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3D modeling of a complex alluvial aquifer for efficient management – application to the lower valley of Var river, France

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ABSTRACT. – Complex alluvial aquifers are frequently essential resources for cities that are looking for reliable and safe resources for water supply. The location of such aquifers is frequently combined with intense urban developments that may, potentially, generate pollutants that could affect badly the quality of the resources. An efficient and safe management of water supply system requests to develop a good knowledge of the dynamics of the aquifer and to characterize the exchanges that could exist with free surface flows in associated rivers. The most efficient way to understand the behaviour of the aquifer is to implement a 3D physically-based hydrodynamic model that could represent all physical processes. However, this approach, in order to become an operational tool, requests a structured methodology for data integration and validation. The paper describes the construction of a 3D hydraulic model of groundwater flow in the Var lower valley, on the French Riviera, with FEFLOW modeling system. Despite a very complex geological structure and a limited knowledge on the aquifer itself, the results demonstrate that the model is able to represent the groundwater flows over long chronologies and to provide an accurate diagnostic on various hydraulic structures that are affecting negatively the aquifer conservation.

Key-words: 3D modeling, complex alluvial aquifer, river-aquifer exchange, FEFLOW, Var river.

Modélisation 3D d'un aquifère alluvial complexe pour la gestion des ressources en eau - Application à la basse vallée du Var, France

RÉSUMÉ. – Les nappes alluviales complexes constituent fréquemment des ressources en eau essentielles pour les communes à la recherche d'un approvisionnement en eau durable. De tels aquifères sont généralement localisés dans des zones où le développement urbain est considérable et peut induire une pollution accidentelle ou chronique des eaux souterraines. Afin d'assurer une gestion efficiente de ces ressources, la connaissance de la dynamique de l'aquifère et des échanges éventuels avec les cours d'eau est essentielle. La méthode la plus appropriée s'appuie sur la conception d'un modèle hydraulique tridimensionnel déterministe. La réalisation d'un outil opérationnel requiert une attention particulière pour l'intégration et la validation de multiples données provenant de diverses sources. Cet article décrit les étapes de construction du modèle hydraulique 3D de la nappe alluviale dans la basse vallée du Var (France), à l'aide du system de modélisation FEFLOW. Malgré une structure géologique complexe et une connaissance limitée de l'aquifère, les résultats démontrent que le modèle est capable de représenter avec succès les écoulements souterrains sur de longues périodes. L'outil permet également d'évaluer l'influence négative de certains ouvrages sur la conservation de cette nappe alluviale.

Mots-clés : modélisation en 3D, nappe alluviale complexe, échanges nappe-rivière, FEFLOW, Var.

I. INTRODUCTION

Groundwater is an important drinking water resource in both urban and rural areas. It is currently the primary freshwater source for approximately two billion people all over the world [Alley, 2006; Kundzewicz and Döll, 2009]. The groundwater management emphasizes two major aspects: controlling the quality of the natural water resource and maintaining the quantity of water supply [Gourbesville, 2008; Das and Datta, 2001]. Understanding the exchange of water between the river and its aquifer is a key issue for the long-term water management regarding these two aspects.

Located in the southeast of France (Fig.1), the lower Var river valley [Potot, 2011; Guinot and Gourbesville, 2003]is the last section of the Var river that drains the water from the mountainous area to the Mediterranean. The groundwater in the unconfined alluvial aquifer is a main water resource for around 600,000 inhabitants who live in the cities and towns near the river mouth such as Nice and St Laurent du Var [Potot et al., 2012]. A Previous study indicates that the shallow aquifer interacts strongly not only with the river, but also with the bedrock underneath the alluvium [Guglielmi, 1993].

During the last century, the human activities affected the environment of this area by developing the land of the natural flood plain. Thus, artificial modifications have significantly reshaped the river morphology and, therefore, the riverbed is strictly limited within a smaller width than its natural form. This led to an increase of the flow velocities associated with a stronger sediment transport. Hence, the riverbed level decreased gradually because of this long-term erosion process. At the same time, the groundwater withdrawal was observed in the shallow aquifer. In August 1967, a severe drought happened in the Riviera area and it had a considerable impact on the drinking water supply in lower

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Var river valley. In some places, the groundwater table fell by 8 meters. Thus the agricultural activities were widely affected by the poor water condition. From 1971 to 1986, 11 weirs have been constructed within the riverbed in order to, as initial purpose, prevent the groundwater table withdrawal.

Several studies have been carried out to estimate the water exchange between the river and its aquifer. Guglielmi [1993] studied the hydrogeological characteristics including hydraulic conductivity of the Holocene deposits and the Pliocene conglomerate. He inferred also the direction of exchange in different sections of the valley, based on the instantaneous iso-contour map of the groundwater level. The result, however, indicates only qualitatively the instant direction of the exchange of water, which is less applicable by the water management services. Likewise, similar analysis has been carried out by Guglielmi and Mudry [1996], Emily et al. [2010]. These researches focused on different hydrological periods and different sections of the lower Var river valley, so they complete the knowledge of the function of the aquifer. Nevertheless, the methods used by these studies are based on the instantaneous iso-contour map of groundwater level, which lacks of continuity over time.

In the current case, a further knowledge of the hydraulic function of the aquifer is needed by the authority to conduct an integrated management of the groundwater. 3D physically based hydraulic model is the most appropriate and feasible tool to study the physical process and to ensure an efficient groundwater management. However, the model-based operational tool requests a structured methodology for data integration and validation. The objectives of this study are: (a) to develop a hydraulic model of the groundwater flow of the lower Var river valley using FEFLOW; (b) to calibrate the unmeasured parameters according to a sensitivity analysis; (c) to evaluate the impacts of the weirs on the groundwater table.

