Geologia Croatica	64/1	17–30	19 Figs.	3 Tabs.		Zagreb 2011

Impact of rainfall infiltration on groundwater recharge of a deep quaternary aquifer

Kosta Urumović, Željko Duić and Branko Hlevnjak

Faculty of Mining, Geology and Petroleum Engineering, University of Zagreb, Pierottijeva 6, 10000 Zagreb, Croatia; (kosta.urumovic@rgn.hr; zeljko.duic@rgn.hr; branko.hlevnjak@rgn.hr)

doi: 104154/gc.2011.02

Geologia Croatica

ABSTRACT

Verification of recharge of groundwater in the deep Quaternary sand layers covered with leaky deposits of silt and clay has been implemented using mathematical modelling of the aquifer system in conditions of test pumping and during the 10 year exploitation of the Osijek well field Vinogradi. An analytical method for the identification of local parameters of the leaky aquifer is shown using the Excel program that enables usage of data from the relatively short term testing period, while at the same time suppressing external impacts. Boundary conditions and regional aquifer parameters have been verified by a numeric aquifer model, calibrated on the basis of three years of data from functioning field wells. The 96 mm average yearly recharge of rainfall infiltration into the aquitard was identified by validation of the model, based on information obtained from observation of groundwater levels in piezometers and in the pumped wells over a ten year period. At the end of the validation period, there was a 20 month break in pumping of the entire well field, and the measured recovery of groundwater levels provided an additional check on the results. Variation in the groundwater levels from particular layers, and water quality during exploitation are explained by the combined effects of closing of some of the well screens and the heterogeneity of the aquifer system.

Keywords: groundwater recharge, groundwater modelling, leakage, aquifer, aquitard, well field, groundwater quality

1. INTRODUCTION

There are two main reasons for emerging interest in water leakage to deep aquifers. One is linked to groundwater recharge and the long term prognosis of groundwater behaviour under conditions of intense exploitation for public water supply and irrigation. The other is linked to issues relating to the filtering of current and potential pollutants into deep aquifers. Many different techniques of quantification of groundwater recharge and periods of retention are being used to significantly reduce the uncertainty in some evaluations (SCALON et al., 2002; HEALY & COOK, 2002). Nevertheless, the most reliable methods are based on usage of data obtained from long term measurement of groundwater levels, because they show the effects of actual groundwater recharge. Research on, and measurement of the activation and the first ten years of production of the new water supply for the City of Osijek, from the Vinogradi well field, is one such documented example.

In the north-eastern parts of the Republic of Croatia, groundwater supplies occur in aquifers of the Quaternary basins. The main issues in identification of supplies and protection of groundwater are linked to quantification of rainfall infiltration into the shallow aquifers, and leakage through layers of silt and clay. This is also the main topic of research into the exploitation and vulnerability of aquifers throughout the region of Slavonia, and also in other regions of the Pannonian basin. Identification of the boundary conditions and parameters, as well as verification of groundwater replenishment in the Osijek well field Vinogradi are especially interesting. The field is located about ten kilometres from the city itself. The captured aquifer emerges in Quaternary layers consisting of sand, with intercalations of silt and clay. There are 18 wells in the Vinogradi well field (W1-W18), 200 m apart (Fig.1) that capture sand layers at depths of 40–170 m (Fig. 2). Detailed research was carried out prior to well field construction and activation at the end of July 1984. Ten years later, the well field was shut down for 20 months during pipeline reconstruction connecting it to the water processing and supply station 8km away. Data pertaining to the behaviour of groundwater during these periods enabled verification of groundwater recharge of the aquifer, groundwater behaviour and identification of the qualities of the leaky layers.

2. GEOLOGICAL STRUCTURE OF THE AQUIFER

Superficially, the geological structure is very simple, both chronostratigraphically (composed of the latest Holocene and the latest Pleistocene), and lithologically (because of an admixture or alternations of sand, silt and clay layers). Nevertheless, there are systematic differences both in morphology and lithostratigraphy that increase with depth. Also lithologic differentiation of the latest layers shows notable reflections of deep structures.

