Regionalization in Hydrology (Proceedings of a conference held at Braunschweig, March 1997).
IAHS Publ. no. 254, 1999.

Effects of spatial data resolution on the calculation of regional water balances

HELGE BORMANN, BERND DIEKKRÜGER
University of Bonn, Geographical Institutes, Meckenheimer Allee 166, D-53115 Bonn, Germany
e-mail: h.bormann@uni-bonn.de

OTTO RICHTER
Technical University of Braunschweig, Institute for Geography and Geoecology, Langer Kamp 19c, D-38106 Braunschweig, Germany

Abstract. Regional simulation demands the reduction of the amount of data and computer time required. In this study the effect of data availability on the annual water balance is investigated. It was found, in a given drainage basin, that the areal mean evapotranspiration and groundwater recharge can be calculated using soil maps at different resolutions. In this way an upscaling of these hydrological processes is possible, in principle without a bias in the simulation results. But in the drainage basin investigated distinct spatial patterns of the differences in hydrological behaviour were observed which correspond to the geological origin. The runoff production therefore showed a strong dependence on the natural variability of soil hydraulic properties and the concept of effective parameters fails in this case. Assuming a lognormal probability distribution of the decisive parameter "saturated hydraulic conductivity" however, and choosing a suitable sampling scheme (Latin Hypercube method) this heterogeneity can be taken into account in modelling the hydrological behaviour.

INTRODUCTION

Scaling up distributed hydrological models often leads to constraints concerning the computer time and data availability required (Kirkby, 1999). This is often compensated for by a change in the model concept (distributed *vs* lumped models). When changing the model concept the problem arises of how to aggregate input data and how to transfer parameters between the different concepts (Wood, 1995). Considering these difficulties the aim of this study is the application of a hydrological model at different scales ranging from the point scale via the basin to the regional scale, without changing the underlying principles. If a model concept is to be applied for a large range of scales, then the model input has to be scaled up appropriately. If available, this can be accomplished by using effective parameters on the upper scale. Effective parameters are defined so that the mean output of the heterogeneous system matches the output of the homogeneous system of the upper scale (Blöschl, 1996). If the system is nonlinear, as in our case, the determination of effective parameters is not always possible. This is complicated by the fact that scaling parameters from soil maps, like texture and bulk density, fail because this results in unrealistic model parameters.

Assuming that the same processes act at different scales, a physically based description of the processes is chosen to be able to separate the hydrological components. This is realized by using a physically based SVAT-scheme (Soil-Vegetation-Atmosphere-Transfer) which calculates the water fluxes of homogeneous areas (homogeneous concerning physical soil properties, land use and morphology) coupled with routing routines. Increasing the size of the simulated area results in an increase in the number of simulation units, so that it can become difficult to manage the amount of data and simulation runs. Some means for the reduction of CPU and data demand is required with a minimal deterioration of the simulation results.

Often only a coarser resolution of input data is available for larger areas. In this case it may be possible to disaggregate the data in order to generate a small-scale pattern. This fails in the case of soil maps because each mapping unit has one representative soil profile. Because this is not a real soil profile it does not have coordinates and therefore cannot be used for interpolation. Although effective parameters may be used for calculating transport processes in the soil as well as for calculating evapotranspiration (Yeh *et al.*, 1985) it is well accepted that this concept fails for runoff generation (Sivapalan & Wood, 1986; Goodrich, 1990; Smith *et al.*, 1990). This is because the time of ponding and the total infiltration depend not only on soil properties but also on the rainfall intensity.

The problem is the dependence of scale and information density: heterogeneity of relevant input data increases with scale but information density decreases (Table 1). So we investigate whether a decrease in information density leads to a bias in the simulation results or to non-applicability of the model concept for a given scale. We examine the spatial resolution of soil data in this study, in relation to the temporal and spatial resolution of meteorological input data published by Bormann *et al.* (1996). The main results of that work are:

(a) no relevant information loss regarding evapotranspiration and groundwater recharge occurs if disaggregated weather data are used (from daily to hourly resolved values), and

(b) the transfer of meteorological data from nearby stations (precipitation and air humidity in particular) to a local station may lead to systematic deviations in the simulation results. These deviations have to be eliminated before a data transference is feasible.

Table 1 Scale dependency of information density.

