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Effect of the spatial distribution of physical aquifer properties on water table depth and stream discharge in a headwater catchment

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Abstract

Water table depth and its dynamics is often poorly predicted upslope despite they control both water transit time within the catchment and solute fluxes at the catchment outlet. The paper analyses how relaxing the assumption of lateral homogeneity of physical properties can improve simulations of water table depth and dynamics. Four different spatial models relating of saturated hydraulic conductivity to topography have been tested: a simple linear relationship, a linear relationship with two topographic indexes, two domains with a transitional area. The Hill-Vi model has been modified to test these hypotheses. The studied catchment (Kervidy-Naizin, western France) is underlain by schist crystalline bedrock. A shallow and perennial groundwater highly reactive to rainfall events mainly develops in the weathered saprolite layer. The results indicate that 1) discharge and the water table in the riparian zone are similarly predicted with the four models, 2) distinguishing two domains constitutes the best model and slightly improves prediction of the water table upslope, and 3) including spatial variations in the other parameters such as porosity or rate of hydraulic conductivity decrease with depth does not improve the results. These results underline the necessity of better investigation of upslope areas in hillslope hydrology.

1 Introduction

In catchments underlain by crystalline bedrock with a deep, thick, weathered aquifer, runoff during storm events as well as during baseflow periods is often controlled by shallow groundwater. Water table depth and its dynamics control both water transit time within the catchment and solute fluxes at the catchment outlet. Water transit time depends on the slope position, from a few days in the riparian zone to a few years in uplands. Transit time in the uplands can be so long because of a much thicker unsaturated zone and also much lower hydraulic gradients on plateaus compared to midslope regions and riparian zones (Freer et al., 1997; Molenat and Gascuel-Odoux, 2002).

While mean transit time can be estimated by water dating techniques or lumped models using input and output signals, modelling the internal flow within the catchment is necessary because of considerable uncertainty with these methods. Coupling water and solute transport requires a good knowledge of flow velocity and geochemical processes, which may vary laterally within the shallow groundwater. In such cases, solute fluxes such as nitrate are also controlled by hydraulic gradients within the groundwater that determine the contribution of different spatial domains and their connectivity (Ocampo et al., 2006a, b). As an example, in upslope areas shallow groundwater can store nitrate, and in downslope positions can act as a nitrate buffer, with low N concentrations due to denitrification (Molenat et al., 2002; Steinheimer et al., 1998). Therefore, reasonable predictions in space and time of the water table depth and its dynamics are urgently needed for shallow groundwater catchments.

Experimental studies have shown that water table depth is more or less correlated with topography, in some cases strongly represented by topographic indexes (Moore and Thompson, 1996; Thompson and Moore, 1996) and in other cases less predicted by topographic indexes (Myrabo, 1997; Seibert et al., 1997). Some studies underline that this topographic dependence is not true for all storm events (Jordan et al., 1997) or for the whole hillslope (Molenat et al., 2005; Martin et al., 2006). The relationship is particularly weak in upslope areas where the water table depth is independent of topography (Molenat et al., 2005). Otherwise, few studies have investigated the dynamics of water table depths in upslope areas because of the difficulties and costs associated with measuring relative deep water tables in areas not important for groundwater resources. The clear lack of correlation between topography and water table depth in some cases indicates that soil and aquifer properties cannot be assumed to be homogeneous for the entire hillslope and the water table dynamics may also depend on variations of the soil physical properties.

Modelling studies have investigated the effects of vertical variations in soil properties on water table dynamics and discharge. Different conceptualisations were proposed for modelling vertical variation of soil properties. Different functions of hydraulic conductivity that decrease with depth have been tested by Ambroise et al. (1996). More recently, assumptions about the drainable porosity (the pore space between field capacity and saturation) decreasing with depth or about the preferential flow have also been proposed and tested (Weiler and McDonnell, 2004, 2006). The occurrence of shallow groundwater flow in a catchment is governed by a decrease of soil hydraulic conductivity and porosity with depth. Under these conditions, percolation is stopped or limited and a saturated zone develops that generates lateral groundwater flow in response to a downslope hydraulic gradient. In these models and most other hillslope models, soil and aquifer properties were assumed to be laterally homogeneous, whereas many field observations show lateral variations in soil depth, porosity and conductivity. These variations are local (Tromp van Meerveld et al., 2007) or linked to topography with respect to soil (Curmi et al., 1998) or weathered materials (Dewandel et al., 2003, 2006). Few modelling studies have investigated the effect of these lateral variations. Saulnier et al. (1997) showed that variation in soil depth with topography does not really affect discharge. In contrast, local variation of soil depth can have a major effect on flow connectivity and thus discharge, as demonstrated by field observations and modelling approaches (Tromp van Meerveld and Weiler, 2008). Lateral variation in transmissivity was modelled using a Topmodel approach and calibrated with field observations (Lamb et al., 1997; Seibert et al., 1997). They produced slightly better agreement between simulated and observed water table depths. These studies focused on the bottom domain. In saturated conditions, different pedogenic processes (e.g., lixiviation and redox) can induce topographic variation in the physical properties as observed in these studies and others (Curmi et al., 1998). Other kinds of processes have not been investigated. Particularly, detailed measurements of hydrodynamic properties as well comprehensive and functioning models of the hydrodynamic structure of weathered layers in crystalline rocks have just been started (Chilton and Foster, 1995; Taylor and Howard, 2000; Marechal et al., 2004; Dewandel et al., 2006). They show that hydrodynamic properties are not homogeneous in space but structured with the topography due to weathering processes.

There is still much debate about which modelling approach is appropriate to predict water table dynamics simultaneously with runoff in catchments dominated by saturated subsurface flow. One of the main questions is how far we can simplify the geometry of the system, the soil and sub-soil, as well as the hydraulics of groundwater dynamics, while still achieving an acceptable simulation of water table variations as well as transit time and solute fluxes. Fully distributed field studies and data are lacking for conceptualising these spatial variations and for testing them in subsurface flow models. Observations of water table dynamics should not be limited to the riparian zones or toeslopes as has primarily been the case so far. In upslope areas, we still lack a satisfactory understanding of the relationships between spatial and temporal variations in water table geometry and the geometric properties catchments (topography, soil depth and spatial distribution of physical properties).

The aim of this work is to analyse how relaxing the assumption of homogeneity can improve simulations of water table dynamics. We focus on spatial variation in upslope position, and we test different spatial dependence of hydraulic conductivity with topography. The study site is a catchment underlain by schist crystalline bedrock, where the shallow and perennial groundwater develops mainly in the weathered saprolite layer. On this site, different modelling approaches based on the homogeneous hypothesis were unable to efficiently simulate the water table depth in an upslope position (Molenat et al., 2005), thus it was considered challenging to search for an improved prediction of water table depth by considering different spatial models of physical properties, in agreement with the structure of soil and bedrock properties.

2 Methods

2.1 Study site

The Kervidy-Naizin catchment is located in Brittany, France, and covers an area of 4.9 km^2. The slopes are gentle, less than 5%, with the northern part being particularly flat (Fig. 1). The land use is intensive farming, with 32% of the surface area covered by meadows, 30% by maize, 23% by winter cereals and the remainder by leguminous plants, fallows and colza. Intensive soil, hydrologic, agronomic and geologic studies have been undertaken during the past ten years (Bruneau et al., 1995; Crave and Gascuel-Odoux, 1997; Curmi et al., 1998; Franks et al., 1998; Molenat et al., 2002, 2005, 2008; Pauwels et al., 2000). It belongs to the ERO AgrHyS (French Environmental Observatory on transfer time in agricultural catchments), an experimental catchment network for studying the response time of hydro-chemical fluxes to the evolution of agricultural practices (http://www.inra.fr/ore_agrhys).

