Atmospheric Research, 27 (1991) 169–189
Elsevier Science Publishers B.V., Amsterdam

Extreme storm rainfall and climatic change

Paolo Burlando and Renzo Rosso
Department of Hydraulic, Environmental and Surveying Engineering, Politecnico di Milano, 32 Leonardo da Vinci, Milano I-20133, Italy

(Received December 2, 1990; accepted after revision January 15, 1991)

ABSTRACT

Burlando, P. and Rosso, R., 1991. Extreme storm rainfall and climatic change. Atmos. Res., 27: 169–189.

In recent years General Circulation Models (GCMs) have been widely developed in the attempt to simulate global climate, and to predict future climate scenarios under conditions of enhanced CO_2 concentration in the atmosphere. Unfortunately, these models provide estimates of he hydrological balance, which are integrated over a very coarse spatial computing grid. Significance and usefulness of such estimates are then limited with respect to the needs of water resources analyses at the basin scale. Accordingly, in this paper, coupling of GCMs' results and hydrology modeling is tackled, by investigating the precipitation process. A theoretical approach for linking modified scenarios simulated by GCMs to local changes in the precipitation patterns under global change is proposed. A stochastic model based on the theory of point process is used to describe the temporal process of precipitation jointly with scaling properties of rainfall, for the purpose of linking analytically changes predicted by GCMs to local modifications of the precipitation patterns under the global change. A sensitivity analysis of the proposed approach is also reported, jointly with an application to real-world data, in order to assess the validity of the procedure, and to address some implications of the global change on the precipitation process.

INTRODUCTION

In recent years, research on climatic processes has been showing the increasing greenhouse warming of the earth to be a potential source for a global change of present climate (see e.g. Bolin et al., 1986; Schneider, 1989). Climate and hydrology have very close links, and for quite a while it has been recognized that their mutual influence has to be accounted for in the exercise of modeling the climatic change at the global scale (see e.g. Manabe and Wetherald, 1985). Hydrology can contribute to the improvement of the techniques of climate modeling, as well as climate modeling can contribute to the improvement of hydrologic prediction. It has been recently pointed out by Dooge (1990) that "such an interchange would not only require a clear delineation of the interface between hydrology and climatology but would probably also call for a clearer understanding of the relationship between theory

0169-8095/91/\$03.50

and practice in each of the two areas". Coupling global climate and hydrologic modeling requires a number of unsolved hydrological problems to be approached, and possibly solved; within this context, scale problems play a major role in both the exercises of entering a sound description of land-surface processes into climate models, and of using large-scale climatic predictions to obtain reliable hydrological estimates of the possible impacts of climate change at the basin scale.

INVESTIGATING THE EFFECTS OF GLOBAL CHANGE ON HYDROLOGIC SYSTEMS

Following Gleick (1989), the methods of analysis used to assess the effects of global change on hydrological systems can be grouped into three approaches: (1) the direct use of General Circulation Models (GCMs); (2) the use of paleoclimate analogues; and (3) the analysis of recent climate analogues. Because several things limit the usefulness of paleoclimatic scenarios (i.e., the farther going back, the more difficult to recover hydrologic data; geologic shifts over geologic time may differ considerably from the anthropogenetic change now anticipated; past changes predate human activity, so that there is no evidence of how they might affect society), the use of GCMs and the analysis of recent climate changes provide the most promising routes to investigate hydrological impact of global change.

GCMs' outputs also include large-scale hydrological predictions: for instance, Manabe and Wetherald (1987) reported large-scale changes of soil wetness induced by an increase in atmospheric carbon dioxide, and Mitchell and Warrilow (1987) investigated summer dryness in northern mid-latitudes due to increased CO_2. However, these results can only provide general tendencies at most, but they have a very minor use within the range of scales of hydrologic interest for both scientific or practical purposes. Therefore, the outputs of GCMs must be disaggregated and used to input hydrological models. For instance, this route has been pioneered in the recent works by Lettenmaier and Gan (1990), who used different GCMs results to analyze hydrologic sensitivity of Sacramento valley, and by Vehvilainen and Lohvansuu (1990), who used GISS (Goddard Institute for Space Studies) model outputs to assess a strong increase of winter discharges of rivers in Finland.

The analysis of recent climate changes has been alternatively used to develop climate and hydrology scenarios. For instance, Camuffo (1984) analyzed long-term precipitation recorded in Padua from 1725 to 1981; Rowntree (1985) developed an annual precipitation index for Southern California from 1769 to 1834 using agricultural records from the Spanish period; Wigley et al. (1984) looked at precipitation records in England and Wales for the period 1766–1980 to evaluate whether there had been a change in the frequency of wet and dry rainfall extremes, and Jones (1984) then used these data to reconstruct streamflow in order to investigate the effects of climate on water availability. Using recent climate analogies has the advantage that these rely on human experience, but the fact that anticipated greenhouse climatic changes may have a cause unlike those that led the past climatic variability is a major drawback to this approach.

Another approach to the problem is to analyze the sensitivity of hydrologic systems to hypothetical climate scenarios (e.g., +2, +3, +4°K temperature, +10 or +20% precipitation or evapotranspiration). This route, which was pioneered by Stockton and Bogges (1979) and Nemec and Schaake (1982), has been recently followed among others by Kutchment (1990) for snowmelt runoff of rivers in the northern regions of USSR; by Mimikou and Kouvopuolos (1990) for three basins in Greece; by Beran and Arnell (1989) to assess the effect of climatic change on quantitative aspects of United Kingdom water resources. However, Gleick (1989) states that "although these scenarios are the easiest to develop, they are not particularly realistic, and they often lack of internal consistency"; therefore a lot of care must be taken in interpretating these results.

In the field of urban hydrology, the pioneer work by Niemczynowicz (1989) assumed hypothetical increases in storm precipitation to investigate the possible impacts of modifications in rainfall patterns on the sewerage system of the city of Lund, Sweden. To perform this exercise, areal intensity-duration-frequency (IDF) curves were simply shifted by superimposing 10, 20 and 30% of rainfall intensity increase, so that the internal consistency of the hypothetical change in rainfall pattern with climatic change is not well assessed. However, the results indicate the great vulnerability of sewerage systems to the results of climatic changes.

ANALYZING THE EFFECTS OF CLIMATIC CHANGE ON STORM RAINFALL

GCMs simulations show the general behavior of the equilibrium rainfall patterns under given future scenarios of greenhouse gas concentrations, i.e., the general tendency of precipitation to increase or decrease on average during a given season after the new climatic equilibrium has been reached (see, e.g. Manabe et al., 1981; Wilson and Mitchell, 1987; Manabe and Wetherald, 1987). An excellent review of the most recent achievements which have been obtained by GCMs, jointly with the present uncertainties involved in predicting the global change has recently presented by Mitchell (1989). Despite of these uncertainties Mitchell states that "annual mean precipitation and runoff increase in high latitude, and most simulations indicate a drier land surface in northern mid-latitude in summer."

