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Application of a distributed physically-based hydrological model to a medium size catchment

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Abstract

Physically based distributed models are rarely calibrated and validated thoroughly because of lack of data. In practice, validation is limited to comparison of simulated and predicted discharges in a catchment, or of simulated and observed piezometric levels in some calibrated wells. Rarely, internal noncalibrated wells or discharge stations are included in model evaluation. In this study, the fully distributed physically based MIKE SHE model was applied to the 600-km² catchment of the Grote and the Kleine Gete, Belgium. Firstly, the MIKE SHE model was calibrated against both daily discharge measurements and observed water levels and then validated using a simple split-sample test. The observed discharges were simulated successfully in both the calibration and the validation period, while results for the piezometric levels differed considerably among the wells. In addition, a multi-site validation test for 2 internal discharge stations and 6 observation wells showed inferior results for the discharge stations and comparable results for the water table wells. As in the calibration and the split-sample test validation, water table fluctuations were predicted well in some wells, but with little agreement in others. This may be due to scale effects and to the poor quality of the data in certain areas of the catchment. Mainly, the lack of data made it difficult to simulate time series of internal catchment variables with acceptable accuracy so that even the calibrated and validated model could not provide reliable predictions of the water table over the entire catchment.

Keywords: integral hydrological modelling; distributed code; MIKE-SHE; model performance; model calibration; model validation

Introduction

In recent years, considerable progress has been made in the conceptual approach and mathematical description of water flow and transport processes (Abbaspour *et al.*, 1997). Several mathematical models of diverse complexity have been developed. The rapid increase in computer power and improved insight into hydrological processes has led to the development of models with increasingly complex equations based on physical processes, along with an increasing density of calculation nodes allowing representation of the spatial variability of the system. In most traditional hydrological studies, only one component of the hydrological cycle is studied in detail while the other components, if considered, are simplified or approximated. In physically-based distributed models, the various components of the hydrological cycle are integrated and transport processes are represented by the partial differential equations of conservation of mass, energy and motion. Since the first description of a physically-based spatially distributed model by Freeze and Harlan (1969), several codes have been

developed such as MIKE SHE (Refsgaard and Storm, 1995), IHDM (Beven *et al.*, 1987) and THALES (Grayson *et al.*, 1992). The distributed nature of these models implies that they are data intensive; the lack of data of good quality may limit the application of these models to large catchments.

The physical characteristics of hydrological catchments, such as soils, land use and topography, are inherently spatially variable. In distributed models, the spatial variability of the processes, inputs, boundary and initial conditions, and the hydrogeological characteristics, can be incorporated through a network of raster points. Integration in, or coupling with, a GIS makes it possible to subdivide the catchment into a large number of nodes and to consider spatial variability to the level of the available data. In principle, spatially distributed models can accept experimental data at each grid element or calculation node. In practice, because of heterogeneity of parameter values, differences between measurement and model grid scales, and experimental and financial constraints, specification of parameter values is difficult. These constraints also apply to

the validation of distributed model predictions by using measurements of internal system response. Conventional strategies for distributed model validation rely on comparing simulated model variables with observations for specific points representing either external boundaries or intermediate locations on the model grid. Traditional validation by comparing simulated with observed outflows at the basin outlet still remains the only attainable option in many practical cases. However, this method is inconsistent with spatially distributed modelling (Rosso, 1994).

The power of integrated, fully distributed physically-based models is at the same time their weakness, since the representation of spatial variability of hydrological characteristics requires a vast amount of data. The problem of overparameterisation is considerably heightened so that Beven (1989, 1996) considers these models as still being lumped conceptual models with an increased number of parameters that can be changed during the calibration process. Refsgaard and Storm (1996) emphasise that a rigorous parameterisation procedure is crucial to avoid methodological problems in the subsequent phases of model calibration and validation of these models: spatial patterns of the parameter values have to be specified so that the parameters reflect only significant and systematic spatial variations, inherent in the available data. Thus, the parameterisation process effectively reduces the number of free parameter coefficients to be adjusted in the subsequent calibration procedure (Refsgaard, 1997).

Model description

MIKE SHE is a deterministic, distributed physically-based

hydrological model, which integrates the entire land phase of the hydrological cycle. The model is based on the SHE (Système Hydrologique Européen) modelling concept (Abbott *et al.*, 1986). The SHE model has been developed over the years into what is now known as the MIKE SHE model (DHI, 1993), comprising interception, infiltration, evapotranspiration, subsurface flow in the unsaturated and saturated zones, surface flow and flow in channels and/or ditches (Fig. 1). The water movement module forms the main part of the model; it can be coupled with other modules such as the advection-dispersion or geo-chemical module.

Rainfall interception is modelled using a modified Rutter procedure (Kristensen and Jensen, 1975), which considers the maximum storage capacity of the vegetation canopy; the amount of water retained in the canopy is modelled as a function of the canopy storage capacity, which depends on the leaf area index. A combination of potential evapotranspiration estimates, land use, leaf area indices, transpiration from the root zone, and canopy storage determine the amount of actual evapotranspiration. The zero-inertia approximations to the St. Venant equation are solved numerically in two dimensions for overland flow and in one dimension for channel flow. The one-dimensional Richards' equation is applied to a representative grid square and solved numerically for pressure head variation, which in turn is converted to soil moisture content through the soil moisture retention curve. A percentage of water arriving at the soil surface is allowed to pass directly to the groundwater through a simple macro pore flow simulation. Saturated zone computations are performed using the 3-D Boussinesq equation for ground water flow. Coupling of the unsaturated and saturated zones is realised by an accounting

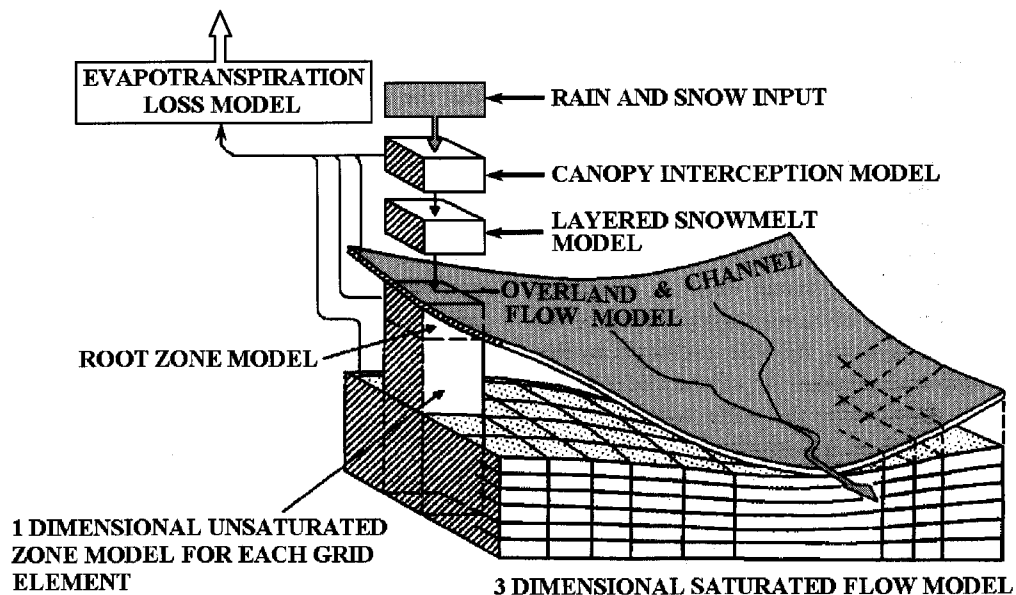


Fig. 1. Schematic presentation of the WM module of MIKE SHE (Refsgaard and Storm, 1995).

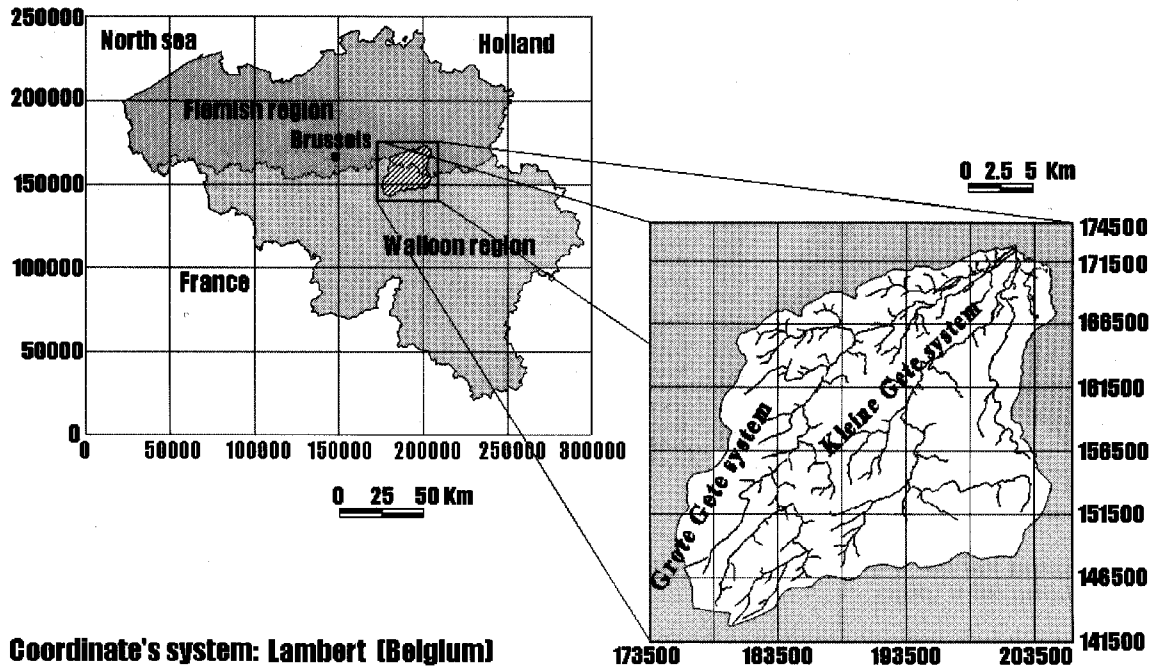


Fig. 2. Location of the catchment of the Kleine and the Grote Gete.

procedure, which takes into account water balance errors resulting from keeping the water table constant during the unsaturated zone computations.