The mean annual precipitation over the last 30 years is 909 mm, whereas the mean annual evapotranspiration and runoff from 1994 to 2004 are 709 and 400 mm, respectively. Three main formations are identified from the soil surface to the bedrock composed of Brioverian schist: the soil, the unconsolidated weathered bedrock, and the fissured and fractured weathered bedrock. The soils are silty loams, depths ranging from 0.5 to 1.5 m (Curmi et al., 1998). The soil system comprises an upland well-drained domain, with an average saturated hydraulic conductivity of 10^{-5} m/s, and a poorly drained bottomland with an average hydraulic conductivity of 10^{-6} m/s. The thickness of the unconsolidated material varies greatly in space from a few metres to 30 m. A model of the extension of the weathering processes was proposed for this region (Wyns et al., 1999; Dewandel et al., 2006). In this model, the unconsolidated weathering areas are more extended upslope than downslope. The hydraulic conductivity of the unconsolidated material is similar to that of the soil (Molenat and Gascuel-Odoux, 2002). A shallow, permanent and unconfined aquifer develops over the whole catchment area in the soil of bottomlands along the stream channel and in the unconsolidated weathered bedrock in the entire catchment.

The catchment is equipped with stream gauge stations at the outlet, a meteorological station and transects of wells that intercept the permanent and shallow groundwater that develops in the soil and unconsolidated weathered bedrock. Stream discharges were recorded every 1 to 6 min at the gauging station. The meteorological station recorded hourly rainfall, air temperature, soil temperature at 50 cm depth and other variables (wind speed, global radiation and relative humidity) required to calculate potential evapotranspiration using the Penman method. Wells ranged from 1.5 to 20 m in depth and are described in detail in Molenat et al. (2005, 2008). Water table depths were monitored with bubble sensors or shaft encoders with an integral data logger. Errors in water-table measurements with sensors ranged from 1 to 5 mm. The present work focuses on one transect named G (Fig. 1).

The observed stream discharge followed a classical pattern for humid and temperate climates underlain by shallow impervious bedrock (Fig. 2): the discharge is at its maximum in winter (December–January) and decreases until early autumn (September–October) until it dries out in most years. The water table depths follow different temporal patterns (Molenat et al., 2008) (Fig. 2). During a first period, with a very shallow water table in the bottomland and extending from late autumn to spring, the water table in the riparian zone ranges from the soil surface to around 0.5 m below the soil surface and water table rise occurs rapidly with each rainfall event. During winter periods without any rainfall, the water table remains in the upper 50 cm and no recession can be observed. In the hillslope shoulder and plateau area, the groundwater table varies widely. The water table varies from 1 m to 3 m below the soil surface in the shoulder (PG4 and PG5) and from 4 to 6 m in the plateau area (PG6). Water table response to rainfall is much slower and smoother compared to the riparian zone. The water table falls as soon as rainfall ceases. During a second period which corresponds roughly to the summer period (July to September), the water table is far below the surface along the entire hillslope. Depletion in the riparian zone drives the water table down to 1 m below the soil, with very weak temporal dynamics in contrast to the winter period. In the shoulder and plateau area, recession of the water table is rapid and the final water table is very deep. An abrupt shift is observed from the first to the second period. Consequently from this temporal variation, the hydraulic gradient can vary by a ratio of 1:5 along the year in upslope areas, whereas it is almost equal to the topographic slope in the midslope area and the riparian zone. Thus, the hydraulic heads in the upslope areas were not proportional to the topographic slopes and no clear relationship could be observed. A previous modelling study comparing water table simulations with three models – Topmodel, a kinematic model and a diffusive wave model – showed the difficulty in correctly predicting water table depth in the upslope area with all three models. Despite a slight improvement with the diffusive model, differences between observations and simulations remain large (Molenat et al., 2005).

2.2 Model

We used the physically-based hillslope model Hill-vi (Weiler and McDonnell, 2004, 2006; Tromp-van Meerveld and Weiler, 2008; Anderson et al., 2009) for runoff generation by simulating water fluxes in the saturated and unsaturated zone of the hillslope. The model is based on the concept that two compartments define the saturated and unsaturated zone for each hillslope grid cell, based on topography and soil depth. The unsaturated zone is defined by the depth from the soil surface to the water table and its time-variable water content. The saturated zone is defined by the depth of the water table above an impermeable interface and total porosity n. Lateral flow in the saturated zone is calculated using the Dupuit-Forchheimer assumption. Routing is based on the grid cell-by-grid cell approach (Wigmosta and Lettenmaier, 1999). Hydraulic conductivity is defined by an exponential function to reproduce the changes of hydraulic conductivity with depth due to soil development. Since the active hydrologic zone at the studied hillslope is relatively deep (up to 10 m), we included constant hydraulic conductivity with depth in our model. The transmissivity T is then given by

$$T(z) = \int_z^D K_s(z)dz = \int_z^D \left(K_o \exp\left(-z(t)/m\right) + K_c\right)dz \quad (1)$$

where K_s is saturated hydraulic conductivity, K_o is the saturated hydraulic conductivity at the soil surface, K_c is the constant saturated hydraulic conductivity with depth, m is the rate of hydraulic conductivity decrease with depth and z is the depth into the soil profile (positive downward). In addition, the model includes a depth function for drainable porosity, taking into consideration that drainable porosity declines with soil depth (Weiler and McDonnell, 2004). Actual evaporation from the unsaturated zone is calculated based on the relative water content in the unsaturated zone and potential evaporation. Drainage from the unsaturated zone to the saturated zone is controlled by a power law relationship between relative saturation in the unsaturated zone and saturated hydraulic conductivity at water table depth (Weiler and McDonnell, 2004). Further details and extension to Hill-vi can be found in the work of Weiler and McDonnell (2004, 2006) and Tromp van Meerveld and Weiler (2008). In total, 8 parameters were used to simulate the dynamics of the saturated and unsaturated zones.

A factor of variation $\alpha(x)$ has been introduced into the Hill-vi model to simulate spatial variation in porosity, hydraulic conductivity or a m parameter. This factor is a multiplicative factor which can be added to these two variables:

$$\mathrm{Var}(x) = \alpha(x)^{*}\mathrm{Var}_{o} \tag{2}$$

where x is the distance to the stream and Var_o can be the hydraulic conductivity, the drainable porosity or a parameter m.

We have considered topographic variation in this factor α to be a first step in implementing heterogeneity. Topographic dependence can be due to saturated conditions in the toeslope. It can be positive, as previously described, with hydraulic conductivity or porosity decreasing toeslope, such as in gleyic soils, or negative, instead increasing toeslope, such as in peat soils. This topographic dependence can also be due to spatial variation in weathering processes.

Four spatial models (i.e., four ways the factor α could vary with topography) have been tested:

- the *linear model*, varying linearly with distance to the stream; the simplest model of the four;

- the *threshold model*, which includes two areas with constant alpha joined by a short transitional domain, describing spatial variation linked to the plateau or bottom domain;

- the *mono index model*, with variation according to a topographic index, computed with a unidirectional scheme, i.e., where the flow coming from one cell goes to a single neighbouring cell, mostly downslope; we assume that spatial variation of the physical properties follows water drainage flow pathways;
- the *multi index model,* varying with a topographic index, computed similarly to 3) but with a multi-directional schema, such that the flow coming from one cell goes to different neighbouring cells according to the slope gradient, assuming similar spatial variation due to water drainage.

For the two last models, α was calculated on the 3 dimensional Digital Elevation Model (DEM). It was been normalised to compare to the four models.