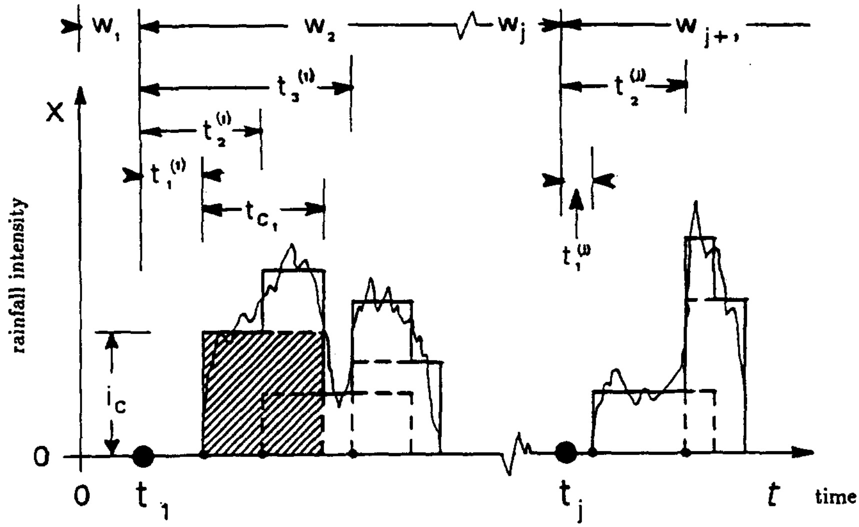
Burlando and Rosso (1990) have recently suggested to combine GCMs simulations with the analysis of rainfall records in order to transfer the results of these simulations to the field of scientific and operational hydrology. To this effect, stochastic models can provide a deeper insight of precipitation process than simply shifting global rainfall data at a point in space. Moreover, those stochastic models which describe precipitation in terms of its major features (e.g., duration and intensity of storm cells, interstorm arrival time, and so on) involve physically based model parameters, which directly resemble the internal structure of rainfall process. Using fine resolution precipitation data (i.e., daily or, better hourly and sub-hourly) can further provide the knowledge of those mutual fluctuations of model parameters, which can give a deeper insight of the internal structure of non-stationarity of the precipitation process. This approach can be also regarded as an alternative approach to the analysis of recent precipitation changes, since it is aimed to detect the internal structure of rainfall process in order to achieve realistic future scenarios.

Hypothetical changes in the frequency of storm rainfall have been recently studied by Gregory et al. (1990) using a daily precipitation model for areal average precipitation, which his based on a first-order Markov chain for the renewal process to decide whether each successive day is "wet" or "dry", and a probability density function for daily precipitation in each wet day, independently chosen according to a best-fit analysis. Their results point to the possibility of dramatic changes in the frequency of daily extremes corresponding to realistic projections of changes in lower order moments. However, their approach does not completely overcome those problems arising from the simple shifting the IDF curve, because the model is able to describe "external" properties of precipitation.

Alternatively, Burlando and Rosso (1990) used a simpler model based on Poisson arrivals of storm events, which are characterized by random duration, and depth of the rainfall amount delivered by each event. Although this model is able to capture the effective features of precipitation for a limited range of scales in time (see, e.g. Rodriguez-Iturbe, 1986; Burlando and Rosso, 1989), it relies on effective storm characteristics, so that it is able to reflect the physical meaning of changes in lower order moments of integrated precipitation (i.e., annual or seasonal rainfall totals). Moreover, Burlando and Rosso (1990) introduced the scaling concept to analyze non-stationarity in temporal rainfall patterns; this concept can usefully overcome the present uncertainties involved in predicting future patterns. In the present this method is applied to the analysis of extreme storm rainfall, and to the derivation of IDF curves from those scenarios which are provided by the results from GCMs simulations.

PREDICTING FUTURE SCENARIOS UNDER THE GLOBAL CHANGE

Following the approach recently introduced by Burlando and Rosso (1990) local modifications of rainfall patterns can be investigated through combining GCMs outputs and non-stationary rainfall analysis by means of a stochas-



$t_c^{(j)}$ = displacement of the j-th cell from the cluster center
w_j = interarrival time of events (pulses)
i_c = intensity of the i-th pulse
t_c = duration of the i-th pulse

Fig. 1. The Neyman-Scott rectangular pulses model.

tic model and scaling concepts. In order to overcome the inadequacies of the simple Poisson model used by Burlando and Rosso (1990) with respect to the extreme storm rainfall, a Poisson cluster model has been adopted to reproduce the temporal process of rainfall, $X(t)$. This is the Neyman-Scott Rectangular Pulses (see Fig. 1), which an extensive analysis has shown to be able to capture the properties of both the continuous and the extreme process (Burlando, 1989; Burlando and Rosso, 1989). It is based on Poisson arrivals of storms, and to each arrival a cluster of rectangular pulses of random height and duration is associated, being these randomly displaced from the cluster origin. The superposition of these pulses provides the description of the storm profile. It is often assumed that both the intensity and the duration of a pulse (i.e. a storm cell) are iid exponentially distributed, being displaced from the cluster origin according to an exponential distribution. Under these hypotheses second-order properties of the temporal rainfall process aggregated at scale $T, X_T(t)$, can be written as (Rodriguez-Iturbe, 1986):

$$E[X_T(t)] = \lambda \mu \delta \nu T \tag{1}$$

$$\mathrm{var}[X_T(t)] = 2\lambda\mu^2\delta^3\nu\left(\frac{T}{\delta} - 1 + \exp\left(-\frac{T}{\delta}\right)\right)\left[2 - \frac{\beta^2\delta^2(\nu-1)}{(1-\beta^2\delta^2)}\right] + \frac{2\lambda\mu^2\delta^2\nu(\nu-1)}{\beta(1-\beta^2\delta^2)}[\beta T - 1 + \exp(-\beta T)] \tag{2}$$

$$\mathrm{cov}[X_T^i, X_T^{i+k}] = \lambda\mu^2\delta^3\nu\left[2 - \frac{\beta^2\delta^2(\nu-1)}{(1-\beta^2\delta^2)}\right]\left[1 - \exp\left(-\frac{T}{\delta}\right)\right]^2 \times \exp\left(-\frac{T(k-1)}{\delta}\right) + \frac{\lambda\mu^2\delta^2\nu(\nu-1)}{\beta(1-\beta^2\delta^2)} [1-\exp(-\beta T)]^2 \exp[-\beta(k-1)T] \quad k \geqslant 1 \tag{3}$$

where λ is the poissonian rate of storm arrival, μ is the mean intensity of a pulse, δ is the mean duration of a pulse, ν is the mean number of cells in a cluster, and β^{-1} is the mean displacement of a cell from the cluster origin.

Furthermore, the scale of fluctuation θ of the process $X(t)$ which gives the time interval required to obtain stable (low variance) estimates of the mean of the fluctuating process of rainfall intensity, can be expressed as:

$$\theta = 2\delta \frac{(\nu+1)}{2 + \frac{(\nu-1)\beta\delta}{(1+\beta\delta)}} \tag{4}$$

In order to derive an analytical framework for disaggregating GCMs' results, these second-order properties can be combined with scaling properties of the actual temporal rainfall process. To this purpose, as first introduced by Gupta and Waymire (1990) and developed by Burlando and Rosso (1990), the temporal rainfall process can be described as a multi-scaling process. This allows for deriving a scaling relationship between the moments of the original and of the rescaled process, which can be written in the logarithmic form as (Gupta and Waymire, 1990):

$$\log E[X_{\beta T}^h] = s(h)\log\beta + \log E[X_T^h] \tag{5}$$

where $X_{\beta T}^h$ represents the process at the temporal level of aggregation T rescaled by β, h is the order of the moment, and $s(h)$ is the scaling exponent, which is a non-linear function of the order of the moment.

