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A quasi three dimensional model of water flow in the subsurface of Milano (Italy): the stationary flow

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Abstract

A quasi three-dimensional model is developed to simulate the behaviour of the aquifer system which is the resource of drinkable water for the town of Milano (Italy). Non continuous semipermeable layers locally separate permeable levels in a multilayered system, consisting of a phreatic and three confined aquifers. The numerical model is a conservative finite difference scheme based on the discretisation of the water balance equation for stationary flow. The grid spacing is 500 m and has been chosen, taking into account the distribution of the data in an area of about 400 km².

The model has been calibrated with a 'trial and error' procedure, by comparison of the results of the model with the observations for three years (1950, 1974 and 1982) which correspond to different flow situations. Once calibrated, the model has been used as a predictive tool, to forecast the behaviour of the aquifer system for other years of the 20th century; the comparison between the model forecasts and observations is good.

The model is capable of describing both the strong drawdown of the water table in the 1970s, when the water demand for domestic and industrial needs was very high, and the rise of the water table in the 1990s, when water extraction decreased. The results of the model confirm that the phreatic level is controlled largely by the local extraction of water; moreover, the aquifer system reacts to an increasing water demand with a small increase of the inflow and with a strong decrease of the outflow from its boundaries.

Introduction

The town of Milano (Italy) draws water for both domestic and industrial purposes from groundwater resources located beneath the urban area. The rate of water pumping from the aquifer system has varied throughout the 20th century (Fig. 1), depending upon the number of inhabitants and the development of industrial activities. This caused variations with time of the depth of the water table below the ground surface, as shown in Fig. 1, and in turn several emergencies: the two most prominent episodes occurred in the middle 1970s, when the water table in the city centre was about 30 m below the undisturbed natural conditions, and in the last decade, when the water table rose at a rate of approximately 1 m/year and caused flooding in deep constructions (garages and building foundations, underground railways, etc.).

To understand and quantify the factors affecting the variation of the phreatic head, a finite-difference flow model has been developed. Other models of this aquifer system were proposed at different scales. A model at the scale of a

single pumping station of the Acquedotto Comunale-Municipal Water Works (Ponzini *et al.*, 1989) is useful to evaluate the local effect of the activation of a pumping well in a limited area (0.25 km²) where a good data set is available. Giura *et al.* (1995) proposed a model at the regional scale for an area of more than 3500 km², bounded by the Prealpine hills to the North and by the Ticino, Po and Adda rivers on the other three sides. In this paper, groundwater flow is modelled at an intermediate scale for an area of about 400 km² that includes the whole surface area of the town of Milano.

After a description of the study area, together with information on the geological setting, the use of groundwater resources and the available data set the scale length of the model is determined; then the identification of the hydrogeological scheme of the aquifer system at the chosen scale length is described. Thereafter the basic equations of the numerical model are introduced, some computational problems are discussed and the model is calibrated: i.e., the boundary conditions are identified as are the source terms and the physical parameters used in the model. Then the

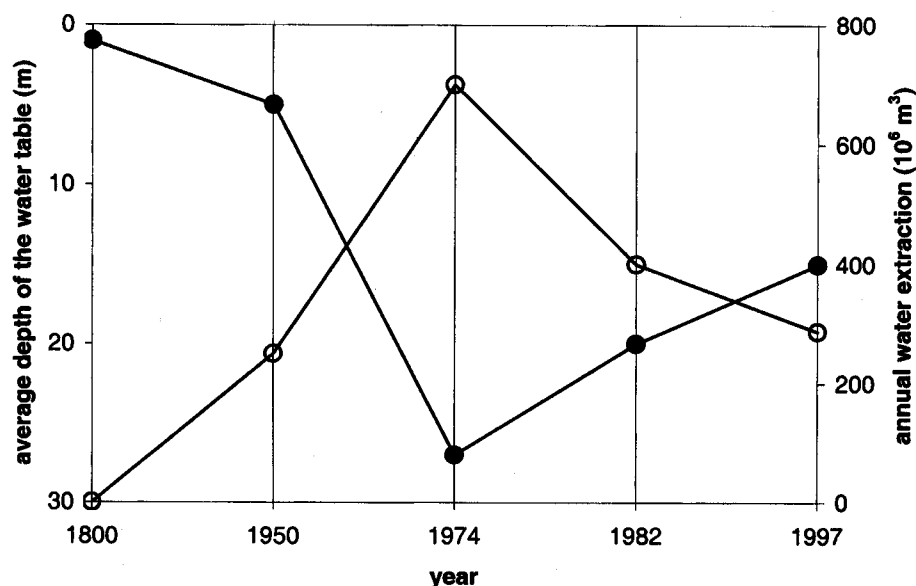


Fig. 1. Annual water extraction (open circles) and average depth of the water table below ground surface observed in the city centre (closed circles) for the 20th century.

predictions of the model are compared to observations, based on data that were not used in the calibration, so that this can be considered as a model validation. In addition, the main results of the model are shown, namely the reconstruction of the time variation of the water table depth during the 20th century, the correlation of this time variation with the volume of water extracted and the hydrological balance for the modelled area. The paper ends with a critical review of the results of the model and with some perspective on future work.

Site description

The area investigated (Fig. 2) is centred in the Comune di Milano—Municipality of Milano—and covers about 400 km^2 .

Water supply in this area comes almost entirely from groundwater. More than 600 wells for extracting drinkable water have been drilled by the Municipal Water Works since its establishment in 1888 (Motta, 1989). About 200 of these wells are not presently connected to the distribution network, because of the presence of contaminants; 70 of these wells are working as purge wells, i.e., they are continuously pumping water to remove pollutants from the aquifer system.

The wells are clustered among 34 pumping stations, each of which includes from 4 to 25 wells; water pumped from the wells is collected in pools located inside the main building of each pumping station and then injected into the distribution network.

More than 400 private wells have also been drilled; they

are more densely located in some industrial areas at the boundary of the town. Recently, most large factories were relocated far from the urban area so the water extraction from private wells decreased markedly from $335 \cdot 10^6 \text{ m}^3$ in 1967 to $25 \cdot 10^6 \text{ m}^3$ in 1997.

