Journal of Soil Science and Environmental Management Vol. 3(12), pp. 297-307, December 2012
Available online at http://www.academicjournals.org/JSSEM
DOI: 10.5897/JSSEM11.156
ISSN 2141-2391 ©2012 Academic Journals

Full Length Research Paper

Temporal evolution of the hydrodynamic behaviour of sandy deposits in the Sahelian part of Burkina Faso

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Accepted 24 July, 2012

Sahelian areas of Burkina Faso are confronted with various natural and anthropogenic processes that can lead to degraded areas where the vegetation is localized mainly in units called "sandy deposits". Those deposits are subject to temporal dynamics that greatly impact their hydrodynamic behaviour; the initial drying crust evolves frequently to a drying crust in transition, and ultimately to an erosion crust (according to the crust classification established by Casenave and Valentin (1992)). To better apprehend the temporal evolution and the hydrodynamic behaviour of the sandy deposits, various field measurements have been performed on seven monitoring sites installed in three contrasting zones corresponding to the three crust types. The sites located on erosion crusts and on drying crusts in transition are characterized by a low infiltration capacity which favours run-off (run-off coefficients between 50 and 80%). The sites situated on drying crusts are characterized by a high hydraulic conductivity and infiltration capacity, as well as, relatively low run-off coefficients (smaller than 40%); the water storage in the root zone is more important and water infiltrates deeper into the soil and, occasionally, drains below the depth of 50 cm. The study confirms that soil properties are subject to important temporal evolution, resulting in a strong degradation.

Key words: Aeolian sandy deposits, infiltration, run-off, soil hydrodynamic, soil physical properties, soil surface crusting, Sahel.

INTRODUCTION

The increase in the demographic and animal pressure in the Sahelian zone of Burkina Faso has induced for a few decades, profound changes in the management of natural resources. These anthropic disturbances have led to a partial clearing of the native vegetation, a degradation of the soil (wind and water erosion), a deterioration of the soil water regime, as well as, a decrease in the effectiveness of water for the vegetable production. The impact of low inherent fertility and degradation of Sahel soils is aggravated by rainfall scarcity and irregularity, population increase and the consequent reduced length of fallowing periods (Lahmar et al., in press).

In this context, it is essential to understand the behaviour of surfaces that are still little degraded in order to be able to define measures capable to preserve them. This is the case of sand patches of wind origin which play a key role in the ecological functioning of the gentle-slope glacis of the Sahelian part of Burkina Faso (Ribolzi et al., 2000). Those patches that present a high infiltration capacity and an herbaceous cover that can reach 90% of

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the surface are particularly exposed to degradation processes. They are spatially discontinuous and form micro-dunes, generally composed of two main horizons: (i) a top layer (from 2 to 5 cm deep) made of loose sands with numerous macropores formed by plant roots and soil fauna; and (ii) a sub-surface horizon (maximal thickness 50 cm), more compact and less permeable, composed of sandy layers alternating with plasmic microlayers. Such a context leads to usually dry soils (Davis et al., 2010) highly susceptible to wind erosion (Toure et al., 2011). According to Leprun (1999), soils and vegetation are subject to a continuous evolution under the cumulated effects of wind and water erosion. Other studies performed in the area (Pieri, 1989; Young and Onstad, 1978; Rajot et al., 2003; Ribolzi et al., 2006) confirmed that soils are very sensitive to erosion and degradation facilitated by generally low organic matter content. Reszkowska et al. (2011) have shown that intensive overgrazing can lead to serious grassland degradation and deterioration of soil structure.

In view of the fact that rainfall is characterized by important fluctuations and the number of rainy days is decreasing (Carbonnel and Hubert, 1992; Albergel et al., 1995; Descroix et al., 2012), it is important to gain a better insight into the run-off/infiltration partition of the rain. The main objective of this study was to better understand processes involved in the degradation of aeolian formations and the hydrodynamic behaviour of sandy patches at various evolution stages. Such a study is of prime importance in view of fighting efficiently desert encroachment.

MATERIALS AND METHODS

The study area is located in the Sahelian part of Burkina Faso on an experimental field located close to the village of Katchari, 15 km West of Dori. It is a non-cultivated pastoral area overgrazed by livestock. Vegetation is of Sahelian steppe type with thorny species, characterized by discontinuous grassland and woodland areas. Two main vegetal units that can be distinguished (Zerbo, 1993):

1. Dune and sandy formations with a dense vegetal cover of up to 90% in the rainy season;
2. Pediment (glacis) formations where vegetation is virtually absent, with only sparse wooden species at a density of 1 to 2 trees/ha.

On the whole, vegetation is subject to a constant degradation caused by numerous physical and socio-economical factors. In woodland areas, the mortality rate can be as high as 200 to 300 trees/ha/year (Thiombiano, 2000). The cumulated annual rainfall presents a strong inter-annual variability with an average of 513 mm and a standard deviation of 123 mm (period 1925 to 2005). The 3 observation years (2003 to 2005) were characterized by two years with a high rainfall (624 mm in 2003 and 875 mm in 2005) and a rainfall deficit year (369 mm in 2004). Rainfalls occur essentially between August and May (rainy season), the other months (September to April) being characterized by an almost total absence of precipitations.

