IN: P. BALABANIS, A. BRONSTERT, R. CASALE AND P. SAMUELS (EDS.), 1999. *THE IMPACT OF CLIMATE CHANGE ON FLOODING AND SUSTAINABLE RIVER MANAGEMENT*. PROC. SECOND RIBAMOD WORKSHOP, 26-27 FEBRUARY 1998, WALLINGFORD, UK, 297-308.

Impacts of trends and uncertainties in river flooding due to climate change

M.J. Booij and C.B. Vreugdenhil

University of Twente, Dept. of Civil Engineering,
P.O. Box 217, 7500 AE Enschede, The Netherlands
E-mail: m.j.booij@sms.utwente.nl, c.b.vreugdenhil@sms.utwente.nl

Abstract. Projected climate changes will have an effect on frequencies and duration of river flooding and therefore on design criteria for dikes or on risk assessment. In addition to existing sources of uncertainty, extremes and variability of climatological input will change. To deal with this problem the purpose of this project can be split into two main parts. First, to identify possible effects of climate changes on extreme discharges of rivers and particularly the uncertainty involved. Second, to determine the appropriate level of modelling needed to predict such effects taking into account the uncertainties. The major subsystems are climate, catchment and river. Important aspects are the additional uncertainty introduced by each subsystem and the appropriate level of modelling a subsystem. In this paper some preliminary excersises to address these questions with respect to catchment and river are shown, based on very schematic models not representing any particular catchment.

1) Introduction

Projected climate changes will have an effect on frequencies and duration of river flooding and therefore on design criteria for dikes or on risk assessment. In addition to existing sources of uncertainty, extremes and variability of climatological input will change.

Existing sources of uncertainty are related to climate and its variability, catchment hydrology, river behaviour, land use, modelling techniques (accuracy), physical data, risk appreciation and socio-economic conditions. Additional sources of uncertainty due to climate change are related to cause and importance of each change and feedback mechanisms of each component. Important questions are how uncertainties propagate across the system and what the relative importance of several uncertainties is. A related question is how good the various components (climate, catchment and river) have to be modelled in order to get the propagation of uncertainty right.

The purpose of our project is twofold:
1. Identify possible effects of climate changes on extreme discharges of rivers (both high and low flows) and particularly the uncertainty involved.
2. Determine the appropriate level of modelling needed to predict such effects, taking into account the uncertainties: models should be neither too coarse nor too detailed.

We want to avoid using climate change scenarios; rather we will use results from climate model predictions which become available at an increasing rate and level of detail. The major subsystems and related research questions are:
- Climate model. Data from detailed models resulting in precipitation data including frequencies, spatial scales and temporal and spatial correlations. How good are these data, particularly for extreme values, and how large are the uncertainties?
- Catchment model. What is the appropriate level of modelling to produce runoff, taking into account spatial and temporal data of precipitation? What is the additional uncertainty introduced by hydrologic modelling?
- River model. What is the appropriate level of modelling to produce water levels, flooding characteristics, low-flow frequencies etc. What are the uncertainties added by river modelling?

In this paper we show some preliminary exercises to address the latter two questions, based on very schematic models not representing any particular catchment.

2) Subsystems

a. Climate

Current spatial resolutions of general circulation models (GCMs) of about 100-300 km are still too coarse to be used to provide input data for climate change impact studies on a regional scale. A number of approaches have been proposed to arrive at the very high resolution (10-50 km) ideally required for impact studies.

One possibility may be to employ a global GCM with variable horizontal resolution (e.g. Déqué and Piedelievre, 1995) with promising results. Otherwise, methods of downscaling output from coarse-grid GCMs will be needed. One approach is to represent the effect of local forcings empirically, by developing statistical relationships which can be used to link large scale variables from the GCM to local surface variables of interest (e.g. von Storch *et al*., 1993; Bárdossy, 1997). Alternatively, the downscaling can be performed using a modelling technique, in which a high-resolution atmospheric regional climate model (RCM) is nested inside the global GCM. With this technique horizontal resolutions of 20 km (Rotach *et al*., 1997), 0.44° or about 50 km (Jones *et al*., 1997) up to 70 km (Giorgi *et al*., 1992) are achieved.

