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Runoff generation processes: results and analysis of field data collected at the East Central Supersite of the HAPEX-Sahel experiment

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Abstract

Within the scope of the HAPEX-Sahel experiment, the hydrological functioning of two small nested catchments was studied at two different scales: the plot scale (of the order of 100 m^2) and the catchment scale (0.2 km^2). At local scale, four runoff plots were set-up on the typical soil surface conditions observed on the catchments (plateau bare soil, two plots on fallow grassland) and an additional one was installed on a millet field. Soil moisture investigations at the plot scale have shown that infiltration was limited between 0.6 to 2 m deep on three sites, but was deeper than 3.4 m on the most pervious site (millet). The maximum water storage on all the sites was found to be reached at the maximum activity of the rainy season (late August), and not at the end of the season. During the dry months, the soil was fully dried off by evapotranspiration, resulting in the absence of inter-annual soil water storage. No influence of vegetation cover on runoff was observed on the fallow sites, but runoff generation was found to be very sensitive to tillage on the millet field. The parameter P_u, calculated from a rainfall hyetograph and defined as the rainfall depth that can actually produce runoff, is shown to be relevant to compute runoff on untilled soils, as it explains more than 87% of the variance in runoff depth. On tilled soils, it is necessary to take into account additionally the temporal evolution of the soil surface, especially the days after weeding operations. Simple linear relationships were derived to compute runoff depth from P_u on the plots for the most typical soil moisture conditions observed, and modified SCS equations have been derived for the catchments. Using the linear equations derived at the plot scale in a simple, empirical, semi-distributed model lead to formulate the assumption that the partial source area concept applied on the catchments. Analysis of discharge data at the catchment scale highlights that seepage through the bottom of a gully between two gauging stations leads to the abstraction of non negligible volumes of water.

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Moreover, the water totally infiltrates in a spreading zone downstream from the outlet of the largest catchment showing that discontinuities occur in the surface water transmission within a catena. Such discontinuities constitute a major problem for the concern of aggregation of hydrologic processes.

1. Introduction

In semi-arid zones runoff from hillslopes occurs in the form of Hortonian overland flow (Dunne, 1978). The overland flow yield is strongly non uniform in space, due to high spatial variability in infiltration capacities of the soil surface. Only small areas of the watersheds contribute to storm runoff to the channel and thus, Yair and Lavee (1990) have suggested that runoff generation in semi-arid areas was following a pattern of partial area contribution similar to that identified for humid areas.

Previous studies in the south-western United States conducted at the runoff plot scale with rainfall simulators have shown significant differences in plot hydrologic responses for various vegetation cover (Kindcaid et al., 1964; Lane et al., 1987), and it was concluded that runoff was negatively related with vegetation density. Others (Blackburn, 1975; Tromble et al., 1974) found rock cover and erosion pavement to be positively related to runoff. Studies in Israel (Morin and Benyamini, 1977; Yair et al., 1980), in Mali (Hoogmoed and Stroosnijder, 1984), and in Tunisia (Berndtsson and Larson, 1987) have shown that, apart from the rain characteristics, the surface conditions, such as soil crusts, rock pavement, vegetation cover or geomorphological situation play a major role in runoff production.

In various French-speaking African countries, numerous studies have been undertaken during the last ten years, peculiarly by ORSTOM[1] researchers, to assess the conditions of runoff production as influenced by surface crusts, using rainfall simulation (Valentin, 1981; Albergel, 1987). As a synthesis of these experiments, the hydrological properties of various tropical soil surface types have been presented by Casenave and Valentin (1992) who proposed a classification of the soil surface runoff capability based on a concept of surface features that includes the description of the type of surface crust, the vegetation cover if any, and the soil. Their results confirm the effect of soil crusting in increasing runoff, and highlights the role of vegetation as a runoff limiting factor and improvement of infiltration capacities due to tillage or faunal activity. Therefore, the concept of surface feature reveals itself to be a useful tool for the description of the spatial variability of runoff production over an hillslope.

In the framework of the HAPEX-Sahel experiment (Goutorbe et al., 1994), the main objective of which was to estimate the atmospheric water fluxes at the scale of a 1 × 1° square, it was of concern to quantify the surface runoff and to infer the redistribution of water at large scale. For that purpose, investigations at different scales have been undertaken. A pool network was surveyed on a 600 km^2 study area to assess the role of the pools in the hydrology of the region (Desconnets et al., 1997). Together with regional groundwater studies (e.g. Leduc et al., 1997), it was shown that they constitute a major contributor to the recharge of the upper aquifer.

[1] ORSTOM is the French Overseas Scientific Research Institute.

Concurrently to these large scale studies, hydrological monitoring has been carried out at local scale from 1992 to 1993 on runoff plots (100 to 130 m^2) and small catchments (0.1 to 6 km^2) located in the East Central Supersite (ECSS). The main outcomes are presented in this paper, the specific purpose of which is to give quantitative results on runoff production capacities at both plot and catchment scale under natural rainfall for the main surface features identified and mapped over the ECSS. The effects of soil moisture, soil surface crusting and vegetation cover on runoff are also discussed. The paper focuses on two of the smallest catchments. The hydrological measurements at the catchment scale have led to an understanding of the role of surface runoff in the redistribution of precipitation and its transmission through the drainage network.

2. Study area

The research catchments are located on the East Central Supersite, about 70 km east from Niamey, in the western part of Niger (Fig. 1). The ECSS is characterised by a gentle relief, with elevations ranging from 200 to 260 m above sea level. The landscape is composed of dissected laterite-capped plateaus, with frequently steep edge slopes that dominate wide topographic depressions located about thirty meters below. The main feature of the hydrography of the region is endoreism (Fig. 1), as induced by the gentle slopes, the lack of well developed drainage networks and the small size of the catchments (a few square kilometres for the largest). There is no baseflow: the runoff is intermittent and ends with the rainfall. The floods flowing away from the plateaus either spread into flat sandy areas or reach the bottom of the depressions where the water accumulates in small pools. A detailed description of the pool hydrology over the ECSS is presented in Desconnets et al. (1997).

Precipitation exhibits an irregular distribution over the year: large storms occurring from early May to late September provide 95% of the annual rainfall depth, while the seven other months are dry. The mean annual rainfall for that region is 560 mm, calculated from 1905 to 1989 (Lebel and Le Barbé, 1997). The storms are mostly convective, either from isolated cumulo-nimbus or from organised cloud formations often evolving in the form of squall lines (Lebel et al., 1992). The typical hyetograph generated by a squall line is characterised by a front with high intensities, and a tail of longer duration and lower intensities. Rainfall is highly variable in space. In this region the median rain rate is 35 mm h^{-1} and 35% of the rain falls with an intensity greater than 50 mm h^{-1} (Lebel et al., 1997).

The experimental site is located on a catena including two main geomorphological units: the plateau with loamy-clayey soils (30% clay) and low slope (0–0.5% slopes; 4–8% for the plateau slope), and a 2 km long sandy hillslope (80% sand), leaning against the plateau, and extending down to a valley (2%–4% slopes) wherein is situated the Banizoumbou village (Fig. 1). The drainage network is composed of gullies that concentrate the water running away from the plateau border and the sandy hillslope.