3 Numeric experiments

One transect with six wells (transect G) and hourly data from the year 1999–2000 (Fig. 2) was selected for this study because it was used previously to compare different models of the water table assuming homogeneity of soil and bedrock properties for the entire hillslope (Molenat et al., 2005). The period is 3650 h long (from 12 p.m. on 31 October 1999 to 12 p.m. on 31 March 2000), is representative of an average year and covers seasonal variations, between fall and spring, as well as the quick responses of the water table to rainfall events in winter. The time step of simulation was one hour. The transect is represented as a 10 m wide and 480 m long area with an homogeneous depth of 10 m. Thus, the transect and the period represent a well-investigated starting point to test the improvement of predictions of discharge and water table dynamics when introducing spatial variation in soil and bedrock properties.

A 2 m grid resolution was used for the Hill-vi modelling. We chose mean winter conditions (a daily rainfall of 3 mm per day) and applied them as constant rainfall over

2000 h to the model to develop the initial conditions for each model run based on the chosen parameter. For the four spatial distributions of the physical properties, the factor of variation α was derived from a detailed topographic survey on a 2 m grid size. For the threshold model, the threshold was set to the slope break between the linear slope and the plateau (Fig. 3). For the two topographic index models, the DEM was used as usual, but the general slope from the grid cell to the stream according to the flow pathway was used in place of a local slope, as proposed by Gascuel-Odoux et al. (1998) and Merot et al. (2006). This modification takes into account drainage of the shallow groundwater according to the slope position. The normalised values of the factor of variation are different, particularly in the upslope domain (Fig. 3).

The effect of the spatial distribution of physical properties was analysed with two methods:

- Firstly, a Monte Carlo procedure (1000 parameter sets) was used to get the best set of parameters for each spatial model considering discharge on one hand and the water table on the other hand. We analysed the improvement of predictions when considering spatial variation of the physical properties based on the best set of parameters. A uniform probability distribution was assumed between the lower and upper limits of the parameters. These values were chosen within a relatively small range according to the observed values and a previous study (Table 1, Molenat et al., 2005). As a first step, only spatial variation in K_o was studied, with its variation ranging over one order of magnitude and with K_o increasing from downslope to upslope, according to the weathering processes in the upslope area observed by Wynns et al. (1999) and Dewandel et al. (2006). In addition, the combined lateral variation of K_o with both m and porosity was investigated.

- Secondly, a sensitivity analysis was performed to explore the effects of the direction and magnitude of spatial variation in the physical properties, taking into account using the set of parameters which provides the best fit to the observed water table depth and considering no spatial variation as the reference (Table 1).

Different ranges of variation in K_o were tested, with a magnitude from 1 to 4, and the location of the threshold in the threshold model was moved up and down between 10 and 40 m.

The performance of each model run was evaluated with the following objective functions:

- For discharge, the Nash-Sutcliffe efficiency and the efficiency of the logarithmic discharge values were calculated by:

$$\text{Eff-Q} = 1 - \sigma_{\text{err}}^2/\sigma_{\text{obs}}^2 \quad (3)$$

$$\text{Efflog-Q} = (1 - \lambda_{\text{err}}^2/\lambda_{\text{obs}}^2) \quad (4)$$

where σ_{err}^2 and σ_{obs}^2 are the variance of the simulation errors and observations, respectively, and λ_{err}^2 and λ_{obs}^2 are the variance of errors calculated from logarithmic discharge and the variance of logarithmic observed discharge, respectively.

- For the dynamics of the water table, the difference in the average level (D) and the range of variations (R) between the observed and simulated water table variation for three wells (PG2, PG5 and PG6) was calculated with:

$$D = 1/n\left(\sum_{1}^{n}\left(\text{WTD}_{\text{sim}}(t) - \text{WTD}_{\text{obs}}(t)\right)\right) \quad (5)$$

$$R = \left[\left(\max(\text{WTD}_{\text{sim}}) - \min(\text{WTD}_{\text{sim}})\right)/\left(\max(\text{WTD}_{\text{obs}}) - \min(\text{WTD}_{\text{obs}})\right)\right] - 1 \quad (6)$$

where WTD is the water table depth, and sim and obs correspond to the simulated and the observed values. D and R were computed for all non-missing values during the study period. We have preferred to calculate D and R on the relative values rather than on the absolute values, despite a possible compensation of the errors: these two criteria are simple and keep a physical significance. In our case, the compensation of errors can probably not be involved: in previous work (e.g., Molenat et al., 2005) the errors were mainly due to a shift or a smoothing of the mean level of the water table depth; but this point will be verified in our simulations. D and R allow us to assess whether average water table depth and its range of variation agree with the observed values, to evaluate the error in the studied period without focusing on detailed response dynamics. These two objective functions (R, D) have a target value equal to (0, 0). The Euclidian distances of R and D of the 3 studied wells to the null target value 0 were calculated and named Dist-D and Dist-R, respectively, to resume and facilitate reading the results.

All behavioural simulations (i.e., simulations leading to a Nash-Sutcliffe efficiency greater than 0.4 and to a logarithmic discharge efficiency greater than 0.7) were retained for our analysis.

4 Results

4.1 Hill-Vi model application

The application of the Hill-vi model to the data set without considering any spatial variation of the physical properties results in a best fit with a similar efficiency for discharge as in previous studies. Molenat et al. (2005) calculated efficiencies of 0.87 (discharge calibration) and 0.82 (groundwater calibration) and of 0.76 (discharge validation) and 0.68 (groundwater validation) with a diffusive model compared to Eff-Q=0.59 and Efflog-Q=0.77 with Hill-vi (Table 2). The Hill-vi model seems to produce a similar behaviour, and the effects of spatial variation of the physical properties can be tested.

The best fit of the model to water table depth results in a slight decrease of the discharge efficiency, but the objective functions for the water table depth are improved

(Table 2). The range of variation of the predicted water table depth is higher in the upslope area compared to the simulation with the best fit for discharge (Fig. 4). However, the simulated seasonal variations at PG5 and PG6 are much smaller than the observed variations for both objective functions (Fig. 3). The *R* and *D* criteria are better, particularly in the upslope domain, but are still not satisfactory.

The Hill-vi model, as well as previous modelling approaches, fails to predict the observed dynamics of the water table, particularly in the upslope area. This result also justifies the presented approach of testing the effects of lateral variation of physical properties.

4.2 Effect of lateral variation of saturated conductivity on discharge and water table depth

The two criteria *R* and *D*, related to the water table depth, have been computed for the different spatial models. In Fig. 5, the behavioural model simulations with Eff>0.4 and Efflog-Q>0.7 are plotted for the three wells and compared with the different spatial models. The optimum value of zero for the two objective functions *R* and *D* cannot be reached when no spatial variations are taken into account, while the two criteria are closer to their optimum when considering any of the 4 other spatial models. However, a similar good agreement for all 3 wells simultaneously can never be achieved. The results are rather similar for the two topographic index models and the linear model. Thus, the slight difference between these three models cannot produce any significant differences in estimates of water table dynamics. Water table depth is best estimated with the threshold model, essentially because of a better prediction of PG5 and PG6. The number of behavioural simulations is lower for this spatial model compared to the three other models. The threshold model is the most selective model for estimating water table depth with a small number of parameter sets.

When analysing the result for the best set of parameters (Table 2), similar conclusions can be drawn: the best simulations of water table depth are obtained with the threshold model. The depth of the water table is always estimated well for PG2 (the mean error is lower than 0.1 m for all models), but only estimated with an error lower than 1 m for PG5 and PG6 using the threshold model (D=0.96 for PG6). Considering the best fit for water table depth, the results are slightly improved (e.g., D=0.76 for PG6). The same behaviours are observed for the R criteria. If the dynamics of water table depth are included in identifying the best model, the simulations become better with respect to these criteria, but the efficiency with respect to discharge does not change much.