Simulation of the current climate results in large biases. For example Kittel *et al*. (1998) compared nine GCMs and found an overestimation of the seasonal temperature cycle, on average, by 3.3 °C and, for most regions, a positive precipitation bias in winter (average bias 61 % of observed) and a slightly negative precipitation bias in summer (-4 %). Generally, biases were larger than variation among observations. Christensen *et al*. (1997) compared present-day climate simulations over Europe with seven nested RCMs and found similar biases. Frequently, there is more agreement among models in their response to altered greenhouse forcing than among their simulations of present climate. Besides, a model may simulate the current climate well, but may not simulate future climate with desired quality or vice versa. This may raise the question whether only changes of climate variables should be modelled with acceptable quality or values of variables as well (e.g. Arnell, 1996).

b. Catchment

Transformation of precipitation into stream discharge in a catchment is captured by rainfall-runoff models. Three types have been devised and used: physics-based, metric (or empirical) and conceptual (Wheater *et al*., 1993).

Distributed parameter physics based models include those as the Système Hydrologique Européen (SHE) (Abbot *et al*., 1986) and Institute of Hydrology Distributed Model (IHDM) (Beven *et al*., 1987). The metric approach is characteristically based on the use of observations to define system response and its origin is the unit hydrograph method first developed by Sherman (1932). According to Grayson and Chiew (1994) simple conceptual models are defined as those with fewer than about eight tuneable parameters and complex conceptual models as those with more than eight parameters. Examples of the former are the Surface inFiltration Baseflow model (SFB) (Boughton, 1984) and the TANK model (Sugawara *et al*., 1983). The Stanford Watershed Model (SWM) (Crawford and Linsley, 1966) and MODHYDROLOG (Chiew and McMahon, 1994) are examples of the latter.

In literature numerous model comparisons can be found, but only few address a broad suite of model types and performance on the same catchment (e.g. Loague and Freeze, 1985; Chiew *et al*., 1993). One of the main problems in catchment modelling is overparametrization. Several studies report that a limited amount of parameters is sufficient to represent the transformation of rainfall to streamflow (e.g. Beven, 1989; Jakeman and Hornberger, 1993).

c. River

Flood routing in rivers is a classical subject for which many methods of different levels of complexity are available, ranging from kinematic wave and Muskingum methods through completely dynamic 1-d models (e.g. Jansen, 1994). Further developments include multibranch models (e.g. Estrela and Quintas, 1994), detailed 2-d models (e.g. Vreugdenhil and Wijbenga, 1982). The general tendency is to refine the models to an ever greater extent. However, this does not necessarily lead to more accurate results of the coupled climate-catchment-river system as the weakest link may be elsewhere. Insight in the obtainable accuracy of various models and in the propagation of uncertainties is therefore needed.

3) Spatial correlation and catchment modelling

a. Statistics

When modelling one catchment with a typical size of $L = L_0$ (situation I), it is assumed that within that catchment the precipitation is uniform and thus the spatial correlation coefficient (ρ_s) for the precipitation within that catchment is equal to 1. When modelling the same catchment with two sub-catchments ($L = L_0/2$, situation II), it is assumed that within the sub-catchments the precipitation is uniform and $\rho_s = 1$. The spatial correlation between the two sub-catchments is variable and therefore ρ_s within the total catchment from situation I is variable and can be specified.

It seems reasonable to suppose that if spatial correlation for precipitation increases, annual peak values and standard deviation of daily runoff series will increase as well with the same average annual precipitation. If these annual peak flows are distributed according to the Gumbel extreme value distribution, an estimation of the T-year flood can be made. In the Netherlands, the 1250-year design flood is a measure commonly used in river management. Plotting the annual peak flows on Gumbel paper reveals if they came from a Gumbel distribution, namely if the distribution of the plotted points on the graph is linear. The T-year flood can be derived; analytically and graphically by means of extrapolating the obtained line $T(x) = 1250$. For more information, the reader is referred to Gumbel (1958).