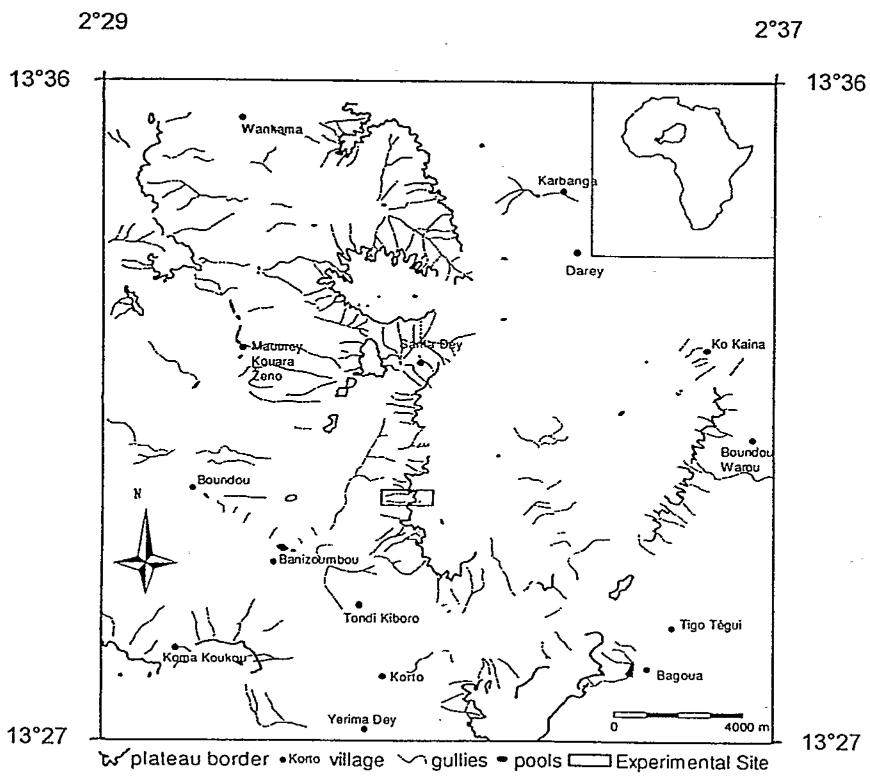

Fig. 1. Location of the experimental site (rectangle) and the Banizoumbou village within the East Central Supersite of HAPEX-Sahel.

Five major surface features have been identified over the catchments (Table 1 and Fig. 2), using both the methodology and the crust typology defined by Casenave and Valentin (1992): bare surface, vegetation strips and slope on the plateaus, piedmont bare surface and fallow on the sandy hillslope. In addition, the cultivated millet field that was not observed on the catchments but in their vicinity, was also surveyed. The

Table 1

Percentage of the total area covered by each surface feature on the catchments

Surface feature	Catchment 1	Catchment 2
Plateau bare surface	27.6	13.5
Plateau vegetation strip	9.5	4.4
Plateau slope	28.7	17.3
Piedmont bare surface	31.6	28.1
Fallow bush/grassland	2.6	36.6
Area (km^2)	0.0468	0.0995

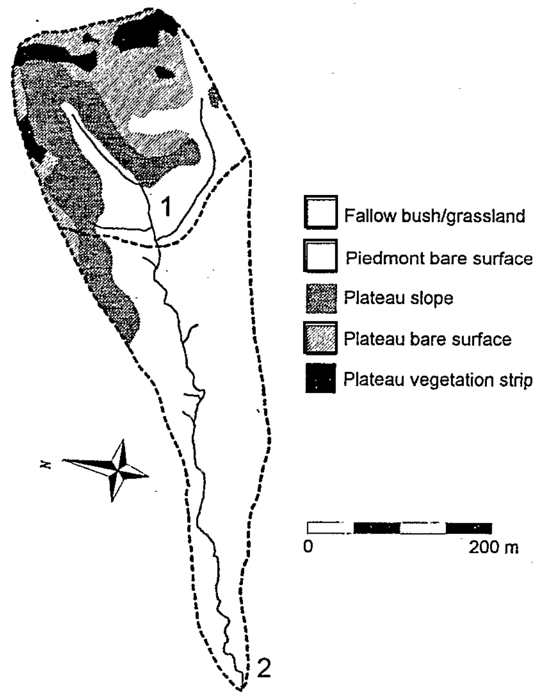

Fig. 2. Soil surface feature map of the two catchments (dashed line) defined by gauging stations 1 and 2, set up along the main gully (solid line).

plateau exhibits an alternating pattern of bare soil areas and vegetation strips roughly running along the contour lines (Ambouta, 1984; Thiéry et al., 1994). This natural system is usually called "tiger bush", due to its tiger skin aspect on aerial photographs (Clos-Arceduc, 1956). The first surface feature (plateau bare soil) is characterised by a typical succession of crusts with, from upslope to downslope, structural crusts (Valentin and Bresson, 1992), erosion crusts with patches of gravel crusts, runoff depositional crusts, and sedimentary crusts partially covered with herbaceous vegetation during the rainy season. The plateau areas enclosed in the catchments are located to a great extent at the edge of the plateau where the last two crusts types are seldom observed. The vegetation strips present a dense canopy cover (80 to 100%) and the soil surface is generally colonised by cryptogamic vegetation (cryptogamic crust). The abundance of litter induces important termite activity, increasing soil porosity and infiltration (Chase and Boudouresque, 1989). The plateau slope is covered with gravel crusts (free laterite pebbles) overlying a highly weathered substratum of loamy sediments, locally outcropping. On the sandy hillslope, the piedmont bare surfaces are covered with erosion crusts, scattered with some sandy or gravely micro-mounds. The fallow areas are a mosaic of two main surface features: small topographic depressions (1 to 20 m^2) without vegetation covered by erosion and structural crusts, scattered in wider areas (vegetation cover ranging from 25 to 40%) with structural and erosion crusts colonised by algae that stabilise the surface. The millet field surface, tilled traditionally, is covered with sandy structural crusts.

3. Experimental design

At local scale, three runoff plots were set up to provide discharge measurements. They are composed of a rectangular area of natural soil surrounded by an iron border, driven 10 cm into the soil and equipped downstream with collector tank for water volume measurements. A ten-fold divider mounted on the first 1 m^3 tank allows excess volume to flow to a second tank. Water volume in the tanks was measured after each rainstorm. Each plot is assumed to be representative of the hydrologic behaviour of a particular surface feature: plateau bare surface (plot area: 26 m by 5 m), fallow grassland and millet field (plot area 20 m by 5 m). The two latter plots were set up on the same geomorphological unit (sandy hillslope), with, however, different soil usage: traditional tillage on the millet and no human intervention on the fallow. Additionally, a 43.4 m^2 natural micro-plot (without any artificial border) equipped with the same type of water collector enclosed two patterns observed on the fallow soil: bare topographic depressions with erosion crust and sandy encrusted area. The bare soil, fallow and micro-plot were monitored from early July 1992 to the end of the 1993 rainy season, whereas only the 1993 data set is available on the millet site.

The two small nested catchments (0.047 and 0.099 km^2), located predominantly on the sandy hillslope but also included part of the plateau, were monitored from 1992 to 1993. The outlet of the larger basin lies about 700 m from the edge of the plateau, a few dozen meters upstream from a flat zone (less than 2% slope) where the sandy sediments carried by the floods accumulate. This sediment spreading zone, located in the upper part of the sandy hillslope, is probably also an infiltration zone for floods running out in the gully. This assumption will be investigated by the measurements. Discharges were measured from two gauging stations (Fig. 2) equipped with Parshall-type flumes and electronic water stage recorders. Several discharge measurements with current meters were performed during the floods to check the theoretical rating curve of each flume.

Rainfall was measured with one tipping bucket recording raingauge with 400 cm^2 cone diameter and five raingauges. Each tip, corresponding to 0.5 mm of rain depth, was recorded to the nearest second on an EPROM cartridge. The raingauges were located near the runoff plots and were read after each rainstorm.

