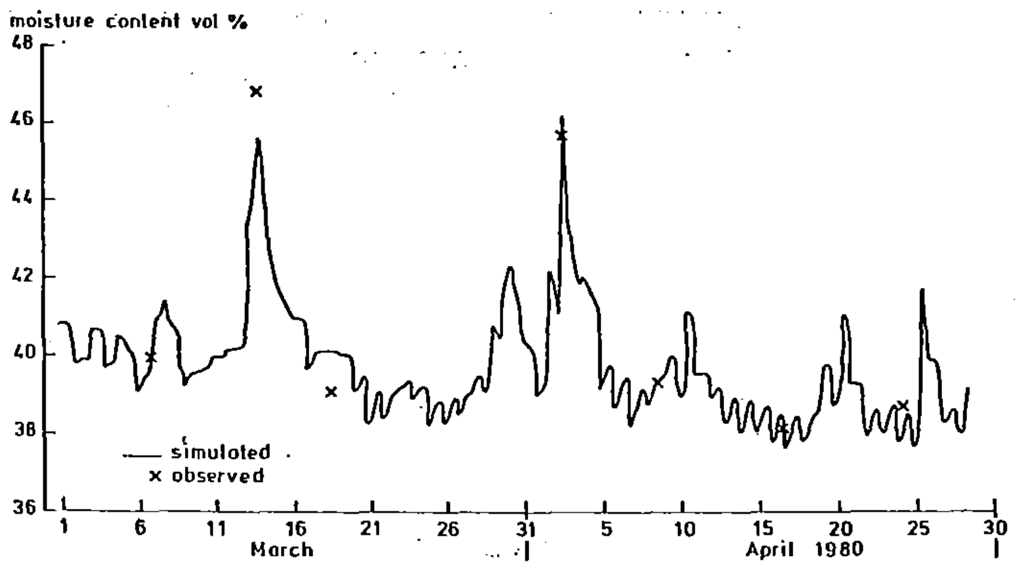

Fig. 3. Soil moisture content at 5 cm depth computed by model FLOW compared with field observations.

10.4.4. Input

As input actual daily values of precipitation and evaporation are fed into the model. Examples are shown from a simple input pattern: 0.1 cm/day as initial condition followed by 15 days of 0.6 cm/day followed by zero rainfall.

Other examples are calculated with the actual wheather data from September 1 to May 31 of the years 1961/1962 and 1974/1975 in the center of the Netherlands (De Bilt, Royal Meteorological Institute).