Scale of information	Soil data	Land use data	Climate data
Field scale	Measurements	Crop calendar	High (spatial and temporal) resolution of a local station
Small drainage basin	Soil map 1:5000	Land use mapping, agricultural statistics	Daily mean values of local stations
Landscape	Soil map 1:50 000	Satellite and aerial pictures, agricultural statistics	Daily mean values
River basin	Soil map 1:200 000	Satellite and aerial pictures, agricultural statistics	Daily mean values

SIMULATION MODEL

The model system used in this study is a one-dimensional SVAT-scheme called SIMULAT. A detailed description is given by Diekkrüger & Arning (1995). SIMULAT considers the following processes represented by the approaches in parenthesis:

- Potential evapotranspiration (*ETP*, Penman-Monteith method). The *ETP* is separated into the potential evaporation and the potential transpiration using an exponential function of leaf area index. The reduction of potential evaporation to actual values is performed using an empirical two stage approach (Ritchie, 1972) and the reduction of potential transpiration to actual transpiration uses the Feddes approach (1978), which is based on the actual soil moisture conditions in the root zone.
- Infiltration (analytical solution of the Richards' equation according to Smith & Parlange, 1978).
- Soil water flow (fully implicit numerical solution of the Richards equation using finite differences). The parameters of the retention and conductivity curves are estimated from basic soil data by using pedo-transfer functions (e.g. Rawls & Brakensiek, 1985) which are well established and tested for various soils (Tietje & Tapkenhinrichs, 1993; Diekkrüger & Arning, 1995).
- Snow melt (degree day method; Smith, 1992).
- Interflow (Darcy's law).
- Baseflow (Dupuit-Forchheimer approximation, Van Schilfgaarde, 1970).
- Plant growth (an empirical approach to provide the parameters necessary for the Penman-Monteith equation like plant height, leaf area index and rooting depth). Plant development is described by a linear interpolation between given values determined from measurements and literature data.

In the model the soil is divided into a number of soil horizons (2–6) and numerous computation layers (40–60). Lateral connections between the ecotopes (homogeneous areas concerning physical soil properties, land use and morphology) are neglected and flowpaths of neighbouring areas are assumed to be independent. First results of the application of the actual version of the model system in the Upper Leine drainage basin are presented in Bormann *et al.* (1999).

INFORMATION DENSITY: SPATIAL RESOLUTION OF SOIL MAPS

Calculations of annual evapotranspiration and groundwater recharge have been carried out using digital soil maps at different resolutions. The databases used were digital soil maps at scales of 1:5000, 1:50 000, 1:200 000 and a geological map 1:25 000 of the upper 2 m. In Germany digital soil maps include one representative soil profile for each soil mapping unit. This profile consists of different soil horizons and their properties such as thickness, texture, porosity and content of organic matter. These properties are translated into hydrological model parameters by using pedo-transfer functions (e.g. Rawls & Brakensiek, 1985). The 1:5000 soil map is based on a soil survey and is available for agricultural areas only. The other soil maps (at coarser scales) were aggregated from finer scales without field observations and without consideration of special criteria for hydrological modelling. The availability is different: coarser scales (1:1 000 000 and 1:200 000) are available for the whole of West Germany, but soil maps at 1:25 000 and 1:50 000 are only available for some areas.

The calculations were carried out for a small drainage basin in northern Germany (Eisenbach near Uelzen, North Germany), which has an area of 16 km^2 and which is dominated by sandy soils and deep groundwater levels. The main land uses in this region are summer barley, potatoes and grassland. In order to isolate the effect of the resolution of the soil map the assumptions made for this study were a homogeneous land use (summer barley, potatoes) and a homogeneous weather input across the entire drainage basin (e.g. precipitation). The number of mapping units were 442 (soil map 1:5000), 7 (soil map 1:50 000), 4 (soil map 1:200 000) and 8 (geological map 1:25 000). Actual evapotranspiration (*ETA*) and groundwater recharge (*GWR*) were calculated for two years (1990, 1991) based on different soil maps. Due to deep groundwater levels we choose the lower boundary condition as free drainage. Lateral connections between the computational units (ecotopes) were neglected because of the moderate slopes and high vertical hydraulic conductivities of the soils.

The three soil and geological maps investigated lead to the same areal mean values of *ETA* and *GWR* (Table 2). The standard deviation of the differences between certain map scales is an indicator of the similarity of the simulation results. These increase with scale difference. Although the mean values resulting from various databases are similar there are selective differences of up to 15% which show a spatial pattern. Figures 1 and 2 show two examples of spatial patterns of *ETA*-differences between distinct soil maps. The distribution of the annual differences corresponds to the geological origin of the drainage basin which consists of ground moraine in the northern part, sand in the southern part and an end moraine in the middle. So there may be a shift in hydrological properties at different scales due to data resolution.

The interpretation of geological maps (upper 2 m) for hydrological modelling is difficult because the translation of geological properties into soil physical parameters causes problems. As can be seen the standard deviation of the differences between the two soil maps is less than the standard deviation between the basic soil map (1:5000) and the geological map (Table 2) although the scales are less similar. If available, soil data should be preferred for hydrological modelling.

The spatial variability of precipitation data has not been considered in this study. The precipitation data have been assumed to be constant in the small test area. For larger areas this assumption has to be abandoned and spatial variability of rainfall

Table 2 Results of areal mean and standard deviation of *ETA* (mm $year^{-1}$) (calculated from a grid based map) based on different scales of digital geological (1:25 000) and soil maps (1:5000 and 1:50 000).

Scale	Areal mean summer barley 1990	Standard deviation	Areal mean potatoes 1991	Standard deviation
1:5000	538.5	22.9	513.3	26.7
1:25 000	535.9	24.4	508.0	21.9
1:50 000	536.5	14.2	513.9	18.4
Difference (1:5000–1:25 000)	2.6	29.9	5.3	31.1
Difference (1:5000–1:50 000)	2.0	20.7	–0.6	24.9

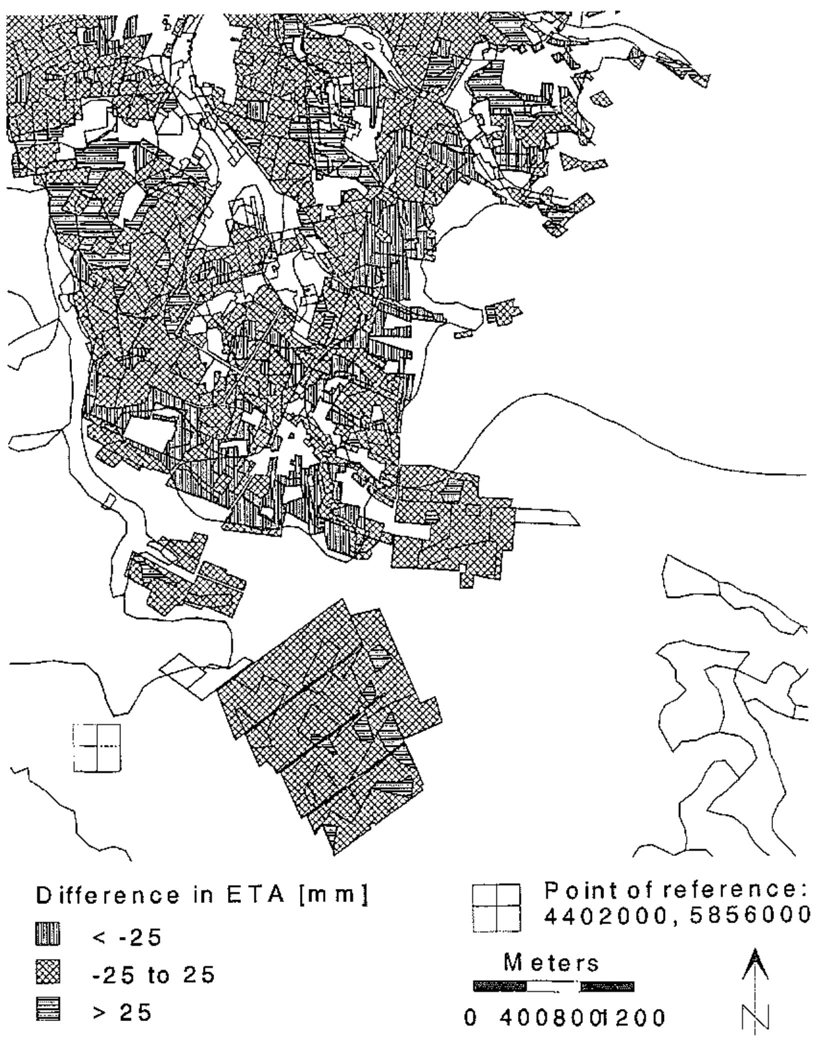

Fig. 1 Spatial pattern of the annual *ETA*-differences between the soil maps 1:5000 and 1:50 000 (1990, summer barley).

has to be taken into account because it is one main reason for the spatial variability of *ETA* and *GWR*. It is possible to create a link between areas with under-/overestimated hydrological quantities and geological mapping units depending on the given scale.

Based on the assumptions of this study (homogeneous precipitation input, homogeneous plant cover, no lateral flow components in the test area and the choice of free drainage for the lower boundary condition) the results for groundwater recharge show the same spatial pattern as actual evapotranspiration does. In this case the use of an upscaled model input reduces the required computer time by about 95%. This is due to the fact that homogeneous land cover and climate data were assumed. In cases where this assumption is not feasible one has to develop other regionalization techniques. One of the possible solutions is presented by Bormann *et al.* (1999).

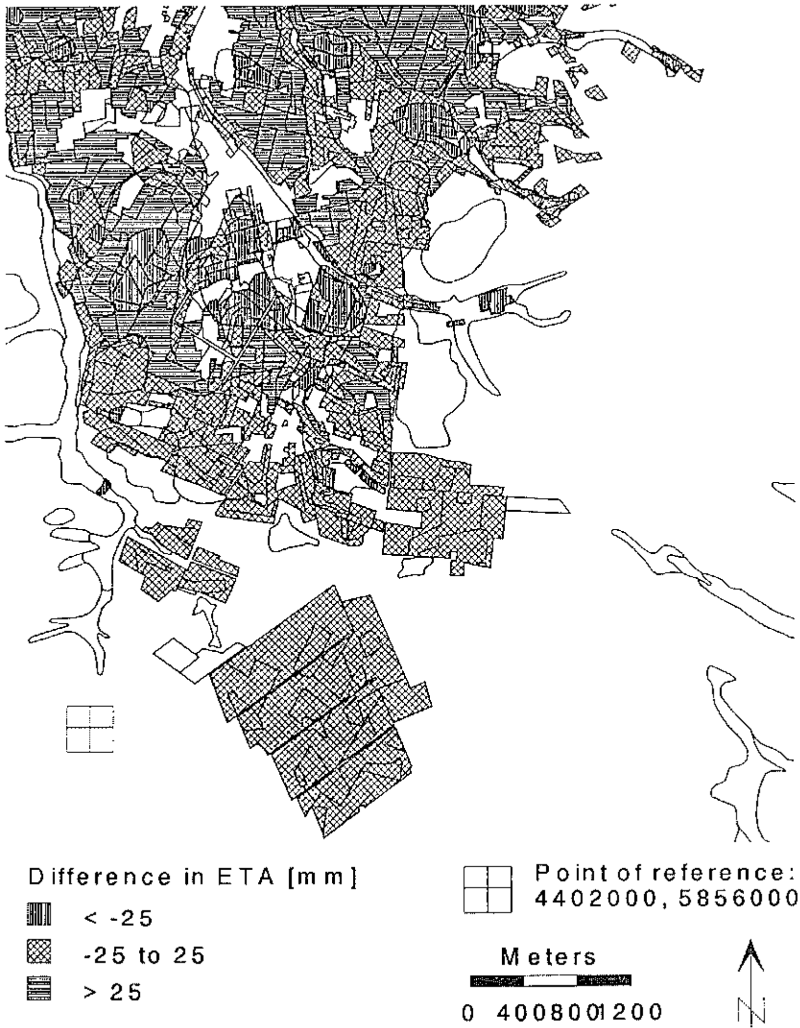

Fig. 2 Spatial pattern of the annual *ETA*-differences between the soil map 1:5000 and the geological map 1:25 000 (1990, summer barley).