If this property is assumed to hold with respect to central moments, one can furtherly derive (Burlando and Rosso, 1990) a scaling relationship between the second-order central moment and the expected value, that is:

$$\sigma_{X_T(t)} = a[m_{X_T(t)}]^\alpha \tag{6}$$

or, equivalently:

$$\sigma^2_{XT(t)} = b[m_{XT(t)}]^{2\alpha} \tag{7}$$

where the characteristic exponent, α, and the two constants a and b have to be estimated from actual data.

The theoretical validity of eqs. 6 and 7, and of the hypotheses under which these are obtained is not herein discussed. Although these relationships do not descend directly from the original framework of a multi-scaling process as introduced by Gupta and Waymire (1990), there is empirical evidence of such relationships to provide a suitable working assumption, as they have been tested on several stations for which long historical series were available. Figure 2 reports the plot of eq. 6 as it was estimated using the above-mentioned data set of the station of Firenze Osservatorio Ximeniano (Italy). In Table 1 it is also shown that eqs. 6 and 7 are verified on monthly basis with values of the scaling exponents, α, ranging from 0.73 to 0.88, and with R^2 ranging from 0.83 to 0.97.

By assuming that the change between the mean daily precipitation in the enhanced CO_2 scenario and in the actual one can be expressed by the ratio:

$$\frac{m_{(n\mathrm{CO_2})}}{m_{(\mathrm{control})}} = K_m \tag{8}$$

and, moreover, by assuming that the parameters of the scaling relationships 5 and 6 do not vary under the new scenario, one can easily get:

$$\frac{\sigma_Y}{\sigma_X} = K_\sigma = (K_m)^\alpha \tag{9a}$$

$$\frac{\sigma^2_Y}{\sigma^2_X} = K_{\sigma^2} = (K_m)^{2\alpha} \tag{9b}$$

where it is assumed, for sake of simplicity, that the index Y stands for the enhanced CO_2 scenario and X for the control one, respectively. In this framework the key factor, K_m, which provides a concise estimate of the likely impact of global change on water resources, can be estimated from the analysis of GCMs' results.

A further relationship, similar to eq. 9, can be obtained also to estimate the rescaling factor for the scale of fluctuation, K_θ, by introducing the assumption of linearity of the system, that is (Vanmarcke, 1983):

$$\frac{\sigma^2_X \theta_X}{m^2_X} = \frac{\sigma^2_Y \theta_Y}{m^2_Y} \tag{10}$$

When actual data are analyzed, this condition is not always verified as one would expect, so that a larger analysis of observed data than only one sample station, like in the present paper, is needed. Nevertheless, eq. 10 still represents a working assumption which allows for analytical derivation, finally

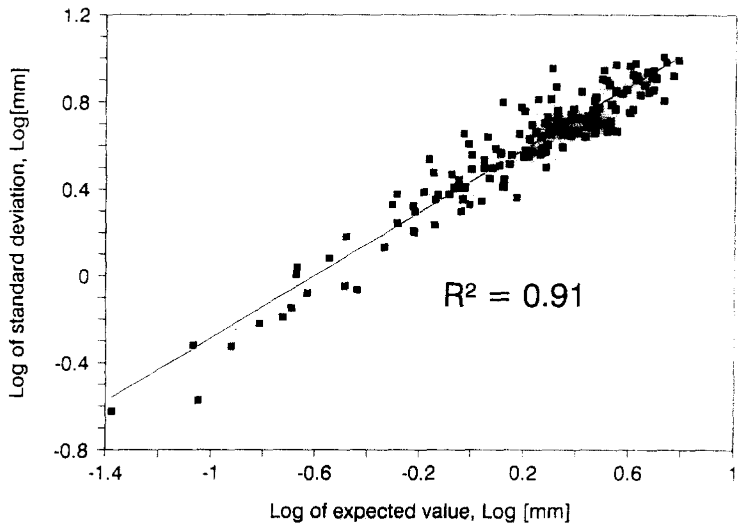


Fig. 2. Scaling property of observed daily data of March 1813–1979 at Firenze Ximeniano, Italy.

TABLE 1

Monthly values of the scaling parameters α for the station of Firenze Ximeniano (Italy)

Month	Scaling exponent α (–)	R^2	Std. error of estimate (–)
Jan	0.763	0.889	0.021
Feb	0.793	0.952	0.014
Mar	0.725	0.906	0.019
Apr	0.776	0.921	0.018
May	0.753	0.827	0.027
Jun	0.882	0.966	0.013
Jul	0.833	0.935	0.017
Aug	0.854	0.954	0.015
Sep	0.813	0.929	0.018
Oct	0.746	0.838	0.026
Nov	0.768	0.848	0.026
Dec	0.747	0.878	0.022

yielding to the relationship, which links for the scale of fluctuation in the enhanced CO_2 scenario to the present one, namely:

$$K_\theta = K_m^{2(1-\alpha)} \tag{11}$$

Relationships similar to eq. 7 can be written also to relate the changes of the NSRP model parameters between the present and the enhanced CO_2 scenario, so obtaining:

$$\frac{\lambda_Y}{\lambda_X}=K_\lambda; \quad \frac{\mu_Y}{\mu_X}=K_\mu; \quad \frac{\delta_Y}{\delta_X}=K_\delta \tag{12}$$

Two further assumptions are needed in order to obtain analytically tractable relationships. These concern (1) the number of cells per cluster, which are assumed equal for both the standard and the enhanced CO_2 scenarios (i.e, $K_\nu=1$), and (2) the mean displacement of cells from the cluster origin, which is assumed to be rescaled with respect to the cell duration, namely:

$$\frac{\beta_Y}{\beta_X}=(K_\beta)=(K_\delta)^s \tag{13}$$

being s a characteristic exponent to be evaluated from estimated parameters.