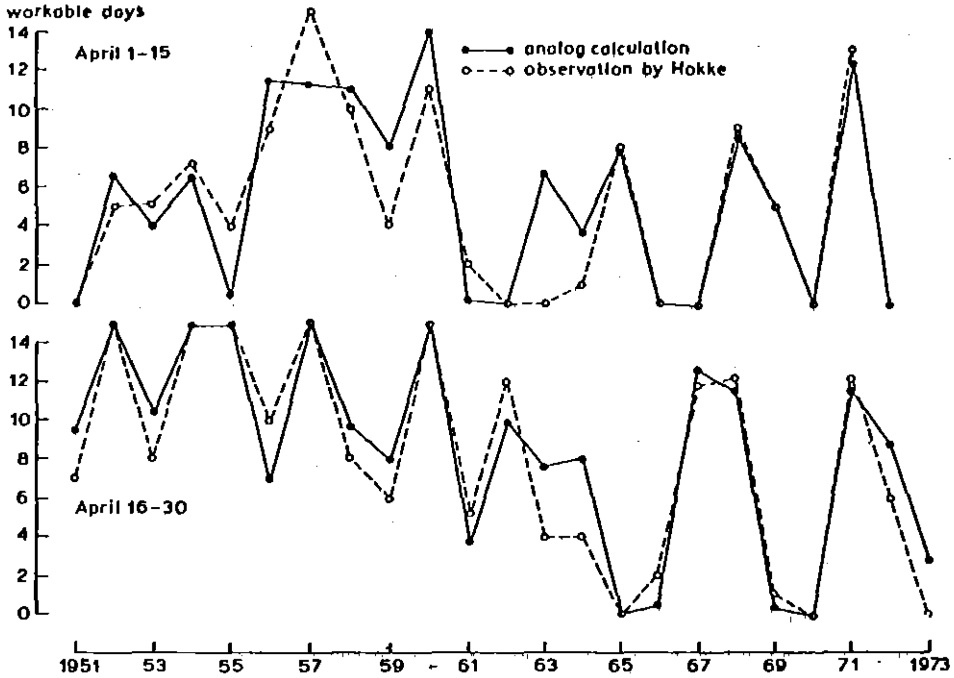

Fig. 4. Observed and computed number of workable days over 23 years

10.4.5. Check of the model

Before the models have been applied in agronomic research they have been checked in several ways. A direct check between observed and simulated discharge rates should be established before application in river flow forecasting. The checks shown in fig. 3 and 4 however, are indicating that the model works well in predicting soil moisture. That is only possible if the discharge function (10) is correct.

10.5. RESULTS OF A SENSITIVITY ANALYSIS OF SOIL AND DRAINAGE PROPERTIES FOR WATER DISCHARGE FROM SOIL INTO RIVER

In order to show how soil physical properties and drainage affect the transformation of precipitation into river alimentation a large number of simulation runs have been carried out. The rain- and evaporation inputs have been described in the previous chapter for the short run and the two long runs.

In the following the effects of drainage intensity, draindepth, soil moisture characteristic and hydraulic conductivity will be shown. The discharge rate mentioned in the figures refers to the sum of sub surface discharge and the surface run off rates.

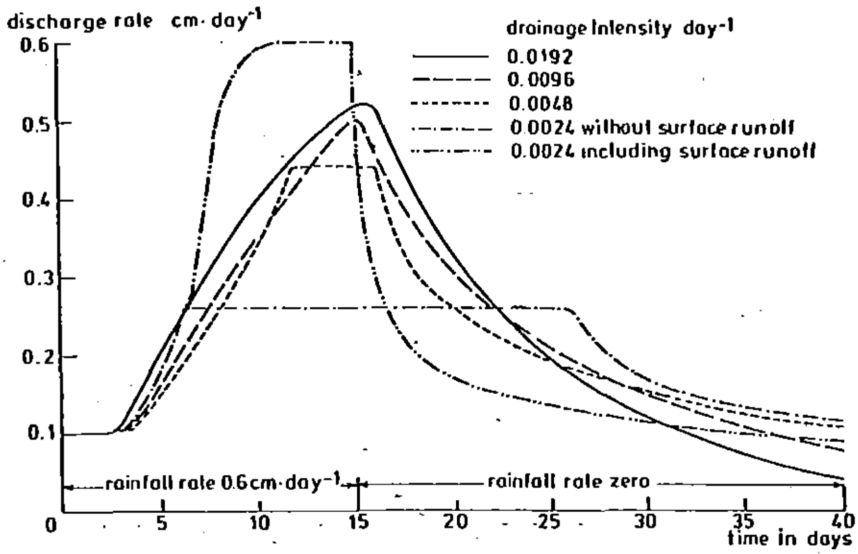

Fig. 5. Drain outflow rates computed by model FLOW for a sandy loam soil. Conditions and input the same as in fig. 1. Drain depth 120 cm. One simulation with surface run-off, according to eq. (9). In the other 4 simulations surface run-off is put to zero.

10.5.1. Drainage intensity

Fig. 5 shows a short run with four drainage intensities (A) for a draindepth of 120 cm in a sandy loam soil. The poorest drainage causes groundwater-rise to surface in 6 days. From then drain-outflow-rate is 0.26 cm.day^{-1}. As precipitation rate is 0.6 cm.day^{-1} the soil surface is ponded with water, causing surface run off. The figure also gives a line of a situation where surface run off was prevented.

Regarding the effect of the largest three intensities one sees that the differences are small. Comparing fig. 5 with fig. 1 one sees that the effect of drainage intensity is heavily overestimated by the approximating model. This is caused by neglecting a part of static and dynamic storage in the very simple model. Another difference is the large time-lag in fig. 5 which lacks in fig. 1.

