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An investigation of the effects of spatial heterogeneity of initial soil moisture content on surface runoff simulation at a small watershed scale

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SUMMARY

In addition to the soil saturated hydraulic conductivity, K_s, the initial soil moisture content, θ_i, is the quantity commonly incorporated in rainfall infiltration models for simulation of surface runoff hydrographs. Previous studies on the effect of the spatial heterogeneity of initial soil water content in the generation of surface runoff were generally not conclusive, and provided no guidance on designing networks for soil moisture measurements. In this study, the role of the spatial variability of θ_i at the small watershed scale is examined through the use of a simulation model and measurements of θ_i. The model combines two existing components of infiltration and surface runoff to model the flow discharge at the watershed outlet. The observed values of soil moisture in three experimental plots are combined to determine seven different distributions of θ_i, each used to compute the hydrographs produced by four different rainfall patterns for two initial conditions classified as "dry" soil and "wet" soil. For rainfalls events typically associated with floods, the spatial variability of θ_i at the watershed scale does not cause significant variations in surface runoff for initially dry or wet soils. Furthermore, when the main objective is to represent flood events a single ground point measurement of θ_i in each area with the same land use may suffice to obtain adequate outflow hydrographs at the outlet.

1. Introduction

Theoretical and experimental investigations of soil moisture content as a function of time have been recently carried out to achieve an acceptable representation of many hydrological processes at different spatial scales (Corradini, 2014; Korres et al., 2015), from the local, or point, (Romano, 2014) to the field (Vereecken et al., 2008, 2014; Penna et al., 2009; Zehe et al., 2010; Ojha et al., 2014; Martini et al., 2015) to the basin scale (Fang and Lakshmi, 2014; Schröter et al., 2015). Soil moisture influences, for example, the water vapor supply to the atmosphere through the evaporation and evapotranspiration processes from the earth surface, recharge of aquifers, sub-surface transport of pollutants, timing of irrigation, and rainfall-runoff transformation. This paper is focused on the role of heterogeneity in initial soil moisture in the production of surface runoff. For light to moderate rainfall events, surface runoff is generally affected by a significant loss due to infiltration, which is typically expressed as a function of rainfall rate, r, soil saturated hydraulic conductivity, K_s, and initial soil moisture content, θ_i, prior to a rainfall event. In this context, in the mathematical representation of the rainfall-runoff transformation at the field/watershed scale the infiltration process should be described considering the spatial heterogeneity of r, K_s and θ_i.

Several studies showed that K_s can be represented as a random field characterized by a lognormal probability density function (Warrick and Nielsen, 1980; Sharma et al., 1987), and this variability influences the hydrological response of a slope to a uniform rainfall rate (Binley et al., 1989a,b; Saghafian et al., 1995; Corradini et al., 1998, 2011). Furthermore, formulations of the areal-average infiltration for K_s as a single spatial variable (Smith and Goodrich, 2000; Govindaraju et al., 2001; Corradini et al., 2002) and for a joint spatial variability of K_s and r (Wood et al., 1986; Castelli, 1996; Govindaraju et al., 2006; Morbidelli et al., 2006) were also proposed. The dominant role of the heterogeneity of K_s in the latter studies was also emphasized for frontal rainfalls with coefficient of variation of r (CVr) considerably less than CV of K_s (CVK_s).

The role of the spatial variability of K_s with respect to infiltration and runoff seems to be well-understood, while that of θ_i needs

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further investigation though CV of θ_i (CVθ_i) is smaller than CVK_s in most fields. For example, Brocca et al. (2009a, 2010) examined the spatial distribution of θ_i in a few plots under natural conditions and showed that θ_i could be assumed as a random variable characterized by a normal probability density function, limited to positive values, with CVθ_i approximately equal to 0.1. In addition, it is recognized that the values of CVK_s in natural soils are typically in the range (0.3–1.0) (Nielsen et al., 1973; Sharma et al., 1980; Smettem and Clothier, 1989; Ragab and Cooper, 1993). A numerical analysis of the effects of a joint spatial heterogeneity of K_s and θ_i was made by Hu et al. (2015) who showed that runoff was more strongly influenced by the K_s variability.

Grayson et al. (1995) compared runoff simulations for a micro-watershed (2 m^2) using two spatial distributions of initial soil moisture characterized by the same statistical properties. Very different responses to the same rainfall pattern were obtained for a random or an organized θ_i-field. Merz and Plate (1997) examined the dependency of runoff on the spatial organization of θ_i and soil hydraulic properties and found them important for medium rainfall events. The effects of the spatial heterogeneity of θ_i on surface runoff generation at the small watershed scale (6.3 km^2) were also investigated by Bronstert and Bardossy (1999) using different spatial distributions of θ_i obtained through interpolation and stochastic methods. Bronstert and Bardossy (1999) noted that spatial variability of θ_i influenced surface runoff, especially when it was a small fraction of rainfall. On the other hand, experimental investigations using indicative values of initial moisture conditions derived by satellite and/or few ground point measurements were found to be sufficient for use in rainfall-runoff simulation models (Goodrich et al., 1994; Grayson and Western, 1998; Aubert et al., 2003; Brocca et al., 2009b). The role of the spatial variability of θ_i on the estimate of surface runoff at the field scale was also examined by Morbidelli et al. (2012), who found rainfall rate and average initial soil water content to be important factors. For heavy rainfall rates the effects of spatial variability of θ_i on surface runoff could be disregarded, but for rainfall events of low intensity over high average soil moisture contents, the effects could be appreciable, and become marked when the latter reduces to very low values. However, cases with low values of both r and average θ_i produce small amounts of surface water and are generally of minor interest in applied hydrology.