An experiment with a simple climate-catchment-river system has been developed in order to investigate the effect of overestimation of the spatial correlation for precipitation. In the following, system and experiment will be described.

b. Climate modelling: synthetic precipitation

In the climate subsystem the effective rainfall is the only considered variable. A synthetic daily rainfall sequence is generated using a first order Markov chain for rainfall occurrence (e.g. Gabriel and Neumann, 1962) and an exponential distribution (special case Γ distribution) for rainfall amounts (e.g. Stern, 1980).

Markov chains can relate the probability of occurrence of an event (a wet or dry day) to the state of the previous day (first order). This approach is illustrated by the next scheme:

	Next state (t+1)

		Wet (W)	Dry (D)
Initial state (t)	Wet (W)	a	1-a
	Dry (D)	1-b	b

in which $0 < a < 1$ and $0 < b < 1$. If the series of daily rainfall are X_t and probability is P, then:

$$P(X_{t+1} = W | X_t = W) = a$$
$$P(X_{t+1} = D | X_t = W) = 1 - a \text{, etc.}$$

The occurrence chain is obtained using a random generator with a normal distribution and the rainfall amounts are obtained using a random generator with an exponential distribution.

In order to obtain rainfall input with different spatial correlations, two different daily rainfall sequences are routed down two identical, parallel catchments as can be seen in fig. 1. The catchments will be described below. The correlation between these two rainfall sequences can be varied by means of changing the correlation between rainfall occurrence as well as between rainfall amounts. In this way the response of the river catchment to varying spatial correlation can be considered.