Figure 6 synthesises these results and shows that any model is able to simulate discharge relatively well with similar efficiency. The spatial models have no significant effect on predicting discharge. The effect of the spatial models on water table depth is small for the downslope area (PG2). Conversely, the spatial models have a larger effect for the upslope area, particularly for estimating mean water table depth, when fitting to discharge, and for estimating the range of variation of water table depth when fitting to the water table. However, if the criteria are averaged, the effects are much smaller (Table 2).

This analysis indicates that variation in K_o, according to the threshold model, results in the best prediction of water table depth in the upslope and downslope areas. When the model was fitted only against discharge, as it is generally done when no data on water table depth are available and no model of spatial structure is known, the water table depth cannot be reasonably predicted in the upslope area for the observed hillslope. Conversely, knowledge of the spatial structure can improve the estimate of water table depth.

Table 3 shows the effect of the different spatial models on the parameter values for the model with the best fit. The variations among the different models cover all possible ranges when the simulations are fitted to discharge, except for K_o and drainable porosity which are rather constant. The range of the variation is slightly wider when the simulations are fitted to water table depth.

The dynamics of the predicted water table depth for all wells are illustrated in Fig. 7. In the riparian zone, the predicted water table depth is smoother than the observations

(PG1, 2 and 3) independent of the spatial model. Some of the observed positive water table depths also cannot be reproduced by the model, since Hill-vi in this version does not consider exfiltration and flooding. The dynamics of water table depth are better predicted in the midslope and the upslope area if a spatial model is introduced. In particular for PG6, though, the large dynamics of water table variation are not captured by any of the models.

4.3 Sensitivity analysis of the magnitude and direction of the spatial variations of the physical properties

Since all results for testing different spatial models with variation in saturated hydraulic conductivity were not satisfactory, a sensitivity analysis was performed to study the influence of possible additional effects. We tested the effects of the magnitude and direction of the variations in the spatial model, considering the different models as shown in Fig. 8 and using the simulation with no spatial variation as a reference. The value of saturated conductivity was fixed 4, 3 and 2 times higher going upslope, such that it increased from bottom to top, and 2 times higher going downslope, such that it increased from top to bottom, compared to a reference value. Figure 8 shows that the magnitude and direction of the spatial model do not affect the prediction of discharge. For the threshold model, better predictions are obtained when K_o increases from downslope to upslope than the reverse. For this model, larger magnitudes of spatial variation result in better prediction of water table dynamics in the upslope area (PG6). For the other three models, the improved prediction in one well is compensated for by worse predictions of the midslope area (PG5).

In addition, we also tested whether the location of the change between the two domains of the threshold model has any effect on performance, using the best-fit model for the threshold model as a reference. Figure 9 shows that the best estimate of water table depth is obtained with a threshold located just before the break of the slope, 60 m downslope of the initial location. However, the performance increase is small and only visible for PG6.

4.4 Effect of lateral variation of K_o, m and porosity on discharge and water table depth

The improvement considering spatial variation of K_o is real, but the estimates are still far from the observations. Two hypotheses can be generated: 1) spatial variation of K_o does not have a large effect because it concerns layers that are too superficial, compared to the depth of variation in groundwater in the plateau domain, due to a too small fitted value of m; 2) the reactivity is induced by both spatial variations of K_o and drainable porosity. These two hypotheses have been tested (Fig. 10, Tables 2 and 3).

No improvement is observed. The discharge is always predicted as well as previously, when fitting to the discharge or to the water table as well. Water table depth is predicted better than without the spatial model, but not clearly better than when considering spatial variation of K_o. Therefore, the actual results due to only variation of K_s can be considered to be the best solutions in cases where no measurements are available.

5 Discussion

5.1 Effect of spatial variability of physical properties on discharge and water table predictions

Previous studies of the Kervidy-Naizin catchment have shown the difficulties in predicting water table depth and its dynamics in the upslope area (Molenat et al., 2005). Soil physical properties have previously only been modelled as they change with depth but not laterally along the hillslope. This hypothesis of lateral homogeneity has been relaxed in this study in an attempt to better predict water table dynamics by including different spatial models describing variations of the physical properties. All tested spatial models have been related to topography to implement relatively simple spatial variations of the physical properties. The addition of such spatial models has improved predictions of water table dynamics in the studied hillslope. The threshold model, which defines two lateral domains of K_o (one for the plateau and one for mid-slope and toeslope), has been shown to be particularly successful in predicting water table dynamics. This spatial model agrees with Dewandel model (Dewandel et al., 2003, 2006) proposing a structural model for highly weathered hillslope. The weathering is more developed upslope and on the plateau, and therefore the weathered layer is thicker and its hydraulic conductivity is higher. It appears to be a convenient solution to improve predictions of water table depth and its dynamics upslope by accounting for different spatial domains of K_o in relation to the weathering processes along the hillslope. Different spatial models had been previously compared, taking into account vertical variations of physical properties by including different vertical layers on this hillslope (Molenat and Gascuel-Odoux, 2002; Martin et al., 2006). These different hypotheses have been found to be not relevant to improve prediction of the water table. Thus, investigating and testing lateral variations of the physical properties appears to be a necessary step to better describe the complexity of the structure of this hillslope and to propose relevant simple structural models of its hydraulic conductivity.

However, the overall modelling results still remain unsatisfactory, and different reasons can be found. One explanation could be the effect of small-scale local variations of the physical properties. Local characteristics could explain the high reactivity of the water table following intensive rainfall events and also influence lateral transfer in the shallow groundwater. A second reason could be the too low range of spatial variation in K_o. It would be interesting to enlarge the range of variation of the physical properties; since the combination of possibilities is huge, this would need to be done in a very structured way. The last potential reason is that the connectivity of the studied hillslope to the stream is not representative of the other hillslopes of the catchment and is therefore not related well to the discharge dynamics of the whole catchment. However, if that were the case, we would have obtained bad simulations of discharge while achieving good simulations of water table depth when fitting the simulations to them.

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Whatever the reasons are, we can obtain two major conclusions from this study: 1) There is a real lack of physical measurements of the weathered layers (e.g., depth, transmissivity and porosity), particularly on the plateau domain. These measurements are important because of their potential effect on the dynamics of the water table, as shown in this study. 2) The potential of the Hill-vi model as a tool to test different simple hypotheses in different physical environments. The Hill-vi model has been applied here in different conditions than previously (a 10 m deep transect, a high seasonal reactivity of water table depth, etc.). Finally, the prediction of discharge was acceptable, and the model has included relevant spatial structure for the hillslope, which could be confirmed. Hill-vi is an interesting tool to investigate the effect of the spatial structure of the hillslope and its effects on hydrological processes.

5.2 Interest for hillslope hydrology

As very often in hydrology, our experiments and their results are related to a specific place. However, the studied hillslope represents common characteristics of many deeply weathered watersheds, and more general assessments can also be drawn from this study.

As was observed with other spatial models representing variation in physical properties, the prediction of discharge was not very affected, while the prediction of the water table depth or other internal processes can be rather different (see also Weiler and McDonnell, 2006). A complex spatial model to describe the physical properties is often not necessary for improving predictions of discharge. But, it may be necessary for prediction of internal processes, water pathways and water transit time. This study showed that lateral variation seems to be as important as variation in soil depth for correctly representing the fluxes and dynamics in a hillslope. Therefore, in the studied environment, as well as in similar environments where the water table dynamics are very different in the upslope and downslope areas, a simple, process-driven spatial model representing changes in saturated hydraulic conductivity could be a good solution to better predict discharge and water table dynamics when detailed data of water table depth are missing. However, this approach implies the ability to choose a correct spatial model and parameterise it adequately to test it. An inverse model approach (Carrera et al., 2005) cannot be the solution as long as we do not know the correct spatial model. The possible combinations of parameters are too numerous, and the unknown values of the parameters to describe the spatial model amplify the problem. In the past, inverse models have been applied to estimate groundwater flow by fitting them to many observed heads; however, spatial variability has been fixed to the different facies, and only their absolute parameterisation has been changed (Fienen et al., 2009). This spatial model can only be reliable if the correct spatial representation from soil or geological surveys is available. But, particular parameters like hydraulic conductivity are already highly variable within one soil or geological unit, so we can hardly use this information to represent absolute differences among the areas. Only a combination of water table measurements (or other spatial observations, such as soil moisture, if other processes are studied) and an appropriate spatial model, together with discharge observations, can provide the necessary information to parameterise a distributed model (Yeh et al., 2008). As can be seen for the studied hillslope, the predictions based on the different spatial models are rather different from each other and different from predictions without a spatial model. A spatial model needs to be based on field data.

