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# Temporal variability of soil surface crust conductivity

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**Abstract**

The temporal variability of soil surface crust conductivity is studied on two different time scales. Undisturbed soil samples were taken from four different plots in order to study the infiltration-runoff behaviour in the laboratory. A simulation model was applied to compute surface runoff and soil suction profiles. Taking best fit between the computed and experimentally measured data hydraulic properties of the soil and of the crust were estimated. In nearly all of the 22 analyzed soil samples, crust development could be observed in the laboratory. Because sampling was repeated during the vegetation period it was feasible to discern if decreases in hydraulic conductivity of the crust occurred in the long term. The data showed no long-term decreases although in the laboratory, crust development was observed in nearly every case. If short and intensive rainfall events were to be simulated, the initial condition of the crust would be the most sensitive parameter for the runoff amount.

*Key words:* Temporal variability; Soil surface; Conductivity of soil surface crust

## 1. Introduction

A soil surface seal is usually thin (1–5 mm) and does not crack (Sombroek, 1985) and the saturated hydraulic conductivity of the seal is low. McIntyre (1958) found the permeability of the seal to be up to 1/2000 times smaller than the permeability of the underlying soil. As opposed to a seal, however a soil surface crust cracks and is moderately thick (5–15 mm). The hydraulic conductivity of a crust is low compared with that of the underlying soil but does not reach the extreme low values of a seal.

Because of the micro relief resulting from tillage the crust thickness may vary from place

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to place on a small scale ($cm^2$) (Freebairn et al., 1991). In order to describe the infiltration runoff behavior of larger areas ($m^2$) the varying crust thickness has been characterized by a mean value in the present study, and the influence of the crust thickness on the runoff amount is discussed.

Many studies have shown the influence of crusting on the rainfall runoff relationship (Moore, 1981; Morel-Seytoux, 1983; Aboujaoudé, 1991; Mualem et al., 1990; Mualem and Assouline, 1991). Numerical models have been developed capable of describing the increase in runoff due to crust formation. Some of these models have been tested against laboratory investigations in which soil have been irrigated and the runoff-infiltration behavior studied. Usually these experiments last a few hours mostly using disturbed soil. Because crusts may crack, the question arises whether or not the laboratory results can be transferred to field conditions. The most important question is whether or not a decrease in crust conductivity can be observed over a vegetative period. This is important to know because, often, simulation models are applied to single storms without modeling the inter storm period. In this case the initial conditions like water content and crust condition should be considered (Le Bissonais, 1990), but are usually unknown.

## 2. Aim of the paper

In order to provide a few answers to these questions short term laboratory investigations were done repeatedly during a vegetative period on four different plots.

Because undisturbed soil samples were taken from the investigated plots at different time intervals short term as well as long term effects were distinguished in the study. In order to describe the crust effects a numerical model for the simulation of infiltration, runoff and soil water flux was developed. This model was applied to simulate the observed runoff and suction profiles of 22 soil samples. From the best fit of observed and numerically simulated data, the hydraulic characteristics of the soils as well as saturated conductivities of the crusts were estimeated.

## 3. Materials and methods

### 3.1. Site description

The research was conducted in a catchment near Neuenkirchen, situated about 35 km south of Braunschweig, Germany, in the northern foreland of the Harz mountains. This catchment covers an area of about 1 $km^2$ and has an average altitude of 150 m above sea level. The land use is exclusively agricultural comprising namely sugar beets, winter barley and winter wheat. Silty, clayey and stony quaternary sediments and soils dominate this catchment, described in detail by Bork and Rohdenburg (1986) and Othmer and Bork (1989). Four different plots were investigated, chosen on the basis of soil properties and crop type. Details of the soil properties are shown in Table 1.

Two of the four plots consisted of loess material (redeposited by fluvial processes in late Wurmian). During early and middle Holocene Orthic Luvisols developed. Since the Middle

Table 1
Sediment type and soil properties of the investigated plots

| Plot Crops | Sediment type | Clay[a] (% by weight) | Silt[a] (% by weight) | Sand[a] (% by weight) | $C_{org}$ (% by weight) | Number of soil samples |
|---|---|---|---|---|---|---|
| M1 sugar beets | redeposited upper cretaceous clayey marl | 29.6 | 60.8 | 9.6 | 1.3 | 6 |
| M2 sugar beets | redeposited loess | 21.2 | 71.9 | 6.9 | 0.9 | 5 |
| M3 winter wheat | redeposited upper cretaceous clayey marl | 38.4 | 60.7 | 0.9 | 1.2 | 7 |
| M4 winter wheat | redeposited loess | 14.6 | 80.3 | 5.1 | 1.4 | 4 |

[a]Texture in % of humus-/carbonate free fine soil according to the German particle size classes (clay < 0.002 mm < silt < 0.063 mm < sand < 2.0 mm).

Ages the clay eluviated A-horizons were eroded. Silty material that was eroded upslope was deposited at the investigation sites (recent clay contents in the ploughing horizon: 14 and 21%). The other two plots consisted of redeposited clayey marl with clay contents of 30 and 38%. Sugar beets and winter wheat were planted. Details of sampling dates and plant cover can be seen in Table 2.

Undisturbed soil samples were taken from the ploughing horizon of the study plots. The sample size was 0.33 × 0.3 m with a depth of 14 cm. The soil was extracted by pressing a sample box horizontally into the soil using a lifting jack as illustrated in Fig. 1. Tensiometers were installed in this sample box at depths of 1.5, 6.5 and 11.0 cm. The box was perforated at the bottom and placed on a sand bed in which tensiometer cups had been installed. After connecting these tensiometer cups with a pump, a constant soil suction at the lower boundary could be obtained. The percolation rate, the runoff and the water splashed off the probe were measured every 5 to 10 min. The soil suction was registered at 2 to 10 min intervals.

