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Assessing the effect on flood frequency of land use change via hydrological simulation (with uncertainty)

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Abstract

The present work aims to assess the effects that land use change, which have affected a wide portion of Italy in the second part of the last century, have induced on the flood frequency regime. The analysis has been carried out by applying a spatially distributed rainfall–runoff model to generate synthetic river flow series. Reference is made to different historical land use scenarios of the Samoggia River basin (178 km^2 in area), located in the Apennines mountains (Italy). The man induced extensive land use changes, which have affected the drainage basin during the last decades, have been assessed by using land use maps which refer to the years 1955, 1980 and 1992. By simulating the Samoggia River flood flows for the different land use scenarios, the variations of the peak flow regime, due to the increasing human activity, have been investigated. Finally, these variations have been compared with the magnitude of the simulation uncertainty. The results indicate that the sensitivity of the floods regime to land use change decreases for increasing return period of the simulated peak flow. By considering peak flows with return period ranging from 10 to 200 years, the effects of human activity seem to be noteworthy. However, these results are to be interpreted cautiously, in light of the uncertainty that affects hydrological simulation studies.

Keywords: Land use change; Flood frequency; Peak flow; Distributed models; Stochastic processes; Uncertainty

1. Introduction

In the last decades many inundations have occurred in Europe causing loss of human lives and financial damages, which have been aggravated, in several cases, by the intense urbanisation of flood prone areas. Considering the Italian case and some of the most significant events only, the Po River has been affected by two remarkable floods (1994 and 2000) in a few years, and catastrophic inundations have occurred in Piedmont (1994 and 2000), Valle d'Aosta (2000), Tuscany (1996), Liguria (2000) and Calabria (2000).

After these disasters, the question has often been raised about the possibility that they are due, at least partially, to an increased vulnerability as a consequence of the land use change that took place in northern Italy in the last five decades. These concerns have therefore urged the hydrologic community to

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investigate the effect of human activities on the river flow regime.

Since the end of the 1960s, the scientific literature reported the results of studies about the possible effects on the fluvial regime of intense deforestation and urbanisation, which had occurred in some drainage basins of the United States (Patric and Reinhart, 1971). Similar analyses were performed in order to assess the effects that intensive land use change, such as the construction of roads in mountain lands, can induce on the river runoff (Bannister, 1979). These studies highlighted, as expected, that the hydrological effects of human activity are strictly dependent on the extension of the area affected by the man made intervention. For instance, referring to a tropical watershed, Costa et al. (2003) found that a deforestation of about 30% of the basin induced a 24% increment of the annual mean discharge. Clearly, the effects of land use changes on flood peaks depend on the nature of the examined flood event. In this regard, Hollis (1975) underlined that the effect of the urbanisation on the peak discharge is less remarkable for increasing return period of the event. Such a conclusion is justified considering that the extreme flood events are caused by storms which induce soil saturation, and therefore the reduction of the soil storativity affects the surface discharge to a smaller extent. In this regard, an interesting result was obtained by Niehoff et al. (2002), who found that the influence of land use conditions on storm runoff generation is only relevant for convective storms with high precipitation intensities, in contrast with long lasting advective storms with low rain intensities. The effect of the climate was studied also by Bultot et al. (1990), who analysed the influence of the land use on the water balance of the near surface soil layer by applying a conceptual rainfall–runoff model to a Belgian river basin. They concluded that the presence of vegetation can induce effects on the river flows that are more evident in arid climates, where the vegetation cover causes a reduction of the river discharge that is more marked for the lower river flows. The higher sensitivity to land use change of the low flows has been confirmed by simulations studies performed by Brath and Montanari (2000). Recently, Naef et al. (2002) came to a similar conclusion by referring to a river basin located in Germany. They in fact found that the flood runoff reduction by land use change can be remarkable in the presence of rapid runoff production only.

In the last two decades, several analyses have been carried out to estimate the effects on the hydrological cycle induced by changes in the vegetation cover, with particular emphasis on the consequences of deforestation (Bosch and Hewlett, 1982), conversion of wooded areas to pasture (Peck and Williamson, 1987), construction of roads in forests, compaction of soils caused by the transit of heavy agricultural vehicles (Luft et al., 1982; Ranzi et al., 2002) and urbanisation of bottom valley areas (Beighley and Moglen, 2002). Briefly, these studies came to the conclusion that the sensitivity of the river flows to land use change appeared to be strongly dependent on the climatic behaviour and the geomorphologic characteristics of the river basin.

It is interesting to note that most of the above studies agree in recognising that the effects on peak flows of human activity have often been less significant than expected. Since their magnitude strictly depends on the local behaviour of the watershed, climate, and man made interventions, they can be evaluated by developing an at site analysis only, which should be carried out by using modelling tools that are capable of exploiting the hydrological effects of land use change.

In light of these considerations, the present work aims to propose a methodology for the assessment of the effects on the flood frequency regime of human activity, by developing a case study which refers to the Samoggia River basin, located in northern Italy. The wide availability of historical rainfall, hydrometric and land use data enabled an accurate calibration of the models that have been used for performing a long-term simulation of hourly river flows, referring to different historical land use scenarios.

