Understanding hydrological processes with scarce data in a mountain environment

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FULL TITLE: **UNDERSTANDING HYDROLOGICAL PROCESSES WITH SCARSE DATA IN A MOUNTAIN ENVIRONMENT**

SHORT TITLE: **HYDROLOGICAL PROCESSES IN A MOUNTAIN ENVIRONMENT**

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INTRODUCTION

Mountainous regions concentrate more than half of the earth's fresh water (Klemes, 1988; Rodda, 1994; Weingartner et al., 2003). In the current context of global environmental degradation and climate change, preservation of this fragile environment is a high priority and requires a good understanding of the different physical processes that affect the dynamics of these complex systems. However, the hydrology of mountainous areas is poorly known; Klemes (1988) qualifies it as "the blackest of the water cycle's black boxes". Observation networks are often very limited even though their density should be higher than what is available in the floodplains in order to capture the high variability in space and time of the water fluxes.

Thanks to the High Atlas range, the semi-arid regions of the south of Morocco receive an important amount of precipitation that sustains both agriculture and urbanization (Schulz and de Jong, 2004). However, whereas water demand is increasing due to demographic pressure, water resources are expected to decrease both in quality and quantity due to environmental changes. Appropriate hydrological modeling is thus required to understand the dominant processes controlling the water balance in the basin so that local authorities can be provided with science-based elements to carry out decisions on the management of water resources. This study focuses on a head-watershed (227 km²) in the Central High Atlas for which a large proportion of the precipitation falls as snow from November to April. Understanding the hydrology of head watersheds, the supplying zone, is a first step towards improved water management but it is also a challenge since many possible transfer mechanisms can be activated in theses watersheds: evapotranspiration, surface runoff, sub-surface runoff, groundwater flow, and snow dynamics.

Because of the harsh observation conditions of most mountain environments, gathering information through a combination of remote sensing, ground measurements, geographical information system (GIS) and models is a necessity. If many improvements have been made through the growing availability of remote-sensing data, selecting a model adapted to mountain hydrology remains a challenge. For instance Krishna (2005) combined remote sensing images, GIS techniques and ground measures to assess the snow and glacier cover in the Himalayas. For mountainous catchment-scale hydrological modeling Holko and Lepistö (1997) used TOPMODEL in Slovakia. Limitations of their modeling performance were due to their assumption of a homogeneous soil profile and inadequate snow subroutine. Andréassian et al. (2004) used two simple, continuous lumped watershed models (GR4J- a 4 parameter model and TOPMO –a 8 parameter modified version of TOPMODEL) on 62 watersheds in France. Both lumped models give good results but cannot provide any insight into the different processes. To learn more about the streamflow generation mechanisms in the Western Ghats region (mountains located in South India) Putty and Prasad (2000) used a lumped conceptual model based on the variable source area theory. Results show that flow from dynamic subsurface saturated zones contributes substantially to quickflow: field work and a modified version of the model were necessary. Globally, as stated by Winiger et al. (2005), remote sensing techniques and runoff models can lead to a better understanding of the mountain hydrology but they need to be improved and adapted to the region of interest.

Because the main characteristic of mountain environments is their complexity, one would be tempted to choose a detailed, physically based model to represent its water cycle. But more information than what is available is required to implement a distributed conceptual hydrological model leading to over-parametrization and a need for calibration. When conceptual rather than lumped modeling is implemented, special attention has to be paid to the calibration process since a better fit does not always mean that the model represents the reality of the processes. This problem is amplified over complex watersheds where almost all hydrological processes (superficial runoff, groundwater contribution, evapotranspiration, snow melt …) can occur simultaneously and their impact on streamflow are space-time scale dependent. In this context, the key questions are:

a- Does a good simulated streamflow guarantee a realistic representation of intermediate water balance processes including storage terms (e.g. snowmelt) and lateral flow redistribution (e.g. groundwater flow) ?

b- How can we get insight into processes without very intensive field work, or in other words, is it possible to develop a limited measurement strategy of some key variables that ensure proper assessment of the water balance components?

In this study, we first addressed the issue of parameter inter-correlation in the context of conceptual hydrological modeling over a semi-arid mountainous watershed. Then attempts were made to address the realism and accuracy of the simulated components of the water cycle such as groundwater flow and snow cover depletion once the model had been calibrated against observed streamflow only.

The main scope of this paper is to demonstrate that comparing the observed and the simulated streamflow is not sufficient to validate a model; intermediate processes have to be analyzed since compensation effects on the resulting outlet flow can be important. Obtaining good modeling results for the wrong reasons is indeed a well-known problem in the modeling community (Andréassian et al., 2004; Batchelor et al., 1998; Beven and Quinn, 1994). In this context, the underlying question is: can correct global streamflow restitution coincide with incorrect simulated processes?

This paper is organized as follows: we first present the object of our study, the Rheraya watershed. We then briefly describe the distributed conceptual model used to assess the hydrological cycle of this catchment, the Soil and Water Assessment Tool (Arnold *et al*., 1993). In the third section, the model performance, is analyzed in terms of streamflow simulation and the problem of parameter intercorrelation is identified throughout the optimization process. In the fourth section, we analyze the methodology and instrumentation developed to characterize two intermediate processes, namely the deep drainage and the contribution of snowmelt to streamflow. Finally, the results and the approach are discussed with respect to the objectives and questions raised above.

SITE LOCATION AND AVAILABLE DATA

The Haouz plain in south-central Morocco is made up of several intensively irrigated districts and is fed by nine head-watersheds located in the High Atlas range. The Atlas range and the Haouz plain belong to a larger watershed called the Tensift watershed covering 20450 km² (see Figure 1). This region is characterized by scarce water resources and is subject to frequent drought.

The study took place in one of the nine Atlasic head watersheds: the Rehraya catchment. This head watershed covers a surface area of about 227 km² and is characterized by a semi-arid and mountainous climate. Indeed, the mean measured annual precipitation at the outlet was 363 mm for the period 1971-2002 and the closest meteorological station registered a mean annual potential evapotranspiration of 1816 mm from a COLORADO pan for the period 1984-2001. The watershed altitude ranges from 1084 to 4167m, precipitation occurring as snow in the upper parts of the watershed. The main geological formation is granite but some clay inclusions are present north of the watershed as well as limestone-marl formations. Overall, the bedrock is shallow and fractured. In terms of geomorphology, rockfaces as well as scree slopes, debris fans and gravelly riverbeds are found. Slopes are very steep with an average grade of 19 % and soils are shallow. Based on these features, quick flow response is expected in this basin: its concentration time, estimated from geometric and geomorphological data, is 4 hours. Figure 2 presents the mean monthly rainfall and streamflow data at the watershed's outlet between 1971 and 2002. Rainfall and streamflow data are characterized by high inter- and intra- annual variability which is typical of semi-arid areas. The average streamflow is 19.3 mm/ month which represents 1.67 $m^3.s^{-1}$. A 2- to 3-month shift is observed between the pattern of annual rainfall and that of streamflow. This delay characterizes slow flow processes which can be caused by the existence of different geomorphological units as well as important groundwater and/or snow component in the water balance.

The Rehraya watershed is thus a complex terrain in which many hydrological processes are likely to be involved: i.e. surface runoff on the steep slopes and shallow soils under convective rainfall events, shallow sub-surface flow in the fractured bedrock, high evaporation rates under semi-arid climatic conditions, slow processes due to snow melt and/or to groundwater flow. Besides, few data are available: the only measurements are streamflow and rainfall collected at the outlet (Tahanaoute, see Figure 1) which are insufficient to provide insight into the hydrological processes occurring in the watershed.

Precipitation in a semi-arid mountainous watershed as in most other mountain watersheds is spatially and temporally highly variable (Holko and Lepistö, 1997; Krishna, 2005; Tani, 1996; Weingartner et al., 2003; Winiger et al., 2005). Figure 2 shows that temporal variability of precipitation and streamflow at the watershed's outlet is such that the standard deviation is of the same order of magnitude as the mean. As a result of the watershed morphology (high elevation gradient and the co-existence of different climatic influence: semi-continental from the north, oceanic from the west and Saharan from the south) important spatial variations of precipitation exist. The preliminary analysis of the data collected for a raingauge network installed in the watershed in 2003-2004 shows that annual precipitation (from 1st of April 2003 to 1st of April 2004) varies from 241mm to 562mm across the watershed (see Chaponniere, 2005a). Since only one raingauge is available before 2003, it is necessary to select years during which this raingauge is representative of the rainfall events occurring over the whole watershed, if not, hydrological modeling should not be attempted since modeling results depend on the quality of input data (Holko and Lepistö, 1997). We assume that a good correlation between the timing of peak flow and rainfall throughout the year attests that the outlet raingauge has been representative of the events which occurred on the watershed.

THE SOIL AND WATER ASSESSMENT TOOL

We chose, based on the climatic and topographical characteristics of the Rehraya watershed combined with the poor data set, a model that accounted for most of the hydrological processes while maintaining a simple approach,. Based on a comprehensive literature review, the Soil and Water Assessment Tool (Arnold *et al.*, 1993) turned out to be most suitable.

The Soil and Water Assessment Tool (SWAT) was developed to assess the impact of land-use and climate changes on water balance at the watershed scale. It operates at a daily time step, is physically based, spatially distributed and takes into account a large number of processes. SWAT has been extensively used and validated at various spatial and temporal scales. Hernandez *et al.* (2000) and Muttiah & Wurbs (2002) proved the validity of streamflow simulation under semi-arid climates. Arnold & Allen (1996) demonstrated the validity of major hydrological processes on three Illinois watersheds.

The elementary spatial unit of the model is the Hydrological Response Unit (HRU) which is defined by a unique combination of geology, land use and soil type in a given subbasin. Figure 3 illustrates the way the hydrological cycle is simulated by SWAT on each HRU; the processes are calculated sequentially. Major inputs for the model are topography, land use, soil type, groundwater characteristics and climatic data. Outputs are available at different spatial scales namely the Hydrological Response Unit, the subbasins and the basin; they mainly consist of water fluxes (evapotranspiration, surface runoff, infiltration, lateral runoff, percolation, streamflow in the stream network …) and vegetation variables (yield, root water uptake, ..).

Following an intensive sensitivity analysis (see Chaponniere, 2005a), eight model parameters were selected: the altitudinal gradients for precipitation and temperature (spatialization of the climatic dataset), the soil depth and available water capacity, the delay coefficient for groundwater and subsurface flow, a parameter used to convert snow water equivalent into snow coverage (parameter 'cov_2' in equation 4) and the snowfall temperature. These parameters will be calibrated in the following section.

When implementing the model on the Rehraya watershed we took into account three HRUs based on geomorphological *in situ* observations and soil and geological maps (see Figure 4): one unit is the valley bottom (referred to as "soil 2") characterized by deep soil with a texture dominated by sand (the river bed is gravel) and clay and the main land cover is a dense agricultural vegetation, the second unit is located in the upper part of the watershed ("soil 1a") and is made up of bare shallow and sandy soil and finally the third unit ("soil 1b") is in the lower part of the watershed and is characterized by intermediate shallow soil with a loamy sand texture and sparse trees (pines and juniper). All soils are considered to be constituted of a single layer. Once the spatial distribution of HRU is taken into account, twelve parameters must be calibrated.