### 3.2. Rainfall simulation

A schematic diagram of the rainfall simulator is given in Fig. 2. This equipment was installed in the staircase of the Institute of Geography and Geoecology (Technical University of Braunschweig). The height of fall was about 23 m, enough to reach the natural fall velocity: 8.9 m/sec for a drop size of 3.9 mm (Laws, 1941). In order to exclude wind effects 19 of the 23 m were covered by a plastic tube. The rainfall simulator consisted of a plate in which 524 capillary tubes with diameters of 0.1 mm (324), 0.5 mm (100) and 1 mm (100) were installed. These tubes produced drop sizes in the range of 2.4 to 4.1 mm. This range could be extended by using different wire nets. For every capillary type, a separate water reservoir was installed. Because the water level above the capillary tubes was separately adjustable for each reservoir, the rainfall intensity as well as the drop size

Table 2
Sampling dates, rainfall intensities and calibrated *Cr* and $K_s$ values

| Soil sample | Date | Days after tillage | Rainfall intensity (mm/min) | Plant cover (%) | *Cr* | $K_s$ of non-crusted soil (cm/day) |
|---|---|---|---|---|---|---|
| M1-2 | 03/14 | 115 | 1.15 | 0.0 | 0.16 | 55.0 |
| M1-3[a] | 06/05 | 62 | 0.75 | 5.0 | 0.01 | 16.0 |
| M1-4 | 06/20 | 77 | 0.68 | 0.0 | 0.10 | 55.0 |
| M1-5[a] | 08/28 | 146 | 0.92 | 95.0 | 0.40 | 80.0 |
| M1-6 | 09/11 | 160 | 0.82 | 30.0 | 0.90 | 30.0 |
| M1-7 | 11/13 | 23 | 0.58 | 0.0 | 0.28 | 80.0 |
| M2-2[a] | 03/14 | 115 | 0.85 | 0.0 | 0.95 | 27.0 |
| M2-3 | 05/23 | 49 | 0.91 | 5.0 | 0.10 | 30.0 |
| M2-4 | 06/12 | 69 | 0.71 | 0.0 | 0.15 | 20.0 |
| M2-6[a] | 09/19 | 168 | 0.77 | 30.0 | 0.35 | 50.0 |
| M2-7 | 11/27 | 37 | 0.45 | 0.0 | 0.52 | 70.0 |
| M3-1 | 02/29 | 103 | 1.17 | 3.0 | 0.33 | 83.0 |
| M3-2[a] | 05/04 | 168 | 1.12 | 5.0 | 0.27 | 80.0 |
| M3-3 | 05/17 | 181 | 0.89 | 75.0 | 0.37 | 120.0 |
| M3-4[a] | 07/06 | 231 | 0.72 | 80.0 | 0.65 | 100.0 |
| M3-5 | 07/06 | 231 | 0.77 | 30.0 | 0.32 | 70.0 |
| M3-6 | 07/17 | 242 | 0.75 | 5.0 | 0.10 | 80.0 |
| M3-7 | 10/18 | 1 | 0.79 | 0.0 | 0.12 | 80.0 |
| M4-4 | 04/24 | 166 | 0.88 | 0.0 | 0.40 | 15.0 |
| M4-5 | 05/10 | 182 | 0.92 | 55.0 | 0.07 | 90.0 |
| M4-6 | 06/26 | 229 | 0.63 | 75.0 | 0.23 | 150.0 |
| M4-7 | 07/10 | 243 | 0.87 | 5.0 | 0.18 | 60.0 |

[a]Significant differences between observed and simulated soil suctions.

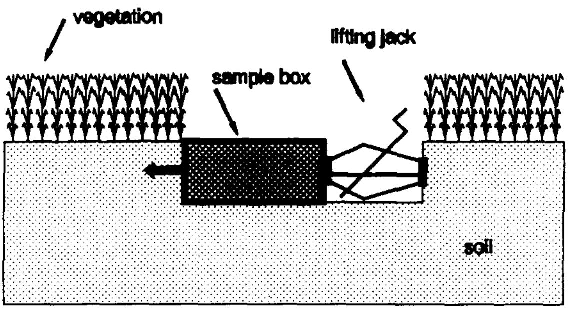


Fig. 1. Extraction of undisturbed soil samples.

distribution could be regulated. Further details on this simulator are given by Bork (1983) and Henk (1989).

For this investigation the rainfall intensities varied in the range of 0.58 to 1.17 mm/min (cf. Table 2). In order to standardize the investigation, total irrigation time was at 180 min.

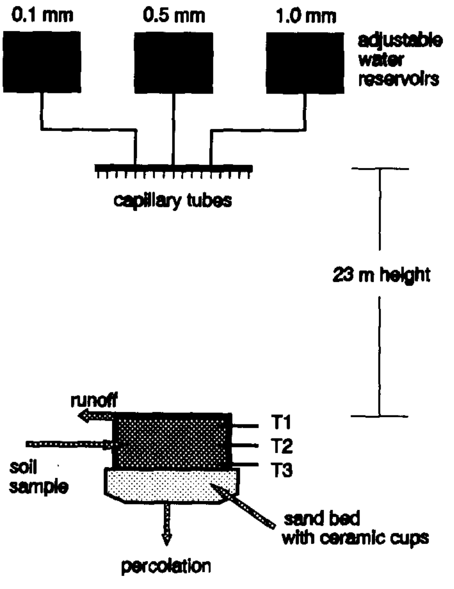


Fig. 2. Rainfall simulator used in this study.

### 3.3. Model description

The vertical water flux in the soil can be computed by the Richards' equation:

$$\frac{\partial \theta}{\partial t} = -\operatorname{div}(q) \tag{1}$$

where $q$ can be defined by Darcy's law

$$q = -K(\Theta)\ \operatorname{grad}(\psi - z) \tag{2}$$

in which $\theta$ = volumetric water content; $t$ = time; $\psi$ = soil matric potential; $K = K(\Theta)$ = hydraulic conductivity; $z$ = depth from surface; $q$ = water flux; $\Theta = (\theta - \theta_r)/(\theta_s - \theta_r)$ = relative saturation; $\theta_s$ = saturated water content; $\theta_r$ = residual water content.

The relationship between $\theta$ and $\psi$ is defined by a retention curve and can be characterized by various approaches. In this study, the often used function of van Genuchten (1980) was chosen:

$$\Theta = \begin{cases} \dfrac{1}{(1 + (\alpha|\psi|)^n)^m} & \text{for } \psi \leq 0 \\ 1 & \text{for } \psi > 0 \end{cases} \tag{3}$$

with $\alpha, n$ = curve parameter; $m = 1 - 1/n$.