2. The Samoggia River basin

The Samoggia River is located in northern Italy and it is a left bank tributary to the Reno River (Fig. 1). The total area of the basin, closed at the river cross-section of Calcara, is 178 km^2. The drainage basin is constituted mainly by mountain areas; the maximum altitude is 850 m above sea level (a.s.l.),

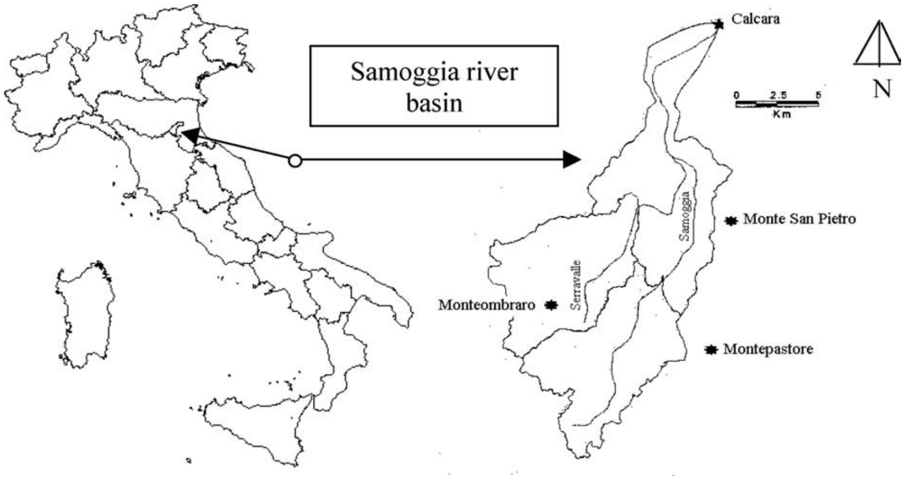

Fig. 1. Location of the Samoggia River basin and the rainfall and river flow gauging stations.

while the main stream length is 60 km. The mountain areas are mainly composed by soils and rocks of sedimentary origin, which are mainly covered by broad leaved woods. The bottom valleys are mainly floodplains for the most part covered by farmlands and urbanised. Because of their low permeability and their extension with respect to the total drainage basin surface, mountain areas give a remarkable contribution to the formation of flood flows, which are principally generated by infiltration excess runoff. The maximum peak discharge observed at Calcara in the period 1938–1997 is 452 m^3/s (year 1940); the annual rainfall depth averaged over the basin area and over the period 1959–1977 is 938 mm, while the average runoff coefficient, for the same period, is equal to 0.37.

The topography of Samoggia River basin is described by a Digital Elevation Model (DEM), shown in Fig. 2, whose resolution is 250×250 m. The hillslopes are significantly steep. In fact, 44% of the contributing area has a slope comprised between 10 and 20%.

Historical available data are constituted by hourly rainfall depths that have been observed for the four-year period 1994–1997 in three different raingauges located at Monte San Pietro (317 m a.s.l), Montepastore (596 m a.s.l.) and Monteombraro (727 m a.s.l.), as shown in Fig. 1. 59 observations of annual maximum rainfall depth for storm duration of 1, 3, 6, 12 and 24 h are also available for Monteombraro raingauge and have been collected during the period 1938–1997 (except 1943). Historical data of hourly river discharges recorded at Calcara are also available for the years 1994–1997, together with 33 annual maximum peak flows observed in the period 1938–1997 (many observations are missing). At last, hourly temperature data recorded at Monteombraro in the years 1994–1997 are at disposal. An extensive data base of soil texture, relative permeability and organic substance content has been derived from the soil map, at 1:250,000 scale and in digital format, provided by the local administration.

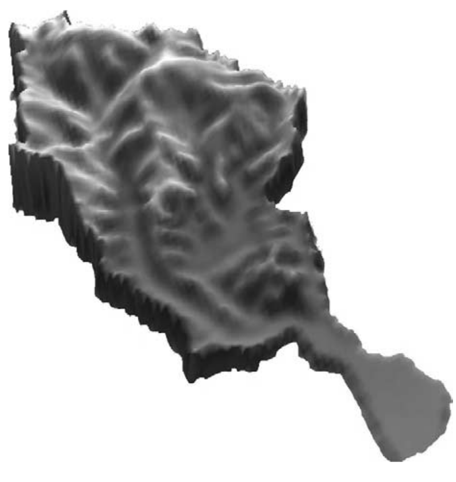

Fig. 2. Digital elevation model of the Samoggia River basin.

The soil use of the Samoggia River basin has been derived, at each DEM cell scale, from a series of surveys which refer to the years 1955, 1980 and 1992. The first two surveys have been carried out by means of field investigations, which were developed as part of the RAISA project (Advanced Research for the Agricultural System Innovation, see Bertozzi et al., 1992) of the National Research Council of Italy. Two soil maps have been produced, both at 1:50,000 scale, which cover the district of Bologna. The third map, which refers to the year 1992, has been produced within the CORINE Land Cover project, funded by the European Community, by means of photo interpretation of remote sensing images.