In summary, we would like to stress the real need for more water table observations in the entire watershed. Generally, wells are drilled in the riparian zone or the lower hillslope zone, because we hope to measure something and we would need deeper wells in the upslope area to ensure success in continuously observing water table dynamics. But, observations in the downslope area are often strongly correlated to discharge and are therefore not sufficient to provide independent observations for benchmarking a model (Lyon et al., 2006). This study also shows that predictions of PG2 are close to the observations, because they are closely related to the dynamics of discharge. Upslope information while it is not as correlated to discharge, provides us additional information to calibrate a model. This independent information is lacking in most experimental catchments (see Lyon et al., 2006 for an example). Up until now, many hillslope studies have focused on an extent of 20–50 m. Extending the area of these studied slopes may help us to better characterise the behaviour of the entire hillslope. It is also important to ensure that our distributed models can predict the behaviour of the entire hillslope since the resulting transit times could be very different if the unsaturated zone in the upslope area is 1 m or 10 m deep.

6 Conclusions

Water table depth and its dynamics are often poorly predicted in upslope areas. We have analysed how relaxing the assumption of lateral homogeneity of physical properties can improve simulations of water table dynamics. Four different spatial models relating hydraulic conductivity to topography were tested for a well-studied catchment (Kervidy-Naizin, western France). We used the Hill-vi model to represent the shallow and perennial groundwater that develops in the weathered saprolite layer. The results indicate that discharge and water table depth in the riparian zone are similarly well predicted by the four models, as well as with a model not considering the spatial variability. However, a spatial model including higher conductivity in the upslope area improves the prediction of the water table in this area. There could be more hypotheses tested with this approach, but we should constrain them with field observations.

This study underlines the real need to better investigate the upslope areas in watersheds and the hydraulic properties of the weathered layers, particularly when questions of residence time and coupling water with solute transport are involved. Upslope information provides additional, independent information to calibrate a distributed model that is not highly correlated to discharge. This additional information has to be better analysed, particularly with a cross-analysis of discharge and water table dynamics in different hillslope positions. However, the purpose should always be to choose the simplest spatial structure of the hillslope, taking into account soil, weathered layers and bedrock.

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Table 1. Values of the parameters in the Monte Carlo analysis and in the sensitivity analysis.

Values of the parameters	Monte Carlo Analysis		Sensitivity Analysis
	Min	Max	Reference
Total porosity (%)	22	32	23
Drainable porosity (%)	3	5.5	4.4
b, parameter of exponential depth function of drainable porosity (m^{-1})	2	5	4.2
K_o, saturated hydraulic conductivity at soil surface (m/h)	0.1	1.5	1.43
m, parameter of exponential depth function of conductivity (m^{-1})	0.4	1	0.86
K_c, constant conductivity in depth (m/h)	0.002	0.03	0.013
Drainage coefficient of the unsaturated zone	20	35	35

Table 2. Mean error (M) and normalized relative range of variations (R) of water table depth during the testing period for three wells – PG2 (bottom), PG5 (midslope) and PG6 (plateau) – and their average distance to the target (mean M and mean D), as well as discharge efficiency and Log discharge efficiency, with different spatial models of variation in physical properties.

Models		K_o decreasing with slope position				K_o and m decreasing with slope position		K_o and porosity decreasing with slope position	
Best fitting regarding discharge	no spatial	linear	threshold	top index (mono)	top index (multi)	linear	threshold	linear	threshold
D–PG 2	−0.09	0.08	−0.08	0.08	0.09	0.09	0.03	0.06	−0.02
D–PG 5	−0.66	−0.59	−0.60	−1.00	−1.25	−0.91	−0.22	−0.63	−0.25
D–PG 6	1.71	2.03	0.96	1.43	1.41	0.96	1.14	1.84	1.44
R–PG 2	−0.21	−0.48	−0.23	−0.49	−0.52	−0.51	−0.48	0.55	0.64
R–PG 5	0.46	0.05	0.30	0.27	0.22	0.44	0.08	1.14	1.03
R–PG 6	−0.39	−0.58	−0.49	−0.45	−0.54	−0.29	−0.51	0.54	0.44
Mean D	1.83	2.11	1.13	1.75	1.88	1.33	1.16	1.94	1.47
Mean R	0.64	0.75	0.62	0.72	0.78	0.73	0.70	0.66	0.67
Eff(Q)	0.59	0.65	0.57	0.63	0.57	0.57	0.22	0.66	0.48
LogEff (Q)	0.77	0.88	0.70	0.88	0.88	0.90	0.63	0.87	0.66
Best fitting regarding water table	no spatial	linear	threshold	top index (mono)	top index (multi)	linear	threshold	linear	threshold
D–PG 2	−0.05	0.12	−0.03	0.10	0.06	0.14	0.03	0.07	−0.04
D–PG 5	−0.69	−0.97	−0.69	−1.01	−1.07	−0.80	−0.22	−0.92	−0.67
D–PG 6	1.52	0.83	0.76	0.80	0.98	0.86	1.14	0.97	0.77
R–PG 2	−0.27	−0.67	−0.33	−0.68	−0.51	−0.71	−0.48	0.49	0.67
R–PG 5	0.55	0.17	0.40	0.28	0.13	0.14	0.08	1.20	1.37
R–PG 6	−0.35	−0.47	−0.35	−0.39	−0.44	−0.45	−0.51	0.59	0.67
Mean D	1.67	1.28	1.02	1.29	1.45	1.18	1.16	1.34	1.02
Mean R	0.49	0.70	0.39	0.68	0.47	0.85	0.70	0.69	0.60
Eff(Q)	0.43	0.45	0.51	0.40	0.57	0.43	0.22	0.57	0.51
LogEff (Q)	0.73	0.64	0.60	0.73	0.77	0.71	0.63	0.71	0.62

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Table 3. The number of acceptable simulations of observations (Eff-Q>0.4 and Efflog-Q>0.7) and the values of the parameters of the best-fitted simulation of discharge and water table depth, with different spatial models of variation in physical properties.