STREAMFLOW MODELING PERFORMANCE AND IDENTIFICATION OF INTERCORRELATED PARAMETERS

The accuracy of the routinely available hydrological data is questionable: precipitation observations are rarely representative of the whole catchment, and streamflow itself is subject to measurement errors. Streamflow at the outlet is calculated on the basis of the measured water depth and the stage-discharge rating curve. In semi-arid and mountainous areas, intense convective rainfall events cause important fluvial erosion and deposits. The streambed is frequently modified and this affects the rating curve. Rating curves are valid for several years under stable environmental conditions, but must be updated several times a year in the Rehraya watershed. Rating curve updating is an expensive and time-consuming task when done so frequently and the hydrological services in the Tensift watershed have to monitor a large number of stations under these conditions. Consequently, rating curves are not always updated and streamflow measurements are thus subject to errors. We analyzed streamflow measurements and corrected them when possible to allow reliable time-series.

To ensure that rainfall events are representative of the whole catchment, we selected hydrological years for which there is a good correlation between peak flow and precipitation at the outlet. The years 1980-1981 to 1983-1984, 1990-1991 to 1998-1999 and 2001-2002 were considered.

Streamflow data at the outlet is an important measure since it integrates all the hydrological processes active in the watershed (Winiger et al., 2005). It is also the most commonly available data. As a consequence, model calibration is generally achieved by minimizing the distance between the measured and the simulated streamflow estimates. In this study, calibration is carried out by maximizing the Nash and Sutcliffe efficiency (Nash and Sutcliffe, 1970) which is defined by equation (1) where i is the time-step, N the total number of simulated time-steps, obs_i is observation on time-step i, sim_i is simulation on time-step i and obs_{my} the mean of the observations for the simulated period.

$$Eff = 1 - \frac{\sum_i^N \left(obs_i - sim_i\right)^2}{\sum_i^N \left(obs_i - obs_{my}\right)^2} \quad (1)$$

When implementing a distributed conceptual hydrological model, uncertainty can come from different sources (Pellenq & Boulet, 2004): boundary conditions, initial conditions, model parameters, forcing variables, and model formulation. Concerning the parameters, available data are mostly insufficient to reduce this uncertainty (e.g. uncertainty in the parameters' spatio-temporal distribution and their representative mean; uncertainty in the value of the parameters themselves). When the degree of freedom of the simulation is higher than the degree of constraints –defined by all the available observations- the observation system is under-determined or over-parameterized (Ambroise, 1999). Over-parameterization is common when distributed conceptual models are used. Through model optimization, the issue of over-parameterization is linked with the issue of parameter intercorrelation and parameter dependence. Two parameters are fully independent if the optimal value obtained for one parameter is independent from that of the other parameter. On the other hand, when two parameters are correlated, the same model output can be simulated with different combinations of these parameters. This means that when this output is used to compute the cost function, a large number of solutions represented by best fit couples of parameters are obtained. Under these conditions, improvement of model performance is as likely to result from parameter intercorrelation (existence of many best-fit realistic values) as from any improvement of the representation of physical processes (Batchelor *et al*, 1998). Indeed, Beven and Quinn (1994) showed that a wide range of parameter sets can be fitted so that models are able to reproduce observed data. This has been named the “equifinality” issue (Beven, 1996). Consequently, in order to assess the hydrological modeling performance, several simulated processes should be analyzed individually and not only the resulting streamflow. Moreover, priority should be given to the analysis of processes which are simulated using parameters for which intercorrelation issues have been identified. However, investigating the performance of the model in terms of intermediate processes is often difficult and requires heavy investment in measurement networks. In fact, model outputs represent processes that are controlled by physical properties which vary both in time and space. Apart from discharge data which integrates basin-scale processes, most of the other measurements (such as soil moisture, leaf area index, groundwater recharge, …) represent instantaneous physical properties at a single point/grid. Different solutions have been suggested in the literature to overcome this scale problem. Regarding soil moisture for example, Schmugge and Jackson (1996) show that the simple mean of soil moisture values can be sufficient, and most authors put high hopes on remote sensing techniques that provide large-scale area-averages values. Despite the existence of techniques such as high spatio-temporal resolution satellite sensors and large aperture scintillometers which give access to the spatial and temporal distributions of the process, their implementation remains confined to very specific studies and teams. Most hydrological studies do not have the required budget and skills to implement such instrumentation and rely on more classical measurement devices.

The problem of parameter intercorrelation can be identified by analyzing best fit parameter ranges when one, two or more parameters are tuned to reduce the difference between the simulated and the observed daily streamflow. If the optimal values of one particular parameter are within the same range no matter how many parameters are calibrated, this parameter is not subject to intercorrelation. However, if the optimal values are scattered or if the optimal range changes when the number of tuned parameters increases, intercorrelation should be expected. Sorooshian and Gupta (1983) identify three sources of equifinality issues during calibration: 1) the structure of the model, 2) the inadequacy of the model in representing reality and 3) the data and its associated measurement or scaling error. We have already mentioned issues related to the third source of equifinality. In the following sections we focus on the first two factors.

During the calibration period, a good simulation performance can be achieved with SWAT (see Figure 5: a Nash efficiency of 0.83 is obtained). The good-fit values (values of the parameter that produce 80% of the global maximum efficiency) were analyzed throughout the calibration of an increasing number of parameters. As stated above, the range of best-fit parameters during successive calibrations is a good illustration of parameter intercorrelation. Behaviors of pairs of parameters along the successive calibrations are displayed in Figure 6, Figure 7 and Figure 8. Figure 6 presents the altitudinal gradient of precipitation on the x-axis (“gdt pcp”), the altitudinal gradient of temperature on the y-axis (“gdt tmp”) and the Nash efficiency on the z-axis (“efficiency”). Each point represents a good-fit simulation: a specific combination of both gradients leads to an efficiency greater than 80% of the maximum efficiency. In the first graph (“opt 2p”) only two parameters are calibrated (the two gradients), while in the following graphs, additional parameters are calibrated (three “opt3p”, four “opt4p” up to 8 “opt8p”). Altogether, up to 12 parameters have been calibrated but all graphs are not displayed here. We analyzed the consequence of the calibration of these additional parameters on the values obtained for the altitudinal gradients. In Figures 7 and 8, the same representation is adopted for different parameters. In Figure 7, the depth of one soil type (referred to as "Depth(soil1a)") is reported on the x-axis of the graphs located on the left part of the figure and the depth of another soil ("Depth(soil2)") is reported on the x-axis of the graphs located on the right hand side of the figure. On the y-axis the depth of the third soil type ("Depth(soil1b)") is reported on all graphs. In Figure 8, the x-axis displays the value of the groundwater delay (referred as "Gw delay") whereas the y-axis displays the value of the snowfall temperature ("$T_{snowfall}$").

To identify possible intercorrelation issues, one can interpret the extent of the good fit parameter space. The three figures show different behaviors: in Figure 6 the scatter plot occupies a very limited space; when the fifth parameter is calibrated the range of possible values increases, but the cluster of points is centered around the same space as when only four parameters are tuned. Altitudinal gradients of temperature and precipitation do not seem to be subject to intercorrelation. However in Figures 7 and 8, the scatter is much more important and remains so throughout the successive calibrations. Different parameters have been analyzed following the same approach: the parameters related to the soil compartment such as soil depth (see Figure 7) and available water capacity (not shown), to groundwater (see groundwater delay in Figure 8) and -to a lesser extent- snow modules (see Figure 8), all show a large scatter of points in the two-parameter space and thus evidence of potential intercorrelation. Given the lack of available observation tools for this compartment, at least at the catchment scale, subsurface water transfer has not been addressed in this study. Concerning the groundwater and snow modules, geochemistry and remotely sensed data have been used to understand the processes and make direct comparison between the modeled and measured processes. How close to "reality" are the simulated intermediate processes? We attempt to answer this question in the next sections, alongside a description of the tools and methodologies specifically used or developed to answer it.

In what follows, the calibrated parameters are set according to the calibration exercise and remain unchanged. After calibration, a validation exercise was performed for various hydrological years displaying contrasting hydrological conditions. Table 1 presents the mean streamflow and simulation performance (the correlation coefficient, the Nash efficiency and the Root Mean Square Error -RMSE) for the six selected years. The mean streamflow ranges from 0.64 $m^3.s^{-1}$ to 1.83 $m^3.s^{-1}$. Correlation coefficients range from 0.43 to 0.75 but the Nash efficiencies are mostly negative. This demonstrates a poor simulation performance. Figure 9a presents precipitation together with the measured and simulated streamflow for the years 1981-1982. The simulation reproduces the seasonal pattern of streamflow fairly well (correlation coefficient=0.64) but some major events are missing at the beginning and at the end of the simulation period. This explains the negative Nash efficiency (-0.12) and the high RMSE (2.09 $m^3.s^{-1}$) of that particular year. Figure 9b presents the simulated streamflow and the measured precipitation and streamflow for 1995. The simulated hydrograph, although reproducing the shape satisfactorily (correlation coefficient=0.59), is shifted by about one month: the high flow period begins and ends one month earlier in the simulation compared with the observations. With no major peak flows missing throughout the simulation, the Nash efficiency is positive (0.11). In both cases we observe poor modeling performance for different reasons (either peak flows omitted or a one-month shift). Semi-arid mountainous environments face extreme spatial variability of meteorological conditions which cannot be taken into account if an appropriate observation network is not available. The hydrological community should put considerable effort into developing observation network but this remains a difficult and expensive task rarely fulfilled by national hydrology services especially in developing countries.

INTERMEDIATE PROCESSES: METHODOLOGY AND RESULTS

1. Geochemical analysis of groundwater contribution to streamflow

1.1. Methodology

A river's geochemical signature gives some information on the origin of its water. Its composition depends on the reservoir it originates from (whether it is the surface, sub-surface or deep reservoir). Geochemical sampling can thus be used to identify the contributing reservoirs during a flood and to quantify their relative contribution to annual streamflow. In this study, we analyzed the silica and Dissolved Organic Carbon (DOC) content of water with respect to streamflow. These two elements were chosen for two reasons. Firstly they are not significantly influenced by atmospheric contamination, since precipitation contains almost no silica or dissolved organic carbon (Probst *et al*., 1990). Secondly they have different origins (Idir *et al*., 1999, Tardy *et al*., 2004): whereas silica originates from mineral alteration and its concentration increases when water flows through the deepest soil horizons, dissolved organic carbon is released from leaching of superficial soil layers rich in organic matter. Therefore, these elements can be used as tracers for deep and superficial reservoirs respectively. Water coming from deep reservoirs is characterized by high silica and low dissolved organic carbon concentrations whereas water coming from the superficial reservoir presents the inverse geochemical signature. To separate the hydrograph into contributions from both reservoirs, we applied the method of Pinder and Jones (Pinder and Jones, 1969) which has been used in many other studies (Idir *et al*., 1999, Pilgrim *et al*., 1979, Probst, 1992, Tardy *et al*., 2004). Following this method, geochemical analysis is only required at the outlet and the contribution of the different reservoirs can be related to streamflow which has to be measured continuously. The method determines, at each sampling time *t*, the individual contribution of each reservoir *k* ($Q_k(t)$ in $m^3.s^{-1}$) to the total streamflow ($Q_t(t)$ in $m^3.s^{-1}$) as specified by equation (2). The main assumption of the method is that the contribution of the different reservoirs (from 1 to *k*) changes with time independently of its geochemical characteristics (C^i_k is the concentration in geochemical element *i* of reservoir *k*) (see equation 3).

$$Q_t(t) = \sum_{k=1}^{K} Q_k(t) \qquad (2)$$

$$Q_t^i(t) \times C_t^i(t) = \sum_{k=1}^{K} C_k^i \times Q_k^i(t) \quad (3)$$

Water samples were collected on a fortnightly basis at the outlet of the watershed from April to December 2003. Common assumptions are that the contribution of the superficial reservoir is dominant during the high flow period (Probst, 1992) whereas the deep reservoir is the only one contributing to streamflow during the low flow period (Smakhtin, 2001). The geochemical composition of each reservoir can thus be directly characterized by the geochemical signature of streamflow at these specific periods of time. Once the geochemical composition of the reservoir is known, the relative contribution of each reservoir can be calculated at each sampling time and then extended throughout the year by establishing a relation between the contribution of one reservoir and the total streamflow. This method reveals the contribution of deep and superficial reservoirs to streamflow. In the following section, these contributions were compared to the deep reservoir contribution simulated by the SWAT.