3. Analysis of the land use change on the Samoggia River basin in the last five decades

In the last decades, considerable changes in land use have occurred in the various regions of Italy. As a quite general rule, from the early 1950s, the meadows and pastures, which characterised most of the upland regions at that time, decreased as increasing industrialisation and urbanisation led to the abandonment of cattle farming. Hence, these upland areas were left fallow and unproductive. A similar behaviour occurred in the considered basin. In order to investigate the land use change that occurred in the Samoggia River basin in the last five decades, one may consider the two historical land use maps of the district of Bologna, referred to the years 1955 and 1980, which are comparable since they were derived from field surveys carried out with similar procedures. The original 10 land use classes adopted by the RAISA project were merged into seven major classes: surface water; fallow and unproductive areas; forestlands and woods; meadows and pastures; crops; arboreal cultivation; urbanised areas. The land use changes are summarised in the bar diagram of Fig. 3. The height of each bar segment is proportional to the relative areal extent of a given land use type so that the heights of the different sectors sum to 100%. While only slight changes can be observed in Fig. 3 when considering the areal extent of surface water and woods and crops, meadows and pastures fell from 23.7% in 1955 to 3.0% in 1980. This reduction was balanced by the increase in arboreal cultivation, unproductive areas and, only marginally, by the growth of urbanised areas. Although these latter increased by more than 500% over the 25 years, their total extent was still less than 5% of the entire district in 1980. More details can be found in Brath et al. (2002).

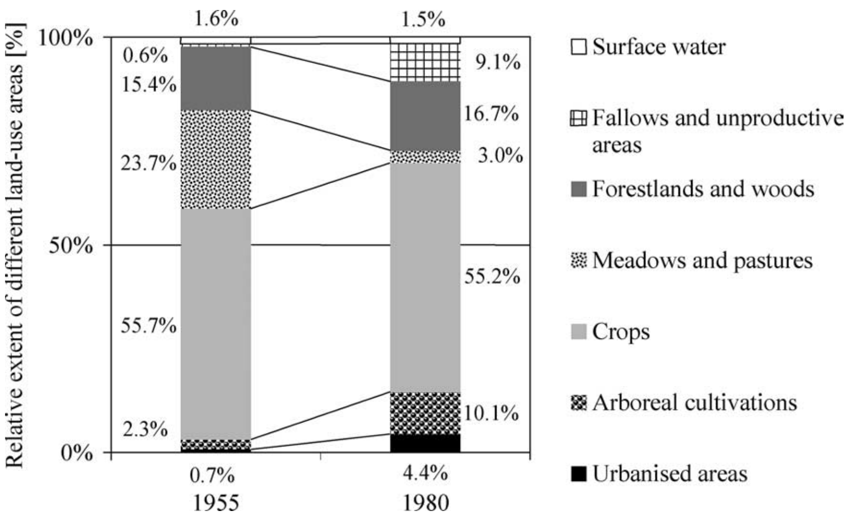

Fig. 3. Land use changes in the district of Bologna between 1955 and 1980 (data derived from the RAISA of the National Research Council of Italy).

4. Brief outline of the simulation procedure

The simulation study is based on the use of a spatially distributed rainfall–runoff model, which operates continuously in time and whose parameters were calibrated by using the observed rainfall and river flow.

As a first step of the simulation procedure, the available four-year historical hourly rainfall and temperature series have been used as input data into the model. By performing two successive simulations, which refer to the land use scenarios of 1955 and 1992, respectively, preliminary indications on the effects of the human activity on peak discharges have been drawn. Nevertheless, the limited extension of the historical series does not allow one to exhaustively assess the effects on the flood frequency distribution. Therefore 1000 years of synthetic hourly temperature and rainfall data have been generated with stochastic models, whose parameters have been calibrated by means of the historical available data. Using the synthetic series as input to the rainfall–runoff model, 1000 years of river flows have been simulated, making reference to the two soil use scenarios. This allowed us to derive indications about the effects of land use change on the flood frequency distribution.

The synthetic hourly rainfall data have been generated by means of the generalised multivariate Neyman–Scott rectangular pulses model (Cowpertwait, 1996). This approach has been selected because of its efficiency in the reproduction of the statistical properties of the rainfall records, which has been proven by numerous previous works. The model was calibrated on a monthly basis therefore preserving the seasonality of the model parameters. A detailed description of the model is given by Cowpertwait (1996).

The generation of the synthetic hourly temperature record referring to Monteombraro, where the historical data are available, was carried out by applying a fractionally differenced ARIMA model (FARIMA). On many occasions, this class of models turned out to be able to fit the autocorrelation structure of temperature series, that is very often affected by a slow decay, which may suggest the presence of long-term persistence, implying this way the presence of the Hurst Effect (Montanari, 2003). The seasonality of the temperature record was preserved by performing a preliminary deseasonalisation procedure, which consists in removing from the data a periodic seasonal component, which is added back at the end of the simulation procedure. More details on FARIMA models and the simulation procedure herein applied can be found in Montanari et al. (1997).

5. The rainfall–runoff model

The rainfall–runoff process has been modelled by means of a spatially distributed continuous simulation approach. The model has been set up at the University of Bologna, utilising an existing algorithm for the extraction of the river network (Mancini, 1990). In order to limit the information needed for the calibration of the model parameters, many of the hydrological processes have been schematised by using conceptual approaches. Consequently, parameter calibration has to be carried out on the basis of some historical hydrometeorological records. However, recent researches (Brath et al., 2001) proved the efficiency and the robustness of the model even when applied to catchments characterised by a limited availability of historical data sets. Such efficiency is believed due to the capability of the model to take advantage of the spatially distributed description of the basin topography, soil type and use.