GEOLOGICAL FRAMEWORK

The site area, which is part of the Po river plain, is characterised by thick sedimentary deposits, which have been the subject of regional geological studies (Airoldi and Casati, 1989; Airoldi *et al.*, 1997; Barnaba, 1998). These studies are based mainly on lithological logs of drilled wells and show the following principal units starting from the ground surface.

1. Gravel-sandy unit, corresponding to recent alluvial and fluvio-glacial deposits (Olocene, upper Pleistocene); its thickness is approximately a few tens of metres and contains a phreatic aquifer.
2. Sandy-gravel unit, deposited in glacial and fluvio-glacial environments during the medium Pleistocene; the thickness of this unit varies between 50 and 90 m and consists of sandy-gravel layers separated by silt and clay lenses, which subdivide this unit into minor subunits from the hydrogeological point of view. In some places, it includes sandstones and cemented facies from the Lower Pleistocene. Units 1 and 2 constitute the 'traditional aquifer', which is the portion of the aquifer system which is most exploited for the water supply to the town of Milano.
3. Sandy-clay unit, deposited in a deltaic environment

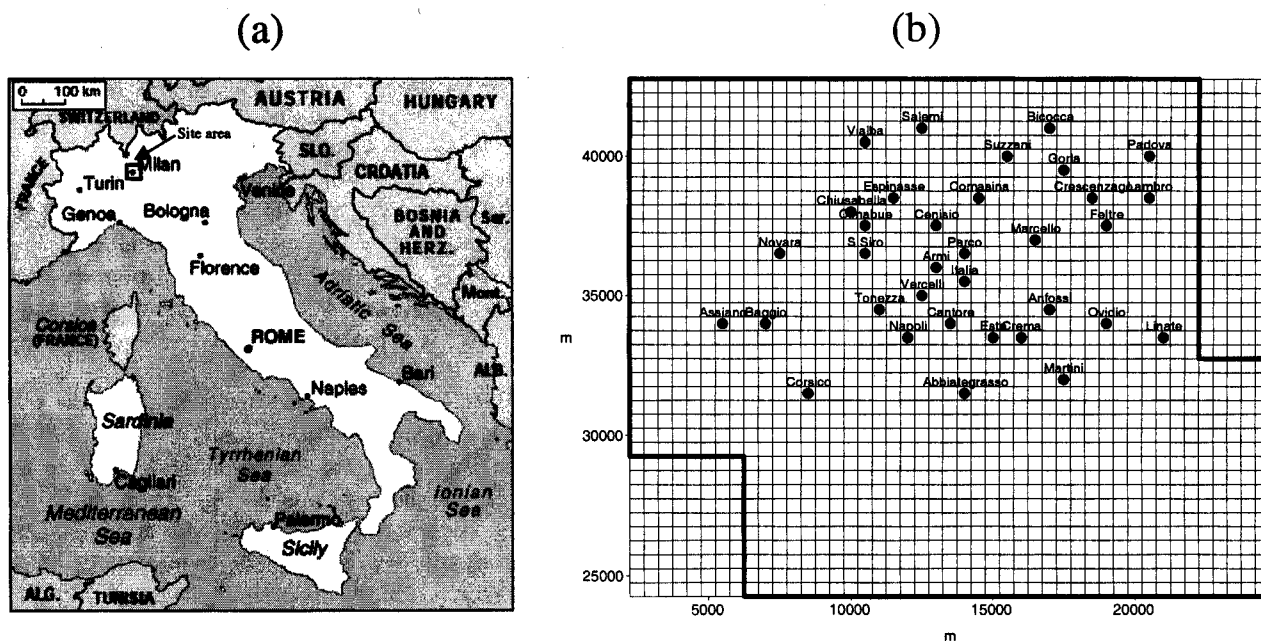


Fig. 2. (a) Location of the site area. (b) Numerical grid with the position and name of the pumping stations of the Municipal Water Works; the thick line shows the boundary of the modelled domain.

during the lower Pleistocene. This unit has a thickness greater than 100 m, is continuous throughout the whole area and, from the hydrogeological point of view, is generally impermeable; this unit is the 'impermeable base' of the 'traditional aquifer'. Some sandy levels of this unit constitute fresh water reservoirs that are partially used for water supply and can be considered as confined aquifers, because the silty covers are continuous. The conductivity of these deep aquifers is much lower than that of the traditional aquifer because of the presence of fine grained sediments.

4. Silty unit, deposited in a marine environment during the lower Pleistocene and Pliocene. The boundary between fresh and salt water lies within this unit.

AVAILABLE DATA

The data used for this study come from several sources: in particular, data published in the national literature (Martinis *et al.*, 1976; Maccagni *et al.*, 1979; Andreotti and Ponzini, 1983; Airolidi and Plos, 1983; SGA, 1981) have been integrated with unpublished data collected by public institutions.

These data can be grouped into three categories.

Cartographic and topographic data

The geological survey of the area is represented at the scale of 1:100000 in the Foglio 45—Sheet 45—(Milano) of the

Carta Geologica d'Italia—Geological Map of Italy (Servizio Geologico d'Italia, 1965).

To the topographic maps published by Istituto Geografico Militare—Military Geographic Institute (scale 1:25000) and by Regione Lombardia (scale 1:10000) the positions and levels of the pumping stations of the Municipal Water Works have been added.

Lithostratigraphic and well information

In addition to the information obtained from the geological literature which is useful in identifying the large scale geological structure, the collection of lithological logs of the public wells (Airolidi and Casati 1989) has been used.

Some wells were used for pumping tests (Martinis *et al.*, 1976; Maccagni *et al.*, 1979; Ponzini *et al.*, 1989).

Piezometric data and elements of the hydrological balance

The water levels, measured in some wells by the Municipal Water Works are referred to as piezometric heads, although it is not possible to attribute these values to specific depths because the screened intervals are positioned at different depths which do not take into account the layering of the hydrogeological system. Moreover these data were not collected simultaneously, but on different days of each month so that a statistical analysis could not be made. The piezometric heads used in the present work are those represented as contour plots for October 1950, September 1974 and December 1982 from Andreotti and Ponzini (1983); these data are representative of the situation of the piezometric head of the respective year. Similar data are

available for 1950, 1975 and 1989 from Airoidi and Casati (1989).