An extensive neutron probe access tube network provided soil moisture data. Two 3.4 m deep tubes were located on each runoff plot and seven similar tubes were placed along a transect on a sandy hillslope in the vicinity of the catchments. A first measurement was performed as soon as possible after the end of each rainstorm (typically five to ten hours) and following measurements were taken one, two, and four days later. After four days, without rainfall, measurement spacing was increased to one week. General planning of soil moisture investigations during the HAPEX-Sahel experiment are presented in a companion paper by Cuenca et al. (1997). The soil moisture measurements began mid-July 1992. As all the access tubes were not set up, the very beginning of the 1992 season was missed.

In order to get a better knowledge of the infiltration and runoff processes related to soil crusts, hydraulic conductivity and sorptivity of different types of crusts were measured using TRIMS infiltrometers. The detailed methodology is presented by Vandervaere et al. (1997) but we underline here the major outcomes.

4. Results

There were no significant differences in the seasonal total rainfall recorded with the six raingauges of the network, but at the event time step, a high spatial variability of rain depth was observed, particularly for the heaviest storms: differences of more than 25% have been recorded from one raingauge to another e.g. 56.2 mm on the plateau bare soil versus 40.2 mm on the fallow on the 13 September 1993. A total of 35 and 29 rainstorm events were observed in 1992 and 1993, respectively, with a maximum depth of 52 mm and 44.5 mm recorded with the tipping bucket raingauge. Moreover, the distribution of intensities is also highly variable from one storm to another (Lebel et al., 1997).

The total amount of rainfall observed in 1992 and 1993 was respectively 430 and 470 mm. The seasonal rainfall depth was thus not very different from one year to another, but the temporal distribution of storms within each season presented various patterns. In 1992, the storms were rather regularly distributed in time, with a maximum occurrence on

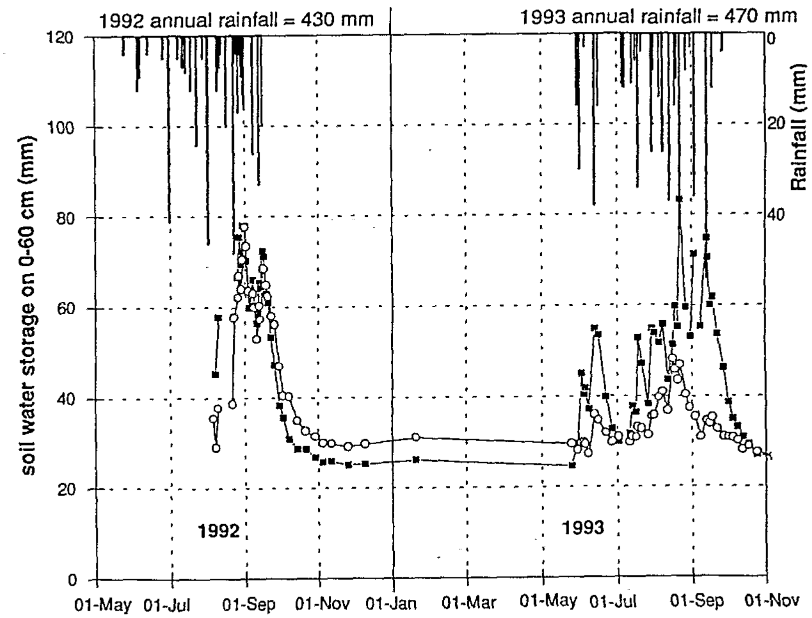

Fig. 3. Time distribution of storms (upper x-axis) and time variations of soil water storage during 1992 and 1993 rainy seasons for millet (closed squares) and fallow (open circles).

late August (a 110 mm rain depth, that is about one fourth of the annual depth, was observed in ten days). In 1993, the rainy season had a bimodal distribution (Fig. 3) with a 105 mm rain depth precipitated in early June, followed by a twenty-day dry period. Both years, the maximum ten-day rainfall depth was observed by the end of August.

In order to allow for the comparison of the different rainfall–runoff events observed, it is thus necessary to take into account the shape of the storm and the intensities that have occurred. For that purpose, we used the concept of "Pluie utile" (P_u), defined as the depth of rain precipitated at intensities higher than a threshold value. P_u is obtained by subtracting from the total rain the amount that, on average, will not produce runoff because of too low intensities. For a given rainstorm, practical calculation of P_u is performed by removing from the rainfall hyetograph the five-minute rain depths lower than the intensity threshold. This calculation method derives from that proposed by Lamachère (1994) and differs from other concepts used in the past, such as CAM (Mean Absorption Capacity, Roche, 1963) or ∅-index (Morel-Seytoux, 1981). As a matter of fact, only the part of a storm that totally infiltrate is neglected in P_u. The infiltration flux observed when runoff occurs is accounted for in the parameter P_u, which is thus higher than the net rainfall depth, defined as the depth of water that actually runs off the surface (equivalent to the runoff depth). The threshold value is determined from experimental data, and equals the maximal five-minute rain depth below which runoff was never observed on each runoff plot. It is dependent on the surface features, and was therefore calculated separately for each plot. The threshold values, presented in Table 2, were used to calculate P_u for each observed rainfall event. The whole methodology is detailed in Peugeot (1995). A specific study have shown that P_u was very little sensitive to the threshold value. As a matter of fact, for most rainfall events, P_u takes into account the convective front of the storm with high intensities, and the tail that, on average, does not generate runoff is neglected. However, when the storm shape is not well structured and when the distribution of intensities are more erratic, P_u becomes sensitive to the threshold value, particularly when the intensities vary around the value of the threshold, as observed for example on 14 September 1992.

At the square meter scale, hydrodynamic properties of both the crust and the subsoil were measured by tension infiltrometry (Vandervaere et al., 1997) showing significantly lower saturated conductivities (K_s) of the crusts as compared to K_s of the subsoil. The measured *crust conductivities* (Table 3) are consistent with the observations made by Casenave and Valentin (1992) who computed, for different types of crusts on 1 m^2 plots under simulated rainfall, the critical value of rainfall intensity (I_c) below which no

Table 2
Intensity thresholds used to compute P_u, and defined by the maximal five-minute rain depth (converted in mm h^{-1}) below which runoff was never observed on the plots and the catchments

Site	Intensity threshold (mm h^{-1})
Bare surface on plateau	7
Fallow micro plot	7
Fallow grassland	18
Millet field	18
Catchment 1 and 2	12

Table 3
Measured values of saturated hydraulic conductivity (K_s, Vandervaere et al., 1997) and critical intensity value (I_c, Casenave and Valentin, 1992) for the main crusts type observed on the plots and the catchments (STR: structural crusts; SED: sedimentation crusts; ERO: erosion crusts) and for the sandy and clayey subsoils

	Plateau bare soil			Fallow		Millet
	STR	SED	Subsoil	ERO	Subsoil	Subsoil
K_s (mm s^{-1})	8.5 × 10 − 4	5.2 × 10 − 4	2.8 × 10 − 3	19 × 10 − 4 [a]	5.4 × 10 − 2 [a]	1.7 × 10 − 2 [a]
I_c (mm s^{-1})	10 × 10 − 4	6 × 10 − 4	− (+)	8 × 10 − 4	− (+)	0.3 × 10 − 2

[a] J.P. Vandervaere, personal communication, 1994.

(+) Due to the continuous soil crusting during the rainfall simulations, the value of I_c for the underlying soil, that is the uncrusted soil, is not available.

runoff occurred on wet soils. The I_c values measured *for the crusts, which were shown to* be close to the saturated conductivity (Valentin, 1981), are of the same order of magnitude as the K_s values presented by Vandervaere et al.