A lithological profile of the Quaternary layers was explored on the basis of data from research drilling. The shallower parts were examined by wells drilled for exploratory and water capture purposes as part of the hydrogeological research, while the deeper parts were explored as part of oil related geological research. Analysing the data collected from boreholes throughout the Northern region of Croatia, URUMOVIC et al. (1976) note two contrasting levels in the otherwise general alternation of fine and coarse clastic Quaternary sediments. The boundary between the two is labelled as a conditional marker Q'. The upper layers above this marker are characterised by their unconsolidated material, relatively high proportion of permeable coarse clastic layers and quantity of fresh water quantity. Below the Q' marker, consolidation is higher, the proportion of permeable, coarse clustered layers is smaller and there are significant differences in the mineral composition of waters from different layers both vertically and horizontally. Water mineralisation is generally higher in these layers in comparison to the upper layers. The layers above the Q' marker (URUMOVIĆ et al., 1976, 1978) are dated as being of mid – late Pleistocene, Holocene ages, while those below the marker are dated as being of lower Pleistocene and Plio–Pleistocene age. The fact that above the Q' marker there is a sequence of layers in which coarse clustered sediments are predominant leads to the conclusion that these are a result of regional changes in the properties of sedimentary material caused by change in palaeoclimatic conditions.

In the middle of the Quaternary layers above the Q' marker, there is another lithological horizon that resembles the Q' marker. It is marked as X in the regional correlation profile (Fig. 2). The coarsest sediments in the entire Eastern region of the River Drava depression (URUMOVIĆ, 1982) occur above this marker. Immediately above this marker, singular, small gravel granules occur over most of this area.

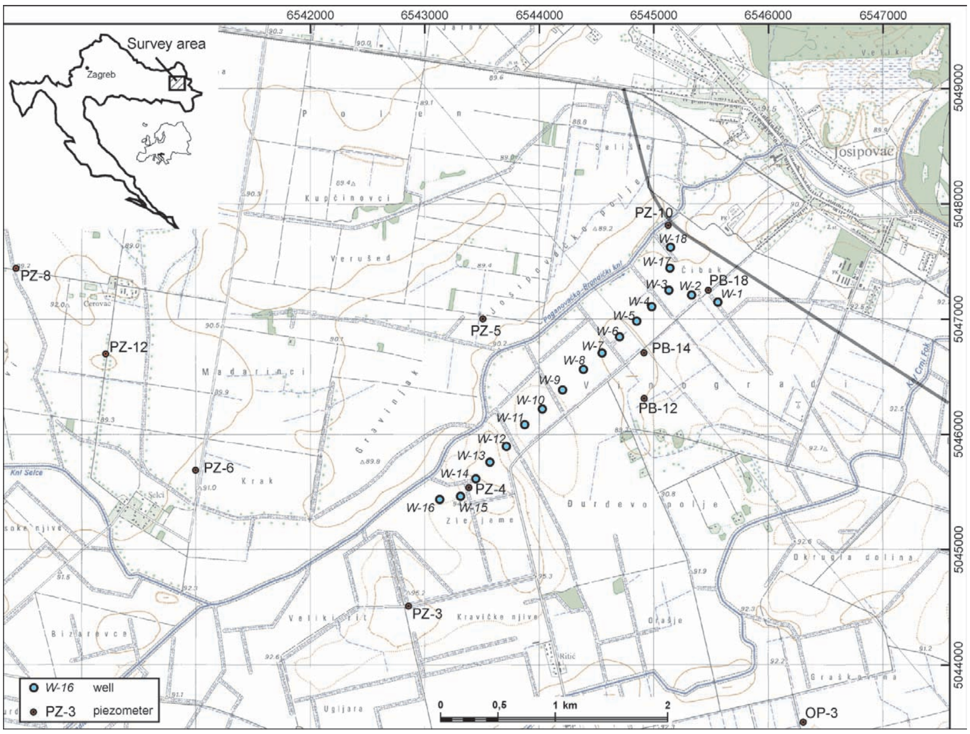

Figure 1: Location of the Osijek Vinogradi well field and tested piezometers.

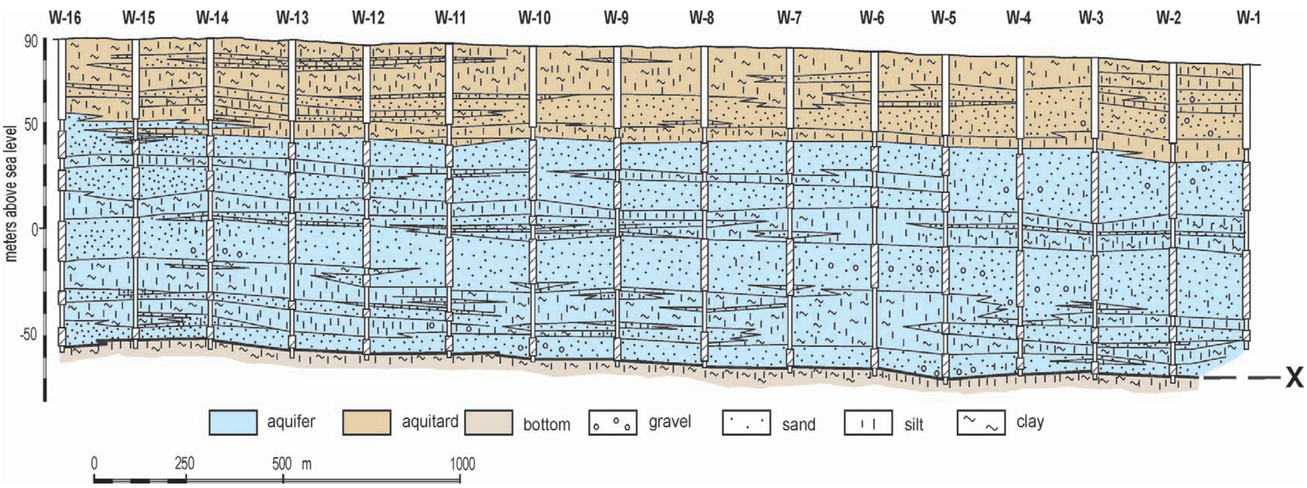

Figure 2: Cross-section of the Quaternary layers of the Vinogradi well field.

The impression is that these layers are a consequence of a powerful widespread transgression when in Baranja, in the inundated part of Podunavlje; gravel layers 50 m thick were deposited. At this level, southeast of Osijek, at 60 m depth, there are bored pebbles of gneiss granite that are 120 mm in diameter. The diameter of these pebbles is extraordinary for the area; however they have also appeared in many other boreholes. Wireline log correlation confirms the regional nature of their occurrence representing the coarsest sediment deposition in this area.

Coarse grained materials of the wider Vinogradi well field area are mostly represented by uniform small – medium sand grains 0.05 to 0.3 mm in size, of up to 85% quartz grains. The fine-grained sands contain a high proportion of micas. Sometimes there is so much of it that the grains shine in the borehole core and appear much larger than they really are. The sands also contain fragments of carbonate rock; feldspars, amphibole, epidote and granite, suggesting that the original rocks were predominantly part of the Alpine massif and to a lesser degree, the Slavonian Mountains.

Fine grained materials are represented by silt, sandy-silt, silty to fine clay. Colouration ranges from gray, silver-blue to green. There are also concretions as fragments, angular to round in shape, usually light gray to green and greenish-white in colour. While the layers of silt and clay are moist, concretions are usually dry because of suction, and in wireline logs they appear as a contrasting deflection. The mineral composition is predominantly quartz, with subordinate micas and feldspars. In some parts there are considerable quantities of calcite, particularly among grains coarser than 0.06 mm, while montmorillonite forms >10% of the finest fraction.