In order to estimate K_λ, K_μ and K_δ these must be expressed as function of K_m, which concisely represents a quantitative although rough estimate of climate change impact on water resources. This can be obtained by combining the NSRP second-order properties—that is eqs. 1, 2 and 3—with eqs. 8, 9 and 11 based on scaling considerations. After some manipulations one can finally write K_λ, K_μ and K_δ as function of K_m, being parameters the mean pulse duration computed for the present scenario, δ_X, the scale of temporal aggregation, T, and the scaling exponent, α, so obtaining:

$$K_\lambda=\frac{K_m}{K_\mu K_\delta} \tag{14}$$

$$K_\mu=\frac{K_m^{2\alpha-1}}{K_\delta}\frac{(\Phi_1+\Phi_2)}{(\Phi_3+\Phi_4)} \tag{15}$$

$$K_\delta=\frac{1+K_\delta^{(s+1)}\beta_X\delta_X}{2+K_\delta^{(s+1)}\beta_X\delta_X(\nu+1)}=K_m^{2(1-\alpha)}\frac{(1+\beta_X\delta_X)}{2+\beta_X\delta_X(\nu+1)} \tag{16}$$

where eq. 16 has to be solved numerically, and Φ_1, Φ_2, Φ_3 and Φ_4 are functions of δ_X, ν, β_X and T, and are given by:

$$\Phi_1=\left\{\delta_X\left[2-\frac{\beta_X^2\delta_X^2(\nu-1)}{(1-\beta_X^2\delta_X^2)}\right]\left(\frac{T}{\delta_X}-1+\exp(-\frac{T}{\delta_X})\right)\right\} \tag{17}$$

$$\Phi_2=\frac{(\nu-1)}{\beta_X(1-\beta_X\delta_X)}(\beta_X T-1+\exp(-\beta_X T)) \tag{18}$$

$$\Phi_3=\left\{K_\delta\delta_X\left[2-\frac{K_\delta^{2(m+1)}\beta_X^2\delta_X^2(\nu-1)}{(1-K_\delta^{2(m+1)}\beta_X^2\delta_X^2)}\right]\left(\frac{T}{K_\delta\delta_X}-1+\exp\left(-\frac{T}{K_\delta\delta_X}\right)\right)\right\} \tag{19}$$

$$\Phi_4=\frac{(\nu-1)}{K_\delta^m\beta_X(1-K_\delta^{(m+1)}\beta_X\delta_X)}(K_\delta^m\beta_X T-1-\exp(-K_\delta^m\beta_X T)) \tag{20}$$

From eqs. 14, 15 and 16 it is then straightforward to estimate the new NSRP parametrization for the n-CO_2 scenario.

SENSITIVITY ANALYSIS AND APPLICATION TO REAL WORLD

A deeper insight of eqs. 14, 15 and 16 shows that that K_λ, K_μ and K_δ account for the internal storm characteristics, represented by the mean duration of the pulse, δ_X, and for the temporal scale of aggregation, T, at which the NSRP model is considered. A further property of the rainfall process is also included in terms of the scaling exponent, α, which describes how the variance is rescaled with respect to the mean. In this view, sensitivity analysis of eqs. 14, 15 and 16 has been performed to assess the capability of this framework to describe the influence of the climatic change on the internal rainfall patterns.

To this effect, a scale of aggregation of 24 hours has been selected for the model, being this homogeneous with the scale with K_m is referred to. The three coefficients, K_λ, K_μ and K_δ have been then computed as function of K_m, for two different values of the mean pulse duration, namely δ_K equal to 3 and 8 hours. These correspond to convective and frontal storms, respectively. The scaling factor α is assumed to range from 0.7 to 0.9, which are values that normally characterize observed data, as it is reported in Table 1, showing data analysis for the station of firenze Ximeniano. The results of this sensitivity analysis are reported in Figs. 3 and 4, where the difference of the three factors from the unit value corresponding to the standard simulation, ΔK_λ, ΔK_μ and ΔK_δ, are plotted versus the variation of the average precipitation factor, ΔK_m.

In the case of front precipitation (Fig. 3), the two factors, K_λ, and K_μ, are shown to decrease when decreasing K_m, and symmetrically, to increase when increasing K_m: conversely, K_δ decreases for increasing K_m, and viceversa. Furthermore, for a given value of K_m each parameter factor shows different gradients depending upon the scale parameter α, being all of the three factors more sensitive for low values of the scaling parameter.

When a convective precipitation set of parameter is examined (Fig. 4), a behavior similar to the previous case is obtained. Weaker variations than in the previous case are reported for K_μ and K_δ, but stronger gradients are shown for K_λ. Accordingly, the number of storm arrivals, i.e., the frequency of events, seems to be more sensitive to climatic change during late spring and summer periods, when convective nature of storm prevails, than in winter time, while internal storm patterns—intensity and duration—seems to show the opposite behavior.

The results shown by the sensitivity analysis can be also detected from the application of the presented approach to real world. This has been performed for the station of Firenze Ximeniano, which is located in Central Italy. This region lies in the area of Southern Europe which has been studied under climate change by Manabe and Wetherald (1987). From the GCMs' results pre-

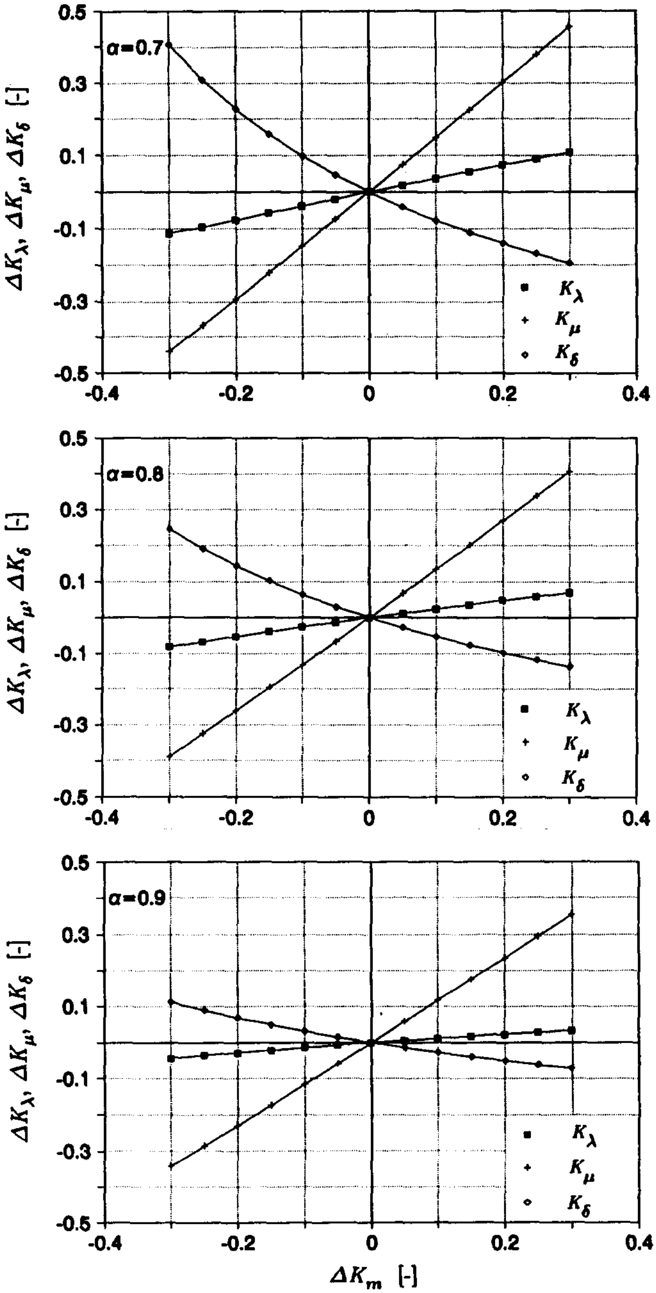


Fig. 3. Variations of K_λ, K_μ and K_δ vs. variations of K_m. Sensitivity analysis for the case of frontal precipitation (δ_X=0.5 hours, $(\beta_X)^{-1}$=6.67 hours, ν=7).