The largest peak discharge in fig. 5 is caused by the poorest drainage intensity due to surface run off. The second largest peak discharge is caused by the best drainage. The differences in peak discharge however, are not very large. This is confirmed by the real time runs of 1961 and 1974. both autumns with very large rainfall.

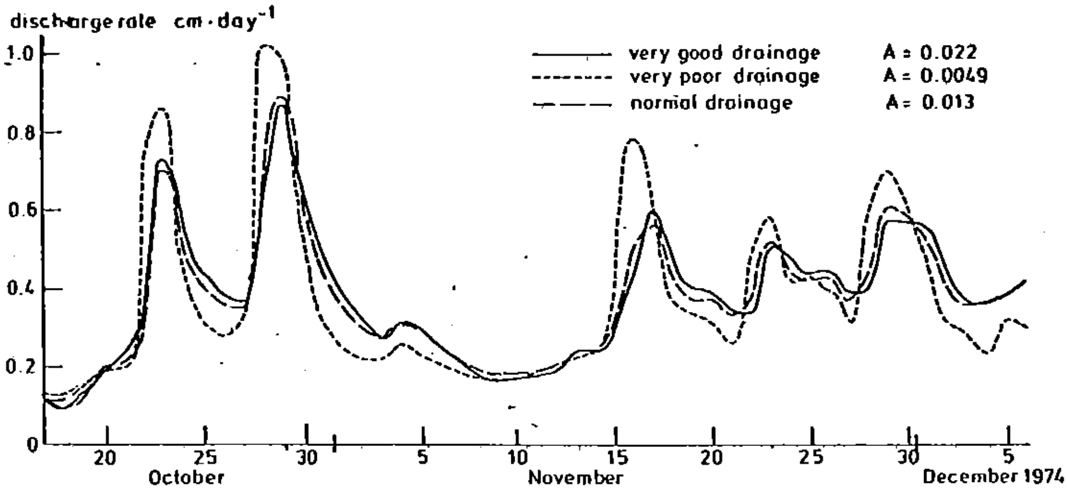

Fig. 6. Effect of drainage intensity on discharge rates (including surface run-off) out of a sandy loam drained at 100 cm depth

Fig. 6 shows the effect of drainage intensity on discharge rate for a part of 1974. The normal drainage and the very good one are causing nearly the same discharge rates. The very poor drainage shows peaks which are 1 or 2 $mm.day^{-1}$ higher.

So the effect of drainage intensity on discharge rate is fairly small; the better the drainage the smaller the peak discharge.

10.5.2. Drain depth

Drain depth has a pronounced effect on total discharge rate. The effect is even so large that the lines in fig. 7 seem to come from different rainfall patterns. But it is the same for all 5 draindepths. The differences between discharge rates are caused by the differences in storage capacity. This is very low for the shallow drainage and very high for the deep one. In the course of October storage capacity in the deeper drainages is decreasing due to the large rainfall and the small discharge rates. One sees that the peak discharges of draindepth 100 cm are increasing during the month of October whereas the peaks of drain-depth 50 cm are decreasing.