1.2. Results

The geochemical compositions of both reservoirs are displayed in Table 2. The superficial reservoir is characterized by a high content in DOC and low content in silica whereas the deep reservoir presents the inverse geochemical signature. The linear regression established to estimate the contribution on a continuous basis is presented in Figure 10. Consequently, the reservoirs contributions to streamflow were identified and plotted in Figure 11. The deep reservoir contribution (thick line) is very stable and equal to the summer streamflow. The superficial reservoir (diamond symbols) shapes the hydrograph. The partition of the contributions is consistent with the basin features, i.e. a steep mountainous watershed with shallow soils. The geochemical measures show that the deep reservoir contribution could be simulated as a constant value throughout the year.

SWAT simulations of groundwater contribution to streamflow present consistent patterns from one year to the next (see Figure 12.a to c). Very little groundwater contribution is simulated during low flow period but it increases slightly during high flow period. The groundwater contribution to streamflow is expressed in SWAT as an increasing function of an average soil water content with a proportionality factor; in our example it follows the streamflow signal pattern with no contribution from summer to December. In some cases though, what is simulated is a small increase in late winter and generally a gradual increase in spring followed by gradual decrease one or two months later. Overall however, the simulated groundwater contribution is small since the soils are shallows and exhibit a very low water storage capacity. This pattern is not consistent with the results from geochemical analysis. The formulation adopted by SWAT's groundwater module is thus not appropriate for the characteristics of the mountainous watershed under investigation: even if the amplitude of the groundwater signal were to increase through increased soil storage capacity, it would remain mathematically impossible to simulate a constant groundwater release with a strict application of an average Darcy-like equation. Other authors (Conan *et al*., 2003; Sophocleous & Koelliker, 1999) reported similar conclusions in different climatic and geographical conditions.

2. Remote sensing for snow surface identification

2.1. Methodology

As mentioned earlier, the watershed culminates at 4167 m. Each year, snow covers the upper parts of the basin. There is no snow cover monitoring network on the site. Implementing one would represent significant investment due to the remoteness of the region and the cost of installing snow water equivalent measuring devices or snow height monitoring stations. Remote sensing is thus a useful tool for snow surface monitoring in high mountainous area where field instrumentation is difficult to implement and expensive to maintain. In semi-arid mountainous zones where snowmelt dynamics are very rapid (see fig. 4 in Schulz and de Jong, 2004) and the terrain is highly heterogeneous (slope and aspect, vegetation cover and types), sensors presenting a high temporal frequency and high spatial resolution are required. However such sensors do not yet exist. To overcome this difficulty, we developed a method combining high resolution (LANDSAT-TM images, 30m of resolution) and low resolution (SPOT-VEGETATION images, 1km resolution) satellite images (Chaponnière *et al*., 2005b). This method establishes a relationship between the snow index at low resolution and the snow surface derived from the classification of high resolution images. In this research, a new snow index specifically adapted to the Atlas Mountains was developed. It corrects for the influence of the soil spectral signature/noise in the original snow index. However, information on snow depth cannot be retrieved from these wave lengths. The main limitations of the methodology are (see Chaponniere *et al*., 2005b): the coarse spatial resolution of the images (1 km), the high scatter of low values in the index-to-area relationship and the temporal intervals of the low resolution images (SPOT-VGT), which are often longer than one or two days. As a consequence, uncertainty of the snow surface estimates are difficult to quantify. The features related to snow cover dynamics that can be estimated with the highest degree of confidence are i) the start and end dates for the snow cover and ii) the maximum values of snow cover during the season. In the next section the temporal profiles obtained from satellite imagery will be compared to the profiles simulated by the SWAT model.

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2.2. Results

Time series covering the hydrological years 1998-99 to 2001-02 were processed with the above-mentioned remote sensing methodology (see Chaponniere *et al*. 2005b for details). The snow cover (in km^2) over the basin obtained from satellite images was considered as the reference (or "ground truth") but was subject to a large number of uncertainties as stated above. The data was compared to the snow-cover simulated by SWAT to investigate how the model is able to reproduce this secondary process. In SWAT the snow water equivalent SNO for a given HRU is converted to snow surface sno_{cov} via Equation 4. SNO_{100} is the threshold snow water equivalent for which 100% coverage is reached, cov_1 and cov_2 design the shape of the curve that characterize the distribution of snow within the HRU. Runoff is not affected by SNO_{100} but snow surface is highly sensitive to it.

$$sno_{\text{cov}} = \frac{SNO}{SNO_{100}} \times \left(\frac{SNO}{SNO_{100}} + \exp\left(\text{cov}_1 - \text{cov}_2 \times \frac{SNO}{SNO_{100}} \right) \right)^{-1} \quad (4)$$

Modification of SNO_{100} (from the default value to a measured or calibrated value) greatly influences snow dynamics. For the year 1998-99, this parameter was calibrated ('simu opt' on Figure 13) to fit the observed snow cover. Figure 13 shows time series of snow cover simulated by SWAT using the default parameter of the snow-distribution equation ('simu ref', continuous line); simulation with the calibrated parameter ('simu opt', dashed line) and satellite observation ('satellite', diamond symbols). As far as the seasonal pattern is concerned (snow accumulation, maximum cover and snow disappearance periods), both simulations are close to satellite observations. Snow surface is present from early December to late June. Despite differences between the simulated and the observed time series in the calibration data set, the calibration of the snow distribution equation improved the results in terms of simulated snow surface and dynamics. The calibrated equation was then applied for the additional time-series in the validation data set. Figure 14.a to c. shows simulated snow cover using either calibrated or default parameters, as well as satellite observation for years 1999-2000, 2000-01 and 2001-02, these three years representing our validation data set. The calibrated equation shows an overall satisfying fit with all observations and a significant improvement compared to the default equation in terms of estimated snow surface and timing of snowfall and snowmelt. We thus conclude that the formalism adopted in SWAT for the simulation of snow cover dynamics is appropriate for our watershed. Of course, important differences remain between simulated and satellite-retrieved snow cover temporal profiles for the reasons presented above (uncertainty of "observation" and simplicity of simulation).

The correlation between the remotely sensed snow cover and SWAT's outputs presents some limitations. Indeed, some snowfall events are not simulated at all (the first and last events in figure 13, the last events in Figure 14 a and 14 b, the first event in Figure 14 c) due either to a unobserved precipitation event in the rainfall data set or to high generated atmospheric temperature preventing rainfall from falling as snow. Also, the typical decline of snow cover is not well reproduced in the model especially in the spring time. The model presents a scale-shaped curve characteristic of models using thresholds. This is the case for SWAT's snow module, based on the degree-day method. Comparison of its performance against that of an energy balance model can be found in Chaponniere (2005a).

As predicted by the sensitivity study, the use of either the default or the calibrated equation has very little impact on the runoff simulation. Increased accuracy in the simulation of this intermediate process does not affect the runoff simulation quality.

DISCUSSION

The watershed studied here is complex and different storage and redistribution processes ("intermediate processes") contribute to the streamflow signal at the outlet. Significant surface runoff is expected on the steep slopes and shallow soils during intense rain events. Rapid overland flow contributes to streamflow within a few hours to a few days. In terms of "slow processes", groundwater and snow melt contribute to streamflow at a time scale of several weeks to months. There are thus at least three major contributions to streamflow which are characterized by seemingly distinctive temporal signatures. The modeling exercise shows that the processes which might present high equifinality issues include soil, groundwater and snow modules. The problems inherent to the evaluation of soil parameters have not been addressed here, but could definitely provide some important insight in the rapid-streamflow response of the basin. The groundwater and snowfall/snowmelt processes have been analyzed in more detail in order to estimate how realistically these "intermediate processes" are simulated by the model. These processes have similar temporal signatures on the streamflow time series and are intercorrelated: it is thus important to be able to analyze how realistically they are modeled. We found that the algorithms used in the SWAT to compute the groundwater contribution to streamflow are not appropriate. Modeling them as a simple constant contribution would be enough in our zone of interest. Regarding the snow cover, it is fairly accurately reproduced by the model once the distribution equation is calibrated.

The modeling presented in this study considered three HRU for the whole watershed. This division was chosen based on geomorphological observations and the analysis of soil and geological maps. It reflects the major geomorphological units of the watershed and the number of units seemed reasonable to us given the lack of information on the hydrological cycle of the basin. However, this number is low and does not reflect the complexity of the watershed. Considering a higher number of HRU would be more realistic but would bring up equifinality issues. Additional details in the modeling would first require a better hydrological characterization of the watershed.

The study takes place in a semi-arid environment where high water losses are expected: via evapotranspiration for the vegetation and via sublimation for snow. Snow penitentes (pointed peaks of hard snow) are signs of a high sublimation rate. On the southern slopes of the High Atlas, Schulz and de Jong (2004) observed penitents. Modeling snow ablation with an energy balance model, they found an average 44% of snow removed by sublimation. The modeling was conducted at two sites located at 3250m and 2960m during 39 days and 20 days respectively. Sublimation is thus an important component to take into account. However, we believe the sublimation rate in the Rehraya watershed to be less important because northern slopes experience less arid climatic conditions than southern slopes and because we observed only a few penitents above 4000m (whereas Schulz and de Jong observe more and at lower elevation). In SWAT, evapotranspiration is simulated with the Penman equation and potential evaporation applies directly to snow which sublimates to fulfill the demand (which equals half the potential evaporation) unless the quantity of snow is below a given threshold. The impact on the runoff retrieval accuracy of using in situ climatic data or generated climatic data from long-term historical mean is low, as demonstrated in Chaponniere (2005a). However, the potential and actual evapotranspiration have not been compared to ground measurements via scintillometry for example. Andréassian et al. (2004) discuss the impact of potential evaporation errors on model efficiency and provide a very complete literature on this issue. Further investigation of this compartment in the Rehraya watershed would definitely provide a valuable input.

CONCLUSION

In Southern Morocco, large irrigated districts rely on the water coming from the Atlas Mountain. However, water transfer mechanisms from the mountains down to the plains remain poorly understood. An understanding of these mechanisms is of crucial importance for developing management strategies that ensure the sustainability of irrigation under the currently changing environment. In this paper, the water cycle of a poorly instrumented semi-arid mountainous watershed is simulated. The main aim of this work was to analyze whether optimal parameters sets –calibrated using a cost-function based on the streamflow data- are consistent with a realistic process representation or whether they only reflect parameter intercorrelation. The Soil and Water Assessment Tool was implemented for the Rehraya watershed. We found that evaluating the model performance by comparing simulated and observed streamflow only is insufficient especially when parameter intercorrelation is identified. The analysis of best-fit parameter ranges when tuning an increasing number of parameters shows that compensation effects seem to take place in the soil, the groundwater and the snow modules, i.e. for the quickest and the slowest water transfer processes. The evaluation of soil parameters was not addressed in the present study. To be able to accurately separate the influence of the slowest processes on streamflow, especially during low-flows, special tools and methodologies tailored to the situation (remote access, low level of instrumentation, high variability, etc) have been used. A geochemical analysis was performed for the groundwater module and a remote-sensing methodology was developed for the snow module. The outputs from the model were compared to the “observations” corresponding to the individual processes, and the quality of the simulation of the process was assessed. The geochemical method shows that the groundwater contribution to streamflow of this watershed is low and constant (equal to low flow) throughout the year whereas the model simulates this contribution as a portion of the total streamflow thus following the general streamflow pattern. As for the snow module, the simulated evolution of snow cover is compared to the satellite-retrieved snow extent. The distribution equation of the model used to compute snow cover is adapted to the watershed. We showed on the one hand that the groundwater module was not appropriate for our mountain watershed and on the other hand, that simulated snow distribution better reflects the observed patterns after calibration of the snow distribution equation. To be taken further, the study would need to analyze soil parameters, which would provide valuable insight into the rapid-streamflow response of the basin. Also a detailed analysis of the evapotranspiration module would bring valuable inputs. Benefiting from the information on different individual processes, modification of the model would lead to realistic and fully validated modeling. The multi-disciplinary approach adopted here to increase the insight into the hydrological processes is supported by the hydrological community (Sivapalan *et al*., 2003) but still uncommon. More efforts should be dedicated to these kinds of approaches since they give access to information which is not available in major parts of the world where measurement networks do not exist.