The study was focused mainly on three types of crusts (by reference to the crust classification established by Casenave and Valentin (1992)) frequently observed in the study area, namely:

I. Drying crusts (Dry): Surfaces characterized by a massive single sandy microhorizon, presenting a high porosity and supporting most of the vegetation.
II. Drying crusts in transition (Dry/Ero) on which the soil proportion occupied by surface features of erosion crust type increases regularly in importance;
III. Erosion crusts (Ero): Smooth surfaces made of a single seal of fine cemented particles and characterized by a very low porosity; no vegetation grows in these areas.

The soil surface classification used here stems from the morpho-genetic typology established by Casenave et Valentin (1989, 1992) from data obtained under simulated rainfall conditions in Sahelian areas. Soil surface features are prone to spatial and time-dependent variations (Valentin, 1991; Valentin et al., 2004) influenced by the combination of successive or simultaneous climatic, faunal and anthropic processes causing soil superficial re-organizations susceptible to the formation of surface crusts. The experiments were performed on seven sites located in three distinct areas:

i. An unprotected grazed area undergoing a continuous degradation due to natural and anthropic processes. Three measurement sites were installed in this area: one on an erosion crust (S1); a second on a drying crust (S3) and a third on a drying crust in transition (S2).
ii. A restored fenced area with restoration practices carried out in 1998, consisting in the application of branches to increase the roughness of the ground and to accelerate the processes of trapping the aeolian sands. Two sites of measurements were set up in this area: one (S5) on a drying crust and the other (S4) on a drying crust in transition.
iii. A protected area, isolated from anthropic activities since 1985, of approximately 20 ha, in which two sites of measurements were installed on a drying crust (S6 and S7). Some characteristics of the various measurement sites are summarized in Table 1.

Pedological studies and grain size distribution analysis reveal a clear prevalence of leached ferruginous soils. They present a loamy-sandy to sandy-loam texture (according to the USDA Soil Textural Triangle) close to the soil surface, evolving towards a sandy clay loam below 30 to 40 cm. These tropical ferruginous soils are characterized by a poor structural stability of the surface horizons attributable to their high silt and fine sand contents and low organic matter content. They are subject to a very active dynamics, largely determined by water and wind erosion. The perpetual shaping leads to the formation of new surface features as pointed out by Casenave et Valentin (1989). Such an evolution is accompanied by the emergence of more recent sandy aeolian deposits and by sand losses from the surface horizon which may reveal a drying crust in transition. When all the sand has disappeared, the soil evolves towards an erosion crust. Some other basic physical characteristics (soil density, organic matter content and porosity) have been determined on undisturbed soil samples collected before the beginning of the rain season 2003. The values of bulk density range between 1.49 and 1.74 g/cm^3. The organic matter content is always lower than 2% with most values ranging between 1.3 and 1.8%.

Each experimental site was equipped with devices making it possible to estimate the main components of the soil water balance, namely run-off, spatio-temporal variations of the soil water content and water pressure head, as well as the drainage at the depth of 50 cm.

The run-off was estimated using 1 m^2 microplots delimited by a metallic frame from which the surface water was transferred to a barrel located downstream (Podwojewski et al., 2008; Malam et al., 2009); the soil water content was measured up to a depth of 80 cm using a neutron probe; the values of the soil water pressure head were given by tensiometers installed at various depths (10, 20, 30,

Table 1. Important features of the measurement sites.

Location	Numbering of sites	Surface crust	Slope (%)	Vegetation
Unprotected grazed area	S1	Erosion crust (ERO)	2.5	Absence of vegetation
	S2	Drying crust in transition (DRY/ERO)	2.5	Sparse vegetation
	S3	Drying crust (DRY)	2.8	Abundant vegetation
Restored fenced area	S4	Drying crust in transition (DES/ERO)	1.0	Sparse vegetation
	S5	Drying crust (DRY)	1.8	Abundant vegetation
Protected area	S6	Drying crust (DRY)	2.5	Abundant vegetation
	S7	Drying crust (DRY)	1.0	Abundant vegetation

40 and 60 cm) and connected to a mercury pressure gauge. The measurements of water content and pressure head were taken daily during the rainy season and monthly during the dry season. The drainage was estimated at the depth of 50 cm from tensiometric values measured at 40 and 60 cm after determination of the unsaturated hydraulic conductivity function from internal drainage experiments. The rain was measured with 3 hand rain gauges and an automatic recording rain-gauge installed in the study area. Subtraction of the run-off volume from the rainfall volume gives the volume of infiltrated water. The evolution of the soil surface properties was monitored by means of tests carried out with:

i. A double-ring (of respective diameters 25 and 33 cm) infiltrometer; the tests performed under a constant water head of 3 cm provide the values of the steady infiltration rate.
ii. A tension disk infiltrometer (100 mm diameter) implemented under a pressure head at the disk of - 40 mm.