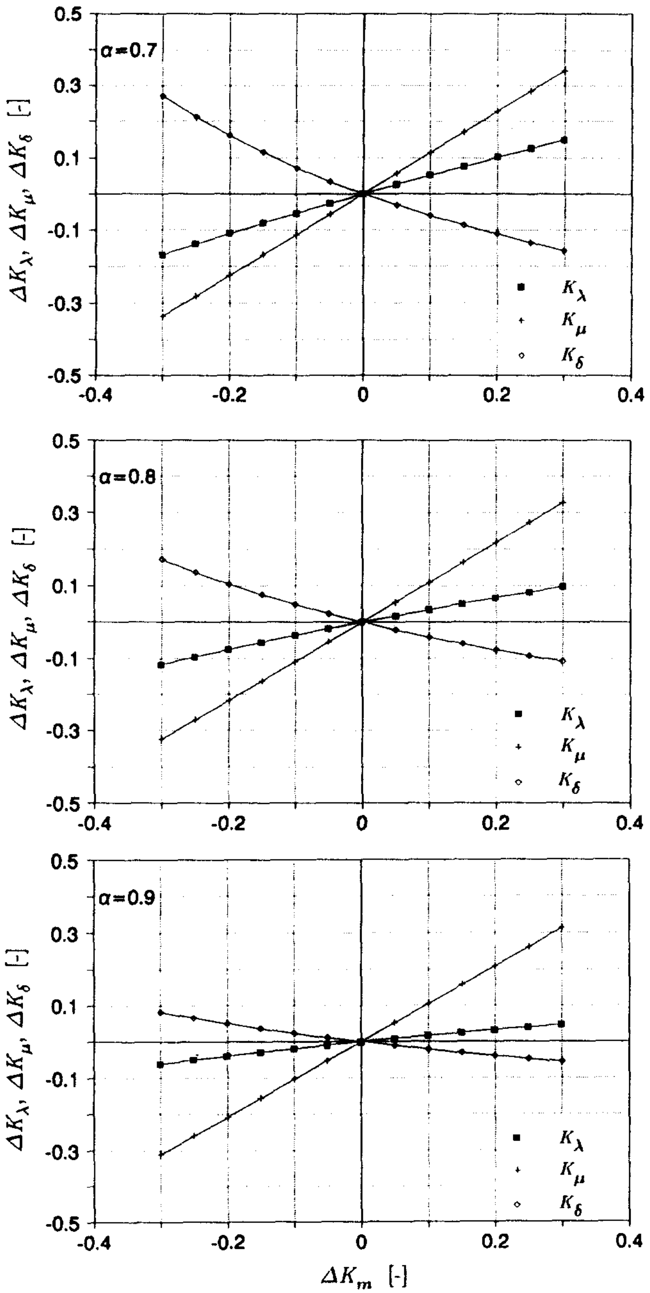


Fig. 4. Variations of K_δ, K_μ, and K_δ vs. variations of K_m. Sensitivity analysis for the case of convective precipitation ($\delta_X = 0.25$ hours, $(\beta_X)^{-1} = 4$ hours, $\nu = 3$).

sented for Southern Europe by these authors, the monthly factor, K_m, which should characterize the change of the average daily rainfall intensity under $4CO_2$ scenario, has been derived. A plot of the values of it are reported in Fig. 5, showing a decrease of the daily intensity during the June to September period, and an increase throughout the rest of the year. The largest increase of the precipitation rate is recorded for the period of transition from winter to spring. Accordingly, wetter winters and drier summers than at present are expected to occur.

Using the values of K_m derived from Manabe and Wetherald (1987), local changes of precipitation patterns can be quantitatively evaluated through the new parametrization of the NSRP model, which can be obtained from the equations derived in the previous section. This is reported in Figs. 6–8, where the values of the NSRP model parameters variations are plotted. It can be observed that two of the three examined parameters of the NSRP (the mean displacement from the cluster origin, β, is rescaled with respect to the mean duration of a cell, δ, while the mean number of cells, ν, is assumed not to vary), namely the rate of arrivals, λ, and the mean intensity of a cell, μ, are shown to decrease when decreasing K_m during summer months. The mean intensity exhibits the largest percent variation ranging from more than 85% increase in February to about 30% decrease in July. The rate of arrivals (i.e. the number of events) varies from about 7% decrease (July) to about 19% increase (March). On the contrary, the mean duration of a cell decreases during winter months when $K_m > 1$, and increases during summer time, ranging from about -30% (March) to about $+9\%$ (July).

Accordingly, if four times of the concentration of CO_2 in the atmosphere should occur, severe dry conditions are expected on the average for summer

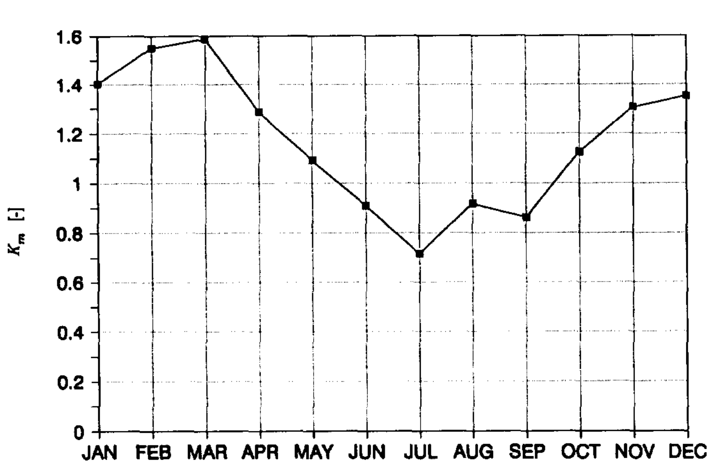


Fig. 5. GCM predicted values of K_m over Southern Europe under $4CO_2$ scenario (derived from Manabe and Wetherald, 1987).

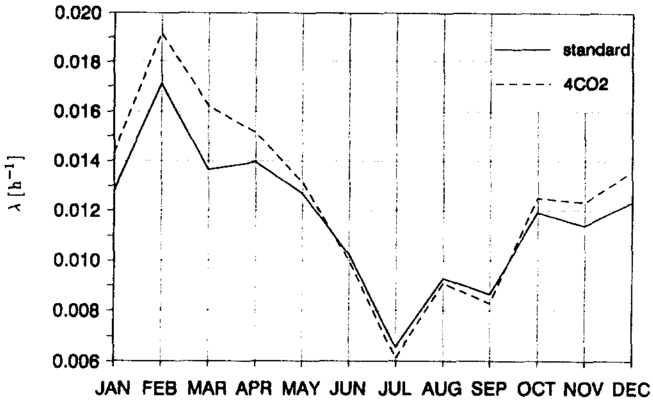


Fig. 6. Monthly variations of the NSRP model parameter representing the number of storm arrivals, λ, for the station of Firenze Ximeniano (Italy) under a $4CO_2$ scenario.