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FIGURE LEGEND LIST

Figure 1 : The Tensift watershed situated in southern Morocco and the Rehraya sub-watershed situated south of the Tensift watershed, in the Atlas mountain range. Tahanoute, the outlet of the watershed, is located north of the Rehraya.
Figure 2: Mean monthly precipitation (in mm) and streamflow (in $m^3.s^{-1}$) and standard deviation at the outlet of the Rehraya watershed on the period 1971-2002.
Figure 3 : SWAT's modeling scheme for the water balance. Initial soil water is known (SW_0). During a rainfall event (PCP), infiltration (1) and percolation (2) are the two first simulated processes: surface runoff ($Q_{surface}$) is then established and soil water actualized (SW_{act}). A second step consists in calculating transpiration (3) and evaporation (4) fluxes in response to the atmospheric evapotranspiration demand (Potential Evapo-Transpiration or ETP). Soil water is updated (SW'_{act}) and sub-surface runoff is calculated ($Q_{sub-surface}$). Finally, the updated soil water provides the percolation flux towards the deep reservoir and the contribution of this reservoir to the streamflow (Q_{base}).
Figure 4 : HRU distribution on the Rehraya watershed.
Figure 5 : Simulated (dashed line) and measured (thick line) streamflow ($m^3.s^{-1}$) together with precipitation (mm) for year 1990-91
Figure 6: Values of temperature ("gdt tmp") and precipitation ("gdt pcp") altitudinal gradients leading to a modelling performance higher or equal to 80% of the maximum efficiency when 2, 3, until 12 parameters are calibrated
Figure 7 : Values of soil depth ("Depth(soil1b)", "Depth(soil1a)", "Depth(soil2)") leading to a modelling performance higher or equal to 80% of the maximum efficiency.
Figure 8 : values of snowfall temperature ("$T_{snowfall}$") and groundwater delay ("Gw delay") leading to a modelling performance higher or equal to 80% of the maximum efficiency.
Figure 9 : Simulated (dashed line) and measured (thick line) streamflow ($m^3.s^{-1}$) together with precipitation (mm) for year a. 1981-82 and b. 1995
Figure 10: Linear relationship between total streamflow (Qtotal) and streamflow from the superficial reservoir (Qsuperficial) established from point measurements in order to define Qsuperficial throughout the year.
Figure 11: Rehraya watershed hydrograph (in $m^3.s^{-1}$) between April and December 2003. The total streamflow is represented by a dashed line, the superficial reservoir contribution to streamflow is represented by diamond and the groundwater reservoir contribution is represented by a thick line.
Figure 12: Total streamflow (dashed line) and groundwater streamflow (thick line) simulated with the SWAT model for year a. 1981-82, b. 1990-91 and c. 1996-97.
Figure 13: Estimated surface snow temporal profiles (in km²) for year 1998-99 by satellite images ('satellite', diamond symbols), reference SWAT simulation ('simu ref', continuous line) and calibrated SWAT simulation ('simu opt', dashed line)
Figure 14: Estimated surface snow temporal profiles (in km²) by satellite images ('satellite', diamond symbols), reference SWAT simulation ('simu ref', continuous line) and calibrated SWAT simulation ('simu opt', dashed line) for year a. 1999-2000, b. 2000-2001, c. 2001-2002

TABLE LIST

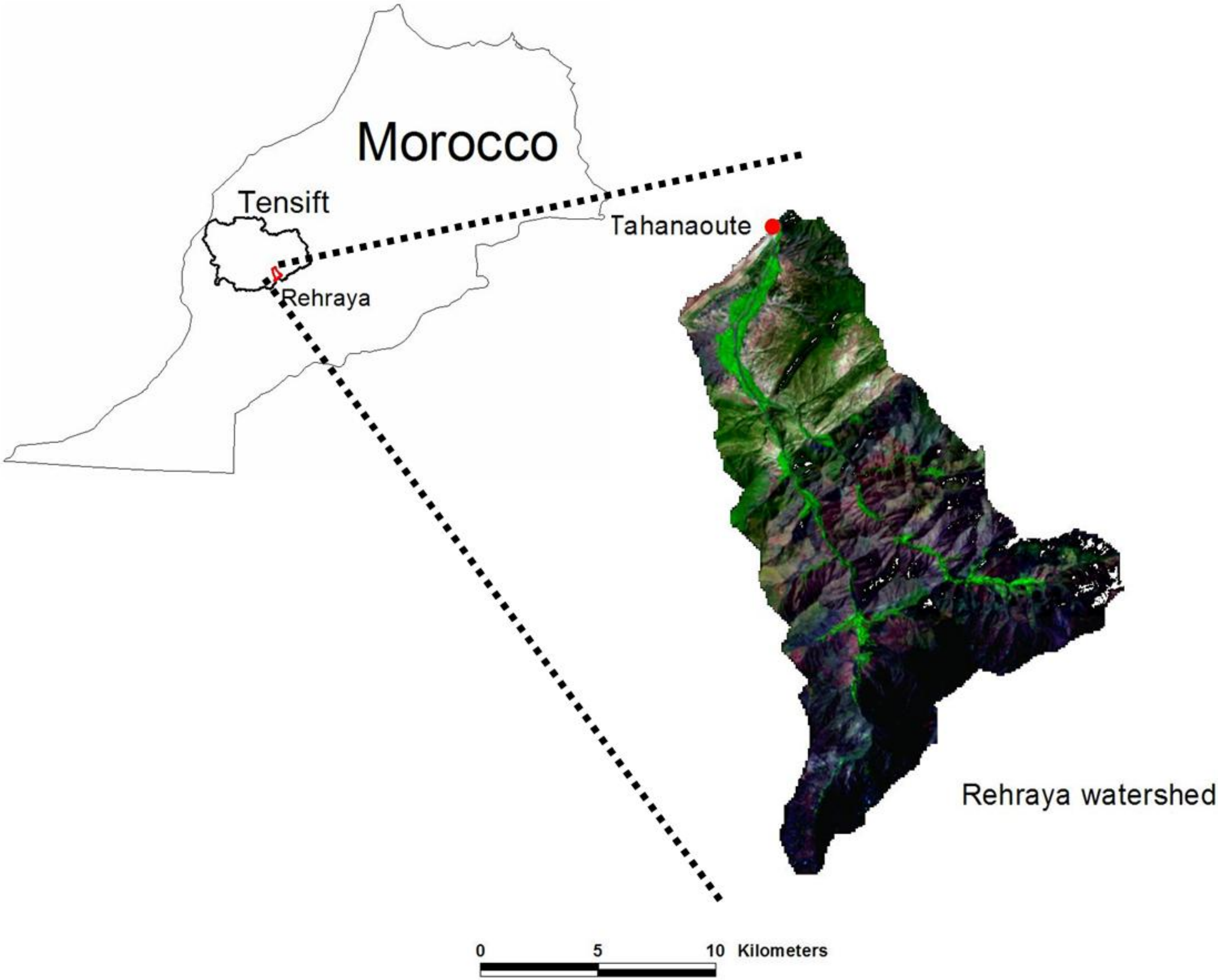
Table 1: Simulation criteria (correlation coefficient, Nash efficiency and RMSE) obtained on validation years
Table 2: Geochemical profile (sodium and dissolved organic carbon) of both reservoirs

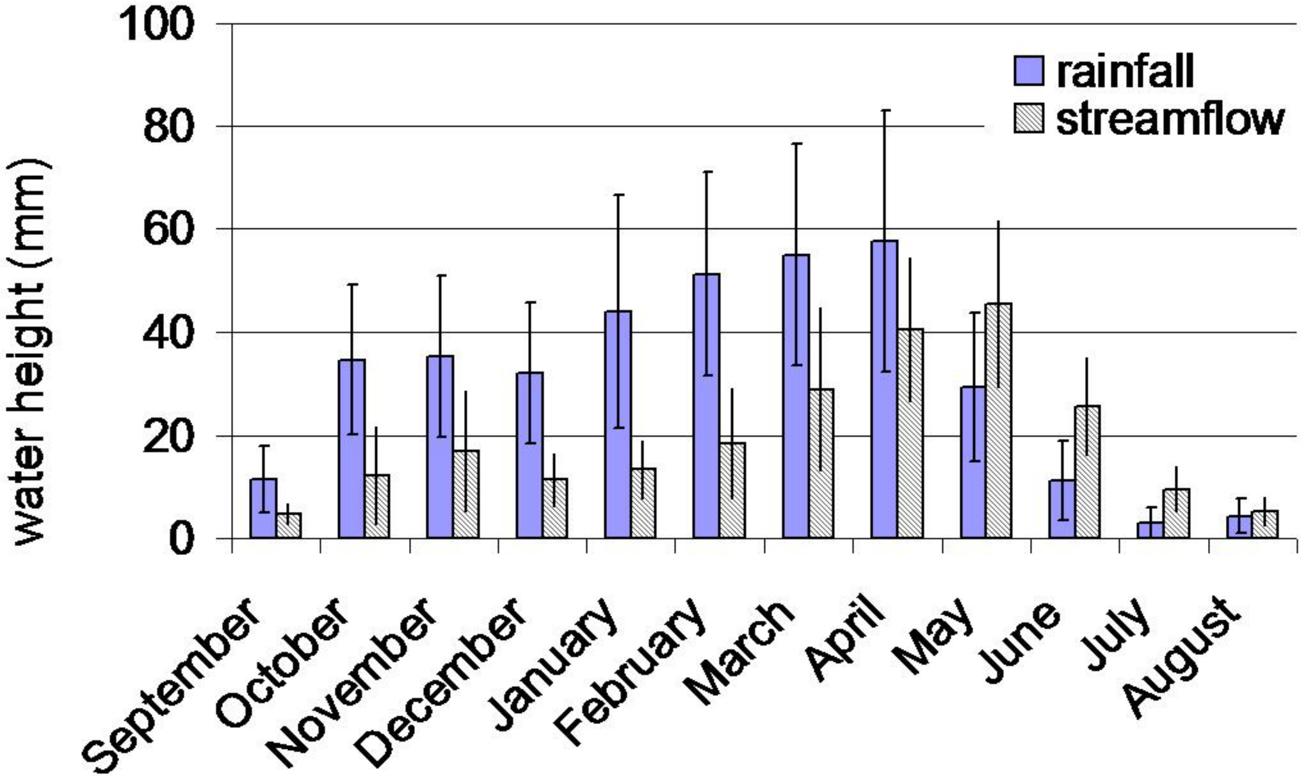
Year	Correlation coefficient	Nash efficiency (-)	RMSE ($m^3.s^{-1}$)	Mean streamflow ($m^3.s^{-1}$)
1980-81	0.43	-0.07	1.42	1.18
1981-82	0.64	-0.12	2.09	1.25
1983-84	0.75	-0.53	1.30	0.64
1993-94	0.43	-1.19	3.65	1.83
1995	0.59	0.11	1.23	1.33
1996-97	0.66	-7.34	2.72	0.88