The model discretises the basin in square cells coinciding with the pixel of the Digital Elevation Model (DEM). The river network is automatically extracted from the DEM itself by applying the D-8 method (Band, 1986). A preprocessing step fills the pits by raising the elevation of the corresponding cells of an appropriate amount. The network determination is then carried out by assigning a maximum slope pointer to each DEM cell, which enables the determination of the outflow direction. Following the water paths, from each cell to the basin outlet, the river network can be identified and the contributing area to each cell is computed. Distinction between hillslope rill and network channel is based on the concept of constant critical support area (Montgomery and Foufoula-Georgiou, 1993). Accordingly, rill flow is assumed to occur in each cell where the upstream drainage area does not exceed the value of 0.5 km^2.

The interaction between soil, vegetation and atmosphere is modelled by means of a spatially distributed conceptual scheme. The model firstly computes the local rainfall $P_l[t,(i,j)]$, for each cell of coordinates (i,j), by interpolating the observed rainfall relative to each raingauge through an inverse distance method. It is assumed for each cell that a first rate of the rainfall depth is accumulated in a local reservoir (interception reservoir), which simulates the interception operated by the vegetation. The capacity of such interception reservoir is equal to $\boldsymbol{C_{inf}}S(i,j)$, where C_{inf} is a parameter (constant with respect to both space and time) and $S(i,j)$ is the local soil storativity in correspondence of the cell. This latter is computed accordingly to the Curve Number method (CN method, Soil Conservation Service, 1972), depending on soil type and use, permeability and soil texture.

The exceeding rainfall reaches the ground once the interception reservoir is full of water. Then the separation between surface and sub surface flow and the computation of the infiltration are carried out at local scale by applying a modified CN approach, which enables one to simulate the redistribution of the soil water content during interstorm periods and, therefore, to perform continuous simulation. In detail, it is assumed that a linear reservoir (infiltration reservoir) is located in correspondence of each DEM cell and at the soil level. This reservoir collects the infiltrated water and empties by means of the bottom outflow, from which the water flows from the surface soil layer to the subsurface one. This latter is assumed to be crossed by a subsurface drainage network that links the cells and follows the same paths of the surface network. The rate of the rainfall depth which contributes to the local surface runoff is computed in accordance with the relationship

$$P_n[t,(i,j)] = P[t(i,j)]\frac{F[t,(i,j)]}{HS(i,j)} \tag{1}$$

where $P[t,(i,j)]$ is the intensity of rainfall which reaches the ground at time t, $P_n[t,(i,j)]$ is the intensity of surface runoff, $F[t,(i,j)]$ is the storage at time t of the infiltration reservoir located in the correspondence of the cell (i,j) and $HS(i,j)$ is the capacity of the infiltration reservoir itself. This capacity is computed by multiplying a parameter H, constant with respect to both space and time, for the above defined soil storativity $S(i,j)$. The outflow from the infiltration reservoir is computed by means of the linear relationship

$$W[t,(i,j)] = \frac{F[t,(i,j)]}{\boldsymbol{H_s}} \tag{2}$$

where $\boldsymbol{H_s}$ is a parameter that is assumed to be constant with respect to both space and time.

The water contents of both the interception and infiltration reservoirs are updated in order to account for the evapotranspiration process. Let us denote with $\mathrm{IF}[t,(i,j)]$ the water content of the interception reservoir at time t in the cell (i,j). First, the model computes the hourly intensity of potential evapotranspiration $\boldsymbol{E_p}[t,(i,j)]$, for each DEM cell, as a function of the local temperature by applying the radiation method (Doorembos et al., 1984). If $\mathrm{IF}[t,(i,j)] \geq \boldsymbol{E_p}[t,(i,j)]$ then $\mathrm{IF}[t,(i,j)]$ is updated accordingly to the relationship

$$\mathrm{IF}[t+1,(i,j)] = \mathrm{IF}[t,(i,j)] - \boldsymbol{E_p}[t,(i,j)] \tag{3}$$

If $\mathrm{IF}[t,(i,j)] < \boldsymbol{E_p}[t,(i,j)]$ then $\mathrm{IF}[t+1,(i,j)]=0$ and $F[t,(i,j)]$ is updated accordingly to the relationship

$$F[t+1,(i,j)] = F[t,(i,j)] - \max\{E[t,(i,j)], F[t,(i,j)]\}, \tag{4}$$

where $E[t,(i,j)]$ is varying linearly from 0, when $F[t,(i,j)]=0$, to $\boldsymbol{E_p}[t,(i,j)] - \mathrm{IF}[t,(i,j)]$ when $F[t,(i,j)] = HS(i,j)$.

By combining Eqs. (1) to (3), the mass balance equation for the infiltration reservoir can be written as

$$\frac{\mathrm{d}F[t,(i,j)]}{\mathrm{d}t} = -\frac{F[t,(i,j)]}{H_s} - E^*[t,(i,j)] + P[t,(i,j)]\left\{1 - \frac{F[t,(i,j)]}{HS(i,j)}\right\} \tag{5}$$

where $E^*[t,(i,j)] = \max\{E[t,(i,j)], F[t,(i,j)]\}$ if $\mathrm{IF}[t,(i,j)] < \boldsymbol{E_p}[t,(i,j)]$ and $E^*[t,(i,j)]=0$ otherwise. Eq. (5) is numerically solved by applying the second-order Runge–Kutta method.