An important characteristic of the Quaternary layer structure is the alternation of coarse and fine layers, evident at two scales. Texturally, thin, almost millimetre alternations of silt and sand, are sometimes laminated and angled. These are notable where fine grained sand and silt layers are mostly present. In the thicker sand layers, some sedimentary cycles typically start with sand >0.2 mm grain size which then fine upwards, so granulometric parameters illustrate sedimentation cycles in almost every thicker layer.

These lithological characteristics generate hydraulic anisotropy evident in laboratory testing and a general heterogeneity that may have effects of anisotropy.

3. ANALYTICAL IDENTIFICATION OF HYDROGEOLOGICAL PARAMETERS AND BOUNDARY CONDITIONS

At the Vinogradi well field, all exploitation wells have had screens installed into all layers of pure, uniform sands. This has practically excluded the effects of anisotropy on groundwater flow, and in these conditions radial flow was horizontal, as it is the flow in a homogeneous and isotropic aquifer. The process of drawdown during well pumping under these conditions, shows recognisable effects of all the hydrogeological parameters of the aquifer, in accordance with the Hantush theory of leaky aquifers (HANTUSH, 1960). The effects of the parameters of the pumped layer are easily recognised, even at the very early stages of well pumping at constant discharge. Estimation of the transmissivity and storage coefficient, by means of a derivative method using the early time drawdown (STRAFACE, 2009), can be derived even for $t>r^2S/4T$. Longer pumping results in a widening cone of depression and the percolation effects begin to be integrated through aquitard leakage and are detected in diagrams showing time-drawdown. In practical terms, very long term pumping is often affected by external impacts. In such cases it is necessary to use data from the early phase of pumping that is dominated by the influence of pumping of the tested well. It was exactly this kind of difficulty that arose in the interpretation of the test pumping of wells W-7 and W-15 at the Vinogradi well field that had been conducted prior to their connection to the water supply system.

Analytical identification of boundary conditions and aquifer parameters used data from the graphic image of the groundwater level drawdown in the semi-log diagram of time-drawdown for all the measured wells (Figs. 3 and 4).

The non-stationary distribution of water level drawdown around the lone standing well pumping from the semi-confined aquifer can be described as a relatively simple HANTUSH-JACOB (1955) solution:

$$s = \frac{Q}{4\pi T}\int_u^\infty e^{-x-\frac{r^2}{4\lambda^2 x}}\frac{dx}{x} = \frac{Q}{4\pi T}W\left(u, \frac{r}{\lambda}\right) \quad (1)$$

Where: $W(u,r/\lambda)$ well function for leaky aquifers, $u = \frac{r^2 S}{4Tt}$, s is drawdown, Q is constant discharge of the well, T is transmissive coefficient, S is storage coefficient, $\lambda = \sqrt{T\, b'/k'}$ is leakage factor, b' and k' are the thickness and hydraulic conductivity of the aquitard bed through which leakage occurs, r is the distance from the centre of the well to any point in the field, and t is the time since pumping began. Simplification in formula (1) relates to the dismissal of elasticity effects of the leaky layer and the occurrence of source layer drainage. The latter may become very significant in long term pumping, the simplification is therefore more appropriate in the beginning of test pumping than in the final phase.

The slope of the theoretical drawdown curve s in relation to time t in semi-log scale shows HANTUSH (1956) by derivation of equation (1) in relation to $log\ t$:

$$\delta s = \frac{\Delta s}{\Delta \log t} = \frac{2.3Q}{4\pi T}e^{-u-\frac{r^2}{4\lambda^2 u}} \quad (2)$$

This applies to any point on the curve, and for a point of inflection, the second derivative is equal to zero:

$$\frac{\partial^2 s}{\partial(\log t)^2} = 4.6\frac{tQ}{4\pi T}e^{-u-\frac{r^2}{4\lambda^2 u}}\left(-\frac{\partial u}{\partial t} + \frac{r^2}{4\lambda^2 u^2}\frac{\partial u}{\partial t}\right) = 0 \quad (3)$$