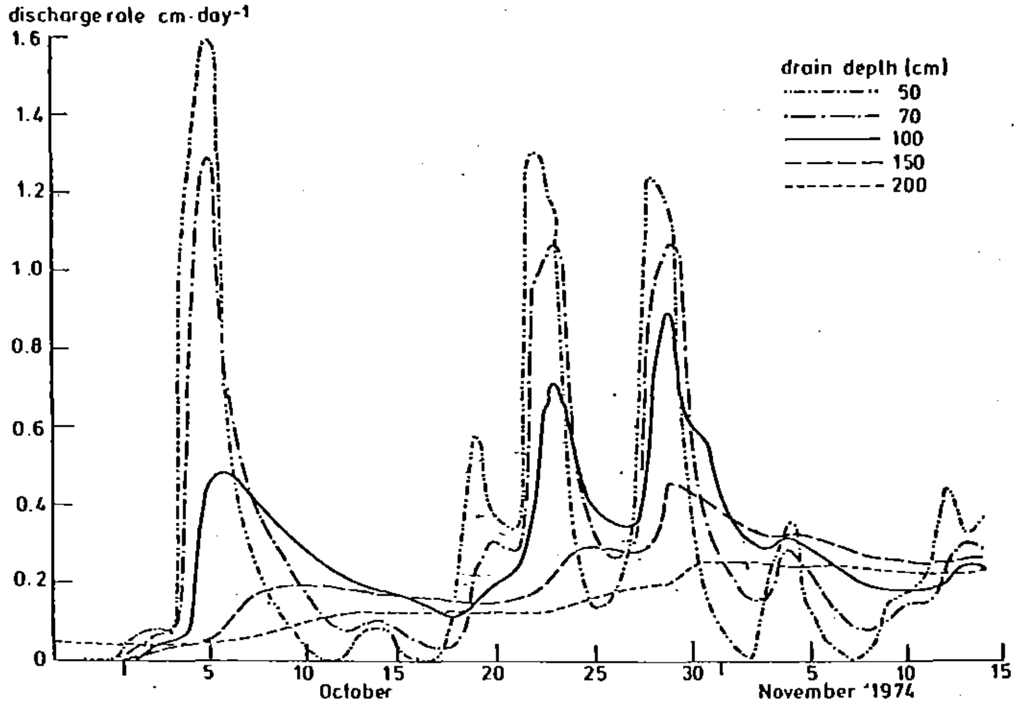

Fig. 7. Effect of drain depth on discharge rates out of a sandy loam with equivalent drainage intensities

The peak discharges are the larger the shallower the drainage is. Very deep drainage causes a very smooth discharge pattern without peaks. In table I number and discharge rates of all peaks in 1961/62 and 1974/75 are given for 6 drain depths. Both years were exceptionally wet in the period September to February which was simulated by the model for a sandy loam soil. For 8 soils simulation runs have been recently made over 30 years between 1950 and 1980. An example of the occurring discharge rates for a silt loam is given in figure 8.

Table I. Number of peaks and belonging discharge rates in autumn and winter of the years 1961/62 and 1974/75 for a sandy loam soil

Discharge rate $cm.day^{-1}$	Draindepth (cm) 50	70	100	150	200	300
>2.0	1	1	-	-	-	-
1.8-2.0	-	-	1	-	-	-
1.6-1.8	1	1	1	-	-	-
1.4-1.6	2	-	-	-	-	-
1.2-1.4	10	3	1	2	-	-
1.0-1.2	10	6	1	1	-	-
0.8-1.0	7	11	5	1	1	-