The distributed nature of the model implies that spatial variations in catchment state variables and characteristics are integrated. The model provides a comprehensive pre- and post-processing package, which facilitates to a great extent the handling of the enormous input data requirements and the presentation of output results. The data format required by the MIKE SHE pre-processing programmes is generally not directly compatible with accepted data formats of the popular geographical information system (GIS) packages. Thus, in this study, the spatial data were stored in ARC/INFO and exported to data formats compatible with MIKE SHE through some routines written in PERL (Practical Extraction and Reporting Language).

Description of the study area

The 600-km² GETE catchment is located to the east of Brussels in the sandy loam region of Belgium and comprises two sub-basins, the Grote Gete (335 km²) and the Kleine Gete (265 km²), up to their confluence point in the community of Budingen. The upstream portions of the two subcatchments are located in the Walloon region of Belgium, with the downstream areas situated in Flanders (Fig. 2). There is a discrepancy in the availability, quality and volume of hydrological information between the two regions of the catchment. The catchment boundary defined

by the topographical divide does not necessarily correspond to the actual hydrological divide, which may extend beyond the topographical divide because of some of the aquifers of the hydrological system of the Gete catchment. However, no extra information, such as piezometric measurements, was available to assess the exact location of the hydrological divide, so that it had to be assumed to be coincident with the topographical divide of the catchment.

The altitude of the area varies from 26 m in the northern portion of the catchment, where the Kleine and Grote Gete merge, to 173 m in the southern parts (Fig. 3). Thiessen polygons based on 7 precipitation stations describe the spatially variable precipitation within the basin. The river network is assumed to run along the boundaries of the grid squares (Fig. 3). This implies that the resolution of the grid determines the detail of the river in the model set-up. The profile definition of the tributaries is based on interpolation and extrapolation of a few measured profiles. Strickler roughness coefficients for the rivers are based on profile descriptions and values from the literature. Land use in the area is mainly agricultural, both pasture and cultivated fields, with some local forested areas (Table 1). The spatial distribution of the ten classes corresponds to observations for the period May–August, 1989. Thus, the land use conditions for such periods were assumed to be steady throughout the whole modelling process. Leaf area index and root depth time series were generated from the literature (Allen *et al.*, 1998) on the basis of the crops described in the land use information. Roughness coefficients for overland flow are based on values taken from the literature for different land use types.

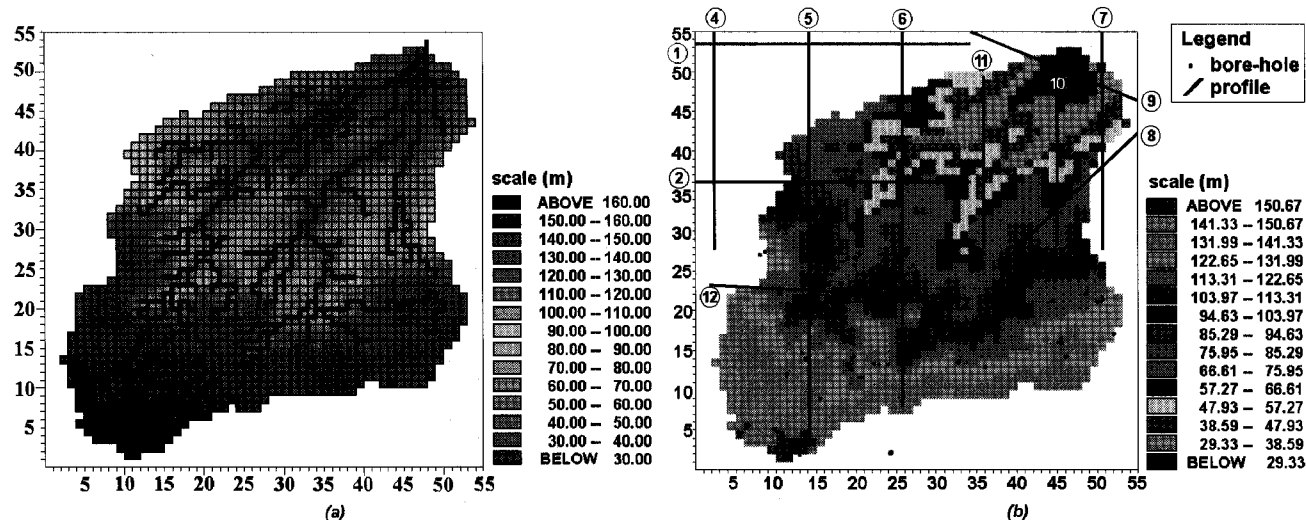


Fig. 3. (a) Topography and river network upstream of Budingen; (b) geology-monitoring features with the interpolated Quaternarian layer as background (resolution: 600 × 600 m).

Hortonian overland flow is based on the difference between the rain intensity and the soil infiltration capacity. The use of daily precipitation values does not allow consideration of single short duration storms so that information of the “real” rain intensity of a storm in a particular day is lost. Hourly precipitation values would be more appropriate for examining this difference between rain intensity and soil capacity. However, data were not available for applying shorter time steps. On the other hand, the material of the unsaturated zone is mainly loamy sand; Hence, Hortonian flow is small as the soil is not so impermeable. Consequently, subsurface flow is more important for transporting storm water to streams. In some local parts of the catchment, loamy clay material is present in the unsaturated zone and Hortonian flow can be the

dominant storm water transport process. There was, however, no information available about the location of these areas with less permeable soils. The one-dimensional Boussinesq equation for unsaturated flow is solved for individual soil columns that constitute the connection between the surface flow and the ground water component. This requires the soil-retention characteristic $[\Psi(\theta)]$, the relation between hydraulic conductivity and soil moisture content $[K(\theta)]$, including the residual $[\theta_{res}]$; the effective $[\theta_{eff}]$ and the saturated $[\theta_{sat}]$ soil moisture content for every soil layer of the different soil types within the simulation domain. The soil characteristics are retrieved from two databases, AARDEWERK-BIS (Van Orshoven *et al.*, 1991), which contains statistical information on soil types in Belgium and AARDEWERK-SISBIS (Huybrechts *et al.*, 1997), which documents individual soil profiles in Belgium. Nine different soil types were defined, according to the legend on the Belgian soil map, e.g. loamy soils (Aba, Ada and Adc), sand-loamy soils (Lca and Lda), clay soils (Eep and Uep) and soils with stony mixtures (Gbb). The van Genuchten equation (1981) and the Averjanov equation (1950) describe the soil retention characteristic and the relation between hydraulic conductivity and soil moisture content, respectively. The parameters for both equations are calculated with pedotransfer functions (Vereecken, 1988) based on information about the soil texture and organic matter content of the soil layers. The results are shown in Fig. 4 for the four soil layers that constitute the Aba soil type, which is the dominant soil type in the catchment.

The geology of the basin comprises nine geological units, namely the Quaternarian (Holocene and Pleistocene), the Borgloon formation (Tongeren Group), the formation of Sint-Huibrechts-Hern, the Brusselian formation (Zenne Group), the formation of Kortrijk (Ieper Group), the formation of Tienen and Landenian (Landen Group), the

Table 1. Land use in the Gete catchment

Land use description	LU class	%
Continuous urban fabric, industrial or commercial units, airports	1	13.0
Port and leisure facilities	2	0.3
Non-irrigated arable lands	3	67.7
Fruit trees and berry plantations	4	1.2
Pastures	5	1.9
Complex cultivation pattern	6	11.1
Land principally occupied by agriculture, with significant areas of natural vegetation	7	3.1
Broad-leaved forest, mixed forest	8	1.1
Water bodies	9	0.1
Paved areas	10	0.4

Legend: LU = Land use.