In both cases, the measurements were repeated three times, representing nearly 300 measurements in total. The tests performed by means of the disk infiltrometer were analyzed according to the procedure described by Vandervaere et al. (2000a) who proposed an expression similar to the equation of Philip (1969) to characterize the transitory uni-dimensional axisymmetric infiltration starting from a circular source on the ground surface, namely:

$$I = C_1\sqrt{t} + C_2 t \quad (1)$$

where I is the cumulated infiltration depth and t is time. C_1 and C_2 are coefficients which can be estimated with the equations suggested by Haverkamp et al. (1994):

$$C_1 = S \quad (2)$$

$$C_2 = \frac{2-\beta}{3}K + \frac{\gamma S^2}{r(\theta_0 - \theta_i)} \quad (3)$$

where S is the capillary sorptivity, K is the hydraulic conductivity, γ is a constant equal to 0.75, β is a parameter ranging between 0 and 1, depending on the type of soil and the applied pressure head, r is the radius of the disk, θ_i and θ_o are the initial and final volumetric water contents. The advantage of the method is that it does not require any estimation of the permanent flow and thus takes less time. On the other hand, it provides only an interval of values for hydraulic conductivity K, between K_{min} for $\beta = 0$ and K_{max} for $\beta = 1$. Vandervaere (1995) proposes to use a value of β equal to 0.6 (assuming a lognormal distribution law for the hydraulic conductivity) for the calculation of the hydraulic conductivity.

Our tests were analyzed by the "differentiated linearization method" (Vandervaere et al., 2000b). This method consists in differentiating the cumulative infiltration data with respect to the square root of time. Performing this differentiation on Equation 1 gives:

$$\frac{dI}{d\sqrt{t}} = C_1 + 2C_2\sqrt{t} \quad (4)$$

Thus, plotting $\Delta I/\Delta t^{1/2}$ vs. $t^{1/2}$ should be linear, with C_1 equal to the intercept and C_2 the half-slope of the regression line. The values of the sorptivity S and of the hydraulic conductivity K can be deduced from Equations 2 and 3. These values have been used by certain authors (Philip, 1985; White and Sully, 1987) to define the capillary length λ_c which expresses the relative importance of the capillary and gravitational forces acting on the penetration of water into the soil. Its mathematical formulation is:

$$\lambda_c = \frac{bS^2}{(\theta_0 - \theta_i)K} \quad (5)$$

where b is a parameter depending on the form of the relationships between the hydraulic conductivity and the pressure head, on the one hand, and between the water content and the pressure head, on the other hand; a value of 0.55 has frequently been considered as an appropriate approximation for the majority of soils (Warrick and Broadbridge, 1992). By using the elementary laws of capillarity, Philip (1985) introduced the average dimension of the hydraulically functional pores, λ_m, given by the following equation:

$$\lambda_m = \frac{\sigma}{\rho_w g \lambda_c} \quad (6)$$

where σ is the water surface tension, ρ_w the water density and g the acceleration due to gravity.

By introducing into Equation (6), the values of σ (0.072 N/m at 25°C), ρ_w (1000 kg/m^3) and g (9.81 m^2/s), one obtains, expressing λ_m in µm:

$$\lambda_m = 13.3\frac{(\theta_0 - \theta_i)K}{S^2} \quad (7)$$

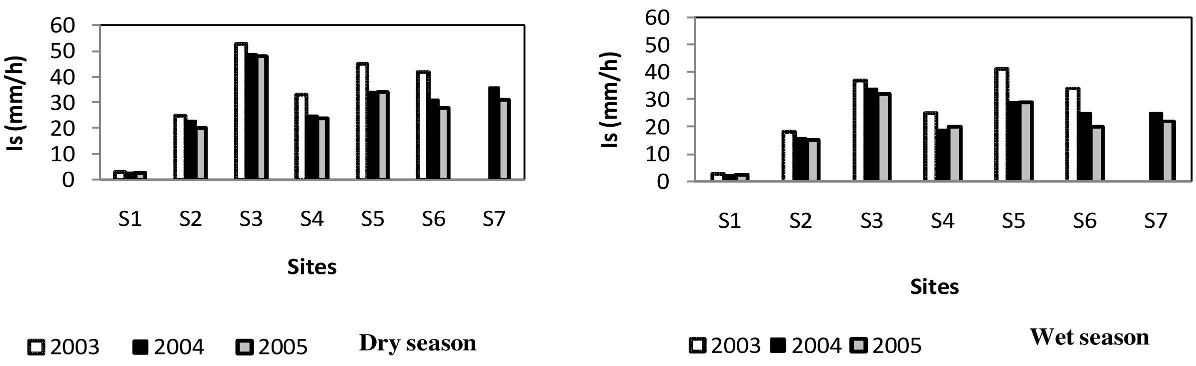

Figure 1. Evolution of the steady infiltration rate on the various measurement sites during the dry and wet seasons 2003, 2004 and 2005

RESULTS AND DISCUSSION

Evolution of the steady infiltration rate

Figure 1 illustrates the evolution of the steady infiltration rate of the soil during the three years of measurements. The figure shows the following:

i. Great differences depending on the properties of the soil surface. The highest values are observed on the drying crusts (S3, S5, S6 and S7), the lowest on the erosion crust (S1), whereas the drying crusts in transition (S2 and S4) present intermediate values. Those differences are probably attributable to a reorganization of the soil surface features and changes in the poral system of the surface layer (Vandervaere, 1995; Niang, 2000; Ndiaye, 2001).
ii. A seasonal variation characterized by generally higher values in the dry season than in the wet season. Such a reduction can be explained by several factors, in particular the destruction of the surface crust by animals in the dry season and the decrease of the dimension of the channels which connect the largest pores in the rainy season.
iii. A similar pattern of evolution of the steady infiltration rate on the different sites in the dry and wet seasons and a decreasing tendency from one year to another.