According to van Genuchten (1980) the hydraulic conductivity can relate to the water content by using the model of Mualem (1976) who calculated the relative conductivity $K_r$ from the retention curve. This leads to the following expression:

$$
\begin{aligned}
K_r &= \Theta^{1/2}(1-(1-\Theta^{1/m})^m)^2 \\
K(\Theta) &= K_s K_r
\end{aligned}
\tag{4}
$$

in which $K_s$ = saturated hydraulic conductivity.

Analytical solutions of Eq. 1 are available only under certain initial and boundary conditions. For the calculation of the runoff-infiltration behaviour these analytical solutions are often used (e.g. Smith, 1983). For all other cases a numerical solution of Richards' equation has to be obtained. Although numerical methods are time consuming, especially when infiltration fronts have to be simulated, the method of Finite Differences was chosen in this study to calculate the infiltration rate. Because unsaturated as well as saturated domains occur during the simulation it was necessary to select a fully implicit approximation of Eq. 1. The numerical scheme could be found in various publications (e.g. Aboujaoudé et al., 1991) and, therefore, will not be presented here. For an accurate solution, a fine spatial discretization near the soil surface was considered indispensable. On the basis that fine spatial discretization leads to a fine temporal discretization, the model was made to adjust time steps automatically.

For calculating the infiltration rate at the soil surface, the following cases had to be distinguished:

$$
\begin{aligned}
q_{\mathrm{infil}} &= r - f && \text{for } \psi_n < \mathrm{d}z_{n+1}/2 \\
q_{\mathrm{infil}} &= K(\Theta)\left(\frac{\psi_n - \psi_{n-1}}{\mathrm{d}z_n} + 1\right) && \text{for } \psi_n \geq \mathrm{d}z_{n+1}/2
\end{aligned}
\tag{5}
$$

According to Eq. 5 the infiltration capacity equals the rainfall ($r$) minus the interception ($f$) until saturation occurs at the soil surface. Usually $f$ is the canopy interception but in this experiment $f$ is equal to the amount of water splashed off the soil probe. Assuming a constant gradient of unity the soil suction at the middle of the upper computational layer ($n$) equals half the depth of this layer at this time of ponding. In this case the boundary condition changes from a prescribed water flux (Neumann boundary condition) to a constant soil suction in the upper layer $\psi_n = \mathrm{d}z_{n+1}/2$ (Dirichlet boundary condition). In the latter case the infiltration rate and the runoff can be calculated from the solution of Eq. 1. The Dirichlet boundary condition will not change until the net rainfall ($r-f$) drops below the infiltration rate.

At the lower boundary, either continuously measured soil suction data, the measured depth of the groundwater table or the water flux in a certain depth has to be provided. To simulate this laboratory experiment, a constant soil suction was employed to characterize the lower boundary. In order to prevent an artificial numerical influence of the lower boundary condition on the calculated infiltration rate, the length of the simulated soil column was extended until the wetting front could not reach the bottom of the column during the simulation period.

Detailed descriptions of the water flux model are given by Rohdenburg et al. (1986) and Diekkrüger (1992).

Several approaches to describe the relationship between crust formation and kinetic energy of rainfall have been published (e.g. Morel-Seytoux, 1983; Ahuja and Ross, 1983). Most of them assume an exponential decay of crust conductivity. It has been recognized that, however, rainfall kinetic energy is not a sufficient measure of crust condition (Ahuja and Ross, 1983). Therefore, in addition to the energy, the intensity of the rainfall has to be considered. Using the concept of storm erosivity (kinetic energy times intensity), the decay of the crust conductivity can be described by the following relationship (Smith, 1985):

$$\frac{K_i}{K_s} = Cr + (1 - Cr)e^{(a\,EI)} \tag{6}$$

in which $EI$ = rainfall erosivity (kJ/m$^2$ mm/h); $Cr = K_{iu}/K_s$; $K_s$ = saturated hydraulic conductivity of the non crusted soil (cm/day); $K_{iu}$ = ultimate saturated conductivity of the crust (cm/day); $K_i$ = actual saturated conductivity (cm/day); $a = -0.075$.

According to Wischmeier and Smith (1978) the kinetic energy $E$ (kJ/m$^2$) of the rainfall can be calculated as follows:

$$\begin{aligned} E_i &= (11.89 + 8.73 \log I_i) N_i\, 10^{-3} && \text{for } 0.05 \le I_i < 76.2 \\ E_i &= 0 && \text{for } I_i < 0.05 \\ E_i &= 28.33\, N_i\, 10^{-3} && \text{for } I_i \ge 76.2 \end{aligned} \tag{7}$$

where $I_i$ = intensity of the rainfall (mm/h); $N_i$ = rainfall amount (mm); $I$ = intensity of rainfall (mm/h); i = period with constant rainfall.

The total energy of a storm is calculated by summing up the energy of all periods with constant rainfall. The erosivity $EI$ for a given rain storm equals the total storm energy $E$ times the maximum 30-min rainfall intensity $I_{30}$.

## 4. Results and discussion

### 4.1. Model application

It is not possible to measure the crust thickness at every place and to calculate a mean value over a certain area although a spatial variability of the crust thickness due to the microrelief can be expected. Aboujaoudé et al. (1991) compared a two-dimensional simulation of infiltration and runoff considering a spatial variable crust thickness with an one-dimensional approach assuming a constant crust thickness. They found a close comparison between both simulations and concluded that the one-dimensional approximation which integrates spatial variability at a small scale was useful. The question arises how the model results will be influenced by the crust thickness.