c. Catchment modelling

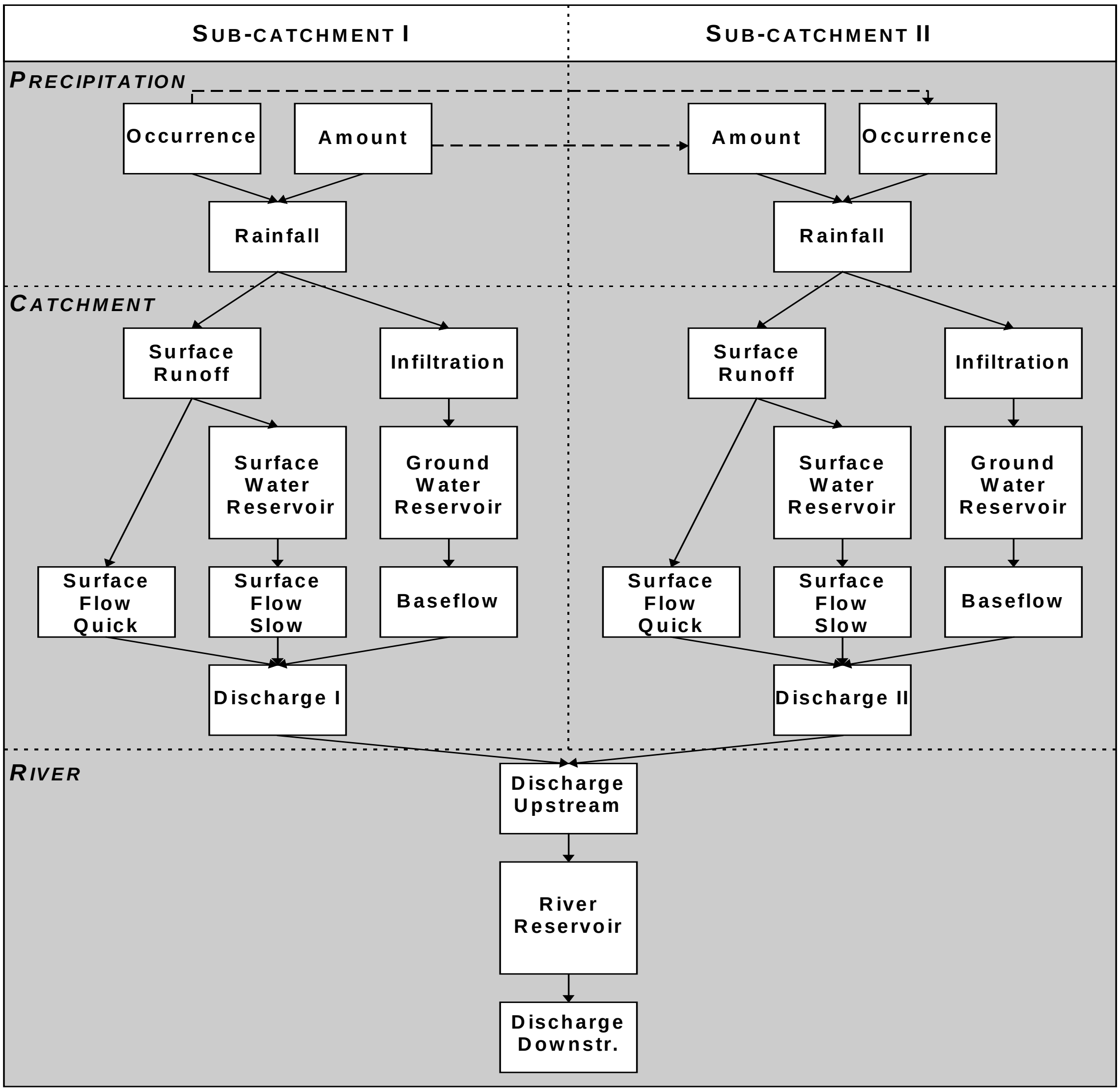

Fig. 1 Climate-catchment-river system flowchart

Two identical catchments have been modelled to transform an effective rainfall into a river discharge. This is achieved by routing that rainfall down a pair of parallel linear reservoirs as illustrated in fig. 1. The relationship between the storage S of a linear reservoir and the outflow Q is

$$S = KQ \tag{1}$$

where K is a time constant for the storage.

The distribution in infiltration and surface runoff is obtained by means of a non-linear distribution coefficient dependent on previous rainfall and the groundwater storage deficit. This latter parameter is the difference between the actual groundwater storage and a certain groundwater storage threshold.

To allow for a sufficiently fast response of the catchment to effective rainfall, a quick surface runoff component directly flowing into the river, besides a relative slow component flowing down a linear reservoir, has been added. The distribution of total surface runoff in a quick and slow component is achieved by applying a constant pre-scribed distribution coefficient.

The river system is represented by a linear reservoir described by eq. (1). The output of this reservoir is the discharge at some distance downstream the confluence of the sub-catchment outflows. This distance is dependent on the time constant K for the storage in the river reservoir.

d. Simulations

To study the effect of different spatial correlations (different spatial resolutions) on the river system, an experiment has been performed. In this experiment, daily rainfall sequences into each sub-catchment are generated; one pair of sequences with $\rho_s = 0.5$ ($L = L_0/2$) and one pair of sequences with $\rho_s = 1.0$ ($L = L_0$; i.e. two identical daily rainfall sequences). For each case ($\rho_s = 0.5$ and $\rho_s = 1.0$) 100 years of 365 daily rainfalls are generated to obtain for each case 100 annual peak discharges Q_p.. With these peak discharges, histograms or frequency distributions are constructed and compared. If these peak flows are distributed according to the Gumbel extreme value distribution an estimation of the T-year flood can be made, as we have seen. We have used a daily time step. The choice of time step is essentially a function of catchment size, data availability and research objective. Flood prediction requires a small time step while for water balance studies larger time steps can be used.