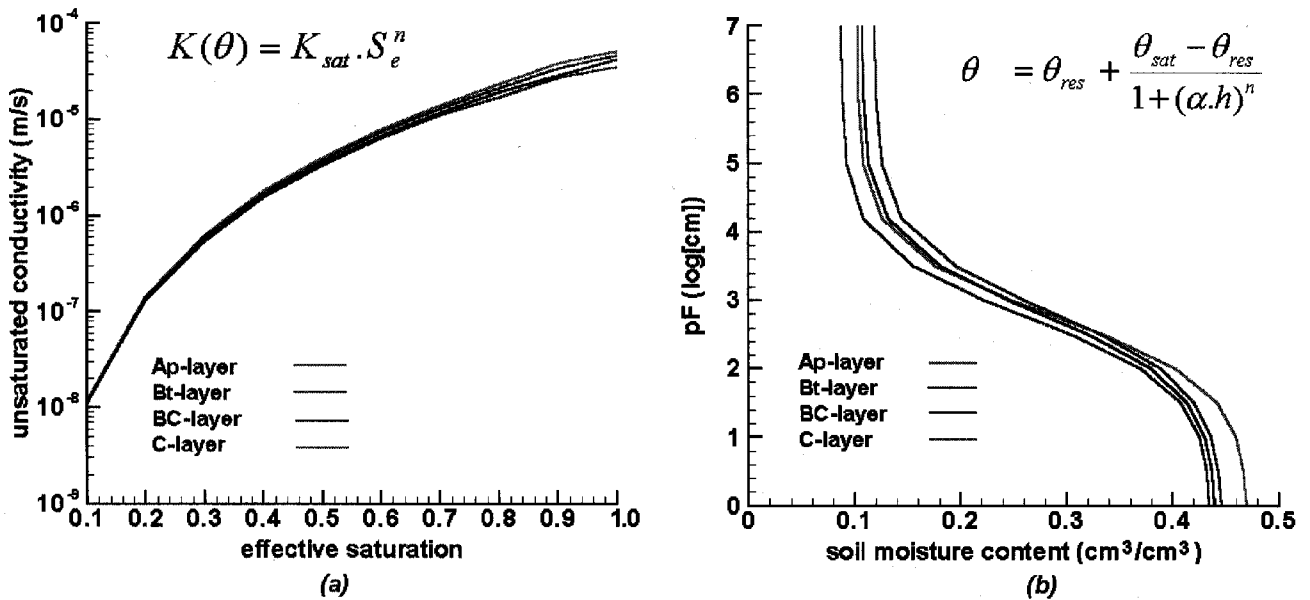


Fig. 4. (a) Relation between unsaturated hydraulic conductivity and effective saturation; (b) soil retention characteristic (pF-curve), for soil layers of the dominant soil type in the catchment.

formation of Heers, the Cretaceous chalk unit and the Paleozoic rocks, some of which occur only in isolated parts of the catchment. The complexity of the geological system indicates that a three-dimensional groundwater model is necessary to simulate flows and water tables. The model was constructed based on 12 geological profiles, digital information about the base of the upper layer, and 160 bore hole descriptions in the Walloon region of the catchment (see Fig. 3). Comparison of the same data from different origins shows such a wide disparity that the credibility of the geological data is questionable, which should be kept in mind when evaluating the results. Despite the extensive amount of available data, this lack of confidence in the quality of the geological data creates a potential source for poor simulation results. Initial values for hydraulic conductivity and specific yield used for the different geological layers are based on values from the literature for comparable formations. Actual evapotranspiration was

calculated using the Kristensen-Jensen model (Kristensen and Jensen, 1975) based on information on land use, crop characteristics and potential evapotranspiration. The empirical parameters of the model were assessed from the literature. Potential evapotranspiration (Fig. 5) was calculated with the Penman-Monteith method, which considers as a reference surface a hypothetical crop with an assumed height of 0.12 m having a surface resistance of 70 s m^{-1} and an albedo of 0.23 (Raes *et al.*, 1986; Allen *et al.*, 1998). The spatial distribution of the potential evapotranspiration was defined on the basis of the location of two meteorological stations.

Calibration procedure

The use of hydrological models for a broad range of applications has increased rapidly during the last few

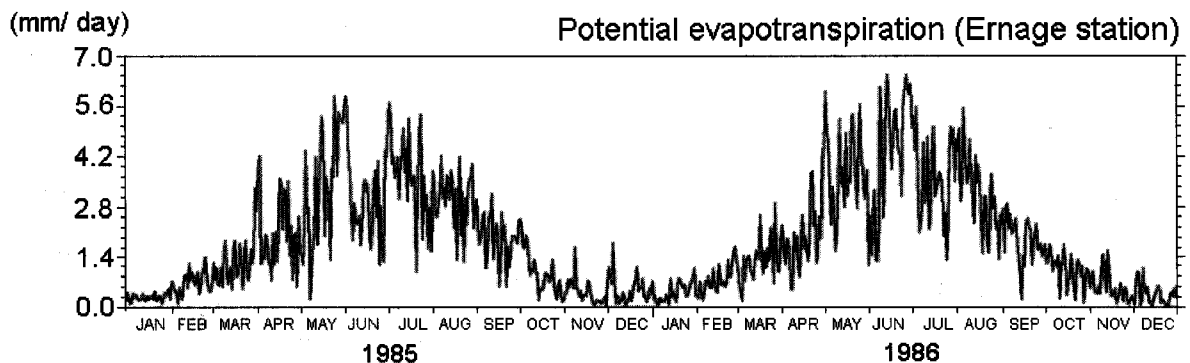


Fig. 5. Potential evapotranspiration (PET) for the meteorological station at Ernage.

decades. The ease with which they are applied often leads to a less critical evaluation of the basin being modelled. Often parameter values are selected without taking into account their physical nature and variability, considerations that are essential for physically-based distributed models such as MIKE SHE. In lumped conceptual models, the parameters do not have a physical meaning and the parameterisation is an optimisation process not restricted to physical boundaries. By definition, a fully distributed physically-based model contains only those parameters that can be assessed from field measurements. This implies that, in principle, calibration is unnecessary when sufficient data are available. However, these models are frequently applied to a scale different from that at which the model equations were deduced and validated. As a consequence of this discrepancy in scales, lack of data and measurement errors, these models usually have to be calibrated. The goal of the calibration process for physically based models is to find an optimal set of physically realistic parameter values that simulate the behaviour of the catchment as accurately as possible (Sooroshian and Gupta, 1995). To accomplish this, physically based constraints for the parameters' values have to be specified, based on the information available.

Figure 6 describes the concept of the calibration process for deterministic models. Once the parameters are defined, these models provide the same model output for a given set of input conditions. Both observations in the field and laboratory experiments on the physical system, as well as mathematical models, are subject to errors. A model is only an approximation to a complex reality and the equations are often applied at a scale different from that assumed in their development. Moreover, it is not possible to characterise all the spatial and temporal variability of the parameters since they represent spatially averaged values rather than point values obtained from field measurements. The relevance of the output of the hydrological model is highly dependent on the quality and quantity of input data. Optimising the parameters in the calibration process can improve the results.

It is advisable that the most uncertain or critical parameters, (i.e. the parameters with the most significant impact on the results) are defined at the beginning of the

calibration process. If the model fails to produce adequate results, an error analysis can possibly reveal the causes of the model failure. The following six sources of error can be distinguished (Spitz and Moreno, 1996; Refsgaard and Storm, 1996): (i) random and/or systematic errors in the input data (parameters and variables) and calibration data (variables); (ii) errors as a result of a non-optimal parameter set; (iii) mathematical errors in the model, in relation to the physical and mathematical basis of the computer code; (iv) conceptual errors in the model, e.g. incorrect interpretation of the processes, boundary conditions and dimensionality of the system; (v) numerical errors inherent in the solution algorithm, e.g. numerical dispersion as a result of discretisation; and (vi) interpretation errors of the predicted results, especially in models lacking post-processing tools and misconception of expected results.

During the calibration process of a hydrological model, only the uncertainty due to the second source of error can be minimised. Therefore, it is important to distinguish between the various sources of uncertainty so as not to change the values of the parameters outside physically acceptable boundaries when compensating for other error sources. Finding an optimal set of parameters, or solving the inverse problem, can be accomplished with many different techniques, e.g. manually through 'trial and error', automatically through numerical parameter optimisation, or a combination of both techniques. The 'trial and error' method is still the most common, especially for complex cases. This manual method requires hydrological insight and experience and, due to its subjective character, it is likely that different model users will find different parameter sets for the same model area subject to identical stresses. In the automatic optimisation process, a mathematical algorithm is used to optimise a given objective function. The purpose of this technique is to search through as many combinations and permutations of parameter values as possible to achieve the parameter set that is optimal in terms of satisfying the assumed criterion of accuracy (Refsgaard and Storm, 1996). To calibrate the parameters of a model by optimisation of an objective function means adopting some statistical technique (e.g., least squares, linear or non-linear regression, maximum likelihood, etc.) based on an analysis of residuals, often neglecting the physical characteristics of the model. As a result, the procedure is subject to the uncertainty inherent in every statistical analysis (Todini, 1988).

Model performance in the calibration and validation process can be evaluated both qualitatively, based on visual graphical techniques, and quantitatively, based on some statistical measures. In this study, both methods were combined but with emphasis on the statistical appreciation of the model performance. A first idea of the accuracy of the results was based on visual inspection of their graphical representation; then, statistical parameters for the simulations were calculated. The statistical criteria used in the analysis are the Relative Root Mean Square Error

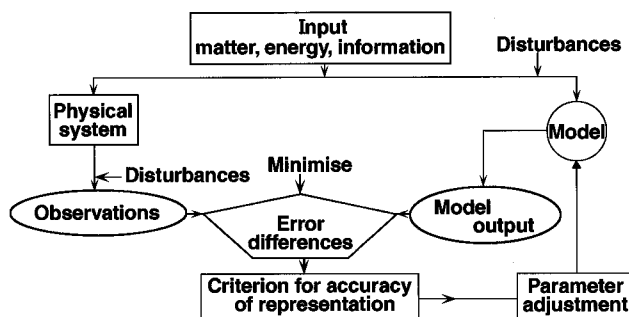


Fig. 6. Schematic representation of the calibration procedure for deterministic models, after Fleming (1975).