The depth of the water table and, therefore, the phreatic head, has been measured monthly in some shallow piezometers by the Sistema Informativo Falda (SIF)—Information Service on the Aquifer—of the Provincia di Milano. In the present study some of these data have been used for the period starting from January 1990. SIF has also monitored the phreatic and piezometric heads in other wells in the region surrounding the area considered in this work.

Total annual water extraction rates from public wells are known for the whole century; the monthly water extraction rates for each pumping station are available from 1986 onwards at the Municipal Water Works. Estimates of water extracted from private wells are based on annual surveys from 1979 onwards. Annual rates are available between 1979 and 1988; half-yearly rates for administrative zones of Milano are available from 1989 to the present. Estimates of the total amount of water extracted for the years 1950, 1974 and 1982 are given by Andreotti and Ponzini (1983).

The average annual rainfall is almost 1000 mm.

Hydrogeological scheme

The identification of the hydrogeological scheme depends strongly upon the goals of the model and in particular upon the scale length at which the groundwater flow is modelled; the model scale length (MSL) at which water flow is considered, is linked to the spacing of the grid for a finite difference model. The hydrogeological scheme and the MSL are also strictly linked to the choice of the flow model (2D, 3D or quasi-three dimensional).

CHOICE OF THE MODEL SCALE LENGTH

The choice of the MSL is of paramount practical importance for several reasons. The identification of the hydrogeological scheme must consider the hydrogeological structures relevant at the MSL.

The values of the physical parameters of the model equations must also be relevant at the MSL; the experimental data collected by field or laboratory experiments that involve flows at different scales cannot be used directly to obtain the values of the model parameters significant at the MSL.

Furthermore, the comparison between observations and results of the model must take into account that the model is reproducing flow at the MSL and therefore observations relevant at the same scale must be considered.

The main purpose of the present work is to simulate the evolution of the water table during the 20th century in the area of the town, neglecting fine scale features such as the local effect of a single pumping well. Therefore, the area modelled has been subdivided in 45×37 square cells of side

500 m (Fig. 2), which corresponds to the typical 'diameter' of a pumping station, i.e. to the maximum distance between two wells belonging to the same station. As a consequence, each cell contains at most one pumping station of the Municipal Water Works.

This choice implies that the heads computed with the model represent an average value over the cell; moreover, since the source term for each cell includes the total extraction rate of the wells of a pumping station, the model accounts for the influence of the whole station on the aquifer system, rather than the influence of the single pumping wells.

The choice of smaller cells, to permit the flow to be modelled at a finer scale and possibly with a fully 3D scheme, would require knowledge of the location of each well of the pumping stations and its extraction rate at different depths and the piezometric heads at many depths. Unfortunately, the hydrogeological scheme and the heterogeneity structure can not be identified at a finer scale because of uncertainty and location of the available data. These considerations suggest the choice of a quasi three-dimensional model rather than that of a fully 3D model.

THE HYDROGEOLOGICAL SCHEME OF THE AQUIFER SYSTEM

The hydrogeological structure of the aquifer system has been identified using the collection of lithological logs of the public wells (Airoidi and Casati, 1989), with the following steps.

1. Identification of the local hydrogeological scheme (LHS) for each pumping station of the Municipal Water Works. In this phase, it is necessary to assess the lateral continuity of less permeable units (silt and clay lenses) at the MSL; this means that lenses of linear dimension smaller than the MSL can be neglected, whereas those which are continuous over an area comparable to that of the cell are taken into account.
2. Correlation of the LHSs along 15 sections with geometric criteria only, i.e. neglecting the geological interpretation of the sedimentary structures. This correlation permits cross-checking of the LHSs proposed for the single pumping stations.
3. Interpolation of the LHSs over the whole modelled area.

An example of the resulting hydrogeological scheme is represented in Fig. 3 for a W-E oriented section. From the ground surface down: (a) phreatic aquifer (*ph*); (b) first aquitard (*s1*); (c) first confined aquifer (*c1*); (d) second aquitard (*s2*); (e) second confined aquifer (*c2*); (f) third aquitard (*s3*); (g) third confined aquifer (*c3*); (h) impermeable base, corresponding to the bottom of the so called 'traditional aquifer'.

In this work the traditional aquifer only has been considered and water exchanges between it and the deep