A numerical experiment was carried out in which a constant rainfall intensity of 0.833 mm/min during 180 min was assumed. In this experiment the crust thickness varied between 1 and 15 mm in steps of 1 mm and the $Cr$ value (ultimate crust conductivity/conductivity of the underlying soil) between 0.1 and 1 in steps of 0.1. The results are given in Fig. 3. The lowest relative runoff (runoff/precipitation) was obtained with a thin crust and a $Cr$ of 1 (non crusted condition), and the highest value with a low $Cr$ and a thick crust. It may thus be concluded from this figure that a relative runoff of 50% can be acquired with several combinations of $Cr$ values and crust thicknesses. Therefore, the runoff amount alone is not a sufficiently good index of crust development even though the conductivity of the under-

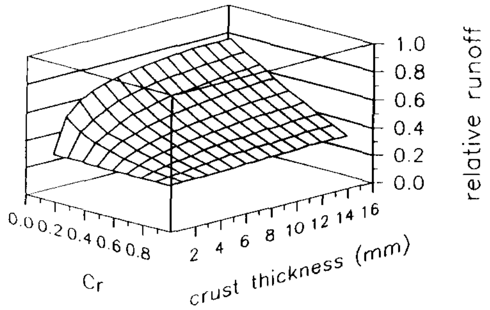


Fig. 3. Simulated relative runoff (runoff/precipitation) under variable crust thickness and *Cr* values (ultimate crust conductivity/conductivity of the underlying soil).

lying soil is known. In the model application employed in this study, the crust thickness was fixed to 15 mm although this value seems quite high. This value has been chosen because the upper tensiometer is situated in this depth and the *K*-values can be directly calculated from the flux measurements and the tensiometer data. The crust thickness has not been modified because it is necessary to standardize the simulation of all 22 plots otherwise the calibrated parameters can not be compared.

For the simulation of the experiments, time courses of the tensiometers, runoff rates and percolation rates were available. For the calculation of the infiltration rate, the splashed amount of water had to be subtracted from the precipitation. The soil properties were taken as constant below the crust. Because it was assumed that only the saturated hydraulic conductivity of the crust would change with time, the parameters of one retention curve, the saturated hydraulic conductivity and the *Cr* value were determined by calibration. The saturated and the residual water contents had been fixed in this calibration procedure because the initial soil suctions were moderately wet (about 20–50 hPa). In this case according to Eq. 3 the unsaturated pore volume ($\theta_s - \theta_i$) mainly depended on the bubbling pressure ($1/\alpha$). The remaining parameters for calibration were $\alpha$, $n$, $K_s$ and *Cr* (cf. Eqs. 3 and 6). The calibration was carried out by minimizing the deviation of the observed and simulated time courses of runoff and soil suctions. No automatic parameter estimation was employed in this calibration procedure although this would have given further information about the statistics of the estimated parameter. It was only for six of the investigated 22 soil samples that no good agreements were found between measured and calculated soil suctions. For the remaining 16 samples, a moderate to good comparison was obtained during the calibration procedure.

Based on two representative examples Fig. 4 shows the simulated and measured time courses for the soil suctions at depths of 1.5, 6.5 and 11 cm of plot M2, sample 3. Further, the observed and calculated cumulative runoff is given. The simulated runoff as well as the simulated soil suction at the depth of 1.5 and 6.5 cm compared well with the observations. As could be concluded from the soil suction at a depth of 11 cm, the experimental wetting front was faster than the simulated front. Nearly the same runoff could be simulated without crust. The results are also given in Fig. 4. The simulation was carried out by using the same retention curve, but by calibrating the conductivity of the whole soil sample. In the example