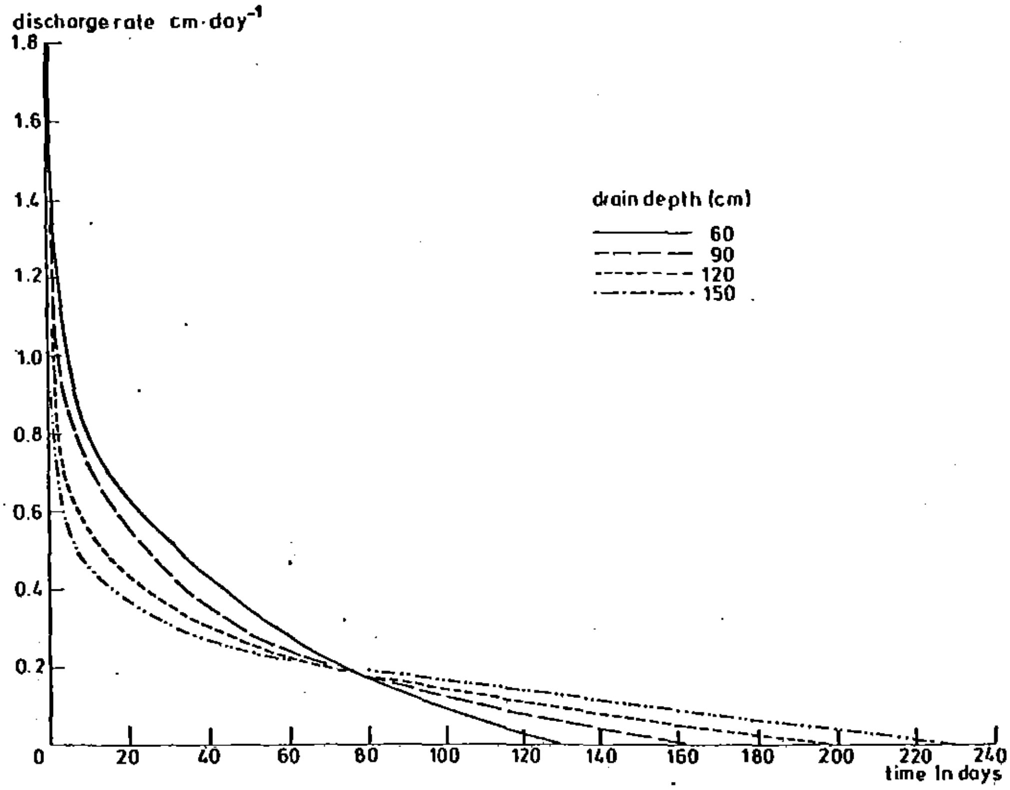

Fig. 8. Cumulative frequency distribution of the average yearly discharge rates for a silt loam soil in the Netherlands. Results of 30 year simulation between 1950 and 1980 with 4 drain depths.

These figures and table show that the deeper the drainage, the more smooth is the discharge pattern. In practice however, it is known that reclamations are causing high floods. This controversy is shown in figure 9. The system called 'swamp' in this figure has no subsurface drainage; c = 0,01 in eq. (9). One sees that the discharge pattern of the swamp is more smooth than that of the poorly drained soil but less smooth than the well drained soil.

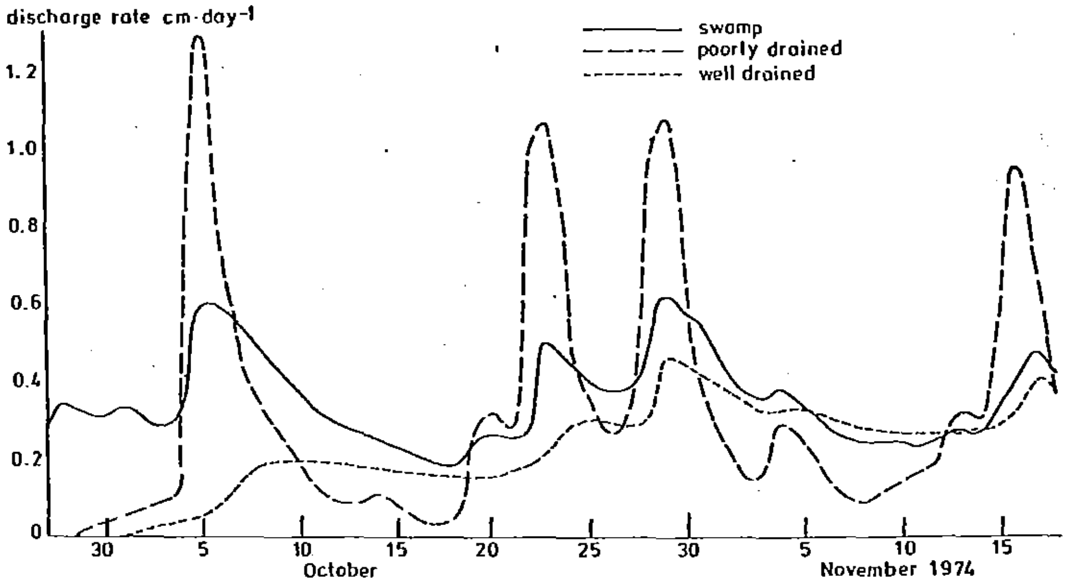

Fig. 9. Comparison of discharge patterns of a 'swamp' (no sub-surface drainage, in run-off eq. (9) (c = 0,01), a poorly drained (D = 70; A = 0,028) and a well drained sandy loam (D = 150; A = 0,0054). Both 'swamp' and well drained soil have smaller discharge peaks than poorly drained.

So the first reclamation of natural or waste land will cause high floods. Subsequent ameliorations will decrease number and heights of the discharge peaks.