Evolution of the unsaturated hydraulic conductivity

On the whole, 136 tests of infiltrometry with imposed pressure head (h_0 = -40 mm) were carried out on the seven sites during the three years of experiments. Figure 2 illustrates the evolution of the unsaturated hydraulic conductivity during the period considered. As in the case of the steady infiltration rate, differences between the three types of crusts emerge clearly; the sites located on the drying crusts (S3, S5, S6 and S7) present the highest values, whereas the soil surface of the erosion crust (S1) is characterized by a low conductivity and the sites in transition (S2 and S4) by intermediate values. Here again, higher values are obtained in the dry season than in the wet season. This may be explained by several factors among which the presence of a vesicular porosity which is caused by air trapping after strong downpours. Some authors like Valentin (1981) and Casenave and Valentin (1989) have shown that in semi-arid areas the hydrodynamic properties of the microlayers of surface crusts can be affected by the appearance of unconnected occluded pores which reduce the infiltration capacity (Valentin, 1981) in the rainy season. Such a reduction in the infiltrability and the presence of vesicular pores was confirmed by Albergel et al. (1986) who highlighted a good relationship between the abundance of vesicular pores and run-off generation. Other factors, such as the slaking phenomenon related to soil surface structural degradation under the action of rains can explain the reduced values of infiltrability in the rainy season. During the three measurement years, a significant decrease of the unsaturated hydraulic conductivity was observed on all the sites, with the exception of S1 installed on an erosion crust. To try and explain the causes of this reduction, a study was conducted on the evolution of the sorptivity and of the average dimension of the functional pores of the surface layer.

Characterization of the average dimension of the functional pores

Table 2 presents the values of the sorptivity S (Equation 2) obtained from the infiltration tests performed during the three years at a pressure head h_0 = -40 mm. The sorptivity of the drying crusts is higher than that of the drying crusts in transition, itself stronger than that of the erosion crust. The latter is close to the value (6 mm/h$^{1/2}$) as obtained by Cook and Broeren (1994) on an alluvial loamy soil at an identical pressure head (h_0 = -40 mm). The values corresponding to the drying crusts are close to the value 13 mm/h$^{1/2}$ measured by White and Sully (1987) on a sandy loam soil at a pressure head h_0 = - 30 mm.

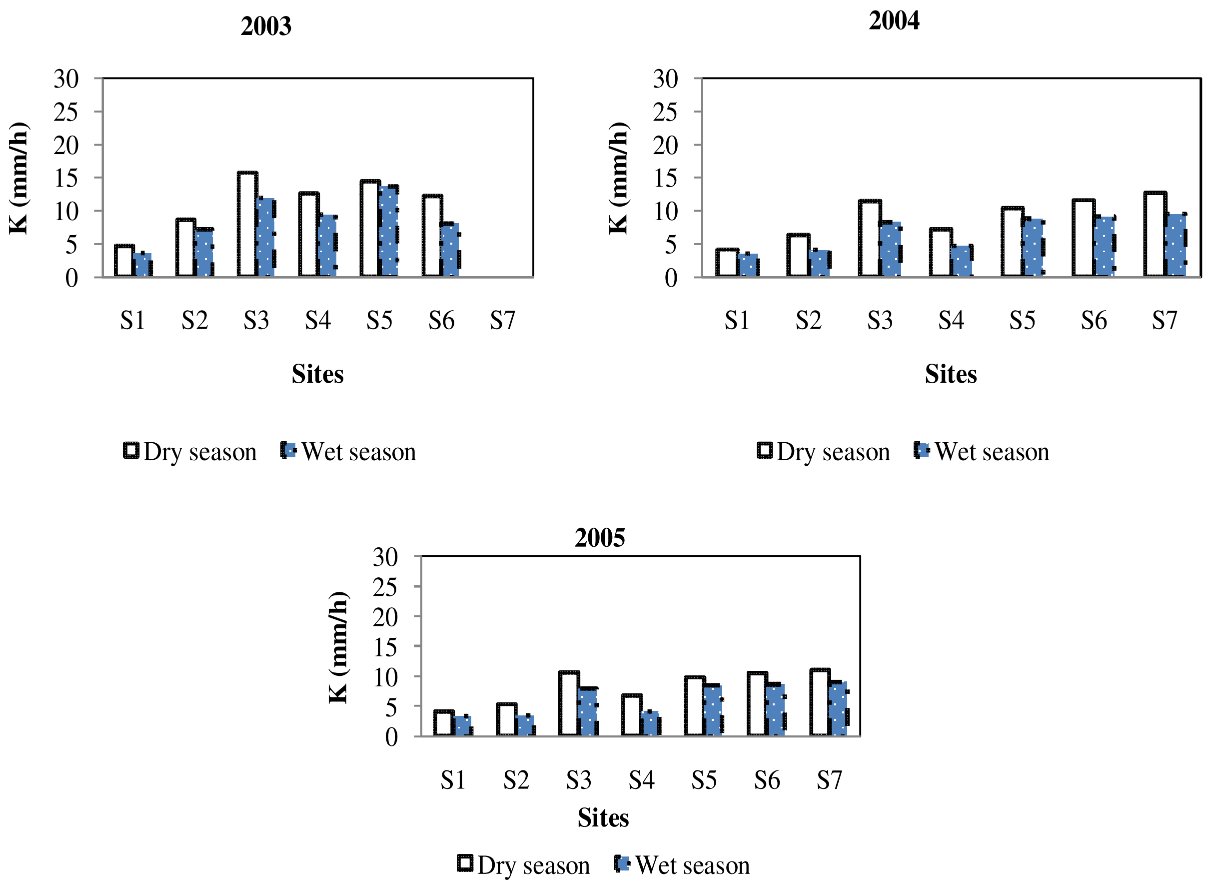

Figure 2. Values of the surface hydraulic conductivity K measured under a pressure head of -40 mm during the dry and wet seasons 2003, 2004 and 2005.