(RRMSE), the Coefficient of Determination (CD), the Coefficient of Efficiency (EF) and the Mean Absolute Error ($ABSERR$) and can be depicted by the following expressions (Loague and Green, 1991; Gupta *et al.*, 1998; Legates and McCabe, 1999).

$$RRMSE = \sqrt{\frac{\sum_{i=1}^n (P_i - O_i)^2}{n}} \cdot \frac{1}{\bar{O}} \quad 0 \leq RRMSE; \quad (1)$$

$$ABSERR = \frac{\sum_{i=1}^n |O_i - P_i|}{n} \quad 0 \leq ABSERR; \quad (2)$$

$$CD = \frac{\sum_{i=1}^n (O_i - \bar{O})^2}{\sum_{i=1}^n (P_i - \bar{O})^2} \quad 0 < CD \leq +\infty; \quad (3)$$

$$EF = \frac{\sum_{i=1}^n (O_i - \bar{O})^2 - \sum_{i=1}^n (P_i - O_i)^2}{\sum_{i=1}^n (O_i - \bar{O})^2} \quad -\infty < EF \leq 1; \quad (4)$$

where P_i is the i -th simulated value, O_i is the i -th observed value, \bar{O} is the average of the observed values, n is the number of observations in the considered period.

All the statistics except $ABSERR$ are dimensionless. The $RRMSE$ is a measure of the overall spread of residuals with respect to its mean value; it may be either positive or negative and its optimum value is zero. The $ABSERR$ is a measure of the average error of the modelled time series; it is always positive with zero as its optimum value. The CD index describes the ratio of the scatter of the simulated values and the observed values around the average of the observations. A CD value of one indicates that the simulated and observed values match perfectly; it is positive and its minimum value is zero. The coefficient of efficiency, defined by Nash and Sutcliffe (1970), describes the deviation from one of the ratio of the squared difference between the observed and simulated values, and the variance of the observations. It ranges from minus infinity to 1 with higher values indicating better agreement. Physically, EF is the ratio of the Mean Square Error (MSE) to the variance of the observed data, subtracted from unity. If the square of the differences between the model simulations and the observations is as large as the variability in the observed data, then $EF = 0$; if it exceeds it, then $E < 0$. Thus, a value of zero for the coefficient of efficiency indicates that the observed mean is as good a predictor as the model, while negative values indicate that the observed

mean is a better predictor than the model (Wilcox *et al.*, 1990). The mean absolute error is a measure of the average error of the modelled time series. As for the $RMSE$, lower values of the $ABSERR$ indicate a better performance of the model.

Calibration of the model

Before the actual calibration process begins, a few initial simulations are performed in which the model is constructed progressively to evaluate to what level of detail the model should reflect reality (e.g. number of geological layers to be included in the geological set-up). This process also includes the parameterisation of the model. To reduce the number of free parameters during the calibration process, the hydrological characteristics of the unsaturated zone are assumed to be fixed. This can be justified by the relatively large amount of soil data available in the two databases and by supposing that the calibrated pedotransfer functions represent the relation between the textural information and the hydrological characteristics of soils correctly in the unsaturated zone. Also, the parameters of the modified Rutter model to calculate actual evapotranspiration are assumed to be constant. Comparison of the results from the Rutter model with calculations of a modified Penman-Monteith method showed little difference, indicating that the values of the parameters of the Rutter model are acceptable.

As mentioned previously, the geology of the area consists of nine geological units. Incorporating a large number of geological units increases the complexity of the model, which leads to enormous computational requirements. Initial simulations demonstrate that the model can be simplified by transforming the nine geological units into five hydrogeological model units. This decreases the simulation time without influencing the results significantly. Selection of the geological units to be excluded from the model set-up was based on (i) the model performance statistics for the different model set-ups, (ii) the simulation running time, (iii) the extent of the geological layers with regard to the catchment area, (iv) the presence of pumping activity in the layers and (v) the hydro-geological properties of the different materials present in the layers. Finally, the geological model was simplified and only six layers were considered, namely the Quaternarian, the Brusselian, the Tienen and Landenian formation, the Heers formation, the Cretaceous chalk unit and the impermeable Paleozoic layer. During calibration, it is important to realise that the geological data are far from complete and reliable. In consequence, the authors did not strive above all to accomplish a close match of the piezometric levels because of the risk of obtaining parameter values that are physically unrealistic. All the aquifers were assumed to be unconfined after considering the composition of the layers and the pumping activity. Since the topographical divide was

assumed to coincide with the groundwater divide, all the aquifers have no-flow boundary conditions.

The choice of the resolution of the model grid depends on several factors, namely (a) the degree of heterogeneity of the hydrological parameters and the boundary conditions; (b) the extent of the flow domain defined by computational limitations; and (c) the predicted resolution necessary for the objectives of the modelling study. For the last case, there exists a scale difference between the observed point measurements and the simulated values, which are the average values of a raster cell. This difference in scale is inherent in the concept of numerical discretisation and should be considered when evaluating the results. Grid dimensions greatly influence the simulation time (Abbott *et al.*, 1986) and trade-offs between accuracy of the results and the calculation times must be evaluated. Some initial simulations were performed with resolutions of 2 km, 1 km and 750 m. The results presented in this paper belong to a model set-up with a resolution of 600 m, which results in a square matrix of 55 rows and columns. In the vertical direction, the division in soil layers was based on the soil profile descriptions. In the unsaturated zone, the resolution increased progressively from 5 cm at the top to 30 cm in the lower parts of the soil.

In hydrological models, one can distinguish between two types of time intervals, namely the stress period during which no change in boundary conditions occurs, and the time steps that define the length of the simulation steps. The time increment for the stress periods was taken as one day, in recognition of the lack of more precise meteorological issues and calibration data. The length of the time step is influenced by numerical stability, the variation in time of the boundary conditions, and the time related objectives of the project. In principle, smaller time steps lead to more accurate results. Too small time steps, however, lead to excessive computational times. Large time steps will result in an excessive number of iterations necessary for the solution to converge and possible numerical dispersion or instability.

In the MIKE SHE code, the simulation time steps for the different components have to be defined; they can be defined explicitly in the code but depend on some parameters. The simulation time step dt_{most} for the unsaturated zone (UZ), overland and channel flow (OC) and evapotranspiration (ET), is controlled by the parameters P_{max} and P_{alfa} , and the maximum time step for the unsaturated zone and overland and channel flow $T_{\text{maxUZ,OC}}$. When precipitation exceeds the maximum allowable amount of precipitation within a time step (P_{max}), the time step is reduced with the ratio $P_{\text{actual}}/P_{\text{max}}$; the time step will be progressively enlarged with the factor P_{alfa} until the corresponding maximum time step is attained, when precipitation decreases again. The simulation time step dt_{SZ} in the saturated zone (SZ) is mainly controlled by the maximum time step for the saturated zone T_{maxSZ} . The ratio $dt_{\text{SZ}}/dt_{\text{most}}$ is always kept equal to $T_{\text{maxSZ}}/T_{\text{maxUZ,OC}}$ (DHI,

1993). The values used in all the simulations are $P_{\text{max}} = 1 \text{ mm}$, $P_{\text{alfa}} = 0.05$, $T_{\text{maxUZ,OC}} = 4 \text{ h}$ and $T_{\text{maxSZ}} = 12 \text{ h}$.

The location of the observation wells and limnigraphic stations used in the calibration and validation of the model are shown in Fig. 7. To evaluate whether the model is capable of simulating internal state variables, it was decided not to include all the available data in the calibration process. Two internal discharge stations located on the two main river branches of the catchment, and 6 observation wells were not used in the calibration. The model performance at these locations was evaluated only for the validation period. From Fig. 7, it can be seen that the observation wells are located mainly in the southwestern and northwestern parts of the catchment, and that no discharge station exists in the Walloon part of the catchment. For the calibration on discharge measurements, only the station on the Gete is used. Since it is located where the two rivers merge, it represents the total discharge of the catchment. The calibration period was chosen based on the availability and continuity of data. It includes the period from August 1, 1984 until December 31, 1986. The first six months were used as an initialisation period. Statistical analysis was limited to 1985–1986.