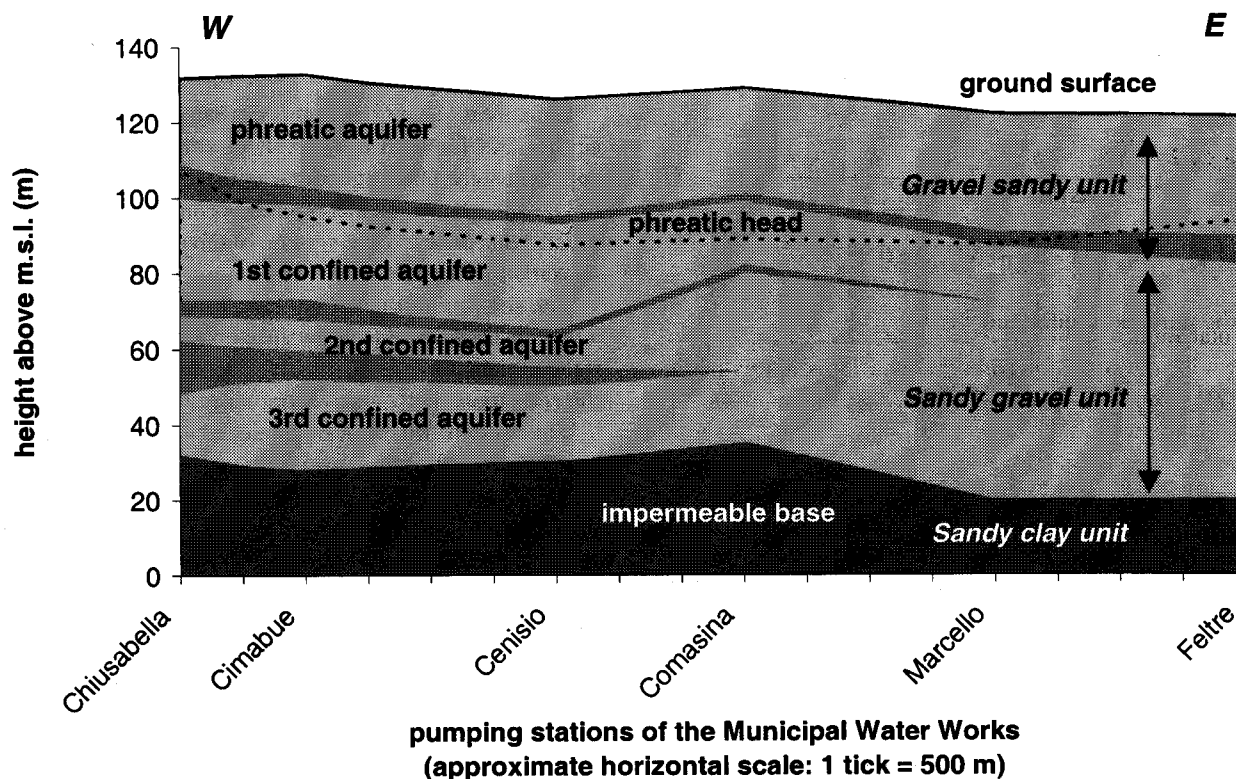


Fig. 3. Hydrogeological scheme for a section approximately oriented in the W-E direction. Areas in dark grey correspond to semipermeable layers (aquitards); areas in light grey correspond to permeable sediments (aquifers). The dashed line represents the observed level of the phreatic head in September 1974.

aquifers have been neglected because (a) the clay layer at the base of the traditional aquifer appears to be continuous and thick (tens of metres) in the modelled area and (b) tests conducted on deep wells show very small transmissivities for the deep aquifers.

As is apparent from Fig. 3, the aquitards, above all the second and third aquitards, are not continuous over the whole area; therefore, the distinction among the confined aquifers is correct only where the aquitards' extension is greater than the area of a cell of the model. However, the distinction among the aquifers is kept throughout the whole domain to take into account the variation of conductivity with depth; in fact, the compaction and cementation of the sediments and the presence of fine-grained materials increase with depth and cause a decrease of conductivity.

The mathematical model

The groundwater flow has been modelled as a succession of quasi-steady states. In fact, a rough estimate of the volume of water stored in (or released by) the aquifer system during one year is almost ten times smaller than the volume of water extracted by pumping wells, so that the storage term

of the time-varying water balance equation can be considered negligible in comparison with the source term due to water extraction.

Moreover, the classical assumptions of 2D horizontal flow in aquifers (with the Dupuit assumption for the phreatic aquifer) and 1D vertical flow through aquitards have been applied, so that the model is 'quasi three dimensional'.

The following notation has been used:

- lower indices i and j identify a node;
- upper indices $ph, c1, c2, c3$ represent the four aquifers;
- upper indices $s1, s2, s3$ represent the three aquitards;
- T and K represent internode transmissivity [$m^2 s^{-1}$] and conductivity [ms^{-1}], respectively;
- h represents the phreatic head [m];
- ϕ represents the piezometric head [m];
- σ represents the leakage coefficient [s^{-1}];
- N represents the source terms [$m^3 s^{-1}$];
- $\Delta x = \Delta y$ represent the spacing of the cell [m].

A conservative finite differences scheme, based on the mass conservation principle and the Darcy's law in the discrete form (Bear, 1979; de Marsily, 1986), is applied. In particular, for each cell of the area, the following balance

equations can be written:

$$\begin{aligned} & \Delta y \left[T_{i-1/2,j}^{ph} \frac{h_{i-1,j} - h_{i,j}}{\Delta x} + T_{i+1/2,j}^{ph} \frac{h_{i+1,j} - h_{i,j}}{\Delta x} \right] + \\ & \Delta x \left[T_{i,j-1/2}^{ph} \frac{h_{i,j-1} - h_{i,j}}{\Delta y} + T_{i,j+1/2}^{ph} \frac{h_{i,j+1} - h_{i,j}}{\Delta y} \right] + \\ & \sigma_{i,j}^{s1} (\phi_{i,j}^{c1} - h_{i,j}) \Delta x \cdot \Delta y + N_{i,j}^{ph} = 0 \end{aligned} \quad (1)$$

for the phreatic aquifer and

$$\begin{aligned} & \Delta y \left[T_{i-1/2,j}^{c1} \frac{\phi_{i-1,j}^{c1} - \phi_{i,j}^{c1}}{\Delta x} + T_{i+1/2,j}^{c1} \frac{\phi_{i+1,j}^{c1} - \phi_{i,j}^{c1}}{\Delta x} \right] + \\ & \Delta x \left[T_{i,j-1/2}^{c1} \frac{\phi_{i,j-1}^{c1} - \phi_{i,j}^{c1}}{\Delta y} + \right. \\ & \quad \left. T_{i,j+1/2}^{c1} \frac{\phi_{i,j+1}^{c1} - \phi_{i,j}^{c1}}{\Delta y} \right] + \\ & \sigma_{i,j}^{s1} (h_{i,j} - \phi_{i,j}^{c1}) \Delta x \cdot \Delta y + \sigma_{i,j}^{s2} (\phi_{i,j}^{c2} - \phi_{i,j}^{c1}) \Delta x \cdot \Delta y + \\ & \quad N_{i,j}^{c1} = 0 \end{aligned} \quad (2)$$

for the first confined aquifer

Equations similar to (2) can be written for the second and third confined aquifers. The coefficients $T_{i-1/2,j}$, $T_{i+1/2,j}$, etc. are called internode transmissivities and have been computed as the harmonic mean of the transmissivities of the two adjacent cells, e.g.:

$$T_{i-1/2,j} = 2T_{i-1,j}T_{i,j}(T_{i-1,j} + T_{i,j})^{-1}.$$

If the sediments were deposited in such a way that some preferential paths were created, e.g., buried river-beds, then the aquifer could be considered anisotropic in the horizontal plane at the MSL. However, the deposition of the sediments in the aquifer system studied is so complex that no preferential path is indicated at the MSL; therefore, each aquifer is assumed to be isotropic in the horizontal plane. On the other hand, the quasi three dimensional scheme accounts for the difference between horizontal and vertical conductivity at the MSL, which is created by the approximately layered structure of the 'traditional aquifer'.