An overall analysis of the aforementioned results suggests that the effects of the spatial variability of θ_i on surface runoff generation are not clearly understood because of the differences in the selected simulation approaches as well as in the spatial scales and rainfall patterns.

The main objectives of this paper are (1) to study the link among the simulation approach, spatial scale and rainfall characteristics, (2) to examine the errors that could be incurred due to scarce sampling of θ_i and how that affects the hydrologic responses of a small watershed to different rainfall patterns. A conceptual/semi-analytical model that combines a point infiltration model for erratic rainfall (Corradini et al., 1997) with a kinematic wave model based on a similarity profile for flow depth over overland regions and stream reaches (Govindaraju et al., 1999) is used here. These components were tested individually and provided accurate results. Simulations have been carried out starting from the results obtained by Morbidelli et al. (2012) at the field scale using measurements of θ_i performed by Brocca et al. (2010) at the local scale. The hydrological response at the watershed scale has been obtained by schematizing the watershed by a network of planes and channels as in Fig. 1 (Hager, 1984; Melone et al., 1998). Simulations have been mainly performed considering θ_i as a random variable and K_s constant through the watershed, besides for the sake of completeness the role of a joint spatial heterogeneity of θ_i and K_s has been shortly investigated.

2. Modeling approach

Simulation of hydrological response at the watershed outlet requires the representation of infiltration, effective rainfall-surface runoff transformation and water routing through the channel networks.

The basic model is set up schematizing a real watershed by a network of planes and channels (see Fig. 1) with θ_i uniform in each plane, but varying from plane to plane, and K_s and r spatially invariable. The models for point infiltration and surface runoff were described in previous studies, and are summarized below for completeness.

To emphasize the specific role of the spatial variability of θ_i, as a first approximation, the random variability of K_s at the field (plane) scale is disregarded. At this scale, Morbidelli et al. (2012) showed that the surface runoff hydrograph at the outlet can be well-approximated simplifying the field of θ_i through the value observed in a site characterized by temporal stability or using, in cases of practical hydrological interest, a value of θ_i observed at the field scale. Thus, to investigate the role of the spatial heterogeneity of θ_i on surface runoff production at the small watershed scale, a spatially uniform value of θ_i is assumed in each plane.

2.1. Point infiltration equations

Following Corradini et al. (1997), the infiltration process is represented combining the depth-integrated Darcy law and the continuity equation under the assumptions of θ_i invariant with depth, z, and dynamic wetting profile, $\theta(z)$, having the shape of a distorted rectangle characterized by a parameter p and a shape factor β ($\leqslant 1$) linked with the water content, θ, at the surface. The resultant ordinary differential equation applicable at each time, t, for any rainfall pattern is:

$$\frac{d\theta_0}{dt} = \frac{(\theta_0 - \theta_i)\,\beta(\theta_0)}{I'\left[(\theta_0 - \theta_i)\frac{d\beta(\theta_0)}{d\theta_0} + \beta(\theta_0)\right]}\left[q_0 - K_0 - \frac{(\theta_0 - \theta_i)\,G(\theta_i, \theta_0)\,\beta(\theta_0)\,p K_0}{I'}\right] \tag{1}$$

where the subscript 0 stands for quantities at the soil surface, q denotes the downward water flux, I' is the cumulative dynamic infiltration depth and G is the net capillary drive depending on the capillary head, ψ, and hydraulic conductivity, K, as:

$$G(\theta_i, \theta_0) = \frac{1}{K_0}\int_{\psi(\theta_i)}^{\psi(\theta_0)} K(\psi)\,d\psi \tag{2}$$

The functional forms of $K(\psi)$ and $\psi(\theta)$ are:

$$K(\psi) = K_s\left[1 + \left(\frac{\psi - d}{\psi_b}\right)^c\right]^{-(3\lambda+2)/c} \tag{3}$$

$$\psi(\theta) = \psi_b\left[\left(\frac{\theta - \theta_r}{\theta_s - \theta_r}\right)^{-c/\lambda} - 1\right]^{1/c} + d \tag{4}$$

where ψ_b is the air entry head, given for soil texture classes by Rawls et al. (1983); θ_s and θ_r are the volumetric soil water contents at natural saturation and residual, respectively; and c, λ and d are empirical coefficients. Starting from rainfall with $r > K_s$ over an unsaturated soil surface, when $\theta_0 = \theta_s$, because $q_0 = r$ and $d\theta_0/dt = 0$, Eq. (1) provides time to ponding, t_p, as:

$$\int_0^{t_p} r\,dt = \frac{(\theta_s - \theta_i)\,G(\theta_i, \theta_s)\,\beta(\theta_s)\,p K_s}{r - K_s} \tag{5}$$

For $t > t_p$, Eq. (1) is solved to obtain the infiltration capacity, $f_c = q_0$, until $f_c \leqslant r$ as: