Computational Methods in Water Resources XIII, Bentley et al. (eds) © 2000 Balkema, Rotterdam, ISBN 90 5809 123 6

Coupling of parallel river and groundwater models to simulate dynamic groundwater boundary conditions

G.Carabin & A.Dassargues
L.G.I.H., University of Liège, Belgium

ABSTRACT: A method is proposed to simulate, with physical consistency, interactions between groundwater and a river when variations of water levels and solute concentrations are computed in both groundwater and river models. It is based on a parallel run of the Groundwater Model (GM) and the River Model (RM). A junction manages the data exchange between the models using the PVM library. Tests on a real case study, show the importance of these interactions on the computed groundwater quantity and quality. The junction river-aquifer is being used in the scope of a research project entitled 'Integrated modeling of the hydrological cycle in relation to global climate changes', and it has already been used for the SALMON project (Sea Air Land Modeling Operational Network).

1 INTRODUCTION

For groundwater modeling, interactions with rivers are usually taken into account with direct or indirect (Dirichlet) prescribed external input/output boundary fluxes. This is often considered as convenient in most of the studied cases in steady state conditions or when the exchanges can be estimated. However, an accurate assessment of these exchanges is certainly not easy in many cases : (1) when the exchanged fluxes are sensitive to the river level fluctuations ; (2) when the groundwater level fluctuations are such that the direction of the water exchanges between the river and the aquifer can be often reversed ; (3) when previsional simulations are concerned. In order to assess accurately the water exchanges between the river and the aquifer, coupling of parallel river and groundwater models can be chosen as a solution. For each model, it allows to simulate dynamic boundary conditions using results of the other model.

The coupling of parallel models has been developed in the scope of the SALMON project (Sea Air Land Modelling Operational Network), an IBM environmental research program supported by the IBM International Foundation. Three existing models (ocean, river and groundwater) are connected adequately in order to form one single model. This model should be able to simulate physically all water and contaminant fluxes in a whole system at regional scale including marine, river, groundwater and atmospheric inputs. In practice, the connection of the different models is done through a specific interface, called « Junction ». As each model has its own time and space discretization, the junction must organize the data exchanges, including various time and space interpolations schemes (Dassargues et al. 1996). The PVM (Parallel Virtual Machine) software is used for the data exchange management between the tasks running on the different processors of a IBM SP2 parallel computer.

This concept of coupling parallel models is now used for developing an integrated model of the hydrological cycle (soil, river and groundwater models) and simulation of the impact of climate changes in different basins. These basins have been chosen for their different hydrogeological characteristics (quartzite-phyllite, sand, chalk or limestone).

In this paper, only the interactions with the river model are considered. The reader interested by groundwater modeling with ocean interaction can be referred to another paper (Carabin & Dassargues, 1999). The junction between the River Model (RM) and the Groundwater Model (GM) is presented with time and space interpolations schemes and the organization of data exchanges. Different applications have been carried out on a part of the Scheldt basin and the results are shown to prove importance of using dynamic boundary conditions in the groundwater model.

2 INTERACTIONS WITH THE RIVER

2.1 *Theoretical background in groundwater modeling*

The usual river boundary conditions have been presented in a previous paper (Carabin et al. 1998). Two classical ways exist to represent the interaction with a river :

1) a Dirichlet boundary condition (boundary condition of the first kind) ; the total head values (H_g) are prescribed equal to the elevation of the water level in the river (H_r) :

$$H_g = H_r \tag{1}$$

2) a Fourier boundary condition (boundary condition of the third kind) ; the exchanged fluxes (q) between the river and the aquifer are computed depending on the difference existing between the total head in the aquifer and the water level in the river :

$$q = \frac{K_r}{e_r}(H_g - H_r) \tag{2}$$

where K_r and e_r are respectively the hydraulic conductivity [L/T] and the thickness [L] of the river bottom.

The choice of the boundary condition is usually guided by the characteristics of the contact existing between the aquifer and the river.

The Fourier condition is adopted because of its general applicability. It can be used on the whole length of the water courses (McDonald & Harbaugh 1988, Nawalany 1994) even if an actual low permeable layer is not present in some places of the interface with the aquifer. Moreover the nature of this last boundary condition allows to simulate the situations where the water level in the aquifer can fall below the river bottom (Fig. 1). In this case, H_g is automatically replaced by the level of the river bottom (b_r) in equation (2) in order to calculate the seepage from the stream. The ability to simulate such situations is needed because they happen often along some courses in the basins chosen to study the impact of the climatic changes on the hydrological cycle.

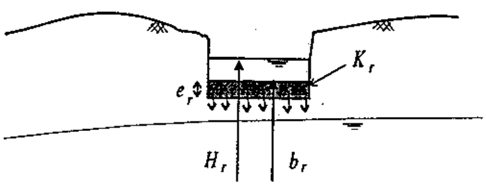

Figure 1. Schema of a losing stream which cannot be simulated with a Dirichlet condition but only with a Fourier condition.

2.2 *The junction between the river and the groundwater models*

The junction is implemented to manage the data exchange between the models running in parallel and to ensure a good simulated mass conservation.

2.2.1 *Computation of the exchanged fluxes*

The exchanged fluxes are computed in a common way in both the groundwater and the river models. For the water fluxes, the Fourier boundary condition (equation 2) is used. For any dissolved substance (C_i), the associated mass adective flux q_{ci} is then simply given by the product of the water flux and the appropriate concentration (if adsorption and degradation processes and chemical reactions are neglected in the streambed):

$$q_{ci} = \begin{cases} qC_{i,g} & \text{if } q > 0 \\ qC_{i,r} & \text{if } q < 0 \end{cases} \tag{3}$$

where $C_{i,g}$ and $C_{i,r}$ are, respectively, the concentrations in the groundwater and in the river.

2.2.2 *Spatial interpolation scheme*

The two models have their own spatial discretization, conditioned by the length scales of their respective processes and topologies. Moreover the GM uses a tridimensional discretization allowing 1D or 2D interface with the rivers whereas the RM uses an unidimensional discretization. As a consequence, the correspondence between nodes from the two models representing the same river has to be determined. Each node of one model has to be associated with a series of nodes of the other model. A mapping algorithm has been implemented to carry out automatically this correspondence in preprocessing. The result of the mapping process is the following:

- for each 'river' node of the GM, the direct downstream node of the RM and the direct upstream node are found and the spatial interpolation coefficient is calculated
- for each node of the RM, the set of the upstream 'river' nodes of the GM located downstream the previous node of the RM (the gray area in Figure 2) is found.

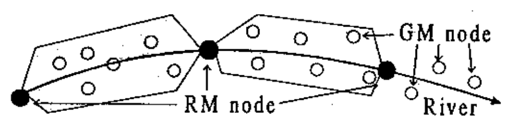

Figure 2. The spatial mapping between the different nodes of the river model (RM) and the groundwater model (GM).

2.2.3 *Time interpolation scheme*

The two models have also their own time scale and resolution time step. The river dynamic is very fast in comparison with the groundwater dynamic and consequently the RM time step is shorter. The GM time step is chosen sufficiently short to have a significant interaction between the models and equal to a multiple of the RM time step. So the interaction time step is the GM time step. No real time interpolation is necessary and it is avoided to ensure the mass conservation in the exchanges between the models.

2.2.4 *Organization of data exchanges*

The mass conservation between the models is ensured by the fact that the models receive identical fluxes computed by the junction at the beginning of every interaction time step and that these fluxes are constant during the time step. It means that an explicit time integration schema is used for the Fourier boundary condition in GM and RM. However, such a schema is conditionally stable with a very restrictive criterion for the GM. Too short time steps are needed to meet the stability criterion. In order to limit the computational cost, an implicit time integration schema must be used in the GM and consequently the presented scheme of data exchanges must be left. Some changes have been done and the used scheme is the following (Fig. 3):

- at the beginning of the interaction time step (equal to GM time step, see paragraph 2.2.3):
  1. The RM sends the river water height (h_r^r) and the river solute concentration (C_r^r) computed for each RM node.
  2. The junction adds the absolute level of the river bottom to the river water height and carries out the spatial interpolation to compute the river water level for each 'river' GM node (H_r^g).
  3. The GM receives the computed river water level and begins the non linear flow computation.
- at the end of the GM flow computation:
  1. The GM sends the water exchanged flux (q^g) for each 'river' GM node to the junction.
  2. The junction computes the solute mass flux for each 'river' GM node (q_c^g) using the equation (3) with the computed concentrations at the end of the previous time step. The sums of the water and the solute mass fluxes for the 'upstream' GM nodes are respectively the water and the solute mass fluxes (q^r and q_c^r) for each RM node.
  3. The junction sends the respective computed solute mass fluxes to GM and RM and it sends the water fluxes to the RM.

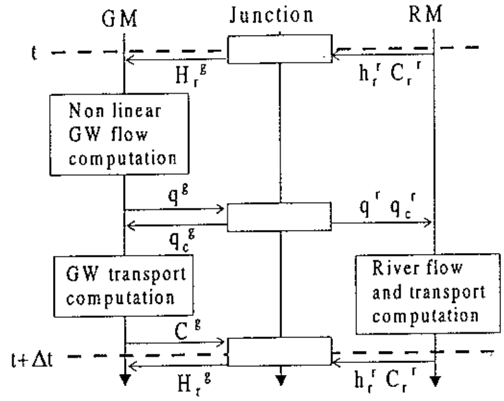

Figure 3. Scheme of the data exchanges between the river model (RM) and the groundwater model (GM)

- at the end of the parallel runs of the models, GM sends the computed concentrations (C^g) at the 'river' GM node to the junction.

This scheme is repeated for each interaction time step. In using a such scheme, only processes with a time scale longer than the interaction time step can be represented.

3 APPLICATION ON AN ACTUAL SITE

3.1 *General description of the site*

The SALMON model have been tested on an actual site in Belgium. The Scheldt basin has been chosen. It needs an integrated management of the water quality because of high industrial and demographic stress as in the main river basins in Europe. The selection has been done taking into account the availability of data and the experience of the research teams in the area.

The Scheldt river and its main tributary rivers is modeled by the river model, the North sea by the ocean model and a part of the Antwerp province by the groundwater model (Fig. 4). For the project time, it was impossible to model explicitly all the aquifers located in the 17,000 km² area of the Scheldt basin and only a 650 km² area is covered by the groundwater model.

The basic geology in the North of Belgium (Flanders) consists in a pile of Tertiary and Secondary sedimentary layers with tabular structures sloping northwards or northeastwards. For hydrogeology, this tabular structure sloping northwards induces a superposition of confined (or semi-confined) aquifers inter-layered with semi-pervious layers (aquitards). The catchment areas of these aquifers are fully connected with Quaternary layers, especially in the alluvial plains.

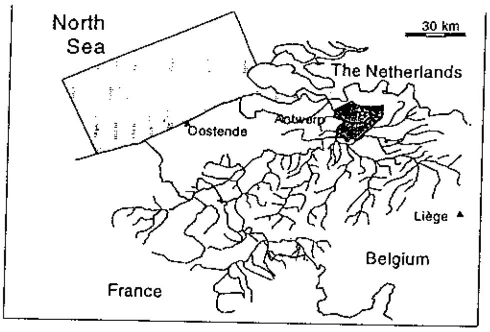

Figure 4. Application of the SALMON model: the Scheldt basin, the Belgian coastal zone and a part of the Antwerp province.

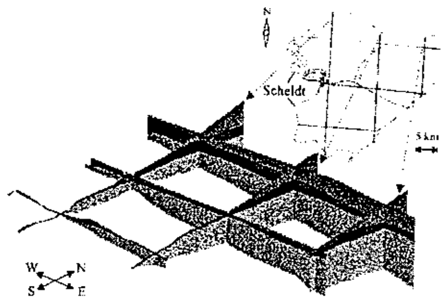

Figure 5. North-South and East-West cross-sections showing the three modeled layers in the groundwater domain.

3.2 *Description of the groundwater model*

The groundwater model domain is bounded by natural boundaries: the Scheldt river on the west, a groundwater divide and a river on the south, a river on the east, the groundwater divide on the north.

The geology of the domain consists in sands of Neogene and Quaternary ages located on the right bank of the Scheldt. These sand layers are overlying a thick clay layer (clay of Boom) which slopes northeastwards. Consequently the thickness of the aquifer increases to reach more than 200 meters at the Belgian northern border. The sands do not form an homogeneous sand formation and many layers can be defined with different lithologies and various values for hydrodynamic properties. All these local properties and thickness variations can not be taken into account in the modeling approach. A conceptual schema was chosen in order to simplify as adequately as possible, the representation of the actual hydrogeology in the concerned domain. One of the main hydrogeological features is the lithologic change of the Lillo formation which becomes more clayey in the North and consequently makes the underlying aquifer semi-confined. Thus, this formation has to be modeled.

To keep a limited number of elements, the domain has been discretized with only three layers (the sandy formations under the Lillo formation, the Lillo formation in itself and the formations overlying). In the southern part of the domain where the Lillo formation does not exist, only two layers are used to describe the Quaternary sands and the Tertiary sandy formations. Cross-sections on the figure 5 illustrate the chosen conceptual model.

The mesh is made of finite elements whose the length of which can vary from 50 to 250 meters and the thickness from 1 to 100 meters. The whole domain is discretized using 136211 finite elements (135921 wedge elements and 290 tetrahedral elements). The total number of nodes is 95415.

The calibration of the groundwater model is based on comparisons of the model results with measured data. The complexity of the modeled domain and the lack of direct measurements for the hydrodynamic parameters do not allow an automatic calibration procedure. A trial and error method is preferred here in order to be able to introduce all the needed 'soft data', i.e. interpretative geological data, to insure a reliable simulation of the actual groundwater system. 51 measured piezometric heads are used to calibrate the model. Results of the model running in steady state conditions are relatively good and the calculated total head are generally close to the measured mean annual values with an accuracy within 0.25 meter (the annual fluctuations is not more than 2 meters in all observed piezometers). As very few piezometric data are available in the southern part of the groundwater domain, it is difficult to evaluate the calibration accuracy in this zone. The figure 6 shows the computed piezometrics head in the bottom layer of the model.

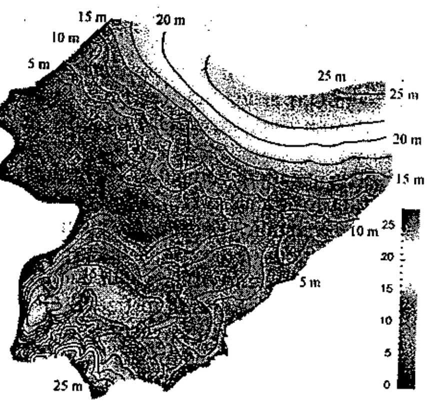

Figure 6. Computed piezometric map for the year 1992 in the bottom layer of the groundwater model

For each modeled layer, the lithologic changes in the sandy formations have been taken into account in the spatial distribution of hydraulic conductivity resulting from calibration. The hydraulic conductivity values vary from 3.8 10^{-5} to 2.2 10^{-4} m/s in the sandy formations but they decrease down to 10^{-7} m/s for the Lillo formation in the far North of the model.

3.3 *Tests results*

Several simulations have been carried out to validate the different components of SALMON and the coupling of the different sub-models. Only the results concerning the GM and the RM are discussed hereafter. Results from the ocean model can be found in another paper (Delhez & Carabin, 2000).