10.5.3. Soil moisture characteristic

Fig. 10 shows the discharge patterns of a sand, a sandy loam soil and a clay soil. These soils were given the same draindepth (D = 150 cm) the same drainage intensity (A = 0.0054 day^{-1}) and the same hydraulic conductivity (k_o = 2 $cm.day^{-1}$ and α = 0,02 cm^{-1}). So the sole effect of the moisture characteristic is shown.

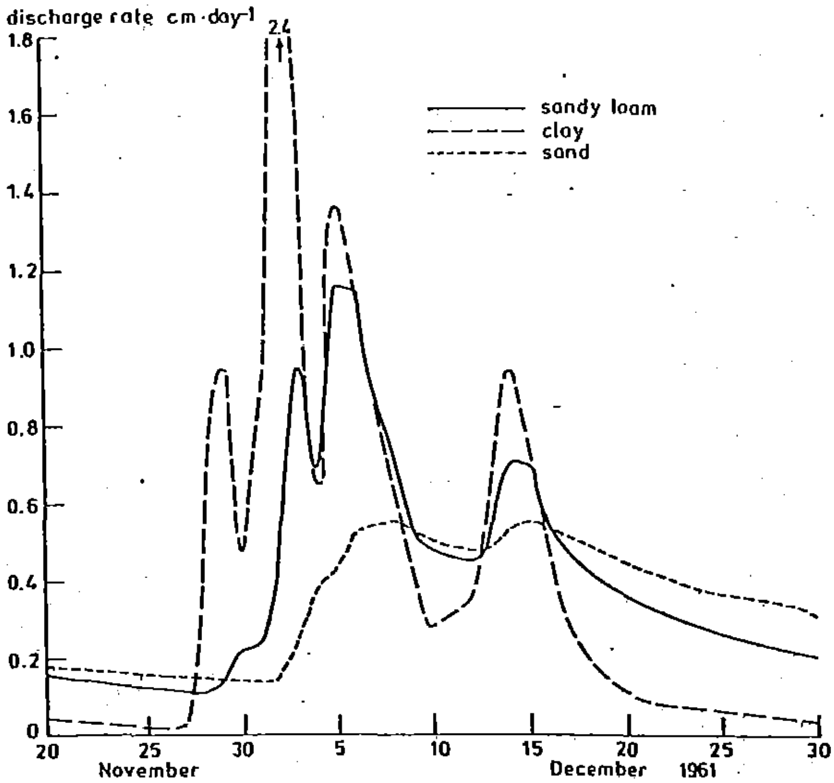

Fig. 10. Discharge rates out of clay, sandy loam and sand. Each soil has the same drainage (D = 150 cm; A = 0.0054 day^{-1}) and the same conductivity (k_o = 2 cm.day^{-1}); α = 0.02 cm^{-1})

The clay gives a very irregular discharge pattern, the highest peak being 2.4 cm.day^{-1}; the sand has a fairly smooth pattern in which the discharge rate remains below 0.6 cm.day^{-1}. The sandy loam soil lies in between with a highest peak of 1.2 cm.day^{-1}.

Here also it is the storage capacity of the soil that causes these differences. As fig. 2 shows this decreases in the direction sand-sandy loam-clay.

10.5.4. Hydraulic conductivity

Hydraulic conductivity is acting in the saturated horizontal flow. The effect of it has been treated implicitely in that of drainage intensity. But it also acts in the unsaturated vertical flow. There hydraulic conductivity is characterized by two factors, k_o the conductivity at zero pressure head and α the rate of decrease with decreasing pressure heads, see eq. (6).