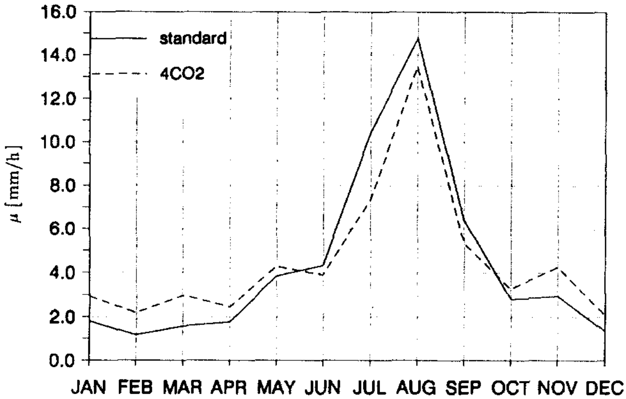


Fig. 7. Monthly variations of the NSRP model parameter representing the mean intensity of a cell, μ, for the station of Firenze Ximeniano (Italy) under a $4CO_2$ scenario.

months. From the values presented in Figs. 6–8, the expected value of rainfall depth for July can be estimated to decrease from about 28 mm in the standard scenario to about 20 mm under the enhanced CO_2 one, i.e., a reduction of about 28% might occur. A reduction of the storm duration and of the storm depth are also expected, of about 5 and 28%, respectively. On the other hand, wetter scenarios than the present ones are expected during winter months. The largest differences are recorded for the month of February, with an increase of about 22% for the storm duration and of about 38% for the storm depth, so generating an increase of the monthly rainfall depth of about 70%.

An important aspect of the modifications of local precipitation patterns is concerned with the extremal properties. It is intuitive to expect that the new parametrization of the NSRP due to the global change will have influence also

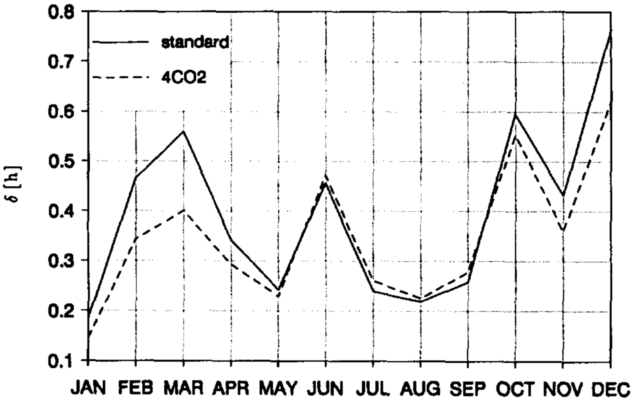


Fig. 8. Monthly variations of the NSRP model parameter representing the mean duration of a cell, δ, for the station of Firenze Ximeniano (Italy) under a $4CO_2$ scenario.

on the extreme storm rainfall. In this view, an assessment of the patterns of the extreme storm rainfall under the global change has been also performed with respect to the same data set. A comparison has been therefore performed between the risk curves predicted by the NSRP model under the standard and the enhanced CO_2 climate, based on the capability that this model showed in capturing the extremal properties of the temporal rainfall process (see Burlando, 1989). Accordingly, 200 years of continuous rainfall have been simulated using both the standard parametrization and the $4CO_2$ one. For different rainfall durations these provided the extreme values, which have been fitted by a Gumbel distribution so allowing for the estimation of the Depth-Duration-Frequency (DDF) curves for both the climate scenarios. These are reported for durations ranging from 15 min to 24 hours in Figs. 9–12.

It can be observed that for short durations the DDF curves under the new climate display lower values than under the standard climate (Figs. 9 and 10). Such behavior is limited to durations lower than three hours, being the gap between the standard and the enhanced curve larger for shorter durations. When the storm duration increases the maximum depths show to be higher under the enhanced CO_2 scenario (Figs. 11 and 12)displaying a larger difference for increasing durations. In any case the sensitivity to return period is shown to be weak for the considered durations, as it is displayed by the curves reported in Fig. 13. Therefore, the risk should be reduced for storm durations approximately lower than 2 to 3 hours, and increased for higher durations, let alone any inadequacy of the model to reproduce the extremes of very short durations (Burlando, 1989), and the uncertainties embedded in the approach. For instance, a return period (RP) of 50 years for a duration of 30 min corresponds to about 33.1 mm of rainfall in the standard climate and to about 29.8 mm under the global change, so indicating about 10% re-

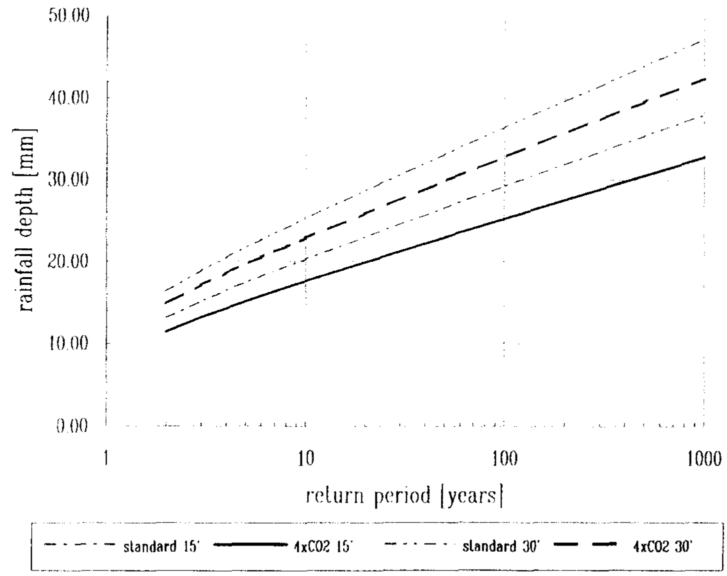


Fig. 9. Depth-Duration-Frequency curves for the standard and the $4CO_2$ scenarios obtained via simulation of the NSRP model at the station of Firenze Ximeniano. Storm durations are 15 and 30 min.

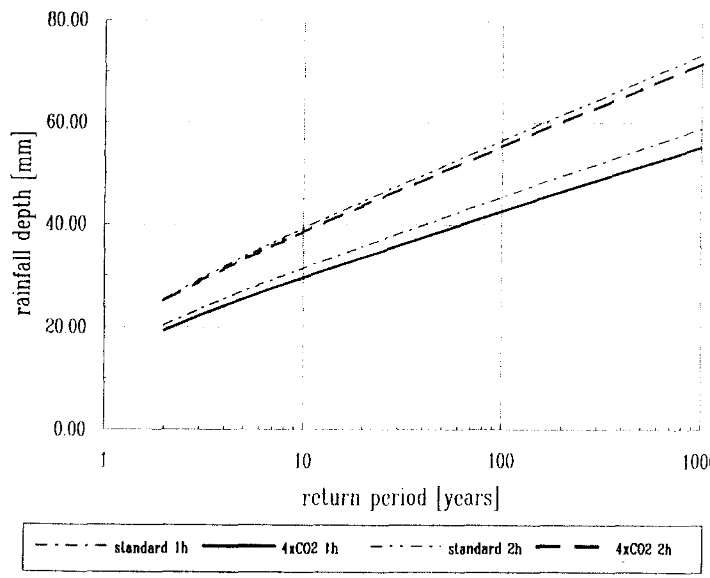


Fig. 10. Depth-Duration-Frequency curves for the standard and the $4CO_2$ scenarios obtained via simulation of the NSRP model at the station of Firenze Ximeniano. Storm durations are 1 and 2 hours.

