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Simulating Stream Networks: An Alternative Approach for Model Evaluation in Ungauged Basins

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Diplomarbeit unter Leitung von Prof. Dr. Markus Weiler

Freiburg i.Br., Oktober 2008

Albert-Ludwigs-Universität Freiburg i. Br.
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Ehrenwörtliche Erklärung:

Hiermit erkläre ich, dass die Arbeit selbständig und nur unter Verwendung der angegebenen Hilfsmittel angefertigt wurde.

Freiburg i. Br., 2. Oktober 2008 Sebastian Stoll

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'[…] the single best recommendation on the issue of rainfall-runoff modelling of ungauged catchments may be to install a stream gauge!'

BLÖSCHL, G. 2005

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Symbols

A	[m^2]	Area
a		Thornthwaite coefficient
b		Decay coefficient
β		Water table slope
CF_{max}	[mm/(°C d)]	Degree-day-factor
CFR		Refreezing coefficient
D_d	[1/m]	Drainage density
E_{act}	[m]	Actual evapotranspiration
E_{pot}	[m]	Potential evapotranspiration
$\mathrm{Error}_{\mathrm{ETP}}$	[%]	Relative evapotranspiration error
$\mathrm{Error}_{\mathrm{Q}}$	[%]	Relative discharge error
HHQ	[m^3/s]	Highest observed discharge
I	[°C]	Heat index
K		Kappa coefficient
k_c	[m/s]	Constant saturated hydraulic conductivity
k_d	[1/s]	Empirical coefficient
k_0	[m/s]	Saturated hydraulic conductivity at spoil surface
k_s	[m/s]	Saturated hydraulic conductivity
$\sum L$	[m]	Total stream length
LLQ	[m^3/s]	Lowest observed discharge
$\log n_{eff}$		Logarithmic Nash-Sutcliffe model efficiency
$MELT$	[mm/d]	Amount of snowmelt
MHQ	[m^3/s]	Mean of the highest discharges
MLQ	[m^3/s]	Mean of the lowest discharges

MQ	[m^3/s]	Mean annual discharge
n		Number of days in the month
n_{avg}		Average drainable porosity
n_d		Drainable porosity
n_{eff}		Nash-Sutcliffe model efficiency
n_0		Drainable porosity at soil surface
p		Relative frequency
P	[mm]	Precipitation
p_a		Proportion of agreement
p_c		Proportion of chance agreement
Q	[m^3/s]	Discharge
Q_{obs}	[m^3/s]	Observed discharge
$\overline{Q_{obs}}$	[m^3/s]	Mean observed discharge
Q_{sim}	[m^3/s]	Simulated discharge
$REFR$	[mm/d]	Amount of refrozen water
S_0	[h]	Average sunshine duration
T	[m^2/s]	Transmissivity
T_a	[°C]	Air temperature
T_m	[m^2/s]	Modelled Transmissivity
$\overline{T}$	[°C]	Mean monthly air temperature
TT	[°C]	Treshold Temperature
V_{sat}	[m^3]	Water volume in saturated zone
W	[m]	Water table depth
w	[m]	width of water flow
z	[m]	Depth in soil profile
Z	[m]	Total depth of the subsurface

Abbrevations

DWD	German national meteorological service
EVP	Evapotranspiration
GLUE	Generalised Likelihood Uncertainty Estimation
IAHS	International Association of Hydrologists
m. a.s.l.	Elevation above sea level
PUB	Predictions in ungauged basins
REC	Recharge to saturated Zone
SMELT	Snowmelt
SSF	Lateral subsurface flow
WaBoa	Water and soil atlas Baden-Württemberg

Abstract

Rainfall-runoff modelling in ungauged basins is one of the greatest challenges in recent hydrological research. The lack of runoff data necessitates the establishment of new innovative approaches for parameter identification and model validation. Besides the transfer of calibrated parameters from similar gauged catchments, the application of distributed data like flooding patterns as an alternative hydrological response seems to be promising.

Therefore, a new approach for model and parameter evaluation based on spatial stream network modelling was tested in five different catchments in Baden-Württemberg. In a first step, spatial explicit modelling of stream networks was performed by the modified hillslope model Hill-Vi by using climate data. Those networks were afterwards compared to a map and finally, the grade of agreement was evaluated. The Kappa goodness-of-fit statistics, a discrete multivariate technique which has been widely used for assessing the accuracy of remotely sensed data, has been selected as objective function.

Significant differences between good and poor simulations could be distinguished and the corresponding parameter sets seemed reasonable. Those parameters were subsequently used to run the model with observed time series of daily precipitation and air temperature. The performances were evaluated with respect to the reflection of the discharge dynamics and the corresponding water balances.

The simulations have shown some promising results but also some evident limitations. Although the model’s actual representation of evapotranspiration and decline of drainable porosity and hydraulic conductivity needs to be improved, satisfying results could be achieved. Stream network modelling, which can be characterised by minimal data requirement, seems to be a reasonable alternative evaluation approach for ungauged basins.

Key words:

Stream networks, predictions in ungauged basins, model calibration, Kappa statistics, distributed data.

Zusammenfassung

Einer der größten Herausforderungen der aktuellen hydrologischen Forschung liegt in der Niederschlag-Abfluss Modellierung von ungemessenen Einzugsgebieten. Das Fehlen von Abflussdaten macht es notwendig neue innovative Alternativen zur Kalibrierung und Evaluierung von Modellen und Parameter zu entwickeln. Neben der Regionalisierung von bereits kalibrierten Parametern aus gemessen Einzugsgebieten, scheint die Anwendung von räumlich aufgelösten Daten (z.B. Überflutungsmuster) als alternative hydrologische Antwort vielversprechend zu sein.

Deshalb wurde die Tauglichkeit eines neuartigen Ansatzes der auf der räumlichen Modellierung von Flussnetzen beruht, in fünf Baden-Württembergischen Einzugsgebieten überprüft. Mit Hilfe des hydrologischen Modells Hill-Vi wurden unter Verwendung klimatischer Daten Flussnetze simuliert und anschließend der Grad der Übereinstimmung anhand von bestehenden Referenzkarten bestimmt Dazu wurde die Kappa Statistik, ein multivariantes Verfahren, das in der Fernerkundung häufig zur Bestimmung der Genauigkeit von erhoben Daten Verwendung findet als Gütemaß genutzt.

Gute Modellierungsergebnisse und die zugehörigen Parameterkombinationen konnten anhand dieses Gütemaßes identifiziert und anschließend zur weiteren Modellierungen unter Verwendung täglich gemessener Niederschlags- und Temperaturdaten eingesetzt werden. Diese Simulationen wurden wiederum anhand ihrer Wiedergabe der Abflussdynamik und der entsprechenden Genauigkeit der Wasserbilanz bewertet.

Diese zeigten einige viel versprechende Ergebnisse, jedoch auch schwerwiegende Probleme und Einschränkungen auf. Die Abbildung der Verdunstung und der Abnahme der hydraulischen Leitfähigkeit mit der Tiefe bedürfen zwar einigen Verbesserungen, aber nichtsdestotrotz könnten noch zufrieden stellende Ergebnisse erzielt werden. Es scheint, dass die Modellierung von Flussnetzen in der Tat ein geeignetes Mittel zur Evaluierung von Modellen in ungemessenen Einzugsgebieten zu sein scheint, das sich vor allem durch einen geringen Datenbedarf auszeichnet.

Stichworte:

Gerinnenetze, Modellierung in ungemessenen Einzugsgebieten, Modellkalibrierung, Kappa Statistik, räumlich aufgelöste Daten.

1. Introduction

Rainfall-Runoff modelling is one of the most important research subjects in hydrology. There has been an enormous progress in model development in the past few decades. Nowadays there is wide range of model types with different specifications in spatial resolution and implementation of physical laws and runoff generation processes.

New challenges for the hydrological community arise from the research initiative Predictions in Ungauged Basins (PUB) of the International Association of Hydrologists (IAHS), launched in 2003. PUB deals with the prediction of the hydrological response of ungauged basins to the impacts of global change. According to SIVAPALAN ET AL. 2003, an ungauged basin is defined as *"[…] one with inadequate records (in terms of both data quantity and quality) of hydrological observations to enable computation of hydrological variables of interest (both water quantity or quality) at the appropriate spatial and temporal scales, and to the accuracy acceptable for practical applications"*.

In those basins the lack of runoff data, which is normally used to calibrate parameters, is the main problem for Rainfall-Runoff modelling. The calibration of parameters is essential to account for the heterogeneities and boundary conditions of the catchment and thus unavoidable for a successful modelling (BLÖSCHL 2005).

As model parameters cannot be obtained through calibration on runoff data in ungauged basins, alternatives for parameter determination are needed.

1.1 Objective and procedure

The aim of this thesis is to develop and test a new approach for model and parameter evaluation, based on drainage network modelling and irrespective to stream flow data. This includes modifications on the conceptual hillslope model Hill-Vi (WEILER & MCDONNELL 2004) with respect to processes affecting the initiation of stream networks, the search for an adequate objective function, which is sensitive to parameter variations, and the testing of the method in several catchments in Baden-Württemberg.

In a first step, spatial explicit modelling of stream networks will be performed by the hillslope model Hill-Vi using climate data of the reference period 1961-1990. Secondly, the modelled networks will be compared with a map and the grade of agreement will be evaluated by the objective function. Finally the model will be run with new time series of input data to test the efficiency of the parameters that have been calibrated before. A conceptual illustration of the procedure is given in figure 1.

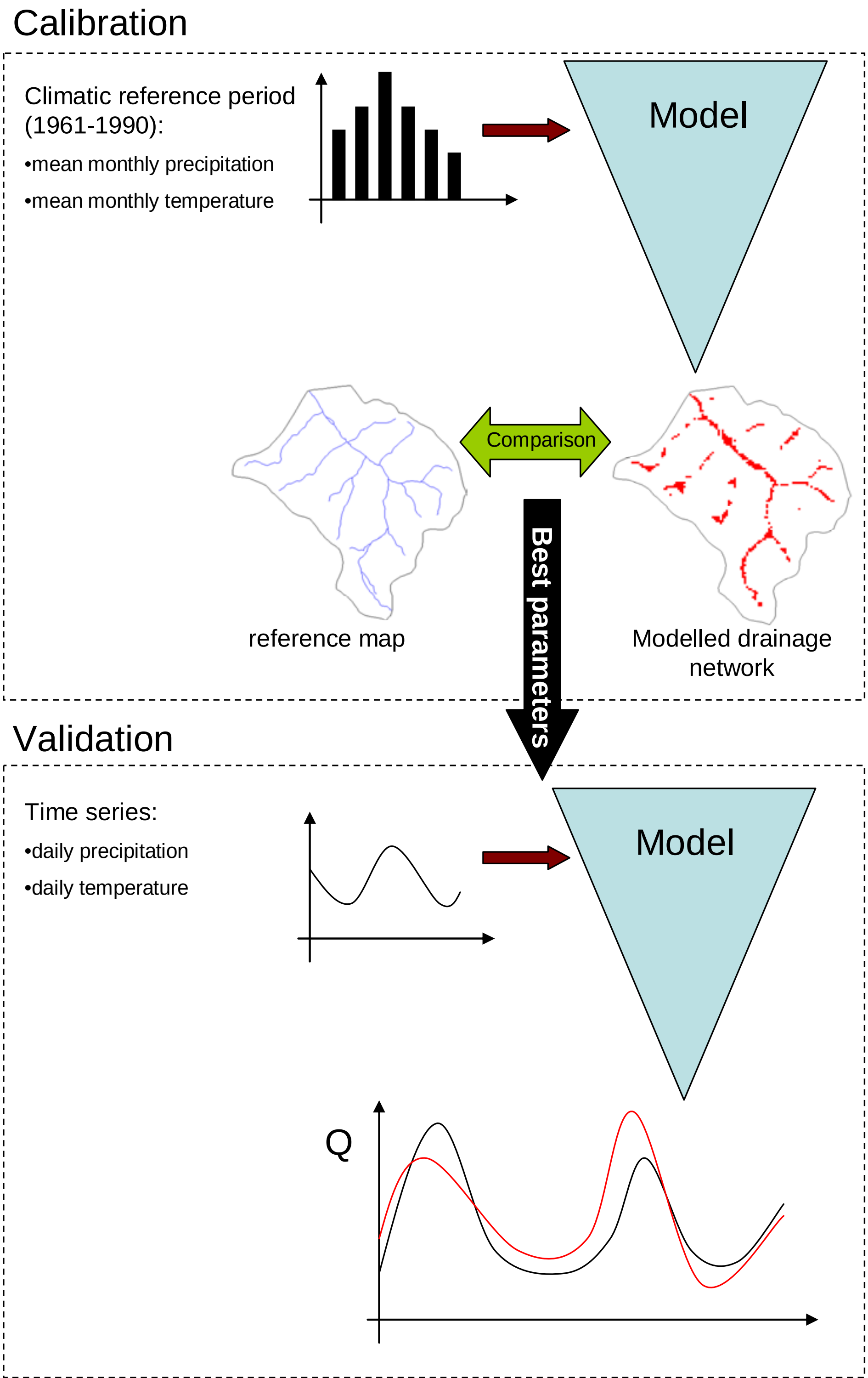


Figure 1: Conceptual procedure of an alternative model evaluation

1.2 State of the art

Due to the lack of runoff data in ungauged basins, alternative methods, providing the parameters of Rainfall-Runoff models were established.

1.2.1 Transfer from hydrologic similar catchments

The concept of transferring parameters from a gauged to an ungauged basin is based on the assumption that two similar catchments will respond identically to the same input. In the past years different specifications of similarity have been presented. BLÖSCHL 2005 describes the traditional concept, saying that catchments which are located close to each other are expected to respond similarly, while it is assumed that the regulation mechanisms (e.g. geology) stay constant. Another method to identify similar basins is the comparison of catchment attributes, e.g. soil types. The hydrological response is assumed to be similar, if the feature is identical. Numerous regionalisation methods (e.g. NATHAN & MCMAHON 1990, ABDULLA & LETTENMAIER 1997, KOKKONEN ET AL. 2003, PARAJKA ET AL. 2005), including the well known hydrological response unit approach (e.g. FLÜGEL 1995) were applied to establish relationships between model parameters and attributes and to transfer them to ungauged basins. Other approaches try to define similarity by the comparison of calculated indices. Their objective is to achieve few values which are representative for the whole catchment. It is assumed that the calculated indices integrate all relevant features. An example is the instantaneous geomorphologic hydrograph introduced by RODRIGUEZ-ITURBE & VALDES 1979, which relates runoff with geomorphologic characteristics.

Several studies (e.g. MERZ & BLÖSCHL 2004, SEIBERT 1999) tested the performance of regionalised model parameters. Even though they reported relative acceptable model efficiencies, they detected poor correlation between catchment attributes and model parameters and a significant decline of model performance when moving from gauged to ungauged basins. Those poor results could be explained by several reasons. BLÖSCHL 2005 refers to the possibility that the usually regionalised catchment features, e.g. land use, may not be relevant for the major hydrological processes, which take place in the subsurface and control the response. Furthermore, he emphasises the problem of uncertainty of the calibrated parameters (e.g. BEVEN 1996). Besides the problem of equifinality, BEVEN 2001a also presents the problem of uniqueness. Exporting parameters from gauged to ungauged basins does not guarantee that they will produce the same good results as under the circumstances they were determined. This is caused by the missing of the balancing effect of the calibration process on specific catchment heterogeneities.

According to the challenges presented above, the application of models using regionalised parameters can be problematic. It should be expected that physically based models, e.g. MIKE SHE (REFSGAARD & STORM 1995), following the blue print by FREEZE & HARLAN 1969, should be able to face the problems of ungauged basins. Unfortunately it has shown that a calibration is unavoidable despite the physical meaning of the parameters. The first difficulty arises from the fact that the measuring scale is not identical to the model scale (BEVEN 1989). Further studies also reported (e.g. JAKEMAN & HORNBERGER 1993, UHLENBROOK ET AL. 1999) that, due to the large number of parameters, an overparameterisation can occur. Thus, HUNDECHA ET AL. 2007 summarise that *'[…] physically based models could in principle do better, but they will not actually do better'*.

A comprehensive compilation on hydrological similarity is given in the recent work of WAGENER ET AL. 2007.

1.2.2 Spatial patterns as alternatives to runoff data

Besides runoff data, alternative hydrological responses can be used to calibrate model parameters. Observed spatial patterns like snow cover and runoff occurrence are suitable for calibrating and evaluating distributed hydrological models (REFSGAARD 2001, GRAYSON ET AL. 2002). Consequently, examples for which different outputs of distributed models were used to calibrate or evaluate model parameters are presented.

BLAZKOVA ET AL. 2002 examined the benefit of the comparison of modelled and observed saturated areas in the small Uhlirska catchment, Czech Republic on the improvement of parameter identification in the conceptual rainfall-runoff model TOPMODEL (BEVEN & KIRKBY 1979). They used the Generalised Likelihood Uncertainty Estimation (GLUE) methodology (BEVEN & BINLEY 1992) to produce parameter sets and evaluated them with respect to discharge predictions and agreement with mapped saturated areas. The evaluation of saturated area information showed that mainly the transmissivity parameter reacted sensitive. Variations of other parameters hardly affected the performance. On the other hand, the transmissivity variations were not significant on the containment of prediction bounds of the discharge. Thus, the authors concluded that the benefit of the evaluation of internal state variables in constraining parameter values is greater than the constraining of discharge predictions.

BAUER ET AL. 2006 studied the impacts of upstream and local activities on the wetlands of the Okavango Delta in Botswana. A coupled physically based surface water/groundwater model was developed to simulate the results of human interventions and climate change on the size of flooded areas. Due to sparse and unreliable runoff data, binary satellite-derived inundation patterns were compared with modelled flooding extents to calibrate the seven model parameters, whereas two different objective functions based on pixel-by-pixel comparison and total flooded area were applied. Additionally, relative water levels at different stations in the Okavango Delta were used to carry out an evaluation of the calibrated parameter (figure 2). The results of the calibration and evaluation were satisfying, since the characteristics of the annual floods in the delta were successfully reproduced. Thus, the model was used to test and analyse scenarios in order to produce decision support for the local stakeholders.

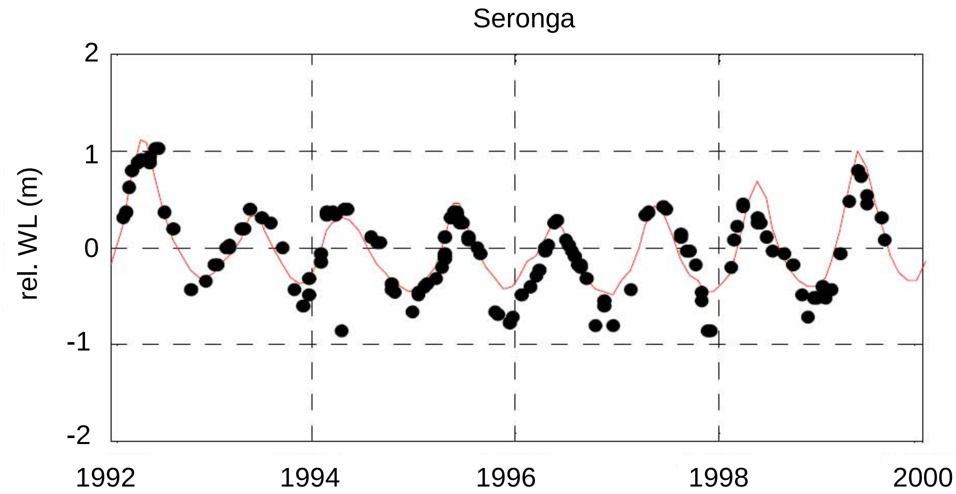


Figure 2: Modelled (red) and observed water levels (black) at the station Seronga (Bauer et al. 2006).

BLÖSCHL ET AL. 2002 presented an approach to calibrate model parameters in the highly karstified Schneealpe catchment in the Austrian Alps. Focussing on snow processes, satellite-derived patterns of snow cover were used for parameter identification of the snow melt module. Based on the energy balance and a wind drift factor, the module was able to simulate the development and melting of the snow pack, which was compared to satellite images. While snow albedo and threshold temperature were calibrated globally, the refinement of the wind drift parameters was done individually for each pixel cell. Afterwards the model was tested in a split sample test in which a significant improvement on snow melt and minor effects on the bulk runoff were noticed.

VERTESSY & ELESENBEER 1999 applied the process-based, distributed Topog_SBM model in the micro (0.75ha) research catchment of La Cuenca in western Amazonia, Peru, to examine the impacts of the spatial distribution of saturated hydraulic conductivity on the occurrence of runoff. A huge amount of undisturbed soil cores for hydraulic measurements and 72 overland flow detectors were available. Eight parameter realisations with different spatial distribution of saturated hydraulic conductivities and the corresponding shape parameters were used to simulate patterns of overland flow, which were compared to maps that derived from the analysis of the flow detectors. Additionally, the total runoff, the peak runoff, the time of rise, and the lag time were evaluated. With the best parameterisation the model was able to predict an overland flow frequency distribution that agreed with the results of the detectors although the spatial extent was overestimated in many cases. The total runoff amount was modelled with a good accuracy but the peak discharge was often underestimated.

The four examples have shown that spatial patterns of different hydrological variables are suitable, respectively recommendable, to evaluate or calibrate distributed hydrological models (GRAYSON ET AL. 2002, BLÖSCHL 2005).

2. Study areas

To test the methodology, five small catchments representing the most important geological units in Baden-Württemberg were selected. Those are the Zastlerbach, the Haslach, the Eyach, the Fichtenberger Rot and the Roggelshäuser Bach. The main data input derives from the spatial data of the digital version of the water and soil atlas of Baden-Württemberg (WaBoA). In figure 3 an overview of the locations is given.

Figure 3: Overview of the catchments' locations in Baden-Württemberg.

2.1 Zastlerbach

The Zastlerbach catchment with an area of 17.9 km^2 is located in the southern Black Forest Mountains in south-western Baden-Württemberg. The elevation ranges from 548 to 1493 meters above sea level (mean = 1053 m a.s.l.) with a mean slope of 35.2%. The topography and the locations of the precipitation stations, operated by the German National Meteorological Service (DWD), are displayed in figure 4.

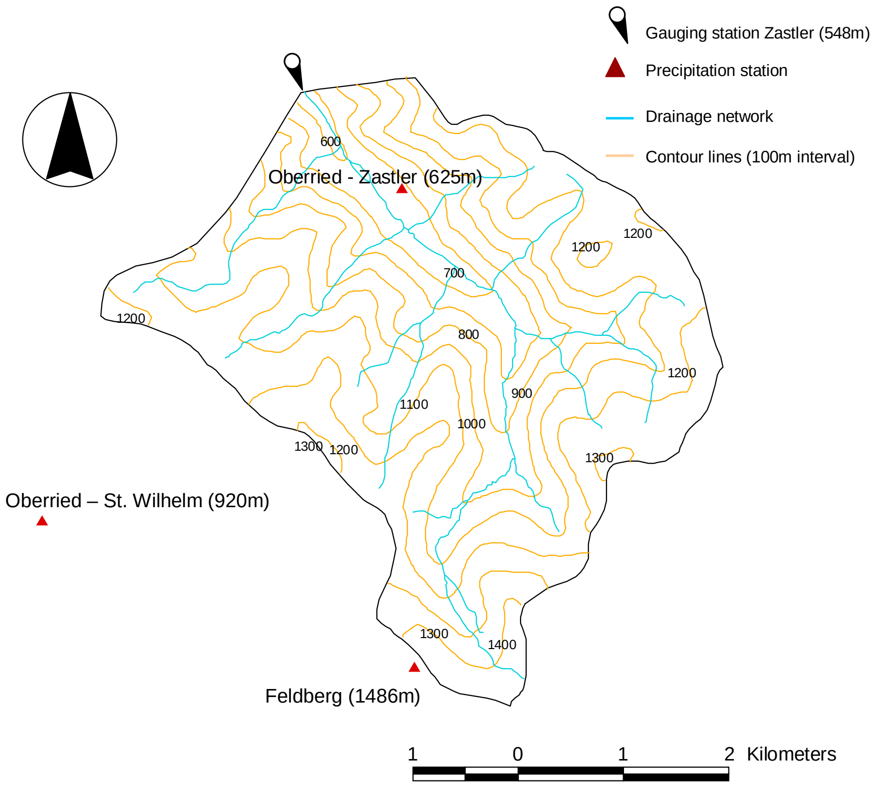


Figure 4: Overview of Zastlerbach basin, southern Black Forest Mountains, Baden-Württemberg.

2.1.1 Climate and hydrology

The local climate is dominated by the distinct topography, which implicates a strong dependence of precipitation and air temperature on altitude. The significant increase in altitude of the mean monthly precipitation at three meteorological stations of the climatic reference period 1961-1990 is illustrated in figure 5. The mean annual precipitation sum adds up to 1654 mm at Oberried–Zastler, to 1814 mm at Oberried-St.Wilhelm and to 1912 mm at Feldberg.

According to precipitation, the annual variability of mean monthly air temperature (1961-1990) at Feldberg and the nearby stations Hinterzarten (883 m a.s.l.) and Freiburg (269 m a.s.l.) shows a strong orographic effect (figure 6). The low temperatures in winter result in a long-lasting snow cover that remains until April (WABOA 2007).