The importance of initial conditions was first emphasised by Stephenson and Freeze (1974). Due to the lack of piezometric levels for interpolating the initial position of the water table within the catchment, the initial conditions were determined through an iterative procedure. From the available piezometric measurements and discharge data, it was concluded that the water balance of the catchment at the beginning of the calibration period was comparable with the situation in August 1980. The deduction of the initial conditions for the calibration period, which is schematised in Fig. 8, was established as follows (Refsgaard, 1997): (1) the model was run for the period August 1980–August 1984 with estimated initial conditions based on the few measurements; (2) the model was run again for the same period using the groundwater results of August 1984 of the previous simulation as initial conditions; (3) if the groundwater results obtained for August 1984 in the two simulations were not comparable, a new simulation was performed based on initial conditions extracted from the second run; etc. With this procedure, convergence to comparable results is not guaranteed but it is expected under the assumption, implicit in every modelling activity, that the model is a good representation of the real system. This procedure was repeated after every significant change of the parameters as the groundwater tables depend on the various model parameters and as consistency between model parameters and initial conditions is important. The use of model generated piezometric levels ensures that the initial conditions are consistent with the model parameters and input. When estimated potential levels, based on observations, are used for initial conditions, the model response will in the first time steps not only be a reflection of the stress factors but it will also reflect the adjustment of the potential

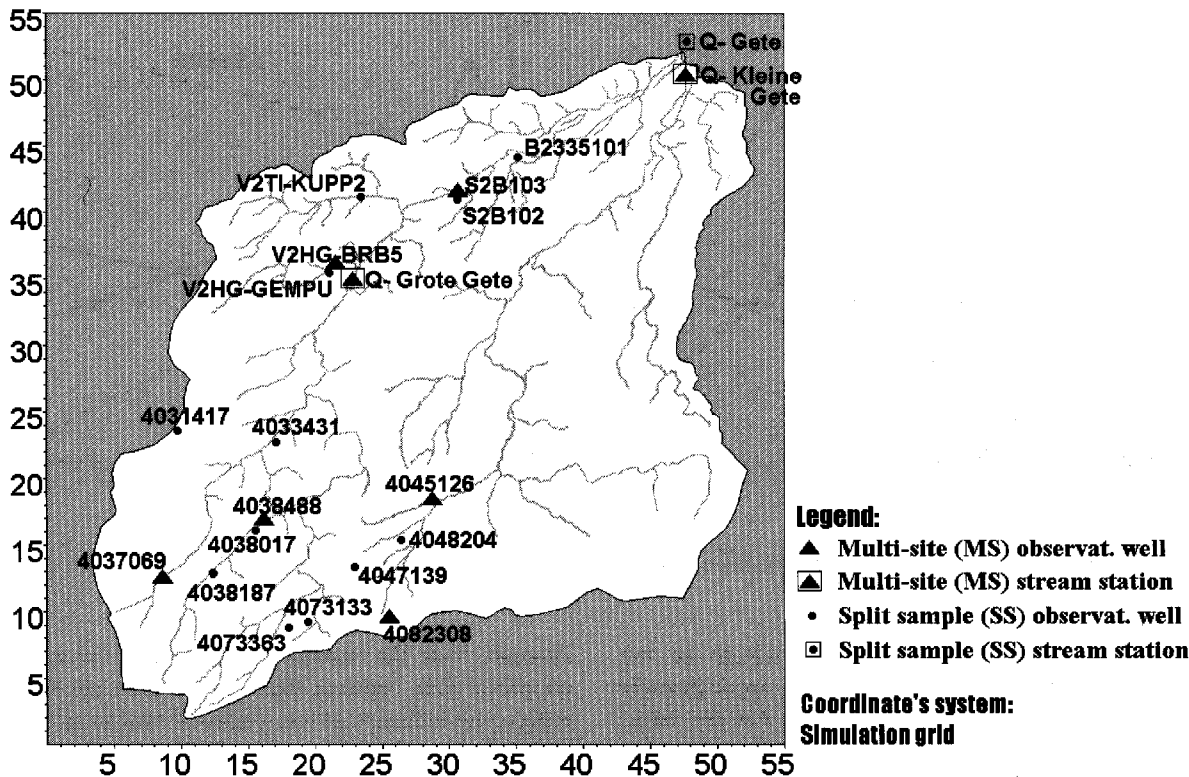


Fig. 7. Location of the discharge stations and the observation wells used in the calibration and validation.

levels as a result of the lack of agreement between the hydrological model input and parameters, and the initial conditions based on observations (Franke *et al.*, 1987).

The calibration was performed using the ‘trial and error’ procedure, in which the influence of the various model

parameters was examined step by step. Prior to calibration as a part of the model parameterisation, a range of physically acceptable values was defined for every parameter based on available data and on the literature (see Table 2). The width of every interval reflects the initial uncertainty of the respective model parameter. In the first phase, the model was calibrated against the overall discharge of the catchment. To account for the small canals and ditches present on a scale smaller than 600 m, drains were specified in the model set-up. Both the *drainage depth* and the *reciprocal time constant*, denominated as *time constant* in the MIKE SHE code, were calibrated. The effect of the drainage depth depends on the average position of the phreatic surface. Drainage starts when the water table rises above the elevation of the drains and is proportional to the difference in level between the water table and the drains. The reciprocal time constant or *drainage coefficient* determines the velocity of the drainage and mostly influences the peak of the hydrograph. The smaller the reciprocal time constant, the smaller are the peaks of the hydrograph. The drainage depth has more influence on the recession of the hydrograph. Refsgaard (1997), after calibration of the medium size Karupt catchment with geology consisting of permeable sand and gravel with few moraine clay lenses, obtained a value of 33 days for the drainage coefficient. For the Gete catchment having an upper geological loamy clay layer, a higher drainage coefficient may be expected. A typical feasible interval for the reciprocal time constant is

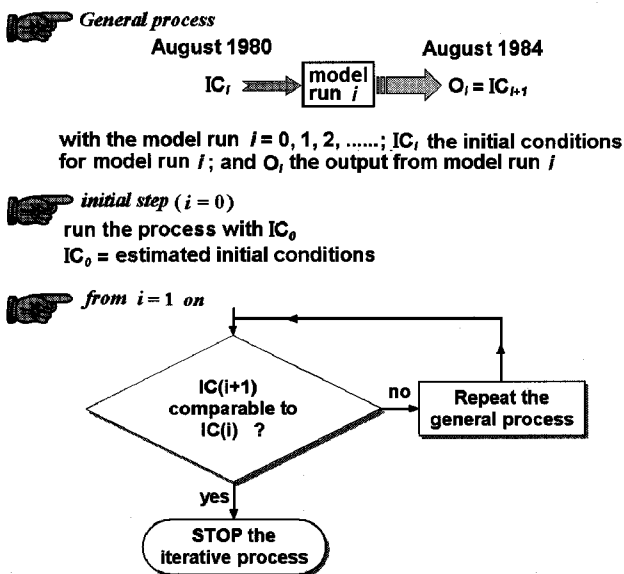


Fig. 8. Process followed for determining the groundwater initial conditions.

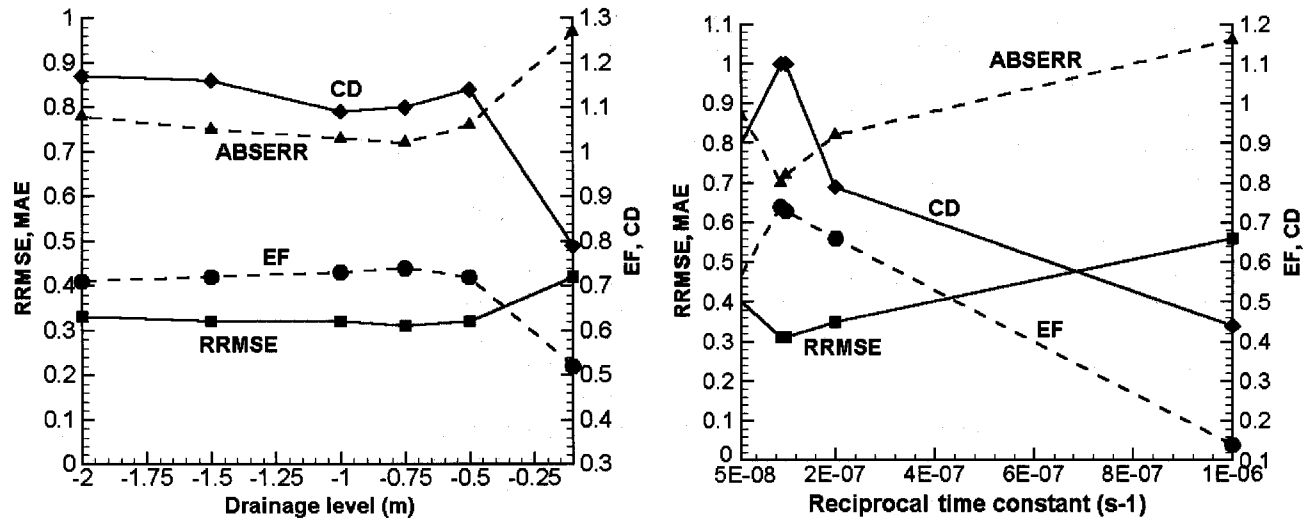


Fig. 9. Statistical parameters for different tested drainage levels and drainage reciprocal time constants.