Models		K_o decreasing with slope position				K_o and m decreasing with slope position		K_o and porosity decreasing with slope position		Statistics		
number of behavioral simulations	99	224	32	208	126	165	0	165	7			
Best fitting regarding discharge	no spatial	linear	threshold	top index (mono)	top index (multi)	linear	threshold	linear	threshold	Mean	Sigma	Coefficient of variation
Total porosity (%)	23.0	30.6	27.3	26.6	27.4	23.2	31.2	24.8	28.2	26.9	2.9	11
Drainable Porosity (%)	4.4	5.3	4.2	5.4	5.2	5.4	5.3	5.2	5.0	5.0	0.5	9
b (m^{-1})	4.17	2.27	4.59	2.68	4.06	2.05	3.62	2.11	4.53	3.34	1.06	32
K_o (m/h)	1.43	1.25	1.43	1.13	1.07	1.08	1.35	1.35	1.46	1.28	0.16	12
m (m^{-1})	0.86	0.53	0.90	0.60	0.68	0.60	0.73	0.50	0.77	0.68	0.14	21
K_c (m/h)	0.013	0.005	0.010	0.011	0.002	0.018	0.029	0.013	0.006	0.012	0.008	67
Drainage coefficient	35.0	30.5	32.9	33.4	34.2	33.6	32.9	25.5	29.8	32.0	2.9	9
Best fitting regarding water table	no spatial	linear	threshold	top index (mono)	top index (multi)	linear	threshold	linear	threshold	Mean	Sigma	Coefficient of variation
Total porosity (%)	24.2	31.3	22.1	27.4	27.9	25.7	31.2	29.4	25.0	27.1	3.2	12
Drainable Porosity (%)	3.3	4.6	5.4	4.0	4.8	5.4	5.3	5.1	3.9	4.7	0.8	16
b (m^{-1})	3.53	3.71	2.36	2.61	4.97	4.72	3.62	4.50	4.66	3.85	0.93	24
K_o (m/h)	1.02	0.42	1.16	0.47	1.14	1.30	1.35	0.82	1.18	0.99	0.34	35
m (m^{-1})	0.99	0.79	0.80	0.68	0.52	0.45	0.73	0.58	0.78	0.70	0.17	24
K_c (m/h)	0.015	0.028	0.029	0.030	0.030	0.014	0.029	0.030	0.029	0.026	0.006	25
Drainage coefficient	32.7	30.2	26.1	27.2	31.3	24.5	32.9	33.9	33.5	30.3	3.5	12

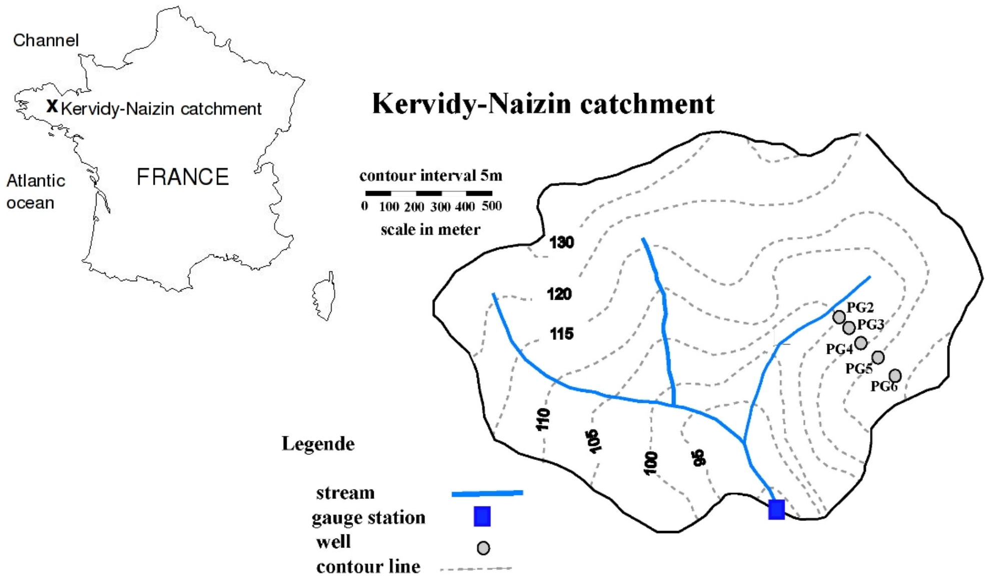

Fig. 1. Kervidy-Naizin catchment (5 km^2, western France).

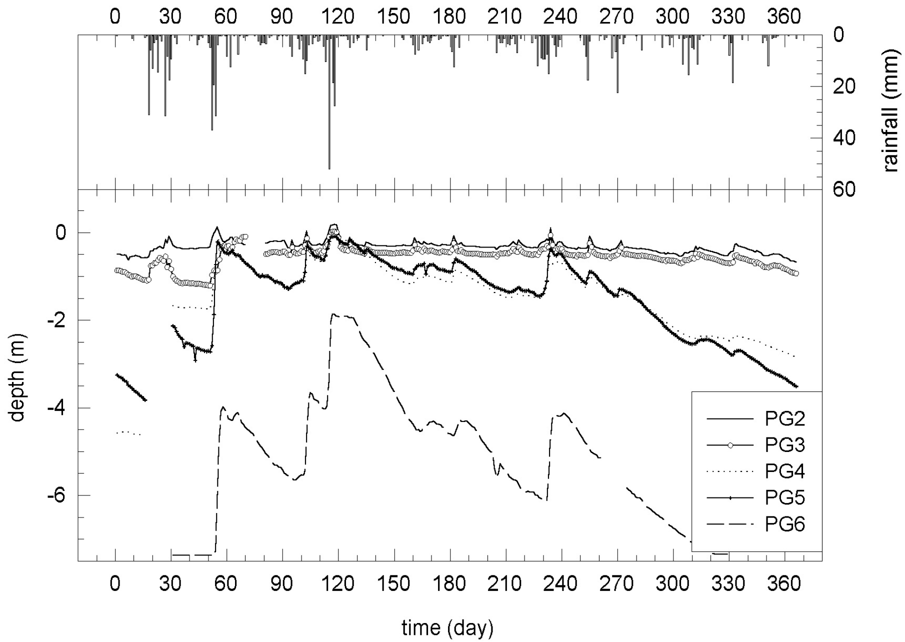

Fig. 2. Temporal variation in precipitation, discharge and water table depth at two wells, one in the bottom domain and one on the summit domain, in the water year 2002–2003.

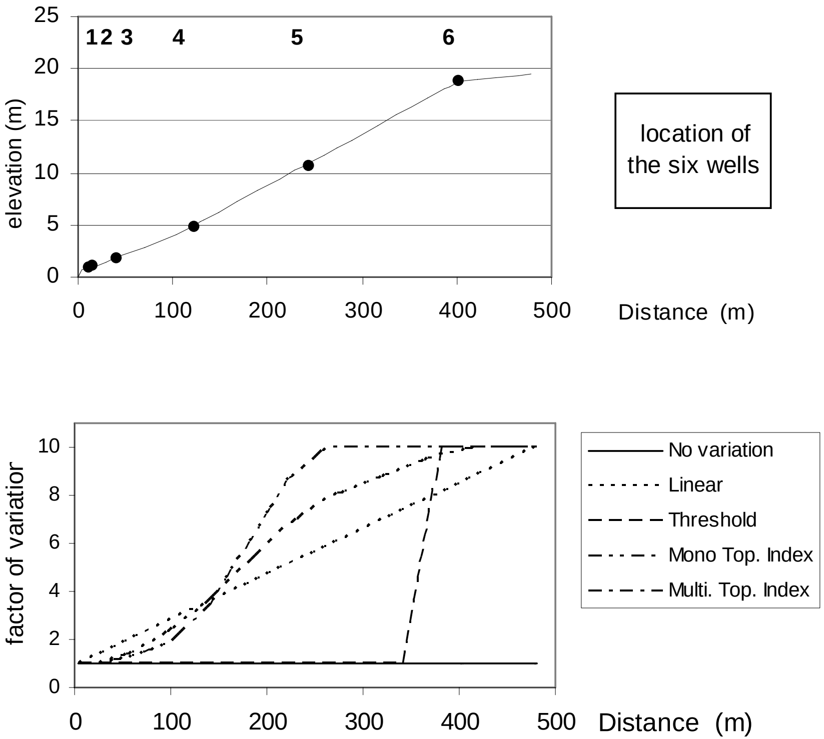

Fig. 3. Normalized factor of spatial variation of the hydraulic conductivity from the bottom to the top domain, according to different spatial models.