Soil water storage was found to be very sensitive to rainfall, increasing after each storm and rapidly decreasing during dry periods. As an example, Fig. 3 shows the soil water storage in the upper 60 cm soil layer, W_{60}, measured for the millet and the fallow plots in 1992 and 1993. This depth was found to allow for the best characterisation of the temporal evolution of soil moisture on the different plots. The evolution of W_{60} is similar for the two plots in 1992. The water content had a sudden increase on August 21st, and reached its maximum on the 31st. After the last rain (September, 15th), W_{60} was divided by two in about 15 days and reached its minimum value on May 26th, 1993, before the first event of the 1993 rainy season. Whatever the maximum value reached, the 60 cm upper layer of soil was fully dried off during the dry season. For all of the monitored access tubes, this minimum value was steady over years. In 1993, the behaviours of the two plots were different. The soil water storage of the millet field reached the 1992 level, but the water content of the fallow area did not exceed 50 mm, although the 1993 total amount of rainfall was higher than in 1992. The same results were observed on the micro-plot and the plateau bare soil. Different was the millet plot behaviour, which highly benefitted of the rainstorms of the end of August, due to the preceding tillage on August 20th, 1993. On all the sites, the maximum value of surface water storage W_{60} was not observed after the last storm of a season, but on late August, when precipitation was more frequent.

In order to get a better understanding of the infiltration capacities, Fig. 4 presents the 1992 annual range of water content measured on the four plots (solid line). This range is defined as the envelope of all the measured soil moisture profiles, and therefore does not correspond to an actually measured profile. The water content varies between 0 and 20% on all sites. The graphs allow for the estimation of the maximum wetted depth (wetting front). The available water decreases with depth and the wetting front does not exceed 50 cm on plateau bare soil, 120 cm on micro-plot, 200 cm on fallow, but is deeper than 340 cm (access tube length) on millet field. Not surprisingly, the minimum soil moisture values were lower on sandy soils (fallow, millet, micro-plot) than on loamy-clayey soils of plateau. As depth increases, the soil moisture range narrows but the minimum and maximum envelope do not merge. The deviation observed under the

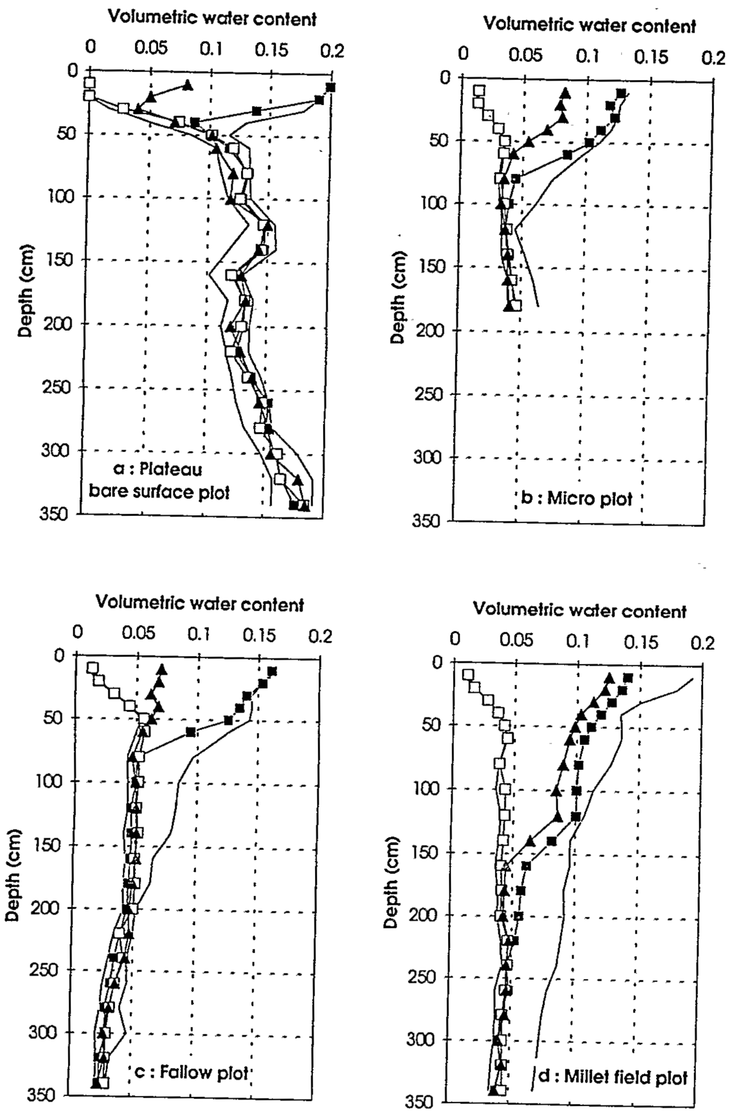

Fig. 4. Comparison of soil moisture profiles for the four runoff plots. Annual range of soil water content (solid line) and three soil moisture profiles measured before rainfall: 29 August 1992 (closed squares), 26 May 1993 (open squares), and 20 August 1993 (triangles).

wetted depth on Fig. 4a, b and c, estimated to about 1.8%, corresponds to the range of measurements errors. On the millet plot, (Fig. 4d), this range can not be estimated due to deep infiltration.

In runoff generation processes, determining is initial water content, that is soil moisture before the beginning of any rainfall. Three particular profiles were therefore plotted on Fig. 4, corresponding to the minimum (May 26th, 1993) and the maximum values of W_{60} for each year (August 29th, 1992; August 20th, 1993) measured before a storm on the four runoff plots. The graphs clearly show that, on all the plots, the dry profile (August 26th, 1993) merges with the minimum envelope of soil moisture if one considers the errors related to measurements. It is thus showed that the soil was at the minimum water content at the beginning of the 1993 rainy season, not only on the 60 cm top layer, but on the entire tube depth. In the first 60 cm of the soil the 1992 wet profile (August 29th) is very close to the maximum values observed, except for the millet plot (Fig. 4d). On this latter site, the maximum envelope corresponds to measurements performed after the first rainstorm following weeding operations, on August 22th, 1993, and profiles on August 29th, 1992 or August 20th, 1993 were clearly lower than the envelope. The maximum surface water content was reached before the outstanding events on late August on the untilled sites. In 1993 the wettest profile (August 20th) is clearly lower than in 1992 for all the untilled sites, but only slightly lower on the millet site. The lower infiltration observed in 1993 is attributed to the irregular distribution of the storms over the season. The early June rain depth was rapidly driven back to the atmosphere by evapotranspiration during the following dry period. Later storms occurred on dry soils, with low hydraulic conductivity, and infiltration was thus reduced.

The consequences of infiltration capacities of untilled soils on runoff are shown on Figs. 5–7. They present the relationships L_r (runoff depth in mm) versus the above-defined parameter P_u for the micro-plot, the fallow and the bare soil plots. The points are

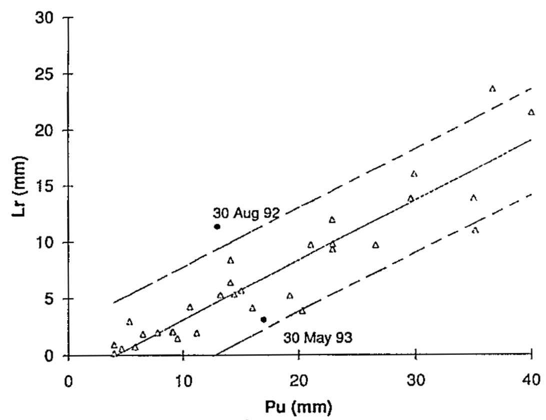

Fig. 5. Lr versus Pu for the micro-plot under typical (triangles) or extreme (circles) soil moisture conditions: 30 August 1992 (very wet) and 30 May 1993 (very dry). The solid line refers to the model Lr(Pu) fitted to the data under typical conditions and dashed lines represent the 95% confidence interval.