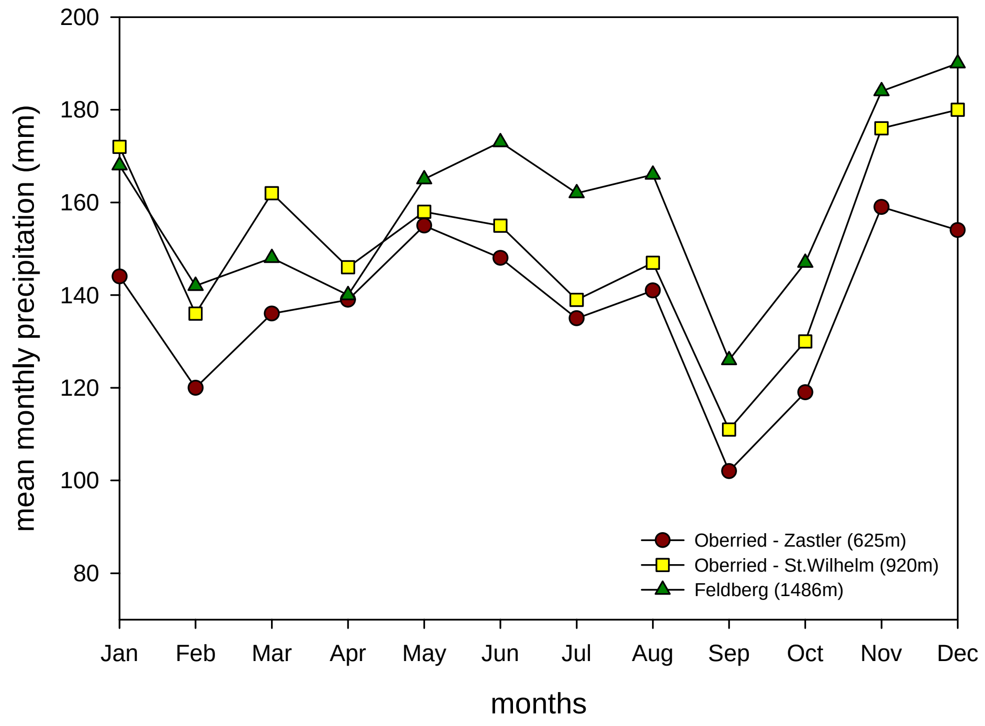


Figure 5: Mean monthly precipitation of stations near Zastlerbach basin for the period 1961-1990.

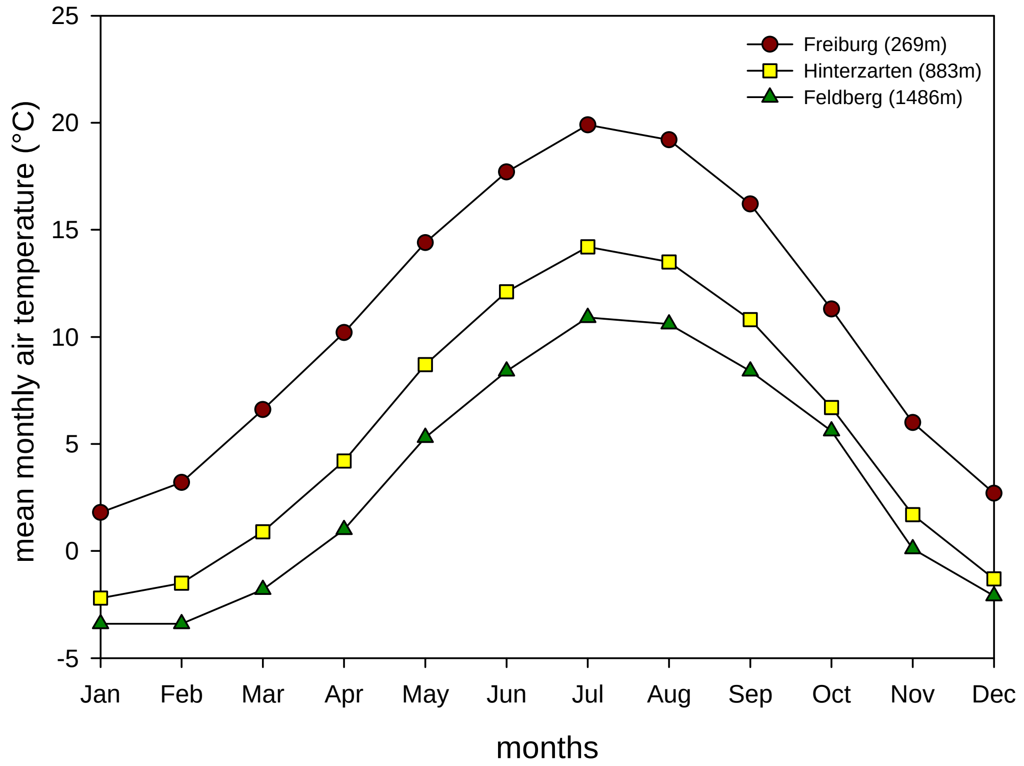


Figure 6: Mean monthly air temperature of stations near Zastlerbach basin for the period 1961-1990.

Based on the conceptual model of ARMBRUSTER 2002, daily interpolated precipitation data of the BONIE approach (REICH & GÜNTHER 1999) and climatic records of the DWD, the actual evapotranspiration was calculated in WABOA 2007 for a 500 to 500 m raster. The mean annual actual evapotranspiration for the Zastlerbach basin averages 531 mm/a.

Due to comparable elevation, the stations Hinterzarten and Oberried-St.Wilhelm were therefore chosen to represent the catchment.

The runoff regime of the Zastlerbach (1961-1990) at the gauging station Zastler, represented by the Pardé coefficient (PARDÉ 1968), is displayed in figure 8. It can be classified as nivo-pluvial with a maximum during April and May caused by snowmelt, minimum discharges during the late summer, and a secondary maximum in winter due to the increased precipitation amounts.

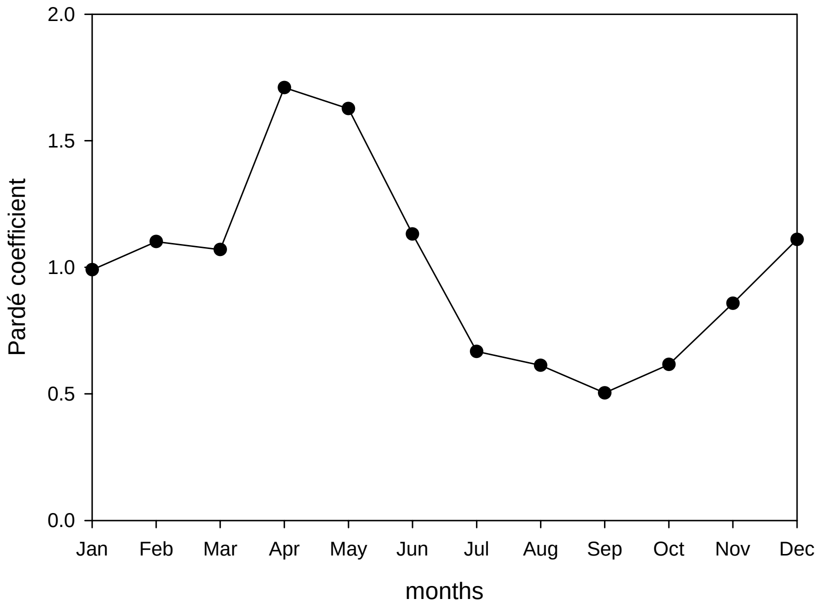


Figure 7: Runoff regime of Zastlerbach for the period 1961-1990.

Table 1 shows some standard runoff characteristics of the Zastlerbach for the climatic reference period. The annual mean runoff (MQ), the mean of the highest runoffs (MHQ), the mean of the lowest runoffs (MLQ) and the highest (HHQ) respectively the lowest runoff (LLQ) at all have been computed for the period 1961-1990.

Table 1: Runoff characteristics of Zastlerbach catchment for the period 1961-1990

	MQ (m^3/s)	MHQ (m^3/s)	MLQ (m^3/s)	HHQ (m^3/s)	LLQ (m^3/s)
Zastlerbach	0.65	4.24	0.14	14.183	0.074

2.1.2 Geology, pedology and land use

The Zastlerbach basin is situated within the basement of the Black Forest Mountains. The underlying bedrock consists of metaphoric rock, mainly gneiss and anatexists. According to WaBoA 2007 the hydraulic conductivity is estimated to 10^{-7}-10^{-6} m/s.

The bedrock is predominantly covered with a sequence of Quaternary material. Those permeable layers, which developed under periglacial conditions, can reach a thickness up to 10 m (Güntner et al. 2004) and can be classified as cambisols.

The land use in the year 2000, illustrated in figure 8, is dominated by forest. 61.9% of the catchment's area consists of coniferous forest and 28.3% by mixed forest. The remaining 9.8% of the basin is used as grassland.

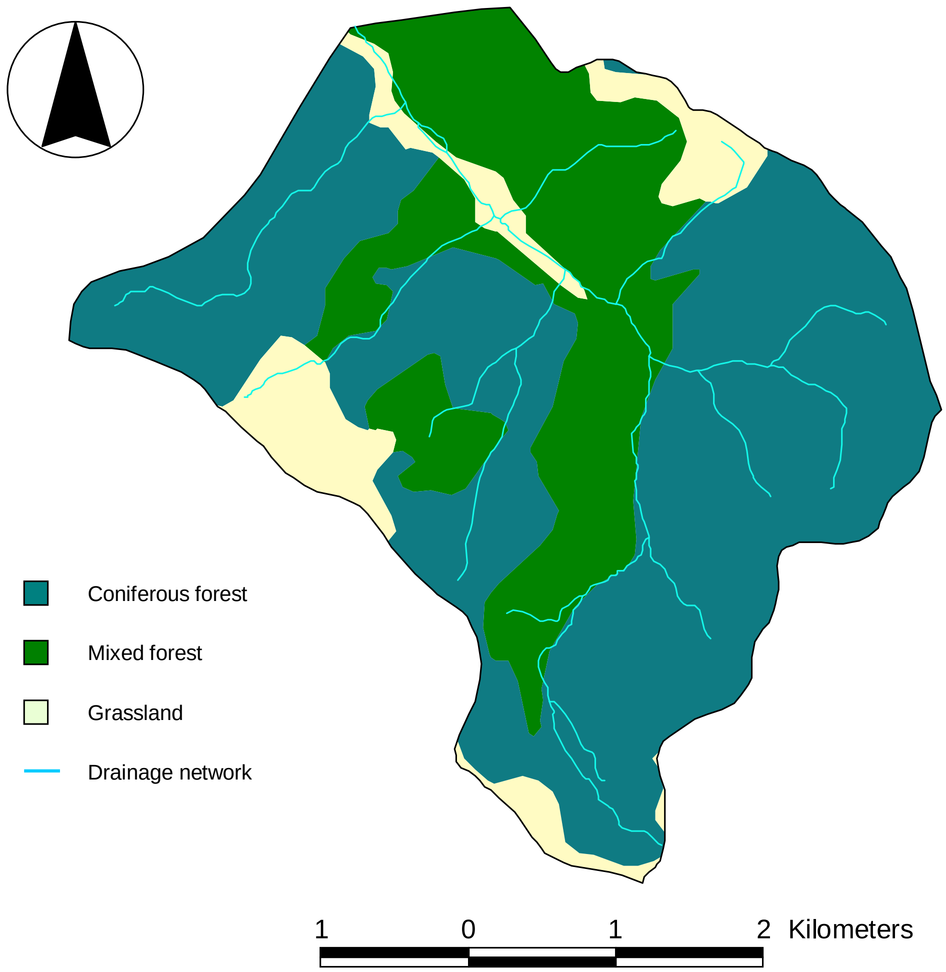


Figure 8: Land use of the Zastlerbach catchment.

2.2 Fichtenberger Rot

The basin of the Fichtenberger Rot is situated in the Frankish-Swabian Forest north-eastern of Stuttgart. It covers an area of 17.3 km^2 and the altitude ranges from 442 to 566 m with a mean elevation of 492 m above sea level. The flat catchment's hills have a mean slope of 8.3%. An overview of the catchment is given in figure 9.

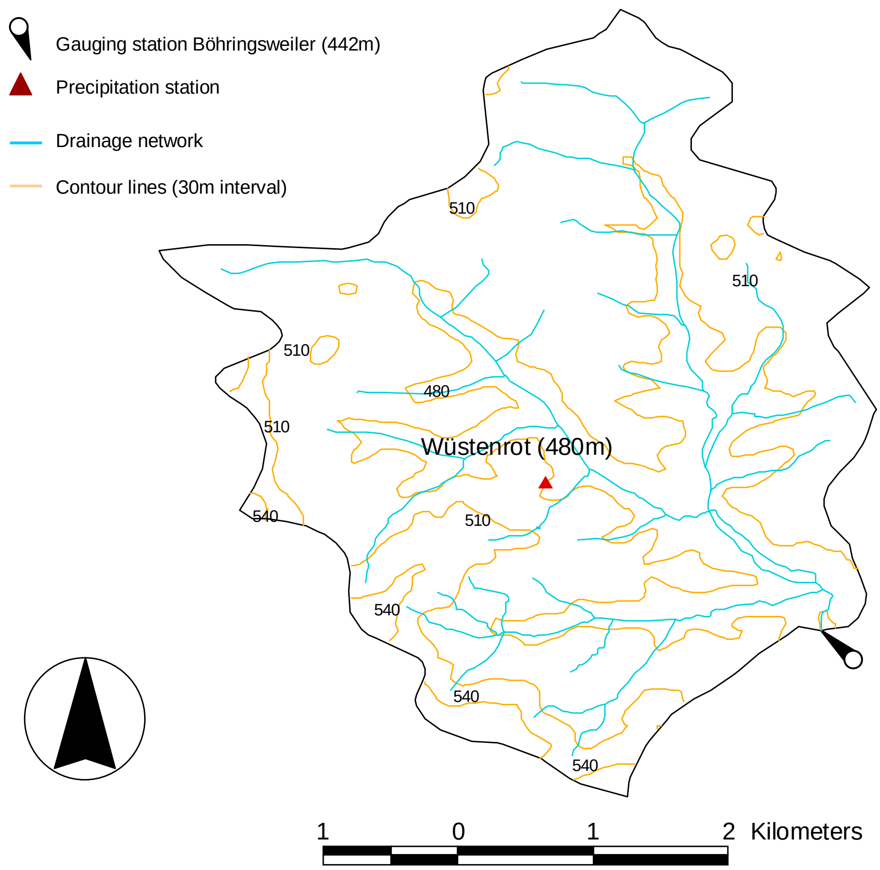


Figure 9: Overview of Fichtenberger Rot basin, Baden-Württemberg.

2.2.1 Climate and hydrology

The climate is strongly influenced by cyclonic west wind situations, delivering a mean precipitation of 1187.9 mm per year (1961-1990). The variability over the year with higher amounts in the summer, which is due to convective events, is illustrated in figure 10. The mean maximum amounts with up to 120 mm are reached in June, whereas September only shows 82 mm.

The mean annual air temperature for the climatic reference period of the station Gschwend (492 m a.s.l.), located about 20 km south-eastern of the outlet, is 7.7°C. Mean monthly temperatures below freezing point occur in January with -1.1°C and February and December each with -0.1°C (figure 11).

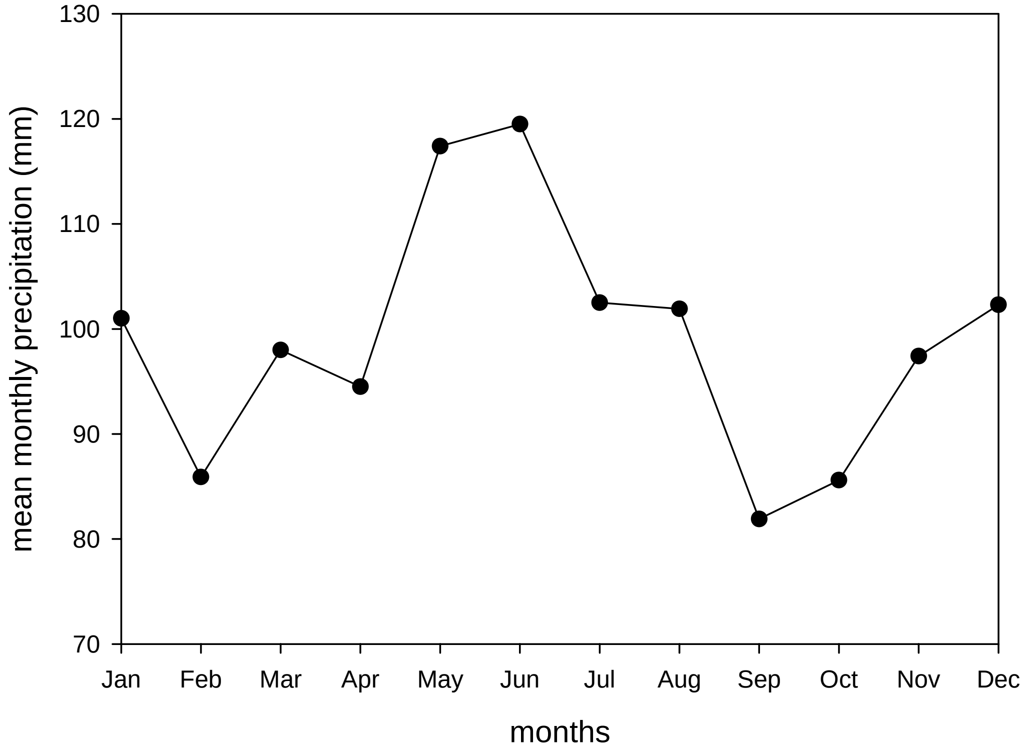


Figure 10: Mean monthly precipitation of the station Wüstenrot for the period 1961-1990.

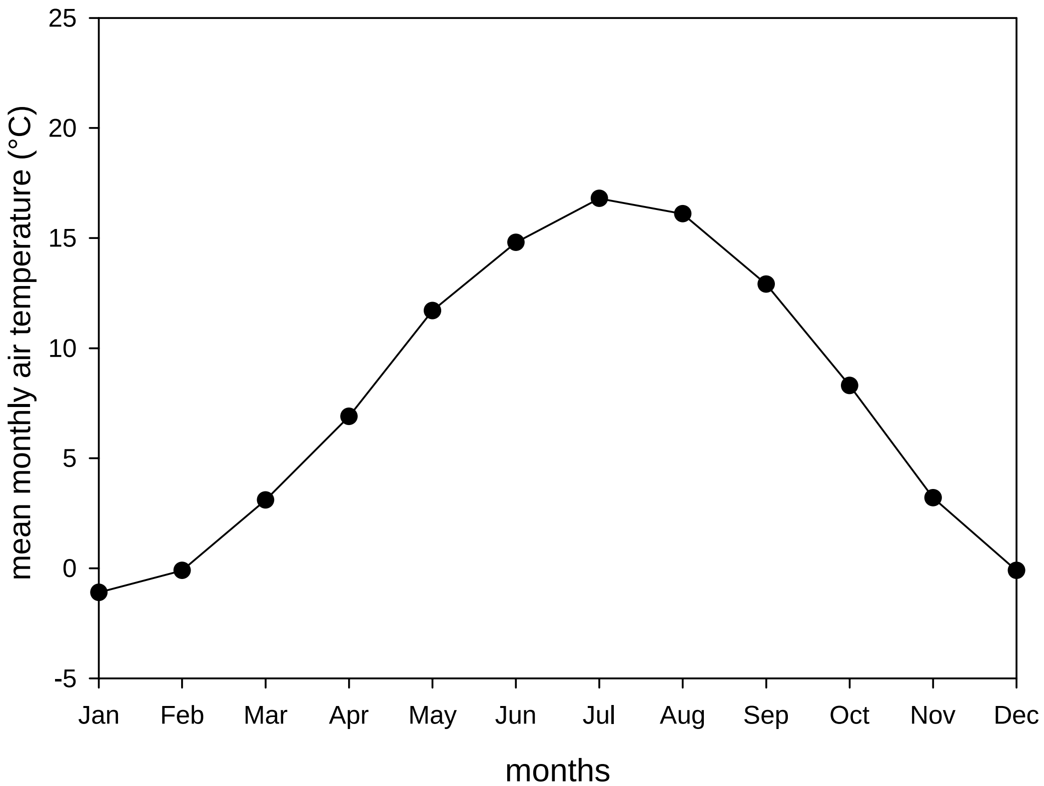


Figure 11: Mean monthly air temperature of the station Gschwend for the period 1961-1990.

The mean actual evapotranspiration of the Fichtenberger Rot basin calculated in WABOA 2007 averages 560 mm per year.

Represented by the Pardé coefficient, the seasonal runoff variation at the gauging station Böhringsweiler for the period 1977-1990 is illustrated in figure 12.

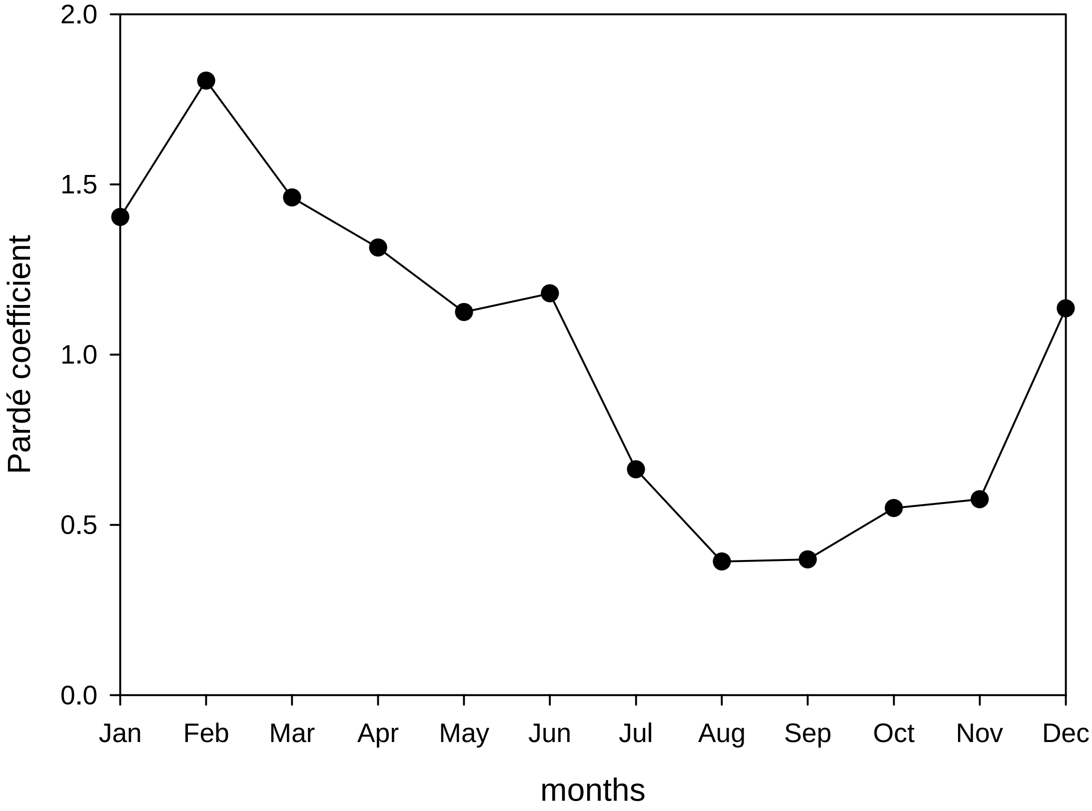


Figure 12: Runoff regime of Fichtenberger Rot for the period 1977-1990.

With the highest discharges in February and the minimum amounts in September, which are due to enhanced vegetation demand and evaporation losses in the late summer the regime can be classified as pluvial. The secondary maximum in June results from high precipitation in this month. The standard runoff characteristics of the Fichtenberger Rot for the period 1977-1990 are displayed in table 2.

Table 2: Runoff characteristics of the Fichtenberger Rot for the period 1977-1990

	MQ (m^3/s)	MHQ (m^3/s)	MLQ (m^3/s)	HHQ (m^3/s)	LLQ (m^3/s)
Fichtenberger Rot	0.29	3.97	0.04	8.164	0.021

2.2.2 Geology, pedology and land use

The catchment is situated within the area of the upper Keuper with sandstones and marls (GEYER & GWINNER 1991). WABOA 2007 estimates the hydraulic conductance to 10^{-5} m/s. Near the main channel highly permeable Quaternary gravels and sands with conductivities up to 10-3 m/s can be found.

The sediments are prevalently covered with podzolic cambisols with moderate conductivities. In the lower part of the basin, WABOA 2007 reports water meadows with gleyic soils.

The heterogeneous land use of the Fichtenberger Rot basin is illustrated in figure 13. The undulating catchment is dominantly used as forest with 44% of the area, followed by grassland with 33%, arable land (7%), discontinuous settlements, and complex cultivations (each 8%).

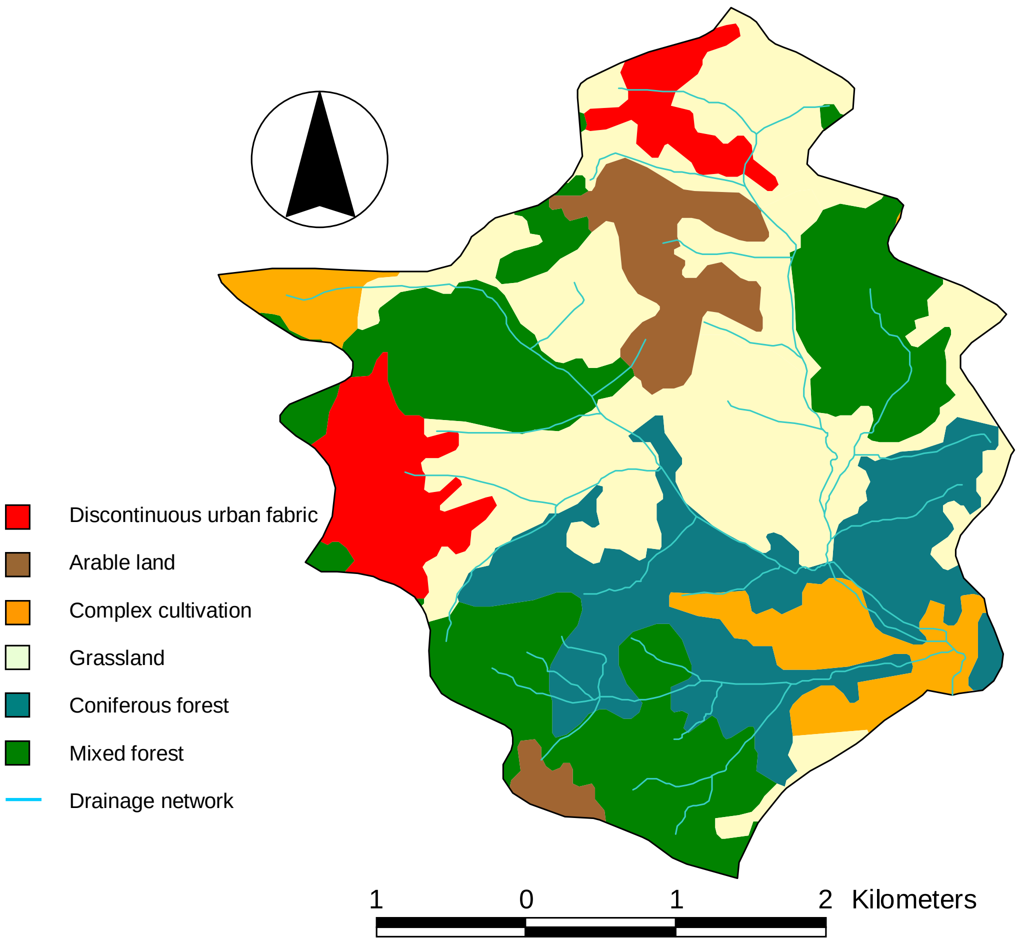


Figure 13: Land use of the Fichtenberger Rot catchment.

2.3 Haslach

With an area of 40.3 km^2 the Haslach basin is located in the region of upper Swabia in the very East of Baden-Württemberg, near of Memmingen. The elevation ranges from 608 to 750 meters a.s.l. with a mean of 688 m and a resulting slope of 5.4%. In figure 14 an overview of the topography and the nearby precipitation station in Bad Wurzach is given.

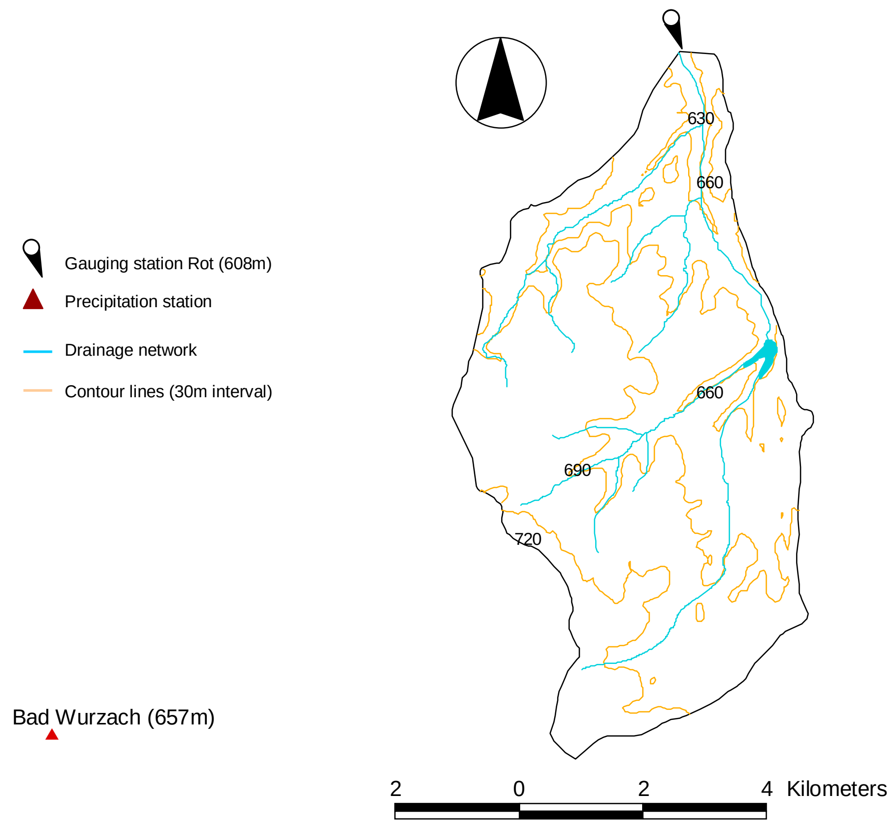


Figure 14: Overview of the Haslach catchment, Baden-Württemberg.

2.3.1 Climate and hydrology

The precipitation regime of the nearby station Bad Wurzach for the period 1961-1990 (figure 15) shows a distinctive maximum during the summer, with 125 mm in June. This is typical for the whole region of upper Swabia and caused by an increased blocking effect of the Alps (WABOA 2007). The precipitation sum adds up to 1024 mm per year.

The annual variability of the air temperature at the meteorological station Memmingen (610 m a.s.l.), which is located in the East of the catchment is illustrated in figure 16. The mean annual temperature averages 7.8°C and mean monthly temperatures below freezing point occur in January, February and December.

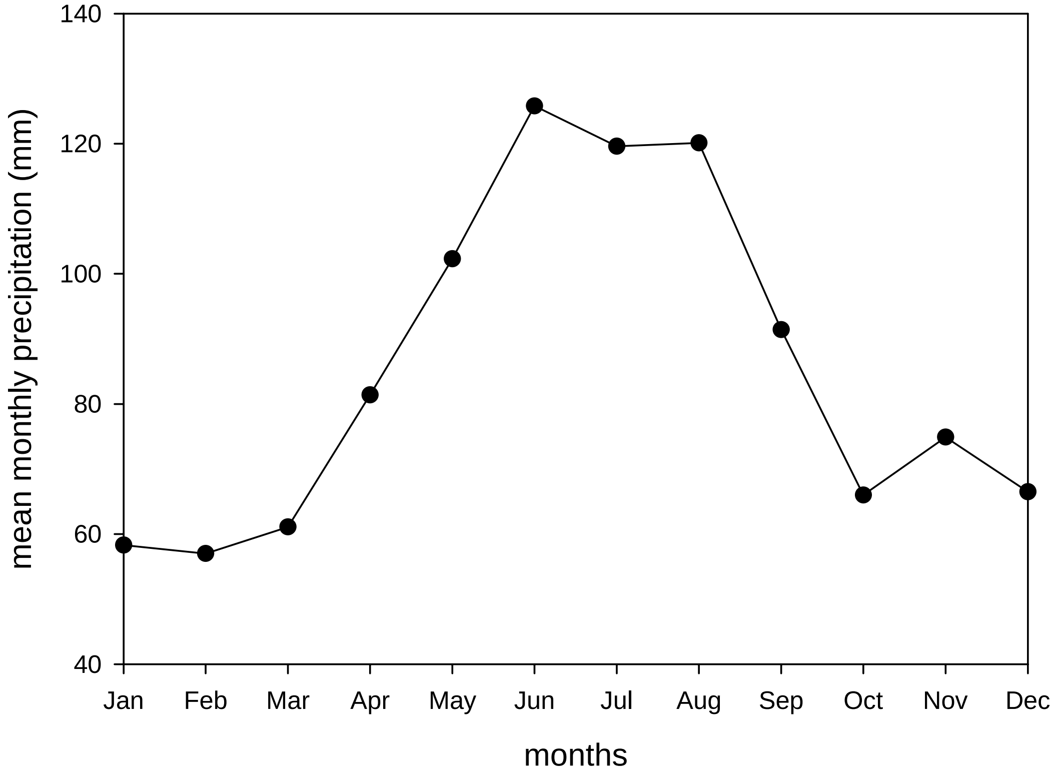


Figure 15: Mean monthly precipitation of the station Wurzach for the period 1961-1990.

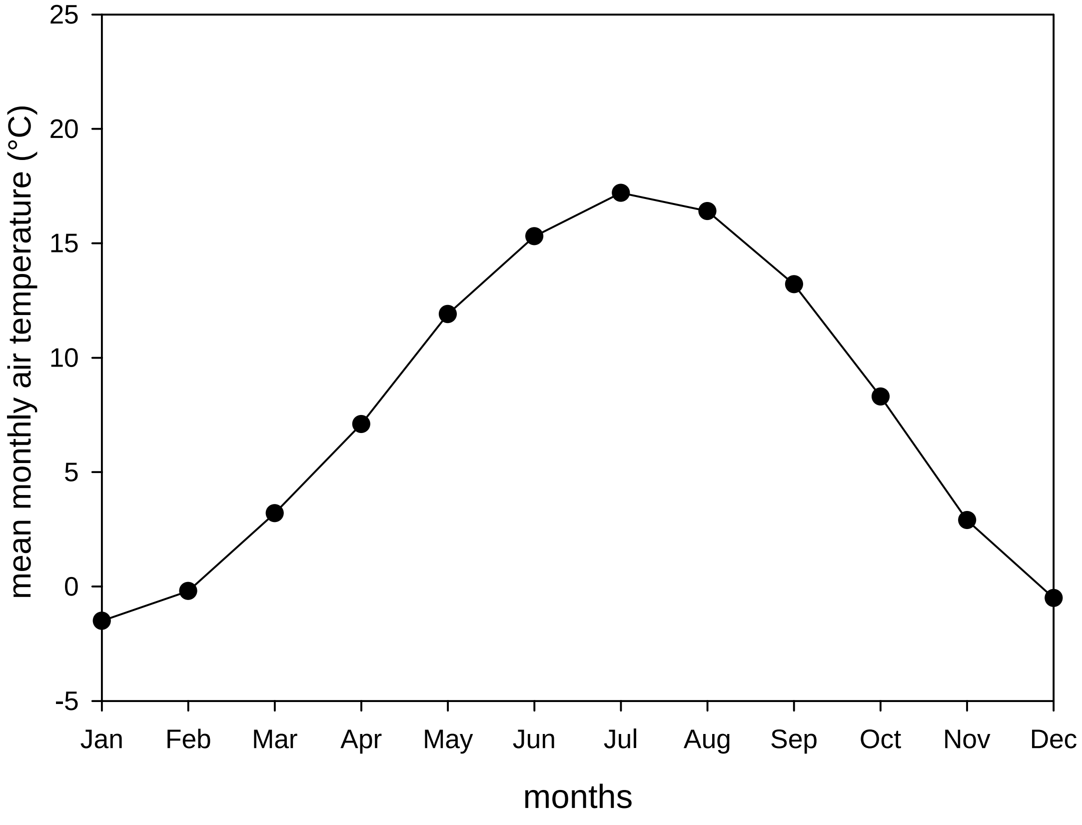


Figure 16: Mean monthly air temperature of the station Memmingen for the period 1961-1990.

The runoff regime of the Haslach at the gauging station Rot is illustrated in figure 17 using the Pardé convention. It shows a small seasonal variability of the runoff, resulting from the existence of a small lake (0.14 km^2), which acts like a buffer on the hydrograph. Nevertheless, a slight pluvial regime with a maximum in February and a minimum in the late summer is apparent.

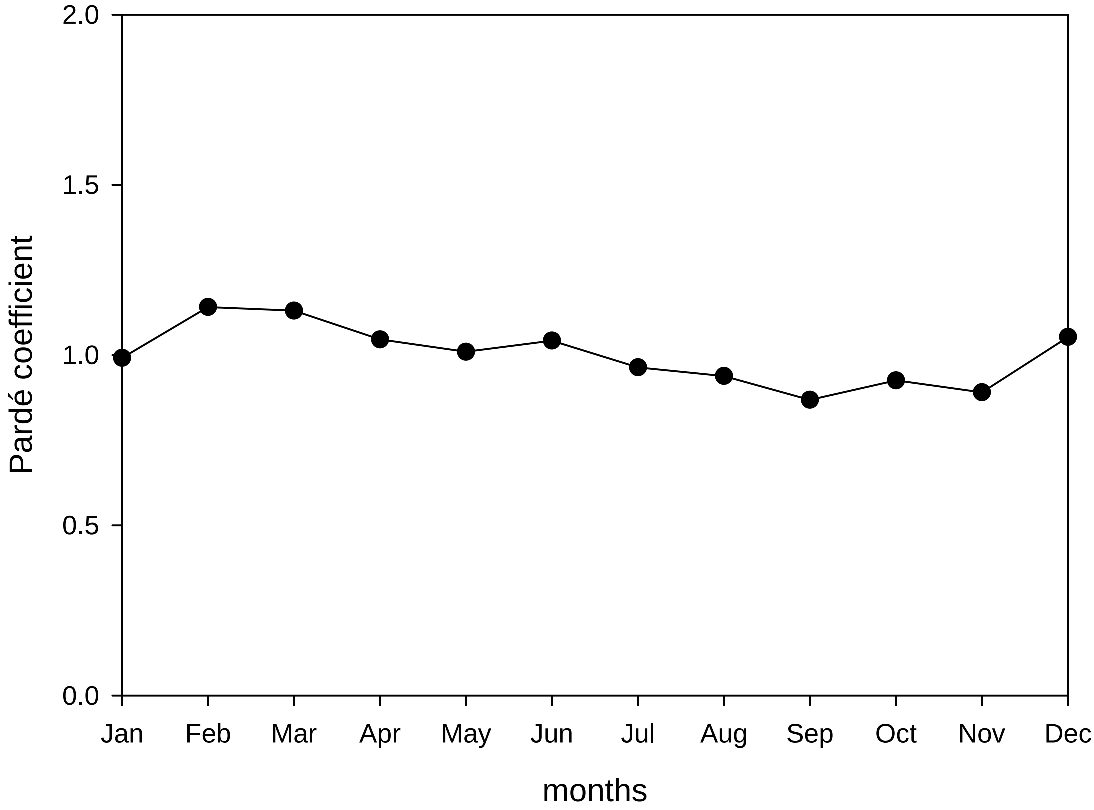


Figure 17: Runoff regime of Haslach for the period 1961-1982.

Some standard runoff characteristics for the period 1961-1982 are listed in table 3. The influence of the lake arises in the relatively high values of MLQ and LLQ, due to a constant outflow.

The reported annual actual evapotranspiration for the Haslach catchment in WABOA 2007 averages 611 mm.

Table 3: Runoff characteristics of the Haslach for the period 1961-1982

	MQ (m^3/s)	MHQ (m^3/s)	MLQ (m^3/s)	HHQ (m^3/s)	LLQ (m^3/s)
Haslach	0.72	5.01	0.36	8.13	0.22

2.3.2 Geology, pedology and land use

The region of upper Swabia was strongly influenced by the ice masses during several ice ages depositing huge amounts of sediments (GEYER & GWINNER 1991). The Haslach catchment is dominated by glaciofluvial debris with hydraulic conductivities up to 10^{-3} m/s and till with conductivities between 10^{-6} and 10^{-7} m/s. Sporadically, poorly permeable molasses (10^{-7} to 10^{-6} m/s) can be found (WABOA 2007).

The quaternary material is mostly covered with luvisols, also occurring as podzolic and pseudogleyic versions. Downriver of the little lake WABOA 2007 maps water meadows with gleyic soils.

The land use, illustrated in figure 18 is dominated by grassland with 54% of the catchment's area and forest with 44%. The remaining 2% of the basin is portioned in each 1% discontinuous settlements and complex cultivations.

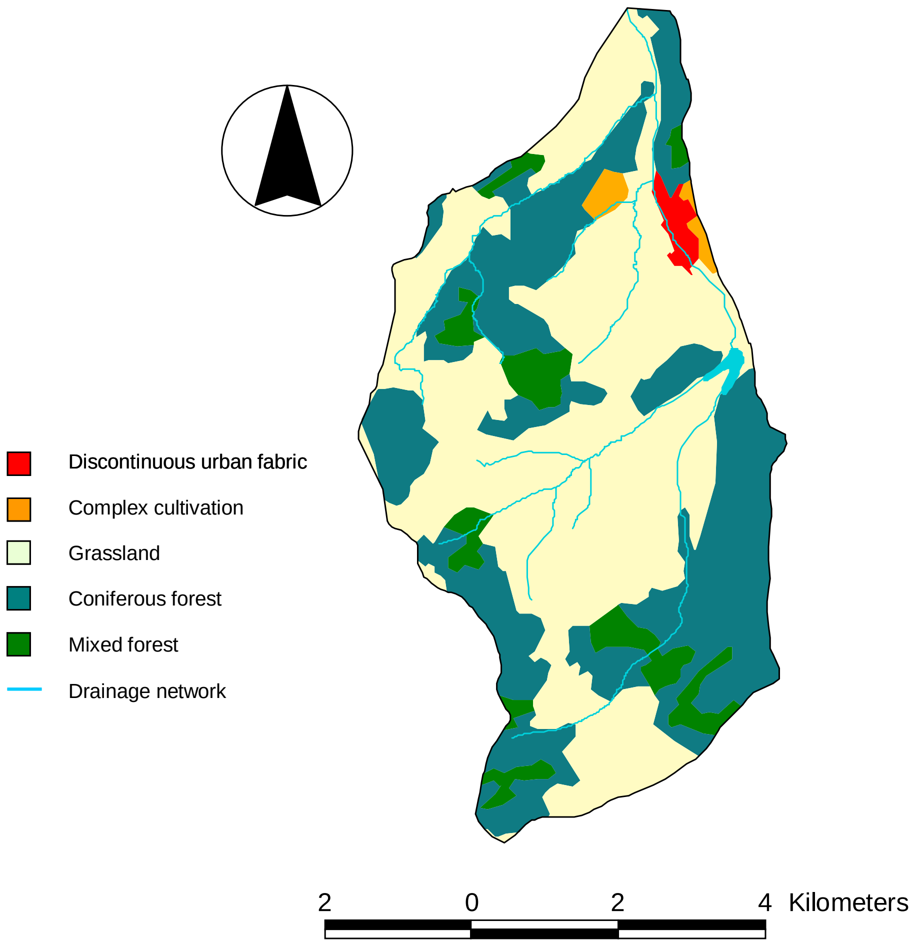


Figure 18: Land use of the Fichtenberger Rot catchment.

2.4 Roggelshäuser Bach

The Roggelshäuser Bach catchment is situated in the Hohenlohe plain in north-eastern Baden-Württemberg. With 13.3 km^2 it is the smallest of the selected basins. The elevation is ranging from 271 at the outlet to 484 meters above sea level. The mean altitude and the mean slope are calculated to 428 m a.s.l., respectively to 10.0%. An overview of the catchment is given in figure 19.

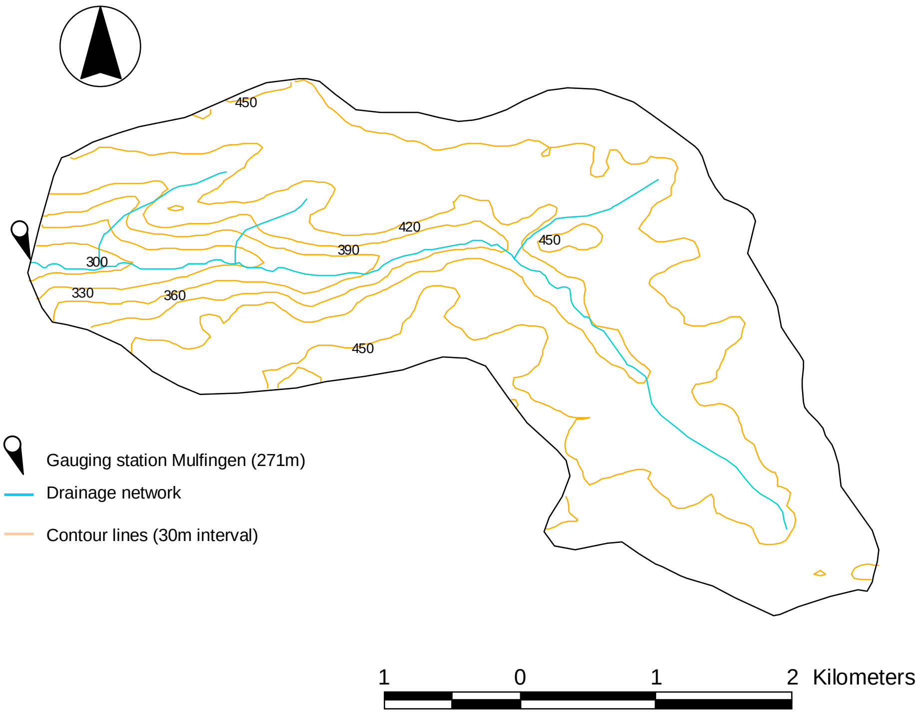


Figure 19: Overview of the Roggelshäuser Bach catchment, Baden-Württemberg.

2.4.1 Climate and hydrology

The seasonal variation of precipitation at the nearby station Schrozberg (455m a.s.l.) for the period 1961-1990, which is displayed in figure 20, is dominated by rising amounts during the summer. Again, June is the month with the highest precipitation, whereas only 51mm are reached in September. With an annual sum of 830.8 mm, the catchment is not only the smallest but also the driest of the selected basins.

The variability of the air temperature at the station Crailsheim-Ingersheim (417m a.s.l.) 30 km to the south-east of the catchment, is illustrated in figure 21, showing only January with a mean monthly temperature below the freezing point. The mean annual air temperature averages 8.0°C.

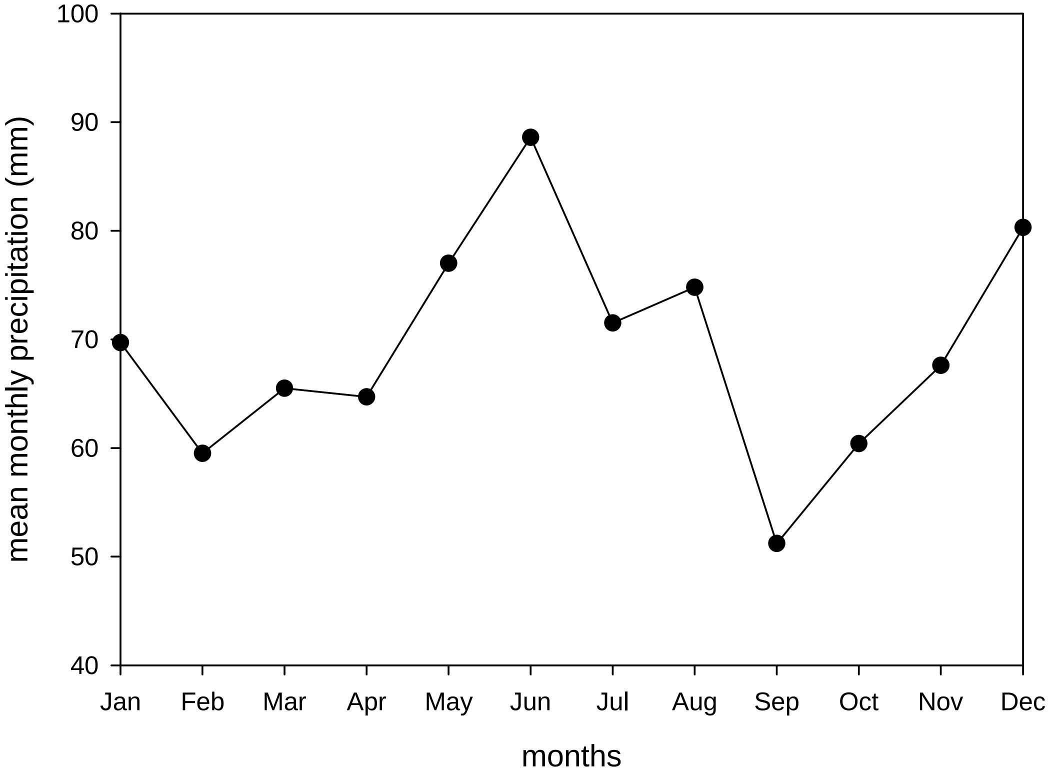


Figure 20: Mean monthly precipitation of the station Schrozberg for the period 1961-1990.

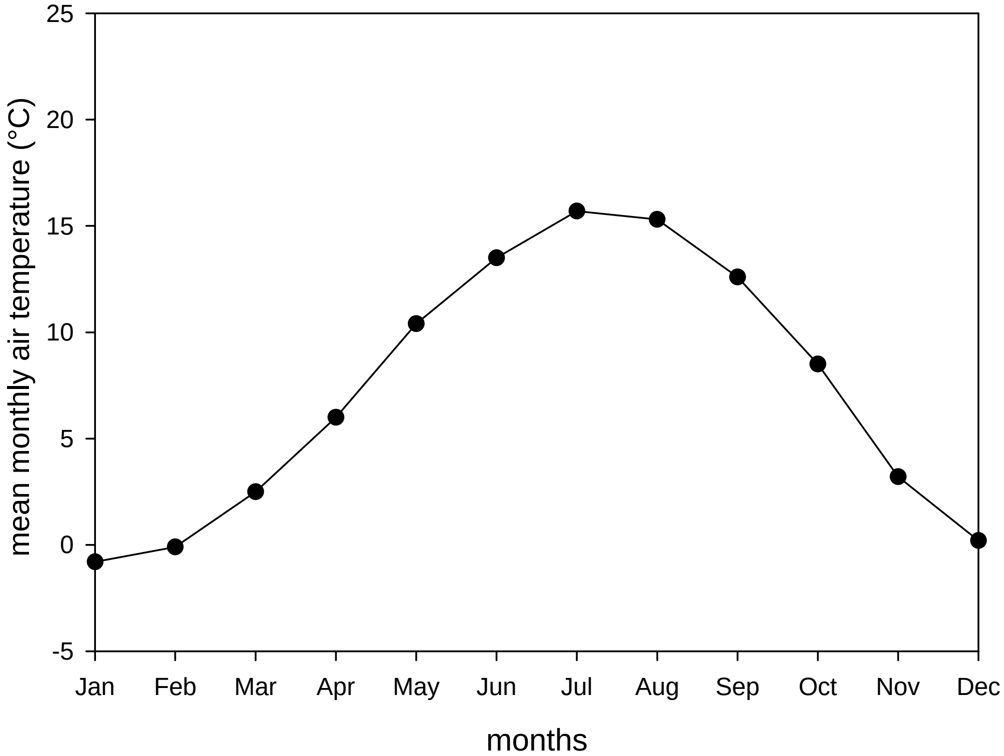


Figure 21: Mean monthly air temperature of the station Crailsheim-Ingersheim for the period 1961-1990.

The regime for the period 1966-1989 is presented in figure 22. It is characterised by an increase in runoffs during winter season with an additional snowmelt peak in April/May and a minimum in late summer so that it can be classified as a moderate nivo-pluvial regime.

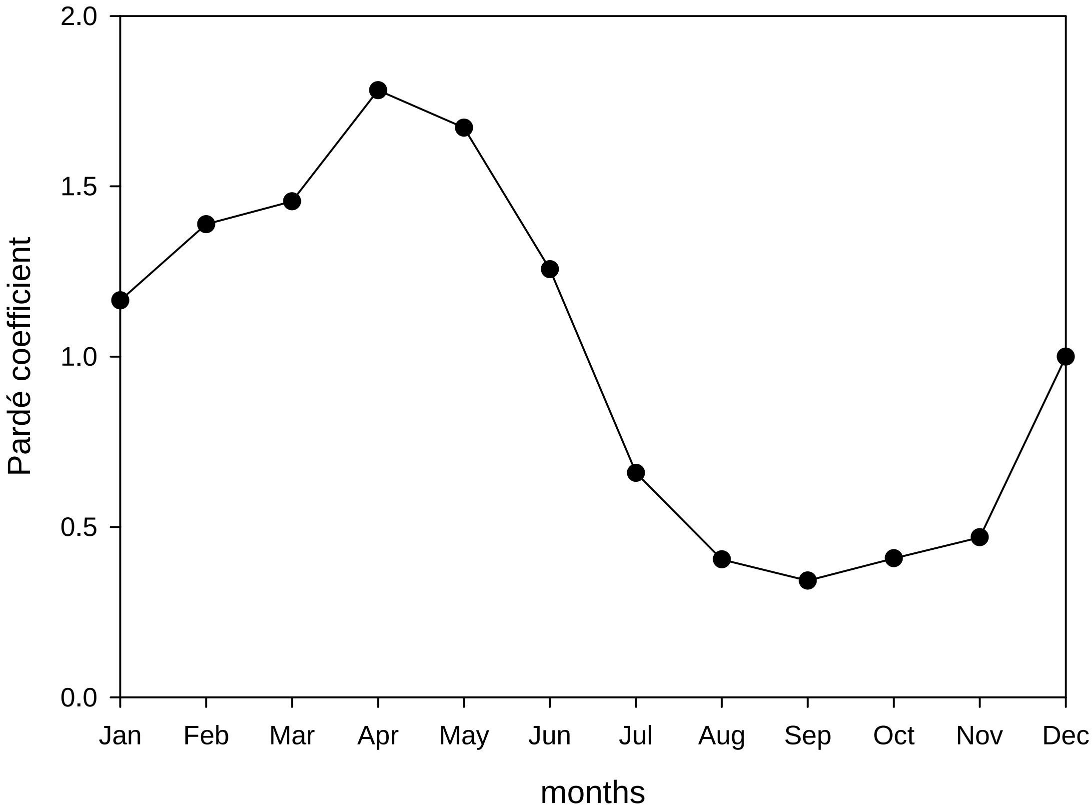


Figure 22: Runoff regime of Roggelshäuser Bach for the period 1966-1989.

In table 4, runoff statistics of Roggelshäuser Bach at the gauging station Mulfingen for the period 1961-1989 are given.

According to WABOA 2007, the mean annual actual evapotranspiration is calculated to 498 mm.

Table 4: Runoff characteristics of Roggelshäuser Bach for the period 1961-1982

	MQ (m^3/s)	MHQ (m^3/s)	MLQ (m^3/s)	HHQ (m^3/s)	LLQ (m^3/s)
Roggelshäuser Bach	0.21	2.38	0.02	4.952	0.003

2.4.2 Geology, pedology and land use

The erosive work of the Roggelshäuser Bach has produced a typical outcrop of the south German escarpments. Starting at the lower Keuper, with an estimated conductivity of 10^{-5} m/s, the Roggelshäuser Bach afterwards flows through the karstified upper Muschelkalk (10^{-4} m/s) and the fissured middle Muschelkalk, where a conductivity of 10^{-6} to 10^{-7} m/s (WaBoA 2007) is reached. Near the outlet, a spot of moderate permeable lower Muschelkalk is present (figure 23).

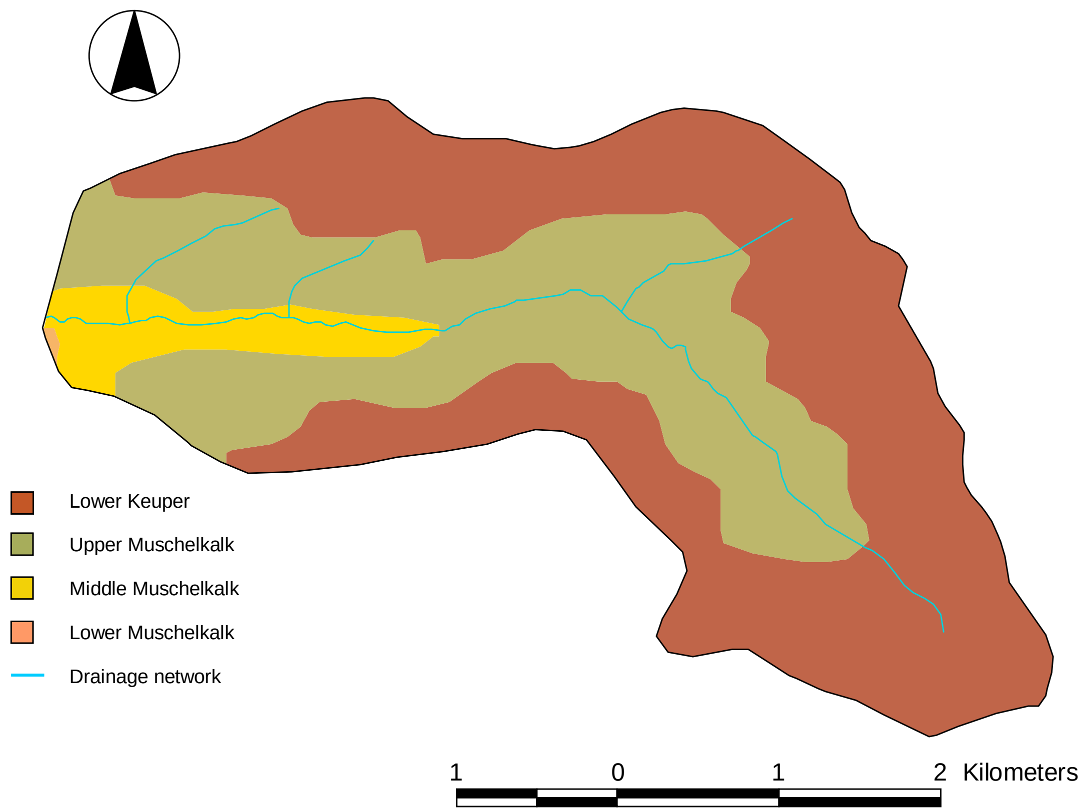


Figure 23: Geology of the Roggelshäuser Bach catchment.

Due to the different Triassic basements the soils vary, too. The lower Keuper is mainly covered with pelosols and gleyic cambisols, whereas the Muschelkalk is dominated by rendzinas. Furthermore, water meadows with gleyic soils occur at the valley bottom (WaBoA 2007).

The land use is dominated by agriculture, with 57% of the area used as arable land and 11% covered with complex cultivations. The remaining area is mainly covered with forests (27%), and isolated spots of grassland and settlements.

2.5. Eyach

The Eyach catchment, including an area of 29.8 km^2 is located in the northern Black Forest about 35 km south of Karlsruhe. The elevation ranges from 476 m at the outlet up to 956 m a.s.l., resulting in a mean elevation of 790 m and a mean slope of 22.0% (WABOA2007). An overview of the Eyach catchment, including the meteorological station at Bad Wildbad-Sommberg, is given in figure 24.

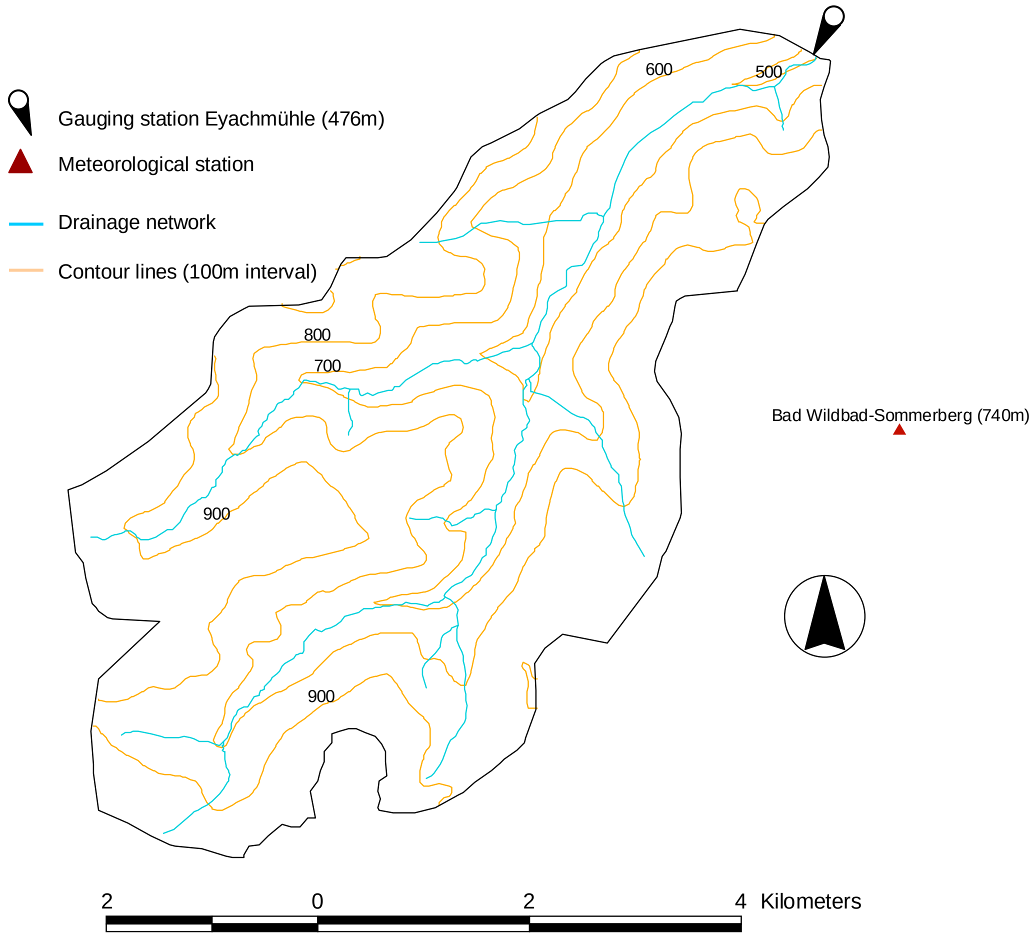


Figure 24: Overview of the Eyach catchment, Baden-Württemberg.

2.5.1 Climate and hydrology

The precipitation regime, illustrated in Figure 25 for the period 1961-1990, is dominated by rising amounts during the winter, adding up to a mean of 1385 mm per year. This is caused by a decrease of the condensation level, resulting in an enhancement of the orographic effect. The maximum amounts of 143 mm are reached in December, whereas September delivers 83 mm only.

The seasonal variability of the air temperature, displayed in figure 25, shows temperatures below freezing point in January and February and the highest mean monthly value with

15.7°C in July. The mean annual temperature averages 7.2°C and the mean actual evaporation is calculated to 574.3 mm per year

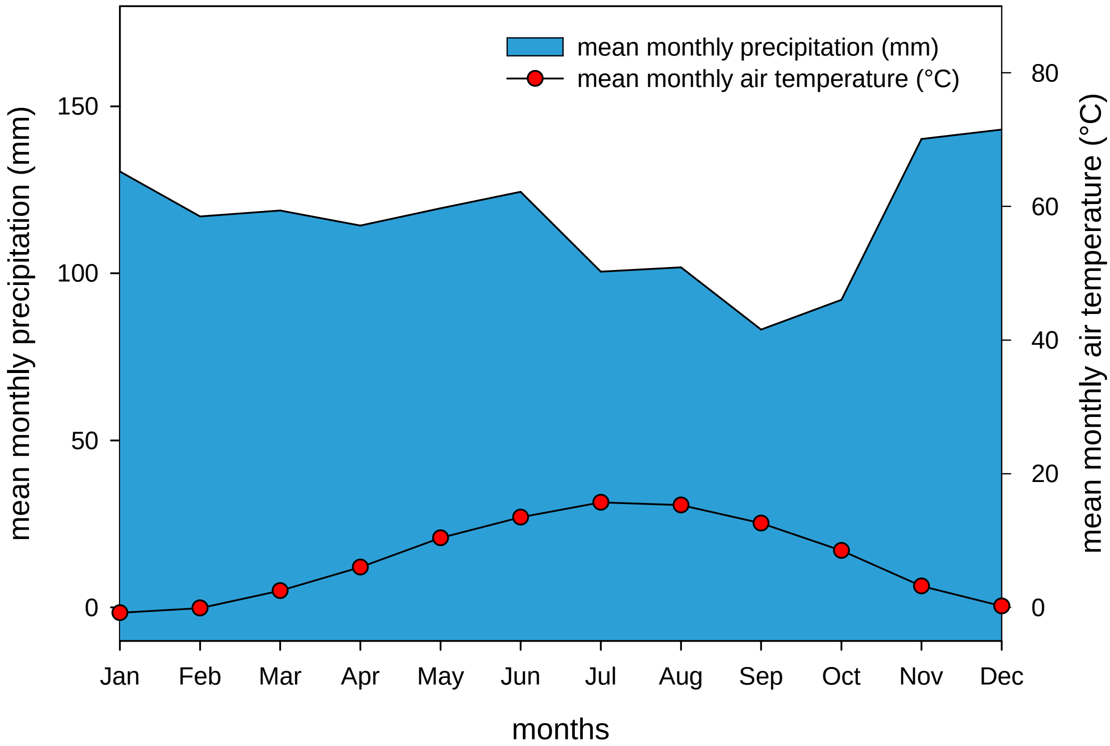


Figure 25: Climate chart of the station Bad Wildbad-Sommerberg for the period 1961-1990.

The daily discharge data at the outlet Eyachmühle for the period 1974-1990 was used to calculate some standard runoff statistics, which are listed in table 5. The runoff regime is presented in figure 26. It is characterised by a peak in February, which is caused by high precipitation amounts and a second peak in April, resulting from snowmelt. According to this, the regime can be classified as balanced nivo-pluvial.

Table 5: Runoff characteristics of the Eyach for the period 1974-1990

	MQ (m^3/s)	MHQ (m^3/s)	MLQ (m^3/s)	HHQ (m^3/s)	LLQ (m^3/s)
Eyach	0.85	5.73	0.33	10.86	0.15

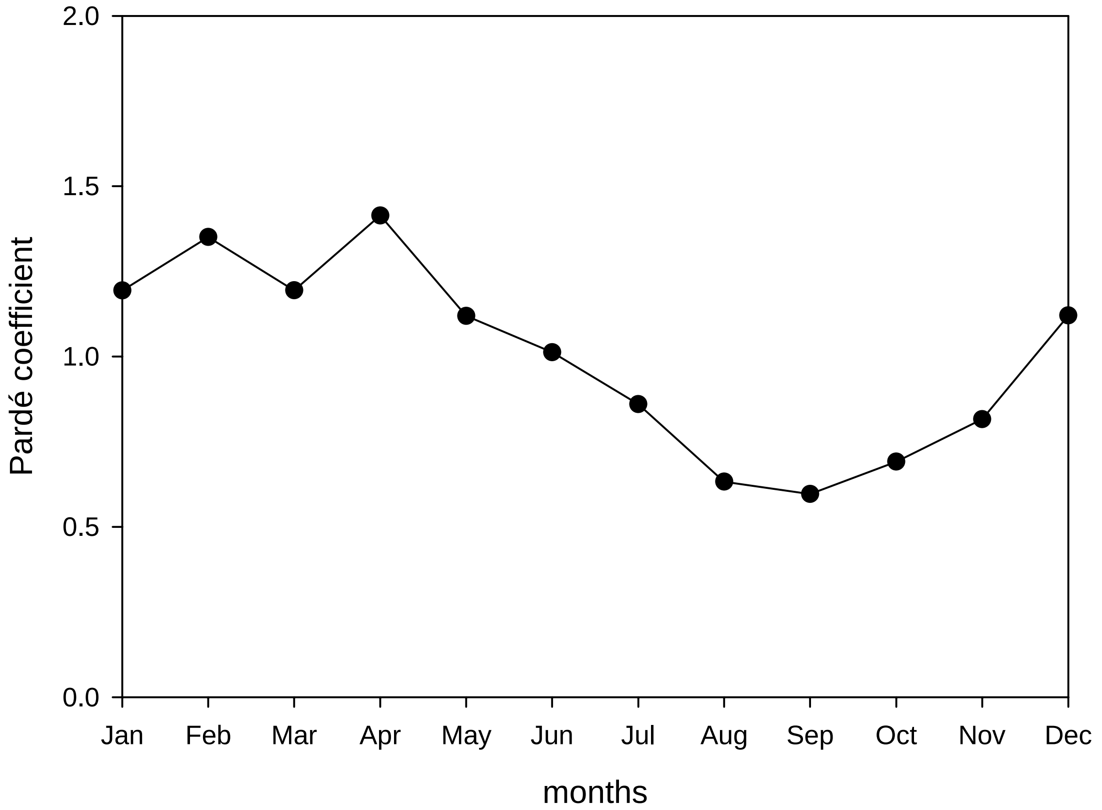


Figure 26: Runoff regime of Eyach for the period 1974-1990.

2.5.2 Geology, pedology and land use

The basin is situated within the region of the middle and lower Buntsandstein which can be characterised by conductivities about 10^{-5} m/s (WABOA 2007). Isolated, the granitic basement and Permian Rotliegend, with conductivities between 10^{-6} and 10^{-7} m/s can be found (CASPER 2002, WABOA 2007).

The Buntsandstein is mainly covered with podzols and podzolic cambisols. Additionally, there are also gleyic soils at the upper catchment. The complete area of the basin is used as forest, with 78% of the area mainly consists of coniferous forest (WABOA 2007).

2.6 Conclusion

Considering the availability of discharge, precipitation, and air temperature data the presented catchments were selected to gain a maximum variability of essential characteristics to test the approach under various conditions. The basin's area was determined by the model run-time and the missing of an overland routing module.

Table 6 shows a compilation of different topographic parameters for the five catchments. The area varies from 13.3 km^2 of the Roggelshäuser Bach up to 40.3 km^2 of the Haslach. The Zastlerbach catchment with the highest altitude difference of 945 m is steepest of the chosen basins, whereas a slope of just 5.4% is found in the Haslach catchment.

Table 6: Topographic characteristics of the five catchments

	Zastlerbach	Fichtenberger Rot	Haslach	Roggelshäuser Bach	Eyach
area (km^2)	17.9	17.3	40.3	13.3	29.8
mean elevation (m a.s.l.)	1053	492	688	428	790
max. altitude difference (m)	945	122	142	213	480
mean slope (%)	35.2	8.3	5.4	10.0	22.0

The climatic and hydrologic characteristics, listed in table 7, show a difference of almost 1000mm on the annual precipitation sum between the Zastlerbach and the Roggelshäuser Bach due to the close connection between precipitation and elevation.

Table 7: Climatic and hydrological characteristics of the five catchments

	Zastlerbach	Fichtenberger Rot	Haslach	Roggelshäuser Bach	Eyach
mean annual precipitation (mm)	1814	1188	1024	831	1385
mean annual air temperature (°C)	5.6	7.7	7.8	8.0	7.2
annual actual evapo-transpiration (mm)	531	560	611	498	574
mean specific discharge (mm/a)	1153	529	562	496	898

In contrast to the precipitation, the highest temperatures are reached in the Roggelshäuser Bach basin, whereas only 5.6°C are measured in Zastlerbach. The calculated actual annual evapotranspiration rates result from the interaction of air temperature, land use and subsurface characteristics (listed in table 8). The highest amounts are reached in the Haslach catchment, which is colder than Roggelshäuser Bach but where also 44% of the area is used as forest, whereas only 27% of Roggelshäuser Bach is covered with forest and karstified rocks occur. The values of the other catchments lie within this range, due to relative low temperatures, good storage qualities and heavily forested regions.

The mean annual discharges for the respective periods were normalised to specific discharges to improve the comparability. The calculated specific discharges are closely connected to the annual precipitation sum and secondary influenced by the slope and the hydraulic conductivity of the underlying rock. The highest discharges are reached in the very steep Zastlerbach catchment where poorly conductive metamorphic rocks are present and the precipitation sum exceeds 1800 mm per year. The Eyach and the Fichtenberger Rot basins have comparable precipitation sums and subsurface characteristics. Nevertheless the specific discharge of the Eyach is nearly twice as much of the one of Fichtenberger Rot. This could result from the steeper hills in the Eyach catchment that influence the runoff generation. The lowest values are delivered by the Roggelshäuser Bach where highly conductive karstified rocks discharge the little precipitation into the depth.

Table 8: Geologic and pedologic characteristics of the five catchments

	Zastlerbach	Fichtenberger Rot	Haslach	Roggelshäuser Bach	Eyach
dominant geology	metamorphic rock	sandstone + marl	quaternary material	limestone + sandstone	sandstone
estimated hydr. conductivity (m/s)	10^{-6} to 10^{-7}	10^{-5}	10^{-3} to 10^{-6}	10^{-4} to 10^{-5}	10^{-5}
dominant soil type	cambisol	podzolic cambisol	luvisol	rendzina + pelosol	podzol
dominant land use + ratio	forest (90%)	forest (44%)	grassland (54%)	arable land (57%)	forest (100%)

3. Methods and basics

3.1 Stream networks

3.1.1 Stream networks as alternative to runoff data

Spatially distributed data is essential to examine the performance of spatially distributed models (e.g. GRAYSON ET AL. 2002). The examples presented in chapter 1.2.2 have shown that it is possible to use those data for model calibration and evaluation, even in macro scale catchments. But it is crucial that the used spatial data is representative to the main processes and the corresponding parameters that control the hydrologic response of the catchment.

According to RODRIGUEZ-ITURBE & RINALDO 1997 reporting that a '[…] *drainage network itself may be viewed as a reflection of the runoff-producing mechanisms occurring in a basin'*, stream networks can meet those conditions.

A second condition arises from the fact that in ungauged basins, per definition, data, especially spatially distributed data, is rare. Drainage networks derived from maps or remote sensing are easily accessible (REFSGAARD 2001) and thus suitable for applications in ungauged catchments.

It seems that stream networks could be an alternative to runoff data in ungauged basins for model calibration. Hence, processes affecting the initiation and development of stream networks should accordingly be elaborated.

3.1.2 Initiation of stream networks

Different approaches to examine the evolution, initiation, and description of drainage networks have been applied in the past (DE VRIES 1994). SHREVE 1966 presented a concept describing the development of stream networks as a stochastic process, whereas RODRIGUEZ-ITURBE ET AL. 1992 chose an alternative thermodynamic approach, in which the network evolution is governed by the principle of minimum energy expenditure. The most famous review on stream networks derives from HORTON 1945. It includes a process-based approach on drainage network development based on overland flow erosion and a large number of indices to characterise drainage networks in catchments, e.g. drainage density.

The drainage density D_d (1/m) is a tool to characterise the drainage ability of different catchments. It is defined as:

$$D_d = \frac{\sum L}{A} \qquad \text{(Eq. 1)}$$

where $\sum L$ is total length of streams (m), which is scale dependent and A is the catchment's area (m^2). Besides the influence of precipitation and relief on the drainage density, HORTON 1945 identified the infiltration capacity of the soil and the resistivity to erosion as additional factors. CARLSTON 1963 reported that the drainage density is controlled by the transmissivity T (m^2/s) and presents an empirical relationship gained from the use of a groundwater model:

$$D_d{}^2 = \frac{k_D}{T} \qquad \text{(Eq. 2)}$$

where k_d (1/s) is a coefficient representing the quotient of groundwater recharge and water table height.

Besides the deterministic concept of HORTON 1945 in which overland flow is seen as the dominant feature, DUNNE 1969 presented additional processes, especially groundwater outflow influencing the initiation of stream networks (see also LOBKOVSKY ET AL. 2007). Groundwater sapping is defined as groundwater outflow (springs or diffusive) eroding material. This is identified as a key process in drainage network development in humid areas. Groundwater flow lines converge at the heads of valleys. This is characteristic for increased hydraulic gradients and result in the development of valley networks (figure 27).

Therefore, in catchments where almost all precipitation excess recharges the groundwater, the stream channels can be classified as outcrops of the groundwater flow system. They represent the interface between groundwater flow and associated surface drainage system. The evolution of the channel network is dominated by subsurface permeability, climate and large scale topography (DE VRIES 1994, DE VRIES, 1995, TROCH ET AL. 1995).

According to the presented theory and equivalent to the variable source area concept of HEWLETT & HIBBERT 1967, DE VRIES 1995 described the initiation of stream network as a self-organising process controlled by seasonal meteorological variations resulting in groundwater level fluctuations. Those fluctuations generate an expanding and contracting of

the stream network with a changing number of participating draining streams (ROBERTS & ARCHIBOLD 1978). An idealised illustration of the concept is displayed in figure 28.
The presented theories indicate that the initiation of stream networks is mainly controlled by processes taking place in the saturated zone. Thus, the applied model should be able to describe the properties of the subsurface, e.g. transmissivity, with as much accuracy as possible.

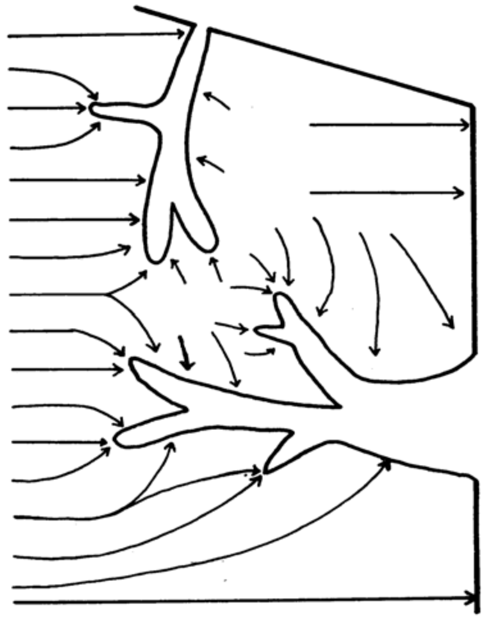


Figure 27: Development of valley network by headward spring-sapping. Flow lines are displayed as black arrows (DUNNE 1969).

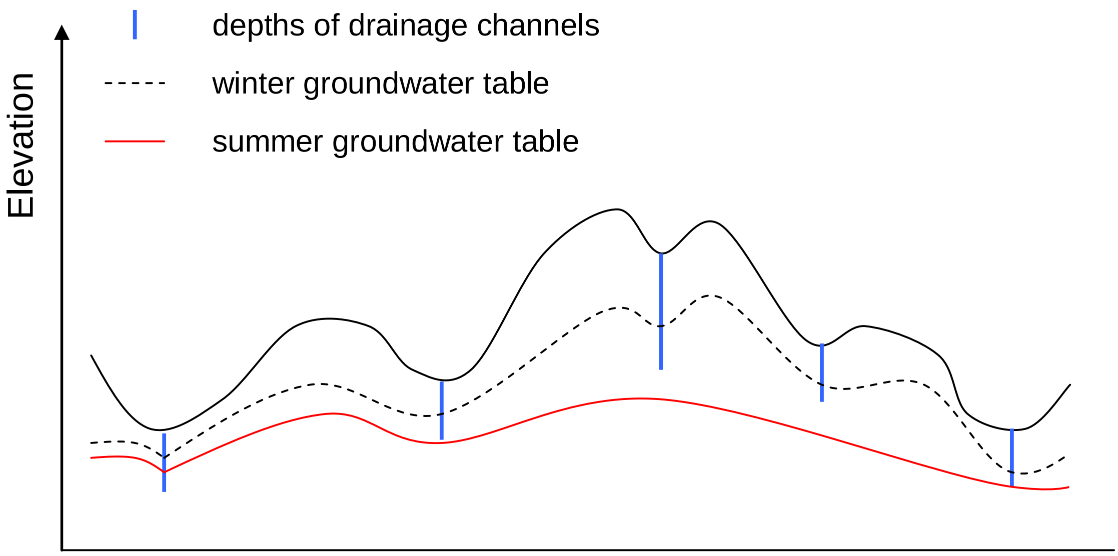


Figure 28: Idealised illustration of the seasonal expanding of the drainage network (modified according to DE VRIES 1995).

3.2. Model description

The conceptual hillslope model, Hill-Vi, was developed by WEILER & MCDONNELL 2004 to test the benefit of virtual experiments to hillslope hydrology. Subsequently, it has been modified with respect to the issues of different projects including nutrient flushing (WEILER & MCDONNELL 2006), the effects of preferential flow networks (WEILER & MCDONNELL 2007) and identification of runoff processes (MCGUIRE ET AL. 2007). For this thesis, further modifications were made with special considerations to processes in the saturated zone, according to the concepts presented in chapter 3.1.2. To assure fast and robust models runs, which is necessary to model five different catchments, processes taking place in the unsaturated zone were removed or transferred to the saturated zone.

3.2.1 Basic concept

The used version can be described as a spatial explicit model based on a storage representing the complete subsurface zone for each grid cell. According to WEILER & MCDONNELL 2004, the water volume in the saturated zones V_{sat} (m^3) is given by:

$$V_{sat} = W\,A\,n_d \qquad \text{(Eq. 3)}$$

with W the water table depth (m), A the area of the grid cell (m^2) and n_d the drainable porosity (analogous to specific yield).

Due to data of different field observations, WEILER & MCDONNELL 2006 implemented a depth function for the drainable porosity:

$$n_d(z) = n_0 \exp\left(-\frac{z}{b}\right) \qquad \text{(Eq. 4)}$$

where n_0 is the drainable porosity at the soil surface, b the decay coefficient, and z the depth into the soil profile (positive downward).

Similar to the power law presented by RUPP & SELKER 2005 a decline of the hydraulic conductivity k_s (m/s) with depth was realized, representing both the compaction of the soil and the transfer to the less permeable bedrock:

$$k_s(z) = k_0 \exp\left(-\frac{z}{b}\right) + k_c \qquad \text{(Eq. 5)}$$

with k_0 the saturated hydraulic conductivity at the soil surface (m/s) and k_c the constant saturated hydraulic conductivity of the bedrock (m/s).

It is assumed that the decline is caused by the same effects also reducing the drainable porosity so that no distinction must be made in the determination of the coefficients.

3.2.2 Water balance of the saturated zone

The water balance of the saturated zone is calculated by the input of precipitation and snowmelt, the lateral in- and outflow as well as the actual evapotranspiration. An overview of the implemented water fluxes is given in figure 29. When the water table rises above the ground level, all water excess is defined as overland flow and the corresponding cell as a stream cell.

Due to the occurrence of regimes controlled by snow melt processes (chapter 2), a snow melt routine was implemented in Hill-Vi. The snowmelt module is equivalent to the degree-day approach of the HBV model (e.g. BERGSTRÖM 1975, 1995). If the air temperature falls below the threshold temperature TT (°C), the complete precipitation accumulates as snow. If, however, the threshold temperature is exceeded, snowmelt takes place:

$$MELT = CF_{Max}\ (T_a - TT) \qquad \text{(Eq. 6)}$$

with the amount of snowmelt $MELT$ (mm/d), the actual air temperature T_a (°C) and the degree-day factor CF_{Max} (mm/(°C d)). The snow storage retains the melt water until it exceeds a certain portion of the snow water equivalent. When the actual air temperature falls below the threshold temperature the melt water refreezes:

$$REFR = CFR\ CF_{Max}\ (TT - T_a) \qquad \text{(Eq. 7)}$$

with the amount of refrozen water $REFR$ (mm/d) and the refreezing coefficient CFR.

Within the saturated zone the Dupuit-Forchheimer assumption (FREEZE & CHERRY 1979):

$$q(t) = T(t)\beta\, w \qquad \text{(Eq. 8)}$$

where T the transmissivity (m^2/s), β the water table slope and w the width of the flow (m) are used to calculate lateral subsurface flow. Based on this concept, Hill-Vi calculates the downslope routing of the subsurface flow by a grid cell by grid cell approach, which was introduced by WIGMOSTA & LETTENMAIER 1999.

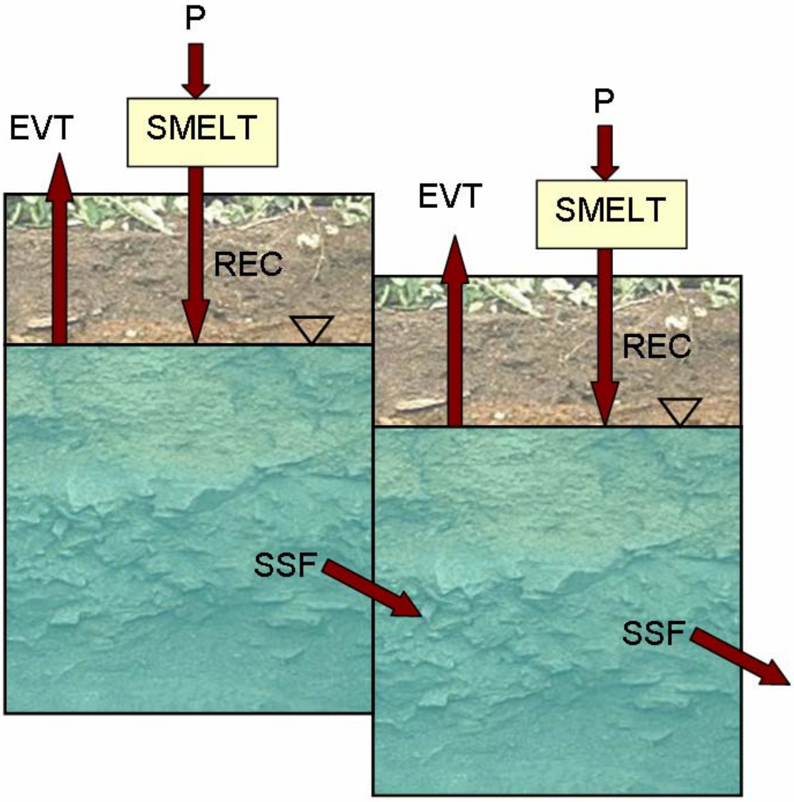


Figure 29: Overview of the implemented water fluxes: Snowmelt (SMELT), evapotranspiration (EVP), recharge to saturated zone (REC) and lateral flow (SSF).

As mentioned above, the unsaturated zone had to be removed to increase the speed and stability of the model runs. As a result, the actual evapotranspiration E_{act} (m) cannot be controlled by the relative water content in the soil any longer. Therefore, a correlation between groundwater table depth and the potential evapotranspiration E_{pot} (m) was established:

$$E_{act}(t) = E_{pot}\left(\frac{1}{2}\exp\left(-\frac{z}{5}\right)+\frac{1}{2}\right) \quad \text{(Eq. 9)}$$

Due to the lack of climatic air moisture data, the approach of THORNTHWAITE 1948, which requires the monthly average air temperature $\overline{T}$ (°C), the number of days n in the month, the average sunshine duration S_0 (h), the heat index I (°C), and the empirical coefficient a, is used to estimate the monthly potential evapotranspiration:

$$E_{pot} = 0.533n\frac{S_0}{12}\left(\frac{10\overline{T}}{I}\right)^a \quad \text{(Eq. 10)}$$

with

$$I = \sum_{1}^{12}\left(\frac{\overline{T}}{5}\right)^{1.514} \qquad \text{(Eq. 11)}$$

and

$$a = (0.0675I^{3} - 7.71I^{2} + 1792\ I + 49239)10^{-5} \qquad \text{(Eq. 12)}$$

As a result of the implemented modifications, the number of the calibration parameters, which are listed in table 9 was reduced to five. The parameters of the snow melt module parameter were estimated according to literature.

Table 9: Calibration parameter

Parameter	Explanation	Unit
Z	total depth of the subsurface	$[m]$
b	decay coefficient	[-]
n_0	drainable porosity at the soil surface	[-]
k_0	saturated hydraulic conductivity at the soil surface	$[m/s]$
k_c	constant saturated hydraulic conductivity of the bedrock	$[m/s]$

3.3 Objective functions

To calibrate the presented parameters, a comparison between the modelled spatial distribution of stream cells and a reference map is needed. The stream network of the WaBoA that derived from a digitalisation of streams exceeding a length of 500 m in the topographic map of Baden-Württemberg (1:200000) was available as a shapefile. This shapefile was transformed into a binary raster file with a resolution of 50 to 50 meters equivalent to the resolution of the binary model output differing in stream cells and no-stream cells.

Different comparison methods were tested to achieve the most sensitive calibration. Finally, the Kappa goodness-of-fit statistics, which was introduced by COHEN 1960 to compare independent psychiatric diagnoses, turned out to be the most adequate. The Kappa value K is the measure of actual agreement compared to a chance agreement. Both derived from an error matrix (table 10). The discrete multivariate technique has been widely used in different disciplines, e.g. in the field of remote sensing to measure the agreement of two maps (e.g. MONSERUD & LEEMANS 1992, CONGALTON & GREEN 1999).

Table 10: Error matrix used for Kappa statistics

		Reference			
		1	2	k	$p_{i\bullet}$
Model	1	p_{11}	p_{12}	p_{1k}	$p_{1\bullet}$
	2	p_{21}	p_{22}	p_{2k}	$p_{2\bullet}$
	k	p_{k1}	p_{k2}	p_{kk}	$p_{k\bullet}$
	$p_{\bullet j}$	$p_{\bullet 1}$	$p_{\bullet 2}$	$p_{\bullet k}$	1

Assuming two maps with two categories (water and no-water), there are agreements of omission (no-water in both maps) and agreements of commission (water in both maps), adding up to the proportion of agreement p_a:

$$p_a = \sum_{i=1}^{k} p_{ii} \tag{Eq. 13}$$

Additionally, a chance agreement indicated by the product of the row and column totals is determined from the error matrix:

$$p_c = \sum_{i=1}^{k} p_{i\bullet}\, p_{\bullet j} \qquad \text{(Eq. 14)}$$

Given the premise of a multinomial sampling model, the maximum likelihood estimate of Kappa can be calculated:

$$\mathrm{K} = \frac{p_a - p_c}{1 - p_c} \qquad \text{(Eq. 15)}$$

Positive values indicate a meaningful classification with an agreement significantly better than a random one. When the two maps are identical, Kappa takes the value of $\mathrm{K} = 1$ (CONGALTON & GREEN 1999).

Besides the simple calculation, Kappa statistics have the advantage of a standardised result that makes a simple comparison possible. However, it is essential that the compared results have the same number of categories that influence the Kappa value (MACLURE & WILLET 1987). LANDIS & KOCH 1977 presented a compilation of possible evaluations of Kappa statistics (table 11).

Table 11: Different degrees of agreement for the Kappa statistics (according to LANDIS & KOCH 1977)

Kappa statistic	Strength of agreement
< 0.00	poor
0.01 - 0.20	slight
0.21 – 0.40	fair
0.41 – 0.60	moderate
0.61 – 0.80	substanial
0.81 – 1.00	almost perfect

However, those subjective thresholds must be approached with caution and can not be transferred to any other application. For example, VIERA & GARRETTT 2005 report that *'for* *rare findings, very low values of kappa may not necessarily reflect low rates of overall agreement*'. They present a case where despite a proportion of agreement of 0.85 the Kappa value is only calculated to 0.04 due to the superposing effects of the accordingly increased chance agreement.

The model efficiency R_{eff} after NASH & SUTCLIFFE 1970 is used to test the performance of the calibrated parameters during the evaluation period. It is calculated as:

$$R_{eff} = 1 - \frac{\sum (Q_{obs}(t) - Q_{sim}(t))^2}{\sum (Q_{obs}(t) - \overline{Q_{obs}})^2} \qquad \text{(Eq. 16)}$$

The values of R_{eff} range from minus infinity to one, reached with a perfect fit. Values below zero indicate that the mean discharge of the used time series would be a better predictor than the model. The squared derivations result in stronger weighting of high discharges whereas low flows are underrepresented. Through a logarithmic transformation of the runoff, more importance to the low flow situations can be attached (LEGATES & MCCABE JR. 1999):

4. Simulating stream networks

In order to take the influence of the climate on stream network development into consideration, temperature and precipitation data of the climatic reference period 1961-1990, which is presented in chapter 2, is used as input for the modelling. A 30-day running average is applied to the data with the intention of reaching a smooth regime. After a one-year warm up period the stream networks were simulated in daily time steps based on a 50 to 50 meter digital elevation model.

As mentioned above, the initiation of stream networks is controlled by seasonal metrological variations, which makes it necessary to select a specific season for the comparison with the observed stream networks on the map. Assuming that only perennial streams that also carry water in the summer were listed in maps, the arithmetic mean of the daily Kappa values, which were calculated for June, July and August, was chosen to evaluate the efficiency of the parameter set.

4.1 Sensitivity analysis and calibration

In each catchment 1000 Monte-Carlo runs were carried out to identify the best parameter sets. The upper and lower boundaries are preset under consideration of typical values. In the following, the results of those runs are presented with a special focus on the sensitivity of the parameter.

The parameter combinations that achieved the best agreements according to the Kappa statistics are listed in table 12. Additionally, the transmissivity T_m (m^2/s) and the average of the drainable porosity over the depth n_{avg}, both dependent on the decay coefficient, are calculated to enable a better evaluation between the individual catchments.

The highest agreement is reached in the Zastlerbach catchment, followed by the Eyach and the Roggelshäuser Bach whereas in the Haslach and Fichtenberger Rot catchment values of only 0.164 and respectively 0.140 respectively were calculated. The Zastlerbach is also identified as the catchment with the lowest transmissivity and average drainable porosity. The Roggelshäuser Bach and the Eyach basin have, analogue to the Kappa statistics, similar average drainable porosities. However, a higher transmissivity, nearly identical to the one in the Haslach basin, is calibrated for the Roggelshäuser Bach. The average drainable porosity in the Haslach catchment was calibrated to 0.073, which is comparable to what was computed for the Fichtenberger Rot.

Table 12: Best parameter sets after calibration

	Zastlerbach	Fichtenberger Rot	Haslach	Roggelshäuser Bach	Eyach
Z (m)	10.50	10.00	8.62	8.41	11.60
b	1.13	2.56	2.20	2.30	2.18
n_0	0.34	0.28	0.29	0.31	0.44
k_0 (m/d)	2.00	1.66	2.59	2.53	1.83
k_c (m/d)	7.9 E-4	1.2E-3	5.2E-4	5.1E-4	2.49E-3
K	0.205	0.140	0.164	0.196	0.199
T_m (m^2/s)	6.29E-4	1.16E-3	1.55E-3	1.57E-3	1.10E-3
n_{avg}	0.037	0.070	0.073	0.083	0.082

The results of the variations of the parameter are given as dotty plots. Each black point represents one model run with the actual parameter value against the result of the Kappa statistic. The combination with the best result is indicated as yellow triangle.

The analysis of the total depth of the subsurface shows a general tendency to lower values for all catchments (figure 30). The combinations with the highest Kappa values all lie below a depth of about 12 m. It is remarkable that the sensitivity of parameter variations is different for the individual catchments. Especially in the Roggelshäuser Bach catchment the sets with high efficiency are close to each other at the very end of the lower boundary. However, it seems that for the remaining catchments the range of combinations with high Kappa values is much wider.

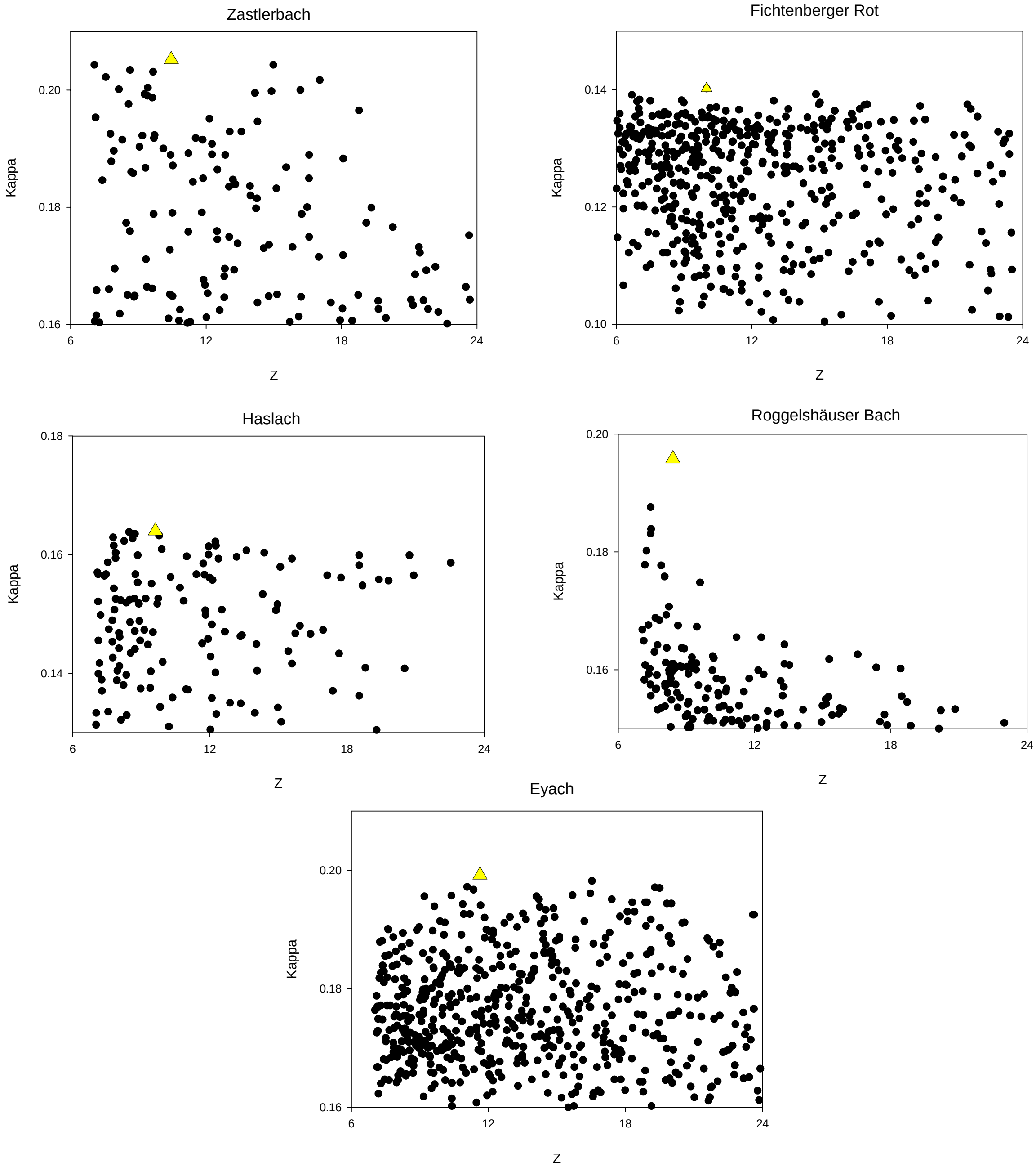


Figure 30: Total depths against the objective function of Kappa. The best values are represented by yellow triangles.

A different situation arises by analysing the sensitivity of the decay coefficient on the model efficiency (figure 31). A clear structure can be seen in the Zastler catchment, where the rising of the Kappa statistic is related to a decrease in the decay coefficient. In the Eyach catchment a pattern with a distinct maximum can be observed. However, it is difficult to identify a significant behaviour in the remaining basins. At least it seems that there are areas around $b=2$ and $b=3.5$ where increased efficiencies can be distinguished.

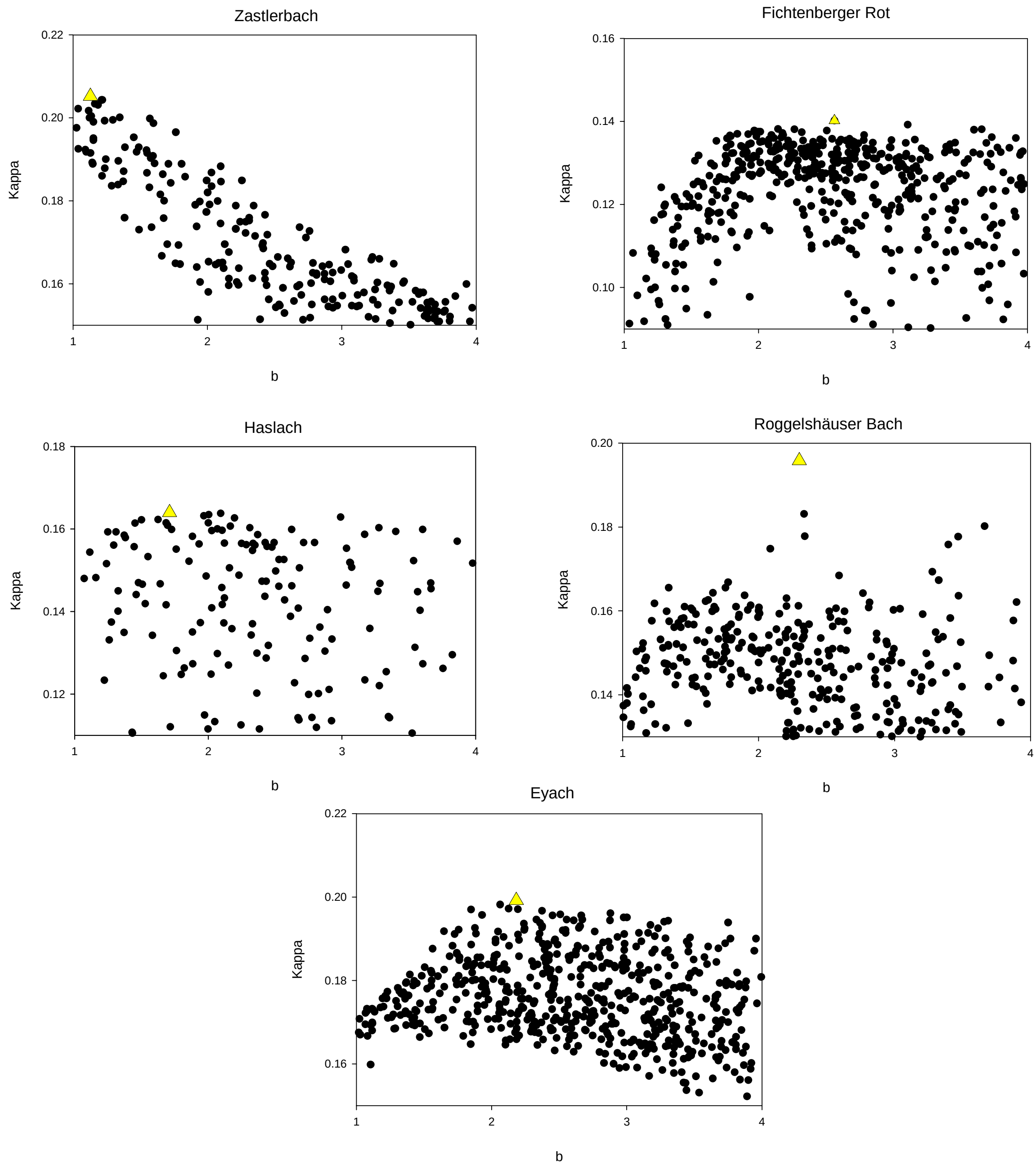


Figure 31: Decay coefficients against the objective function of Kappa. The best values are represented by yellow triangles.

The plots of saturated hydraulic conductivity at the soil surface against the objective function of Kappa (figure 32) indicate that model performance increases with decreasing conductivities, which is especially distinctive in Zastlerbach, the Fichtenberger Rot, and the Eyach. It seems that partially the lower bound constrains further decreasing of the conductivity. Although the Roggelshäuser Bach follows the general tendency, the best performance is reached with relative high hydraulic conductivity. Again, it is difficult to identify a significant pattern for the Haslach catchment.