ASPECTS OF SOIL HYDRAULIC VARIABILITY ON RUNOFF PRODUCTION

In the first part of this study the small-scale variability of soil hydraulic properties was neglected because this produces only minor effects on evapotranspiration and groundwater recharge. This was also supported by the soil properties (sandy soils, high permeability, deep groundwater levels) of the investigated drainage basin. While investigating drainage basins with different geological and topographical conditions, other hydrological processes (e.g. surface runoff production) have to be included. These are strongly affected by the variability of the soil mapping units which are mapped as homogeneous areas.

In our approach the numerical solution of Richards equation is used for calculating water transport in the soil. This model is linked to the infiltration equation of Smith &

Parlange (1978) which was used for calculating runoff production. Both approaches are linked via the water content of the upper horizon. For surface runoff production we assume that the saturated hydraulic conductivity K_s is the dominant parameter (Smith & Diekkrüger, 1996). As confirmed by numerous investigations, the distribution of K_s values of most soils can be assumed to be lognormal.

Applying this statistical distribution function to the analytical solution of the Richards equation a sampling scheme is needed. Because the Monte-Carlo Technique needs a large number of simulation runs a more suitable approach is the Latin Hypercube method (McKay *et al.*, 1979). The distribution of K_s is divided into a number (e.g. 10) of representative segments and the infiltration equation is solved for each segment value. The segmental division is based on dividing the probability density function into equal areas (Fig. 3). The first moment of each segment area is used as the representative value for its respective segment. In this application a one-dimensional application is used. The K_s distribution is represented by 10 segments and thus 10 simulations are necessary. The representative infiltration rate is the average value of the 10 simulations. Because the infiltration equation can be solved much faster than the numerical solution of the Richards equation the approach described here does not increase the required computer time significantly.

The sensitivity of the standard deviation σ_{ks} of K_s with regard to the surface runoff production and to the actual evapotranspiration is given by Fig. 4. The values shown are sums from a 3.5 year period (1993–1996) of a silty clay under northern German conditions. While the runoff production as well as the infiltration is strongly affected by σ_{ks}, the evapotranspiration is not.

It is not possible to calculate the runoff production only from the mean K_s, or any representative/effective value. This is confirmed by the results of a series of simulations shown by Table 3. At low rainfall intensities in particular there are significant differences between the simulation results based on the Latin Hypercube

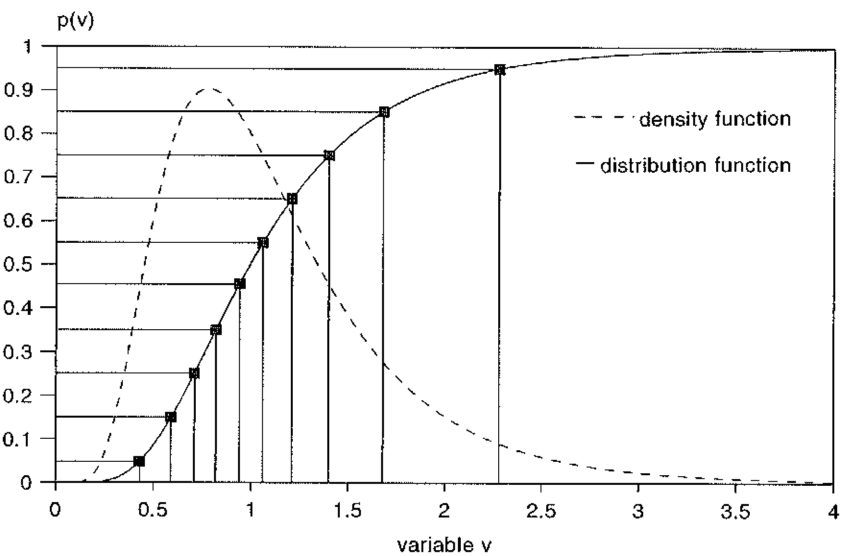

Fig. 3 Graphical description of the Latin Hypercube sampling scheme. Sub-areas of equal mass are created to represent the probability distribution function for each relevant parameter.

