Regional scale hydrology: II. Application of the VIC-2L model to the Weser River, Germany

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Abstract This paper describes the application of a grid network version of the two-layer Variable Infiltration Capacity (VIC-2L) macroscale hydrological model. The VIC-2L model is implemented on a rotated grid, which is compatible with the weather forecast and climate model REMO (Regional Model), a joint project of the German Weather Service (DWD), GKSS Research Centre and the Max-Planck-Institute for Meteorology, Hamburg. Observed surface meteorological data in the Weser River basin are used to force the model off-line on a daily time step. After a 22-month calibration period, simulated and measured streamflow data are compared for a 12-year period. The resulting predictions compare well with observations at daily, monthly and annual time scales. A sensitivity analysis is presented.

Hydrologie à l'échelle régionale: II. Application du modèle VIC-2L sur la rivière Weser, Allemagne

Résumé Cet article décrit l'application du schéma de surface VIC-2L (Variable Infiltration Capacity). Le schéma de surface VIC-2L est implémenté sur un domaine compatible avec le modèle météorologique de prévision et de climat REMO (REgional MOdel), un projet commun du Service Météorologique Allemand (DWD), du centre de recherche GKSS, et de l'Institut Max-Planck pour la météorologie à Hamburg. Les paramètres météorologiques de surface observés sur le bassin versant de la Weser sont utilisés pour forcer le modèle au pas de temps journalier. Après une période de calibration de 22 mois, les débits simulés et observés sont comparés sur une période de 12 ans. Les résultats des simulations se comparent bien avec les observations aux échelles journalière, mensuelle, et annuelle. Une analyse de sensibilité a été faite.

INTRODUCTION

This study focuses on the ability of a grid-net version of the two-layer Variable Infiltration Capacity (VIC-2L) macroscale hydrological model (Liang, 1994; Liang *et al.*, 1994; Nijssen *et al.*, 1997) coupled to the routing model developed by Lohmann *et al.* (1996) to predict streamflow in the Weser River catchment (37 495 km^2) above the gauging station at Intschede and its tributaries (BALTEX, 1995). The model grid spacing in this study is 1/6 degree (*c.* 18 km) on a rotated grid to be compatible with

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the atmospheric regional scale model (REMO) used for numerical studies within the Baltic Sea Experiment (BALTEX) (Majewski, 1991; Karstens *et al.*, 1997; BALTEX, 1995).

Several recent papers (Wetzel, 1994; Nijssen *et al.*, 1997; Abdulla *et al.*, 1996) have tested land surface parameterizations (LSPs) off-line (that is, using observed surface meteorological forcing) and compared the resulting predictions of streamflow with observations. These comparisons have generally been performed by aggregating high frequency simulations and observations to monthly, seasonal or annual time steps. As hydrological processes like evapotranspiration, soil moisture movement, transport of water in rivers and snowmelt occur on much shorter time scales, it could be argued that LSPs should be tested on smaller time steps. Furthermore, in previous papers the routing model has not been checked independently from the runoff production model.

This study compares measured and modelled data on daily, monthly and yearly time steps. The VIC-2L model was run in its water balance mode, i.e. surface energy fluxes other than evapotranspiration were not computed, and the ground heat flux was neglected. A sensitivity study shows the influence on the model results of a variation in the meteorological input data and the soil parameters of the VIC-2L model. The results underline the need for large-scale validation data of evapotranspiration and soil water storage as well as large-scale hydrological parameters.

MODEL SETUP

Basin and data description

The Weser River above Intschede covers an area of 37 495 km^2 in north central Germany (coordinates: 8–11°E; 50.5–53.5°N). The annual mean precipitation in its moist, maritime climate ranges from some 600 mm in the northeastern part to about 1400 mm in the Harz and Rothaargebirge mountains (Fig. 1), the mean annual evaporation ranges from about 350°mm to about 700 mm. The mean annual runoff ranges from about 150 mm year^{-1} to about 800 mm year^{-1}, with an average between 250 and 290 mm year^{-1}, close to the climatology value of 300 mm year^{-1} given in Brutsaert (1982).

Daily data from 185 precipitation measurement stations and 33 climate stations (2 m air temperature, air pressure, 10 m wind speed, relative humidity, daily sunshine duration, precipitation) were used for the hydrological model forcings. In grid boxes with no precipitation or climate station, either the closest station, or an average value of the surrounding stations was used. In grid boxes with more than one station, the values were averaged. To account for the difference in height of the grid box mean elevation and the height of the measurement stations, temperature was lapsed at -6.5°C km^{-1}. Pressure was lapsed using the hydrostatic equation including a temperature dependent scale height. No elevation adjustments were performed for wind speed, relative humidity, sunshine duration and precipitation. To calculate the

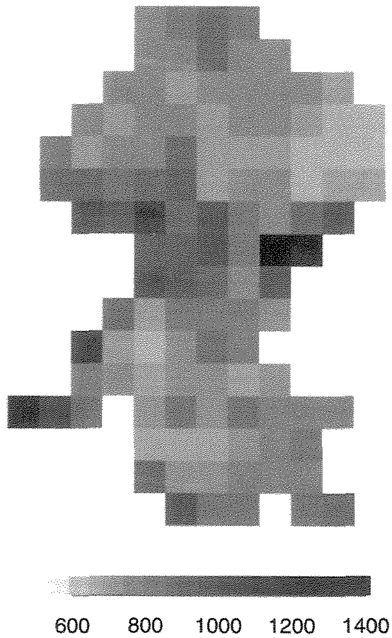

Fig. 1 Mean distribution of precipitation (mm year^{-1}) in the Weser River catchment for the years 1981–1993.

net radiation the suggestions of Shuttleworth (1993) were followed, originally proposed by Prescott (1940) for the shortwave part and Brunt (1932) for the longwave part.

River discharge data were made available by the German Bundesanstalt für Gewässerkunde (BfG), the Niedersächsisches Landesamt für Ökologie (NLfÖ) and the Rheinisch-Westfälische Technical University Aachen (RWTH). A schematized river network of the Weser River showing the location of ten selected gauging stations is shown in Fig. 2.

Model parameters

The hydrological model contains free parameters that have to be determined or calibrated. This section describes the parameter selection and calibration procedure. The model was calibrated for the period January 1993–October 1994 and then tested and verified for the years 1981–1992. The horizontal transport model can be calibrated separately from the VIC-2L model. Its impulse response function is used to transform the VIC-2L direct surface runoff and the baseflow into streamflow, which afterwards can be compared with measured data. The setup for the calibration procedure is described by Lohmann *et al.* (1998). The two routing model parameters b and k are optimized together with the VIC-2L parameters.

Vegetation parameters To be compatible with the current parameterization of

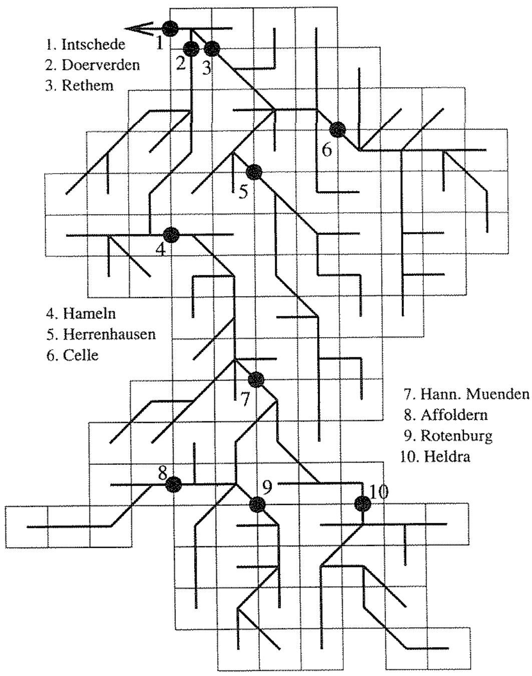

Fig. 2 Schematized flow directions for the rotated 1/6 degree grid showing gauging stations which have been used for the calibration and validation procedure.

the weather forecast models EM/DM of the German Weather Service (DWD) the vegetation parameters of the EM/DM were used whenever possible (DWD, 1995). The Leaf Area Index (LAI) varies between 0.2 in winter and 4.0 in summer. The roughness length in each grid box was determined as the sum of the roughness lengths due to orography and vegetation. There are no root density distributions available, and so the fraction of the roots was chosen to be 0.5 in each soil layer. The specified upper soil layer thickness is approximately one third of the lower one, which reflects the fact that there are more roots closer to the surface. The canopy resistance was set to 200 s m^{-1} divided by the leaf area index as suggested by Shuttleworth (1993). The critical soil moisture value above which transpiration was not affected by soil moisture stress was set to 0.54 of the maximum soil moisture content independently of the soil. The permanent wilting point was set to 0.25 of the maximum soil moisture content. Both values are close to ones reported by Warrilow *et al.* (1986) for medium to fine textured soils. There is no general agreement for the choice of these large-scale parameters (Viterbo, 1996; Shuttleworth, 1993). For the architectural resistance an average value of 4 s m^{-1} was chosen; values reported by Ducoudre *et al.* (1993) range from 2 s m^{-1} for grassland to 50 s m^{-1} for forested sites.

Soil parameters The soil parameters were estimated manually by comparing daily hydrographs of modelled and measured streamflow. However, the parameters were constrained to be within physically realistic ranges. As noted by Liang (1994) and Nijssen *et al.* (1997) the parameters with the largest effect on the hydrograph are the infiltration capacity shape parameter β, the maximum soil moisture in layers 1 and 2, the baseflow parameters d_1, d_2, d_3 and the fraction of maximum subsurface flow W_S.. As there are no detailed data sets available from which some of theses parameters could have been deduced, they were all subject to calibration. Table 1 shows the ranges of all the VIC-2L parameters used in this study.

Table 1 VIC-2L soil parameters for the Weser basin and their influence on the yearly runoff volume. A "+" in runoff influence means that runoff is increased when the parameter is increased. No general statement can be made about K_S and B_p.

Soil parameter	Units	Range	Runoff influence
Infiltration parameter β	-	0.12–0.16	+
Max. soil moisture upper layer	mm	60–90	-
Max. soil moisture lower layer	mm	190–290	-
$T_{1/2}$ lower layer	day	200–300	-
Fraction of maximum soil moisture W_S	-	0.70–0.75	-
Saturated hydraulic conductivity K_S	mm day^{-1}	30.0–200.0	
Pore size distribution index B_p	-	0.25–0.5	

The saturated hydraulic conductivity K_S and the porosity also strongly influence runoff production. These parameters were fixed depending on the soil type, which was selected by the DWD from the FAO Soil Map of the World (FAO, 1978). These parameters were set in ranges which have been suggested by various authors, but on the large scale they must be seen as effective parameters.

Routing model calibration The parameters of the river routing model were optimized using measured streamflow data and precipitation data. As the iterative scheme is not able to include snow processes, the scheme was applied from February to November 1993. For the river routing C and D were either estimated or optimized with measured river discharge data.

Figure 3 shows the sensitivity of the separated slow flow to different routing parameters b and k. the influence on the impulse response function is shown in Fig. 4. The main influence of the parameters b and k on the impulse response function is not the overall shape, but the peak height at two or three days corresponding to a shift of water from the early peaks to the following days. The parameters were chosen in such a way that the agreement between the slow component and the baseflow curve of the VIC-2L looked reasonable (b = 0.03 day^{-1}, $T_{1/2}$ = 40 days, where $T_{1/2}$ = (ln2)/k). For this parameter choice, Fig. 5 shows that the assumption of linearity of the horizontal routing model holds quite well. The iterative scheme had problems in April when snowmelt processes occurred and which cannot be captured by this procedure.

VIC-2L model calibration Table 1 shows the direction of influence of simulated

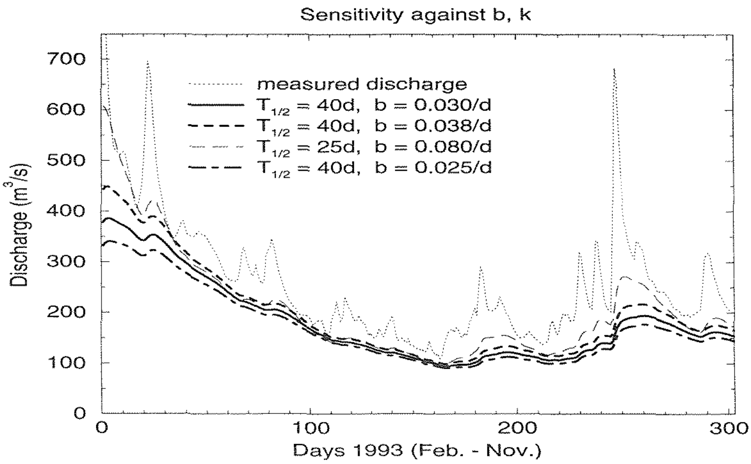

Fig. 3 Sensitivity of the slow flow separation against variations in the parameters b and $k = \ln 2/T_{1/2}$. A higher b puts more water into the slow component, a higher k leads to lower mean residence times in the slow flow storage.

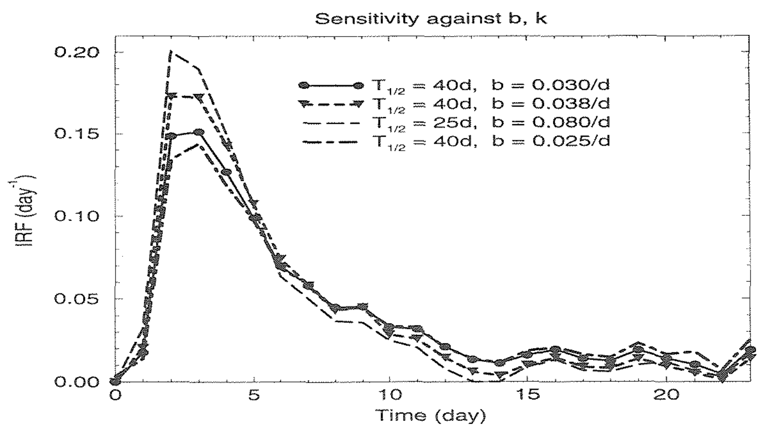

Fig. 4 Sensitivity of the impulse response function against variations in the parameters b and $k = \ln 2/T_{1/2}$.

streamflow to selected VIC-2L parameters. In the absence of constraints the simulated streamflow can vary over a wide range. However, because the routing model is derived independently from the VIC-2L model, those parameters that mainly influence the peak streamflow, such as β and the thickness of the upper soil layer, can be derived based on routing considerations and the requirement that the VIC-2L direct runoff should approximately equal the fast flow $Q^F(t)$ from the slow flow separation. This assumption means that all fast processes should be resolved within the upper layer.

Subsequently the lower layer maximum soil moisture and the factor d_1 are

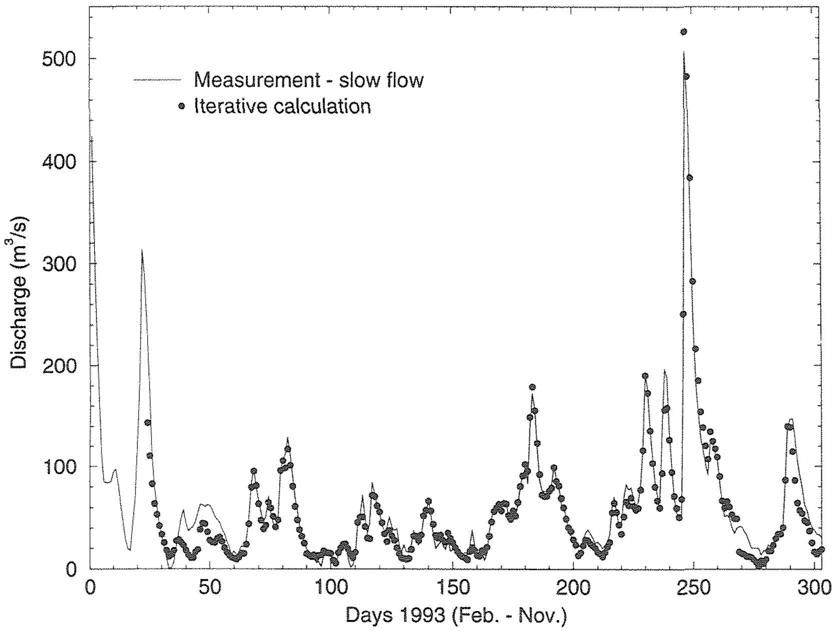

Fig. 5 Time series of the measured flow minus the separated slow flow *vs* the iteratively calculated fast flow from Intschede (Weser).

determined since these parameters mainly control the baseflow from April to October when the soil is not too wet. The parameter d_1 fixes the half-life decay of the water in the lower soil layer under dry conditions, which was determined by calibration to be between 200 and 300 days. The nonlinear part of the Arno baseflow recession curve was then calibrated. It is important mainly in wet conditions (winter/spring). The transition of the baseflow recession curve towards shorter residence times of about 8–15 days was found to start at approximately $W_S = 0.72$ of the maximum lower soil moisture content. The saturated hydraulic conductivity and the porosity were adjusted slightly during the calibration procedure within their physically meaningful ranges. In all cases, the initial soil moistures for both soil layers were set to 0.75 of the maximum soil moisture content. However, the effect of the initial moisture is removed within a few months, and therefore an initial warm-up period of one year was used prior to all comparisons.

Figures 6(a) and (b) compare mean daily modelled and observed discharge at the Intschede and Herrenhausen gauging stations for the calibration period. The Pearson product-moment correlation coefficient r is 0.966 for Intschede and 0.93 for Herrenhausen. Additionally the three day moving average difference between modelled and measured discharge is shown. The largest absolute errors of streamflow prediction occur in the snow accumulation and melt period around February. Whether this is due to inaccurate measurements of precipitation (underestimation of snow amount) or to model errors cannot be concluded.

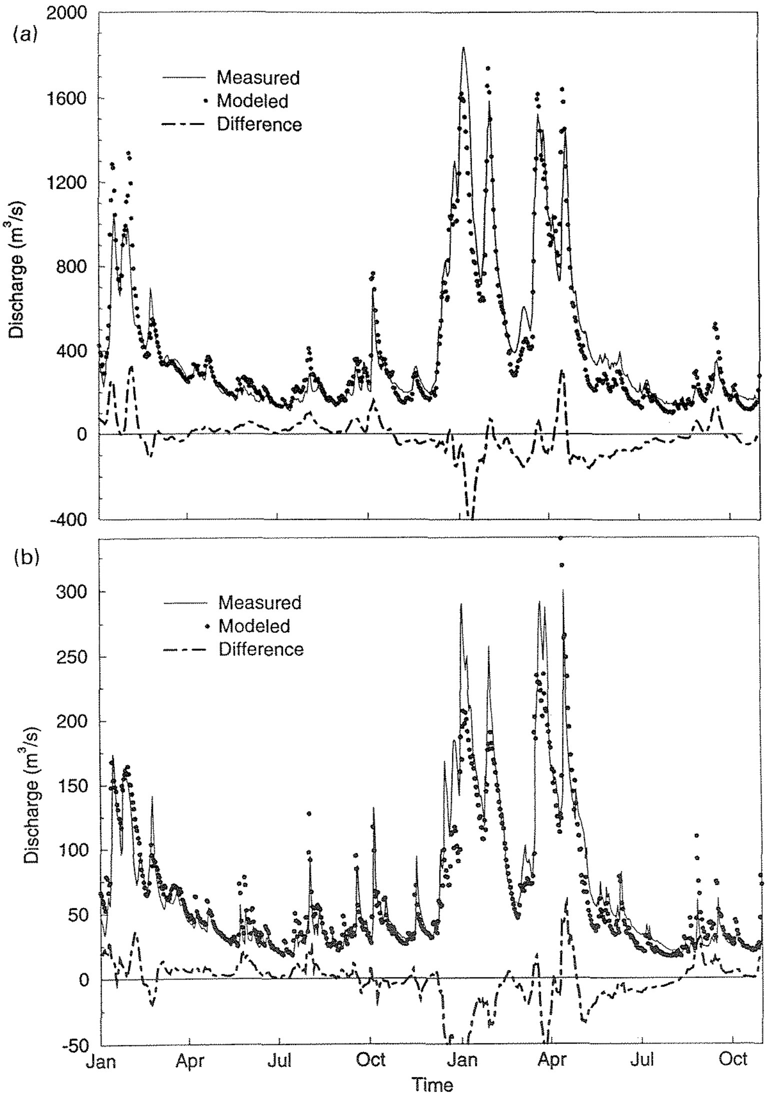

Fig. 6 Mean daily discharge of the Weser River (a) at its last non-tide influenced gauging station, Intschede, and (b) at the Herrenhausen gauging station, for the calibration period 1 January 1993–31 October 1994.