Which is achieved when $\frac{r^2}{4\lambda^2 u^2} - 1 = 0$, so at the inflection point

$$u_i = \frac{r}{2\lambda} = \frac{r^2 S}{4Tt_i} \quad (4)$$

and the slope of the drawdown curve at the inflection point results from the insertion of (4) into (2):

$$\delta s_i = \frac{2.3Q}{4\pi T}e^{-r/\lambda} \quad (5)$$

where $e^{-r/\lambda}$ expresses how the effect of percolation through the aquitard bed is increasing by the distance from the pumped well. When $\lambda \rightarrow \infty$, $e^{r/\lambda} = 1$ the aquifer becomes confined and equation (5) turns into the well known Cooper-Jacob (1946) equation.

For the identification aquifer system parameters on the basis of data from the beginning of the testing period, testing of at least two objects is required. If there is at least 50% of maximum drawdown, and this is achieved when $t > r\lambda S/2T$ the graphic-analytical (HANTUSH, 1956) or the type curve (URUMOVIĆ, 1978) method may be used. However, nowadays it is more appropriate to use computer programmes such as MS Excel that enables the elegant application of comparison of theoretical functions and experimental data. This proved very useful for identification of the range of values of aquifer parameters and the effect of boundary conditions on the basis of test pumping data of the wells situated in the middle (W-7) and on the edge (W-15) of the Vinogradi well field (Fig. 1). Each test pumping lasted for 26 days. The first occurred in autumn 1983, and the second in the winter of 1984. Test pumping was done in the period of recession when the stabilization of the level on maximal drawdown was not expressed, nor was its tendency that could facilitate graphic extrapolation of maximum drawdown (Figs. 3 and 4). However the effects of groundwater recharge are clear through the spatial distribution of the slope of the bi-log curve s(t) in some of the tested wells.

When identifying real boundary conditions and the values of parameters, the slope of the drawdown curve is not directly conseidered but see the function:

$$\left[e^{-r/\lambda}\right]' = \frac{\delta s_i T}{0.183Q} \quad (6)$$

calculated for the slope reading δs_i and the presumed T and λ. The related series of points are compared to the function

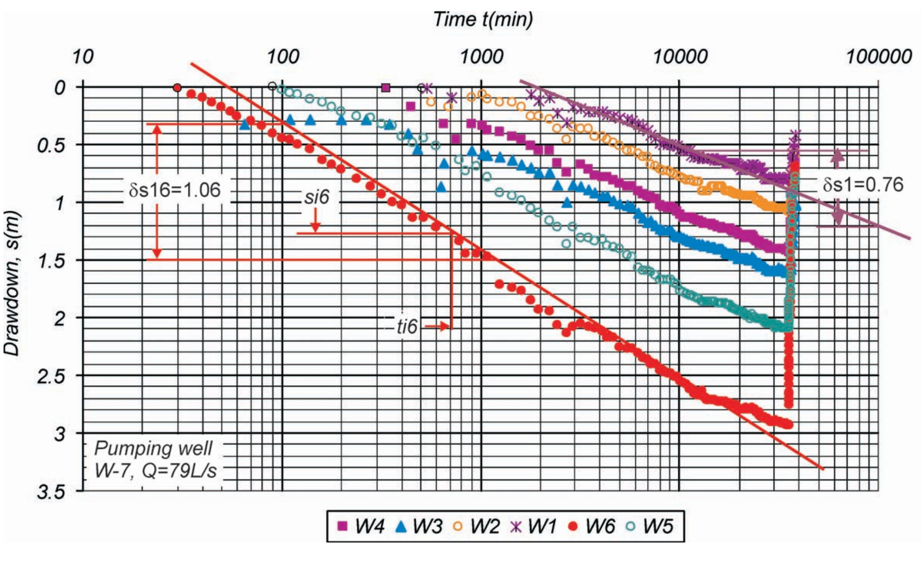

Figure 3: Semi-log plot of drawndown versus time in observation wells during testing of W-7.

graphs r/λ, $exp(-r/\lambda)$ and $exp(r/\lambda=0)$ (Fig. 5 and 6). The last one is a horizontal line that represents the case of a non-leaky aquifer.