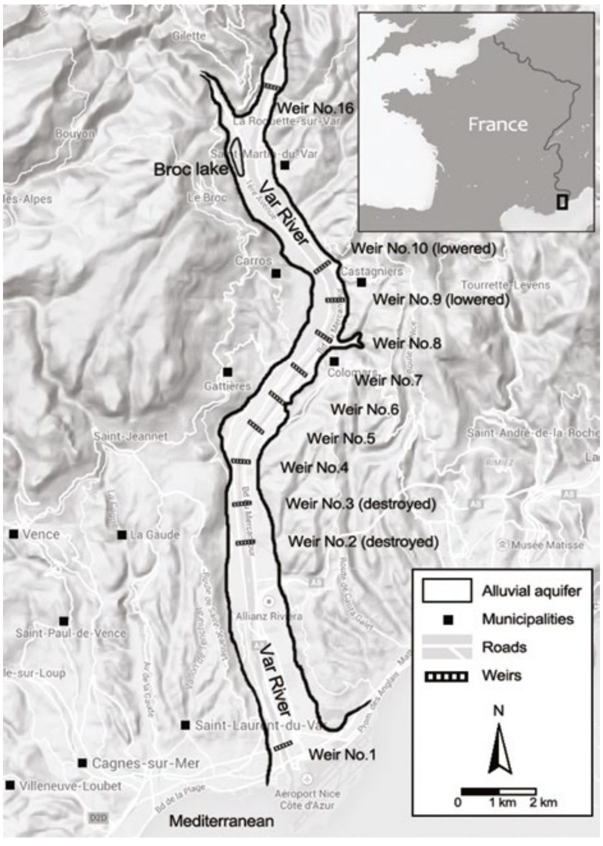

Figure 1: *Study area: the lower Var river valley*

II. METHODOLOGY

II.1. Governing equations

FEFLOW solves the 3D groundwater flow equations in porous media with finite element method. The code has been validated by a numerous case studies [Diersch, 2005] and applied for research and engineering projets [Zhao et al., 2005; Ashraf et al., 2008; Gaultier ,2012]. The governing equations are fluid mass conservation equation and the Darcy equation [Diersch and Kolditz, 1998]:

$$S_s \cdot s(h)\frac{\partial h}{\partial t} + \varepsilon \frac{\partial s(h)}{\partial t} + \nabla \boldsymbol{q} = Q \qquad (1)$$

$$q = -K_r(s)\boldsymbol{K}\left(\nabla \psi_g + \chi \boldsymbol{e}\right) \qquad (2)$$

where, $\psi_g = h + z$ is the hydraulic head (m), z is the elevation (m), h is the pressure head (m), $s(h)$ is the saturation, (s = 1 if medium is saturated), $\boldsymbol{q}$ is the Darcy flux vector (m/s), Q is the specific mass supply (m/s), $S_s = \varepsilon\gamma + (1-\varepsilon)\Upsilon$ is the specific storage due to fluid and medium compressibility, ε is the porosity, γ is fluid compressibility, Υ is the coefficient of skeleton compressibility, $K_r(s)$ is the relative hydraulic conductivity, ($0 < K_r < 1$, K_r = 1 if saturated at s = 1), $\boldsymbol{K}$ is the hydraulic conductivity for the saturated medium, χ is the buoyancy coefficient including fluid density effects, $\boldsymbol{e}$ the gravitational unit vector.

II.2. Model domain and geological layers

The modelled domain is the total area of the hydrogeological catchment that is delimited by geological faults and the scope of permeable geological layers. Within this range, all the precipitated water contributes to recharge the unconfined aquifer in the valley.

Three types of mesh are used to discretize the computational domain : 25 m for the riverbed and pumping stations where the water exchange are the most intensive; 50 m for the alluvial aquifer which is the main studied area; 100 m for the rest model domain. The mesh is smoothed so that no obtuse angles are created.

The northern boundary of the model is set at the weir No.16 where the data of a piezometer is used as the upstream boundary condition. The southern boundary is the Mediterranean sea. Knowing that the alluvial aquifer is connected to the sea water [Guglielmi, 1993], the sea level is used as the downstream boundary condition. It forms a study area of 146.45 km2, with a river length of 22.4 km (Fig.2(a)).

Vertically, the model contains the layers of recent alluvium, alluvial terraces, Pliocene conglomerate, Pliocene marls, impermeable layer from Miocene to Cretaceous, and the Jurassic limestone. Even though the model focuses on the flow in the first three layers where the unconfined aquifer exists, it is important to include the other layers to reduce the bottom boundary influence on the top layers.

As a key material property, the hydraulic conductivity is a decisive input data of the model. In reality, the recent alluvium, alluvial terraces, conglomerate, marl are homogeneous in different flow directions, while in the limestone the flow direction depends on the fissures. Since there is only a small section of the valley where the limestone contacts directly the alluvium, it could be acceptable to assume that the flow in the limestone is also homogeneous in order to simplify the model.

The hydraulic conductivity of the alluvium is measured on some points of interest. So the measured values are interpolated and used directly as an input data. As for the other layers, the value is assigned uniformly, because of their minor importance regarding the groundwater flow in the shallow aquifer. Guglielmi [1993] has estimated that the hydraulic conductivity of the Pliocene conglomerate is 2.6×10^{-6} m/s. For the other soil and rock, only empirical values are available due to the lack of filed measurement. The distribution of the hydraulic conductivity value is shown in Fig.2(b).

II.3. Source and sink terms

Except for the boundary conditions, the main factors that influence the quantity of the groundwater are the water supply or extraction, treated as the source/sink term Q in the groundwater flow equation. In the Var valley, the direct water recharge, surface water body and the water pumping are considered as the predominant source/sink term that may have a strong impact on the groundwater flow (Fig.3).

II.3.1. Direct water recharge Q_1

The direct water recharge depends only on the rainfall and on AET (actual evapotranspiration). The former is measured directly, while the later can be calculated by using the temperature data. The daily observed data are recorded by the meteorological station of Nice airport. The source/sink term can be written as:

$$Q_1 = \alpha \cdot P - AET \tag{3}$$

where, Q is the direct recharge (mm), this value is positive if the groundwater is being recharged, negative if there is a withdrawal, α is the percentage of the surface permeability estimated according to the land use information, P is the precipitation depth (mm), AET is actual evapotranspiration (mm).

Delaroziere-Bouillin [1971] concludes that in the southern France, the Turc formula is the most suitable method to estimate the AET. However, since this formula requires many parameters measured in the field, it is difficult to collect all the data needed to apply this formula. Thus a simplified Thornthwaite formula based on water balance method [Laborde, 2010] is used to meet the availability of observed data in order to estimate the AET in this case study.

II.3.2. River aquifer exchange Q_2

The river water level is simulated by a 1D river hydraulic model of Var river built with MIKE11 software and calibrated in previous studies. As the shallow aquifer is unconfined, the river water level affects significantly the groundwater table. In FEFLOW, the flux of river-aquifer exchange is given by the formula below which is deduced from the Darcy equation:

$$Q_2 = A \cdot \varphi \cdot \left(\psi_s - \psi_g\right) \tag{4}$$

where, Q_2 is transfer flowrate (m³/s), a positive value corresponds to infiltration, and a negative one for exfiltration, A is the concerned area on the river bed, φ is transfer coefficient (s^{-1}), ψ_s and ψ_g are respectively hydraulic heads of the surface and groundwater (m). According to the Darcy equation, the transfer rate φ is defined as:

$$\varphi = K_0 / d \tag{5}$$

where, K_0 is hydraulic conductivity of the clogging layer (m/s), d is the thickness of the clogging layer (m).

Actually, the transfer rate is too difficult to be quantified because, a) the d can be hardly measured; b) the sediment type varies from one place to another along and across the river therefore the K_0 is difficult to be estimated; c) the submerged area of riverbed is different for rainy season and for drought season, hence, both K_0 and d are variable. For all of these reasons, the transfer rate needs to be calibrated.

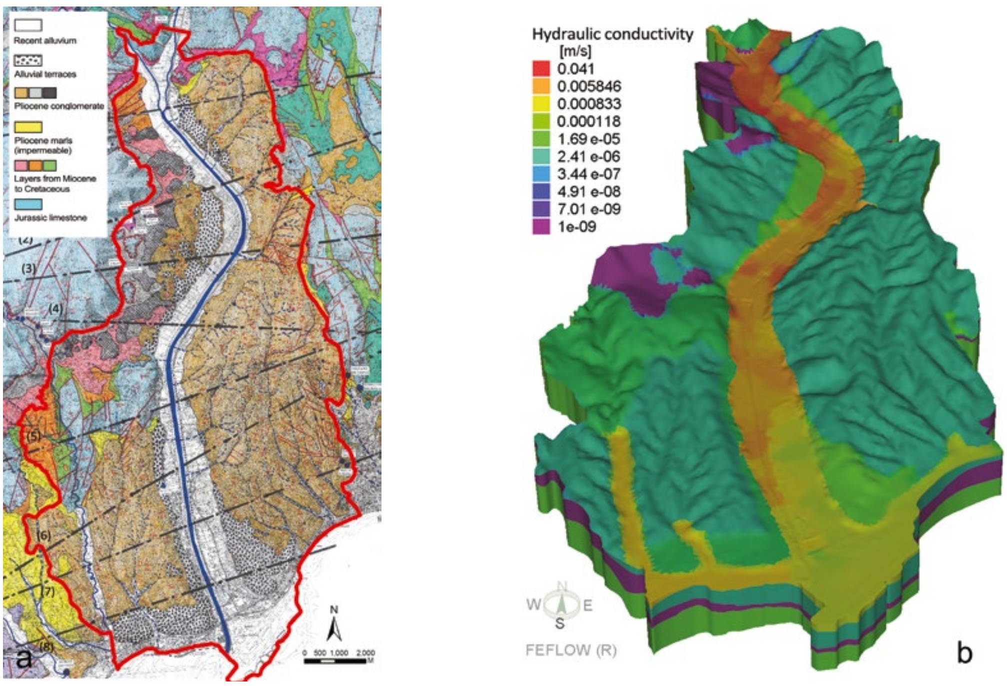

Figure 2: *The model domain (red contour) according to the geological map given by Emily et al. [2010] (a) and the interpolated hydraulic conductivity with 3D view in FEFLOW (b).*

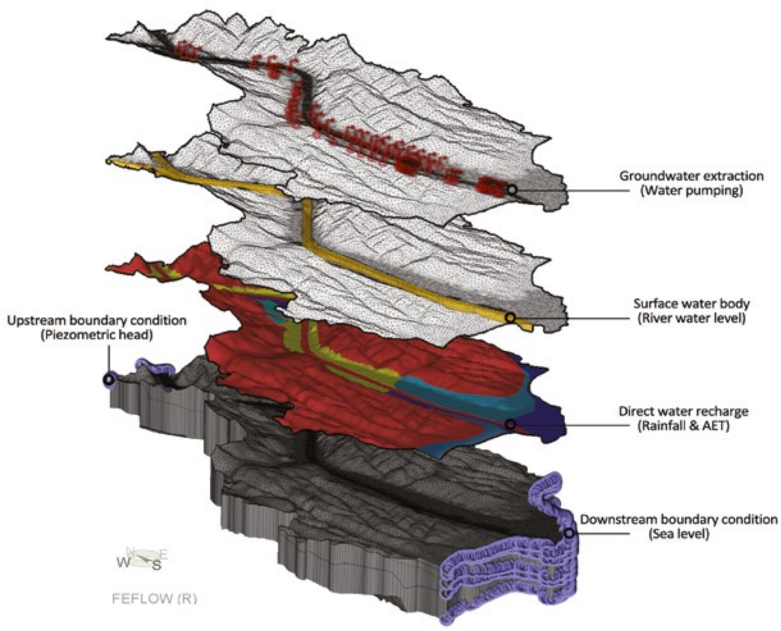

Figure 3: *Boundary conditions and source/sink terms considered in the 3D model.*

II.3.3. Groundwater extraction Q_3

As an important resource of the nearby municipalities, the groundwater extraction influences significantly the groundwater table due to the huge amount of water consumption. Several pumping stations along the Var river have been built to meet the demand of domestic water use. The pumping volume was recorded and controlled by the local water management authority. For the agricultural and industrial water uses, however, the landowners have created only individual and private pumping wells. The pumping volume is not recorded neither by the landowners nor the water management authority.