To have some feeling with physical reality, first the climate-catchment-river system is calibrated with rainfall-runoff data of the Meuse river. This calibration is only qualitative; considering real data may give a better impression about what is going on. The 1250-year design flood for the Meuse river at Maastricht has been determined at 3800 m^3/s (Delft Hydraulics, 1994).

In fig. 2, Q_p as a function of return period $T(x)$ been plotted. For both $\rho_s = 0.5$ and $\rho_s = 1.0$ the distribution of the plotted points is linear with only small departures from the straight line, therefore it is assumed that the annual maxima came from a Gumbel distribution. With the derived sample mean and sample standard deviation of the annual maxima, the 1250-year flood $Q_p(1250)$ for $\rho_s = 0.5$ and $\rho_s = 1.0$ is obtained; 3.9 10^3 m^3/s and 4.9 10^3 m^3/s respectively. This has been repeated qualitatively in fig. 2, by means of the extrapolated straight line (dotted part). When comparing these values of $Q_p(1250)$ with the current one used for the Meuse this result does not seem unacceptable.

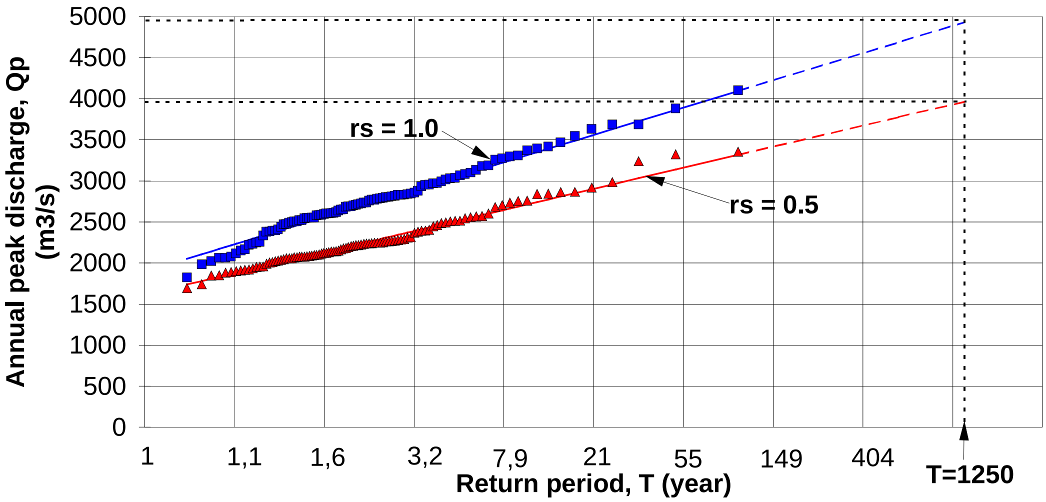

Fig. 2 Annual peak discharge Q_p in m^3/s as a function of return period T in years for spatial correlation coefficient (*rs*) 0.5 and 1.0

The conclusion is that a coarser spatial resolution with corresponding higher internal spatial correlation for precipitation results in a higher 1250-year flood compared to the situation with a finer spatial resolution. The difference is large, even taking into account the relatively large uncertainty in the extrapolated $Q_p(1250)$.

4) River modelling

The question "what is an appropriate model for flood simulation and effects of climate change" can be split into three aspects.

a. Physical effects

The full dynamical flow equations contain a number of physical processes which may or may not be important. Grijsen & Vreugdenhil (1976) presented a way to estimate this (see also Vreugdenhil, 1989). The 1-d equations for mass and momentum conservation are linearized about a reference situation of steady and uniform flow at discharge Q_0. All variables are made dimensionless using reference waterdepth h_0 and discharge as typical scales. A wave with period T is assumed and the space coordinate is made dimensionless using $\sqrt{gh_0}T$ as a length scale. The linearized equations admit a travelling-wave solution, the properties of which depend on three parameters: b = ratio of total to streamflow width (indicating the effect of storage areas not participating in the total flow), the Froude number $F = u_0 / \sqrt{gh_0}$ and a *friction parameter* $K = |u_0| Tc_f / h_0$. Here, u_0 is the reference flow speed and c_f the dimensionless bottom stress coefficient.