As regards the physical meaning of the terms appearing in Eqns. (1), (2) and similar equations for the second and third confined aquifers, the first two lines of (1) represent the horizontal fluxes crossing the four lateral sides of the cell of the aquifer. The terms proportional to the leakage coefficients represent vertical water exchanges between aquifers and couple the equations for the different aquifers. The source terms correspond to the sum of the water extraction rates from the wells located within the cell; for the phreatic aquifer they also include recharge from the surface (rain infiltration; losses from channels, rivers and pipes).

The transmissivity of the phreatic aquifer depends on the

thickness of the aquifer itself and therefore on the phreatic head, so that (1) is a non linear equation with respect to the phreatic head.

Equations (1), (2) and the similar equations for each aquifer and for each cell, result in an algebraic system of approximately 6000 equations with the same number of unknowns when Dirichlet boundary conditions are assigned on all boundaries. This system is solved with an iterative relaxation procedure.

The solution of the system has posed several difficulties. The non-linearity of (1) can cause problems of convergence if the initialisation of the iterative procedure or the choice of the relaxation coefficient are not accurate.

The aquitard windows are modelled introducing values of conductivity for the aquitard of the same order of magnitude as for the connected aquifers and an arbitrary thickness of 1 m in the window cell; in these cells, this causes a perfect coupling of the connected aquifers, so that their heads almost coincide. These aquitard windows cause further problems for convergence of the iterative procedure.

Moreover in some years, when the extraction rates were very high, the water table was below the top of the first aquitard in some areas, so that the phreatic aquifer was actually dry and the first confined aquifer changed into a free-surface aquifer (see Fig. 3); the balance equations have to be adjusted to these physical conditions, e.g. transferring the recharge to the next aquifer.

To handle the above mentioned difficulties, a proprietary code has been developed for solving the system of equations, instead of using the commercial codes available. In particular, the code can modify the system of equations to be solved when the water table is below the top of the first aquitard. Moreover, the code controls the convergence of the iterative procedure to the solution of this highly non-linear system. In particular, the iterative procedure is stopped when the maximum difference between the values of the heads at two successive iterations is smaller than a fixed small tolerance (usually 10^{-5} m); moreover an *a posteriori* check is always performed to verify that the discrete balance equations are satisfied by the approximate solution.

The model calibration

The use of the numerical model as a predictive tool requires the determination of its parameters. This process is known as model calibration and is subdivided into three steps: boundary conditions, source terms and hydraulic conductivities and transmissivities.

BOUNDARY CONDITIONS

Dirichlet boundary conditions are assigned at the border of the domain.

In particular, the southern and most of the eastern and western borders of the domain correspond to a good approximation to the line of depression springs, where the phreatic head coincides with the ground level and is therefore known and constant in time.

On the other hand, the northern boundary corresponds to the line of depression springs observed during the geological survey conducted in 1963–1964, mapped in the Foglio 45 (Milano) of the Geological Map of Italy (Servizio Geologico d'Italia, 1965). Along this boundary, the phreatic heads correspond to the level of the springs for the first decades of the century only; in recent years, the drawdown of the phreatic head caused the disappearance of these depression springs. The phreatic heads on this northern boundary do not correspond to a physical limit of the aquifer system and have been assigned by interpolation of the data described earlier.

SOURCE TERMS

Source terms due to extraction from pumping wells are quite reliable for the extraction rates by the Municipal Water Works but are more uncertain for the extraction rates from private wells.

The recharge from the surface due to losses from buried pipes, rain infiltration and losses from natural and artificial channels has been estimated. In particular, the annual precipitation is assumed constant; since the area is largely covered with roads and buildings, it has been assumed that all the water is collected by the sewage system. Even water which has been used for domestic or industrial purposes, is largely collected by the sewage system. Therefore, the recharge of the aquifer from the surface has been estimated on the basis of the assumed percentage of water flowing into the sewage network with respect to the total extraction and the loss from the network.

The source terms corresponding to rain infiltration and to losses from channels, rivers and buried pipes are not easy to estimate; these values are very uncertain and their estimation is the subject of active research.

HYDRAULIC CONDUCTIVITIES AND TRANSMISSIVITIES

Each aquifer is assumed homogeneous, i.e. with a constant value of conductivity. However, transmissivity is space dependent because of the variations of the aquifer thickness which is determined by the geometrical characteristics of the hydrogeological scheme. The values of K_{ph} , K_{c1} , K_{c2} and K_{c3} , only have to be identified; these must be effective values at the MSL. The results of pumping tests or other field or laboratory measurements cannot be used in a straightforward way, because they usually yield effective values at scale lengths smaller than the MSL. Since the identification procedure must be consistent with the MSL,

it is based on the results of the model itself. The main steps of the procedure are sketched below.

1. The equivalent conductivity of the whole aquifer system, K_{eq} , is assumed to be

$$K_{eq} = (K_{ph} + K_{c1} + K_{c2} + K_{c3})/4 \quad (3)$$

2. As mentioned earlier the compaction and cementation of the sediments and the presence of fine-grained materials increase with depth so that it is assumed that conductivity decreases with depth. The vertical variation of conductivity has been parameterised with the following law:

$$K_{ph} : K_{c1} : K_{c2} : K_{c3} = 1 : s : s^2 : s^3 \quad (4)$$

- where $0 < s < 1$. The values of K_{ph} , K_{c1} , K_{c2} and K_{c3} can be computed in a unique way from K_{eq} and s with (3) and (4). Therefore, the aquifer system is parameterised so that two parameters only, K_{eq} and s , have to be identified.
3. The values of K_{eq} and s are determined by minimising the difference between the observed piezometric heads and the phreatic heads computed with the model for the years 1950, 1974 and 1982. This is done with a 'trial and error' procedure, driven by the analysis of a distinct objective function for each year considered. In particular, the following objective function is defined for each of the above years:

$$f(K_{eq}, s) = \frac{1}{N_{obs}} \sum |h^{(m)} - h^{(o)}|, \quad (5)$$

where: N_{obs} is the number of head observations available for the year; the sum is extended to the available observations; for each data location, $h^{(o)}$ and $h^{(m)}$ represent the observed and the modelled value of phreatic head respectively.

The low quality of the piezometric head data does not permit reliable estimates of the parameters with automatic inversion and the objective function (5) has been used as a guide for a subjective choice.