3.3.1 *Experiment 1 : Continuous pollution source in the river*

A constant release of 50 g/s of a conservative and a radioactive tracers was considered in the river, more precisely in the 'Grote Schijn' river which flows in the middle of the groundwater domain. The model ran for six months starting on April 29th 1993, i.e. with the actual meteorological and hydrological conditions relative to this period of time.

Such a scenario corresponds to the case of a continuous pollution by direct discharge of pollutants in the river through the sewage system or through outlets of some factory. The passive constituents is then transported to the sea through the river system and can possibly infiltrate into groundwater according to the hydrologic conditions.

On figure 7, the computed concentrations of the conservative tracer are shown at a given time (18/08/93) in the river. Classical results are obtained: a maximum concentration at the release point, and a downstream decrease due to the progressive dilution by lateral inputs of water. The concentration varies also in time because of the variable flow rate through the river. Indeed the total mass of tracer in the river decreases when the river flow rate increases and the tracer flux into the ocean model increases with the flow rate.

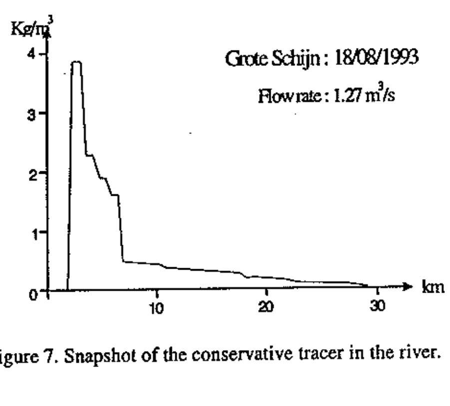

Figure 7. Snapshot of the conservative tracer in the river.

At the end of the simulation, nearly 98 % of the mass of the tracer released in the river have reached the sea. The 2 % left are still in the river, for the most part and only a small part can infiltrate into the groundwater.

Usually, the river drains the aquifer and, even during the summer when the water table is low, no significant mass transfer of the tracer can be observed into the groundwater domain. The very small mass fluxes effectively exchanged from the river into the aquifer during the short periods when the river level is higher than total heads are progressively compensated afterwards by a return flux back to the river, and consequently the computed tracer plume is very limited in the groundwater domain. In the simulation considered here, a pumping well has been considered near the contaminated river (at a distance of 240 meters) to examine the potential transfer of tracers from the river into the groundwater domain. It is observed that a pumping rate equal to 0.02 m³/s is enough to induce a nearly continuous mass flux from the river into the groundwater domain.

The computed tracer mass fluxes exchanged between the river and the groundwater domain near the pumping well are shown in figure 8. Positive values are taken for fluxes from the river into the groundwater domain. The simulated fluctuations of the river water levels and of the piezometric head in the pumping well are drawn too. It can be observed that the general trend of the computed exchanges is imposed by the slight variations of total head in the groundwater domain whereas the short and intense variations of the exchanged mass fluxes are mainly induced by the water level fluctuations in the river.

Figure 9 shows the computed time evolution of the conservative tracer mass which is extracted from the aquifer at the pumping well. The tracer mass fluxes exchanged in the zone influenced by the pumping well are shown again on this figure. In these results, one can observe the important following points :

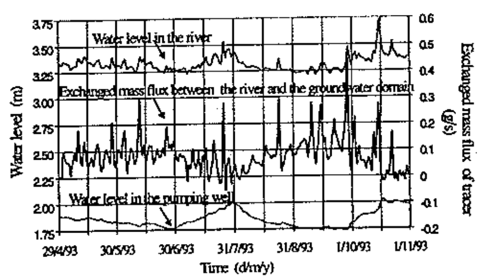

Figure 8. Simulated tracer mass fluxes exchanged between the river and the groundwater domain near a pumping well – water level fluctuations in the two models.