As an example, fig. 3 shows the dimensionless wave speed at b = 1. For small K (small friction or short-period waves), the wave speed approaches the well-known characteristic speed. For large K (long-period waves such as most flood waves, high friction), the wave speed approaches the kinematic-wave speed (1.5 u_0 /b in this approximation). A similar picture can be made for wave damping and other typical quantities. Together, these show in which circumstances a kinematic or diffusive wave approximation or rather the full dynamic equations can be used.

Fig. 3 Dimensionless wave speed (*b* = 1); drawn lines: wave travelling upstream, dashed lines: wave travelling downstream

b. Spatial extension

Many rivers have flood plains playing an important part at high discharges. The simplest approximation is that these are just part of the river cross section and therefore included in the storage width. They can also be schematised as separate river channels, leading to a multibranch model. This assumes that flow directions are fixed by topography, but that flow rates are dynamically determined. If flow directions also change with river stage, a full 2-d model might be needed. The effect of these refinements on the water levels (determining dike design levels) can be estimated in a way similar to the previous section.

For example, for curved flows with curvature radius r, the cross-slope is

$$\frac{\partial z}{\partial r} \approx \frac{u^2}{gr} \qquad (2)$$

leading to a lateral waterlevel variation of

$$\Delta z \approx \frac{Bu^2}{gr}$$

If u = 1 m/s, B = 200 m, r = 5 km, this gives Δz = 0.004 m which is irrelevant for flood levels, but might be important for flow distribution within the river bed.

c. Resolution

Numerical resolution (grid size Δx, time step Δt) can be determined in a similar way (Vreugdenhil, 1989). Suppose, for example that a kinematic wave approximation has been found acceptable from section 4.a.:

$$\frac{\partial h}{\partial t} + c\frac{\partial h}{\partial x} = 0 \tag{3}$$

Using the Preissman/Wendroff/4-point scheme:

$$\frac{h_j^{n+1} - h_j^n + h_{j+1}^{n+1} - h_{j+1}^n}{2\Delta t} + \frac{c}{\Delta x}\left\{\theta(h_{j+1}^{n+1} - h_j^{n+1}) + (1-\theta)(h_{j+1}^n - h_j^n)\right\} = 0 \tag{4}$$

the numerical wave properties can be compared with those from the differential equation. For example, fig. 4 gives the numerical wave speed as a function of time step, assuming the grid size to be sufficiently small. Given some required accuracy, you can read the time step required.

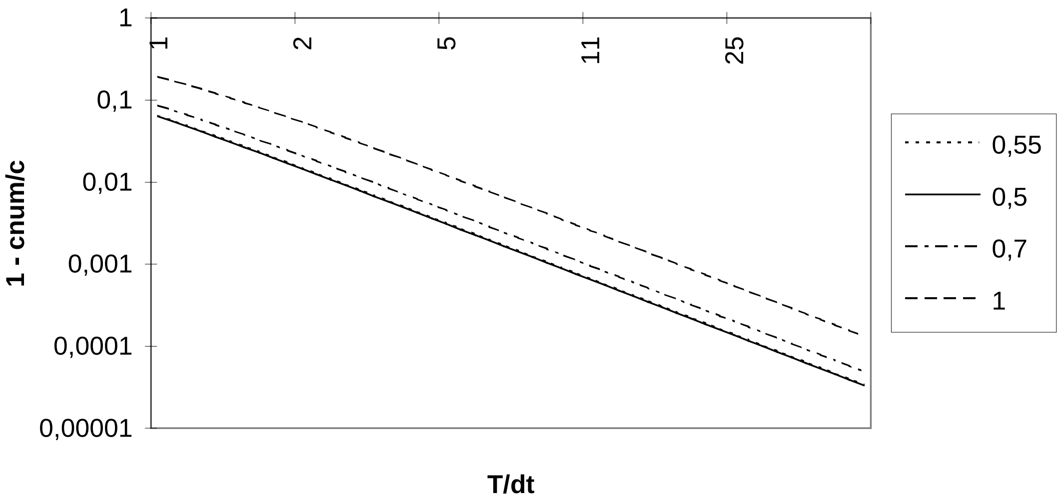

Fig. 4 Error in numerical wave speed cnum as a function of time step for different values of θ