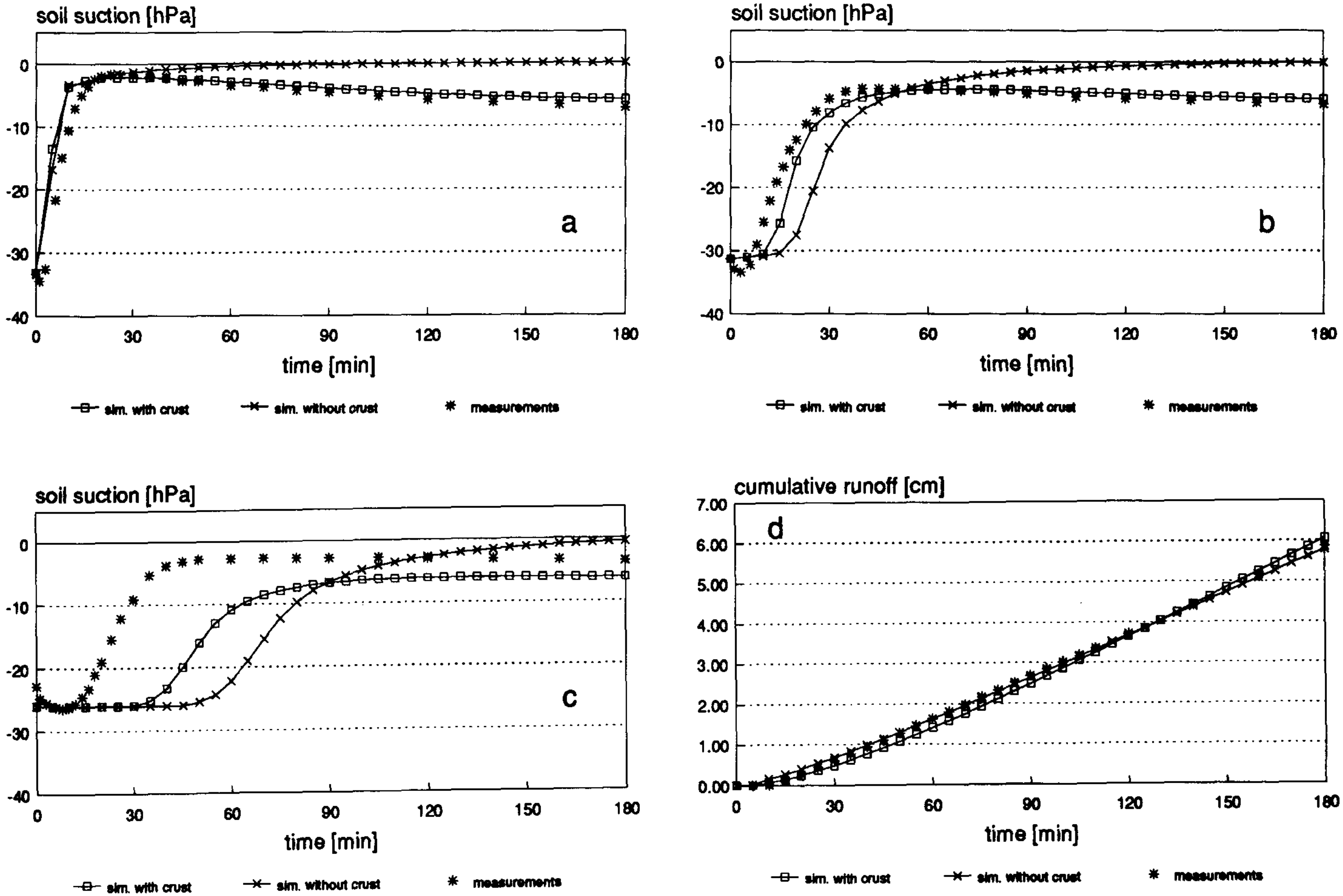


Fig. 4. Comparison of simulated and observed runoff (d) and soil suctions in the depths of 1.5 (a), 6.5 (b) and 11 (c) cm of plot M2 sample 3.

with crusting a saturated hydraulic conductivity of 30 cm/d for the underlying horizon was estimated while the conductivity of the crust was 3 cm/d. In the example without crusting the saturated hydraulic conductivity was 18 cm/d. From the runoff alone, the crust conductivity could not be estimated. The observed soil suction at depths of 1.5 and 6.5 cm increased after reaching a maximum value near saturation. This phenomenon, which is related to a decrease in infiltration after the development of the crust, was exactly reproduced by the simulation considering a crust. In this case, only unsaturated soil suctions were observed and simulated. Neglecting the crust, the simulated soil suction approached saturated conditions and did not increase later.

The crust development is given in Fig. 5. There, the ratio of crust conductivity to conductivity of the underlying soil is plotted against time. From this figure it can be seen that the increase in soil suction at a depth of 1.5 cm started at time when the crust conductivity sank to 50% of the conductivity of the underlying soil.

An example in which the effect of the crust development could not be found in the measurements is given in Fig. 6 (plot M2, sample 7). A good comparison between simulated and observed soil suctions was obtained independent of whether or not a crust is considered. The runoff was quite low compared with the example discussed previously. Only in the example with crusting ($K_s = 70$ cm/d, $Cr = 0.3$) runoff occurred. In the example without crusting ($K_s = 70$ cm/d) no runoff was observed. The total runoff was too low to compute a unique set of parameters. Therefore, no further attempt was made to find a better agreement between simulated and observed runoff.

As previously discussed, $\alpha$ and n of the retention curve were also calibrated. Applying Mualem's approach (Eq. 4) to the van Genuchten function, these parameters influenced also the unsaturated hydraulic conductivity of the soil. Because no further information was available (e.g. measured retention curves) it is interesting to look at the calibrated values for $\alpha$ and $n$. Fig. 7 shows the calibrated soil water retention curve, the unsaturated hydraulic conductivity of the crust and the underlying soil as well as the conductivities calculated directly from the measurements (plot M2, sample 3). The conductivities within the crust and below could easily be distinguished, and were confirmed by the simulation. The retention curve calibrated during the simulation looks like a curve common for sandy soils and

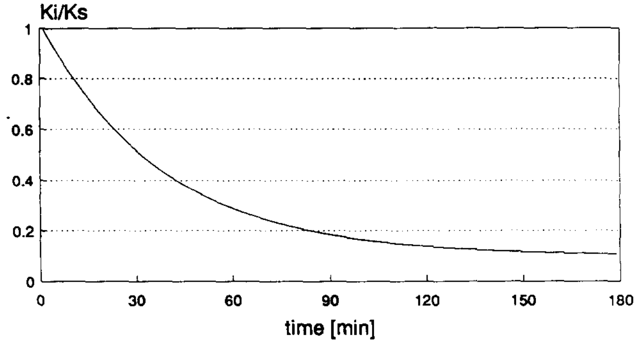


Fig. 5. Decrease of relative crust conductivity with time. Calibrated $Cr$ value = 0.1 (ultimate crust conductivity/conductivity of the underlying soil).

