On the necessity to use 3D groundwater models for describing impact of drought conditions on the stream-flow regimes

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Abstract Integrated hydrological models are used to assess the impact of global climatic changes on the water cycle. Very often, they are supposed to represent with physical consistency the exchanged water flows between the different parts of the hydrological cycle. In drought conditions, it is particularly important to study with accuracy the base-flow groundwater component for a good assessment of the stream-flow. However, in most of the integrated models, the contribution of groundwater in the total stream-flow is calculated by use of an empirical or lumped coefficient (i.e. recession coefficient). Due to the lack of recharge during the long drought periods, a general piezometric drop is induced. In heterogeneous and especially in fissured aquifers, the base-flow is changed due essentially to desaturation of the upper zone. By this way, the geometry of the groundwater basin and the local values of hydraulic permeability and effective porosity are influencing the groundwater contribution. Consequently, 3D groundwater models are strongly recommended in order to be able to describe locally different hydrodynamic characteristics in the aquifer. These models should be able to describe explicitly the river-aquifer exchanges with physical consistency, including space and time variations of the base-flow component.

INTRODUCTION

There is a common belief that climatic changes may have a stronger effect on extremes than on mean values of hydrological variables (Knox & Kundzewicz, 1997). Quantitative results of the impact of climate change on river flow regimes and especially low flow regimes are not easy to calculate. In drought conditions, very high percentages of the stream-flows are provided by the groundwater base-flow. When studying climate change impacts on stream-flows, most of the authors are thinking about rainfall-runoff models, which are depending directly on the climate scenario (some scenarios are predicting longer summer drought periods and of course all of them are predicting more potential evapotranspiration) with very short time delays. One must think also to the so-called second-order effects where the changes in land-use, vegetation and irrigation practices will influence groundwater and river flows. As mentioned by Sefton & Boorman (1997), various aspects of this question have been studied using monthly water balance models or daily models (Nemec & Schaake, 1982; Bultot *et al.*, 1988). Annual and seasonal changes were addressed rather than changes in extreme high or low flows. Physical descriptors such as catchment area, stream length, slopes, drainage density and some index of soil type are used. For groundwater aspects, these models are abusively called physical models, as they are more close to fitted black-box models, using semi-empirical parameters describing globally the physical reality. Often authors are concluding that 'the good agreement between observed and modelled flow duration curves demonstrates the ability of the model to simulate the overall pattern of flow response across the whole range of flows'. However, for the hydrogeological point of view, no confidence can be given to such models when it is foreseen to simulate solicitation conditions lying out of the calibration range.

RECESSION COEFFICIENT: AN OLD-FASHIONED CONCEPT

The groundwater component is generally roughly approximated by only one coefficient for the whole concerned basin. Different simple approximations of the recession curves were proposed since the beginning of the century. In most of the cases, a coefficient is taken equal to the fitted recession coefficient used in the scope of the chosen model. For example, using the Maillet equations (1905), a recession coefficient is usually found by mean of a straight line in a $(t, \log Q)$ hydrograph. There are four main reasons explaining why recession coefficient is a lumped coefficient, which should be disregarded in accurate integrated modelling of hydrological basins. **(1)** The choice or the fitting of its value is very subjective because in the reality, even in a $(t, \log Q)$ hydrograph, a straight line can never be found. This is due to the actual shape (ellipse in a homogeneous medium) of the dropping free surface leading to a curve-type $(t, \log Q)$ hydrograph (Fig. 1). A straight line can only be found if drainage is supposed to occur keeping in the whole aquifer a perfectly horizontal water table (the aquifer is drained just like an emptying bathtub!). This oversimplified concept is needed to use, for example, a Maillet recession coefficient. **(2)** It consists in a lumped coefficient where the 3D reality, the actual geometry of the basin and the actual form of the groundwater/surface-water interface are totally neglected. To describe with accuracy local and regional variations of groundwater contribution to the stream-flow, it should be needed to take into account geometrical characteristics of the actual interface and the different local shapes and variations of the groundwater table. **(3)** Depending on heterogeneous geological conditions, local variations are to be expected for the groundwater flow reaching the stream. The recession coefficient calculated at the basin scale provides only a kind of global averaged value, which dissimulates the strong spatial variations and leads to misunderstanding when sub-basins or local zones are considered. **(4)** In the same way, the local recession coefficient is largely dependent on the local properties. In layered and fissured aquifers, the hydraulic conductivity (K) and the effective porosity (n_e) values have most often decreasing values with the depth. In consequence, during droughts, as deeper layers of the aquifer are concerned the base-flow is strongly influenced by: a) the lowering of the water table, b) the decreasing of (K) inducing a slower time variation of the base-flow rate, and, c) the decreasing of (n_e) inducing a higher time variation of the base-flow rate.

To illustrate this last case together with the effect on the recession curve of heterogeneity, 2D vertical simulations of the drainage of a heterogeneous water table aquifer is carried out using MODFLOW. As mentioned previously (1), the conceptual analogy with the ‘emptying bathtub’ is used. For an aquifer consisting in four horizontal layers of equal thickness and decreasing effective porosities to the bottom, results are given in Fig. 2. In this case, and a fortiori, in more complex heterogeneous cases, the fitting of a unique recession coefficient is difficult to accept. Multiple 2D simulations of this type can be carried out for different kinds of heterogeneous conditions. The results confirm the four main observations mentioned here above.