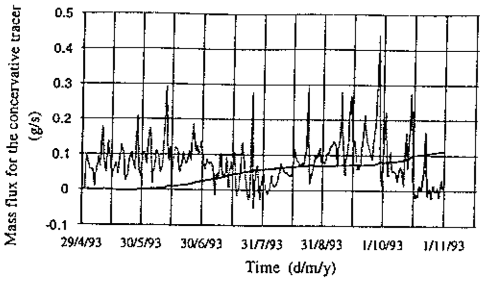

Figure 9. Temporal evolution of the mass flux of the conservative tracer exchanged with the river (light line) and extracted at the pumping well (dark line)

- a delay of about 40 days is necessary before the tracer is detected in the pumping well;
- a first stabilization of the rate of tracer mass extracted at the pumping well arises after 95 days;
- the second increase arises with a delay of about 40 days after the increasing of the exchanged mass from the river into the groundwater domain;
- the rate of the extracted mass of tracer does not increase higher than the value of 0.12 g/s that corresponds to 0.24 percent of the injected mass in the river. This value is the maximum of total mass flux of tracer which definitively goes out of the river into the aquifer in the pumping well influence zone.

From the injection point in the river to the pumping well in the aquifer, a total transfer time varying between 84 and 93 days is calculated using the ratio between concentrations of the degraded and not degraded tracers. A time of about 1.5 days is calculated for the tracer to run 16 km in the river. Then a mean velocity in the aquifer can be deduced as varying between 2.5 and 3 m/d. These values are in the range of those deduced from the classic application of the Darcy's law for a mean gradient of 0.6 %, hydraulic conductivity of about 10^{-4} m/s and effective porosity of 0.02.

It must be noted that this small value of the effective porosity explains the rapid transfer of the tracer for the considered period. This small value is not realistic for the studied sandy formations but it has been chosen to increase the dynamic of the groundwater model for the tests of the SALMON model. Indeed a run of the Ocean Model (OM) has a very long computational time by comparison with the others models (the time steps are equal to 18 seconds, 1 hour and 6 hours for respectively the OM, the RM and the GM) and, contrary to the GM, the transport results for a long simulation period (several dozens of years) become less credible for the OM.

The transport results may be transposed to other situations by simple dimensional arguments; for instance, the conclusions would remain unchanged if an actual effective porosity of 0.2 was used and the space scale and dispersivity values were divided by 10 (the pumping well would then be located at about 24 meters from the river).

3.3.2 *Experiment 2 : Temporary diffuse injection in the aquifer.*

Other kinds of pollution can have their origin far from the river. It was therefore decided to examine the case study of a pollution through the aquifer and the possible removal of this pollutant by the river.

In this case, the tracer is thus injected into the aquifer. Afterwards, it progressively reaches the river by transport processes in the groundwater. The passive tracer is released at the top of the aquifer on an area 500m x 280m during a period of 30 days. The total released mass flux of tracer is equal to 0.7 kg/s. The simulation of one year time starts on January 1 1993.

As mentioned previously, in the studied area, the river generally drains the aquifer and consequently the transfer of tracer from the aquifer into the nearest draining river occurs naturally. The temporal fluctuations of the water and tracer fluxes between the groundwater and the river are plotted on figure 10 for the part of the river (a length of 780 meter) where exchanges are the most significant. On the basis of these results, one can observe the following important points :

- on the studied section, the global exchanged water fluxes have always negative values confirming that in this area the river is draining the aquifer;
- the first arrival of the tracer in the river takes 100 days from the beginning of the diffuse injection;
- the exchanged tracer mass flux between the aquifer and the river is proportional to the exchanged water flux ;
- after one year, the simulated mass transfer is still increasing whereas the center of the plume (maximum of the concentration) has reached the river two months earlier. This is due to the strong increase of the infiltration during this period leading to the increase of the water exchange by a factor three.