In order to estimate the non-recorded pumping rate, we assume that this withdrawal is totally caused by water pumping for irrigation purpose, and the water is then totally evaporated by the crops. Consequently, the flux of water pumping on unit area can be estimated by the gradient of the evolution of the groundwater table level over time, which is -0.015 m/d on this area, where the minus sign is used to indicate that the aquifer is losing water. This value is inflicted on all the agricultural land.

III. SENSITIVITY ANALYSIS

The grid convergence of the numerical model is studied by performing simulations with different cell sizes. A small area in the middle section of the lower valley is chosen to perform the simulations with 4 cell sizes: 100 m, 50 m, 25 m and 10 m, and two points are set to compare the simulation results (Fig.4). Point 1 is located on the upstream area of the weir No.4, where the river feeds the aquifer. Point 2 is located on the downstream side of the weir, where the aquifer feeds the river. Therefore, the simulations results cover the two possible river-aquifer exchange directions. The simulations are performed from 1st Nov. to 30th Nov. 2014, with a time step of 15 minutes. For these 4 simulations, the upstream boundary condition (north) is the groundwater level measured by piezometer P57 and the downstream boundary condition (south) is the groundwater level measured by piezometer P16. The initial condition of the simulations is the interpolated groundwater level between the northern and the southern boundaries. Thus the

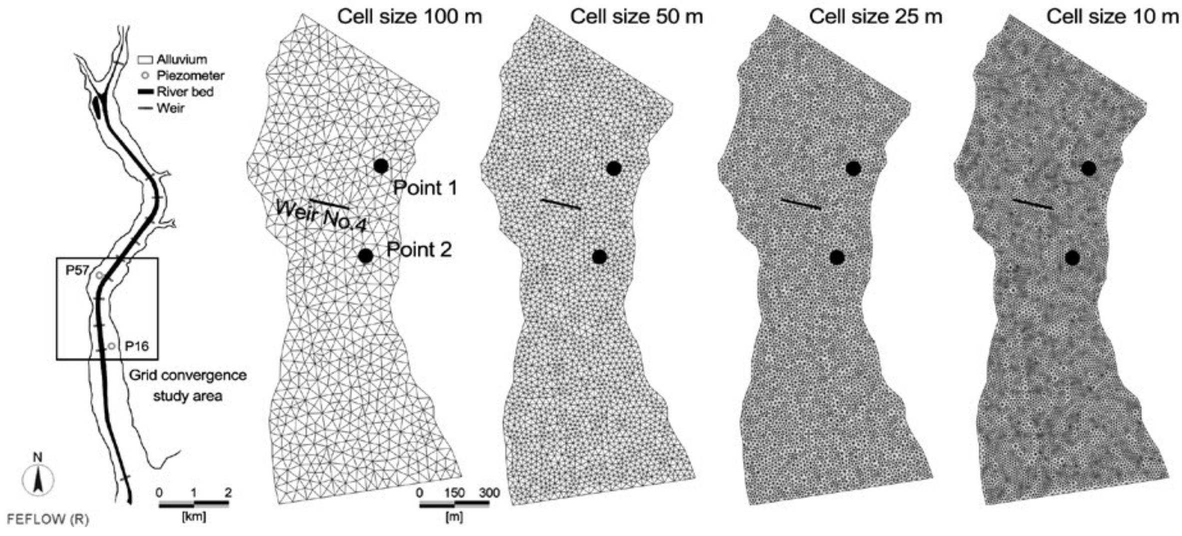

Figure 4: *Location of the area for grid convergence study and 4 cell sizes.*

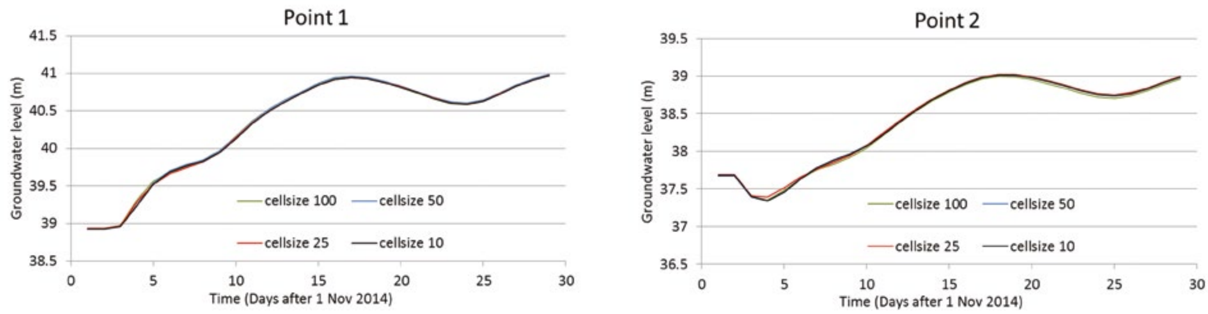

Figure 5: *Comparison of the simulation results for grid convergence study.*

only variable of these 4 simulations is the cell size. The results of the simulations at two points are shown in the Fig.5. It can be seen that, with the same input data, the groundwater level simulated with different cell sizes show almost no difference. Therefore the model is proved to be grid-independent.

Kassem et al. [1997] has also made an appraisal of the transfer rate φ between the river and the aquifer in the lower valley. Only 10^{-4} s^{-1} was given as a rough result for the whole 20 kilometers long river.

Apparently, this conclusion is too coarse to be the input data of the model because they may impact significantly the model output. Since the characteristic of the river changes from the upstream to the downstream, it is obvious that these values must also be spatially distributed along the river. A sensitivity analysis is therefore needed to quantify the in/out-transfer rate.

According to the availability of the observed data, the duration of the simulation is 293 days, from 10th May 2012 to 26th Feb. 2013, which covers an entire drought season and an intensive rainfall event during a hydrological year. In the lower Var river valley, the groundwater level is measured and recorded automatically in daily scale by the national subsurface database of France (BSS-Eau), and published by the access of national groundwater data of France (ADES) [Chery and Cattan, 2003; Chery et al., 2008]. The measurement of groundwater level is based on the reference of benchmarks of general levelling of France (NGF), which is also the geographic reference of the data used to set up the model. The measurement of groundwater level by BSS has a precision of 0.01 m.