Table 1: Simulation criteria (correlation coefficient, Nash efficiency and RMSE) obtained on validation years

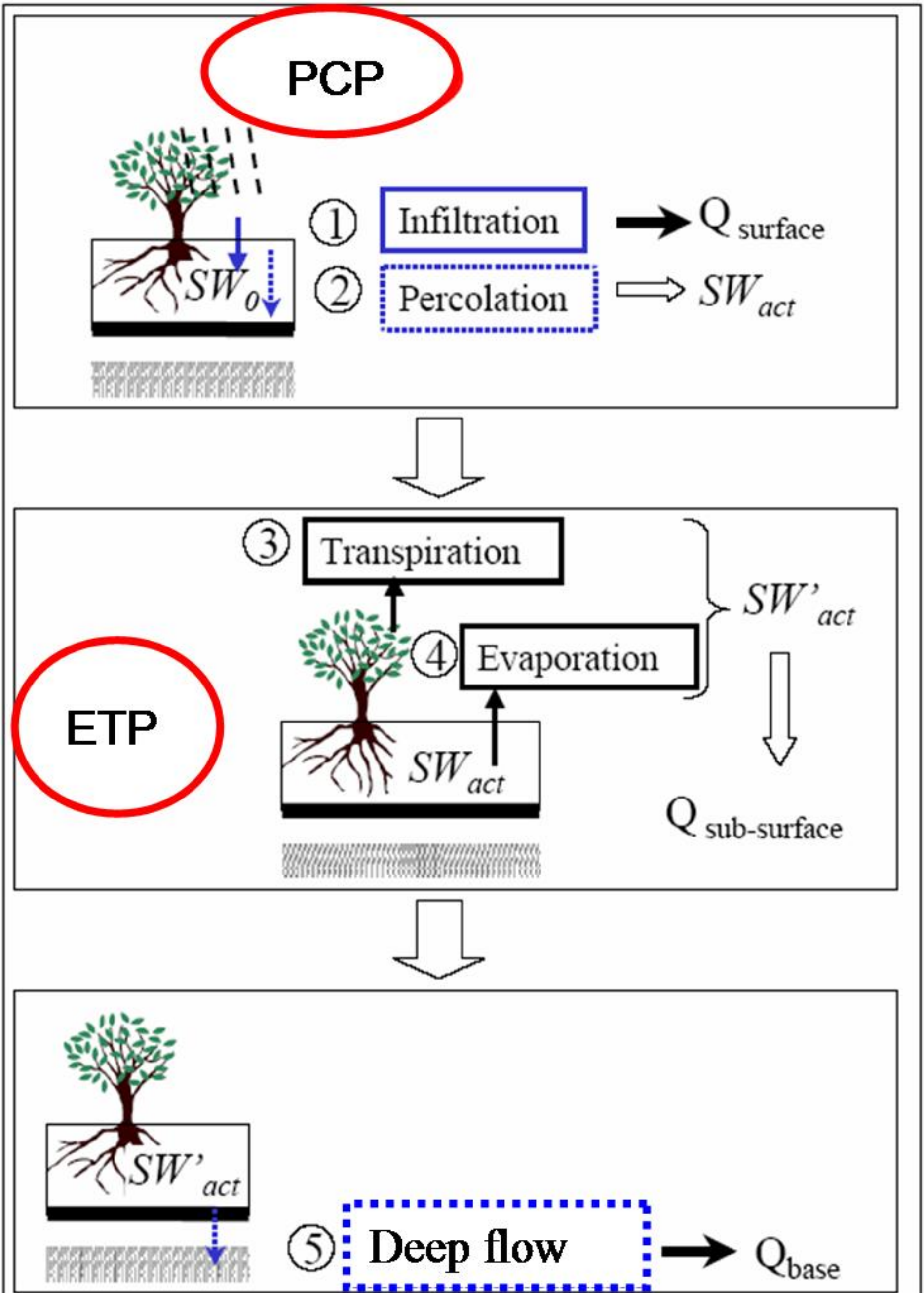
Composition	Si (ppm)	COD ($mg.L^{-1}$)
Superficial reservoir	4.57	2.69
Deep reservoir	5.89	0.87

Table 2: Geochemical profile (sodium and dissolved organic carbon) of both reservoirs

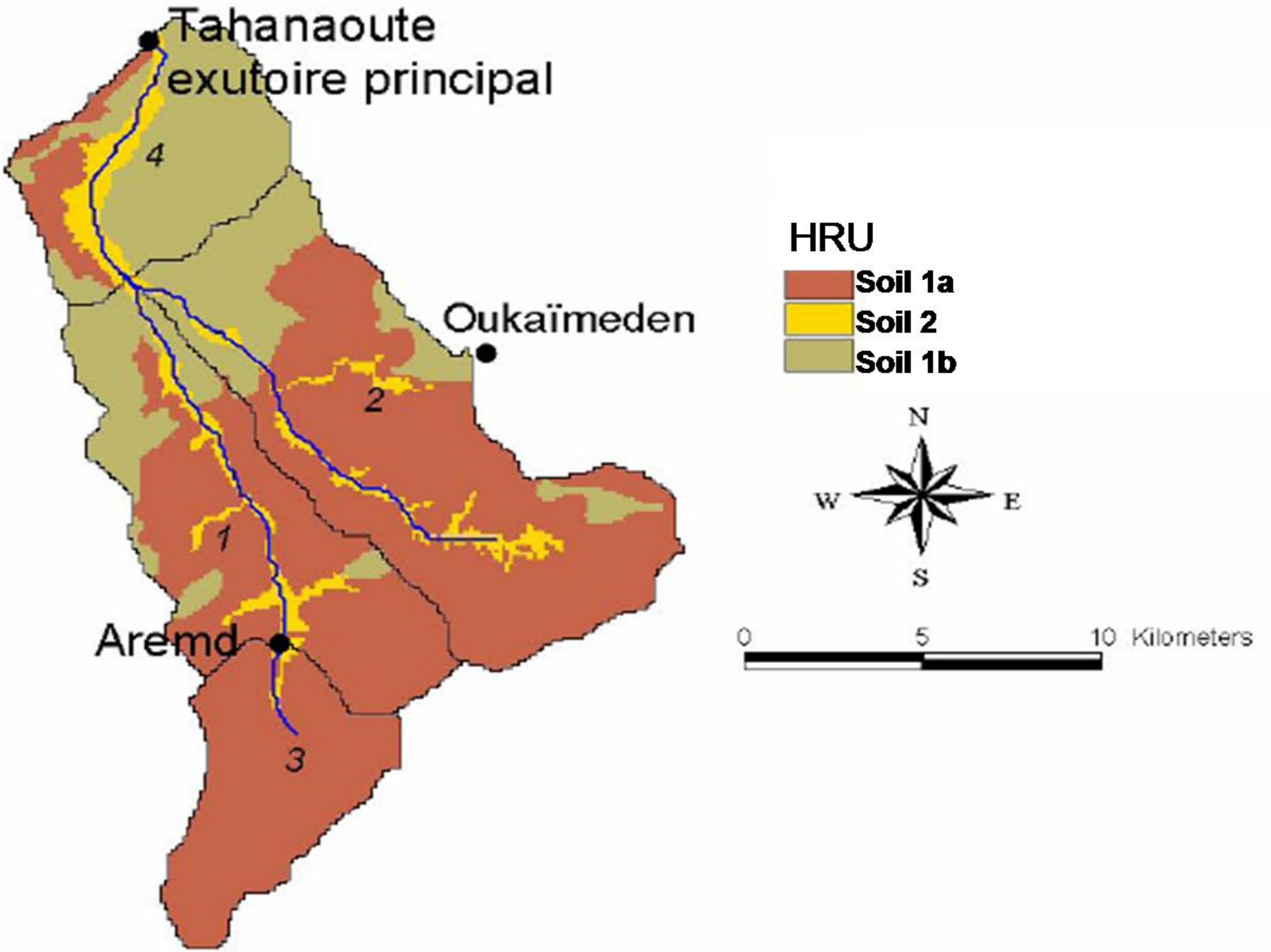

Morocco
Tensift
Rehraya
Tahanaoute
Rehraya watershed
0
5
10
Kilometers

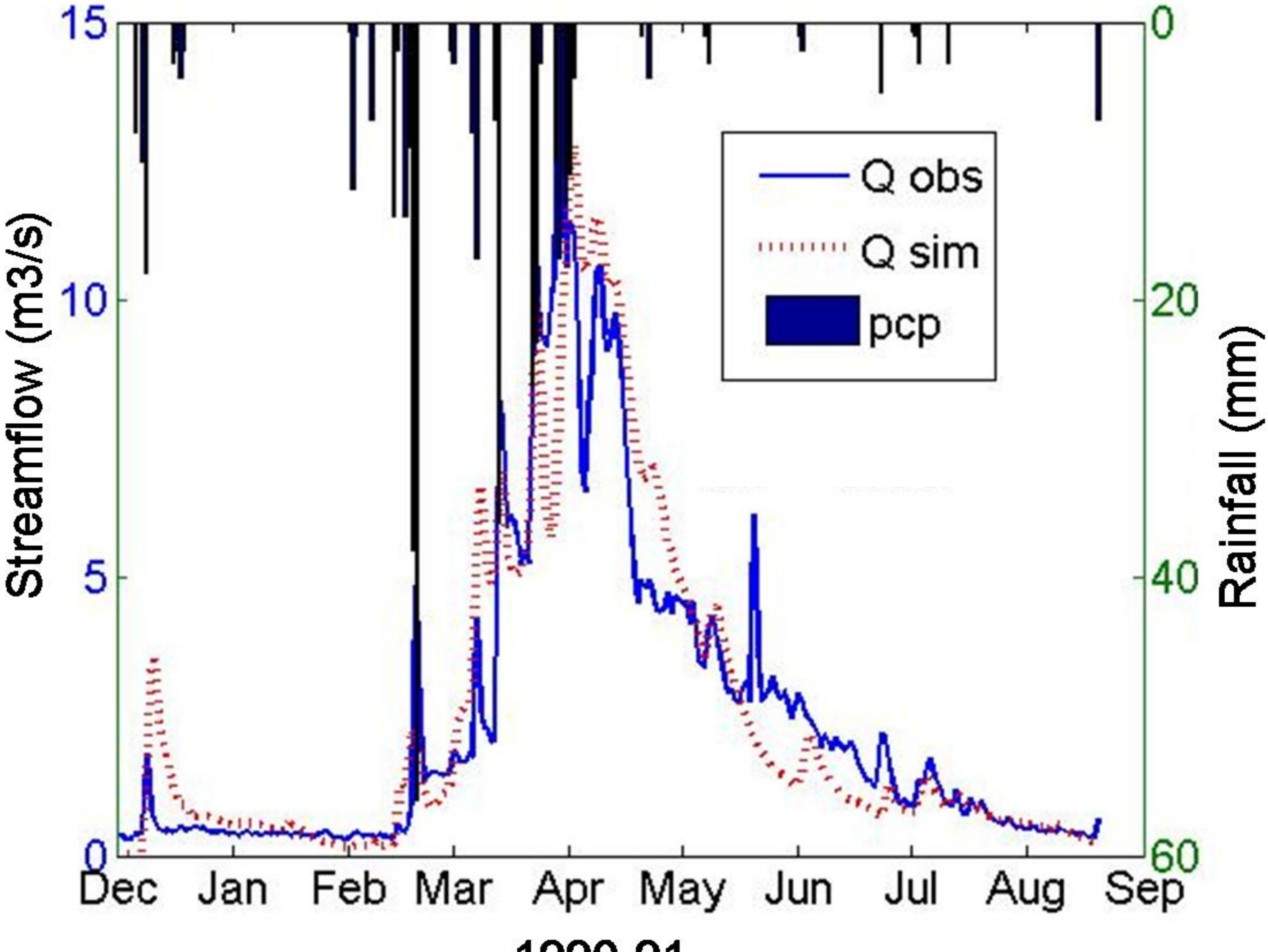


100
80
60
40
20
0
water height (mm)
rainfall
streamflow
September
October
November
December
January
February
March
April
May
June
July
August