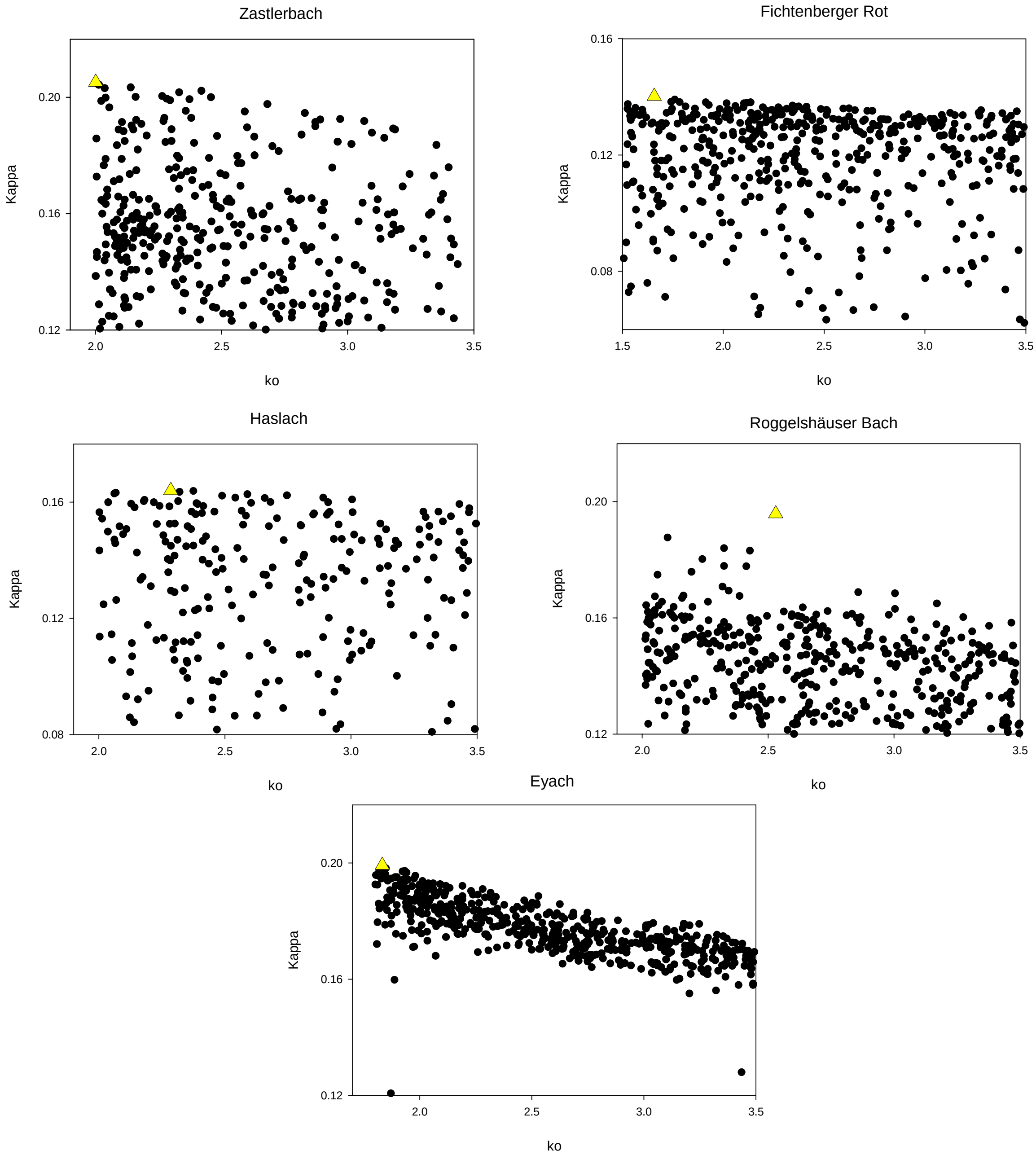


Figure 32: Hydraulic conductivities at the soil surface against the objective function of Kappa. The best values are represented by yellow triangles.

Concerning the drainable porosity at the soil surface two different observations can be made (figure 33). In the Zastlerbach, the Eyach and the Roggelshäuser Bach the variations of the porosity seems to be very sensitive to the model efficiency. On the other hand, in the Fichtenberger Rot and the Haslach catchment the impact on the efficiency by the variation is less significant. A slight maximum in the middle of the range of values can be detected.

Since the constant hydraulic conductivity with depth shows a complete non-sensitive behaviour, no dotty plots are given for the variations of this parameter.

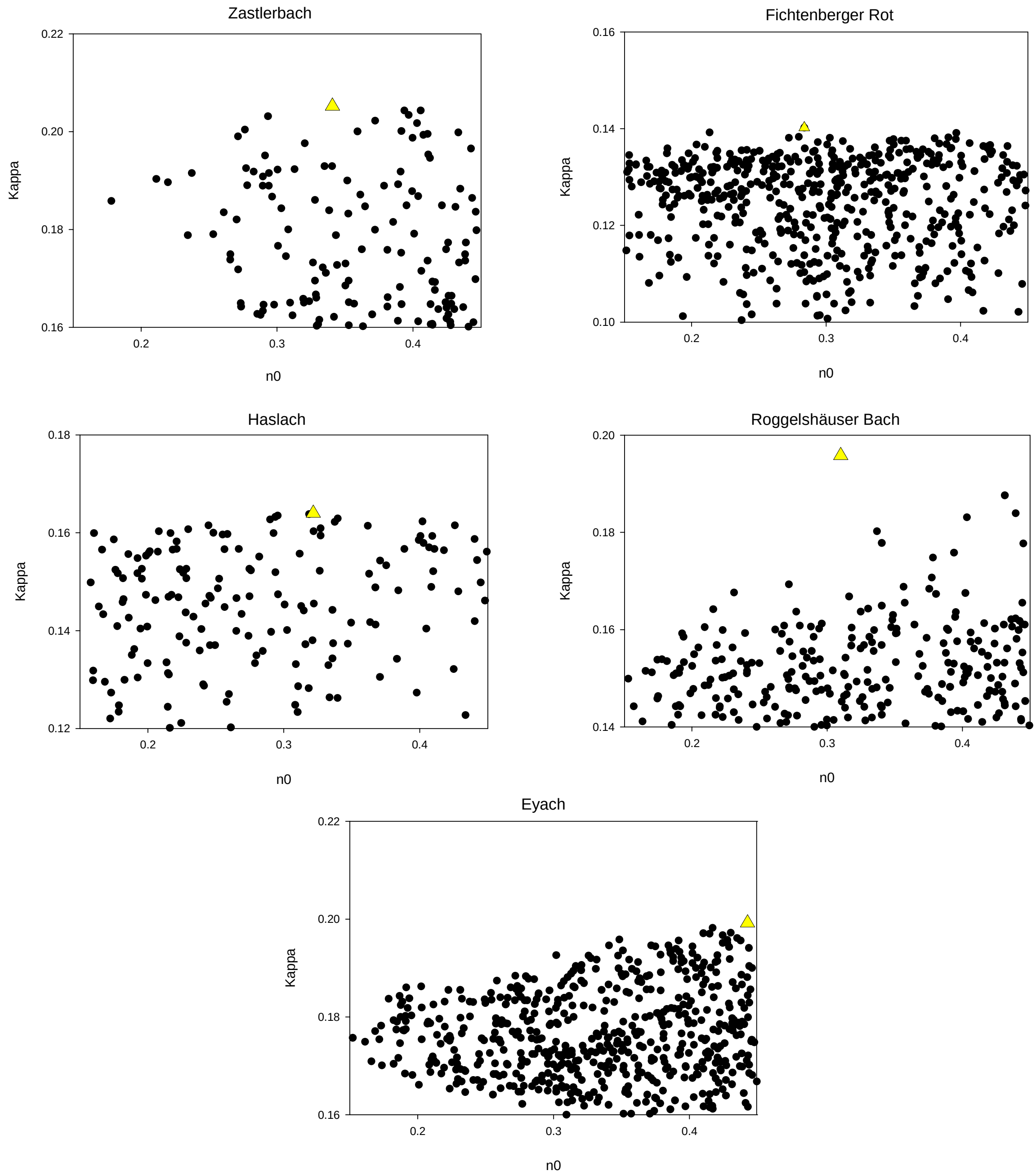


Figure 33: Plots of the drainable porosities at the soil surface against the objective function of Kappa. The best values are represented by yellow triangles.

4.2 Resulting regimes

In order to test if the model, which was driven by average climatic data, has the ability to reflect the long-time hydrological dynamics of the catchments, the simulated daily discharges during the calibration period were used to calculate the corresponding Pardé coefficients. Those were compared to the regimes, which were presented in chapter 2 (figure 34).

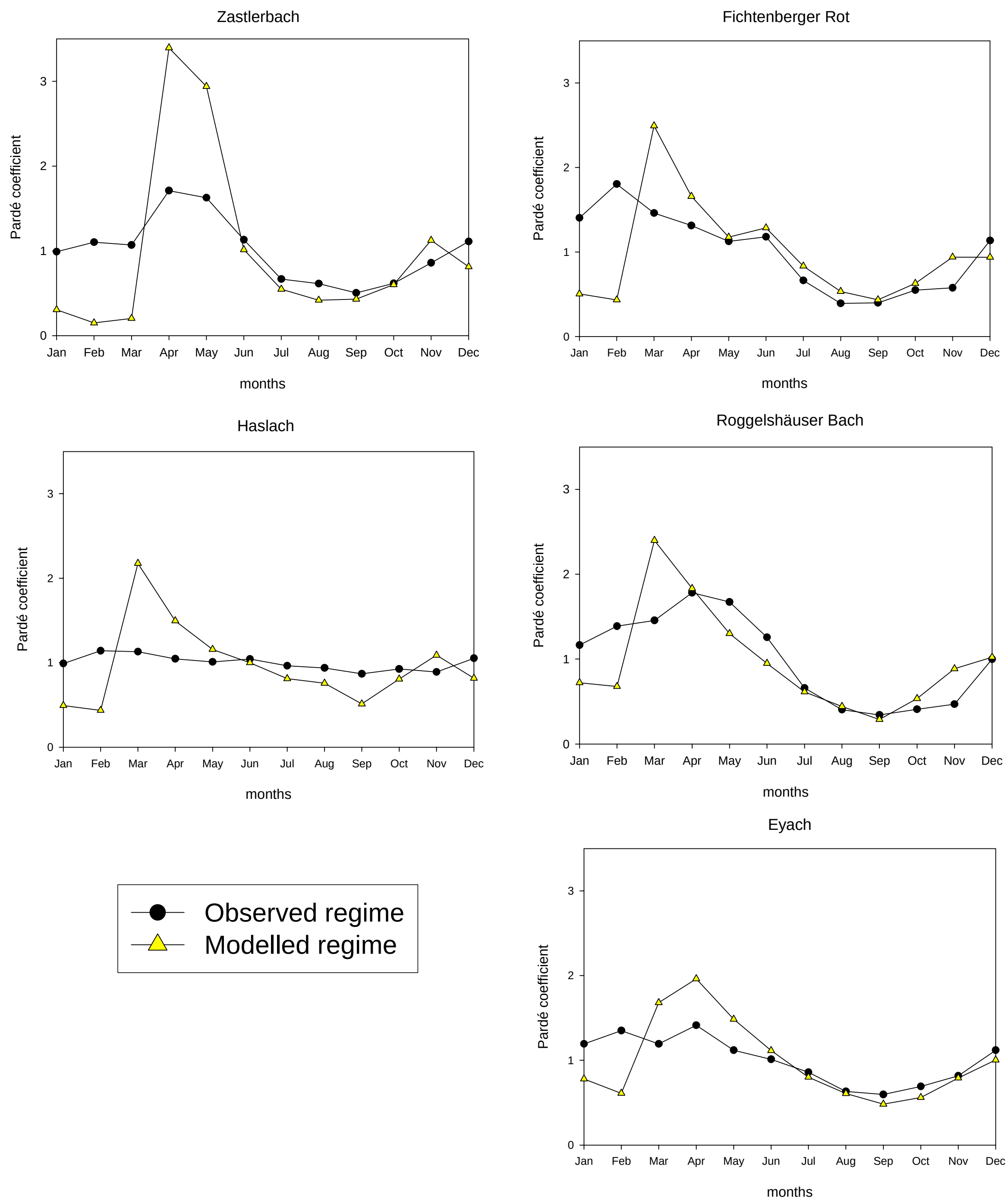


Figure 34: Modelled and observed regimes of the five catchments.

In all catchments poor agreement in the first half of the year with an underestimation of the discharges in January and February can be noticed. However, the accuracy of the simulated Pardé coefficients in the second half of the year seems quite good.

Further differentiation can be made by focussing on the occurrence of the maxima and minima. In the Zastlerbach, the Eyach and the Haslach catchment, the months with the highest discharges were identified correctly even though the amplitude was overestimated. The main peak in the Roggelshäuser Bach catchment was calculated one month in advance whereas the maximum in the Fichtenberger Rot basin was predicted one month too late.

Except for the Haslach catchment, the recession in the late summer was predicted satisfactorily. Moreover, an underestimation of the discharges in the Zastlerbach, the Fichtenberger Rot and the Haslach catchment for December can be detected.

The theories which were presented in chapter 3 indicate that the fluctuations of the groundwater level are controlled by seasonal meteorological variations, resulting in an expanding and contracting of the stream network. Although there is no quantitative information about efficiency, the seasonal comparison of the spatial expansion of the stream network, which is given in figure 35, can be used to evaluate the consistency of the model structure.

Every catchment shows distinctive disagreements between the extents during the periods with much discharge and those during the recession period in the late summer. Particularly within the upper areas of the catchments, the number of draining streams increased significantly during spring. Generally, the stream networks during the late summer agree better with the reference maps than those during spring. However, it can be noticed, e.g. in the Roggelshäuser Bach catchment, that there are discontinuities within individual stream channels.

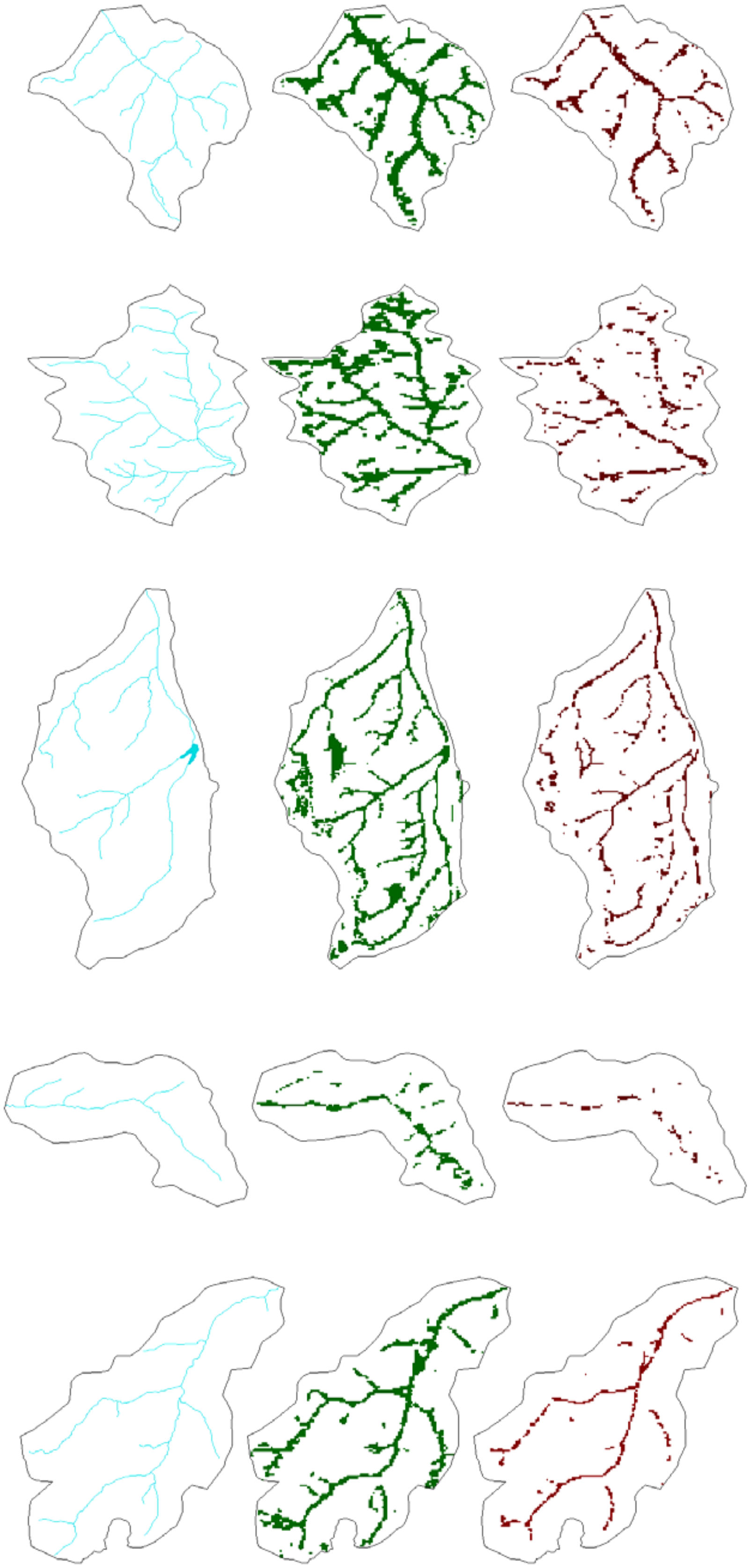


Figure 35: Seasonal development of the spatial extent of the stream networks with the reference maps in blue and the simulated stream networks during spring (green) and the late summer (red).

4.3. Discussion

4.3.1 Sensitivity analysis and calibration

The results of the Kappa statistics show clear differences between the Haslach and Fichtenberger Rot catchment and the remaining basins. This could be related to differences in the geomorphologic features. The three catchments with considerable higher values have distinctive topographies with valleys which are easily to identify. Due to a strong relationship between the slope and the lateral subsurface flow in the model, these profiles promote the concentration of water in the corresponding cells resulting in spatially confined stream networks which can reach higher Kappa coefficients.

According to the classification by LANDIS & KOCH 1977, the agreement between the modelled stream networks and the reference map can be described as fair. But those relative low values of Kappa coefficients need not necessarily indicate that the modelling was unsuccessful. Stream networks are line-like structures and account only little for a catchment's area. Therefore, differences in the resolution, uncertainties of defining the exact position and conversion errors can easily result in deviations between the reference and the simulated stream network (Güntner et al. 2004). Additionally, studies (e.g. VIERA & GARRETTT 2005, FOODY 1992) have shown that through the overestimation of the chance agreement in the calculation of Kappa, an underestimation of the accuracy can occur. This has made a particularly strong appearance for rare findings, like stream cells. However, for the subject of model calibration it is less important to reach absolute high Kappa values than to identify significant relative differences between the individual stream networks.

Since they are dependent on each other it makes sense to compare the transmissivity and average drainable porosity instead of using the individual parameters. The calculated transmissivity, which is based on the calibrated total depth, the hydraulic conductivity and the decay coefficient, seems to be closely related to the prevailing geologies. The transmissivity in the Zastlerbach basin, which is dominated by low conductive metaphoric rocks, is by far the lowest and consistent to previous study results (GÜNTNER ET AL. 2004). Consequently, the transmissivities of the Fichtenberger Rot and the Eyach catchment, where sandstones and marls are present, are significantly higher. Those values agree well with the hydraulic conductivities which were reported in WABOA 2007. The highest transmissivities are calculated for the Haslach and the Roggelshäuser Bach catchment. The highly conductive quaternary material, the karstified limestone respectively fit well with the calibrated parameters. Those results confirm the assumption introduced by CARLSTON 1963 that the transmissivity is strongly related to the drainage density.

It is more difficult to evaluate the results of the average drainable porosities because of the high value ranges occurring for the specific rocks. It is only the value of the Zastlerbach basin that can be clearly related to the geology. Significant differences between the remaining catchments are no longer existent due to the similar storage properties of the rocks. Finally the simulated drainable porosities between 0.07 and 0.083 are consistent with reference values which were presented by MORRIS & JOHNSON 1967, who reported a range for sedimentary rock of 0 to approximately 0.45. In general, the calibrated parameter sets seem to be reasonable for the specific catchments.

The fact that they are dependent on each other makes it also difficult to interpret the sensitivity of the individual parameters, which do not show uniform patterns. Additional challenges arise from the small number of reasonable Monte-Carlo runs. For the individual catchments the model efficiency reacts differently to variations impeding the identification of sensitive parameters. Especially in catchments without distinctive topography the dependencies on the parameter variations are less clear than for example in the Eyach catchment. According to previous considerations, the smaller absolute Kappa coefficients reduce the range of possible positive values, which makes it more difficult to detect significant differences between the individual stream networks.

Although the number of high Kappa values for total depths below 12 meters is relatively high, this tendency is not very distinctive when compared to other parameters. The total depth has to account for both, the storage properties and the subsurface conductance of the catchment. The sensitivity only rises if a variation of total depth enforces a better realisation of the drainable porosity as well as the hydraulic conductivity. Therefore, it can be assumed that, especially for Roggelshäuser a low total depth satisfies the effects of the drainable porosity and the hydraulic conductivity best.

The sensitivity analysis of the decay coefficient shows greater differences. The strong sensitivity of the Zastlerbach basin can be attributed to the very low conductive metamorphoric rocks. In order to satisfy the lower transmissivities under moderate total depths, very low decay coefficients are necessary. A clear maximum is also found for the Eyach catchment whereas the remaining basins struggle with the problem of equifinality (e.g. BEVEN & BINLEY 1992).

Concerning the saturated hydraulic conductivity, there is again the differentiation between catchments with distinctive topography and flat basins. The effects of the conductivity variations become more apparent in steeper catchments because of the strong weight of the slope on the lateral subsurface flow (see Equation 8). A critical remark is necessary on the determination of the lower boundary. In at least two catchments possibly better parameter sets were inhibited because of an overestimation.

The dotty plots of the drainable porosity at the soil surface indicate a moderate sensitivity. Hardly any high efficiency is reached below a porosity of 0.2. Except for the Eyach catchment, no unique maximum can be detected.

The fact that the variations of the constant hydraulic conductivity have no significant effects on the model efficiency is a consequence of the low total depths and the relative high decay coefficients resulting in too high conductivities. The order of magnitude of the constant hydraulic conductivity is not reached.

4.3.2 Resulting regimes

The occurring differences between the modelled and observed regimes in the first half of the year and December can be explained by problems in snowmelt simulation.

The main reason is that only mean monthly averages of the air temperature were available and have to be transferred to daily time steps. This means that if the mean monthly temperature in a month was below the freezing point, all corresponding daily values were assumed to be also below the freezing point accordingly. Thus, according to equation 8, all precipitation during the winter season is retained and can no longer recharge the groundwater, which results in an underestimation of the discharges. When the first month with positive temperatures is reached, all water which was retained over the last months melts resulting in an overestimation of the snowmelt. Although a running average was applied, this effect could not be damped. Additionally, it is important that the snowmelt parameters were not calibrated but estimated to typical values according to SEIBERT 2002.

Those effects are particularly significant for catchments with a distinctive nivo-pluvial regime like the Zastlerbach basin. For catchments with warmer climates like the Roggelshäuser Bach the impacts are less important, naturally.

Except for the Haslach catchment where the impact of the little lake made it difficult to reflect to the seasonal variation of the discharge, the recession during the late summer was modelled with good accuracy. This is also due to the selection of the calibration period (see introduction of chapter 4) that delivered the best agreement for the summer months.

Despite the differences in the early months, the simulations of the regimes are acceptable reflecting the major discharges dynamics.

The seasonal development and extent of the stream networks in the catchments seems reasonable and attest a good realisation of the dominant processes on a seasonal scale. Additionally, discontinuities in the modelled stream networks could deliver information about the location of topographic favoured groundwater infiltration areas (SOPHOCLEOUS 2002).

5. Evaluation

In order to evaluate the results of the parameter calibration, the model was run with conventional input data (mean daily precipitation, mean daily air temperature) over a period of three years while the first one was used for initialisation. Due to the differences in data availability, different years have to be chosen for each catchment.

5.1 Hydrographs

A common approach for model evaluation is the comparison of modelled and observed stream flow. The model efficiency after NASH & SUTCLIFFE 1970 is calculated to quantify the agreement. Additionally, the transformed logarithmic efficiency $\log n_{eff}$ is computed to reduce the sensitivity to extreme values and increase the impact of the low flows.

The hydrographs for the Zastlerbach during 2004 and 2005 are given in figure 36. It can be seen, that the general agreement is quite good and the stream flow dynamics are reflected correctly. The corresponding efficiencies are listed in table 13. The model predicts the low discharges during the dry second half of 2005 very well. Accordingly, the calculated logarithmic efficiencies are increased compared to the standard efficiency.

Table 13: Model efficiencies of the Zastlerbach

	2004	2005
n_{eff}	0.40	0.65
$\log n_{eff}$	0.38	0.65

However, the model tends to underestimate the peaks, especially during the spring, and to overestimate the discharges for the recession period.