In the database of BSS-Eau, there are 21 piezometers installed along the valley to measure the groundwater level of the unconfined alluvial aquifer. Among them, 6 piezometers are chosen to conduct the sensitivity analysis and the model calibration. They are equally distributed from upstream to downstream in order to be representative (Fig.6). Besides, the groundwater level measured at piezometer P37 is used as the upstream boundary condition, assigned on the northern border of the model. As for the downstream boundary condition, the average tidal level 0.3 m is used.

A set of simulations are carried out to evaluate the impact of non-measured parameters on the model output, so as to acquire the value of these parameters by model calibration. The designed combination of parameters is listed in the Tab.1. The values to be tested are within a reasonable range verified by few studies mentioned above. Normally, the φ_{out} is larger than the φ_{in}, because the clean groundwater tends to "flushes" the pore space in the clogging layer.

The sensitivity analysis of transfer rate can also indicate the feeding direction of an area. A place that is more sensitive to the change of in-transfer rate means that the river feeds the aquifer, and vice-versa. According to the results shown in the Fig.7, the point of PZ_LIG is merely affected by the change of in-transfer rate values, because the exfiltration is the predominant feeding direction in this area. Spatially, the impact of in-transfer rate is more prominent in the section of weirs than in the downstream area. Temporally, the impact is more effective during the drought season than the flood season, except for the section weirs where the influence lasts for the whole hydrological year due to the severe groundwater withdrawal.

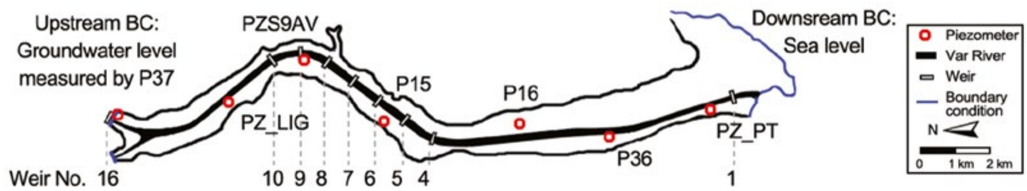

Figure 6: *Location of the piezometers used for calibration*

Table 1: Sensitivity analysis simulations

Target parameter	Values to be tested	Fixed parameters
In-transfer rate (φ_{in})	From $\varphi_{in}=10^{-6}$ s^{-1} to $\varphi_{in}=10^{-4}$ s^{-1}	$\varphi_{out}=10^{-4}$ s^{-1}
Out-transfer rate (φ_{out})	From $\varphi_{out}=10^{-5}$ s^{-1} to $\varphi_{out}=10^{-3}$ s^{-1}	$\varphi_{in}=10^{-5}$ s^{-1}

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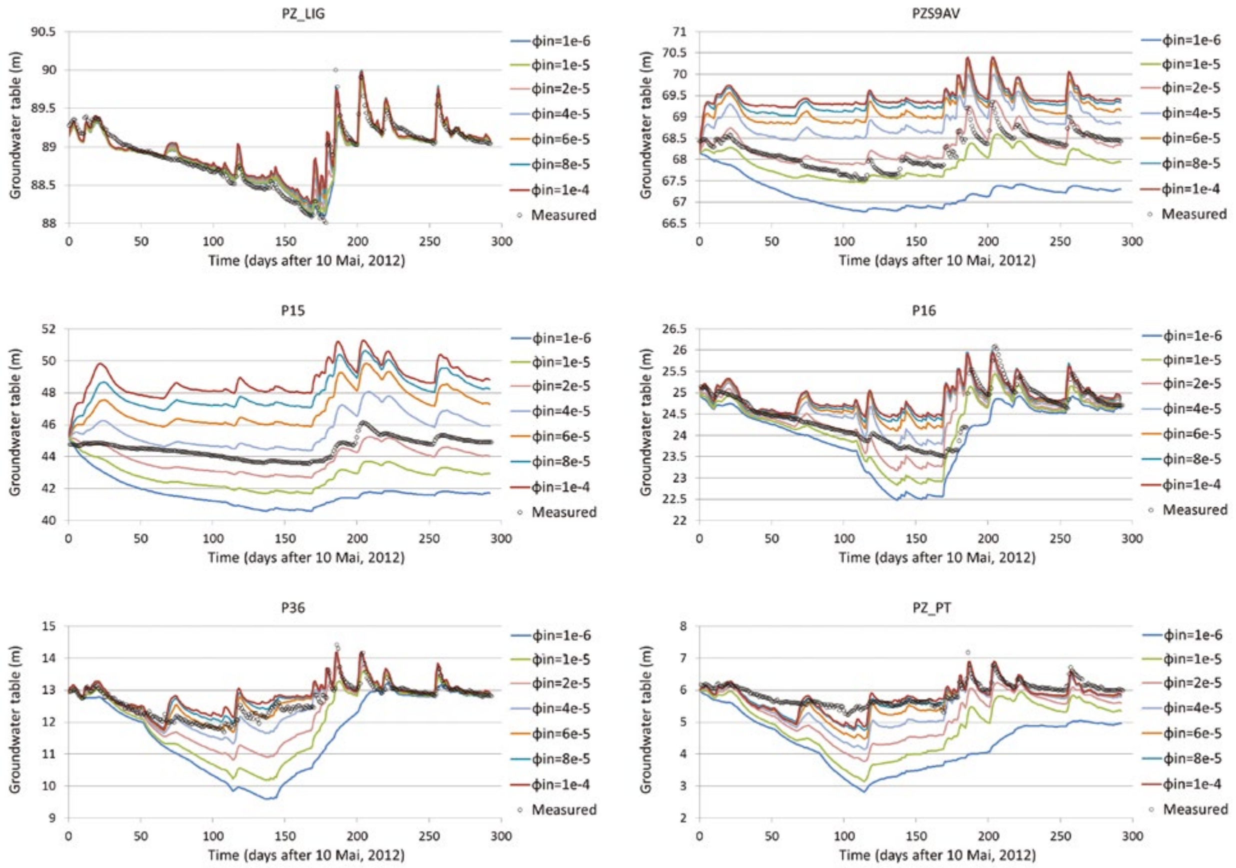