Surface and subsurface flows are propagated downstream, first along the hillslope then through the channel network, by applying the variable parameter Muskingum–Cunge model. Extensive details can be found in Cunge (1969) and Orlandini et al. (1999), for the surface and subsurface propagation, respectively. The kinematic celerity of the surface flows is computed by considering rectangular river cross-section with fixed width/height ratio. This latter parameter and the channel roughness can assume different values along the river network and on the hillslopes. In particular, the channel roughness in the river network is allowed to vary from a minimum to a maximum value depending on the contributing area. For the subsurface flows, the kinematic celerity is instead computed as a function of the saturated hydraulic conductivity and the soil slope.

It is worth noting that the model describes through a simplified scheme the dynamics of the subsurface flow. In particular, it does not distinguish between near surface and deep water flow, and assumes that the calibration parameters H and $\boldsymbol{H_s}$ are constant with respect to both space and time. This simplified description has the advantage of reducing the number of model parameters and, consequently, the amount of historical data required for their calibration. On the other hand, one may expect a reduced accuracy in the simulation of subsurface flows, which means a less accurate representation of lower river discharges. Nevertheless, because the analysis is focused on the flood flows, the above mentioned limitation of the model is expected not to significantly affect the results of the current work.

Moreover, the production of the surface runoff is modelled accordingly to a scheme that is similar to the one adopted by the CN method, which is considered by many authors as an infiltration excess model (Hortonian runoff; see e.g. Beven, 2000). Therefore, one may expect that the proposed model is better suited for basins characterised by low permeability and prevalently impervious hillslopes, where excess of infiltration plays generally a more important role than excess of saturation in generating surface runoff.

Table 1 reports a list of the model parameters and indicates which of them have been estimated by considering the characteristics of the drainage basin or by physical reasoning, and which ones have been derived by manual calibration. The latter has been performed by comparing observed and simulated hourly river flows of the flood event occurred on 8 October 1996, referring to the 1992 soil use scenario, which is closer in time to the calibration event. Fig. 4 shows a comparison of the observed and simulated flood hydrographs. The model has been subsequently validated by simulating the hourly discharges of the whole 1994–1997 four-year period. A dispersion

Table 1
Rainfall–runoff model parameters and relative values. These latter have been partly optimised by calibration (calibrated) and partly estimated by considering the characteristics of the drainage basin or by means of physical reasoning (estimated)

Parameter	Symbol and dimension	Method of estimation	Estimated value
Strickler roughness for the hillslope	k_{sv} ($m^{1/3}$ s^{-1})	Calibrated	0.5
Channel width/ height ratio for the hillslope	w_v (–)	Calibrated	600
Maximum and minimum Strickler roughness for the channel network	k^0_{sr}, k^1_{sr} ($m^{1/3}$ s^{-1})	Estimated	10–25
Channel width/ height ratio for the channel network	w_r (–)	Estimated	20
Constant critical source area	A^* (km^2)	Estimated	0.5
Saturated hydraulic conductivity	K_i (m s^{-1})	Calibrated	0.01
Bottom discharge parameter for the infiltration reservoir capacity	$\boldsymbol{H_s}$ (s)	Calibrated	79,095
Multiplying parameter for the infiltration reservoir capacity	H (–)	Calibrated	0.08
Multiplying parameter for the interception reservoir capacity	$\boldsymbol{C_{inf}}$ (–)	Calibrated	0.65

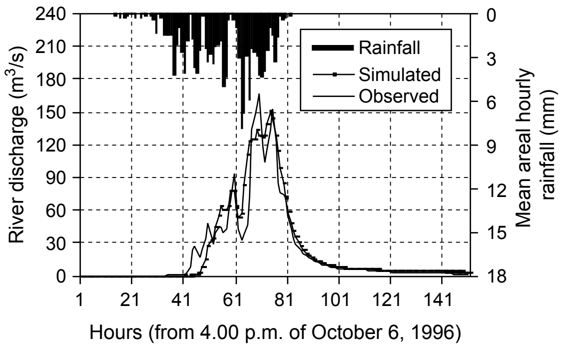

Fig. 4. Comparison between the observed and the simulated hydrographs for the calibration flood event of 8 October 1996.

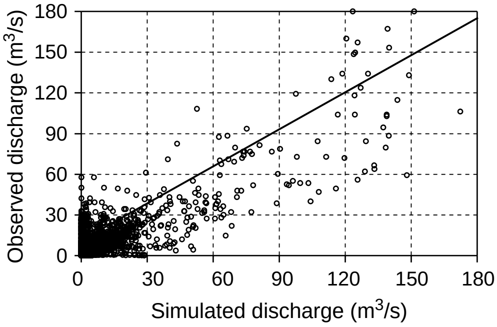

Fig. 5. Dispersion diagram of the observed versus the simulated hourly discharges for the 1994–1997 period.

diagram of the observed versus the simulated 1994–1997 hourly flows is reported in Fig. 5.