In the procedure presented above, a trial method is used to identify values for the transmissivity and leakage factors, and a coefficient of storage according to equation (4), whereby time t_i is read on the diagrams (Fig. 3) for

$$s_i = \frac{Q}{4\pi T} K_0(r/\lambda) \qquad (7)$$

where K_0 is the modified Bessel function of the second kind and zero order.

The performed procedure confirms the boundary conditions for a leaky aquifer, identifies values of the hydraulic parameters of the aquifer, and confirms that during pumping of the tested wells, groundwater flow was the same as for a homogeneous aquifer.

The method performed has two characteristics, the main one being based on the analytical solution of the continuity equation for the semiconfined aquifer. Therefore the hydraulic homogeneity of the aquifer is assumed. The second characteristic is that it uses a technique of interpretation that is classed as a trial method, which is the implicit characteristic of all types of curves and the explicit characteristic of all numerical methods.

4. NUMERICAL MODELLING

The investigation has been initiated under the assumption of a leaky aquifer as the general case of an artesian aquifer, and has been directed on recharge identification. Groundwater flow through such a system can be described by two partial differential equations, one for the aquifer and the other for the covering aquitard. For the aquifer:

$$\frac{\partial}{\partial x}(Kb\frac{\partial h}{\partial x}) + \frac{\partial}{\partial y}(Kb\frac{\partial h}{\partial y}) = Q + S(\frac{\partial h}{\partial t}) - \frac{k'}{b'}(h' - h) \qquad (8)$$

Where in (Fig. 7): $h(x,y,t)$ is the hydraulic head in the aquifer at time; t, $h'(x,y,t)$ is the hydraulic head in the covering aquitard at time t; K equals the hydraulic conductivity of the main aquifer for horizontal flow; k' is the covering aquitard's hydraulic conductivity for vertical flow; b (x,y) is the thickness of the aquifer; b' (x,y,t) represents the saturated thickness of the covering aquitard at time t; $S(x,y)$ is the storage of the aquifer; and $Q(x,y,t)$ is the net rate of pumping at time t. The left side of Eq. (8) represents the horizontal flow in the aquifer, the right side the vertical flow.

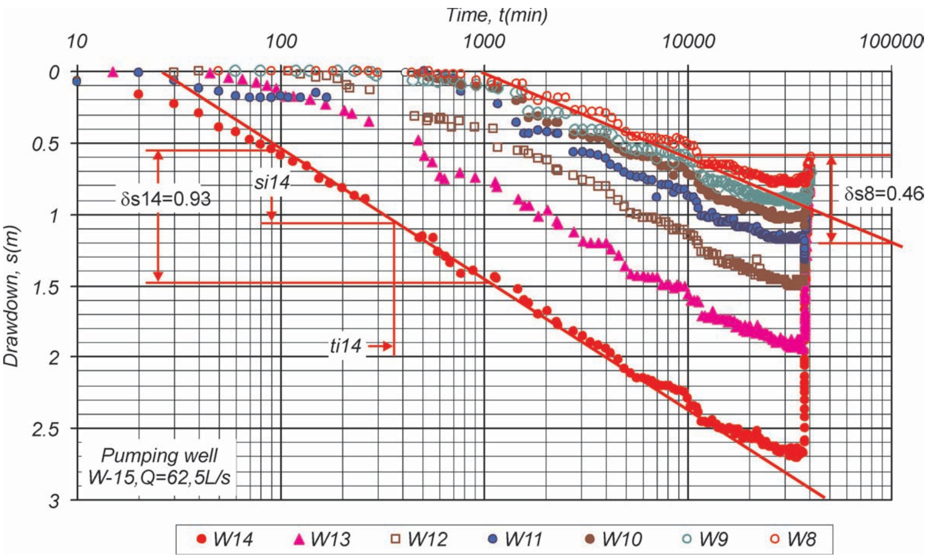

Figure 4: Semi-log plot of drawndown versus time in observation wells during testing of W-15.