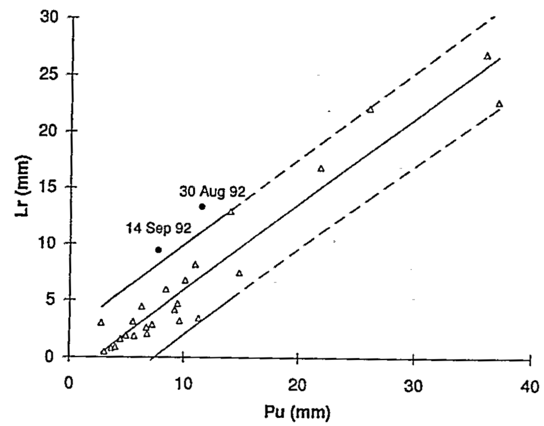

Fig. 6. L_r versus Pu for the plateau bare surface runoff plot under typical (triangles) or very wet (circles) soil moisture conditions on 30 August and 14 September 1992. The solid line refers to the model $L_r(P_u)$ fitted to the data under typical conditions and dashed lines delimit the 95% confidence interval.

moderately scattered around a straight line, except a few one that correspond to the storms that have occurred under extreme soil moisture conditions (August 29th, 1992 and May 26th, 1993). Due to missing data in the data sets, every "very wet" or "very dry" day does not necessarily appear in all the graphs. The outstanding event on September 14th, 1992 on Fig. 6 is not related to particular soil moisture conditions. As the storm hyetograph was not well structured, P_u was very sensitive to the threshold value, resulting in uncertainty in estimating the P_u value. This particular event will not be considered in the study. In the following, "typical conditions" will refer to initial soil moisture content the most frequently observed, that is, neither very wet nor very dry. On very dry soils, runoff

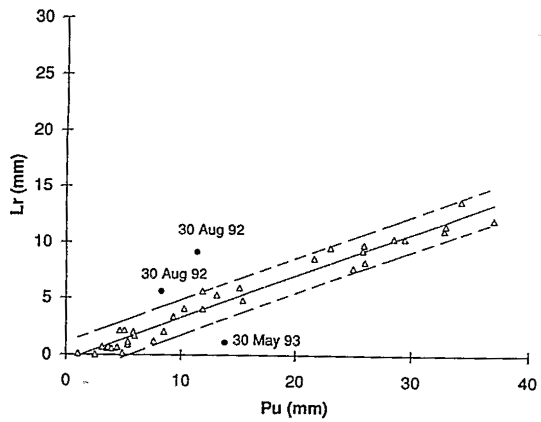

Fig. 7. L_r versus P_u for the fallow runoff plot under typical (triangles) or extreme (circles) soil moisture conditions: two storms on 30 August 1992 (very wet), 30 May 1993 (very dry). The solid line refers to the model $L_r(P_u)$ fitted to the data under typical conditions and dashed lines delimit the 95% confidence interval.

Table 4
Simplified models $L_r(P_u)$, with correlation coefficient, sample size and standard error, corresponding values of soil moisture defining typical initial conditions, and seasonal mean runoff coefficient (Cr) and maximum Cr value for the observed events

Site	Model	Typical conditions: mean soil moisture on a 60 cm deep profile	Runoff coefficient	
			mean	max
bare soil	$L_r = 0.77\,P_u - 1.7$, $r^2 = 0.92$, $n = 25$, $\sigma_r = 2.11$	7–12%	0.47	0.76
micro-plot	$L_r = 0.53\,P_u - 2.4$, $r^2 = 0.87$, $n = 28$, $\sigma_r = 2.29$	3–10%	0.33	0.90
fallow	$L_r = 0.38\,P_u - 0.5$, $r^2 = 0.97$, $n = 35$, $\sigma_r = 0.79$	5–12%	0.23	0.66
millet: stable surface	$L_r = 0.26\,P_u - 0.4$, $r^2 = 0.81$, $n = 16$, $\sigma_r = 1.61$	4–12%	0.17	0.41
millet: evolving surface	$L_r = 0.06\,P_u - 0.2$, $r^2 = 0.53$, $n = 7$, $\sigma_r = 0.64$	4–12%	0.05	0.10
catchment 1	$L_r = P_u^2 / (P_u + 40.6)$, $r^2 = 0.96$, $n = 38$, $\sigma_r = 1.19$ [a]		0.24	0.59
catchment 2	$L_r = P_u^2 / (P_u + 39.1)$, $r^2 = 0.95$, $n = 38$, $\sigma_r = 1.35$ [a]		0.22	0.52

[a] Correlation coefficient and standard error of the relationships L_r computed $= f(L_r$ measured).

production was lower than runoff yield observed under typical conditions, e.g. on Fig. 5 and Fig. 6 the first event of the 1993 season (30 May). Wet initial conditions lead to higher runoff production, e.g. the two storms on 30 August 1992 (Fig. 7) or the second storm on 30 August (Fig. 5 and Fig. 6), that occurred after a succession of heavy rainfalls recorded the days before.

Initial soil moisture at a given time was computed from soil moisture measurements as the mean water content of the upper 60 cm soil layer. The range of soil moisture values defining typical conditions is presented (Table 4). For those typical conditions, a linear relationship $L_r = f(P_u)$ can be fitted to experimental data for each runoff plot on untilled soil. The linear correlation coefficient, varying from 0.87 to 0.93 (Table 4), shows that the runoff depth variance is explained essentially by P_u. The points corresponding to typical conditions are situated inside the 95% confidence interval for the bare soil and the fallow site, but they are a bit more scattered for the micro-plot. This plot is not artificially delimited by a border and its area, albeit precisely determined from a detailed topographic survey, may vary depending on the characteristics of the storms. That introduces a variability that is not represented by P_u and could explain the scattering for heavy storms (Fig. 5) and thus the lower correlation coefficient.

The millet plot exhibits a highly variable runoff production in time. In order to characterise this variability, the cumulative runoff (sum of runoff depth for all the previous events) is plotted versus cumulative P_u (sum of P_u for all the previous events) on Fig. 8 for the 1993 rainy season. The graph shows a steady trend with two sections of lower slope related to rainfall events that occurred after the first and the second weeding, in early July and late August 1993. The sections with a smoother slope are persistent of 60 mm in units of P_u, which corresponds to about 80 mm of natural rainfall depth. The data set was then split into two categories: rainstorms occurring on stabilised surface features where the crusts were established, and rainstorms that occur during the restoration phase of the crusts. For both categories, a linear expression $L_r(P_u)$ was fitted to non cumulative data (Table 4). The results are satisfactory for the data corresponding to established crust ($r^2 = 0.81$), but worse for the second category ($r^2 = 0.53$). By splitting the runoff events into two categories with respect to the date of the last weeding, we made the assumption that the functioning was homogeneous within each category. Indeed, processes are progressive and the crusts evolve over time. The variability of runoff yield related to the evolution of surface features within the second category is not taken into account and induces a low correlation coefficient.

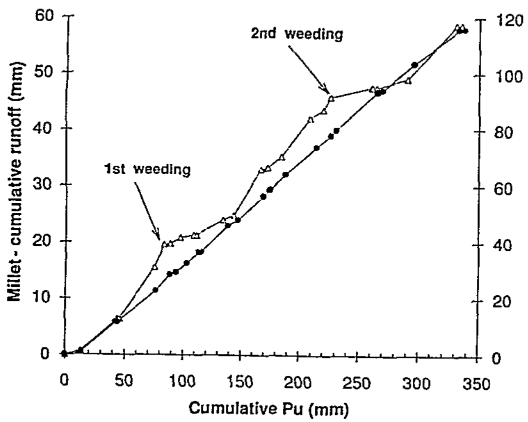

Fig. 8. Cumulative runoff versus cumulative P_u for the fallow bush (circles) and the millet field plot (triangles): sections with lower slope beginning after the weedings operations correspond to the progressive restoration of the crusts under the consecutive rainfalls.