$[1 \times 10^{-7}, 1 \times 10^{-6}] \text{ s}^{-1}$ (DHI, 1993), which for the drainage coefficient or time constant is approximately equivalent to an interval of [120, 10] days. On this basis, as depicted in Table 2, the feasible interval for calibration was fixed as $[5 \times 10^{-8}, 1 \times 10^{-6}] \text{ s}^{-1}$ for the reciprocal time constant with its corresponding equivalent interval for the drainage coefficient being [230, 10] days. Results from this calibration stage are presented in Fig. 9. The graphs

illustrate that the statistical criteria vary more for the reciprocal time constant than for the drainage level. This reflects the influence of the reciprocal time constant on the hydrograph peaks, which often leads to a large difference between the observed and simulated values. The drainage depth has a less pronounced influence on the statistical criteria. The results also indicate that the optimal values for different objective functions do not correspond necessarily

Table 2. Feasible parameter intervals (initial uncertainty) considered during calibration of the Gete model

Model parameters	Geological layers	Lower boundary	Upper boundary
Drainage level (m)		-2.00	-0.10
Reciprocal time constant (s^{-1})		5×10^{-8}	1×10^{-6}
K_h (m s^{-1})	Quaternarian	1×10^{-8}	4×10^{-5}
	Brusselian	7×10^{-5}	2×10^{-3}
	Landenian	1×10^{-5}	5×10^{-4}
	Heers	5×10^{-6}	5×10^{-5}
	Cretaceous	1×10^{-6}	1×10^{-5}
K_{sat} (m s^{-1})	Quaternarian	1×10^{-7}	1×10^{-6}
	Brusselian	7×10^{-6}	7×10^{-5}
	Landenian	5×10^{-6}	5×10^{-5}
	Heers	1×10^{-7}	5×10^{-7}
	Cretaceous	1×10^{-8}	1×10^{-7}
K_v (m s^{-1})	Quaternarian	0.01	0.30
	Brusselian	0.01	0.44
	Landenian	0.01	0.46
	Heers	0.02	0.40
	Cretaceous	0.01	0.46
S_v (-)	Quaternarian	0.01	0.30
	Brusselian	0.01	0.44
	Landenian	0.01	0.46
	Heers	0.02	0.40
	Cretaceous	0.01	0.46

Legend: K_{sat} = saturated hydraulic conductivity, K_h = horizontal saturated hydraulic conductivity; K_v = vertical saturated hydraulic conductivity, S_v = specific yield.

to the same parameter value; therefore, it is advisable not to limit the calibration to the optimisation of one single objective function, but to use multiple criteria in conjunction with a qualitative evaluation. Research in the last two decades has yet to provide a goal function that is more appropriate for calibration of a hydrological model. (Chapman, 1970; Diskin and Simon, 1977; Sorooshian *et al.*, 1983; Yan and Haan, 1991; Yapo *et al.*, 1997; Gupta *et al.*, 1998). Based on the statistical criteria and a qualitative evaluation, it was decided to use a drainage level of -0.75 m and a reciprocal time constant of $9 \times 10^{-8} \text{ s}^{-1}$, which is equivalent to a drainage coefficient of about 125 days. This value indicates clearly that the velocity of the drainage is low, which in general leads to an underestimation of the peak flows.

Next, the influence of some hydro-geological parameters on the simulated discharge was examined. The analysis showed that the hydraulic conductivity of the first, Quaternarian loamy layer, influenced the simulated discharge significantly. Furthermore, the vertical permeability of the top layer seemed to be far more important than its horizontal permeability; the thick loam layer is considerably less permeable than the deeper sandy layers and the vertical conductivity will determine the velocity of percolation through the Quaternarian layer to the deeper aquifers. A low vertical conductivity leads to a diminished vertical flow to deeper zones and to higher surface runoff and drainage flow, which results in higher simulated peak flows. Higher vertical permeability results in lower and flatter peaks. The horizontal conductivity of the deeper hydro-geological units influences mainly the base flow rather than the surface runoff or peaks.

When the overall discharge was reasonably well simulated, the hydro-geological parameters were optimised to get a good agreement between the simulated and observed piezometric levels in the observation wells considered in the

calibration process. Parameters were allowed to vary spatially within the same unit but too large a difference in values within one hydrogeological unit was avoided. During this optimisation process also, the influence of the local changes of the hydrogeological parameters on the discharge was evaluated in order not to get inferior results for the discharge. As described by Gupta *et al.* (1998), optimisation of one parameter will often lead to a diminution of the model performance that was achieved for other parameters in earlier calibration stages. Table 2 shows physically acceptable intervals for the parameters of the different hydrogeological units that were tuned during the calibration process. These values represent effective parameters which partly explains the sometimes broad ranges for parameter values. Information about the yield coefficient intervals for the layers is also included in Table 2.

A qualitative picture of the model performance for the discharge in the Gete station in Budingen is given in Fig. 10. From the hydrographs, it is clear that the base flow is overestimated and the higher peaks underestimated. The cumulative discharge curve shows that the total discharge from the catchment is estimated relatively accurately in the calibration period. Inspection of the hydrographs and of Table 3 indicates that the model can simulate the integrated discharge of the stream basin for the calibration period reasonably well. A model efficiency of 0.69 is similar to the results of analogue studies (Refsgaard and Knudsen, 1996; Refsgaard, 1997), while the RRMSE is sufficiently small. However, the CD index is relatively high because the peak discharges are underestimated. This is expressed strongly in the value of this objective function (Eqn. 3) because the overall spread of observed discharges around their mean value (standard deviation of the historical time series) is much higher than the overall spread of simulated discharges around the mean observed value.

In Fig. 11, the simulated and observed piezometric levels

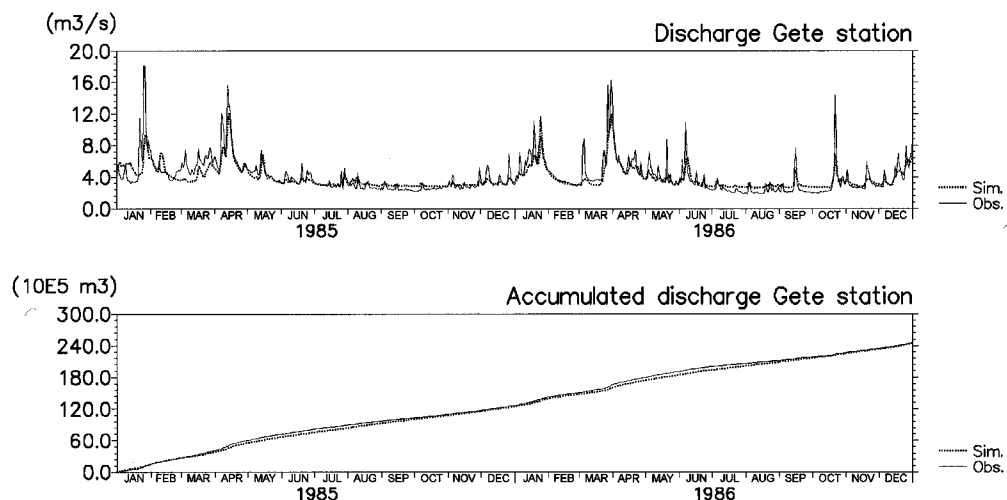


Fig. 10. Simulated and observed discharge versus time, and cumulative simulated and cumulative observed discharge versus time for the limnigraphic station on the Gete in Budingen (calibration period).

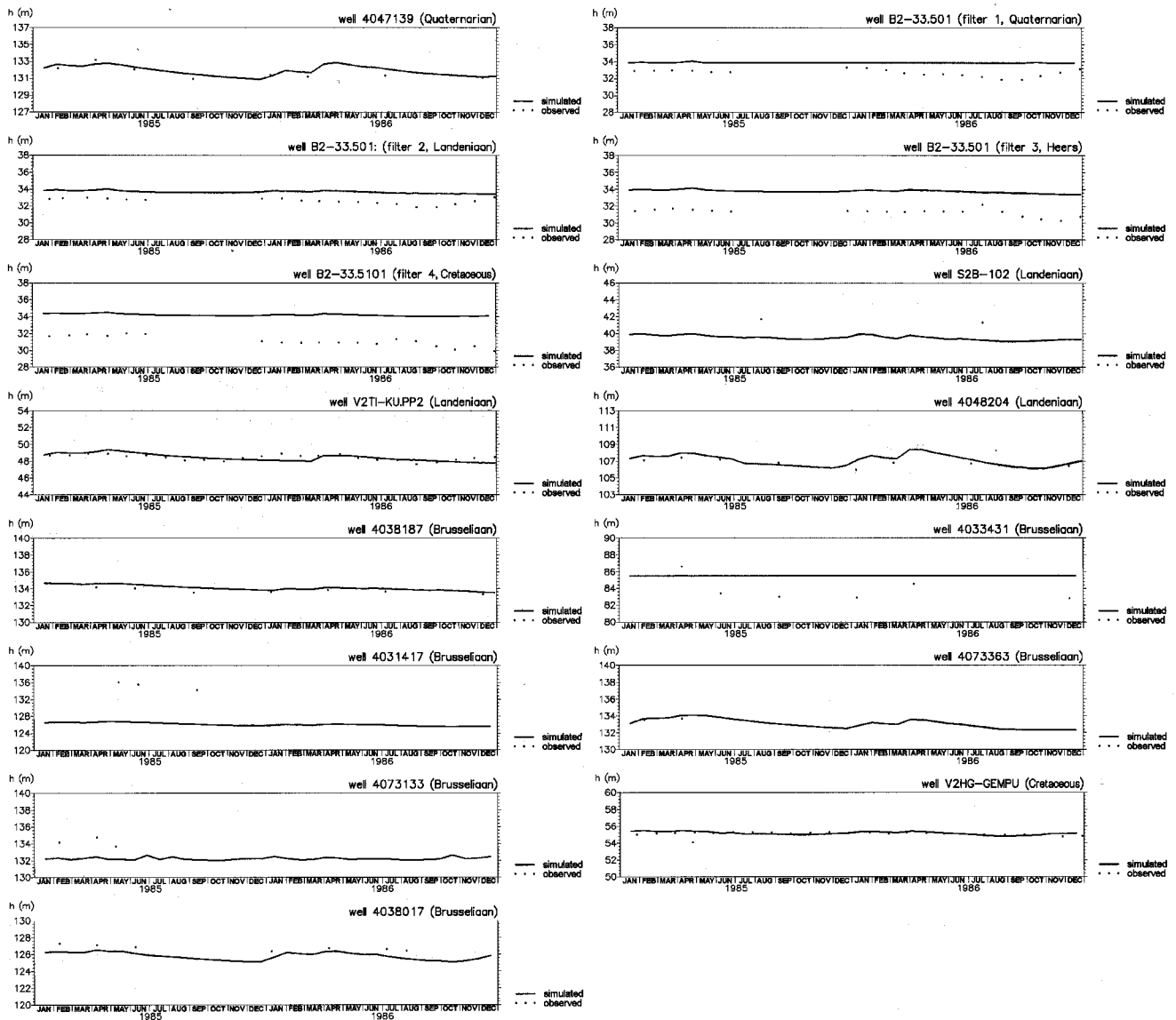


Fig. 11. Simulated and observed piezometric levels in the twelve observation wells used in the model calibration.