The objective functions for the three years behave slightly differently but they all show minima for values of K_{eq} very close to $2 \cdot 10^{-3} \text{ ms}^{-1}$; on the other hand, the value of s for which the minimum is obtained is not well defined because the model has been calibrated considering the phreatic heads only; the uncertainty in the position of the screened intervals of the wells prevents a measured piezometric head from being assigned to a definite depth and to a specific aquifer.

Finally, the following values have been assigned: $K_{eq} = 2 \cdot 10^{-3} \text{ ms}^{-1}$ and $s = 1/5$. They correspond to $K_{ph} = 6.4 \cdot 10^{-3} \text{ ms}^{-1}$, $K_{c1} = 1.3 \cdot 10^{-3} \text{ ms}^{-1}$, $K_{c2} = 2.6 \cdot 10^{-4} \text{ ms}^{-1}$ and $K_{c3} = 5.1 \cdot 10^{-5} \text{ ms}^{-1}$, with two orders of magnitude between the conductivities of the uppermost and lowermost layers.

4. The value of transmissivity at each cell is obtained by multiplying the corresponding conductivity by the thickness of the aquifer.

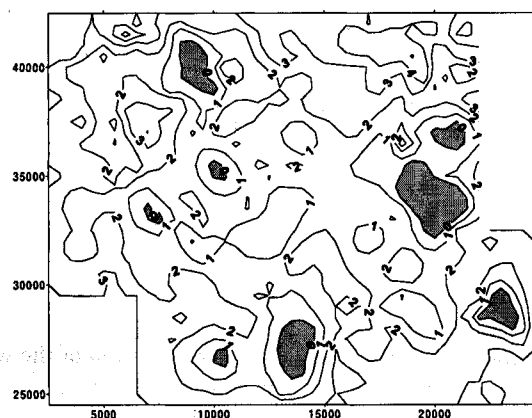


Fig. 4. Contour plot of the depth of the water table below the ground surface; equidistance 1 m. The grey areas correspond to the outcropping of the water table.

5. The hydraulic conductivity is assumed to be constant for the aquitards and equal to $1 \cdot 10^{-7} \text{ ms}^{-1}$, i.e. 50 times smaller than K_{c3} . A sensitivity analysis on these parameters has shown that the results of the model are not very sensitive to them, at least for steady state conditions;

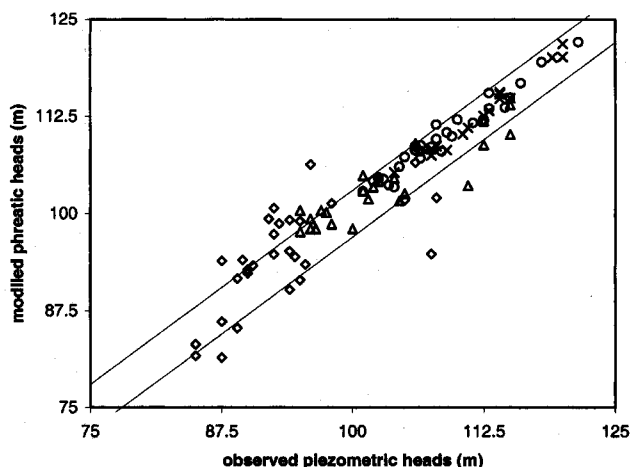


Fig. 5. Modelled phreatic heads versus observed piezometric heads for different years: crosses—1950; diamonds—1974; triangles—1982; circles—1997. The lines envelope the values with an absolute difference between modelled and observed heads less than 3 m.

in fact, the vertical hydraulic gradient is controlled

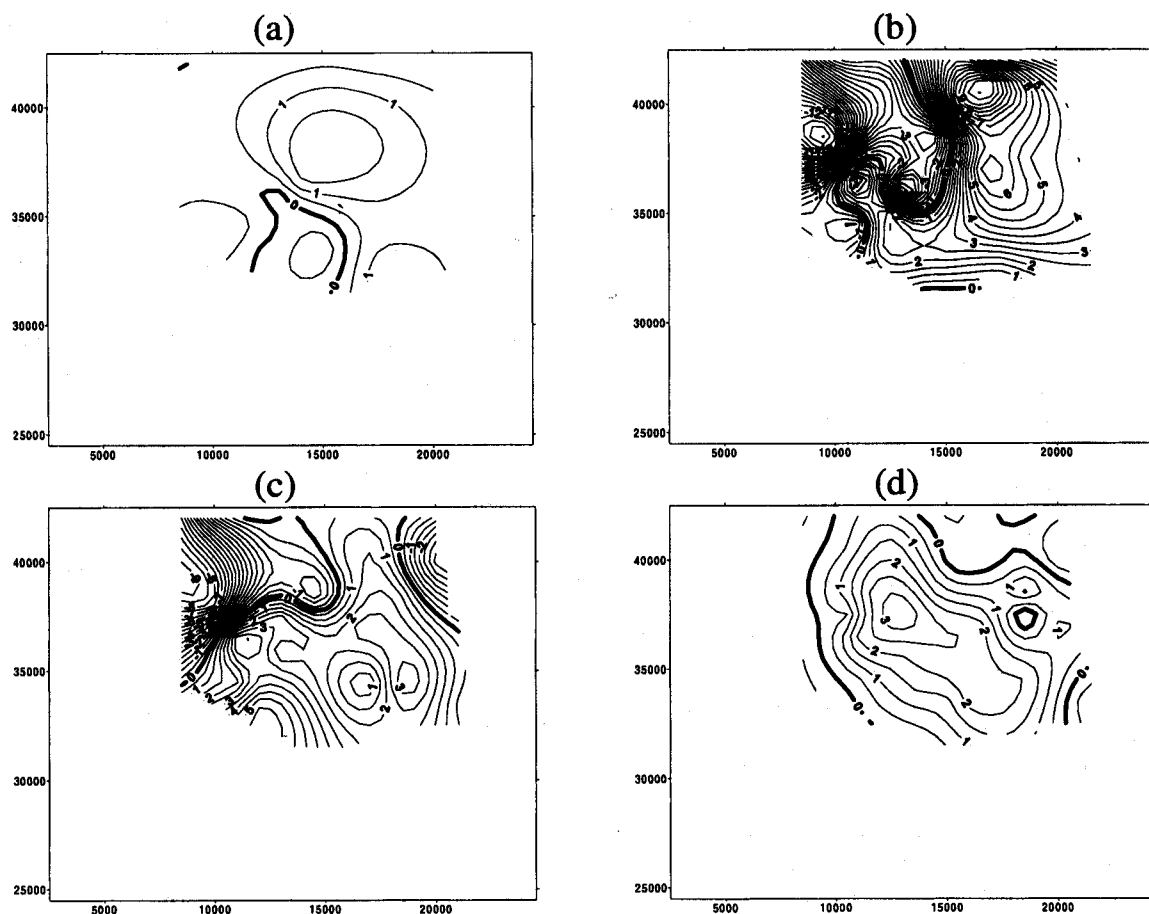


Fig. 6. Contour plots of the difference between modelled phreatic heads and observed piezometric heads for different years: (a) 1950; (b) 1974; (c) 1982; (d) 1997. Contour lines are only plotted in the region where data are available. Equidistance: 0.5 m.

mainly by the position and extension of the aquitard windows.