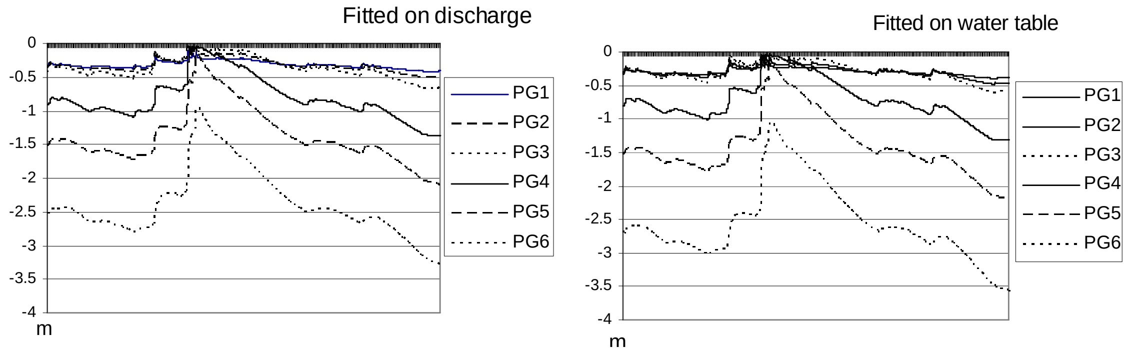

Fig. 4. Temporal variation of water table depth at five wells in the water year 2002–2003 considering no spatial variation of physical properties: **(a)** fitted to discharge; **(b)** fitted to water table depth.

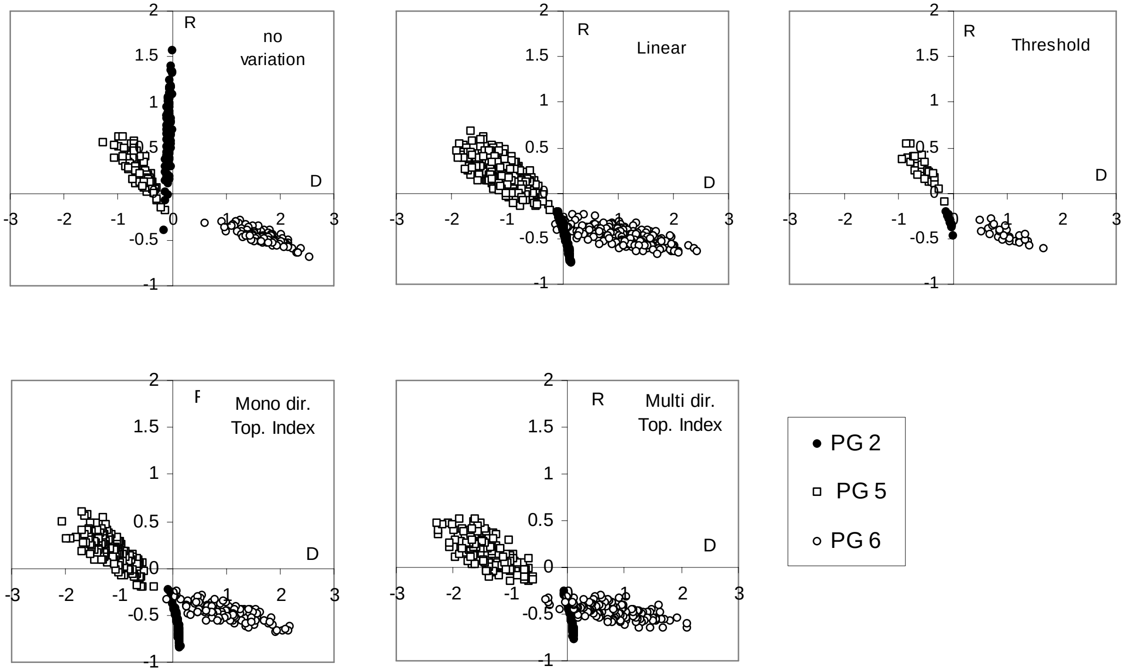

Fig. 5. Mean error (*D*) versus normalized relative range of variations (*R*) of water table depth, computed for the selected simulations (Efficiency>0.4; Log Efficiency>0.7), during the tested period, for three wells – PG2 (bottom), PG5 (midslope) and PG6 (plateau) – and with different spatial models of variation in hydraulic conductivity.

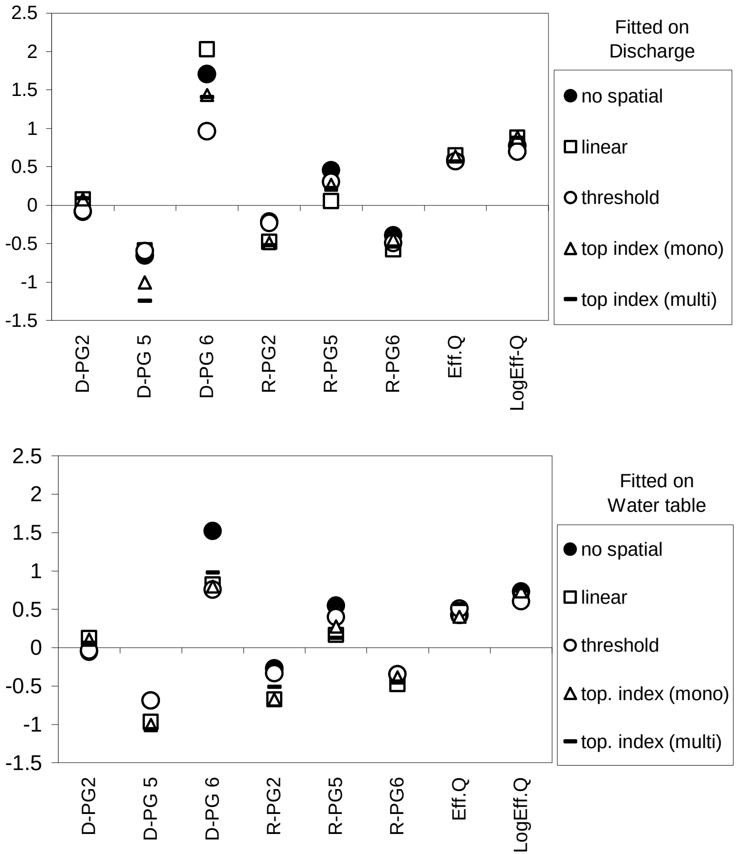

Fig. 6. Mean error (*D*) and normalized relative range of variations (*R*) of water table depth, computed for the best simulation fitted to: **(a)** discharge and **(b)** water table depth, during the tested period, for three wells – PG2 (bottom), PG5 (midslope) and PG6 (plateau) – and with different spatial models of variation in hydraulic conductivity.

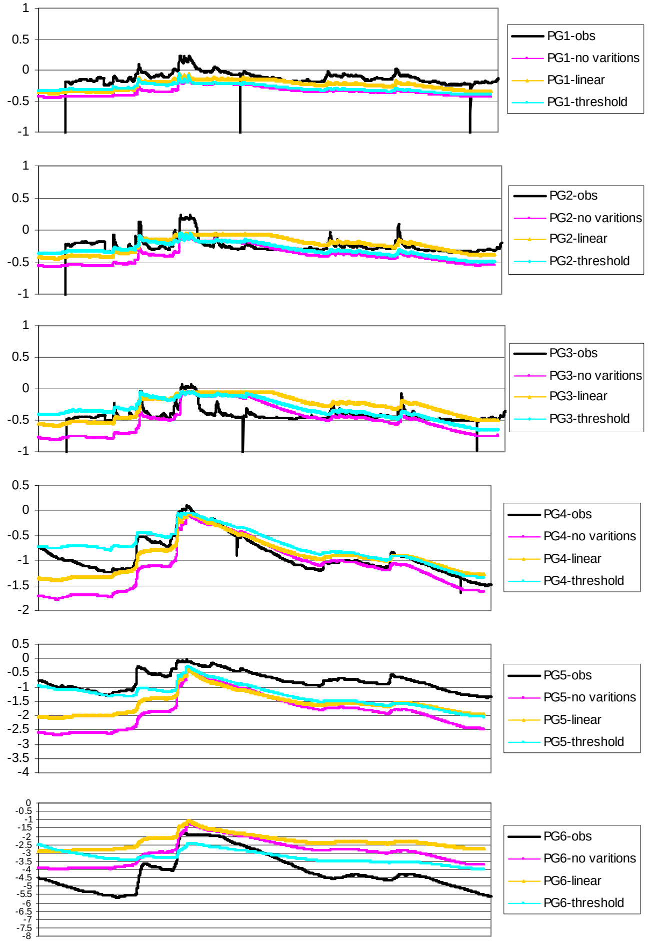

Fig. 7. Daily variation of water table depth for six wells and different spatial models of variation in saturated hydraulic conductivity.