PHYSICALLY CONSISTENT MODELS IN 3D

The groundwater/surface-water interactions are physically driven by the difference of water levels between groundwater and surface-water, hydraulic properties (conductivity and effective porosity) and geometry of the interfering geological layers. The effective porosity has an overwhelming influence in the desaturation zone as it gives the quantity of water (per volume of porous medium) which is drained for a fall of 1 m in piezometric head. In the saturated zone, nearby the groundwater/surface-water interface, the groundwater flow into the stream is largely influenced by local hydraulic conductivity values. These properties are highly variable in space and particularly with depth. In drought conditions, if groundwater levels drop due to the lack of infiltration the gradient is changed but the concerned properties also.

The only way to simulate adequately the physics of groundwater/surface-water exchanges consists in implementing 3D groundwater flow models (GWM's) coupled with river models (RM's). When variations of water levels in the river must also be computed, the proposed solution is based on parallel runs of the GWM and RM. Exchanged water and solute mass fluxes through the contact interface are calculated by a junction on the basis of the received results from each model at each time step. As each model has its own time and space discretization, the junction must organize the data exchanges, including various time and space interpolation schemes (Carabin & Dassargues, 1999). This approach is being adopted in a research project entitled 'Integrated modeling of the hydrological cycle in relation to global climate changes', applied on five sub-basins in Belgium.

An example of results from such 3D coupled simulations (using SUFT3D, Carabin *et al.*, 1998) is given here below. Exchanges between a portion of an actual river (near Antwerp, Belgium) and the alluvial aquifer (loam, sand and gravels) are calculated in transient conditions during the year 1993 taking all other hydrological conditions (infiltration, sink/source terms) into account. The river model and the 3D groundwater model are fully coupled through the expression of an exchange of water, which is depending on the difference of the water levels (the only way to be physically consistent). Simulations have been done with the river draining most of the time the aquifer. The computed time fluctuations of the exchanged water flux are given in Fig. 3. On basis of these results, it is easy to understand that the concept of recession coefficient is an approximation, which leads to smooth the actual description of exchanged flow-rates. In many cases, and certainly when this coefficient is re-used in studies involving new conditions (not lying in the measured range of variation), this smoothing can also create a bias with regards to the reality in function of space and time.

Some fully integrated models have previously been developed (Refsgaard *et al.*, 1995; Van Lanen *et al.*, 1997) but until now they seemed to be limited in their way to discretize spatially the groundwater domain and in the way to describe water exchanges between river and groundwater.

CONCLUSIONS

The effects of global climate changes on the hydrological cycle taken globally and locally are still poorly understood and not well quantified. Very often, the concept of linear reservoirs and corresponding recession coefficients serves as a tool for simulations of base-flow. Buchtele *et al.* (1996) and Panagoulia & Dimou (1996) admitted that these kinds of methods are oversimplifying the description of the interaction between groundwater and stream-flow. It is admitted by hydrogeologists that even if this concept of recession coefficient was very useful in the past, the more powerful computers and the more detailed data-sets allow now to abandon this oversimplified coefficient for computing groundwater contributions in the stream-flow.