are shown for the twelve observation wells used in the calibration process of the model; the results differ a lot among the different wells. Observation well B2-33.501 has 4 filters in different hydro-geological layers. In the upper two filters, the observed potential varies around a value of 33 m, which is about one metre higher than the average observations in the two lower filters of the well. The simulated potentials in the 4 filters show little difference and a low variation in time. This implies that the model represents reality poorly in this region of the catchment. Other unacceptable results can be found in wells 4033431 and 4031417. In the former well, although the average position of the water table is relatively well estimated, the simulated piezometric level shows little variation in time during the calibration period. Varying the parameters does not change this, which means that the model setup is inconsistent with

the actual geohydrological situation. The latter well varies in time, but the levels are, on average, underestimated by 8 m; this might reflect the influence of the adopted boundary condition as this well is located near the boundary of the catchment (Fig. 7). For wells S2B-102 and 4073133, the observed piezometric levels are underestimated by almost two metres. The small numbers of observations in time for these wells make it difficult to assess to what extent the simulated fluctuations follow the observed fluctuations. The other seven wells (4047139, V2TI-KU.PP2, 4048204, 4038187, 4038017, 4073363, V2HG-GEMPU) show a good agreement for the variation of the water table in time. Due to the large discrepancy in the number of observations and the quality of the results for the different wells, no values for the statistical criteria were calculated to quantify the model performance in simulating the water table position.

Table 3. Calculated values of the objective functions for Q-Gete (calibration and validation), Q-Grote Gete and Q-kleine Gete (both only for internal validation)

	EF	RRMSE	CD	ABSERR (m ³ s ⁻¹)
SS-calibration Gete	0.69	0.30	2.20	0.69
SS-validation Gete	0.76	0.24	1.23	0.69
MS-validation Grote Gete	0.11	0.56	0.52	0.55
MS-validation Kleine Gete	0.46	0.26	0.74	0.40

Legend: SS = Split sample test, MS = Multi-site test.

The results of the calibration show that the calibrated model is an acceptable simulator for the integrated discharge of the catchment under study, as well as for some observation wells within the catchment. With additional geological information in certain areas and a more correct hydrological delineation of the basin, the performance of the model might be improved in other observation wells. More spatially distributed observations would enable the model to be optimised in other regions of the catchment. Because of the lack of observations in the eastern parts of the catchment, it is not appropriate to make distributed predictions of the water table position in these parts of the stream basin.

Validation of the model

To validate the model, a simple split-sample procedure was first performed, through which the performance of the model for a period different from the calibration period was inspected. This procedure is generally applied to lumped models where validation is limited to verifying the model performance in relation to the discharge. Other validation

methods regularly applied are the differential split-sample, the simple proxy-catchment and the differential proxy-catchment tests, which are described in detail in Klemeš (1986). Based on data availability, the period 1987–1988 was chosen for validation of the model.

Figure 12 shows the results for the Gete discharge station for this period. Both the qualitative picture and the quantitative measures (Table 3) show that the overall discharge is simulated better than in the calibration period. The lower flows, especially, are estimated more accurately. The cumulative curve shows that the total discharge during these two years is also estimated accurately and the decrease in the coefficient of determination is remarkable.

Figure 13 shows the results and observations for the validation period for the twelve calibrated wells. For well B2-33.501, only the results for the second and the third filter are given since they differ little from filters one and four, respectively. The observations of this well seem unreliable because of discontinuities in the time series, which may well be caused by different people employing an inconsistent datum when measuring the water levels. Results in the other wells are comparable with the results of the calibration period. Wells with poor results for the

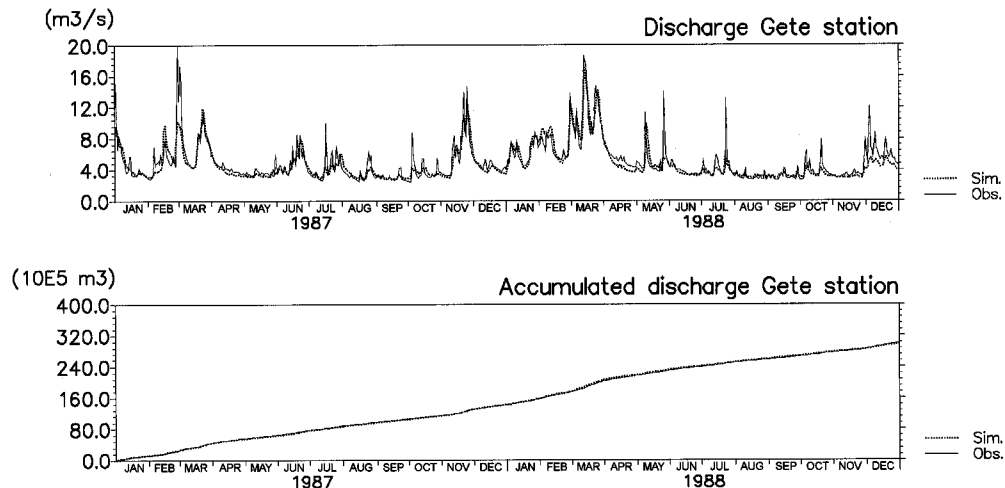


Fig. 12. Simulated and observed discharge versus time, and cumulative simulated and cumulative observed discharge versus time for the limnigraphic station on the Gete in Budingen (validation period).

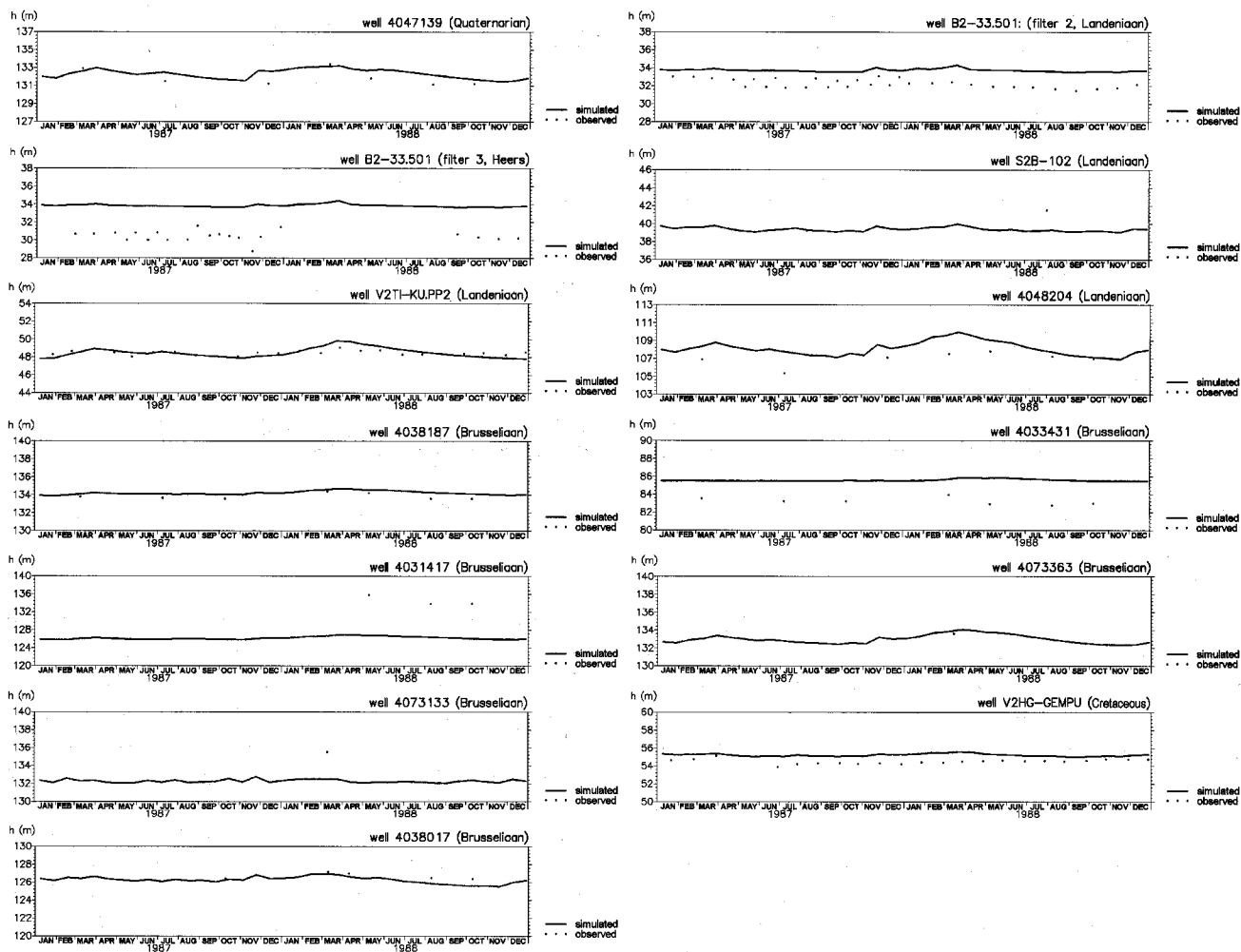


Fig. 13. Simulated and observed piezometric levels in the twelve observation wells used in the model calibration (validation period).

calibration period show the same shortcomings during the validation period. In contrast to the results of the calibration period, the simulated water table shows some variation at the end of the validation period, but not in agreement with the observed fluctuations. Of the wells that showed good results in the calibration process, only well 4048204 and well V2HG-GEMPU show little deterioration in the results. The position of the water table is now somewhat over-estimated, especially during the first 18 months of the validation period, but the simulated fluctuations stay in reasonable agreement with observations.