Table 2. Water balance components on the different sites for the period June 1st to August 28^{t}h, 2004.

Parameter	S1 (Ero)	S2 (Dry/Ero)	S3 (Dry)	S4 (Dry/Ero)	S5 (Dry)	S6 (Dry)	S7 (Dry)
P	355	355	355	326	326	322	322
ΔS	7	5	10	8	9	12	12
R	252	183	145	165	121	120	113
D	0	0	26	0	24	23	26
ETR	96	167	174	153	172	167	171

P, Rain; ΔS, storage change R, run-off; D, drainage; ETR, real evapo-transpiration.

Figure 3 presents the evolution of the average dimension of the functional pores λ_m (Equation 7) during the three measurement years, for both, the dry and wet seasons. The S1 site presents a clearly different pattern than the other sites. Functional pores have a lower dimension which varies little during the three years.

The drying crusts (S3, S5, S6 and S7) present the highest values of λ_m with a maximum for S3; this is probably related to a high porosity of this type of surface formation, accentuated by the destruction of the crust by cattle tramping. The soils located on the drying crusts in transition (S2 and S4) present intermediate values of λ_m, be it on the drying crusts or on the drying crusts in transition, a progressive reduction in the dimension of the functional pores is observed which tends to approach the values of the erosion crusts. This confirms that the transition from a drying crust to an erosion crust results in a decrease of the size of the functional pores due to the concentration and the compaction of the finer particles in a plasmic layer. This layer becomes even more compacted when exposed to the direct impact of rain drops as in the case of the erosion crust where the overlying sand micro-layers have been removed by wind or water erosion.

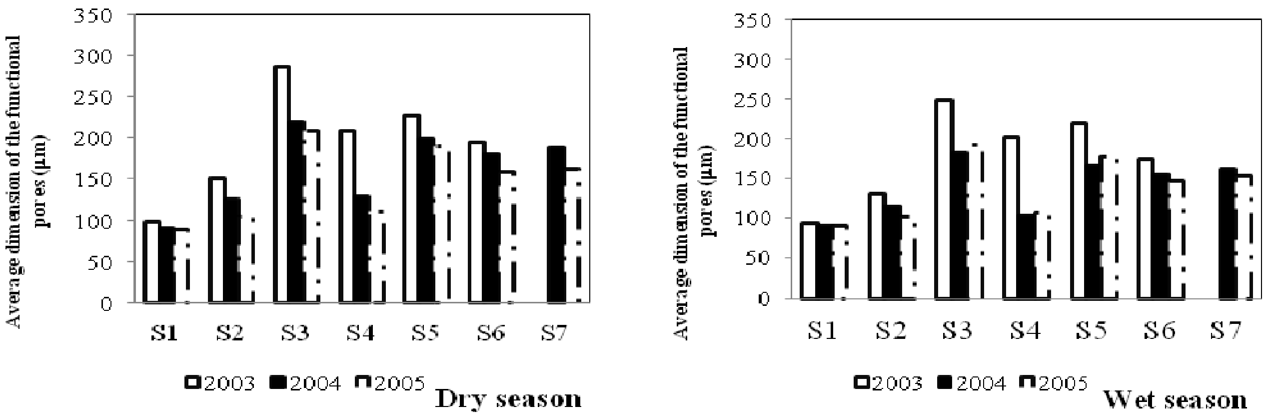

Figure 3. Evolution of the average dimension of the functional pores in the dry and wet seasons on the different sites during the 3 years of experiments.

Run-off

Table 3 presents the values of the run-off coefficients measured on the various sites during the wet season 2004. The results are consistent with those mentioned previously and confirm that run-off is notably higher on the erosion crust and, at a lower degree, on the drying crust in transition. The increase in run-off induces a fast degradation of the physico-chemical properties of the surface horizons and important soil losses by erosion.

Follow-up of the temporal evolution of soil moisture

Figure 4 presents the temporal evolution of the volumetric water content measured at 10, 20 and 50 cm depths on the different sites in 2004. It shows that, for all the measurements sites, the water content variations are important near the soil surface, but they remain weak in the subsurface horizons. Here again, the behaviour of the sites can be attached to two great units. The first one includes the sites located on the erosion (S1) and drying crusts in transition (S2 and S4) on which the water content variations occur mainly in the surface horizons (first 30 cm), whereas below 40 cm, the water content remains almost constant and at a very low level (moisture profiles not presented here). This suggests that the effect of precipitations remains limited to the superficial layers. The second unit gathers the sites located on drying crusts (S3, S5, S6 and S7) on which the water content variations are more important, even at 50 cm where occasional drainage can occur.