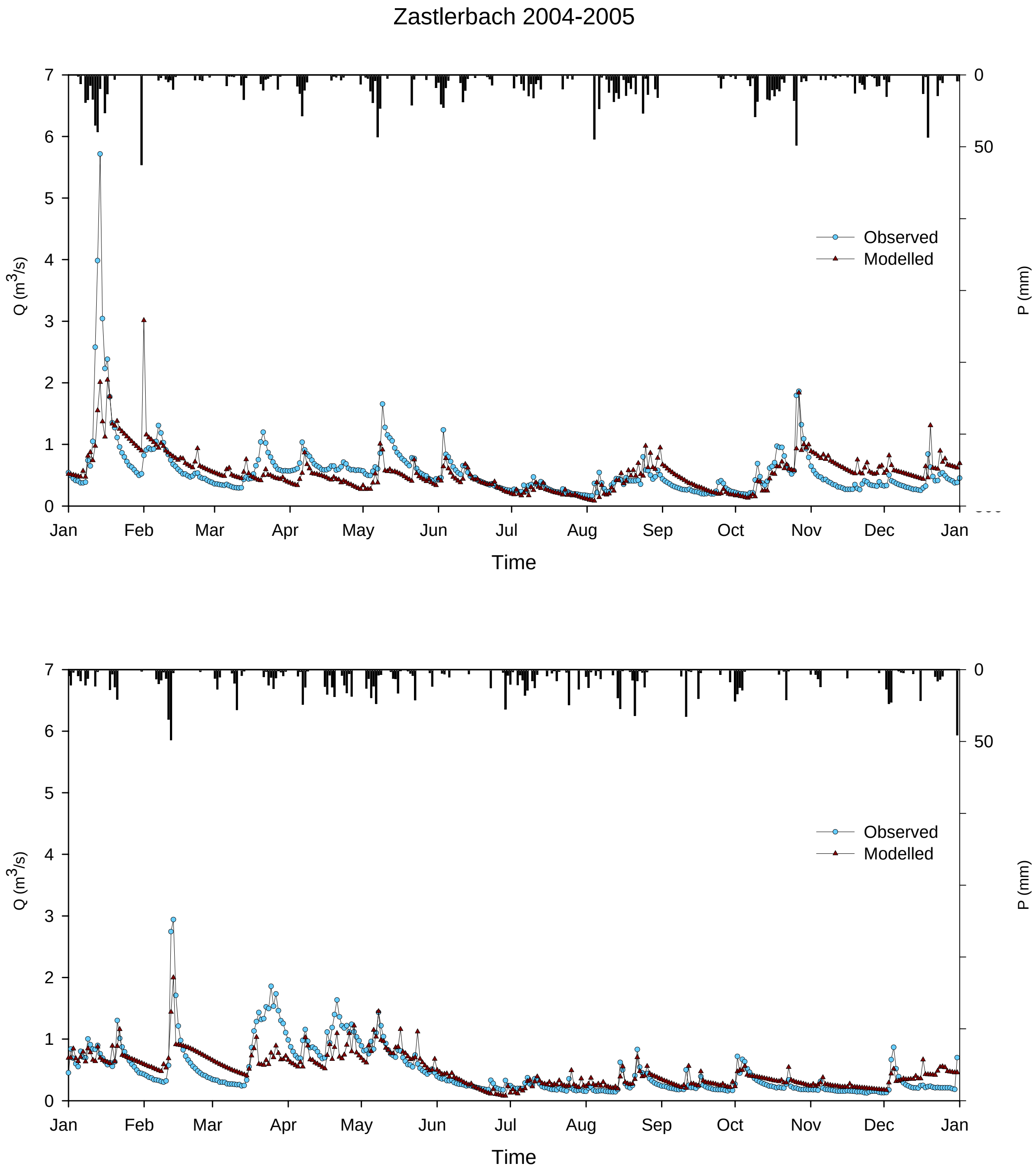


Figure 36: Modelled and observed hydrographs of the Zastlerbach (2004-2005).

The predictions for the Fichtenberger Rot, which are illustrated in figure 37, show something different. There, the model overestimates the discharges almost during the whole year. The extreme events and the recession period in the late summers are underestimated only. As a consequence, the logarithmic efficiency in 1991 amounts only to -0.15. The recession period in 1992 is modelled with higher accuracy resulting in a logarithmic efficiency of 0.42. In general, the modelled hydrographs seem to agree with observed discharges quite well. Thus, the conventional efficiencies are calculated to 0.47 in 1991 and 0.46 in 1992.

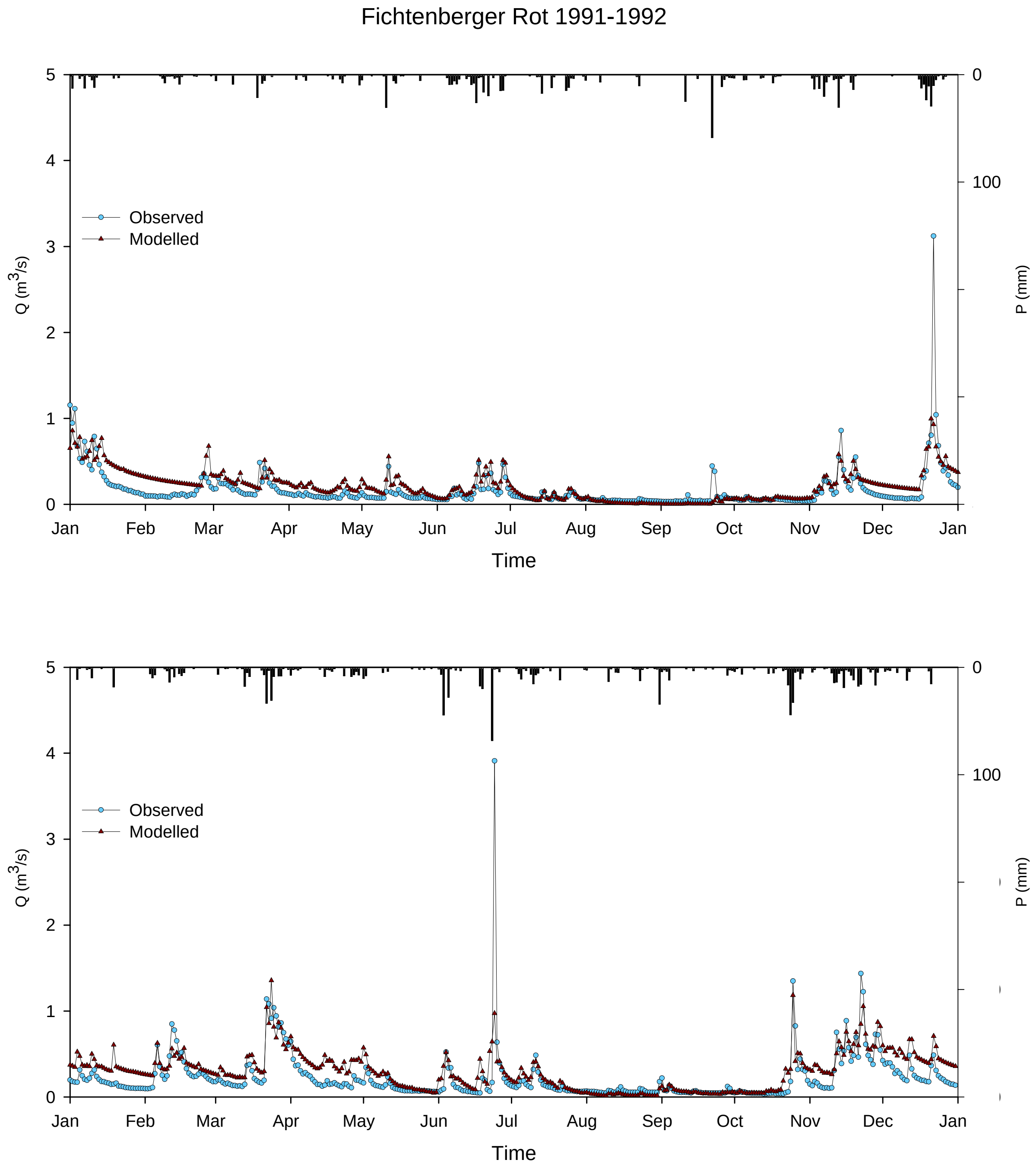


Figure 37: Modelled and observed hydrographs of the Fichtenberger Rot (1991-1992).

The results of the stream flow prediction of the Haslach in 1966 and 1967 are given in figure 38. In general, the hydrological dynamics seem to be reflected realistically showing relative high model efficiencies of 0.53 in 1966 and 0.58 in 1967. However, especially for the low flows, the model shows a tendency to underestimate the runoff, which arises from the little lake buffering the observed discharge at a level of approximately 0.5 m^3/s. According to this, the logarithmic efficiencies are significantly decreased and a value of only -0.08 is reached in 1966. In 1967 the logarithmic efficiency averages to at least 0.18.

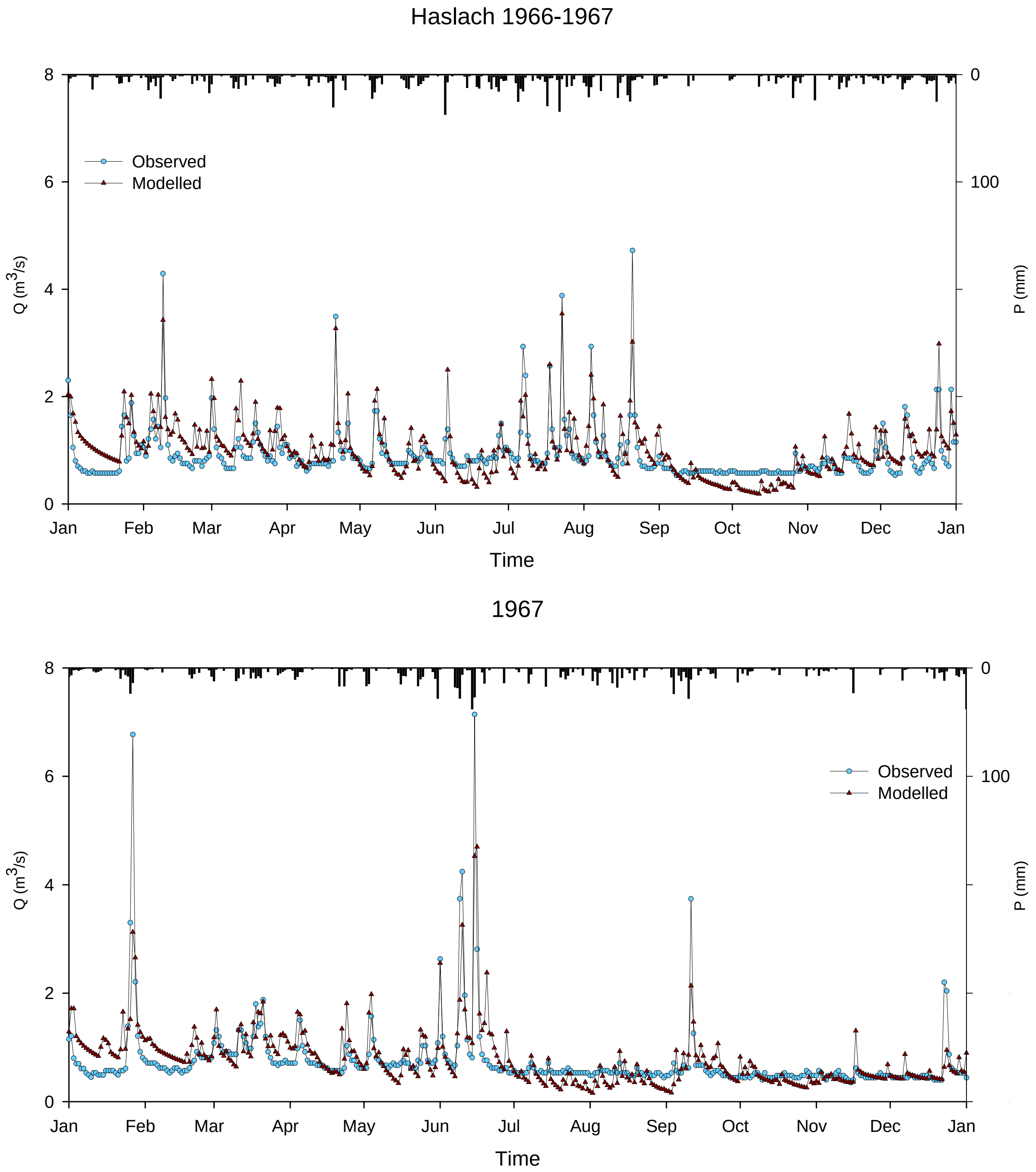


Figure 38: Modelled and observed hydrographs of the Haslach (1967-1968).

In figure 39 the predicted and observed hydrographs of the Roggelshäuser Bach are given. A significant underestimation of the discharges during spring 1967 as well as 1968 can be detected. Additionally, the observed peak discharges are not reached. Thus, the model efficiencies with 0.38 in 1967 and 0.24 in 1968 are, compared to the previous results, relatively low. On the contrary, the low flows are predicted with much accuracy and the resulting logarithmic efficiencies amount to 0.53 in 1967 and 0.49 in 1968.

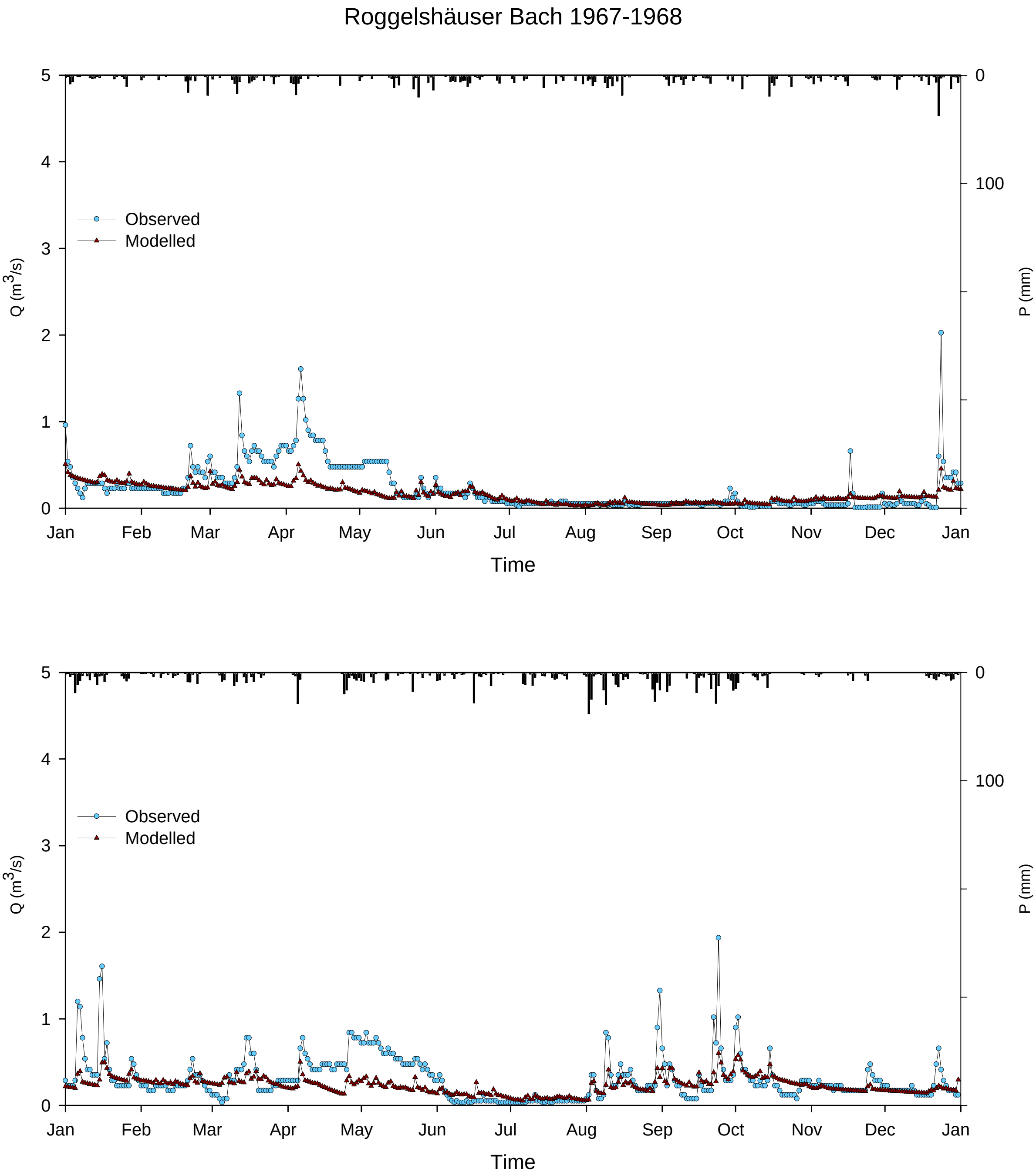
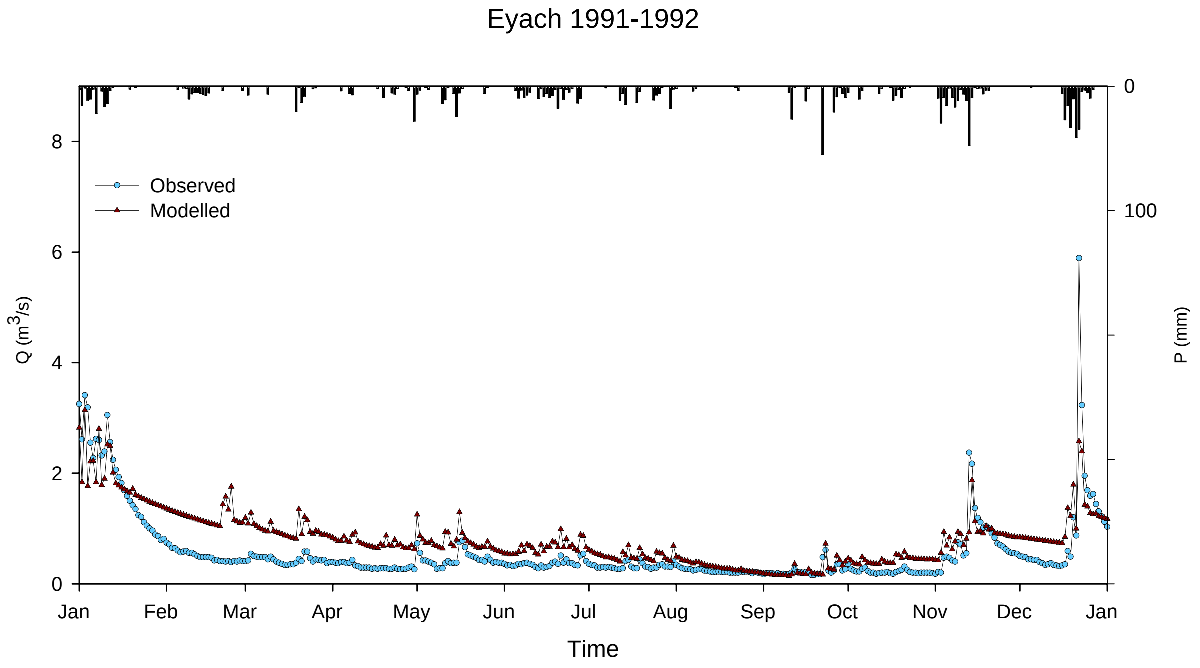


Figure 39: Modelled and observed hydrographs of the Roggelshäuser Bach (1968-1969).

The comparison between the modelled and observed runoffs of the Eyach, which are illustrated in figure 40, shows a good reflection of the hydrological dynamics. Due to the overestimation of the recession in the winter of 1991 the logarithmic model efficiency is computed to only 0.17 and the conventional to 0.53. In 1992 the highest value of conventional efficiencies is reached with an amount of 0.73. Nonetheless, the logarithmic amounts to 0.63, which is due to the ongoing overestimation during the recession period.



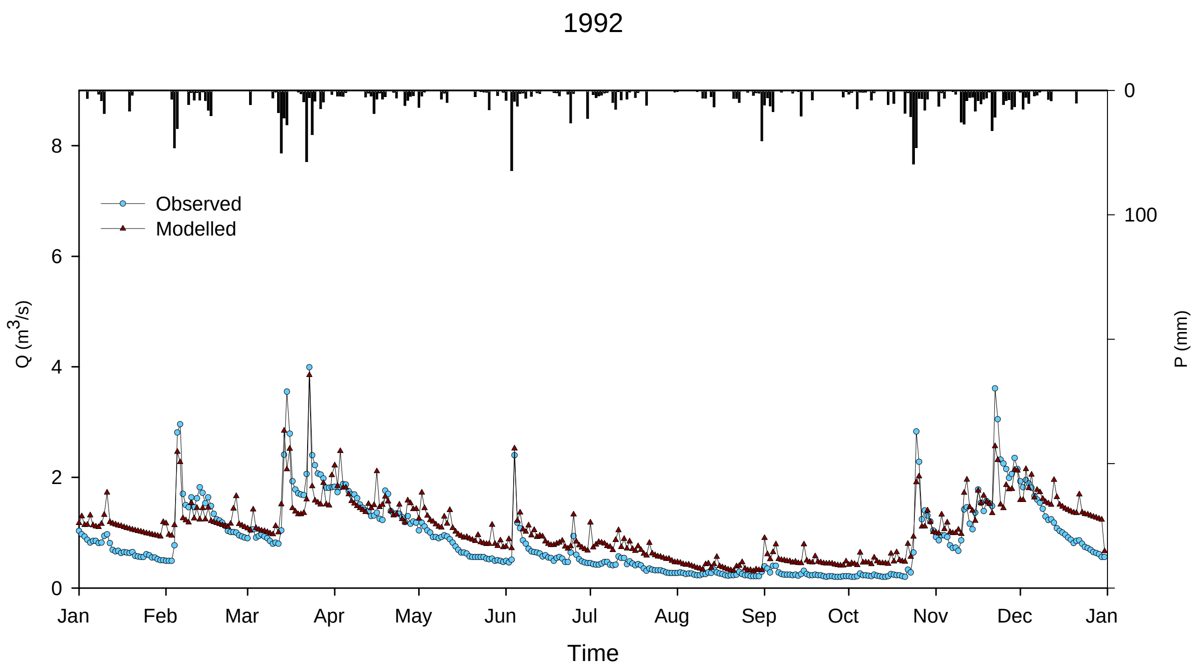


Figure 40: Modelled and observed hydrographs of the Eyach (1991-1992).

5.2 Water balances

Besides the comparison of the hydrograph dynamics, the calculation of water balances can deliver additional, more integrative information on the model's simulations. Therefore, the averages of the precipitation, evapotranspiration and discharge over the modelled two years were calculated. The corresponding water balance is not closed, which is caused by differences in initial and final water levels, numeric calculation errors, and occurring subsurface outflows.

For the comparison, the observed discharges sums were computed and the evapotranspiration was calculated as the remainder term of the water balance equation. This, of course cannot be equalised with measurements, not considering potential subsurface outflows, but can at least estimate the possible evapotranspiration dimension. The results and the relative errors are listed in table 14.

Table 14: Modelled and observed water balances with the corresponding relative errors

	P (mm/a)	mod. Q (mm/a)	obs. Q (mm/a)	$Error_Q$ (%)	mod. ETP (mm/a)	obs. ETP (mm/a)	$Error_{ETP}$ (%)
Zastlerbach	1404	911	913	-0.2	493	491	0.4
Fichtenberger Rot	1018	462	350	32.0	544	668	-18.0
Haslach	1258	694	573	21.1	649	685	-5.3
Roggelshäuser Bach	993	471	636	-26.0	527	357	47.6
Eyach	1468	992	769	29.0	489	699	-30.0

The predictions of summed discharge and evapotranspiration in the Zastlerbach catchment are exceptionally good and nearly identical with the observed values. For the Fichtenberger Rot distinctive deviations of discharge as well as evapotranspiration can be noticed. The results of the Haslach are slightly increased. Although the discharge was significantly overestimated by 121 mm per year, the evapotranspiration was modelled with high accuracy and only a minor underestimation. For the Roggelshäuser Bach, the modelled discharge and evapotranspiration sums differ significantly from the reference values. The discharge sum was underestimated by 165 mm per year whereas the evapotranspiration sum was overestimated by 170 mm per year. The simulations of the Eyach are contrary to the results of the Roggelshäuser Bach. The discharge was drastically overestimated and the evapotranspiration was undercharged by 210 mm per year.

5.3 Parameter uncertainty analysis

A Rainfall-Runoff model can only be a simple approximation of the processes controlling the hydrological response in a catchment. Therefore, the results of the model are always associated with certain degree of uncertainty that makes an analysis necessary (BEVEN 2001b). There are four main sources of uncertainty: Errors in model input and output data, uncertainties which derive from inaccurate model structures and uncertainties due to inadequate model parameters (e.g. BUTTS ET AL. 2004). This chapter deals with the latter.

Due to time constraints no state-of-art methodology (e.g. GLUE) could be applied. Instead of that, ranges of reasonable parameter values were identified with respect to the accomplished sensitivity analysis (see chapter 4). For every parameter a pair of lower and upper values was selected to represent two extreme cases. In the first case, all parameters are set to a minimum which should represent a small storage. This results in low baseflow discharges and heavy reactions to precipitation events. In the other case, the parameters are set to a maximum accounting for a large storage with high baseflow discharges. Those sets are used to predict a lower and an upper bound which should enclose the results of the calibrated parameters. All calculations were executed for the periods which are presented in chapter 5.1.

The best parameter sets after the calibration and the parameters of the lower respectively the upper bound are given in table 15. Additionally, the corresponding ranges of the main runoff characteristics are listed. Due to the difference in sensitivity, varying uncertainty ranges were estimated for the respective catchments. The values of the total depths vary between 5 and 22 meters while the uncertainty range of the decay coefficient amounts only to 3. The range of the drainable porosity at the soil surface was assigned between 0.2 at 0.5 whereas a range of 3 m/d was estimated for the hydraulic conductivity. Due to the complete non-sensitive behaviour of the constant hydraulic conductivity, no uncertainty ranges were specified.

The corresponding results of the comparison of the mean annual discharge (MQ) show that in all catchments the calculations with the best parameters lie within the narrow uncertainty ranges. The same applies to the mean of the highest discharges (MHQ) and the mean of the lowest discharges (MLQ). However, the ranges are much wider than for the mean annual discharge. Additionally, the mean annual sum of the discharge (Q_{sum}) was calculated. For the first time it can be noticed that a value, which was calculated with the best parameter set, does not lie within the computed uncertainty range. The sum of the Haslach amounts only to 694 mm whereas the uncertainty range is specified between 696 and 766 mm. All the other values lie within the uncertainty range. However, there are differences which of the boundaries reflects the upper and which one the lower value.