The model parameters optimised by calibration and reported in Table 1 assumed physically realistic values. The Nash and Sutcliffe (1970) efficiency for the simulation of the 1994–1997 hourly discharges (validation phase) is 0.68. For the periods 1994 - 1995 and 1996 - 1997, the efficiency amounts to 0.60 and 0.81, respectively. An application of the same model to the Samoggia River basin is presented also by Montanari and Brath (2004). Table 2 shows the mean relative absolute error for the simulation of the 1994–1997 flows and for different thresholds values. It is derived by computing the ratio between the absolute error in the simulation of each data point and the corresponding observation; the result is averaged over all the simulated data points and expressed in percentage terms. It can be seen that the model performances are improving with increasing flows.

6. Application of the simulation procedure

The model has been run twice by using the CN parameter values referred to 1955 and 1992. As previously mentioned, primarily the observed hourly rainfall and temperature data for the period 1994–1996 have been input to the rainfall–runoff model. This simulation is named in the following sections as 'historical data simulation', to distinguish it from the 'synthetic data simulation' that have been carried out later by using, as input variables, the synthetic 1000 year rainfall and temperature series.

Table 2
Mean relative absolute error (%) of the simulation of the 1994–1997 hourly discharges (validation data set) for different threshold values of simulated flow

>1 m^3/s	>10 m^3/s	>30 m^3/s	>50 m^3/s
60.5	51.1	38.7	27.9

6.1. Historical data simulation

Simulating the flood event of 8 October 1996, which has also been previously adopted for the model calibration, one notes that the estimated peak flows relative to the 1955 and the 1992 land use scenarios are 112 and 164 m^3/s, respectively. Fig. 6 shows a dispersion diagram of the 1955 versus the 1992 simulated hourly flows for the whole four-year period. As one can observe, the discharges of 1992 are systematically greater than the 1955 ones. As far as real applications are concerned, this implies that the parameterisation of the rainfall–runoff model, considering the 1955 land use, would introduce an underestimation of the current hydrological response of the Samoggia River basin.

The relative deviation $\varepsilon(t)$, which is defined as

$$\varepsilon(t) = \frac{|Q_{1992}(t) - Q_{1955}(t)|}{Q_{1992}(t)} \quad (6)$$

where $Q_i(t)$ indicates the simulated discharge at time t and year i, is shown in Fig. 7, together with an estimate of the simulation uncertainty that will be described in Section 7. The difference between the discharges in the two scenarios is about 20% on average. Moreover, $\varepsilon(t)$ tends to decrease with increasing discharge, as one can expect on the basis of intuitive considerations (see also

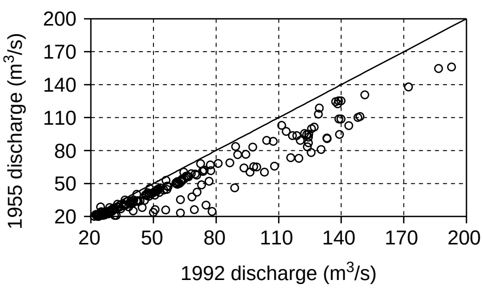

Fig. 6. Historical data simulation. Dispersion diagram of the simulated discharges (>20 m^3/s) for the 1992 versus the 1955 land-use scenario.

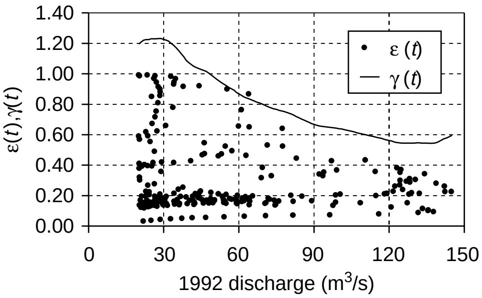

Fig. 7. Historical data simulation. Relative deviation ε (t) (see Eq. (4)) versus the 1992 land-use scenario discharges ($Q_{1992}>20$ m^3/s). The continuous line indicates the width of the 95% confidence bands of the simulation expressed as percentage of the simulated discharge.

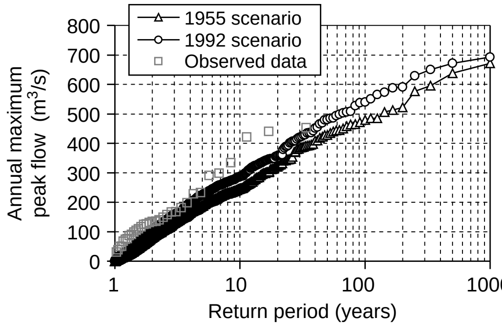

Fig. 8. Synthetic data simulation. Comparison between the 1955 and the 1992 flood frequency distributions (annual maximum peak flows). The squares indicate the observed annual maximum peak flows.

Hollis (1975)). Actually, while the peak flow and therefore the return period of the event increase, one may expect a smaller incidence of the infiltration process on the formation of the flood runoff, which makes the effects of land use change less significant. Nevertheless, even for the highest simulated discharges during the four-year period 1994–1997, the value of $\varepsilon(t)$ is comprised in between 0.1 and 0.2.

However, the relatively short length of the considered simulation period does not allow to thoroughly explore the effects of land use changes on flood frequency. In this regard, one has to consider that the flows of the years 1994–1996 have never been of exceptional nature, about 170 m^3/s being the highest discharge value simulated in that period, which corresponds to a return period of about 2 years. For that reason, the historical data simulation might allow only a partial analysis of the effect of land use change on flood flows, even if it has showed that this effect could be somewhat important in some cases.