Figure 7: *Simulated results with different in-transfer rate values*

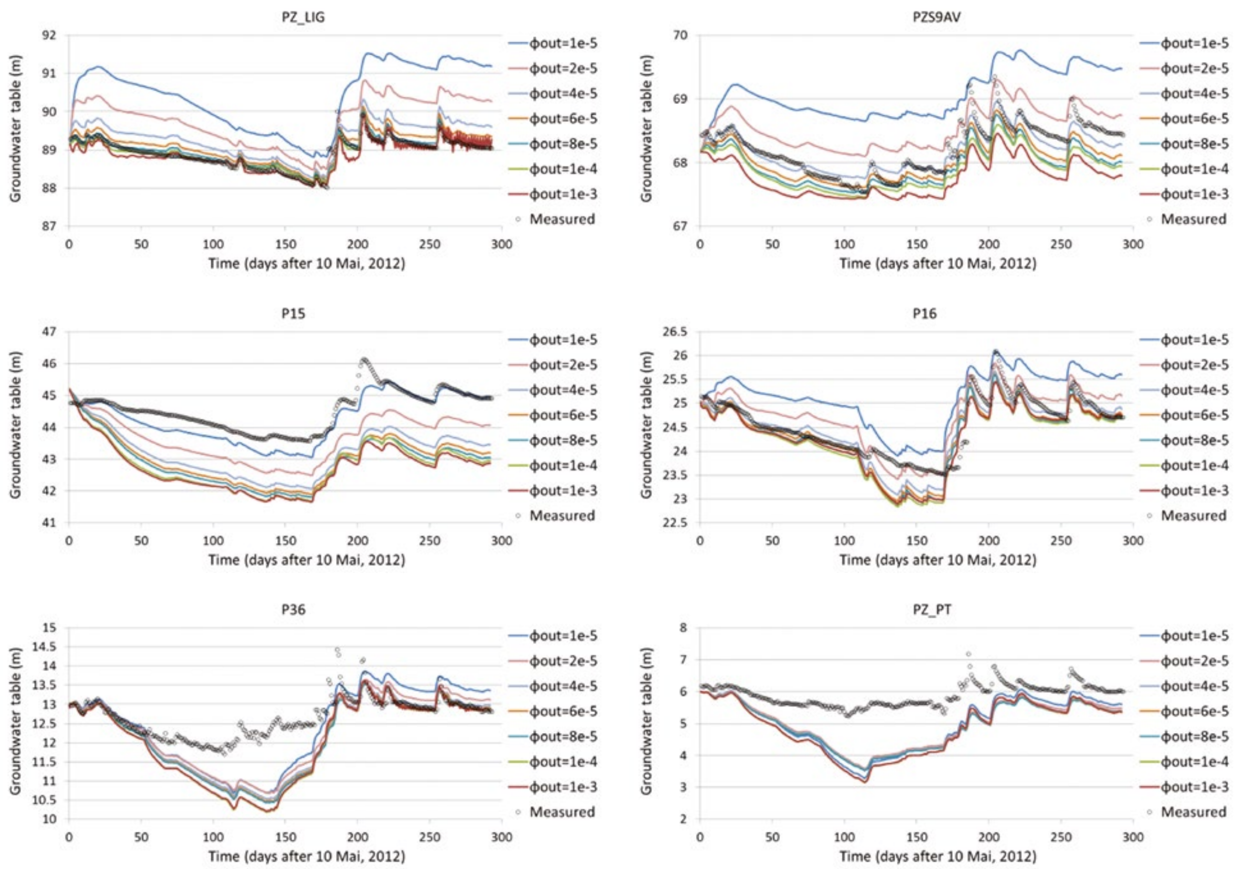

Figure 8: *Simulated results with different out-transfer rate values.*

Similarly, the Fig.8 shows that the upstream section is more sensitive to the variation of the out-transfer rate. It means that the feeding direction is from aquifer to river. The result of the point P15 can be explained by an associated impact of the increase or decrease of groundwater table in upstream area. As for the downstream section after P36, less variation can be observed since the groundwater table is mainly affected by the infiltration.

These tests make it possible to establish a calibration of the in/out-transfer rate based on the result of the sensitivity analysis. Based on the results of these simulations, the range of the in-transfer rate is from 2×10^{-5} to 4×10^{-5} s^{-1} to the north of the P36 piezometer, and this value increases up to around 1×10^{-4} s^{-1} near the river mouth. The out-transfer rate from the northern junction to the section of weirs is between 1×10^{-4} s^{-1} to 1×10^{-3} s^{-1}. It decreases down to 2×10^{-5} s^{-1} near the P16 point.

IV. RESULTS AND DISCUSSION

The calibrated parameters are shown in the Fig.9. Since the transfer rate can be treated as the index of the river-aquifer exchange activity, according to the calibration result, it is inferred that the exchange is more dynamic in the upstream area and downstream area, where no weir has been constructed. The section near the weir No.10 and No.9 have also a higher transfer rate.

The simulated values of groundwater table and the observed ones are compared in the Fig.10. In general, the results match the observed data no matter for the drought season or the rainfall season. The model can correctly

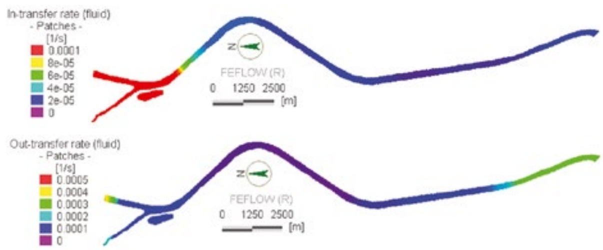

Figure 9: *Calibrated in/out-transfer rate*

represent the trend of groundwater table variation. Given that the peak values are directly related to the flood event in the river, the model is capable to represent the river-aquifer exchange. As for the low level part, the model result has a discrepancy within 0.3 meters except for the PZ_PT monitoring piezometer, where the error of low-level period reaches up to 0.5 m. Nevertheless, knowing the depth of the groundwater table in this area is around 7 m, such a difference makes only a relative error less than 7%. Hence the result is also acceptable.