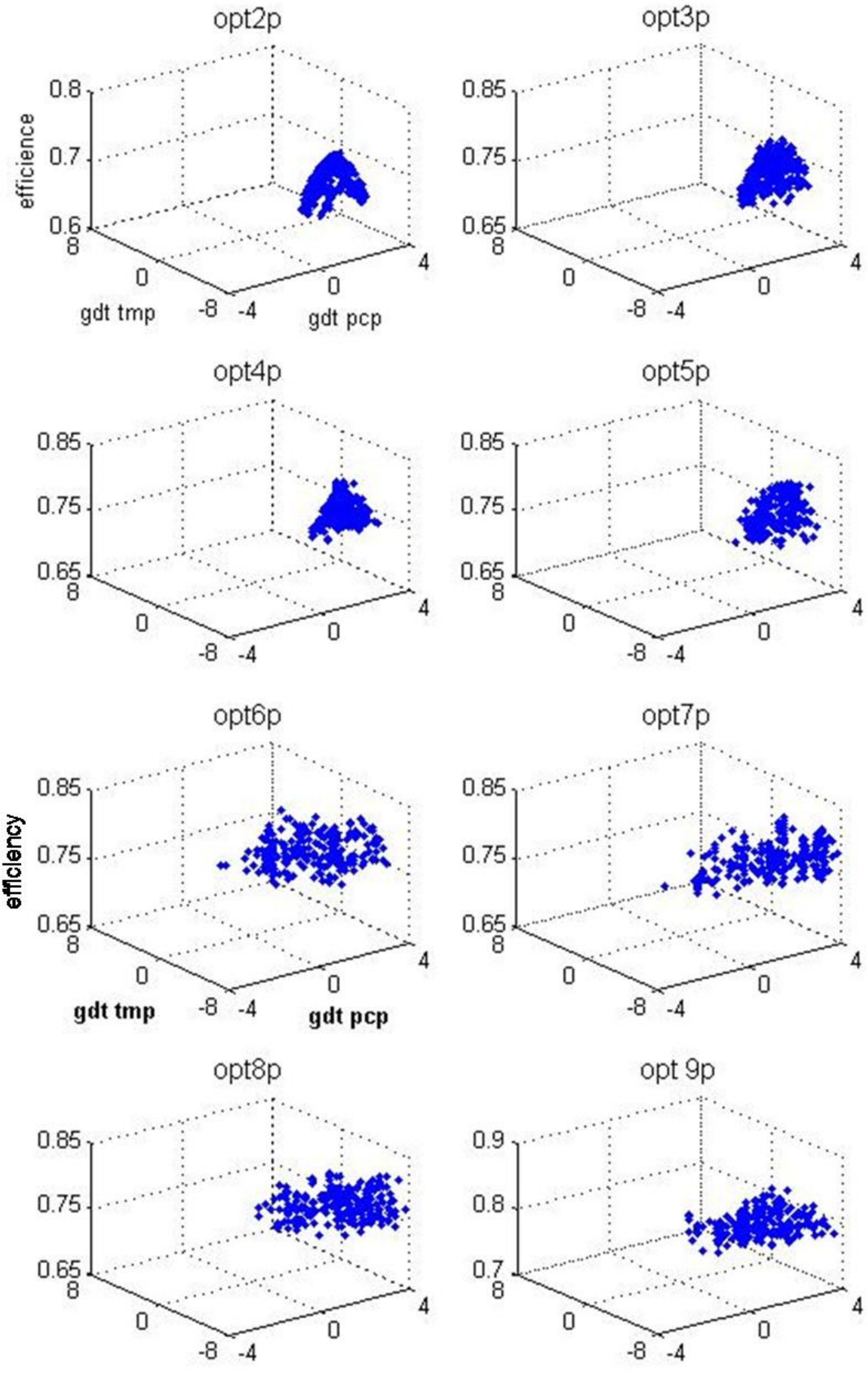


PCP
① Infiltration
$Q_{surface}$
SW_0
② Percolation
SW_{act}
③ Transpiration
④ Evaporation
SW'_{act}
ETP
SW_{act}
$Q_{sub\text{-}surface}$
SW'_{act}
⑤ Deep flow
Q_{base}

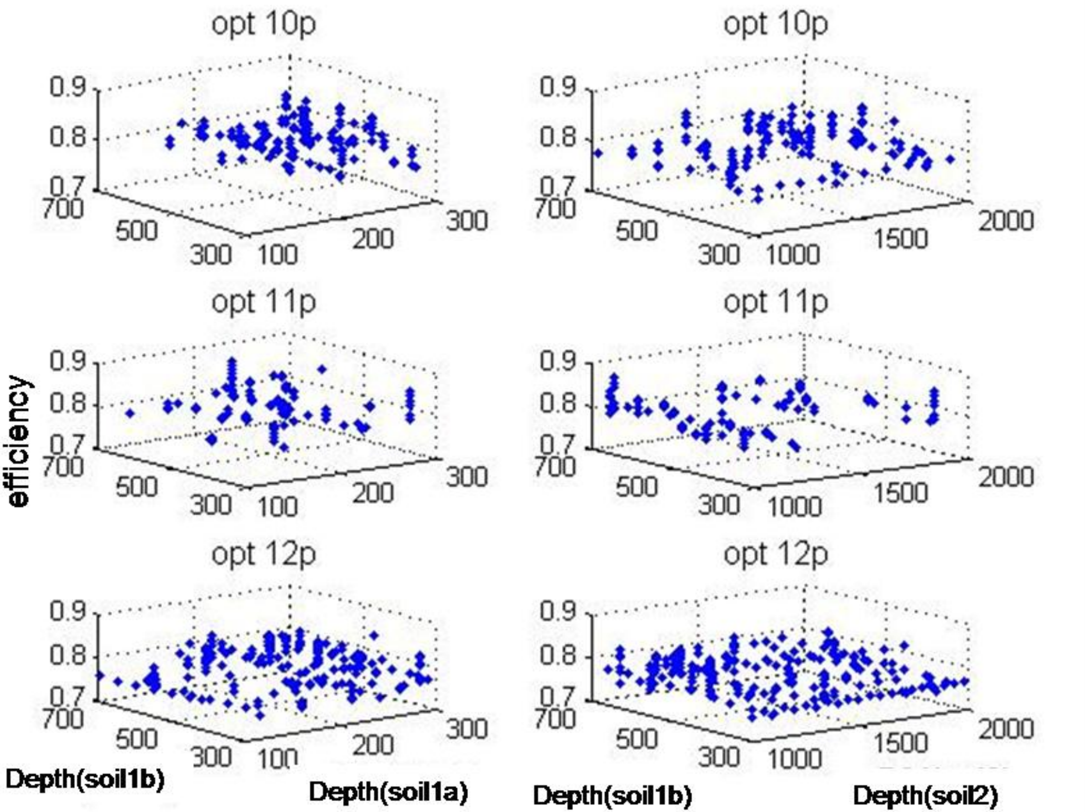

Tahanaoute
exutoire principal
4
Oukaïmeden
2
1
Aremd
3
HRU
Soil 1a
Soil 2
Soil 1b
N
W
E
S
0
5
10 Kilometers



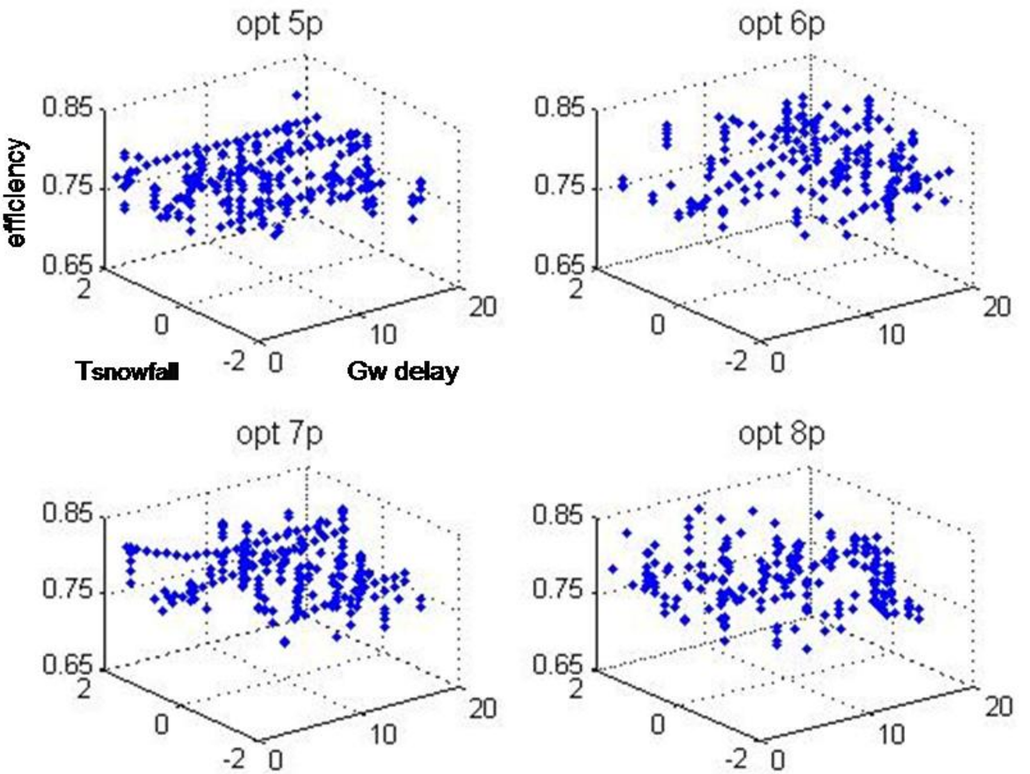
Streamflow (m3/s)
Rainfall (mm)
15
10
5
0
0
20
40
60
Q obs
Q sim
pcp
Dec
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
1990-91