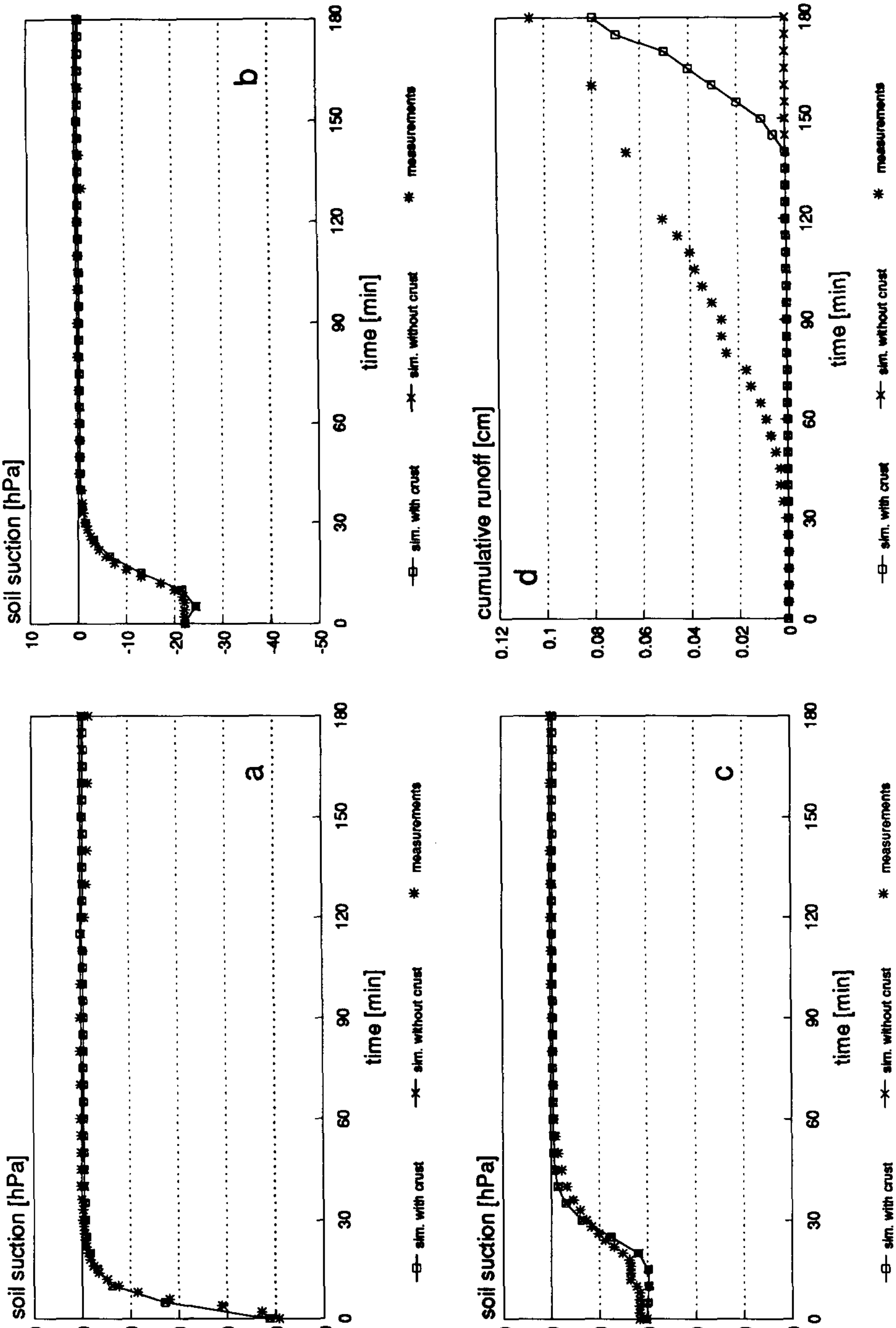


Fig. 6. Comparison of simulated and observed runoff (d) and soil suctions in the depths of 1.5 (a), 6.5 (b) and 11 cm (c) of plot M2 sample 7.

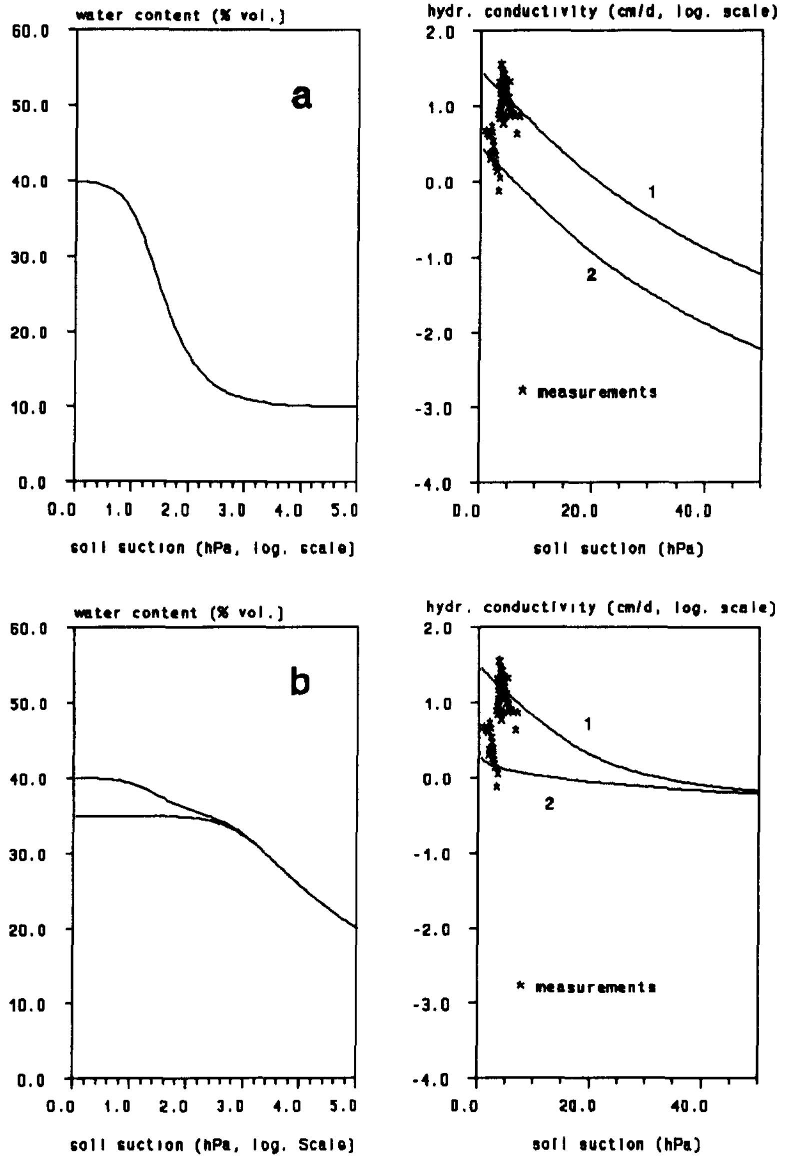


Fig. 7. Comparison of measurements with calibrated retention and conductivity curve using an unimodal (a) and a bimodal (b) approach. (1) conductivity of the underlying soil, (2) conductivity of the crust.

not like a curve supposed for this soil type. The steepness of the retention curve is in the expected range of a loamy soil, whereas the bubbling pressure ($1/\alpha$) corresponds to that of a sandy soil. The bubbling pressure was the most important factor for calculating the unsaturated hydraulic conductivities. A low bubbling pressure resulted in a steep conductivity curve as could be found in Fig. 7a. In nearly all of the 22 soil samples the calibrated