6.2. Synthetic data simulation

The simulation with synthetic data has been carried out in order to achieve more complete indications on the effect of land use change on the flood frequency distribution. Initially, in order to verify the reliability of the simulation procedure based on synthetic data generation, the frequency distributions of the observed and the simulated (both 1955 and 1992 land use scenarios) annual maximum peak flows were compared with the frequency distribution of the observed annual maxima, retrieved from the 33 year long sample of data. The comparison was performed by referring to both 1955 and 1992 land use scenarios because the observed data were collected in the 1937–1997 period. Fig. 8 shows a satisfactory agreement.

In order to perform a quantitative comparison between the main statistics of the observed and simulated samples, we first of all collected in a unique set of data the 2000 synthetic annual maximum flows referred to the 1955 and 1992 scenarios. Then, 1000 set of data with sample size equal to 33 (the sample size of the observed annual maximum river flows) were generated by randomly extracting synthetic annual maxima from the unique sample. The mean values μ_i, with $1\leq i\leq 1000$, of the 1000 generated sets were computed and ranked in ascending order. Finally, we took μ_{25} and μ_{975} as 95% confidence limits for the mean to compare with the mean value of the annual maximum observed sample. The same procedure was followed to obtain the 95% confidence limits for the standard deviation and skewness.

Table 3 shows the obtained 95% confidence limits along with the respective statistics computed on the observed sample. One can see that there is a

Table 3
Comparison between mean, standard deviation and skewness of observed and simulated annual maximum river flow series. Details about the calculation of the confidence limits are given in Section 6.2

	Observed sample	95% confidence limits simulated sample
Mean (m^3/s)	166.49	87.60–168.90
Standard deviation (m^3/s)	113.83	74.13–146.44
Skewness	1.18	0.29–2.10

satisfactory agreement for the standard deviation and skewness, while the mean value of the observed sample is close to its upper confidence limit. This indicates a tendency of the simulation model to underestimate the mean value of the flood frequency distribution. However, from a statistical point of view the fit can be considered satisfactory, as one cannot reject the hypothesis that observed and simulated mean values are not different at the 95% confidence level.

However, one should note that the verification of the goodness of the fit is extended to a return period of about 40 years (see Fig. 8). Therefore, the simulation of considerably higher peak flows could be affected by a relevant uncertainty, which cannot be explored reliably as a results of data unavailability. However, the derived indications are useful in order to detect the response of the rainfall–runoff model when referring to very high flows.

Concerning the assessment of the effects of land use change, Fig. 8 reports the comparison between the frequency distributions of the annual maximum peak flows relative to the 1955 and the 1992 land uses. The significant difference between the low return period flows in the two scenarios confirms the results achieved in simulation with historical data. For example, the 500 m^3/s discharge, which is related to a 200 year return period in the 1955 scenario, corresponds to a return period of about 80 years in the 1992 scenario. Fig. 9 shows the progress of $\varepsilon(t)$ for increasing the 1992 simulated flows and confirms that the human induced land use change is more influent on the lower discharges (Brath and Montanari, 2000). This result is proved also by Fig. 10, which draws the

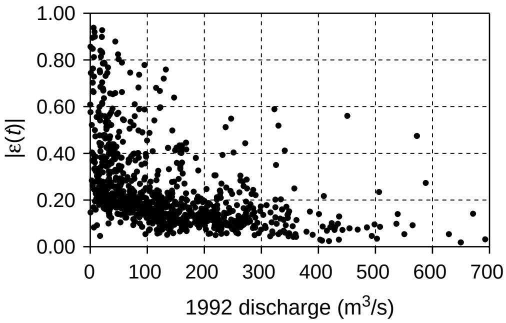

Fig. 9. Synthetic data simulation. Relative deviation $\varepsilon(t)$ versus the 1992 land-use scenario discharges ($Q_{1992} > 20$ m^3/s).

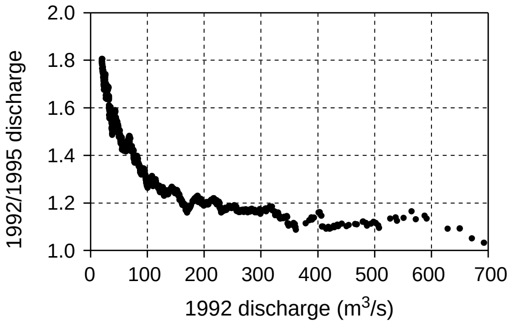

Fig. 10. Synthetic data simulation. Ratio between the 1992 and the 1955 peak discharges versus the corresponding 1992 one ($Q_{1992} > 20$ m^3/s).

ratio between the 1992 and the 1955 peak discharges versus the corresponding 1992 one.

7. Analysis of the simulation uncertainty

The results derived with the simulation approach are to be evaluated in view of their uncertainty. In fact, even if the rainfall–runoff model used here seems to provide satisfactory performances, one should not forget that river discharge simulations are always affected by errors. Uncertainty in rainfall–runoff models output may be originated by several causes, such as input uncertainty, parameter uncertainty and model structural uncertainty. The latter is due to the intrinsic inability of the rainfall–runoff model to provide a perfect simulation of the inherent hydrological processes. Furthermore, a potential source of uncertainty relies in the model output, which may be affected by measurement errors that are influent on model parameters and evaluation of model performances.