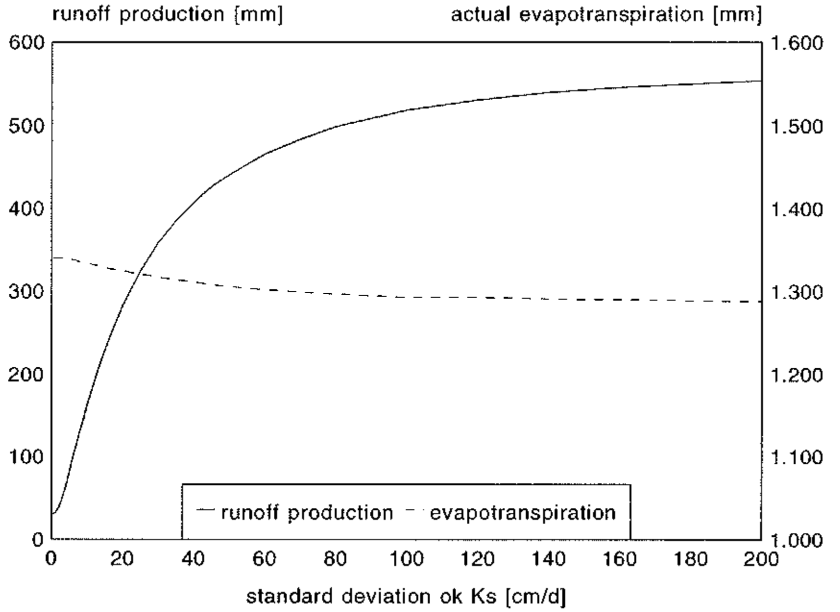

Fig. 4 Sensitivity of runoff production and actual evapotranspiration regarding the standard deviation of the saturated conductivity (σ_{ks}).

Table 3 Results of runoff production simulations using the Latin Hypercube method. Parameters of the probability function are μ_{ks} = 10 cm day^{-1}and σ_{ks} = 10 cm day^{-1}, the design storm had a uniform rainfall intensity and a duration of 2 hours.

Rainfall intensity (cm day^{-1})	Runoff production Latin Hypercube (cm day^{-1})	Runoff coefficient Latin Hypercube (–)	Runoff production -expected value of K_s (cm day^{-1})	Runoff coefficient - expected value of K_s (–)	Effective K_s value (cm day^{-1})
12	0.24	0.0150	0	0	2.80
18	1.32	0.0757	0	0	4.60
24	3.84	0.1614	0.36	0.0135	5.90
30	7.44	0.2491	3.48	0.1163	6.80
36	11.88	0.3290	8.04	0.2248	7.30
42	16.68	0.3972	13.32	0.3164	7.80
48	21.72	0.4533	18.84	0.3918	8.10
54	27.12	0.5027	24.48	0.4525	8.35
60	32.76	0.5457	30.24	0.5046	8.50
72	44.28	0.6144	41.88	0.5821	8.57
84	55.92	0.6661	53.52	0.6369	8.63
96	67.80	0.7058	65.40	0.6819	8.68
108	79.68	0.7377	77.40	0.7172	8.73
120	91.56	0.7635	89.40	0.7455	8.80

method compared to simulation results based only on the expected value of K_s. Using the Latin Hypercube method and consequently taking care of the natural variability of soil properties causes a higher runoff production than considering only average behaviour. Accordingly the runoff coefficients are higher, particularly for low rainfall intensities. This context is endorsed by the effective values of K_s which are defined as the parameters which compute the same runoff volume as computed for the heterogeneous case. During all simulations the effective value of K_s is smaller than the expected value of K_s (10 cm day^{-1}), and the effective K_s decreases as rainfall intensities decrease.

CONCLUSIONS

There is no bias in the simulation results when different databases are used. But the differences in *ETA* and *GWR* increase with scale difference. These differences have a distinct spatial pattern which corresponds to the geological origin. This study therefore enables us to create a link between under- and overestimating areas concerning simulation results and geological mapping units depending on the given scale. Looking at the runoff production it is not sufficient to consider only the expected values of the decisive parameters (e.g. K_s). The natural variability has to be taken into account. This can be considered by the application of a statistical distribution function of the parameters and a suitable sampling scheme. Thus representative parameters are conceived. This enables us to calculate the water balance of large drainage basins by using physically based and distributed simulation models without neglecting the spatial heterogeneity. It considerably reduces the number of simulation runs without reducing the quality of the simulation results. Thus physically based hydrological models can be used at fine and coarse scales, ranging from point scale to the river basin.

Acknowledgements This work was supported by the Deutsche Forschungsgemeinschaft (DFG-Schwerpunktprogramm, Regionalization in Hydrology). The authors also thank the Collaborative Research Programme 179 (Water and matter dynamics in agricultural ecosystems) for providing weather and soil data for the test basin.