In the same manner, the fallow cumulative runoff data were plotted on Fig. 8, showing a steady increase of cumulative runoff. By chance, the slope of cumulative L_r versus cumulative P_u for the fallow site is roughly twice that for the millet site. For both sites the runoff events occurred under the same rainfall conditions and for similar initial soil moistures, the differences in behaviour can be related to the contrasted temporal evolution of surface features. On the millet site, surface crusts were destroyed twice by the weeding tool and were gradually restored by successive rainfalls. On the fallow site, several surface feature mapping surveys have shown that one cause of surface feature temporal variability was the growth of annual herbaceous vegetation (essentially *Zornia gloschidiata*) after mid-July. As shown on Fig. 8, a decrease of runoff due to increase of vegetation can not be obviously observed. The sparse vegetation cover (estimated to 35–40%) probably does not significantly modify the hydrodynamic properties of soil surface. Therefore, the evolution of the fallow surface feature is not likely to influence runoff yield. A similar result is observed on the micro-plot.

The mean seasonal values of the runoff coefficient Cr (volume of total runoff/volume of total rainfall over a season) observed during the two rainy seasons (Table 4) allows the runoff capability of the different sites to be compared. At the plot scale, the mean runoff coefficient, ranging from 5% to 47%, reflects the type of crust prevailing on each plot. The classification of runoff capability of the plots derived from Cr values is consistent with the one that can be assessed from the measured saturated conductivity (Table 3) and from the observations of Casenave and Valentin (1992) with high runoff on plateau clayey soil with erosion crusts and lower on sandy hillslope soils with structural crusts. The mean Cr value is lower than 50% for all the plots, but Cr computed at the event time step (volume of runoff/volume of rainfall for the event) can reach 75% on the plateau and 66% or 90% for the fallow and micro-plot, respectively. The maximum Cr values reported in Table 4 for the micro-plot, the tiger bush and the fallow plot were recorded for the second storm on the 30 Aug. 1992, under very wet soil conditions, whereas the maximum value for the millet site corresponds to a maximum of crust evolution.

At the catchment scale, we proceeded in the same manner as previously by plotting the runoff depth observed at the outlet of each catchment versus P_u. Most of the rainfall runoff events are distributed along a concave curve and, as shown before, points corresponding to extreme soil moisture conditions are offset (Fig. 9). The soil moisture conditions, either typical, very wet, or very dry, were estimated from local values on the plots. In order to provide a simple model for the estimation of runoff at the catchment scale, a modified Soil Conservation Service runoff equation (Soil Conservation Service, 1972), rewritten with P_u instead of the actual rainfall depth P, has been fitted to the data corresponding to typical initial soil moisture conditions for each catchment (Fig. 9), using the Marquardt (1963)

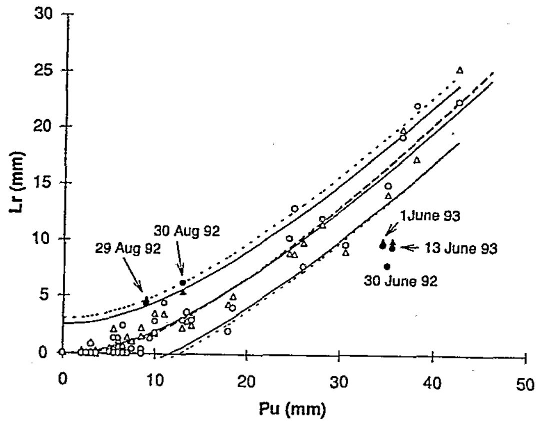

Fig. 9. L_r versus P_u for catchment 1 (open triangles) and 2 (open circles) in typical initial conditions and in extreme conditions (closed symbols), and SCS models fitted to the data set under typical conditions (solid line, catchment 1; dashed line, catchment 2). Additional lines delimit the 95% confidence interval for the individual estimation of runoff depth.

automatic fitting procedure (1).

$$L_r=(P_u-I_a)^2/(P_u-I_a+S) \quad (1)$$

L_r is the runoff depth at the outlet of the catchment, P_u is the rainfall depth calculated by the method detailed above, S is linked to the ability of the catchment to store water and is related to the type of soil, the land use and tillage and the vegetation cover, and I_a represents the initial infiltration, before runoff starts. The best fits were obtained with the value of 0 mm for I_a and respectively 40.6 and 39.1 mm for S for catchments 1 and 2. The value $I_a=0$ for both catchments is due to the fact that the SCS function was written with P_u instead of P. By definition, P_u is related to the quantity of water that can actually produce runoff, and then I_a was no longer useful in the SCS function.

Even though the $L_r(P_u)$ relationships are linear at the plot scale, they are markedly non-linear at the catchment scale, for the same range of rainfall depths. This difference in shape shows that, at the catchment scale, the amount of precipitated water that effectively contribute to runoff, estimated by P_u, is very low for small rainstorms, but increases non linearly with rainfall depth. The graph on Fig. 9 shows that, for a value of P_u greater than 20 mm, the curvature is reduced, suggesting that above this value, the proportion of rainfall depth forming overland flow increases more slightly with increasing rainfall. The results of the fitting are not very different although the surface of catchment 2 is twice that of catchment 1, revealing similar runoff production capacities. As a matter of fact, the ratio of the number of discharges recorded at the outlet of both catchments with respect to the number of rainfall events, computed from a 42 rainstorm sample, is 80% and 59% for catchments 1 and 2, respectively. As both gauging stations 1 and 2 belong to the same gully, this result demonstrates that significant quantities of water infiltrate through the bottom of the gully between the two gauging stations. The underlying deep sandy soil (up to 10 m in the vicinity of the plateau) allows for high infiltration capacity.

In-the-field observations show that all the water flowing away through gauging station 2 totally infiltrates in the sediment spreading zone, a few dozen meters downstream. Indeed, the floods are totally soaked up there, and there is therefore no transmission of water beyond this zone, that is, no water supply to the Banizoumbou river, about one km farther (Fig. 1). In other words, there exists a sharp spatial discontinuity in the surface hydrologic processes within the hillslope. Such spreading zones, occurring with various size in flat, sandy areas within the hillslopes of large watersheds, are frequently observed in the Sahelian zone. As mentioned by Rodier (1964) these flat, pervious areas contribute to the so-called hydrographic degeneration, characteristic pattern of the arid and semi-arid African regions.

To go thoroughly into the understanding of the scale effects on the hydrological processes, a simple empirical, semi-distributed model was used to estimate hillslope overland flow. A detailed description of the model was presented by Peugeot (1995). The soil surface features map (Fig. 2) and distribution (Table 1) were used to split the two catchments into a cascade of rectangular hillslope planes, homogeneous with respect to surface features, referred in the following as hillslope units. These units were organised in order to schematically reproduce the relative location of the actual soil surface feature zones, with plateau units upstream and fallow units downstream. Conceptualising the observed flow path and the Hortonian overland flow process, the model takes into account both runoff and run-on on each unit. For a given unit, the runoff depth is computed at the event time step through the corresponding $L_r(P_u)$ relationship presented in Table 4 by using, instead of P_u alone, P_u increased, if any, by the run-on depth. The relationships corresponding to the plateau slope and the piedmont units (not presented here) were derived from measurements performed in 1994 on two representative plots, following the method presented in this paper for the other runoff plots. This simple model provides an estimation of the hillslope overland flow depth for the two catchment.