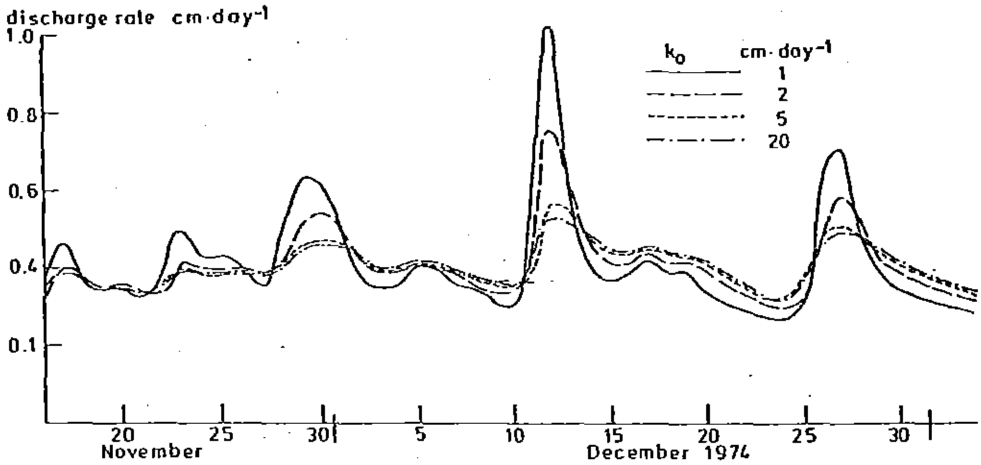

Fig. 11. Effect of the conductivity value k_o on discharge pattern. Sandy loam soil, drainage depth 150 cm, intensity 0.0054 day^{-1}

The effect of k_o is demonstrated in fig. 11 for a sandy loam drained at 150 cm depth with A = 0.0054. There k_o varies from 1 to 20 $cm.day^{-1}$; α is kept constant at 0.02 cm^{-1}. The peak discharge rates of the fairly large k_o-values are nearly the same. Only the low conductivities are showing important differences. That stands to reason because the precipitation rates often exceed the hydraulic conductivity values of 1 and 2 $cm.day^{-1}$. In that case the difference between precipitation- and infiltration rate is discharged as surface run-off.

The effect of α , shown in fig. 12 is more pronounced than that of k_o. Moreover, it occurs over the whole range of used α-values, whereas k_o has no influence above $k_o \doteq 5$ $cm.day^{-1}$. The explanation of the significance of α for peak discharges is that α affects the storage capacity. Soils with large α have very low unsaturated conductivities. That causes vertical fluxes to approach to zero already at fairly moist conditions. High α-soils thus are getting only a low storage capacity in dry periods.

10.6 DISCUSSION

A number of soil- and drainage-properties seem to have influence on discharge rates in periods with heavy rainfall. The difference between the highest and the lowest peak discharge caused by these properties of soil and drainage are given in table IL The other properties of soil and drainage have been kept constant.

Although only 3 peaks are given in table II the variability of the difference between high and low peaks is paramount. But as an average draindepth appears to be the most important factor governing peak discharges. It is directly followed by the moisture charactistic of the soil. Steep moisture characteristics (clay soils) are causing high floods too. The value of α has also an influence on peak discharge but clearly to a less extent. All these three factors are determining storage capacity of the soil.

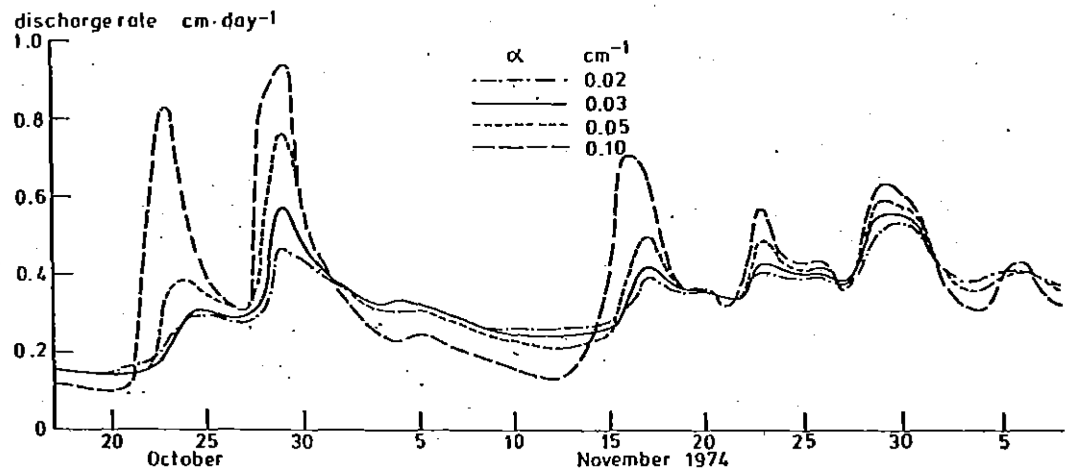

Fig. 12. Effect of the conductivity value α on discharge patterns, for the same conditions as fig. 11.