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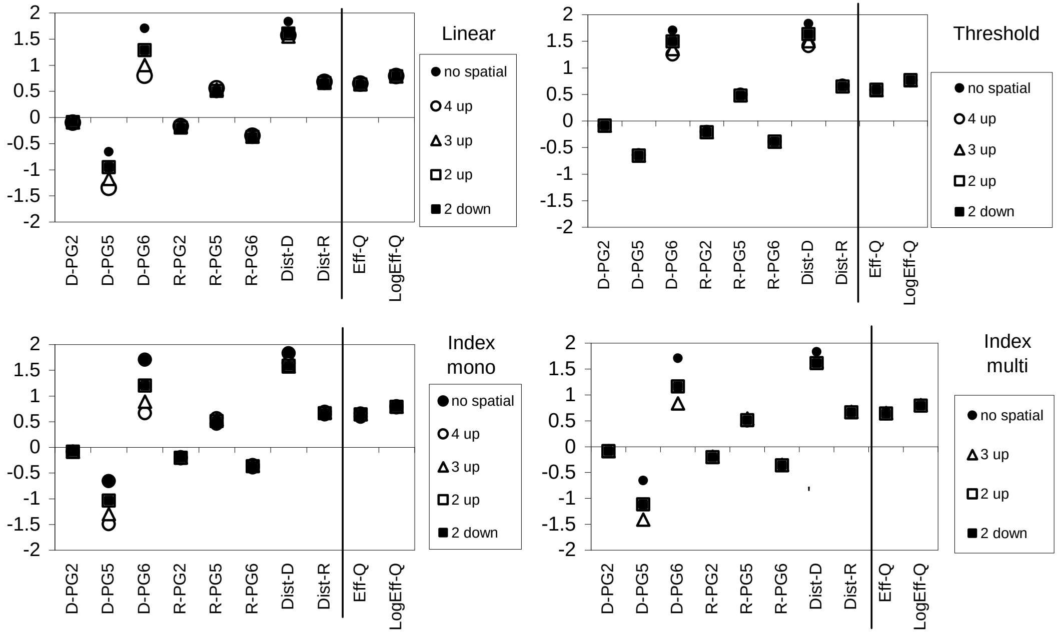

Fig. 8. Mean error (*D*) and normalized relative range of variations (*R*) of water table depth, computed for a reference case considering a different range of spatial variation of saturated hydraulic conductivity from the top to the bottom and reverse, during the tested period, for three wells – PG2 (bottom), PG5 (midslope) and PG6 (plateau) – and with different spatial models of variation in hydraulic conductivity.

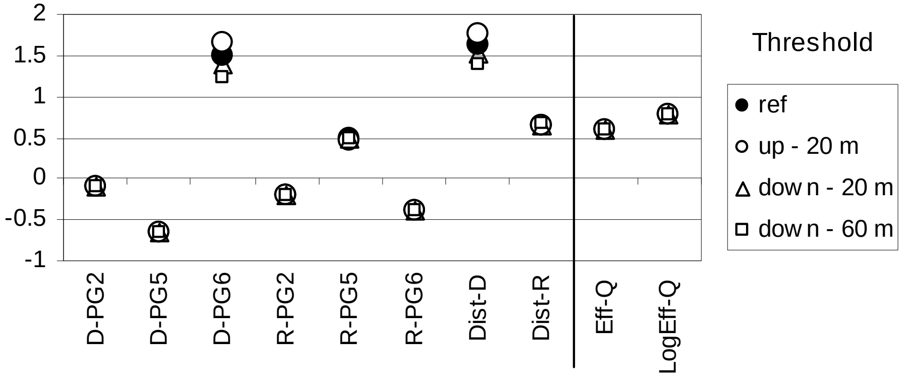

Fig. 9. Mean error (*D*) and normalized relative range of variations (*R*) of water table depth, computed for a reference case considering different slope positions of the threshold, using the threshold spatial model of variation in saturated hydraulic conductivity, for three wells – PG2 (bottom), PG5 (midslope) and PG6 (plateau) – during the tested period.

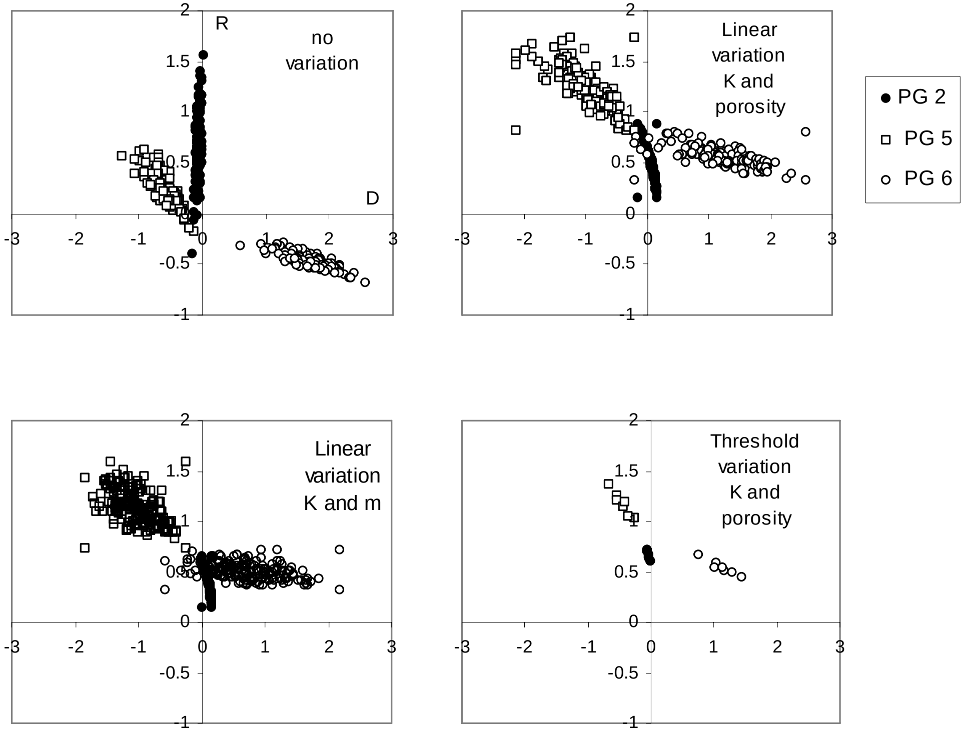

Fig. 10. Mean error (D) versus normalized relative range of variations (R) of water table depth, computed for acceptable simulations (Efficiency>0.4; Log Efficiency>0.7), considering spatial variation of K_s, m and drainable porosity and using two spatial models of saturated hydraulic conductivity, during the testing period, for three wells – PG2 (bottom), PG5 (midslope) and PG6 (plateau).

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