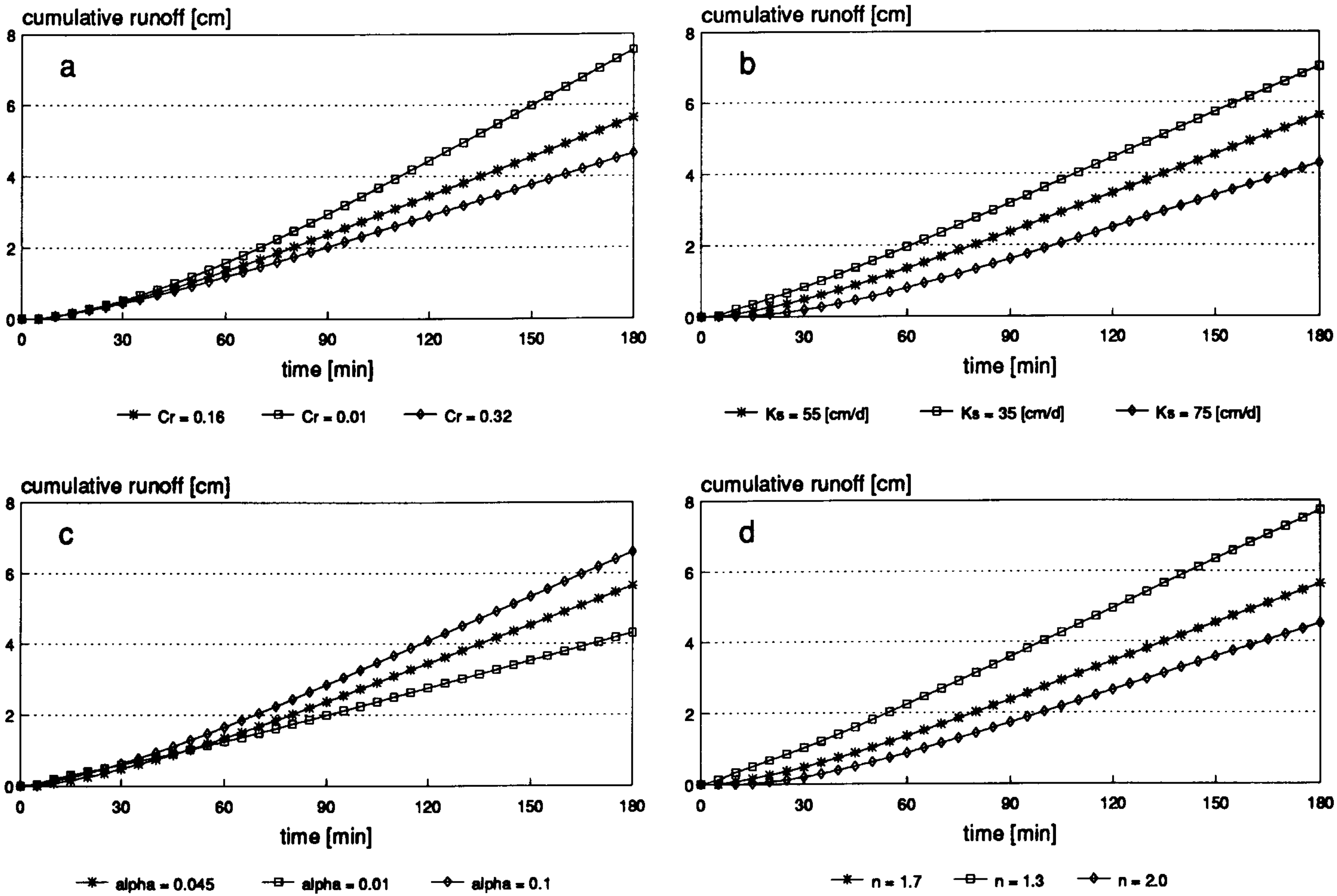


Fig. 8. Variation of runoff due to changes in model parameters. Parameter of the best fit simulation: $Cr = 0.16$, $K_s = 55$ (cm/d), $\alpha = 0.045$ (1/hPa), $n = 1.7$.

values for $\alpha$ were unrealistic high compared with values obtained from measurements of the retention curve.

It can be found in the literature that the description of the retention curve according to van Genuchten may fail near saturation (Durner, 1992; Othmer et al., 1991; Vogel and Cislerova, 1988). Durner (1992) and Diekkrüger (1992) showed that a bimodal frequency distribution of the pore size can not be adequately described by the approach of van Genuchten. Furthermore, they showed that bimodal porosity can be often observed in loamy and clayey soil due to tillage and biological activity. This behavior can be described by a superposition of two van Genuchten type curves (Diekkrüger, 1992) and can be seen in Fig. 7b. Then, the decrease of conductivity due to crusting can be explained by the clogging of larger medium-size pores with fine material. In this case the micro- and small medium-size pores are not blocked and, therefore, only the hydraulic conductivity near saturation is influenced by crusting. This hypothesis cannot be confirmed by our experiments because only a small range of the retention curve is covered. Fig. 7b shows that, assuming a bimodal porosity in which only the saturated hydraulic conductivity of the larger pores decreases, the measured data are described exactly as the unimodal approach. But the bimodal retention curve looks like a curve that is representative of loamy to clayey soils and not like a curve for sandy soils. This observation is confirmed by Othmer et al. (1991) and Diekkrüger (1992). They analyzed similar soils from the same catchment by comparing unimodal and bimodal descriptions of the retention curve. They found not only a closer agreement between measurements and fitted retention and conductivity curves, but also a significant difference between the simulation results using the unimodal and the bimodal approach.

### 4.2. Sensitivity analyses

A unique solution for the calibrated set of parameters could only be obtained when, in addition to runoff, soil suction data from different depths are available. To evaluate the model results, it is important to know the influence of the parameters on these results. This sensitivity analysis was carried out for all experiments. All sensitivity analyses showed the same trend. As an example, this is illustrated in Fig. 8 for the results of the simulation of plot M1 sample 2. Starting from the best fit, the parameters $\alpha$, $n$, $K_s$ and $Cr$ have been modified separately. The general trend is summarized in Table 3. From this, it may be concluded that a decrease in $K_s$ or $Cr$ can be compensated by an increase of $\alpha$.

Table 3
Changes of total runoff due to changes of model parameters

| Parameter | Parameter larger then in best fit simulation | Parameter smaller than in best fit simulation |
|---|---|---|
| $Cr$ | − | + |
| $K_s$ | − | + |
| $\alpha$ | + | − |
| $n$ | − | + |

+, runoff increases compared to best fit simulation.
−, runoff decreases compared to best fit simulation.