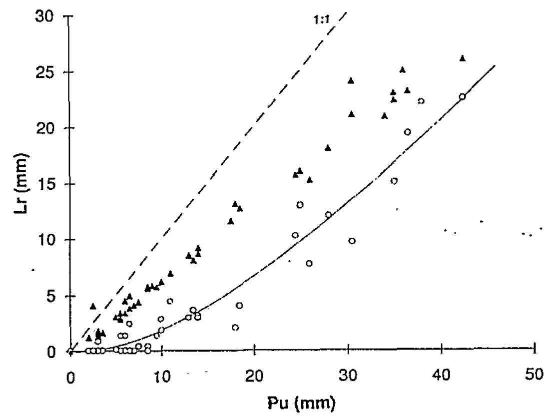

Fig. 10. Overland flow depth computed by the semi-distributed, empirical model (close triangles) and observed runoff depth (open circles) for catchment 2 presented versus P_u. Solid line represents the SCS model fitted to data under typical soil moisture conditions and dashed line is the first bisector.

The model results and the measured runoff depth plotted versus P_u are compared on Fig. 10. It is obvious that (i) the computed runoff depth is proportional to P_u, and (ii) the model systematically overestimate runoff depth. The latter result is easily conceivable if one considers that the model does not account for channel seepage. The difference in shape between the computed and the observed data markedly reminds that the hydrological processes at the catchment scale are not as simple as the linear combination of the processes observed at the plot scale. This well-known result, stated in numerous studies (Amermann and McGuiness, 1967; Beven, 1989 among others) gives here the opportunity to discuss the scale effects in the context of our study area.

5. Discussion

The measured values of hydraulic conductivity of crusts show a ratio of the saturated hydraulic conductivity of the crust to that of the subsoil ranging from 1/3 to 1/6 in the case of the plateau bare soil (Table 3). This result is consistent with other published values. Tackett and Pearson (1965) for example, observed a ratio of 1/5. In order to estimate the runoff increase due to the existence of a crust, the values obtained by Vandervaere et al. on this site were used to compute cumulative infiltration in a two-layer Green and Ampt (1911) model for 5 rain events of the 1993 season for which an important runoff was measured. Both the crust and the subsoil conductivities must be considered in order to describe the evolution in the infiltration capacity of a layered soil. Indeed, the former controls the first stages of the infiltration process while, as time progresses, the resistance to water movement caused by the soil beneath the crust becomes preponderant (Parlange et al., 1984). Runoff is simply calculated by subtracting cumulative infiltration from the total rain amount. Comparison with measured data from the tiger bush bare soil plot showed that the calculated runoff is overestimated by 8% with a 6.5 mm deep sedimentation crust (SED) and underestimated by 5% with a 10 mm deep structural crust (ST3), while it is underestimated by 51% when both layers are considered to have the properties of the subsoil. Therefore, it appears that the crust conductivity is the preponderant variable to consider for runoff prediction under the intensive sahelian rain events. This is a confirmation of the role of soil surface crusting in surface hydrological processes in that region, as previously stated by Casenave and Valentin from simulated rainfall experiments, and strengthens the relevance of soil surface features mapping in the evaluation of runoff capability of the catchments.

The effects of the locally estimated crusts characteristics on infiltration and runoff at the plot scale are obviously shown on Fig. 4. As rainfall depths can be considered similar on all the sites, the difference in infiltration capacities is not due to rain but to surface feature. Trends in soil moisture measurements are consistent with measured crust saturated conductivity values, with deeper wetting front on soils with higher hydraulic conductivity (Fig. 4 and Table 3). Except for the millet site, surface soil moisture was already at the maximum value before the rainstorms on August 30th, 1992. On the contrary, the maximum value was reached after the weeding operations on the millet plot. On this site, the maximum water storage is primarily induced by tillage. The effect of initial soil moisture, probably hidden by the prevailing tillage effects, was not clearly demonstrated here.

Water storage is variable from one plot to another, but the general pattern of soil moisture on all the sites from 1992 to 1993 is very similar, with lower infiltration the latter year on the untilled plots, due to the occurrence of a dry period in early July 1993. This result underlines the role of the temporal distribution of the rainstorms within a season on soil water storage and runoff. Due to tillage, this contrasted behaviour observed the two years was however smoothed on the millet site, where the maximum soil moisture values were only slightly lower in 1993 than in 1992.

The results on the millet plot show a high temporal variability in runoff yield. Traditionally, a first goal of the weedings is to destroy the weeds that compete with the millet for nutrients and soil water. A second one, stated in former studies (Hoogmoed and Stroosnijder, 1984; Valentin, 1993) and once again demonstrated here, is increasing soil water infiltration by destroying surface crusts. The persistency of the "weeding effect" evaluated here to 80 mm is greater than the values presented by Valentin (1981) or Hoogmoed and Stroosnijder (1984) who have estimated that a new structural crust generally forms after 20–30 mm of rain, but is rather lower than other values determined under different tillage conditions. In northern Burkina Faso, Lamachère (1991) has estimated this persistency to 200 mm for a micro-mounds tillage pattern from one squared meter experimental plots under simulated rainfall. The absence of micro-relief on the millet field soil surface induces a faster restoration of crusts than on a field including ridges. These results clearly point out the important role of the weeding operations for water availability for the crop. The role of the vegetation cover on runoff production was stated in previous studies (Dunne et al., 1991; Perez, 1994), but on the fallow plot, where the vegetation cover can reach a maximum value of 35%, there is no clear evidence for that. This result agrees with Thébé (1987) who found no influence of vegetation cover on runoff for densities of 35 to 40%, in the Sahelian northern Cameroon. Moore et al. (1979) found a similar result in southern Kenya for a vegetation cover ranging from 0 to 57%.

At the plot scale, linear relationships were fit with satisfactory results to the experimental (L_r, P_u) data corresponding to typical soil moisture conditions. The coefficients of P_u presented in Table 4 characterise the runoff capacity of each plot, and a fraction of P_u is accounted for in infiltration. Within the range of the observed rainfall depths, it is possible to express the runoff as a linear function of P_u, computed from a unique experimental value of the intensity threshold, steady over time. However, physical considerations assign the slope of the $L_r(P_u)$ relationship to be close to 1 for very large storms, as the runoff depth tends towards the total rainfall depth P, P_u being then close to P. It suggests that the equations used here are valid only for the range of observed rainfall depths, extending to values of 50 mm (actual depth). Time series show that 95% of observed rainfall depths are lower than 48 mm (Lebel et al., 1997): the observed range thus corresponds to the events the most frequently observed. Outside this range, i.e. for very large storms, these equations will probably underestimate L_r. At the catchment scale, P_u also allows to well characterise runoff without such restrictive condition as the SCS function tends asymptotically to the first bisector.

The relationships presented in Table 4 were computed for various rainstorms and a for wide range of initial soil moistures, and are sensitive to extremes of initial water content. Bearing in mind the application range of P_u, this rain-depth-derived parameter consequently appears to be a relevant parameter to compute runoff generation under typical initial conditions and stationary surface features. As it is rather easy to calculate from a rainfall hyetograph, this parameter could be used in simple runoff production modules in hydrologic modelling at local scale or at the field scale, that is, on a few ha. It can be noticed that the success encountered here by using P_u is related to the Horton-type runoff processes. Under different climates with interflow or baseflow, this parameter would perhaps not be so useful.

Upscaling the local $L_r(P_u)$ relationships to the hillslope scale lead to a bad estimation of the actual runoff depth. Two underlying assumptions of the model are that (i) surface water is supposed to be transferred in the form of a sheet flow and (ii) that lumping the processes from the plot to the hillslope hydrologic unit is possible. The linear $L_r(P_u)$ relationships derived for the plots account for the processes and the heterogeneity of hydrologic properties internal to each plot. The second assumption thus implies that this heterogeneity does not increase when the observed area is wider and that the runoff generation processes remain identical. These two assumptions are obviously false in most cases, as the variety of soil patterns, crusts or vegetation is higher on large areas, resulting in additional processes such as water storage in the micro-relief, or infiltration in more pervious areas. Such pervious areas contribute to soak up run-on flow, with particular impact when located downstream from a high runoff potential zone. Relative location of surface features is of high importance in semi-arid areas. Consequently, as stated above, the resulting effect is a decrease in runoff depth at the catchment scale. For small rainstorms, intermediate abstraction by both the hillslope and the channel are higher than for larger events. As a matter of fact, field observations have shown that for small storms, the bare sloping zones located in the vicinity of the gully were predominantly contributing. Due to intermediate abstraction, not the entire catchment area contribute to runoff. This result strengthens the hypothesis of a variable source area pattern for semi-arid watersheds, as mentioned by Yair and Lavee (1990). The effective contributive area increases over the catchment with increasing rainfall. As suggested by Fig. 10, for P_u greater than 20 mm, the modelled and the observed runoff depth seem to evolve following two parallel straight lines. For such rainfall depths, the modelled runoff is generated by the catchment as a whole: consequently, the effective contributive area is close to that underlying in the model, i.e. the total catchment area.

Considering the mean Cr value, the catchments seem to behave like the fallow plot, even though catchment 2 contains a greater surface of plateau and piedmont bare soils (with high runoff potential) than catchment 1 (Table 4). Catchment 2 then has a higher runoff potential, but seepage in the channel alluvium decreases the flow volumes observed at the outlet. Evidence for seepage processes were also strengthened on Fig. 10.

Both intermediate abstraction and channel seepage contribute to decrease the flow at the catchment outlet. The S parameter appearing in the SCS equation precisely describes the maximum abstraction capacity within a catchment.

In order to extend the outcomes of this study to wider watersheds of the study area, Eq. (1) has been fitted in the same manner to (L_r, P_u) data sets related to the drainage basins of the Bazanga Bangou (0.35 km^2), Wankama (2.1 km^2), and Sama Dey (6.3 km^2) described in Desconnets et al. (1997). The values of S obtained on those catchments and that computed on the small nested catchments were plotted against the catchment area (Fig. 11). The graph shows an increase of S, that is, infiltration capacities within the

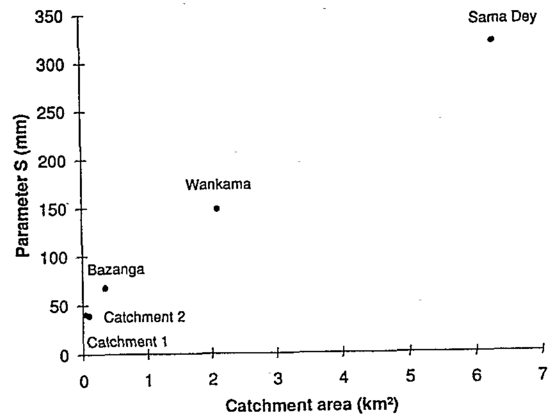

Fig. 11. Evolution of the S parameter from the SCS abstraction function versus the area of different watersheds studied within the ECSS.

catchment, with increasing drainage area. As described in Desconnets et al., the Wankama pool basins include a flat sandy zone, similar to that mentioned above, into which surface runoff can dissipate. Such zones have also been observed on the Sama Dey catchment. These spreading zones, increasing intermediate abstraction within the catchments, are indicators of the degradation of the drainage network. Total infiltration in spreading zones and channel seepage might be frequent over the area, and from our point of view, these processes are of first concern in the limitation of large scale runoff. Forthcoming regionalisation work on surface hydrologic processes would undoubtedly take advantage of an accurate description of them.

6. Conclusion

Soil moisture investigations at the plot scale allow to draw three major results: no deep infiltration was observed, and the wetted depth reached a maximum depth of about 4 m on the millet plot, the most pervious site. Of interest is also the fact that the maximum surface water storage was not reached by the end of the rainy season but both years late August, when storms were more frequent. Lastly, soil water was fully dried off in the dry season, resulting in a minimum water storage at the beginning of the next rainy season.

From runoff data collected on runoff plots, we have confirmed the role of soil crusting in runoff generation in the Sahel. At the very local scale, conductivity measurements have confirmed that infiltration and runoff were driven by the hydraulic properties of the crust. At the plot scale, the parameter P_u, defined as the rainfall depth that can actually produce runoff seems to be relevant to estimate runoff production on either runoff plots or small catchments, for the typical initial soil moisture conditions that have been defined. Wetter or drier conditions induce higher or lower discharge volumes. The influence of vegetation growth on runoff yield did not clearly appear in the case of the fallow sites, but runoff generation was very sensitive to tillage (millet field). For the most typical moisture conditions observed, the rain-depth-derived P_u parameter allowed to confidently estimate the observed runoff depths (L_r) at both scales.

At the catchment scale, the comparison of the observed runoff depths to the results of the simplified model involving linear combination of the relationship obtained on the plots allowed to propose that runoff was generated by a partial source area process, with an increase of the effective contributive catchment surface with increasing rainfall.

Discharge data at the catchment scale highlight important water infiltration through the bottom of the gully, and total infiltration in a flat spreading zone downstream from catchment 2. These flat areas, possibly numerous over the ECSS, induce sharp modifications in the floods features by soaking up on a few dozen meters, if ever, the water flowing away in gullies. Otherwise, the water reaching the depressions accumulate in pools. Between the plot or small catchment scale, where runoff was measured, and the pool scale, the spatial discontinuities of surface flows constitute a major problem in the concern of aggregation of hydrologic processes.

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Estimating hydraulic conductivity of crusted soils using disc infiltrometers and minitensiometers

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Abstract

Although soil crusting has long been recognized as a crucial runoff factor in the Sahel, very few field methods have been developed for the measurement of the crust hydraulic conductivity, which is difficult to achieve because of the small thickness of most surface crusts. A field method, based on the simultaneous use of disc infiltrometers and minitensiometers is proposed for determining the crust hydraulic conductivity and sorptivity near saturation. On crusted soils, the classical analysis of the steady state water flow was found to be inadequate. The proposed method is based on sorptivity measurements performed at different water supply potentials and uses recent developments of transient flow analysis. A minitensiometer, placed horizontally at the crust–subsoil interface, facilitated the analysis of the infiltration regime for the crust solely.

Results are shown for representative soil units of the East Central Super Site of the HAPEX-Sahel experiment: fallow grasslands, millet fields and tiger bush. Non-crusted soils were also considered and validated the transient method as demonstrated by comparison with Wooding's steady state solution. This validation was obtained in the case of fallow grasslands soil but not for the millet fields. In this latter case, the persistent effects of localized working of the soil to remove weeds caused large variations in infiltration fluxes between the sampling points, which tended to dominate over effects of differences in applied potential. For the tiger bush crusted soils, the ratio of the saturated hydraulic conductivity of the crust to that of the underlying soil ranges from 1/3 to 1/6, depending on whether the crust is of a structural (ST) or sedimentation (SED) type. The method also allows the estimation of a functional mean pore size, consistent with laboratory measurements, and 40% less for the crusts in comparison with the underlying soil.

The results obtained here will be used in hydrological models to predict the partition of rainfall between infiltration and runoff.

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