Table 15: Estimated range of runoff characteristics and used parameter sets

		Zastlerbach (2004/05)	Fichtenberger Rot (1991/92)	Haslach (1966/67)	Roggelshäuser Bach (1967/68)	Eyach (1991/92)
Z (m)	lower	6.0	6.0	5.0	5.0	6.0
	best	10.5	10.0	8.6	8.4	11.6
	upper	18.0	15.0	15.0	18.0	22.0
b	lower	1.0	1.0	1.0	1.0	1.0
	best	1.1	2.6	2.2	2.3	2.2
	upper	4.0	3.5	4.0	4.0	4.0
n_0	lower	0.28	0.20	0.20	0.25	0.20
	best	0.34	0.28	0.29	0.31	0.44
	upper	0.50	0.50	0.50	0.50	0.50
k_0 (m/d)	lower	1.0	1.0	1.0	1.0	1.0
	best	2.0	1.7	2.6	2.5	1.8
	upper	3.0	4.0	4.0	3.0	3.0
MQ (m^3/s)	lower	0.50	0.26	0.84	0.19	0.94
	best	0.52	0.27	0.91	0.21	0.97
	upper	0.58	0.28	0.96	0.23	1.04
MLQ (m^3/s)	lower	0.27	0.08	0.19	0.07	0.32
	best	0.31	0.13	0.48	0.13	0.63
	upper	0.46	0.17	0.65	0.17	0.74
MHQ(m^3/s)	lower	1.67	1.66	4.99	0.82	4.79
	best	1.17	0.63	2.34	0.37	1.90
	upper	0.80	0.49	1.66	0.32	1.48
Q_{sum} (mm/a)	lower	873	452	766	408	949
	best	911	462	694	471	992
	upper	1018	467	696	526	1074

Besides the runoff characteristics, the hydrographs are illustrated to have an additional graphic evaluation possibility (figure 41-45). For purposes of clarity only results for one year for each catchment are given. All catchments have in common that the baseflow of the upper bound is significantly increased whereas the peaks are significantly decreased. This is contrary the behaviour of the lower bound. During the rise and the fall of a peak the uncertainty range becomes very narrow due to the crossing over the two bounds. Generally, the hydrographs of the best parameterisation lie within those ranges, and exceeds them rarely.

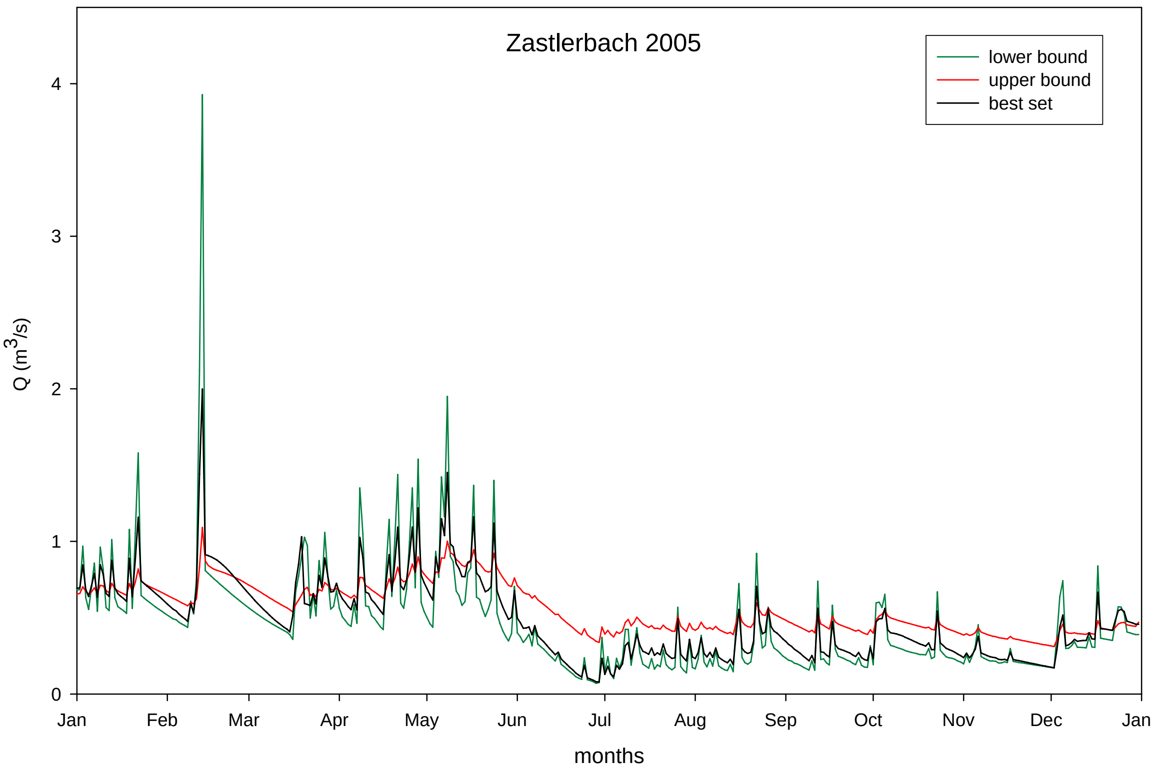


Figure 41: Uncertainty boundaries of the Zastlerbach.

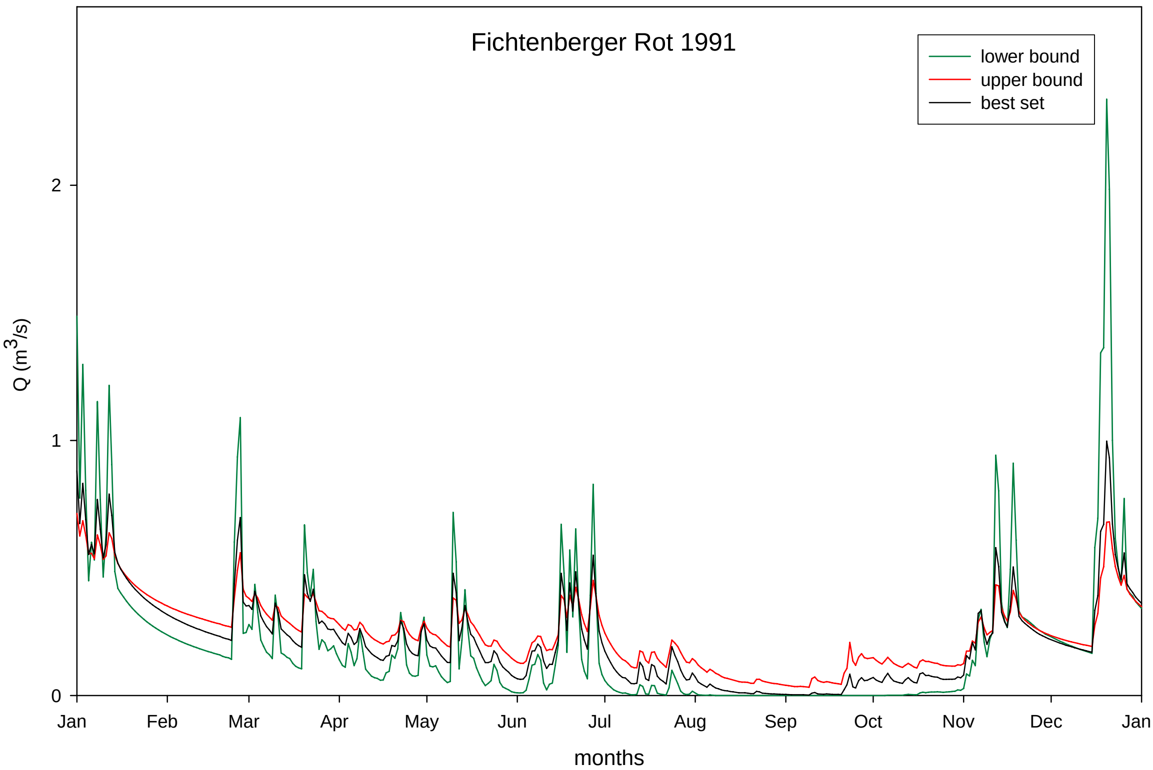


Figure 42: Uncertainty boundaries of the Fichtenberger Rot.

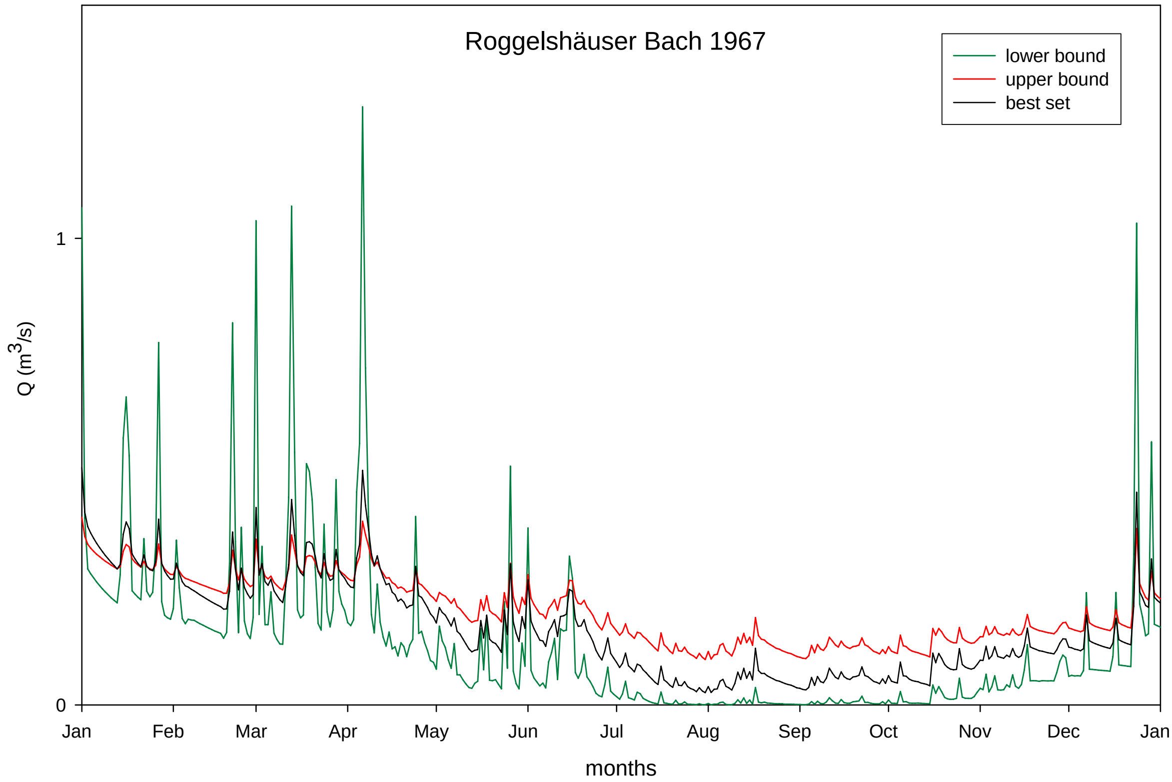


Figure 43: Uncertainty boundaries of the Roggelshäuser Bach.

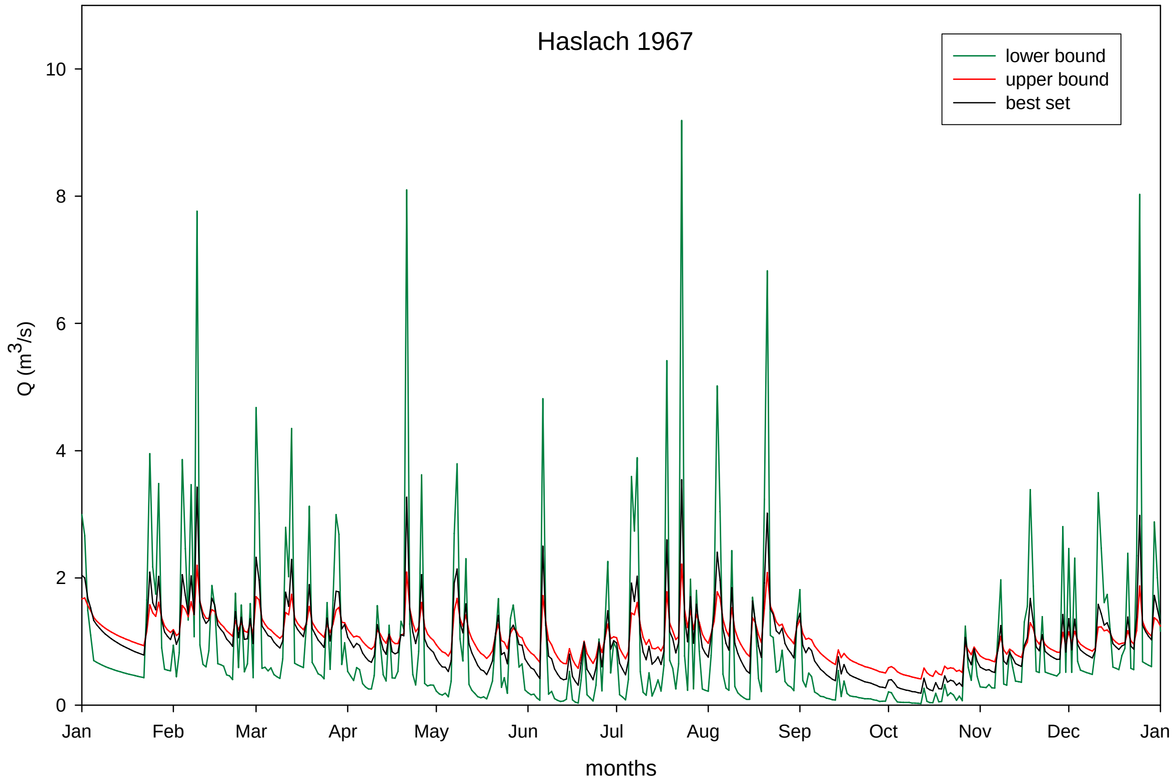


Figure 44: Uncertainty boundaries of the Haslach.

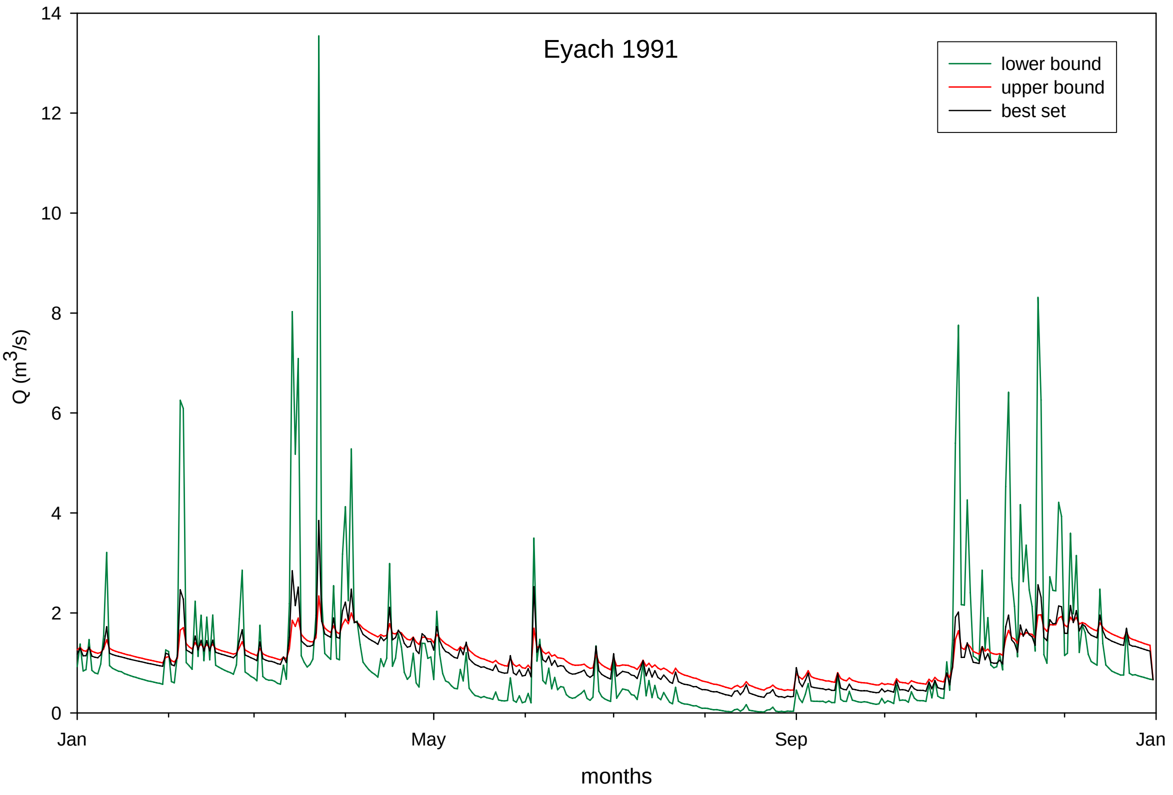


Figure 45: Uncertainty boundaries of the Eyach.

5.4 Discussion

5.4.1 Hydrographs

Two major problems concerning the hydrographs can generally be identified. First, the model is not able to predict the high peaks reliably, especially in the first half of the year. One explanation may be that processes dominating the runoff generation were not implemented in the model. The whole model approach is based on the assumption that the runoff is dominated by subsurface water which agrees with the findings of SKLASH & FARVOLDEN 1979. However, there are additional processes that affect the runoff generation. For example, in the Zastlerbach catchment UHLENBROOK 1999 observed that between December and January precipitation on frozen soil occurs frequently resulting in winter floods. Furthermore, surface runoff on steep slopes, bedrock, and sealed areas are reported (e.g. HOEG ET AL. 2002, UHLENBROOK ET AL. 2002). Those processes, which are not implemented in the model, can also occur in the other catchments. Unfortunately, hardly any hydrological research has been done, which makes it impossible to deliver further details.

Another explanation is related to the used snowmelt parameters. Those were not calibrated on the specific cases but were set to typical values. Especially the degree-day-factor and the threshold temperature can strongly affect the runoff predictions in snowmelt dominated catchments (SEIBERT 2002).

The second major problem is the overestimation of the runoff during the recession periods. It can be observed that the modelled recession does not follow the classical decline, which was described by MAILLET 1905. The modelled decline of the runoff over the time is too small to account for a realistic simulation. The reaction time of the subsurface in the model is mainly controlled by the hydraulic conductivity, the drainable porosity and the corresponding decay coefficient. The problem seems to be related to the latter. If a smaller decay coefficient would be selected, a swifter decline could be realised. However, the drainable porosity would simultaneously be diminished. Thus, the model could not assure a good simulation of the base flow periods any longer. This is definitely a limitation of the model and could be corrected by the implementation of a separate decay coefficient for the drainable porosity as it used to be in the previous versions. According to that, the assumption that the declines are caused by the same effects must be discarded. More attention must be paid to the appearance of preferential flow paths (e.g. FLURY ET AL. 1997, BEVEN & GERMANN 1982).

A minor problem is expressed by the partial low logarithmic efficiencies. The model tends to little over- or underestimations of the baseflow. This could be related to the simple realisation of both the potential and the actual evapotranspiration, which were implemented to account for the rare data availability. However, the bad results of the Haslach catchment are due to the existing of the small lake. It buffers the runoffs and results in an underestimation of the model.

Despite the problems mentioned, the calculated model efficiencies as well as the subjective visual evaluation of the hydrographs of all catchments are acceptable as the model is able to reflect the hydrological dynamics.

According to the evaluation suggestion of the conventional efficiencies by HENRIKSEN ET AL. 2003 (table 16), the simulations of the Zastlerbach can be classified as poor for the first year and very good for the second year.

Table 16: Evaluation of Nash-Sutcliffe efficiencies (according to HENRIKSEN ET AL. 2003)

	Very poor	Poor	Good	Very good	Excellent
n_{eff}	< 0.2	0.2 - 0.5	0.5 - 0.65	0.65 – 0.85	> 0.85

The Haslach shows good results in both years while the evaluation of the Roggelshäuser Bach and the Fichtenberger Rot show continuous poor simulations. The best performances are made by Eyach catchment where good results in the first year and very good results in the second year can be found.

The modelled efficiencies can be compared with the results of previous studies modelling in ungauged basins (see table 17).

Table 17: Comparison of model performances in studies modelling ungauged basins

studies	n_{eff}
this study	0.24 - 0.73
HUNDECHA ET AL. 2007	0.87 - 0.96
PARAJKA ET AL. 2005	0.54 - 0.64
BARDOSSY 2007	0.17 - 0.90

HUNDECHA ET AL. 2007 established a regionalisation method transferring individual parameter values, which are linked to catchment properties, from similar gauged basins. For the simulations a distributed model based on the concept of the HBV (BERGSTRÖM 1995) was used. The same model type was used by BÁRDOSSY 2007. Instead of transferring individual parameters whole sets are assigned in case the previously regionalised runoff and climatical characteristics of the ungauged basin provided are similar to the donor catchment. PARAJKA ET AL 2005 tested different regionalisation methods using the HBV model in 320 Austrian catchments.

The performances of the actual approach can not keep up with the reported results and must be assigned to the low-to-mid range of them. However, it must be considered that the approaches have different starting conditions and were executed by different models. For example, the regionalisation by HUNDECHA ET AL. 2002 requires remarkable information about the ungauged basin (e.g. soils and land-use) which is not necessary for the actual approach.

5.4.2 Water balances

The results of the water balances evaluation are slightly different to those of the hydrographs. The Zastlerbach delivers exceptional good results concerning the water balance but only moderate performances in the hydrograph evaluation. On the contrary, for the Eyach the best hydrograph simulation can be performed while great differences occur in the comparison of the water balances. The remaining catchments show similar bad results. A possible solution could be related to the influence of the evapotranspiration. As mentioned above, the evapotranspiration is only inadequately represented. Since the dependence on the air temperature, a good realisation of the evapotranspiration is more essential to the catchments with lower altitudes. According to that, the impacts of the bad realisations are minor in the Zastlerbach catchment.

The quality of the hydrographs is primarily evaluated in consideration of an adequate reflection of the dynamics. The timing of the rise and fall of the peaks is mainly controlled by the parameters describing the subsurface. The absolute amount of the evapotranspiration is less important. Thus, a good realisation of the hydrographs dynamic need not necessarily implicate good simulations of the water balance.

Besides this possibility, other options also exist to explain the differences. Possible anthropogenic in- and outlets were not considered in the calculations. Furthermore, the reference evaporation was calculated as the remaining term of the water balance equation.

As a consequence, potential subsurface outflows as well as inflows from neighbour catchments could not be excluded. Especially in the karstified Roggelshäuser Bach catchment it is very likely that those connections may exist.

5.4.3 Parameter uncertainty analysis

The calibrated values concerning the runoff characteristics are mainly positioned within the ranges. The summarised discharge of the Haslach is the only one that is out of range. This result as well as the changing of the upper and lower bounds can be attributed to the integrative nature of the summarised discharge, which does not account for the temporal dynamics.

The absolute mean uncertainty amplitudes of the different catchments are controlled by the variations of the individual uncertainty ranges of the parameters and the dimension of the mean discharge. For example, the greatest amplitude is reached for the Eyach, where the greatest parameter intervals as well as the highest discharges can be detected. The position of the calibrated simulations in the uncertainty range is dependent on the offset to the parameter bounds. Since the calibrated parameters of the Zastlerbach are located closer to the lower bound, the corresponding hydrograph is more alike the lower bound, too.

Although relatively wide parameter uncertainty ranges were defined, the corresponding uncertainty of the model seems very small. Usually, small uncertainty ranges are assessed as a success and an indication of the capability of principle of parsimony in conceptual hydrological modelling (e.g. BEVEN 1993, PERRIN ET AL. 2001). However, the parameter uncertainty analysis that has been carried out also reveals some problems. During the peak and the low flow periods clearly defined uncertainty boundaries can be identified. Nonetheless, intersections during the rise and fall occur caused the ambivalent characteristics of the decay coefficient, which affects the drainable porosity and the hydraulic conductivity. Accordingly, no uncertainty range can be detected. As a consequence, the calibrated hydrographs does not always range between the given boundaries. Thus, those boundaries cannot be accepted as an adequate representation of the uncertainty. Further analyses with greater parameter ranges and state-of-art approaches (e.g. GLUE) have to be arranged.

6. Conclusion

The presented approach for model and parameter evaluation in ungauged basins has shown some promising results but also some evident limitations.

The underlying assumption that the development and initiation of stream networks is controlled by the properties of the subsurface could be plausibly confirmed by the dimensions of the calibrated parameters. Generally, those agree with the values which were reported for the specific catchments. The subordinate aim of searching an adequate objective function could be achieved by the application of the Cohen Kappa statistics. Although the absolute values were relatively low, the Kappa statistics seems to be suitable to evaluate stream network modelling since significant relative differences between good and poor simulations were found. However, a connection between the performances' quality and geomorphologic features could be seen.

The modifications on the hill slope model Hill-Vi showed some problems. Since the unsaturated zone was removed to assure fast and robust model runs, the evapotranspiration was represented by a simple correlation based on the ground water table depth. However, the results of the water balances and the hydrographs evaluations showed that this representation does not deliver adequate simulations. The same applies to the combination of the decay coefficients of the drainable porosity and the saturated hydraulic conductivity in just one parameter. This caused significant decreases in the model efficiencies due to a poor representation of the discharge recession and could be corrected by a reversion to the original model structure with two independent coefficients. Despite those limitations, the model performances are able to achieve results close to the efficiencies of regionalisation methods which were carried out with more detailed models. It seems that the differences are caused by the mentioned problems which could be verified by a comparison with a conventional calibration.

Besides the quantification of the impacts of the limited model, further investigation possibilities can be suggested: The model is capable to simulate distributed groundwater depths. Thus, an additional evaluation criterion could be established by comparison with observed groundwater level data. Moreover, it would be interesting to investigate the impacts of different reference data, e.g. higher resolved digital elevation models and stream networks maps on the performance.

In general, the approach of stream network modelling seems to be an alternative evaluation approach for ungauged basins. A major advantage is the low data requirement compared to the regionalisation methods. Besides the climatic forcing, a digital elevation model and a reference map are necessary only. Furthermore, additional information on the subsurface can be gained by the interpretation of physically based parameters.

However, attention must be paid to the selection of catchments in which the approach may be applied. It delivers good results for basins with distinct topographies and closed water balances. Problems occur for basins without clearly defined channel geometries and karstified undergrounds.

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