5) Discussion and conclusions

We found that a coarser spatial resolution with corresponding higher internal spatial correlation for precipitation results in a higher 1250-year flood compared to the situation with a finer spatial resolution. With a too coarse resolution, overestimation of spatial correlation for precipitation and consequently overestimation of the T-year flood could occur. This overestimation can result in design criteria for dikes which are too wide with consequent high costs. Similar things may occur when modelling other subsystems like climate and river.

In reality the spatial correlation for precipitation for two points is dependent on the distance between those two points (L) and the scale of the atmospheric disturbance responsible for the precipitation (λ). In general, the higher the ratio L/λ the lower will be ρ_s between two points at distance L. Thus given a certain atmospheric disturbance with scale λ, the catchment has to be refined until a certain L for L/λ causing ρ_s to approximate 1. Meteorological observations should reveal this appropriate refinement for a specific region. When simulating a catchment with a too coarse spatial resolution, overestimation of the spatial correlation for precipitation and the T-year flood may be a significant result.

The question what is an appropriate river model for flood simulation and effects of climate change, can be split into three aspects; physical effects, spatial extension and resolution. In this paper, these aspects have been discussed

References

Abbot, M.B., Bathurst, J.C., Cunge, J.A., O'Connell, P.E. and Rasmussen, J.L., 1986: An introduction to the European Hydrology System SHE, 2: Structure of a physically-based, distributed modelling system. *J. Hydrol.*, **87**, 61-77.

Arnell, N., 1996: *Global warming, river flows and water resources*. Wiley, 224 pp.

Bárdossy, A., 1997: Downscaling from GCMs to local climate through stochastic linkages. *J. Environ. Man.*, **49**, 7-17.

Beven, K.J., 1989: Changing ideas in hydrology - The case of physically-based models. *J. Hydrol.*, **105**, 157-172.

Beven, K.J., Calver, A. and Morris, E.M., 1987: The Institute of Hydrology distributed model. *Rep.* **98**, Inst. of Hydrol., Wallingford, England.

Boughton, W.C., 1984: A simple model for estimating the water yield of ungauged catchments. *Civ. Eng. Trans. Inst. Eng. Aust.*, CE(**26**), 83-88.

Chiew, F.H.S. and McMahon, T.A., 1994: Application of the daily rainfall-runoff model MODHYDROLOG to 28 Australian catchments. *J. Hydrol.*, **153**, 383-416.

Chiew, F.H.S., Stewardson, M.J. and McMahon, T.A., 1993: Comparison of six rainfall-runoff modelling approaches. *J. Hydrol.*, **147**, 1-36.

Christensen, J.H., Machenhauer, B., Jones, R.G., Schär, C., Ruti, P.M., Castro, M. and Visconti, G., 1997: Validation of present-day regional climate simulations over Europe: LAM simulations with observed boundary conditions. *Clim. Dyn.*, **13**, 489-506.

Crawford, N.H. and Linsley, R.K., 1966: Digital simulation in hydrology - Stanford watershed model IV. *Tech. Rep.* **39**, Stanford Univ., Stanford, California.

Déqué, M. and Piedelievre, J.Ph., 1995: High resolution climate simulation over Europe. *Clim. Dyn.*, **11**, 321-339.

Estrela, T. and Quintas, L., 1994: Use of a GIS in the modelling of flows on floodplains. *River flood hydraulics*, J.W.R. White and J. Watts, Editors, Wiley, 177-189.