Study of the water balance

The study of the water balance was carried out from June 1st to August 28th, 2004, period during which tensiometric measurements were available. All the components were measured (storage change ΔS, rainfall P, run-off R and drainage D), with the exception of real evapotranspiration ETR estimated by means of the water balance equation.

The values of the various terms of the water balance are illustrated in Figure 5 that showed that the real evapo-transpiration ranges from 27 to 47% of the rain on the sites S1, S2 and S4 whereas it is between 49 and 53% on the other sites (S3, S5, S6 and S7). This difference is related to lower initial water content and to the absence of vegetation on the sites S1, S2 and S4 exposed only to evaporation, whereas on the other sites, the extraction of the water from the soil is due to the combined effects of evaporation and transpiration (Descroix et al., 2012). On a daily basis, the average values of ETR vary between 1 and 2 mm/d and are weak compared to the potential values (about 6 mm/d). The difference could be explained by the low pluviometry observed in 2004 which resulted in a reduced availability of soil water to satisfy the evaporative demand. The entire previous elements made it possible to highlight two main different surface types:

i. Run-off type surface corresponding to sites S1, S2 and S4: On these sites, the existence of a thin superficial film (plasmic layer) made of fine compacted particles results in a reduction in the infiltration capacity due to a reduced porosity, assimilating these pellicular organizations to a hydraulic barrier which strongly limits water infiltration water into the soil. Such surfaces are generally deprived of vegetation and generate important run-off resulting in a very weak accumulation of water in the soil. Infiltration affects only the surface horizon where most of the moisture fluctuations occur. Similar results have been obtained by Hiernaux et al. (2009) who showed a 5% yearly decrease in yields related to soil degradation. Within the framework of a simulation of impacts of climate change effects and land use, Séguis et al. (2004) obtained run-off ratios that can vary by up to a factor of three between degraded and undegraded soils.

Table 3. Values of the runoff coefficient (RC) at the experimental sites during the period June 1st to August 28th 2004.

Sites	S1 (Ero)	S2 (Dry/Ero)	S3 (Dry)	S4 (Dry/Ero)	S5 (Dry)	S6 (Dry)	S7 (Dry)
RC (%)	71 (2)[a]	52 (8)	41 (4)	51 (6)	37 (5)	37 (7)	35 (6)

[a] Standard deviation in parenthesis.

