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Initial soil moisture effects on flash flood generation – A comparison between basins of contrasting hydro-climatic conditions

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SUMMARY

The purpose of this paper is to contribute to the understanding of the importance of the initial soil moisture state for flash flood magnitudes. Four extreme events that occurred in different case study regions were analysed, one winter and one autumn flash flood in the Giofiros and Almirida catchments in Crete, and two summer floods in the Rastenberg catchment in Austria. The hydrological processes were simulated by the spatially distributed flash flood model Kampus. For the Crete cases Kampus model was calibrated against remotely sensed soil moisture while for the Austrian case the model was calibrated against observed runoff. Kampus model was then used to estimate the sensitivity of the stream flow peak to initial soil moisture. The largest of the events analysed (in terms of specific peak discharge) was found to have a sensitivity of less than 0.2% flood peak change per % soil moisture change while the smallest event had a sensitivity of more than 3% flood peak change per % soil moisture change. This suggests that initial soil moisture effects on the flash flood response probably depend on event magnitude rather than on the climate or region. Moreover, the Austrian catchment was found to exhibit a more nonlinear relationship between antecedent soil moisture and the peak discharge than the Cretan catchments which was explained by differences in the soil type.

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1. Introduction

Europe has experienced numerous catastrophic flash floods in the last decades with vast social and economic impacts on the affected areas (Hall et al., 2013; Merz et al., 2014). The frequency and the magnitude of the flash flood events have differed between Continental and Mediterranean regions in Europe, with a tendency of the latter to produce more extreme floods (Gaume et al., 2009; Norbiato et al., 2009).

There are a number of factors that affect the severity of floods including precipitation intensity, percentage of sealed catchment area, soil permeability, water holding capacity, topographic slopes and soil moisture content soil at the beginning of the event. In contrast to the other characteristics that do not change much between events, soil moisture can vary significantly, even on a sub daily time scale. Soil moisture can vary from near to the wilting point to saturation. It is considered as the most important soil factor for rapid runoff and flash flooding. Saturated soils obstruct precipitation to infiltrate, resulting in higher runoff regardless other environmental conditions. Soil moisture can, in fact, control whether a given rainstorm produces a major flash flood or not, due to the non-linear nature of runoff response to rainfall (Hlavcova et al., 2005; Komma et al., 2007; Zehe and Blöschl, 2004). In the framework of flood warning systems, the knowledge of soil moisture is crucial (Georgakakos, 2006; Javelle et al., 2010; Lacava et al., 2005; Raynaud et al., 2015; Van Steenbergen and Willems, 2013). Hence, it is essential to capture antecedent soil moisture well for flood forecasting applications (Berthet et al., 2009; Yatheendradas et al., 2008).

Michele and Salvadori (2002) evaluated the influence of antecedent soil moisture conditions on the flood frequency distribution based on derived distribution theory. Based on an analysis of soil moisture from point (in-situ) to footprint (remote sensing) scale (Joshi et al., 2011) concluded that soil properties and topography are the most significant physical parameters that jointly control the spatio-temporal evolution of soil moisture. In a similar study on an experimental catchment, (Nasta et al., 2013) found that, during wet periods, catchment topography is an important factor of

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spatial soil moisture distribution whereas, during dry periods, it depend primarily on soil hydraulic properties. This is consistent with the earlier findings of (Grayson et al., 1997) in the Tarrawarra catchment. Based on model simulation, (Yoo et al., 1998) found that, during the rainfall event, soil texture plays a greater role for the soil moisture evolution than rainfall variability although this finding may be contingent on their particular catchment conditions (Viglione et al., 2010).

In order to assess soil moisture effects on flood magnitudes, soil moisture is needed at the catchment scale. However this is difficult to measure on an in-situ basis. An alternative is soil moisture retrieval by satellite sensors such as the series of passive multi-frequency radiometers (SMMR, Windsat, AMSR-E, etc. – see (de Jeu et al., 2008)) for which a long record of continuous measurements starting in 1978 is available, and the series of active microwave scatterometers, which begun with launch of the ERS 1scatterometerin 1991 (Naeimi et al., 2009; Wagner et al., 1999). The Advanced Scatterometer (ASCAT) started in 2007 (MetOp-A and MetOp-B) and continues to operate to date (Bartalis et al., 2007).

This paper investigates the importance of initial soil moisture for the flood magnitude in catchments of different geomorphology and hydro-meteorological regimes by making use of satellite-retrieved soil moisture and hydrological modelling.

2. Methodology

2.1. ERS scatterometer soil moisture data

The ERS scatterometer data used here allow the retrieval of surface soil moisture information on the top 2 cm soil layer (Parajka et al., 2006). The top layer of hydrological models, however, is usually considered to be the root zone which is much deeper than the satellite beam can penetrate. A transformation of the ERS data was hence used to account for the time delay as the water infiltrates from the surface into the soil based on a simple linear filter in the time domain (Wagner, 1998). In this method, soil water index (SWI) that should reflect root zone soil moisture is defined:

$$\mathrm{SWI}\ (t) = \frac{\sum_i m_s(t_i) e^{\frac{t-t_i}{T}}}{\sum_i e^{\frac{t-t_i}{T}}} \ \text{for } t_i \leqslant t \tag{1}$$

where m_s is the surface soil moisture estimate from the ERS Scatterometer at time t_i and T is the time constant of the filter. T is related to the hydraulic characteristics of the top soil with more permeable soils being associated with smaller T because of the faster infiltration. However, at the pixel scale, this relationship is difficult to identify based on soil characteristics and can best be obtained by backcalculation from terrestrial soil moisture data and/or hydrological models (Parajka et al., 2009; Wagner, 1998). The SWI is calculated if there are at least one ERS Scatterometer measurement in the time interval $[t, t-T]$ and at least three measurements in the interval $[t, t-5T]$. Following (Brocca et al., 2012, 2010; Wagner, 1998; Wagner et al., 1999) a time constant of $T = 20$ days is well suited for the present application.