RESULTS

Model validation

Figures 7(a) and (b) show daily modelled and measured streamflow together with the three day moving average difference during the validation period. The Pearson product-moment correlation coefficient r (see Table 2) shows the generally excellent

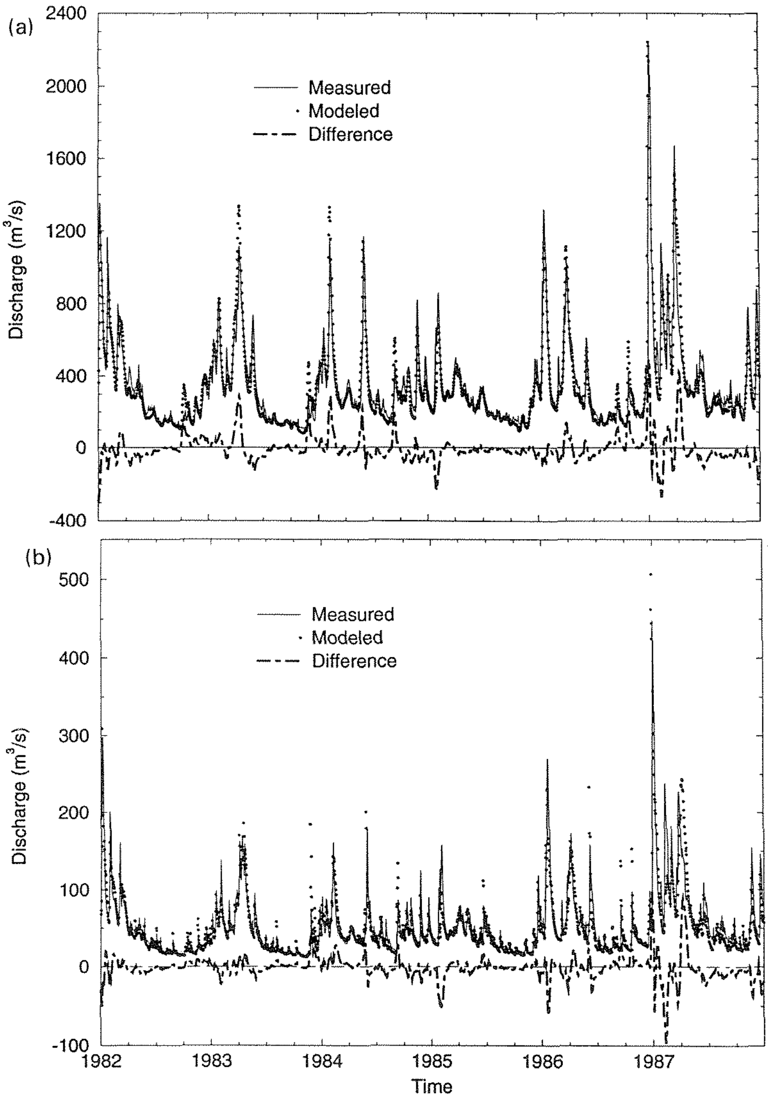

Fig. 7 Mean daily discharge of the Weser River (a) at its last non-tide influenced gauging station, Intschede, and (b) at the Herrenhausen gauging station, within the validation period 1 January 1982–31 December 1987.

agreement between mean daily modelled and observed data for all 10 gauging stations which were used. Only for the catchment above Affoldern (Eder River) is the agreement rather poor. This is due to the presence of a reservoir at Edertalsperre. To account for diversions into the basin, the measured discharge at Affoldern (Eder) was therefore added to the river routing scheme. On average, the amount is about 6.8% of the overall discharge of the Weser River at Intschede. Again, the largest differences between observed and modelled streamflow occur in winter and early spring when snowmelt processes are important. The monthly mean in Fig. 8 for the whole catchment at the Intschede gauging station and the complete time period from 1981 to 1994 also suggests that high streamflow volumes in late winter and early spring due to rain and snow tend to be overpredicted. One cause for that might also be the influences of dams and reservoirs in the Harz mountains (subcatchment Herrenhausen, Fig. 7(b)) and irrigation, which are not yet included in the model.

Table 2 Statistics for modelled and measured daily discharges in the test period (1981–1992) for the 10 gauging stations.

Station	Measured $\overline{Q}$ ($m^3 s^{-1}$)	Modelled $\overline{Q}$ ($m^3 s^{-1}$)	ΔQ (%)	r	Runoff ratio
1. Intschede	337.91	338.46	+0.16	0.933	0.369
2. Dörverden	212.57	211.34	-0.58	0.924	0.379
3. Rethem	117.28	120.53	+2.77	0.912	0.356
4. Hameln	172.00	167.70	-2.50	0.912	0.386
5. Herrenhausen	52.44	52.76	+0.60	0.830	0.400
6. Celle	28.56	28.34	-0.08	0.880	0.310
7. Hann. Münden	125.30	123.88	-1.13	0.889	0.391
8. Affoldern	22.83	19.98	-12.48	0.604	0.480
9. Rotenburg	24.86	24.40	-1.86	0.781	0.404
10. Heldra	40.45	41.92	+3.63	0.863	0.390

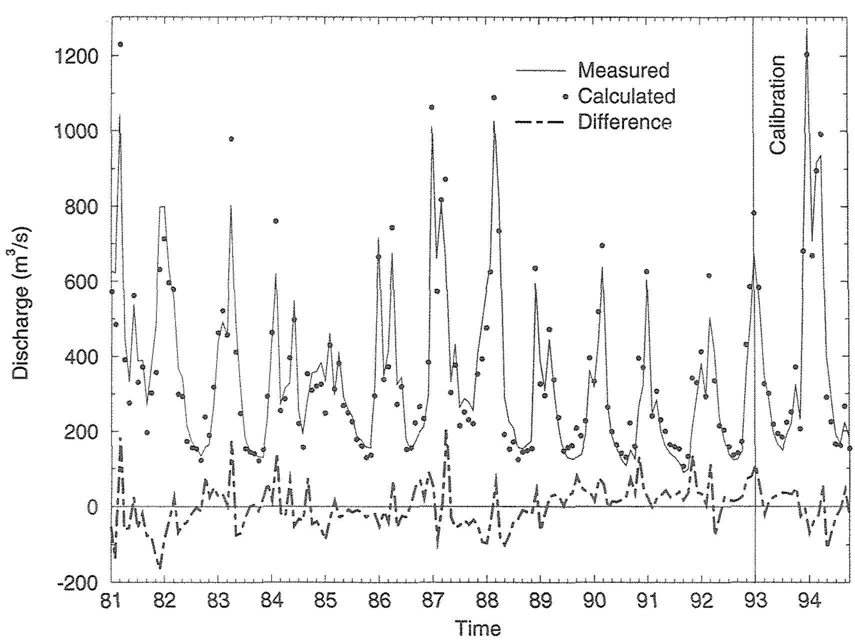

Fig. 8 Mean monthly discharge of the Weser River at the last non-tide influenced gauging station, Intschede, for the period January 1981–December 1993.

Inference about water balance

The monthly division of precipitation into the different components of the water balance (storage change, evapotranspiration and runoff) averaged for the whole catchment is shown in Fig. 9. From April to July the sum of evapotranspiration and runoff exceeds precipitation, while from October to January the opposite occurs. In August/September the soil has the lowest water content and in February/March the highest during the annual cycle, leading to the minimum and the maximum (together with January) runoff rates. Storage change during these months is close to zero. Figure 9 strongly emphasizes the need for an accurate representation of soil processes in atmospheric models. Evapotranspiration in June and July is strongly limited because of the drying soil; during this period evapotranspiration exceeds precipitation by only a small amount, while in May additional water from the soil is used. The uncertainty about the remaining terms in the water balance (storage change and evapotranspiration) can be accessed only with independent estimations of snow amount and soil moisture or evapotranspiration from an atmospheric budget analysis.

On an annual time scale the storage change of the soil is close to zero, so precipitation equals runoff plus evapotranspiration. As the VIC-2L model and the routing model are able to reproduce streamflow from daily to annual time scale, the mean annual spatial distribution of evapotranspiration, runoff and the runoff ratio in Fig.10 reflect the climatological mean for these quantities for the modelled years. Evapotransiration and runoff are clearly correlated with the precipitation distribution. However, runoff (ranging from 160 to 850 mm year^{-1}) shows a much stronger spatial

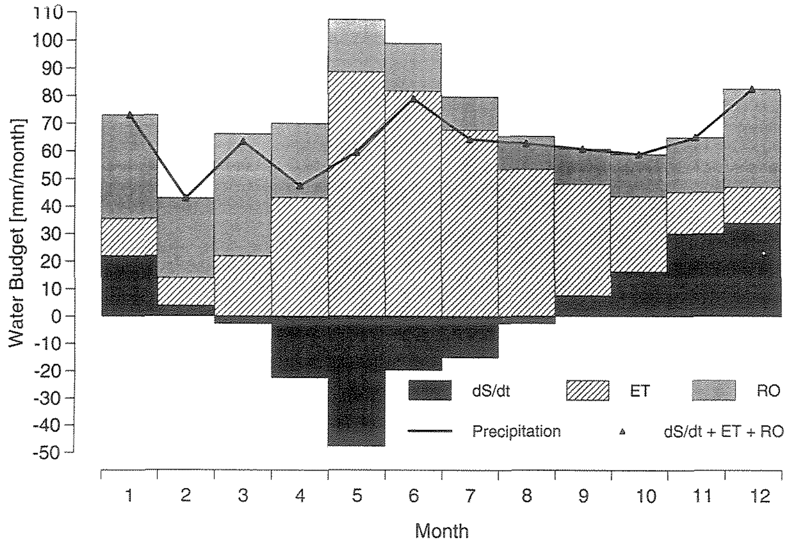

Fig. 9 Mean monthly rates for the whole Weser catchment of precipitation (*P*), evapotranspiration (*ET*), runoff (*RO*) and storage change (d*S*/d*t*, including the snow storage) for 1981–1993. The triangles indicate that the water balance equation is always valid.

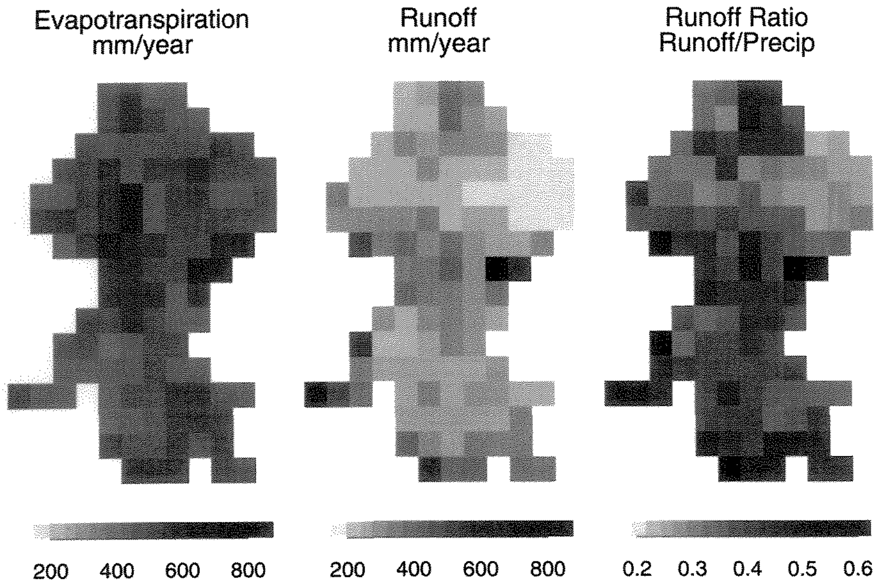

Fig. 10 Mean annual spatial distribution of evapotranspiration, runoff and runoff ratio (runoff divided by precipitation) in mm year^{-1} for all grid boxes within the Weser catchment for 1981–1993.

variability than evapotranspiration (ranging from 380 to 580 mm year^{-1}), leading to runoff ratios between 24% and 62%. This means that spatial variability in precipitation in the Weser catchment is reflected by a strong spatial variability in runoff production, while the evapotranspiration distribution is relatively smooth. The spatial variability of 200 mm year^{-1} in evapotranspiration between wet and dry grid cells mainly results from the precipitation difference in the summer months June and July, where precipitation and soil moisture are the limiting factors for evapotranspiration. On the other hand, spatial runoff variability of about 700 mm year^{-1} is mainly produced in December–March, dependent on how fast the soil storage is filled up again by precipitation. In wet parts of the catchment, the soil can already be filled up again in November/December, while in dry parts it can take up to three months longer, resulting in different runoff production rates.

Sensitivity to VIC-2L parameters and atmospheric data

The VIC-2L runoff production can vary considerably within the physical range of its parameters. Figure 11 shows the sensitivity of mean monthly discharge due to changes in soil thickness, baseflow parameters and infiltration parameters. For the thin soil the thickness of each layer was reduced to one third of the calibrated thickness, the infiltration parameter β increased by a factor of 1.5, and the residence time of the lower soil layer reduced to one third of its calibrated value. For the thick

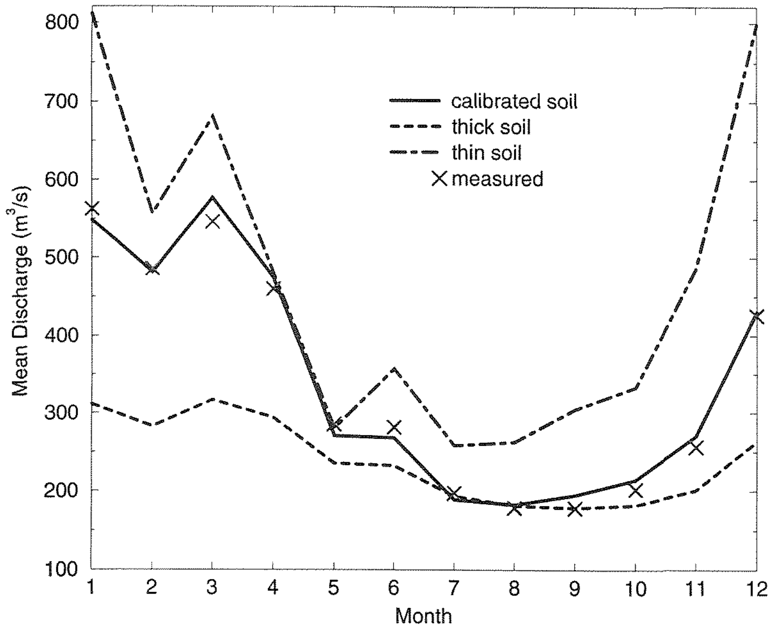

Fig. 11 Mean monthly discharge variability at Intschede (Weser) due to a change in soil parameters. Monthly means were averaged over 1981–1993.

soil the soil capacities were tripled, the infiltration parameter β halved, and the residence time of the lower soil layer tripled. The effect was to change the runoff ratio (see Table 2) to about 0.26 (thick soil) or 0.51 (thin soil). All vegetation parameters (LAI, root distribution, roughness length) were assumed to be fixed in these model runs, although a variation in these parameters also has a significant influence on the runoff production. From this simple sensitivity study it was concluded that applying this (and other) hydrological model(s) in ungauged catchments must presuppose some knowledge about the appropriate parameters. One possible approach is the regionalization method suggested by Abdulla (1995).

The sensitivity of the model output with unchanged VIC-2L parameters to systematic errors in the atmospheric data is shown in Fig. 12. An increase of 10% in precipitation increased the runoff ratio to 43.3%, while a decrease of 10% gave a runoff ratio of 31.4%. A systematic absolute error of 5% in relative humidity produced runoff ratios of 40.5% and 34%. The runoff ratio changed to 35.2% or 39.5% due to a systematic error of 20% in the short wave radiation. The relatively small influence of the systematic error in the short wave radiation comes from the fact that the vapour pressure deficit part of the Penman-Monteith formulation contributes the major part (about 80%) to the evapotranspiration. It was concluded that systematic errors in the atmospheric data can be compensated to a certain extent by the choice of soil parameters. This again stresses the desirability of an independent estimation of the major soil variables. In practice, however, for a macroscale application there appears to be little alternative to some calibration.

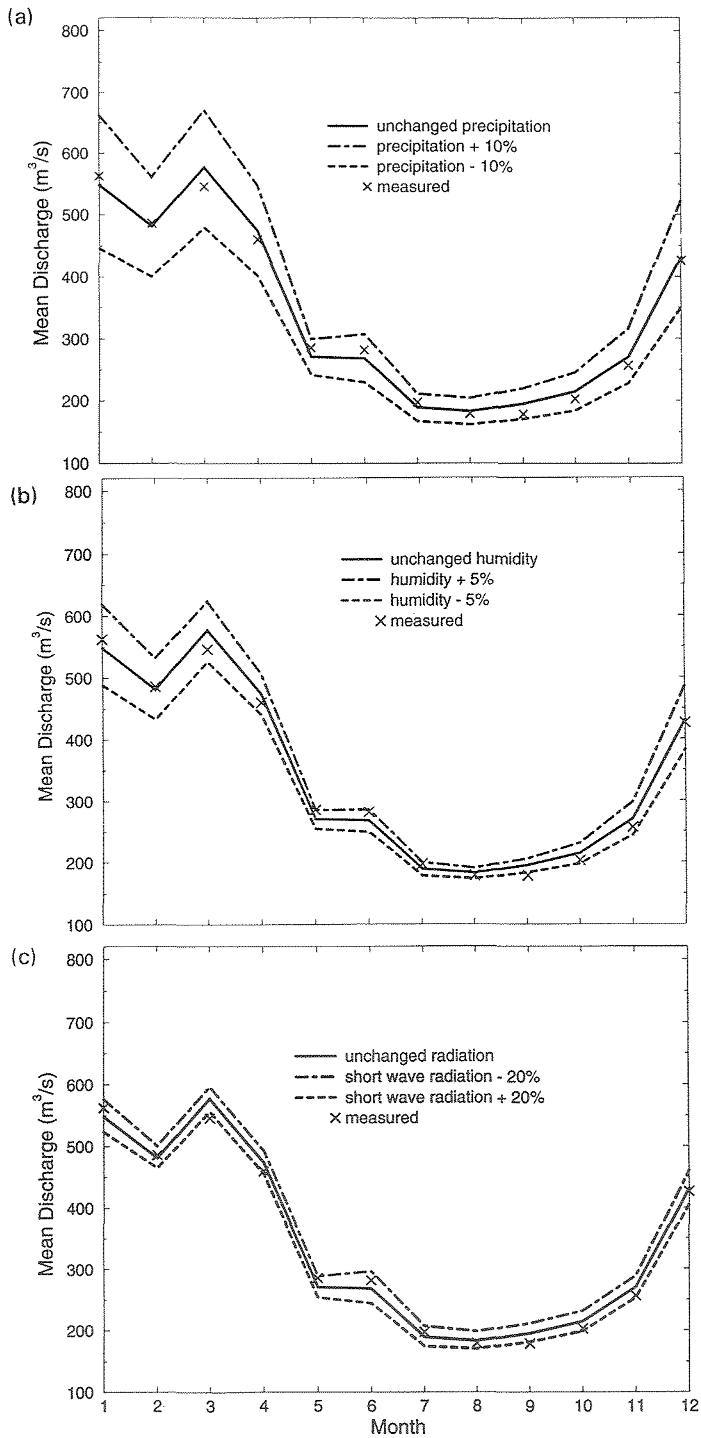

Fig. 12 Mean monthly discharge variability at Intschede (Weser) due to changes in (a) precipitation, (b) relative humidity, and (c) short wave radiation. Monthly means were averaged over 1981–1993.

CONCLUSIONS

The grid net version of the VIC-2L model together with a linear routing scheme was able to predict fairly accurately daily, monthly and annual streamflow in the Weser River and its tributaries. Arnell (1995) states that "the ability of a climate model to simulate the partitioning of energy at the land surface into latent and sensible heat fluxes can [therefore] be assessed using observed precipitation and runoff data". Therefore, because the amount of runoff can be predicted quite accurately on daily, monthly and annual time steps, it is argued that the estimation of evapotranspiration found in this study should also be accurate, although finally an independent estimation is needed. Further, use of the model within NWP models should offer a tool for a systematic study of the resulting performance of the atmospheric part, especially precipitation. A simple sensitivity analysis showed that the runoff volume produced by the VIC-2L model is strongly dependent on the soil thickness, the baseflow recession curve and the infiltration parameter. As there were no measurements available to gain information about soil thickness or soil moisture content, it is concluded that these measurements would be necessary to get a final verification for the modelled soil moisture. Even if all fluxes of an LSP scheme are predicted correctly the internal variables should be compared with measurements to check the reliability of the parameterizations.

As the VIC-2L model has also been validated using one dimensional point data (Liang, 1994; Liang et al., 1994), its parameterizations are a valuable tool to be included in point-scale models. The parameters can be chosen to represent the original point-scale model exactly or to represent a mesoscale (β) to macroscale hydrological model with the VIC-2L parameterizations. In short, they offer an opportunity to calibrate the model with only a few parameters, although conceptual problems with solving the Richards equation for vertical redistribution are apparent. The scaling characteristics of the VIC-2L model are currently under investigation. For spatially distributed applications the model is used so far for scales up to 2° by 2° (approx. 200 km × 200 km) grid cells.

The mean residence time in the lower soil layer is 280–450 days (with half-life decays of the storage between 200 and 300 days). Therefore it is not clear whether feedbacks on much longer time scales are adequately represented. A test of such a hypothesis can only be done with a much longer data set. In principle it would have been straightforward to include more storage terms with longer time constants in the modelling approach, but the necessary parameters are not obtainable with a 15-year data set.

Currently several refinements on the VIC-2L models are being pursued (Liang et al., 1997). One is the inclusion of a thin upper soil layer (VIC-3L) to represent better shorter time scales for the diurnal cycle (Liang et al., 1997). These parameterizations will be included in the research version REMO, a derivative of the operational weather forecast model of the German Weather Service (Majewski, 1991; DWD, 1995) within the BALTEX (1995) project.

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