Uncertainty assessment in rainfall–runoff modelling has been widely investigated by many authors in the recent past (see, e.g. Beven and Binley, 1992; Krzysztofowicz, 2002; Beven, 2002; Vrugt et al., 2003). In particular, a quite general framework for the assessment of the simulation uncertainty of rainfall–runoff models has been recently proposed by Montanari and Brath (2004), by using a meta-Gaussian approach for assessing the uncertainty of rainfall–runoff simulations. The approach is structured accordingly to the following steps: (a) a record S_t of river flows simulated by the model, along with the corresponding record of simulations errors E_t, is collected. The normal quantile transform (Kelly and Krzysztofowicz, 1997) is applied in order to make the marginal probability distribution of S_t and E_t Gaussian. Let us denote the transformed variables with the symbols $\boldsymbol{NS_t}$ and $\boldsymbol{NE_t}$. (b) the cross-dependence between NS_t and NE_t is assumed to be governed by the normal linear equation

$$\boldsymbol{NE_t} = \boldsymbol{\rho_{NE,NS}}(0)\boldsymbol{NS}_t + N\varepsilon_t, \tag{7}$$

where $\rho_{\mathrm{NE,NS}}(0)$ is the lag zero Pearson's cross-correlation coefficient between NS(t) and NE(t) and $N\varepsilon_t$ is an outcome of a stochastic process which is stochastically independent of NS_t and normally distributed with mean zero and variance $1-\rho^2_{\mathrm{NE,NS}}(0)$. Under these assumptions the probability distribution of the transformed model error conditioned to the transformed simulated river flow is Gaussian, with mean and variance given by

$$\mu(\mathrm{NE}_t|\mathrm{NS}_t) = \rho_{\mathrm{NE,NS}}(0)\mathrm{NS}_t \tag{8}$$

$$\sigma^2(\mathrm{NE}_t\,|\,\mathrm{NS}_t) = [1-\rho^2_{\mathrm{NE,NS}}(0)] \tag{9}$$

Therefore, once the model simulation is known, one can estimate the probability distribution of the model error in the transformed space, which allows the computation of the confidence limits of the simulated river flows for a given confidence level. Ultimately, by applying back the normal quantile transform, one derives the confidence limits in the natural domain. The interested reader may refer to Montanari and Brath (2004) for more details on the procedure and its application to the case study of the Samoggia River basin.

By applying the above-described technique to the 1992 simulation of the Samoggia River, one can estimate the 95% confidence limits for the computed river flows. Since it is advisable not to extrapolate the uncertainty assessment procedure outside the range of the river flows which were used for its calibration (Montanari and Brath, 2004), the technique has been applied to the historical simulation only. The results of the uncertainty assessment are summarised in Fig. 7, where the progress of the quantity

$$\gamma(t) = \frac{\max\{|S_t^+ - S_t|, |S_t^- - S_t|\}}{S_t} \tag{10}$$

is reported for the sake of comparison with the corresponding $\varepsilon(t)$ values. In (10) S_t^+ and S_t^- represent the upper and lower 95% confidence limit of the simulated river flow S_t, respectively.

Looking at the plot of Fig. 7, one notes that the simulation error, at the 95% confidence level, has the same order of magnitude of the estimated river flow variation due to land use change from 1955 to 1992, along the whole range of the simulated river flows. This means that the impact of land use change could be potentially hidden by the simulation uncertainty at the 95% confidence level. However, it should be noted that the detected impact is nevertheless significant from a hydrological point of view. In fact, it is a common experience that a change in a hydrological parameter or variable that is not statistically significant is instead extremely important from a physical perspective (Klemes, 1974). The presence of uncertainty does not necessarily imply that a result is useless; however, the uncertainty assessment is essential to know to what extent one may be confident in the results. In the present case, we found that the return period of the 500 m^3/s flow is reduced of more than 50% and therefore underwent through a remarkable variation. Nevertheless, one should be aware that these indications are to be interpreted cautiously, since the uncertainty of the simulation procedure is not negligible.

8. Conclusions

In the present work, the effects of land use change which occurred in the last decades on the flood frequency have been investigated for an Apennine river basin. The results of the analysis have highlighted the remarkable sensitivity of the flood flow regime in response to the occurred land use change, which implies a not negligible increase in the peak discharge. In percentage terms, this increment reveals itself with greater incidence for the lower return period discharges, whereas it seems to be less significant for high return period.

It is worth pointing out that these conclusions cannot be extended to the analysis of the effects on river stages, and hence on the flood risk, of man made changes along the river banks, since in this circumstance even a local and small variation of the river cross section geometry may induce significant drawbacks in terms of vulnerability of flood prone areas.

The authors believe that the Samoggia River basin constitutes a case study representative enough for the Apennine area and the land use modifications that occurred in most part of northern Italy in the last 50 years. However, one should not forget that the results of the analysis strictly depend on the rainfall–runoff model and that the simulation uncertainty appear to be significant. The analysis of other case studies, which may refer to different catchments affected by land use change, by using different rainfall–runoff modelling procedures, could confirm the results of the present study.

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