The authors have two remarks to make about the data used.

Firstly, the data (phreatic heads) used for the calibration have been sampled at the MSL, i.e. with a spacing that is of the same order as the side length of the cell; since heads are usually quite regular functions, these data are representative of the flow behaviour at the MSL.

Secondly, data relative to different years that correspond to very different flow conditions have been used simultaneously. In fact, 1950 corresponds to the beginning of the intensive exploitation of the aquifer system, 1974 corresponds to the maximum drawdown of the water table and 1982 corresponds to a first rise in the water table. The use of data from different flow conditions reduces uncertainty in model calibration (Scarascia and Ponzini, 1972); (Sagar *et al.*, 1975; Ponzini and Lozej, 1982; Carrera and Neuman, 1986a, b; Ginn *et al.*, 1990; Ginn and Cushman, 1992; Parravicini *et al.*, 1995; Giudici *et al.*, 1995; Snodgrass and Kitanidis, 1998). In the present case, since the minima of the objective functions are found at about the same values of K_{eq} for the three years, the values identified can be used to forecast different scenarios in a reliable way, if proper boundary conditions and source terms are assigned.

Model validation and results

MODEL VALIDATION BY COMPARISON OF OBSERVATION AND FORECAST

To validate the model, the code is run with the calibrated parameters to forecast the behaviour of the aquifer system for flow situations different from those used in the calibration.

This is similar to the usual practice when different future scenarios are planned and the model is used as a tool to forecast the behaviour of the aquifer under different hypotheses. Unfortunately, the model prevision can be checked only after some time, e.g. a few decades.

The present approach differs in the sense that scenarios that have occurred already are forecast so that the relative data have already been collected and the model can be validated by comparing 'forecast' with 'observation'. In fact, the situations corresponding to the years '1800' and 1997 are forecast. The year '1800' corresponds to the situation when very limited amounts of water are extracted and the aquifer system is under natural conditions, not perturbed by human activity. Historical data show that the water table was generally a few (1–3) metres below the ground surface and pools existed in slight depressions even inside the town. This situation is well reproduced by the results of the model, which yields the contour plot of the depth of the water table represented in Fig. 4.

Modelled phreatic heads are plotted against observed piezometric heads in Fig. 5 for all the modelled years. The

absolute values of the differences are less than 3 m except at some scattered points for the years 1974 and 1982. The contour lines of the difference between the modelled phreatic heads and the observed piezometric heads for different years are plotted in Fig. 6 and show the spatial distribution of the errors.

EVOLUTION OF THE GROUNDWATER FLOW DURING THE 20TH CENTURY

The model permits simulation of the evolution of the water table in the area of the town of Milano during the 20th century. The 'cartoon' is presented in Fig. 7.

Under undisturbed conditions, the water table followed the ground surface with some smoothing and with a NW–SE trend (Fig. 7a). When abundant water extraction began, the flow field was modified, as shown in Fig. 7b, which corresponds to year 1950, when $250 \cdot 10^6 \text{ m}^3$ of water were extracted from the aquifer system. In 1974 the total annual water extraction was about $700 \cdot 10^6 \text{ m}^3$; this caused an increase in the hydraulic gradient in the northern part of the town, a modification of the flow lines to produce an incoming flow from North–East, a decrease of the hydraulic gradient in the southern region and a strong drawdown in the city centre where the pumping stations are more dense. These effects gave rise to a 'depression cone' in the city centre where the phreatic aquifer was depleted and the first aquitard dried up; as a consequence, the city centre suffered from subsidence, which caused risk of damage to some monumental buildings. Later, the annual water extraction declined to $400 \cdot 10^6 \text{ m}^3$ in 1982 and $287 \cdot 10^6 \text{ m}^3$ in 1997. Hence, the aquifer system is now approaching the situation observed in the 1950s.

The model permits computation of the inflow and the outflow through the boundaries. The values of these flows are plotted in Fig. 8 as a function of the water extraction rate. This figure shows that the aquifer reacts to an increasing water demand mainly by reducing the outflow.

Under undisturbed conditions the outflow is estimated to be greater than $15 \text{ m}^3 \text{ s}^{-1}$, whereas it reduces to less than $4 \text{ m}^3 \text{ s}^{-1}$ when the annual volume of water extracted is equal to $700 \cdot 10^6 \text{ m}^3$.

On the other hand, Fig. 7 shows that an increase in water demand increases mainly the part of the boundary through which water enters and the hydraulic gradient there. Nevertheless, the inflow does not vary very much around a value of $16 \text{ m}^3 \text{ s}^{-1}$ for annual water extraction up to $400 \cdot 10^6 \text{ m}^3$, and it increases to $22 \text{ m}^3 \text{ s}^{-1}$ when the total annual water extraction is $700 \cdot 10^6 \text{ m}^3$.