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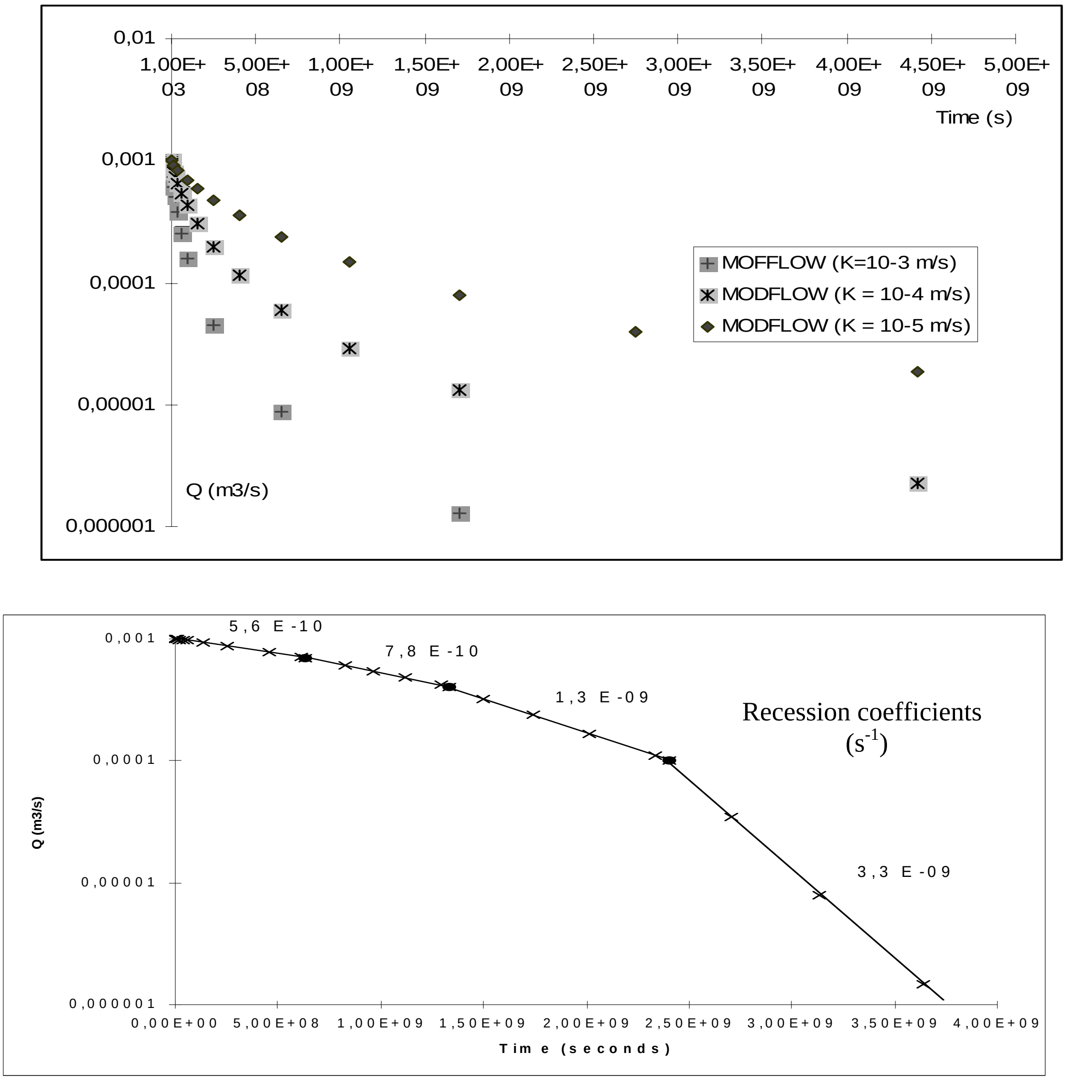
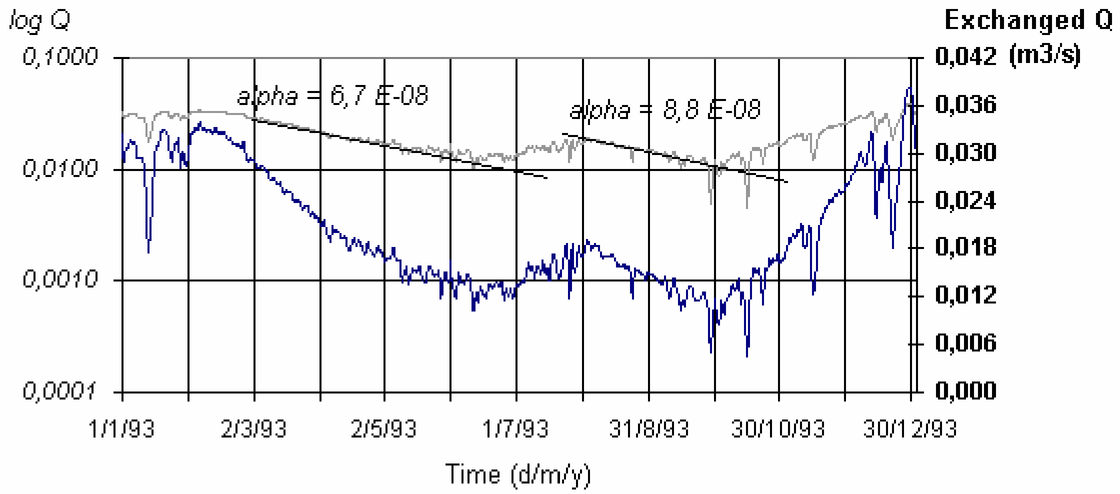
FIGURE CAPTIONS

Fig. 1 2D vertical simulations (using MODFLOW) of the natural drainage of a homogeneous water table aquifer. Results in a $(t, \log Q)$ diagram confirm that a straight line can not be found. Influence of K values on the results can also be distinguished.

Fig. 2 2D vertical simulations of the drainage of a heterogeneous aquifer (4 horizontal layers with decreasing n_e towards bottom) with a horizontal water table. Results in $(t, \log Q)$ diagram confirm that 4 different values of recession coefficient (s^{-1}) can be distinguished depending on each layer effective porosity.

Fig. 3 3D simulations (using SUFT3D). Exchanges between a portion of an actual river and the alluvial aquifer during the year 1993 (water fluxes from the aquifer to the river are positive). Q is the computed exchange flow rate (m^3/s) between the river the aquifer, the fitted recession coefficients (using the Maillet equation) are calculated in s^{-1}.




log Q
0,1000
0,0100
0,0010
0,0001
Exchanged Q (m3/s)
0,042
0,036
0,030
0,024
0,018
0,012
0,006
0,000
alpha = 6,7 E-08
alpha = 8,8 E-08
1/1/93
2/3/93
2/5/93
1/7/93
31/8/93
30/10/93
30/12/93
Time (d/m/y)