Gabriel, K.K. and Neumann, J., 1962: A Markov chain model for daily rainfall occurrence at Tel Aviv. *Q. J. R. Meteorol. Soc.*, **88**, 90-95.

Giorgi, F., Marinucci, M.R. and Visconti, G., 1992: A 2xCO_2 climate scenario over Europe generated using a limited area model nested in a general circulation model. II: Climate change scenario. *J. Geophys. Res.*, **97**, 10,011-10,028.

Grayson, R.B. and Chiew, F.H.S., 1994: An approach to model selection. *Preprints, Water Down Under 94 Symposium, Natl. Conf. Publ. 94-10*, Sydney, New South Wales, Australia, Inst. of Eng., 507-512.

Grijsen, J.G. and Vreugdenhil, C.B., 1976: Numerical representation of flood waves in rivers. *Symp. Unsteady Flow in Open Channels*, Newcastle on Tyne, England, WL publ. 165

Gumbel, E.J., 1958: *Statistics of extremes*. Columbia University Press, 375 pp.

Jakeman, A.J. and Hornberger, G.M., 1993: How much complexity is warranted in a rainfall-runoff model? *Water Resour. Res*., **29**, 2637-2649.

Jansen, P.Ph. (Editor), 1994: *Principles of river engineering*. Pitman, 509 pp.

Jones, R.G., Murphy, J.M., Noguer, M. and Keen, A.B., 1997: Simulation of climate change over Europe using a nested regional-climate model. II: Comparison of driving and regional model responses to a doubling of carbon dioxide. *Q. J. R. Meteorol. Soc*., **123**, 265-292.

Kittel, T.G.F., Giorgi, F. and Meehl, G.A., 1998: Intercomparison of regional biases and doubled CO_2-sensitivity of coupled atmosphere-ocean general circulation model experiments. *Clim. Dyn*., **14**, 1-15.

Loague, K.M. and Freeze, R.A., 1985: A comparison of rainfall-runoff modelling techniques on small upland catchments. *Water Resour. Res*., **21**, 229-248.

Rotach, M.W., Marinucci, M.R., Wild, M., Tschuck, P., Ohmura, A. and Beniston, M., 1997: Nested regional simulation of climate change over the Alps for the scenario of a doubled greenhouse forcing. *Theor. Appl. Climatol*., **57**, 209-227.

Sherman, L.K., 1932: Streamflow from rainfall by the unit-hydrograph method. *Eng. News Rec*., **108**, 501-505.

Stern, R.D., 1980: The calculation of probability distributions for models of daily precipitation. *Arch. Met. Geoph. Biokl. Ser*. **B28**, 137-167.

Sugawara, M., Watanabe, I., Ozaki, E. and Katsuyame, Y., 1983: *Reference manual for the TANK model*. Natl. Res. Cent. for Disaster Prev., Tokyo, Japan.

von Storch, H., Zorita, E. and Cubasch, U., 1993: Downscaling of global climate change estimates to regional scales: An application to Iberian rainfall in wintertime. *J. Clim.*, **6**, 1161-1171.

Vreugdenhil, C.B., 1989: *Computational hydraulics*. Springer, 180 pp.

Vreugdenhil, C.B. and Wijbenga, J.H.A., 1982.: Computation of flow patterns in rivers. *J. Hydr. Div. ASCE*, **108** (HY1), 1296-1310.

Waterloopkundig Laboratorium, 1994: *Investigation of the Meuse flood, report 4: hydrological aspects* (in Dutch). Ministerie van Verkeer en Waterstaat, the Netherlands.

Wheater, H.S., Jakeman, A.J. and Beven, K.J., 1993: Progress and directions in rainfall-runoff modelling. *Modelling change in Environmental Systems*, A.J. Jakeman, M.B. Beck and M.J. McAleer, Editors, John Wiley, 101-132.