The Nash–Sutcliffe Efficiency coefficient (NSE) [Nash and Sutcliffe, 1970] is applied to evaluate the model. The Tab.2 shows the result of the evaluation at each monitoring piezometer. For the PZ_LIG, PZS9AV, P16 and P36, the NSE can all exceed 0.75. Considering the location of the monitoring piezometers, it indicates that the accuracy of the result is related to the weir. The model gives a better result in the area where there is no weir (PZ_LIG and P36) or where there was a weir (PZS9AV and P16).

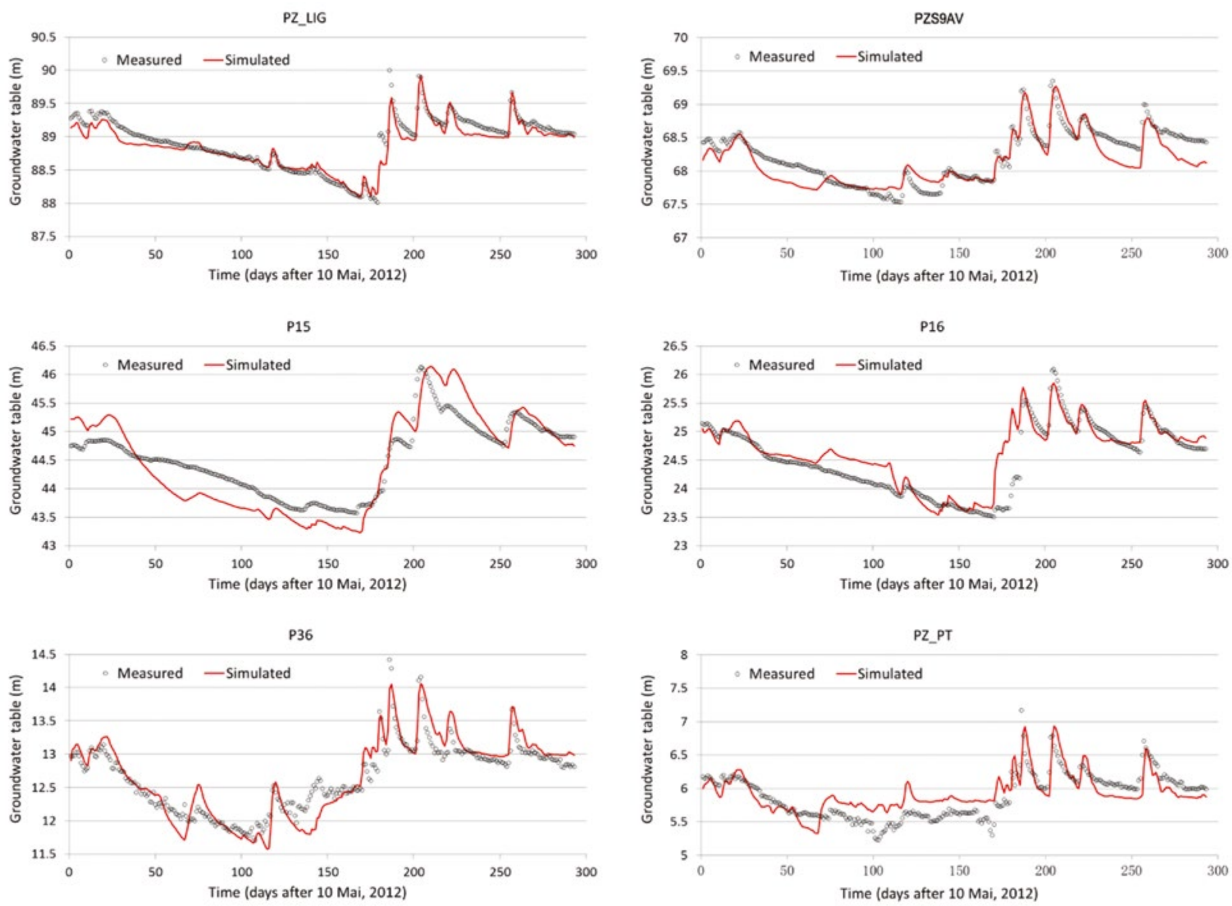

Figure 10: *Calibrated result of the hydraulic model.*

Table 2: Model results evaluation by Nash coefficient

Piezometer	Location characteristics	Principal land use	NSE
PZ_LIG	Upstream sections, without weirs	Industrial zone	0.91
PZS9AV	Upstream sections, weir No.9	Industrial zone	0.75
P15	Midstream sections, weir No.6	Agricultural land	0.67
P16	Midstream sections, site of former weir No.2	Agricultural land	0.75
P36	Downstream sections, without weir	Urban area	0.83
PZ_PT	Downstream sections, weir No.1	Urban area	0.56

The Fig.11 shows a longitudinal profile of the riverbed as well as the groundwater table. It can be observed that the hydraulic gradient is higher in the upstream area (before weir No.4) than in the downstream area. This corresponds also to the slope of the riverbed. It is apparent that the weirs have a negative effect on the groundwater resource conservation. The groundwater table withdrawal is more severe in the weir sections. The restoration appears near the lowered weirs (No.10 and No.9) and the destroyed weirs (No.3 and No.2), because the river-aquifer exchange becomes stronger since the natural river profile has been regained. This explanation matches the long term observation of the sediment and groundwater level near the weir No.2 (Fig.12). The weirs have been built since 1970s, from then, the fine sediments has been accumulated on the river bed and blocked the river-aquifer exchange. The photography taken on 1983 shows that dense vegetation has occupied the river banks, thus the groundwater level kept declining until 1994, when an extreme flood event happened and destroyed weir No. 3 and No. 2 [Guinot and Gourbesville, 2003]. The photography taken after 1995 shows that the river morphology was restored to its natural state. The vegetation on the river bed has been moved away and the river-aquifer exchange is also revived. Hence the groundwater level has increase by 2 m.

Through the difference between the flood season profile and the drought season profile of the groundwater table (Fig.11), it is observed that the river-aquifer exchange is more dynamic in the downstream part. For the weir sections, the aquifer is more stable regarding the season due to the thick clogging layer formed by the fine sediment deposition.