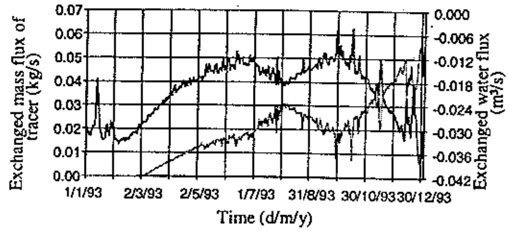

Figure 10. Temporal evolution of the flux of water (dark line) and of the tracer (light line) exchanged with the river along the section (about 720 m) draining the contaminated groundwater.

In this test, the injection lasts only one month and is located far from the river, i.e. 300 m to 700 m. Since the mean velocity in the aquifer varies from 1.75 m/d to 2.2 m/d, it can be considered that the tracer can reach the river during a period of about 250 days without taking into account any dispersion and diffusion effects. After one year of simulation, only three quarters of the total injected mass of the tracer have reached the river. In the reality, when a pollution is detected in the river, it is very difficult to know the exact location of the source, the extent and the intensity of this pollution.

It must be noted that, as in the first test, the effective porosity has a strong influence on the time scale of the interaction. The shown results obtained for an effective porosity of 0.02 can be extrapolated to the case of a larger porosity, e.g. 0.2, by considering appropriate adapted time scales as the transport would be slower (10 x) and, consequently, exchanges would be considerably longer.

4 CONCLUSIONS

The exposed method of coupling groundwater and river models allows to assess dynamically the exchanges between the river and the groundwater. The processes which can be taken into account depend on the chosen interaction time step between the models. A relatively short time step must be set in the groundwater model to obtain a significant influence from the river dynamic.

The tests carried out show that this dynamic boundary condition is very useful in case of transport modeling and above all when the exchanged fluxes are sensitive to the river level fluctuations.

The coupling between the models is now used for simulations in order to study the impact of climatic changes in some basins located in Belgium. This method is particularly recommended in such a study because the groundwater level fluctuations induce water exchanges in both directions between river and aquifer.

ACKNOWLEDGEMENT

Thanks are expressed to the IBM Foundation which supported the SALMON project and to the other teams participating to this project: GeoHydrodynamics and Environment Research (GHER), Centre d'Etudes et de Modélisation en Environnement (CEME). Thanks are also expressed to the Belgian Geological Survey (BGS), to the Flemish Administration for Environment, Water and Infrastructure (AMINAL) and to the Center for Nuclear Power in Mol (CEN/SCK) for their collaboration in the data collection needed to apply the groundwater model to the chosen domain.

Concepts about groundwater–river interactions are also developed within the research project 'Integrated modeling of the hydrological cycle in relation to global climate changes', supported by Prime Minister's Office – Federal Office for Scientific, Technical and Cultural Affairs of Belgium in the scope of the general program 'Global Change Sustainable Development'.

REFERENCES

Carabin G., Dassargues A., Brouyère S., 1998. 3D flow and transport groundwater modelling including river interactions. *Computational Methods in Water Resources XII*, Vol. 1, pp. 569-576

Carabin G. and Dassargues A., 1999. Modeling groundwater with ocean and river interaction. *Water Resources Research*, Vol 35, NO. 8, pp. 2347-2358.

Dassargues A., Brouyère S., Carabin G., and Schmitz F., 1996. Conceptual and computational challenges when coupling a groundwater model with ocean and river models. *Computational Methods in Water Resources XI*, Vol. 1, pp. 77-84

Delhez E. J. M. and Carabin G., 2000. Integrated modelling of the Belgian coastal zone, accepted in *Estuarine Coastal and Shelf Science.*

McDonald, M. G., and Harbaugh A. W., 1988. A modular three-dimensional finite-difference groundwater flow model, Techniques of Water-Resources Investigations of the United States Geological Survey, Vol. 6, chap A1.

Nawalany M., 1994. Combining the analytical and finite element models of the river-groundwater interaction, *Computational Methods in Water Resources X*, Vol 1, pp. 83-90.