The results of the simple split sample test show that the model can be assumed as validated for simulating the rainfall-runoff relation for the whole catchment and for predicting to some degree the water table in a limited number of wells. In some other observation wells, the model is unable to predict, adequately, the position and fluctuation of the water table. It must be stressed that the predictive use of a model is valid only as long as no significant non-stationarities, such as climate changes or changes in land use, occur in the future.

In most cases, model users assume that, after such a validation test, the distributed model can be used to simulate internal flows and spatially distributed water table positions. These suppositions were made by Refsgaard *et al.* (1992) and Jain *et al.* (1992) based on their results of the application of the MIKE SHE model to large catchments in India. To verify whether the model is also capable of simulating internal state variables, a multi-site validation test was applied. Hereby, it was checked to what degree the model could estimate observed variables in internal points of the catchment, i.e. in observation points not used in the calibration. The data used for this validation test are the following (Fig. 7):

- ✓ discharge data for the station on the Kleine Gete in Budingen;
- ✓ discharge data for the station on the Grote Gete in Hoegaarden; and
- ✓ 6 observation wells.

In Fig. 14, the results of the validation period for the discharge stations at Hoegaarden (Grote Gete) and Budin-

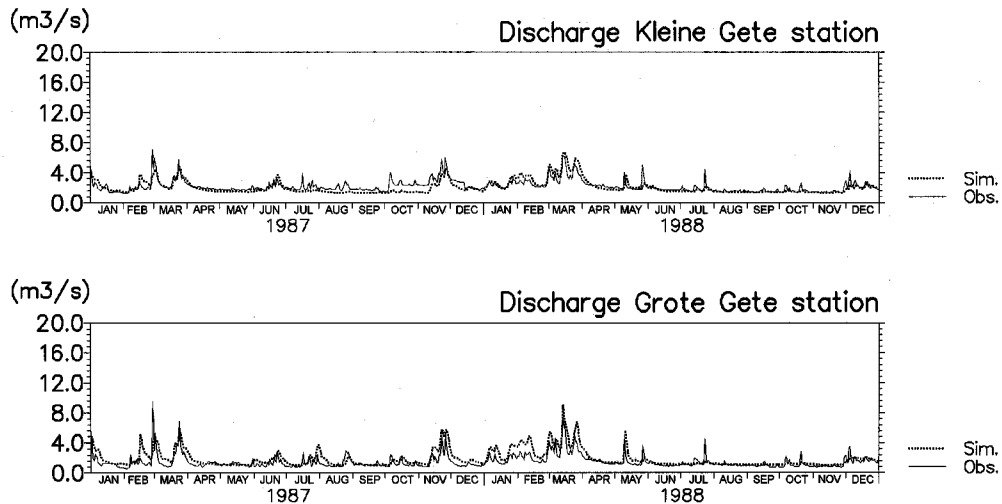


Fig. 14. Observed and simulated discharges versus time for the discharge station on the Grote Gete in Hoegaarden and the Kleine Gete in Budingén (validation period).

gen (Kleine Gete) are presented. The calculated statistical criteria are given in Table 3. The results are considerably worse than those in the calibration and simple split-sample validation. The absolute differences between the observed and simulated daily discharges did not increase in comparison to the calibrated Gete station but, relatively, the difference is increased as a consequence of the lower flows of these two stations. This shows in the lower values for the ABSERR, but inferior results obtain for the other statistical parameters, in particular the modelling efficiency. The simulated time series for the discharge station on the Kleine Gete shows little variation for the period July 1987–November 1987. This causes the discharge for October and November 1987 to be strongly underestimated by the model which is reflected in the values of the statistical parameters. For the Grote Gete, the base flow is relatively well simulated, while the peaks are underestimated and the simulated recession is somewhat too slow in comparison to the observed recession.

In Fig. 15, the simulated and observed heads in the six observation wells not used in the calibration are presented. Three of the wells (S2B-103, 4038488, and V2HG-BR.B5) are located in the vicinity of calibrated wells, while the other wells are further away from the observation wells (Fig. 7). Because they are near the calibrated wells, the first three wells behave similarly to wells S2B-102, 4038017, and V2HG-GEMPU, respectively. In well 4038488, both the position and fluctuation of the water table are estimated relatively well, while in wells S2B-103 and V2HG-BR.B5 the positions are underestimated and overestimated respectively. For the three other wells, the results are also very different. For wells 4045126 and 4082308, relatively close to each other, the piezometric levels are highly overestimated. In the calibrated wells 4047139 and 4048204, which are not far away from these two wells, the results are satisfactory. This strengthens the belief that good results would also have been established for these two wells had they been included in the calibration process. There does not seem to be a

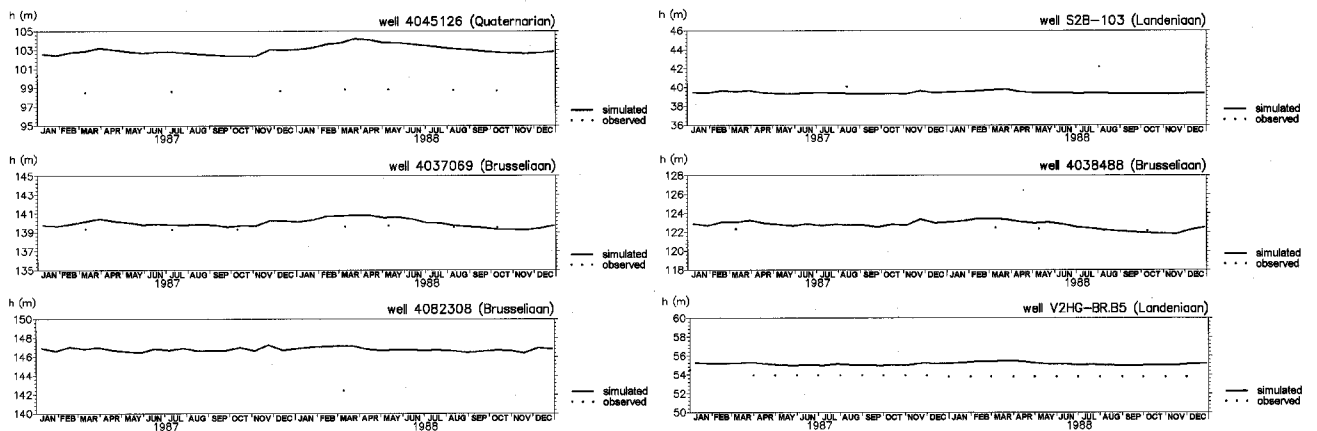


Fig. 15. Simulated and observed piezometric levels in the six observation wells only used in the model validation (validation period).

pronounced pattern in the differences between the simulated and observed piezometric levels.

Conclusions

For the 600-km² catchment of the Kleine Gete and the Grote Gete, the MIKE SHE model was calibrated and validated using the discharge station at the outlet of the catchment, two internal discharge stations and 18 observation wells. Results suggested that the model is capable of simulating an integrated state variable like the discharge with relative accuracy, but that the distributed results show large variance. During the validation process, good results were obtained for the integrated discharge of the stream basin but, for the internal discharge stations that were not used in the calibration process, the results were of inferior quality. The predictions of the piezometric level in internal observation wells differed considerably. The behaviour of wells located in the vicinity of calibrated wells was similar to that of the calibrated wells.

The study concluded that the calibrated model is a reasonable simulator of the discharge of the whole catchment and that the piezometric levels can be estimated roughly at certain internal observation points. To get more reliable and accurate results, the model needs additional data. Information on the moisture content in the unsaturated zone will allow this part of the model to be calibrated, which will improve the estimation of the runoff and the recharge to the water table. Additional hydro-geological information in some areas of the catchment and more distributed observations of the piezometric level in the different hydro-geological layers may lead to a better optimisation of the model and to accurate spatially distributed predictions of the state variables, which is ultimately the goal of distributed physically-based codes like MIKE SHE. However, as long as these models are only provided with little, non-distributed, and uncertain information, one cannot expect reliable spatially distributed results; the model predictions can only be as good as the input to the model. The most important challenge in relation to the applicability of distributed physically-based models consists of gathering, in a consistent way, sufficient, continuous, reliable distributed data.

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