The existence of the weirs has a direct influence on the model output accuracy. It is caused by the infrastructure of

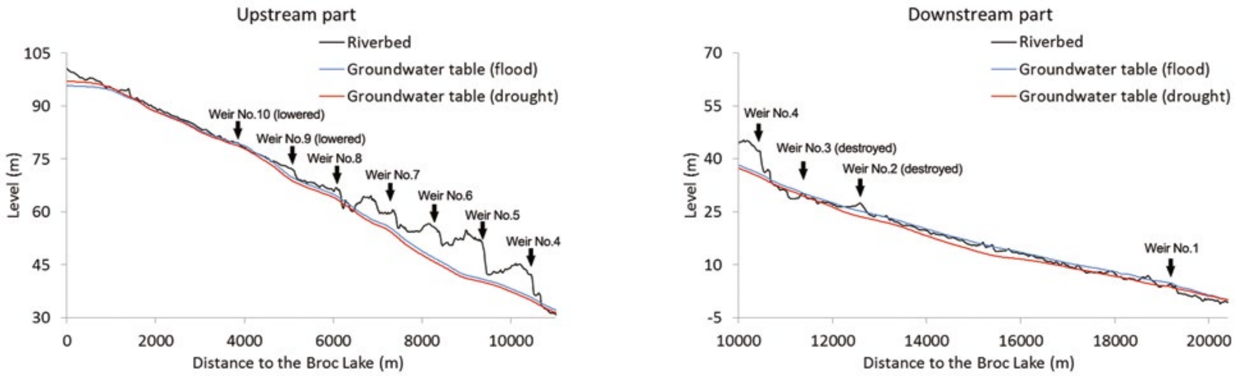

Figure 11: *Longitudinal profile of riverbed and groundwater table*

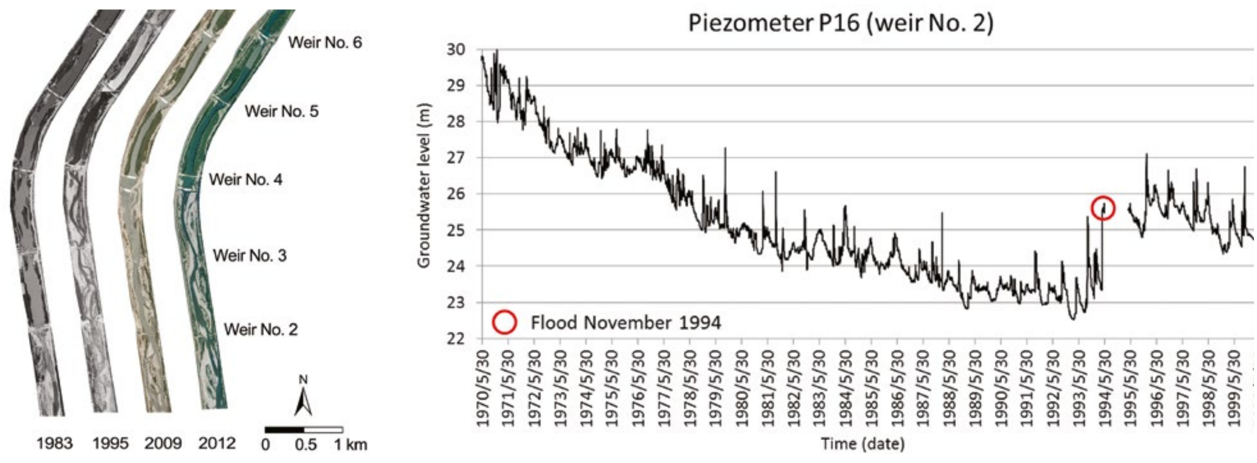

Figure 12: *Long term observation of the river morphology and the groundwater level near the weir No. 2*

the weir, which is a row of piles whose depth is at least 12 m beneath the riverbed. In reality, the infrastructure must impact the local hydraulic conductivity. However, this particular change of the soil hydraulic conductivity is not considered in the model, because the width of the piles is usually much smaller than a computational grid in the model.

On the other hand, the uncertainty of the model input data, such as river water level and the geological layers are as well an important factor of the inaccuracy. As the water level is calculated by a 1D model within the river, even though well calibrated and validated in previous studies, it still cannot be as accurate as the real observed data. Furthermore, the river width during the drought season and rainfall season are different, which leads to more infiltration area on the riverbed. The model is unable to take it into account, because the boundary condition is applied on the fixed cells. The geological layers data used to build this model are the geological profiles drew by geologists. Only 20 cross profiles are given for the recent alluvium [Guglielmi, 1993] and 8 cross profiles for the extended area (Fig.2(a)). Each profile is produced with the result of several drilling tests. There is no doubt that these conclusions are made by hypothesis and it is just a qualitatively correct result. No matter which method of interpolation is used to create the geological layers in a model, the uncertainty always exists. The error of the thickness of the alluvium estimated by the geologists can reach up to 10 m or even more. This error would be ignored on a geological map but it could cause a huge inaccuracy in a numerical model output, especially near the border of the alluvium, where the calculation of the hydraulic gradient is much affected by the thickness of the alluvium (eg. P15 and PZ_PT).

V. CONCLUSIONS

A 3D groundwater hydraulic model has been established for the study case of lower valley of Var river. The rainfall-evapotranspiration recharge, river-aquifer exchange and the water pumping have been considered as the predominant source/sink terms of the model. A sensitivity analysis of the in/out-transfer rate has been carried out in order to estimate the reasonable range of these two parameters. The Nash coefficients of the calibrated results are between 0.56 and 0.91, which indicates that the model is efficient as a predictive model.

The groundwater table withdrawal is more severe in the river section where the weirs are built. This conclusion is in accordance with that inferred from the long term observation at weir No. 2. One possible explanation is that the fine sediment deposition increases the thickness of the clogging layers, so the aquifer can hardly fed by the river. It proves that the construction of the weirs has a negative impact against the groundwater conservation.

VI. ACKNOWLEDGMENTS

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