Less important seem to be the factors which are determining velocity of flow in the soil: the hydraulic conductivity and the drainage intensity. The lower these are the higher are the peak discharges.

Table II. Difference between highest and lowest peak discharge rates in cm.day^{-1} caused by some properties of soil and drainage for three periods with heavy rainfall.

Date of discharge	Properties of soil and drainage				
	drain depth	moisture characteristic	Soil constant	hydraulic conducitivity	drainage intensity
1974 Oct 29	1.04	0.75	0.48	0.25	0.09
1974 Dec 12	1.08	0.79	0.27	0.50	0.32
1961 Dec ±5	1.93	1.76	0.92	0.43	0.66
Average	1.35	1.10	0.56	0.39	0.36

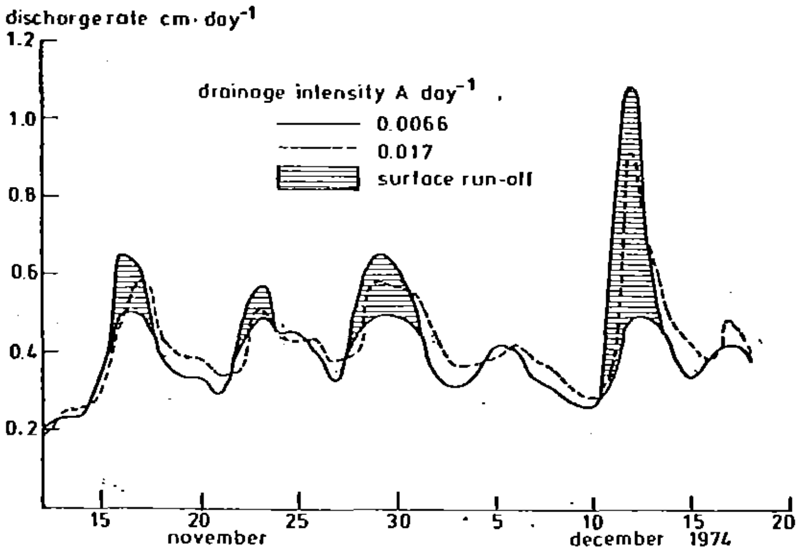

Fig. 13. The low drainage intensity (0.0066) gives lower discharge peaks than the high intensity (0.017), but this difference is overcompensated by surface run-off.

Soil and drainage properties enabling large fluxes are counterbalanced by the system of surface and subsurface drainage. Mostly an overcompensation occurs. Therefore natural conditions and artificial measures to promote large sub surface fluxes do not cause high floods but in the contrary are contributing to a more smooth discharge pattern.

The overcompensation is clearly shown in fig. 13 where a low- and high intensity drainage are compared. Subsurface discharge of the poor drainage is mostly smaller than that of the good one. Because of surface run-off the total discharge rates of the poorly drained field is the largest in wet periods.

10.7 LITERATURE

FEDDES, R.A., P.J. KOWALIK AND H. ZARADNY, 1978. Simulation of field water use and crop yield. Wageningen, Centre for Agr. Publishing and Documentation.

NEUMAN, S.P., R.A. FEDDES AND E. BRESLER, 1975. Finite element analysis of two dimensional flow in soils considering water uptake by roots. Soil Sci. Soc. Am. Proc. 39 (2): 224-230.

WIND, G.P. AND A.N. MAZEE, 1979. An electronic analog for unsaturated flow and accumulation of moisture in soils. J. Hydrol. 41: 69-83.

WIND, G.P. AND W. VAN DOORNE, 1975. A numerical model for the simulation of an saturated flow of moisture in soils. J. Hydrol. 24: 1-20.

ZARADNY, H. AND R.A. FEDDES, 1979. Calculation of non steady flow towards a drain in saturated-unsaturated soil by finite elements. Agric. Water Management 2: 37-53.

ZEEUW, J.W. DE AND F. HELLINGA, 1958. Neerslag en afvoer. Landb. Tijdschr 70: 405-421.