Conclusions and perspectives

The choice of the MSL is important in all the phases of development and application of a numerical groundwater

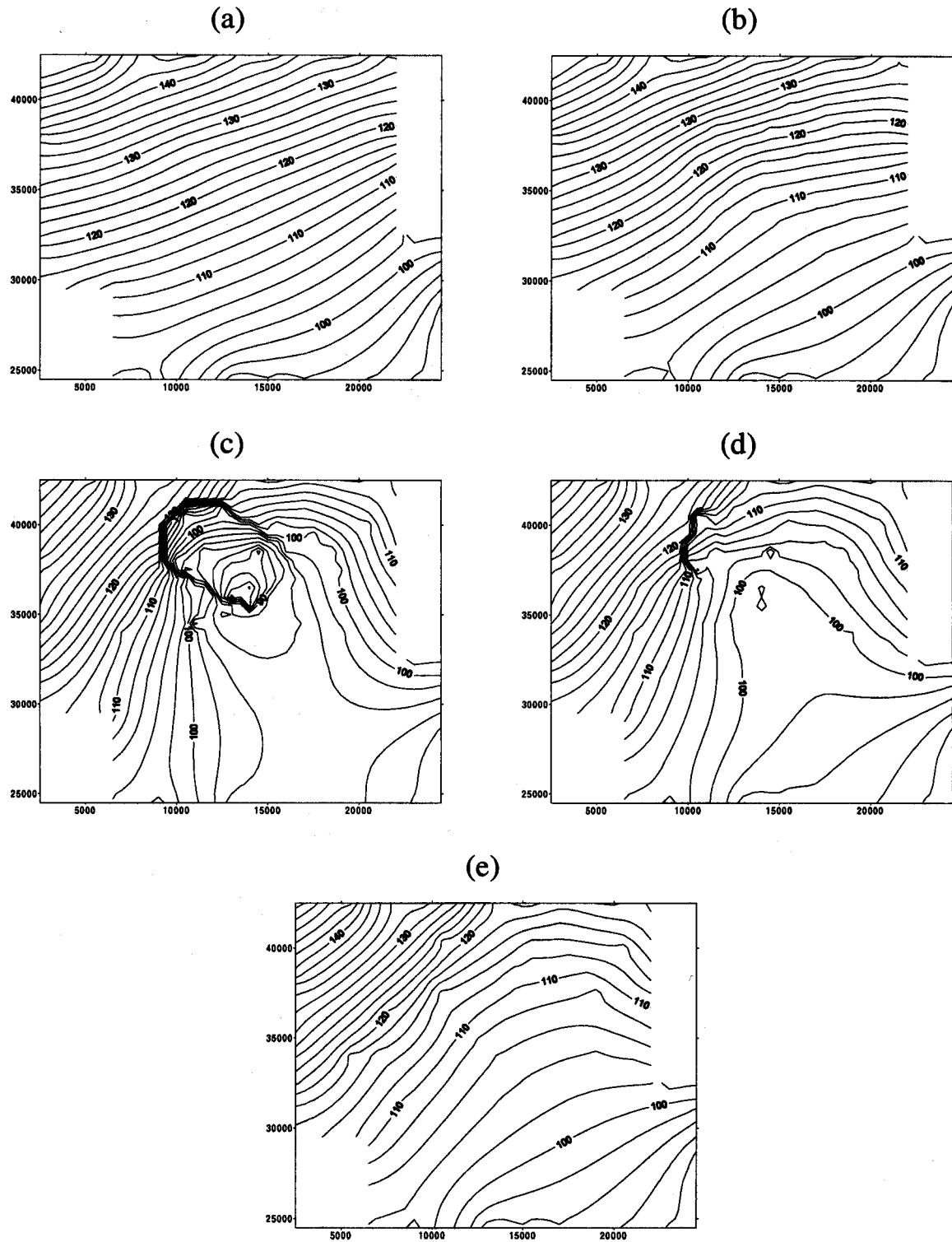


Fig. 7. Contour plots of the phreatic head computed with the numerical model for the following years: (a) "1800", undisturbed conditions with negligible water extraction; (b) 1950; (c) 1974; (d) 1982; (e) 1997. Equidistance: 2 m.

flow model; the chosen MSL is a key parameter that modellers have to take into account since it strongly affects the choice of the hydrogeological scheme, the analysis of the

data, the determination of the values of the physical parameters, etc.

In the double-phase calibration/validation procedure, the

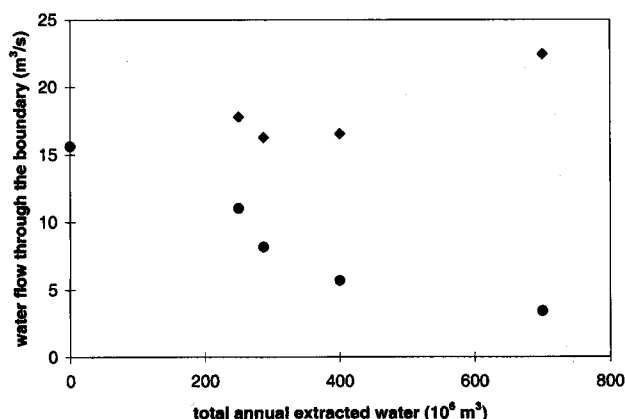


Fig. 8. Water inflow (diamonds) and outflow (circles) through the boundary of the modelled area as a function of the total annual extracted water.

use of data that correspond to very different flow conditions of the aquifer system (years 1950, 1974 and 1982) reduces the uncertainty of calibration. Validation is performed by comparing the model forecast and observation for other years ('1800' and 1997). A subjective 'trial and error' procedure has been used for model calibration, rather than an automatic inversion; nevertheless, the validation shows that the calibration is effective when performed by comparing the model outcome with observations (piezometric heads) that are sampled at the MSL and correspond to different flow conditions.

As for the specific application, the model has reproduced the trend of the phreatic head in the 20th century and shows a strong correlation between variations of the water table depth and the total volume of extracted water. In particular, the response of the aquifer system to a larger demand of water is to increase the inflow slightly and to decrease the outflow strongly. The effects of the cluster of pumping stations in the city centre are also evident from the model, which is able to reproduce the lowering of the water table in the middle 1970s to below the top of the first aquitard in the city centre.

This work opens perspectives for future developments. Present work is on transient flow; preliminary results show that the assumption of quasi-steady state conditions is not arbitrary. It is planned to model contaminant transport and subsidence; the latter phenomenon was relevant in the middle 1970s and was interpreted as an effect of the excessive drawdown of the water table.

The use of a quasi three dimensional model is very important for studies of land subsidence and contaminant transport, since in principle it permits a better prediction of the truly three dimensional flow of groundwater in a complex system. These phenomena can be very sensitive to the hydrogeological scheme, which requires further study because, presently, it is based on geometrical, rather than geological, correlation.

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