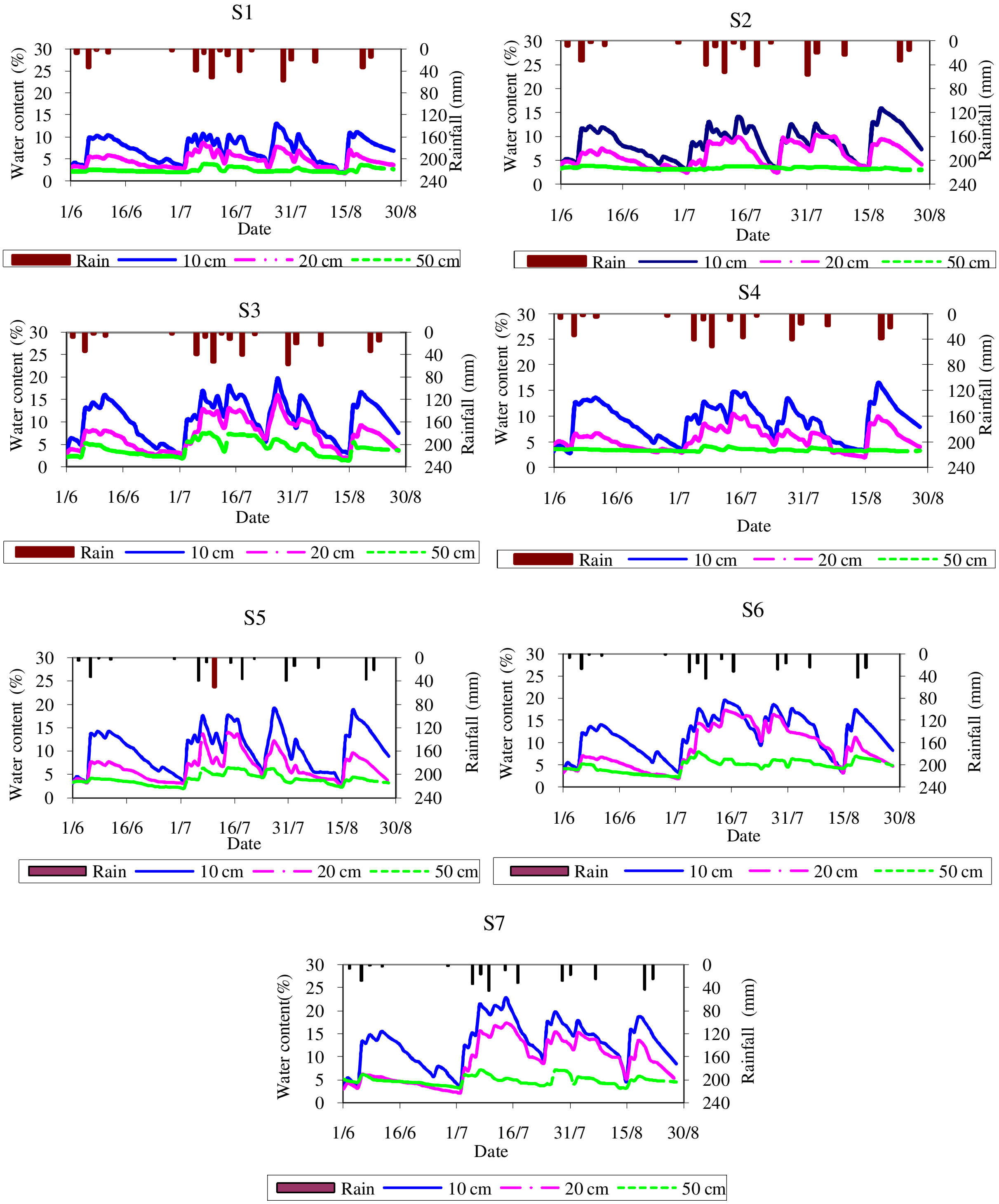

Figure 4. Temporal evolution of the water content at different depths (10, 20 and 50 cm) on the various sites in 2004.

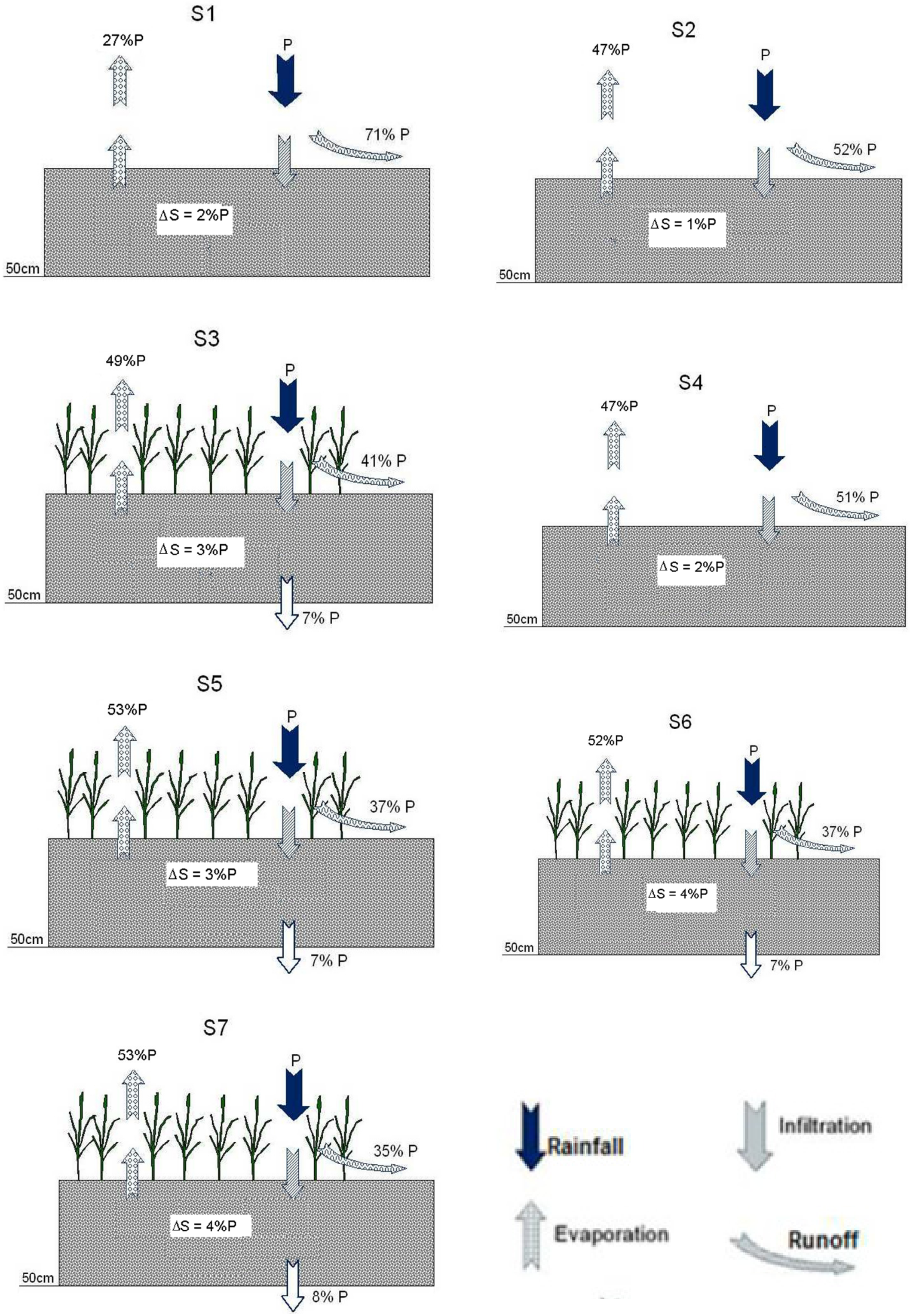

Figure 5. Water balance components on the different sites, for the period of June 1st to August 28th, 2004.