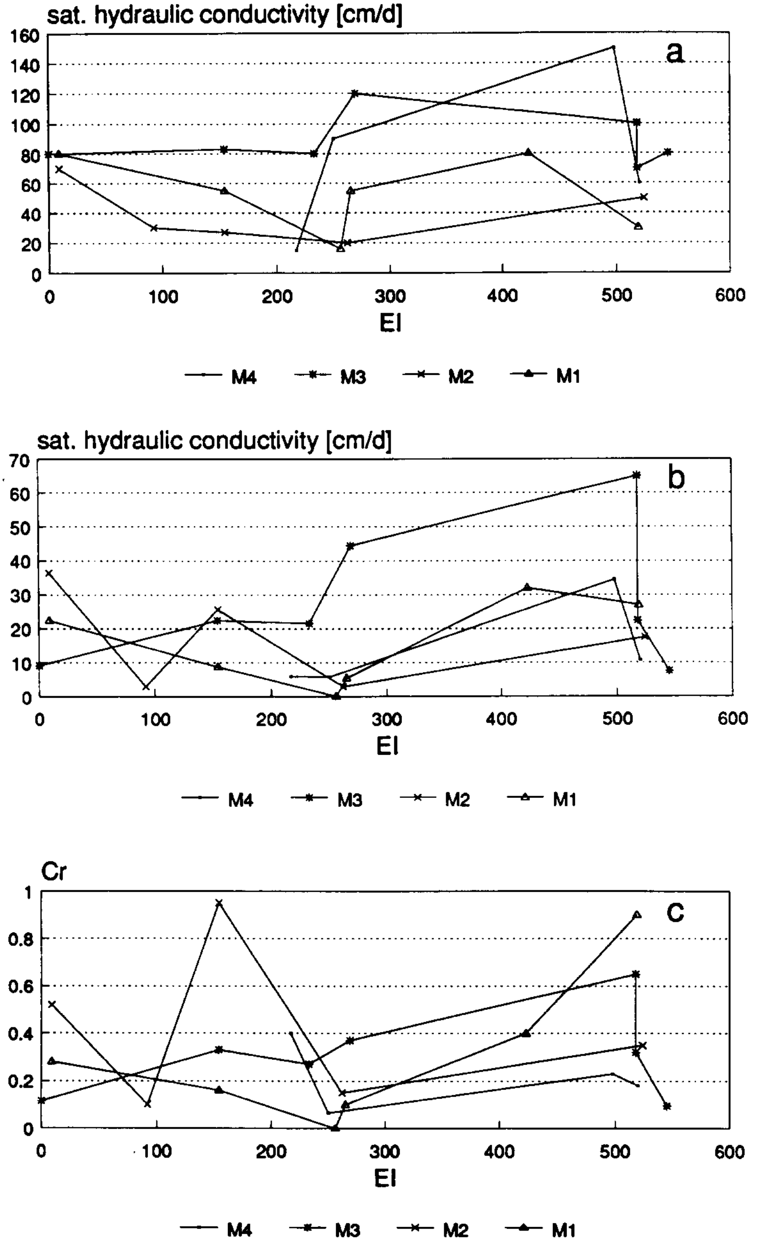


Fig. 9. Variation of $K_s$ of the underlying soil (a), $K_s$ of the crust (b) and $Cr$ (c) in dependence on the rainfall erosivity.

In order to evaluate the calibration results it was considered necessary to install an automatic parameter estimation procedure by linking a simulation model to a statistic package like BMDP (Dixon, 1981). In this case, the simulation model calculates the time courses of the water fluxes, and the statistic package computes from the deviation of the simulated with the measured time courses the next set of parameters. This will be repeated until the deviation between observed and calculated data is less than a given value. At the end of the iteration procedure the statistic package computes the co-variance matrix which is the basis for an evaluation of the results. This procedure will be applied to data from the present study in the next step.

### 4.3. Temporal variability

On every soil sample, a crust was established. Because the laboratory experiments were repeated several times during the year, in addition to the temporal small scale, the large scale temporal variability could be analyzed. The results are given in Fig. 9 in which the saturated hydraulic conductivity of the soil and of the crust, as well as that of the *Cr* value are plotted against the cumulative rainfall erosivity since tillage. The erosivity was chosen because the sampling date or the days after tillage is not a good measure of crusting due to different climate. As can be seen in this figure, the variability is large. Three different variabilities had to be compared: 1) saturated hydraulic conductivity of the underlying soil, 2) conductivity of the crust and 3) *Cr* values of the crust. The variability of the $K_s$ value of the soil was not necessarily due to temporal effects but might also be due to spatial variability which was quite high in these soils. In all three plots, no general trend could be observed. Most of the *Cr* values (16 of 22 samples) fell between 0.1 and 0.4 while, from Eq. 6, a range of 0.2 to 0.4 was expected. The mean *Cr* values over the whole period compared well for plot M1 (measured 0.19, expected 0.21), M2 (measured 0.28, expected 0.27) and M3 (measured 0.25, expected 0.20) when, for each plot, one extreme high value was neglected. The mean *Cr* value for plot M4 was lower than expected (0.22 compared with 0.37). Further analyses of these data (e.g. crop behaviour, plant cover, type of tillage) did not provide any clearer interpretation of the variability. Therefore, it is considered, that most of the variability was due to spatial instead of temporal variability. This conclusion is supported by the fact that the mean *Cr* values compare well with the expected value.

## 5. Conclusions

A simulation model for the calculation of runoff, infiltration and soil water fluxes was applied to simulate the infiltration-runoff behavior of 22 different soil samples taken from four plots over a vegetation period. Although the soil samples used had already developed a surface crust before they were precipitated in the laboratory, nevertheless, a temporal variation of the crust properties during the 3 hour irrigation could be determined by simulating the water fluxes although a crust was already established. This leads to the conclusion that the existence of the crust alone does not give a good measure of the infiltration capacity of the soil because due to biological activity and shrinking the crust may crack. In order to obtain a satisfactory initial condition for a simulation model it is necessary to estimate the destruction of the crust and subsequent re-establishment which occurs quickly.

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