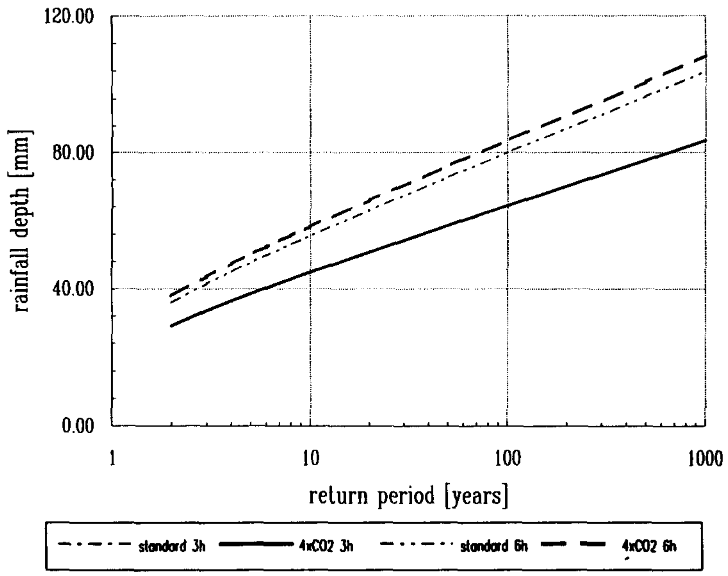


Fig. 11. Depth-Duration-Frequency curves for the standard and the $4CO_2$ scenarios obtained via simulation of the NSRP model at the station of Firenze Ximeniano. Storm durations are 3 and 6 hours.

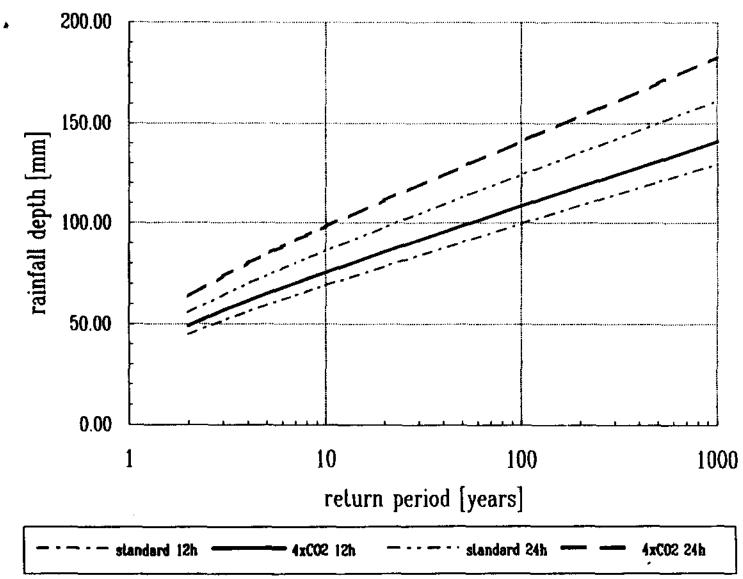


Fig. 12. Depth-Duration-Frequency curves for the standard and the $4CO_2$ scenarios obtained via simulation of the NSRP model at the station of Firenze Ximeniano. Storm durations are 12 and 24 hours.

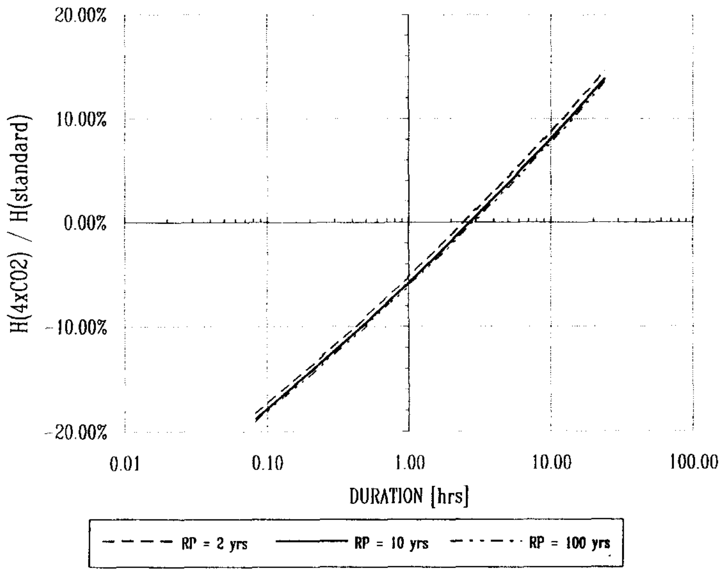


Fig. 13. Sensitivity of the ratio between the maximum depth under the changed and the standard climates to different rainfall durations and return periods.

duction of the risk value for a given RP. Similarly, a RP of 50 years for a duration of 12 hours corresponds to about 90.6 mm of rainfall in the standard scenario, and to 98.7 mm in the enhanced CO_2 one, so indicating a 9% increase under the climate change.

CONCLUDING REMARKS AND RESEARCH PERSPECTIVES

Based on the approach introduced by Burlando and Rosso (1990) the present analysis showed that a considerable impact of the greenhouse effect on the rainfall patterns should be expected. The referred results confirm those obtained by Burlando and Rosso with respect to the continuous process. Therefore, both the sensitivity analysis and the application to real-world put in evidence the importance of considering the internal storm patterns, i.e., disaggregating the GCMs' results, to understand local modifications of the process. In fact, by using this approach the general tendencies indicated by global simulations can be incorporated directly into the parameters of the stochastic model used for reproducing the precipitation process at a point in space. The global outputs from GCMs are so transformed into quantitative and detailed estimates of the local impact of a change in the climate, rather than purely hypothetical hydrological changes (see, e.g., Niemczynowicz, 1989). Such estimates allow to detect at the point scale the considerable impact that the outlined changes in the mean intensity, in the duration and in the number of storms may have with respect to both water availability and hydrogeological hazards. Accordingly, shortage of precipitation during summer months over the region of Southern Europe considered for comparison with actual data, and large rainfall amounts during the winter season have been assessed to occur with a maximum potential reduction of the monthly rainfall depth of about 28%, and a maximum potential increase of about 70%.

Moreover, the use of a cluster model allowed to overcome some of those inadequacies of the simple Poisson model used for that analysis that are related to the capability of the model to capture the extremal properties of the actual process. The NSRP model allowed therefore to estimate the extreme values using both the standard and the enhanced CO_2 parametrization, and to compare the DDF curves derived from such extremes throughout a wide range of time scales. These indicate that a reduction of the risk should be expected for short durations, namely less than three hours, while an increase of the extremes could be expected for longer durations. In both cases the variation is shown to be almost independent of the return period.

Accordingly, one can observe that small urban basins will not be considerably affected by a climatic change, while natural catchments, which are characterized by longer time scales of the hydrologic response, will be. A possible reason for such behavior stems from the modifications in the parameters of the model, which show that the reduction of the average rainfall rate predicted by GCMs generate a decrease of the cell intensity and an increase of the cell duration: therefore, convective storms which are characterized by very short and intense rain appear to be smoothed in the extreme values. The opposite effect is obtained for an increase of the GCMs predicted rate, which generates a strong increase of the cell intensity, being reduced its duration: therefore, the so modified cell pattern produces an increase of extreme storm values.

It must yet be observed that these modifications and the related results are subject to several uncertainties, which still require further investigations. First, one can observe that the uncertainty embedded in the procedure to derive DDF curves are sometimes of the same order of magnitude of the modifications predicted under the climatic change. This suggests that low increases or decreases in the risk values due to the greenhouse effect could be regarded to as already accounted for in most of the current design practices. On the other hand the effective risk modification should be evaluated with respect to the rainfall excess, rather than to the absolute values of the extremes. This implies that the above referred percentages of increase and decrease of the extremes, ranging from about -10 to about $+15\%$, could be enhanced due to infiltration and loss effects, which vary from basin to basin. In this view, the largest increases should be adequately considered as the major source of modification of flood hazards risk.

A number of remarks arise and several questions still remain open. There is an effective need for disaggregation of GCMs' outputs, in order to transform these into quantitative and detailed estimates of the local impact of a change in climate, which could be used for operational needs. To this purpose analytical framework have to be preferred, whichever process of the hydrological cycle is considered. With respect to precipitation, stochastic model and scaling properties of the process seem to provide a reasonable approach. It must yet be stressed that the presented results should be critically regarded, and further analyses are required to point out unsolved questions. These can be finally summarized as:

—the assessment of the parametrization of the model and its adequacy to reproduce precipitation patterns related to past observed climate fluctuations, which were characterized by strong deviations from the average pattern;

—the assessment of the range of validity of scaling properties with respect to modifications in the climate;

—the verification of some of the assumptions herein assumed for analytical tractability (e.g., the "degree of linearity" of the system);

—the verification of the suitability of using the average rate predicted by GCMs as an index for re-parametrization of the stochastic model;

—a deeper analysis of the extremal behavior under climatic change, which in some way was unexpected for short durations.

Some of these aspects, which of course are not exhaustive of the problem, are presently under investigation.

REFERENCES

Beran, M.A., and Arnell, N.W., 1989. Effect of climatic change on quantitative aspects of United Kingdom water resources. Contract Rep. Dep. Environ. Water Directorate, London.

Bolin, B., Doos, B.R., Jaeger, J. and Warrik, R.A. (Editors), 1986. The Greenhouse Effect, Climate Change and Ecosystems. SCOPE Rep. 29.

Burlando, P., 1989. Modelli matematici per la previsione e la simulazione della precipitazione nel tempo. Thesis. Inst. Hydraul., Politecn. Milano.

Burlando, P. and Rosso, R., 1989. Stoachastic models of temporal rainfall: Reproducibility, estimation and prediction of extreme events. NATO Advanced Study Institute (to appear In: J.D. Salas, R. Harboe and J. Marco-Segura (Editor), Stochastic Hydrology in Water Resources Systems: Simulation and Optimization. Kluwer, Dordrecht, in press.

Burlando, P. and Rosso, R., 1990. Investigating temporal rainfall patterns under different climate change scenarios via re-scaling point process models. In: Chapman Conf. Hydrologic Aspects of Global Climate Change. Am. Geophys. Union, Lake Chelan, USA.

Camuffo, D., 1984. Analysis of a series of pecipitation in Padua. Clim. Change, 6: 57–78.

Dooge, J.C.I., 1990. Hydrological models and climate change. In: Chapman Conf. Hydrologic Aspects of Global Climate Change. Am. Geophys. Union, Lake Chelan, USA.

Gleick, P.H., 1989. Climate change, hydrology and water resources. Rev. Geophys., 27(3): 329–344.

Gregory, J., Wigley, T.M.L. and Jones, P.D., 1990. Possible changes in the frequencies of extreme rainfall events. In: Chapman Conf. Hydrologic Aspects of Global Climate Change. Am. Geophys. Union, Lake Chelan, USA.

Gupta, V.K. and Waymire, E., 1990. Multiscaling properties of spatial rainfall and river flow distributions. J. Geophys. Res., 95 (D3): 1999–2009.

Jones, P.D., 1984. Riverflow reconstruction from precipitation data. J. Climatol., 4: 171–186.

Kutchment, L.S., 1990. Estimation of the sensitivity of hydrological systems to possible climate changes. Proc. 15th General Assembly Eur. Geophys. Soc., Copenhagen. Ann. Geophys., Spec. Iss., 85.

Lettenmaier, D.P. and Gan, T.Y., 1990. Hydrologic sensitivities of the Sacramento-San Joaquin river basin, California, to global warming. Water Resour. Res., 26(1): 69–86.

Manabe, S. and Wetherald, R.T., 1985. CO_2 and hydrology. Adv. Geophys., 28A: 131–157.

Manabe, S. and Wetherald, R.T., 1987. Large scale changes of soil wetness induced by an increase in atmospheric carbon dioxide. J. Atmos. Sci., 44: 1211–1235.

Manabe, S., Wetherald, R.T. and Stouffer, R.J., 1981. Summer dryness due to an increase of atmospheric CO_2 concentration. Clim. Change, 3: 347–385.

Mimikou, M.A. and Kouvopoulos, Y.S., 1990. Climate change impacts on water resources, Proc. 15th General Assembly Eur. Geophys. Soc., Copenhagen. Ann. Geophys. Spec. Iss., 88.

Mitchell, J.F.B., 1989. The "greenhouse" effect and climate change. Rev. Geophys., 27(1): 115–139.

Mitchell, J.F.B. and Warrilow, D.A., 1987. Summer dryness in northern mid latitudes due to increased CO_2. Nature, 330: 238–240.

Nemec, J. and Schaake, J., 1982. Sensitvity of water resources systems to climate variation. J. Hydrol. Sci., 27(3): 327–343.

Niemczynowicz, J., 1989. Impact of the greenhouse effect on sewerage systems – Lund case study. Hydrol. Sci. J., 34(6): 651–666.

Rodriguez-Iturbe, I., 1986. Scale of fluctuation of rainfall models. Water Resour. Res., 22(9): 15S–37S.

Rowntree, L.B., 1985. A crop-based rainfall chronology for pre-instrumental record in Southern California. Clim. Change, 7: 327–342.

Schneider, S.H., 1989. Global Warming: Are We Entering the Greenhouse Century? Sierra Club Books.

Stockton, C.W. and Bogges, W.R., 1979. Geohydrological implications of climate change on water resources development. Rep. U.S. Army Coastal Eng. Res. Cent., Fort Belvoir.

Vanmarcke, E., 1983. Random Fields: Analysis and Synthesis. MIT Press, Cambridge, MA.

Vehvilainen, B. and Lohvansuu, J., 1990. The effect of climate change on discharges in Finland. Proc. 15th General Assembly Eur. Geophys. Soc., Copenhagen. Ann. Geophys., Spec. Iss., 86.

Wigley, T.M.L., Lough, J.M. and Jones, P.D., 1984. Spatial patterns of precipitation in England and Wales and a revised homogeneous England and Wales precipitation series. J. Climatol., 4: 1–25.

Wilson, C.A. and Mitchell, J.F.B., 1987. Simulated climate and CO_2 induced climate change over western Europe. Clim. Change, 10: 11–42.