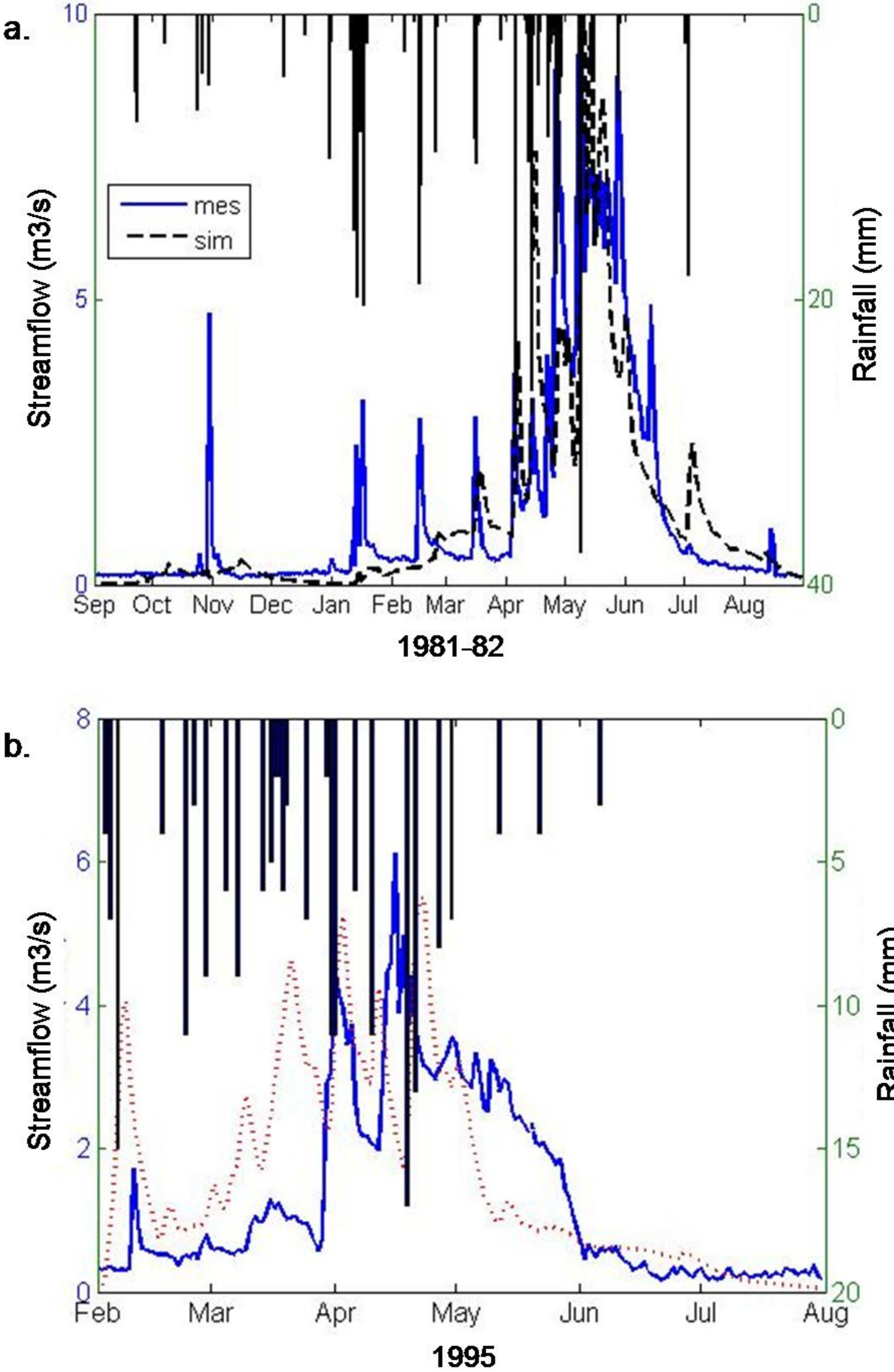
opt2p
opt3p
opt4p
opt5p
opt6p
opt7p
opt8p
opt 9p
efficience
efficiency
gdt tmp
gdt pcp
0.8
0.7
0.6
0.85
0.75
0.65
0.9
8
0
-8
-4
4



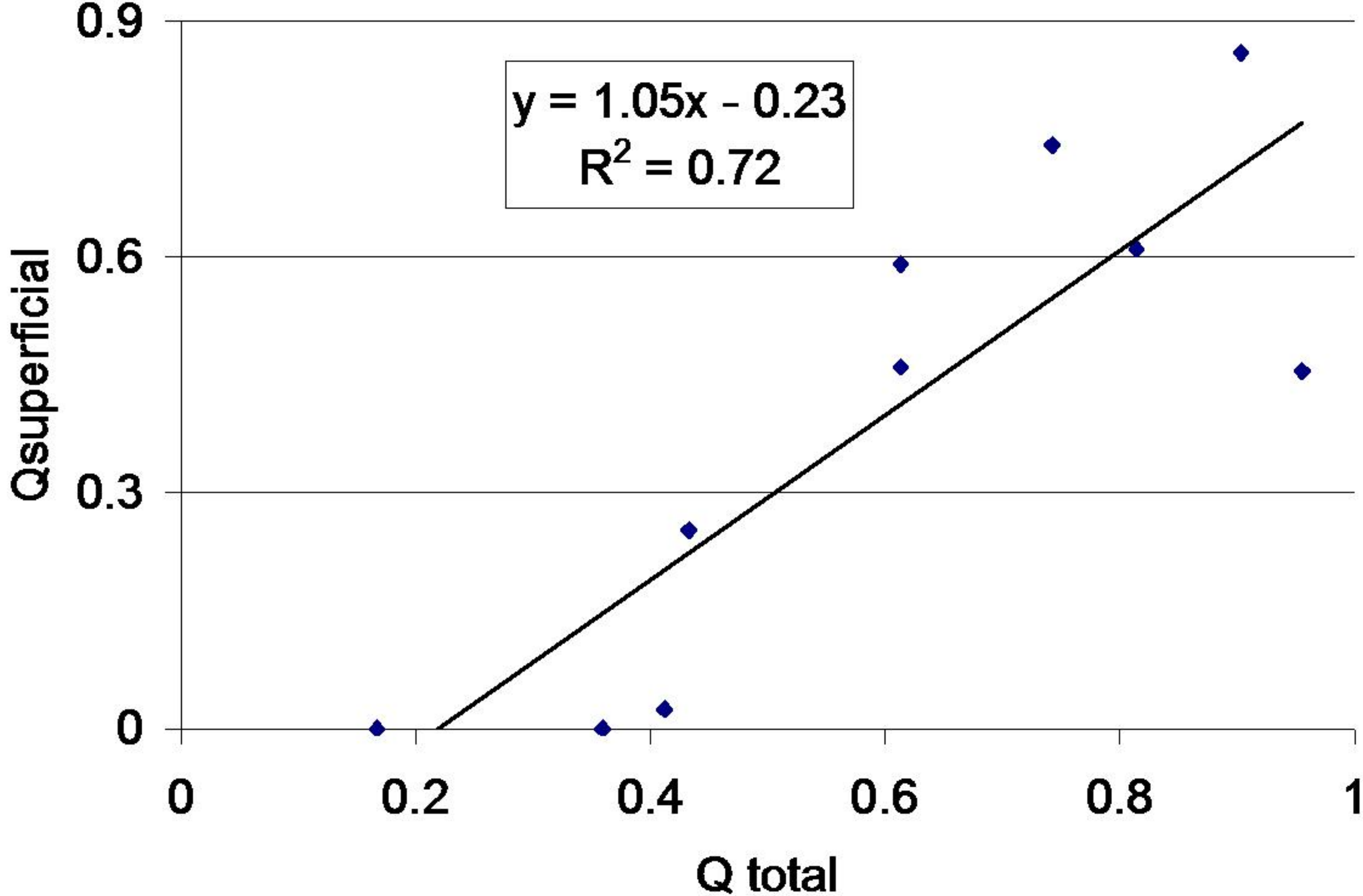
opt 10p
opt 10p
opt 11p
opt 11p
opt 12p
opt 12p
efficiency
0.9
0.8
0.7
700
500
300
100
200
300
1000
1500
2000
Depth(soil1b)
Depth(soil1a)
Depth(soil1b)
Depth(soil2)



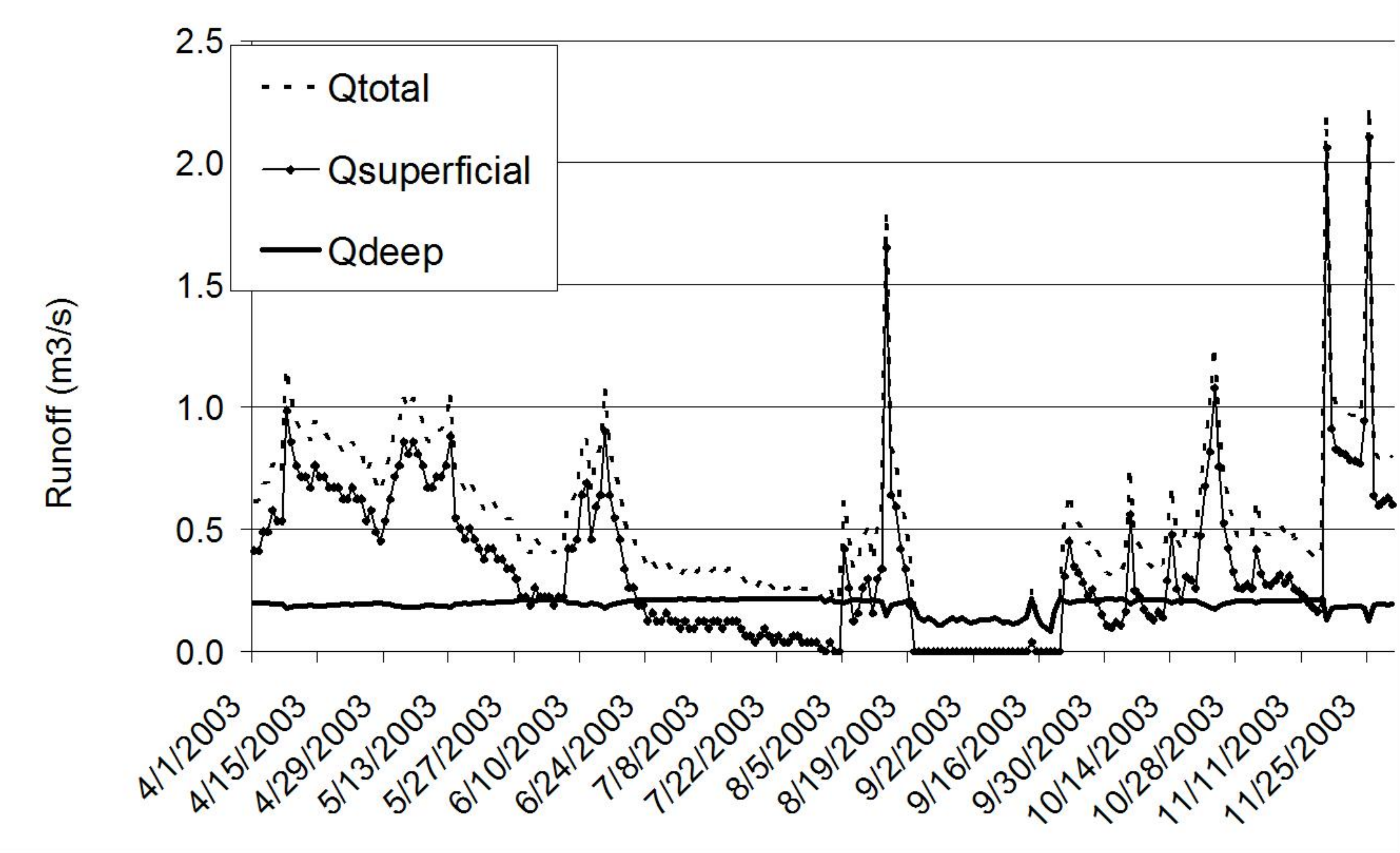
opt 5p
opt 6p
opt 7p
opt 8p
efficiency
Tsnowfall
Gw delay
0.85
0.75
0.65
2
0
-2
0
10
20



a.
mes
sim
Streamflow (m3/s)
Rainfall (mm)
Sep Oct Nov Dec Jan Feb Mar Apr May Jun Jul Aug
1981-82
b.
Streamflow (m3/s)
Rainfall (mm)
Feb Mar Apr May Jun Jul Aug
1995



y = 1.05x - 0.23
R^2 = 0.72
0.9
0.6
0.3
0
Qsuperficial
0
0.2
0.4
0.6
0.8
1
Q total