REFERENCES

Blöschl, G. (1996) Scale and scaling in hydrology. *Wiener Mitteilungen* Band 132.

Bormann, H., Diekkrüger, B. & Richter, O. (1996) Effects of data availability on estimation of evapotranspiration. *Phys. Chem. Earth* 21, 171–175.

Bormann, H., Diekkrüger, B., Renschler, C. & Richter, O. (1999) Regionalization concept for the prediction of large scale water fluxes. In: *Regionalization in Hydrology* (ed. by B. Diekkrüger, M. J. Kirby & U. Schröder) (Proc. Conf. Braunschweig, March 1997). IAHS Publ. no. 254 (this volume).

Diekkrüger, B. & Arning, M. (1995) Simulation of water fluxes using different methods for estimating soil parameters. *Ecol. Modelling* 81, 83–95.

Feddes, R. A., Kowalik, P. J. & Zaradny, H. (1978) Simulation of field water use and crop yield. *Simulations Monograph*. Pudoc, Wageningen, The Netherlands.

Goodrich, D. C. (1990) Geometric simplification of a distributed rainfall-runoff model over a range of basin scales. PhD Thesis, The University of Arizona, Tuscon, Arizona, USA.

Kirkby, M. J. (1999) Translating models from hillslope (1 ha) to catchment (100 km^2) scales. In: *Regionalization in Hydrology* (ed. by B. Diekkrüger, M. J. Kirby & U. Schröder) (Proc. Conf. Braunschweig, March 1997). IAHS Publ. no. 254 (this volume).

McKay, M. D., Beckman, R. J. & Conover, W. J. (1979) A comparison of three methods for selecting values of input variables in the analysis of output from a computer code. *Technometrics* 21(2), 239–245.

Rawls, W. J. & Brakensick, D. L. (1985) Prediction of soil water properties for hydrological modelling. In: *Proc. Symp. on Watershed Management in the Eighties* (ed. by E. Jones & T. J. Ward), 293–299. Denver, Colorado, USA.

Ritchie, J. T. (1972) A model for predicting evaporation from a row crop with incomplete cover. *Wat. Resour. Res.* 8(5), 1204–1213.

Sivapalan, M. & Wood, E. F. (1986) Spatial heterogeneity and scale in infiltration response of catchments. In: *Scale Problems in Hydrology* (ed. by V. K. Gupta, I. Rodriguez-Iturbe & E. F. Wood), 81–106. D. Reidel, Dordrecht, The Netherlands.

Smith, R. E. (1992) OPUS: an integrated simulation model for transport of non-point source pollutants at the field scale. Vol. 1, Documentation, 120, USDA, ARS-98.

Smith, R. E. & Diekkrüger, B. (1996) Effective soil water characteristics and ensemble soil water profiles in heterogeneous soils. *Wat. Resour. Res.* 32(7), 1993–2002.

Smith, R. E., Goodrich, D. C. & Woolhiser, A. (1990) Areal effective infiltration dynamics for runoff on small catchments. In: *Trans. 14th International Congress of Soil Science, Kyoto, Japan*, vol.I, 22–27.

Smith, R. E., & Parlange, J. Y. (1978) A parameter-efficient hydrological infiltration model. *Wat. Resour. Res.* **14**(3), 533–538.

Tietje, O. & Tapkenhinrichs, M. (1993) Evaluation of pedo-transfer functions. *Soil Sci. Soc. Am. J.* 57, 1088–1095.

Van Schilfgaarde, J. (1970) Theory of flow to drains. In: *Advances in Hydroscience* vol. 6 (ed. by V. T. Chow), 43–106.

Wood, E. F. (1995) Heterogeneity and scaling land-atmospheric water and energy fluxes in climate systems. In: *Space and Time Scale Variability and Interdependencies in Hydrological Processes* (ed. by R. A. Feddes), 3–19. Cambridge University Press, Cambridge, UK.

Yeh, T. C., Gelhar, L. W. & Gutjahr, A. L. (1985) Stochastic analysis of unsaturated flow in heterogeneous soils. I. Statistically isotropic media. *Wat. Resour. Res.* **21**, 447–456.