2.2. The hydrological model Kampus

The Kampus model used in this study is a spatially-distributed continuous rainfall–runoff model (Bloschl et al., 2008; Viglione et al., 2010). Kampus model uses a 15 min time step and consists of a snow routine, a soil moisture routine and a flow routing routine. The snow routine represents snow accumulation and melt by the degree-day concept. The soil moisture accounting routine is the main part controlling runoff formation. It represents the runoff generation and changes in the soil moisture state of the catchment and involves three parameters: the maximum soil moisture storage L_s, a parameter representing the soil moisture state above which evaporation is at its potential rate, termed the limit for potential evaporation L_P, and a parameter in the non-linear function relating runoff generation to the soil moisture state, termed the non-linearity parameter *ß*. Runoff routing on the hillslopes is represented by an upper and two lower soil reservoirs. Excess rainfall Q_p enters the upper zone reservoir and leaves this reservoir through three paths, outflow from the reservoir based on a fast storage coefficient k_1; percolation to the lower zones with a percolation rate c_P; and, if a threshold of the storage state L_1 is exceeded, through an additional outlet based on a very fast storage coefficient k_0. Water leaves the lower zones based on the slow storage coefficients k_2 and k_3. Bypass flow Q_{by} is accounted for by recharging the lower zone reservoir (k_2) directly by a fraction of the excess rainfall. k_1 and k_2 as well as c_P have been related to the soil moisture state in a linear way. The outflow from the reservoirs represents the total runoff Q_t on the hillslope scale. These processes are represented on a 1 km $\times$ 1 km grid. Kampus model states for each grid element are the snow water equivalent, soil moisture S_s of the top soil layer, the storage of the soil reservoirs S_1, S_2, S_3 associated with the storage coefficients k_1, k_2, k_3, with $k_1 < k_2 < k_3$.

In the soil moisture routine, the sum of rain and melt, $P_r + M$, is split into a component dS that increases soil moisture of a top layer, S_s, and a component Q_p that contributes to runoff. The components are split as a function of S_s:

$$Q_p = \left(\frac{S_s}{L_s}\right)^{\beta} \cdot (P_r + M) \tag{2}$$

L_s is the maximum soil moisture storage, β controls the characteristics of runoff generation and is termed the non-linearity parameter. If the top soil layer is saturated, i.e., $S_s = L_s$, all rainfall and snowmelt contributes to runoff and dS is 0. If the top soil layer is not saturated, i.e., $S_s < L_s$, rainfall and snowmelt contribute to runoff as well as to increasing S_s through $dS > 0$:

$$\begin{array}{ll} dS = P_r + M - Q_p - Q_{\mathrm{by}} & \text{if } \ P_r + M - Q_p - Q_{\mathrm{by}} > 0 \\ dS = 0 & \text{otherwise} \end{array} \tag{3}$$

where additionally, bypass flow Q_{by} is accounted for. The effect of the soil routine is that the contribution of precipitation to runoff is small when the soil is dry (low soil moisture values), while it becomes larger in wet soil conditions.

Analysis of the runoff data of the Rastenberg catchment indicated that flow that bypasses the soil matrix and directly contributes to the storage of the lower soil zone is important for intermediate soil moisture states S_s. For $\xi_1 \cdot L_s < S_s < \xi_2 \cdot L_s$ (with $\xi_1 = 0.4$, $\xi_2 = 0.9$) bypass flow was assumed to occur as:

$$\begin{array}{ll} Q_{\mathrm{by}} = \alpha_{\mathrm{by}} \cdot (P_r + M) & \text{if } \ \alpha_{\mathrm{by}} \cdot (P_r + M) < L_{\mathrm{by}} \\ Q_{\mathrm{by}} = L_{\mathrm{by}} & \text{otherwise} \end{array} \tag{4}$$

while no by pass flow was assumed to occur for dry and very wet soils. Changes in the soil moisture of the top soil layer S_s from time step $i-1$ to i are accounted for by

$$S_{s,i} = S_{s,i-1} + (dS - E_A) \cdot \Delta t \tag{5}$$

The only process that decreases S_s is evaporation E_A which is calculated from potential evaporation, E_P, by a piecewise linear function of the soil moisture of the top layer:

$$\begin{array}{ll} E_A = E_P \cdot \frac{S_s}{L_P} & \text{if } \ S_s < L_p \\ E_A = E_P & \text{otherwise} \end{array} \tag{6}$$

where L_p is a parameter termed the limit for potential evaporation. Potential evaporation was estimated by the modified Blaney–Criddle method as a function of air temperature.

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