Qtotal
Qsuperficial
Qdeep
Runoff (m3/s)
2.5
2.0
1.5
1.0
0.5
0.0
4/1/2003
4/15/2003
4/29/2003
5/13/2003
5/27/2003
6/10/2003
6/24/2003
7/8/2003
7/22/2003
8/5/2003
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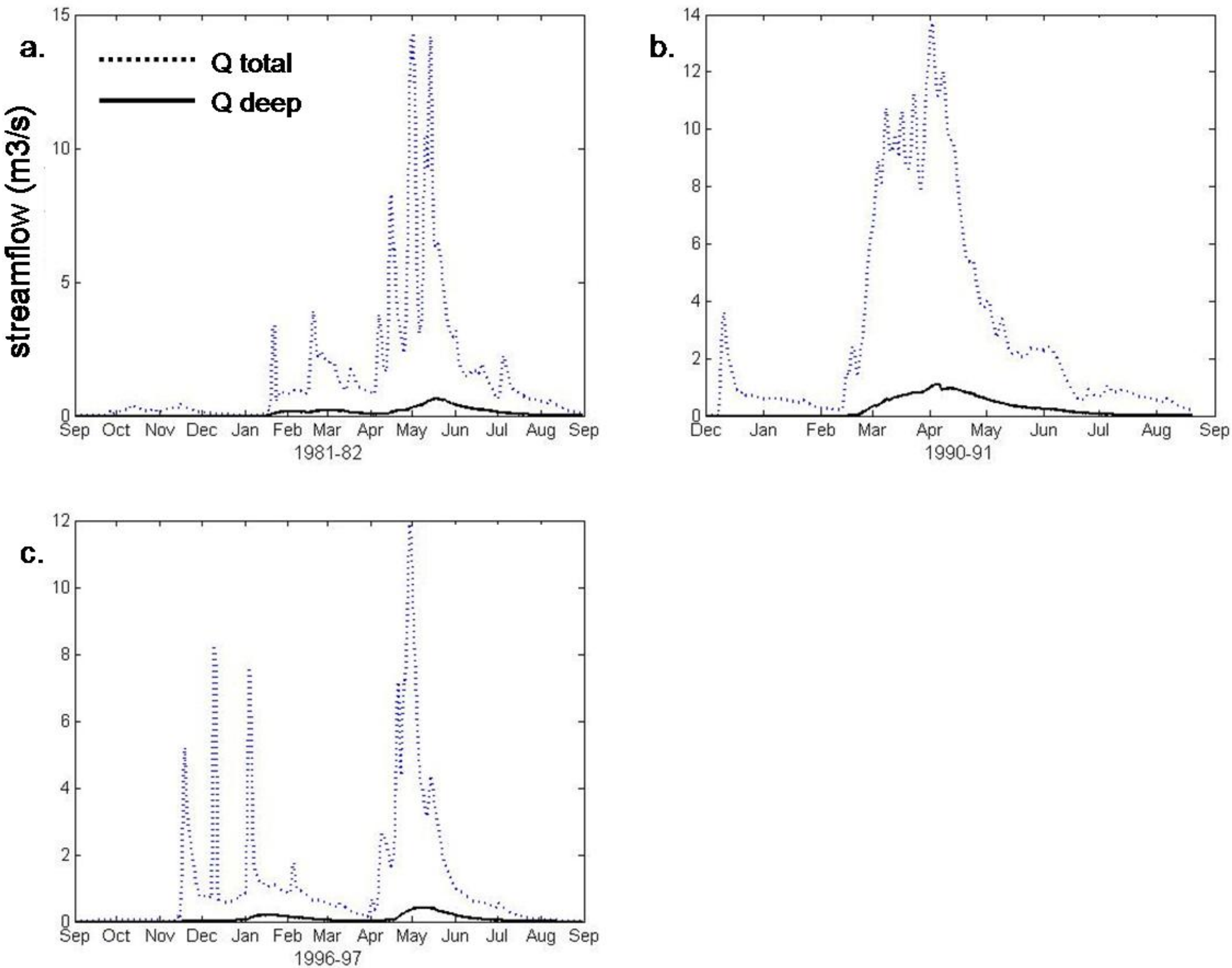
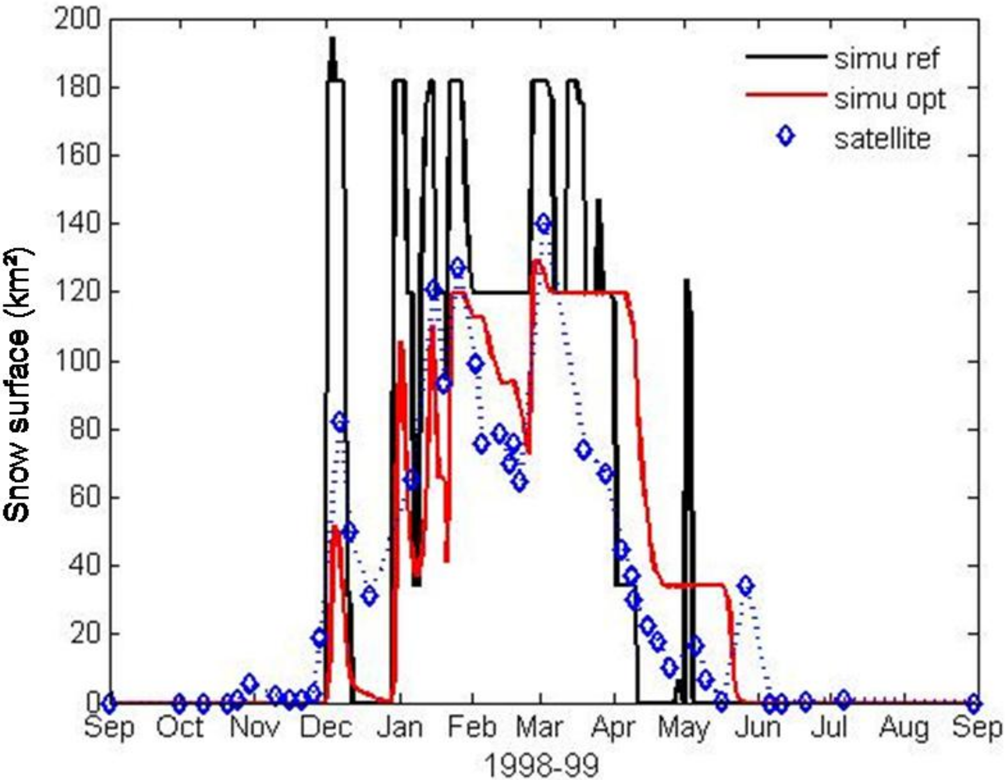
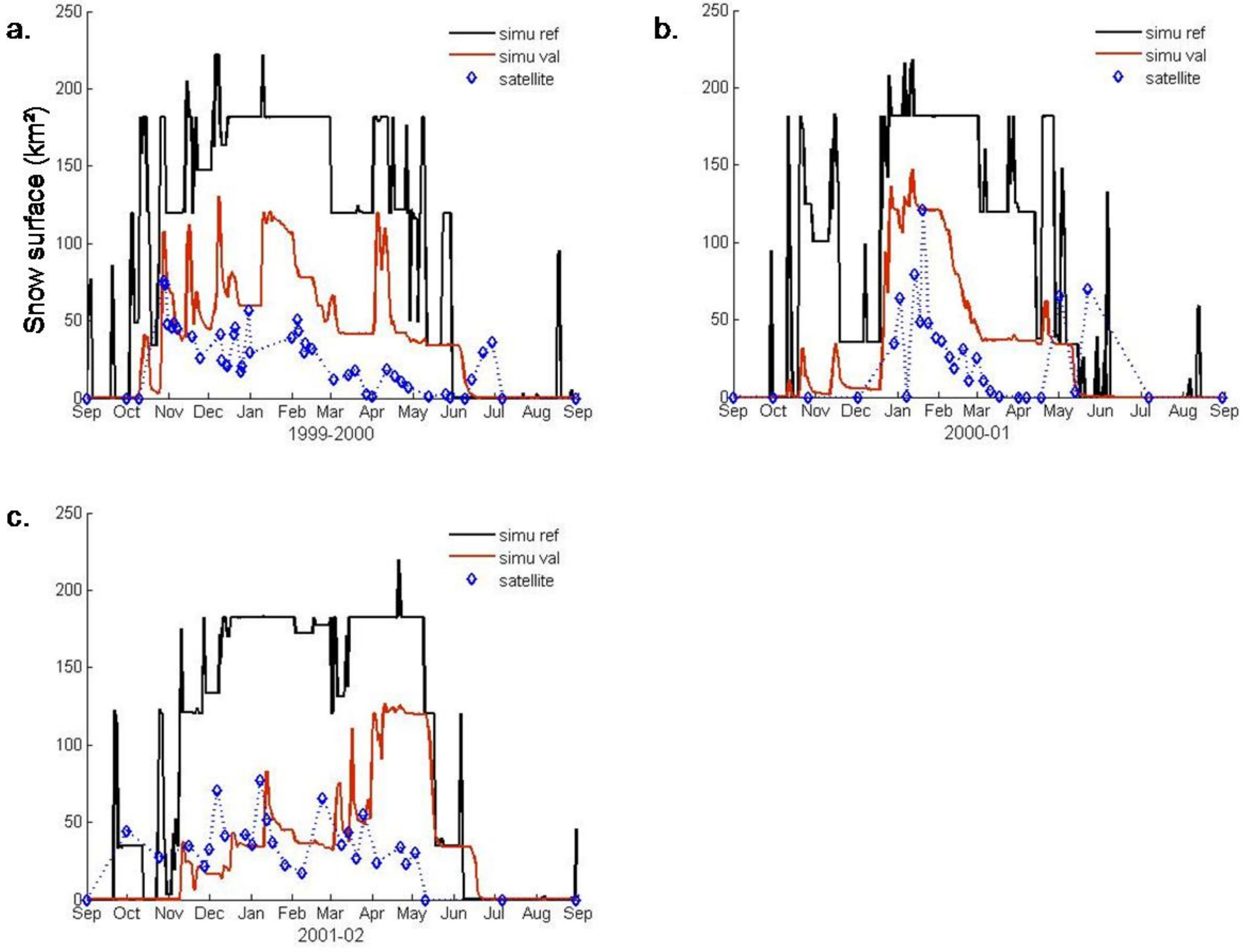
9/30/2003
10/14/2003
10/28/2003
11/11/2003
11/25/2003



a.
Q total
Q deep
streamflow (m3/s)
0
5
10
15
Sep
Oct
Nov
Dec
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
1981-82
b.
0
2
4
6
8
10
12
14
Dec
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
1990-91
c.
0
2
4
6
8
10
12
Sep
Oct
Nov
Dec
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
1996-97


simu ref
simu opt
satellite
Snow surface (km²)
0
20
40
60
80
100
120
140
160
180
200
Sep
Oct
Nov
Dec
Jan
Feb
Mar
Apr
May
Jun
Jul
Aug
Sep
1998-99


a.
b.
c.
Snow surface (km²)
simu ref
simu val
satellite
0
50
100
150
200
250
Sep Oct Nov Dec Jan Feb Mar Apr May Jun Jul Aug Sep
1999-2000
2000-01
2001-02