ii. More permeable surfaces corresponding to sites S3, S5, S6 and S7: The absence of a surface film creates conditions more favourable for infiltration, which results in a more important accumulation of water in the soil. In some cases (sites S6 and S7), the transfer of water in the subsurface horizons is facilitated by the presence of roots and termites galleries (Niang, 2000). Such soils are characterized by an increased soil rugosity related to surface aeolian deposits which create a more permeable epipedon responsible for a signicative reduction of soil losses (Thiombiano, 2000). The differentiated behaviour of these two surface types shows the complex nature of the constitution and functioning of the Sahelian soils. Soil surface properties and plant cover play a central role on the rain water fate; they condition the prevalence of run-off or of infiltration and correlatively, the importance of soil water storage. Several authors like Zhen (2006), Ye et al. (2006), Nicolau and Asensio (2000), Moreno-de las Heras et al. (2008) and Moreno-de las Heras et al. (2009) have showed in similar areas, that the presence of a plant cover tends to increase infiltration, reduce surface flow and delay erosion. Our results showed that in semi-arid areas, the hydrodynamic properties of the superficial micro-horizons determine to a large extent, the soil infiltrability and that the surface hydraulic properties are affected by degradation under the combined effects of climate and anthropic actions. Several authors like Casenave and Valentin (1989); Reynolds et al. (2007); Malam et al. (2009) and Zika and Erb (2009) reported also similar conclusions.

Soil degradation favours erosion (Cotler and Ortega-Larrocea, 2006; Onda et al., 2007) and consequently, a progressive destruction of soils affecting dramatically their fertility (drop of the exchange capacity and of the available elements, in particular) and the water balance (run-off increase, reduction in the available water for plants, modification of the water regime and of the soil-atmosphere exchanges). The sites installed on drying crusts are characterized by a high infiltration capacity, which facilitates the water supply of the root zone and favours the survival of the vegetation. This demonstrates that the water behavior of the studied soils is closely linked to surface features which evolves very rapidly under demographic pressure and the over exploitation of natural and environmental resources (Descroix et al., 2009; Larwanou and Saadou, 2011). Overgrazing plays an especially important role during wet periods; trampling may compact, disturb and loosen the soil surface and render it more susceptible to run-off and erosion. Indirect effects influencing dramatically infiltration and run-off may also result from a reduction of vegetation and damages caused to plants by trampling (Dunne et al., 2011).

The water behavior of the soils under study clearly illustrates one of the desertification processes related to a large extent to human and climate action. In the given area, the population growth by about 3% (INSD, 2010) results in profound changes of management techniques and the use of natural resources and rural areas. Human-caused disturbances (abusive cutting of wood, poor range management, setting of bush fires, overgrazing) generate rarefaction of the vegetation, soil degradation by water and wind erosion and deterioration in soil water regime (Ganaba et al., 1998). Previously, soils were left fallow to preserve their quality (structure and fertility), but presently they are cultivated each year. Such an over-exploitation leads to soil nutrient depletion seldom compensated by addition of fertilizers. The area's climate is characterised by a high degree of rainfall irregularity in space and time. Rainfall is often preceded in June and July by heavy sand storms generating strong wind erosion which includes sand transport, deposition and remobilization. Water erosion causes removal and selective loss of fine elements, the consequence of which is the formation of erosion crusts covering large surfaces of the Sahel region. All of this leads to important soil losses because of low infiltration rate and, consequently, high run-off rate.

The monitoring of the evolution of the sandy deposits reveals a strong spatial and temporal variability. Though, the average height remains relatively constant over the years, lateral dimensions tend to decrease over time. This is probably due to the wind action which, in interaction with water erosion, leads to a progressive diminution of the area occupied by the sandy deposits. Such hypothesis is supported by a research work conducted on the Katchari site (Thiombiano, 2000) during which some sandy deposits disappeared completely. The results obtained on the Katchari site attest the predominant role of wind and water erosion in the modification of the morphological characteristics of the soils.

Conclusions

This work aimed at better understanding, in terms of physical and hydrodynamic characterization, the hydrodynamic behaviour of different types of sandy deposits which are among the rare ecological units met in the Sahelian part of Burkina Faso which are susceptible to support vegetation. It made it possible to highlight important differences in water behaviour of the experimental sites related essentially to the variations of the surface properties which considerably influenced the components of the water balance.

Two categories of soils presenting differentiated behaviour were highlighted, namely, (i) surfaces with erosion and drying crusts in transition, characterized by a low infiltration capacity and a strong run-off attributable to the existence of a plasmic layer playing the role of hydraulic barrier and, (ii) surfaces covered with a drying crust, characterized by a good hydraulic conductivity and a high infiltration capacity related to the absence of a plasmic micro-layer.

The monitoring of surface hydraulic properties of the studied soils showed a progressive deterioration which frequently leads to strongly degraded soils unable to support vegetation. To prevent such a degradation and durably improve soil properties, various measures should be implemented, in particular, the control of the run-off and erosion, the sustainable improvement of the soil structure (supply of organic and/or chemical amendments), the revitalization of the soil surface layer by the addition of manure or other organic substances and the supply of the essential mineral complements for a correct biomass production.

ACKNOWLEDGMENTS

The research was carried out at the Swiss Federal Institute of Technology of Lausanne and at the International Institute for Water and Environmental Engineering of Ouagadougou and financed by a grant from the Swiss Agency for Development and Cooperation (SDC).

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