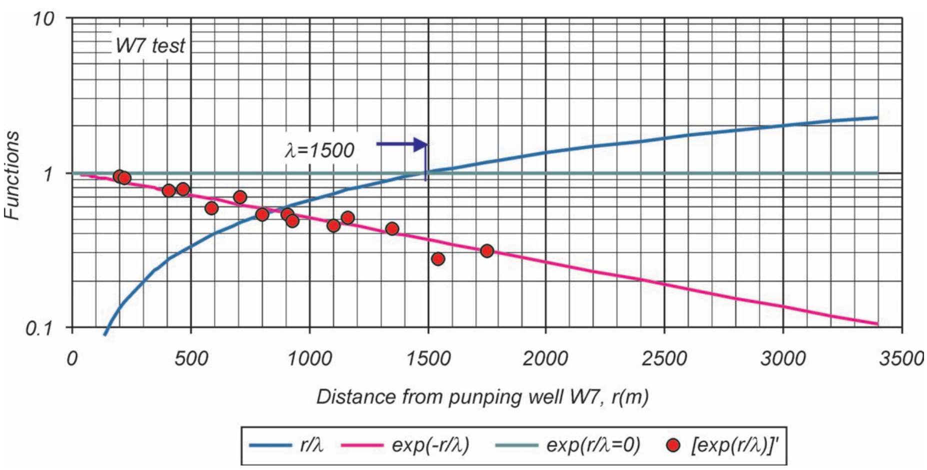

Figure 5: Semi-log graph of theoretical functions and data of the test well W-7 (Q=79 l/s, transmissivity $T=1.05*10^{-2}$ m^2/s, leakage factor $\lambda=1500$ m).

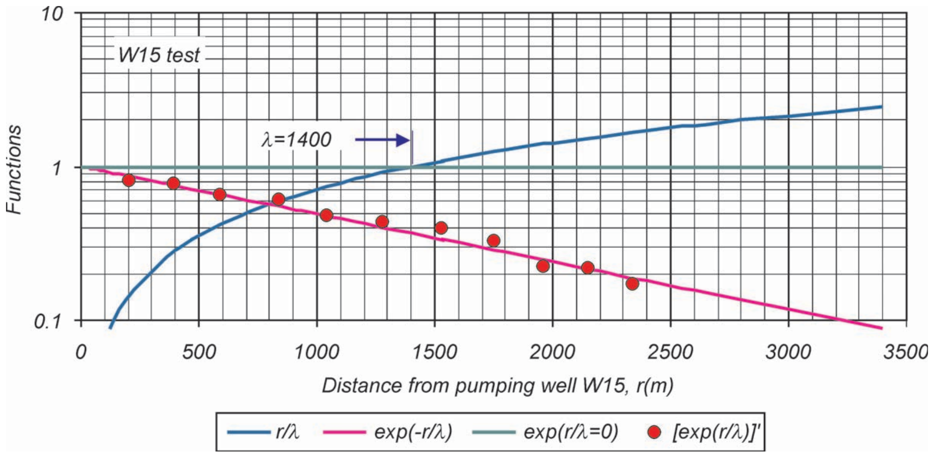

Figure 6: Semi-log graph of theoretic functions and data of the test well W-15 (Q=62.5 l/s, transmissivity T=1.0*10^{-2} m^2/s, leakage factor λ=1400 m).

For the covering aquitard, there is a one-dimensional differential equation. Assuming that the aquitard has a free water table, so that it's saturated thickness *b'* is not constant, but may vary with time. Since both the water received from infiltration, and the water released by a falling water table percolate through the aquitard before reaching the aquifer, the following equation applies:

$$I_w - n'\frac{\partial h}{\partial t} = \frac{k'}{m'}(h' - h) \tag{9}$$

where I_w is the net rate of infiltration, and *n'* is the specific yield of the covering aquitard. Owing to its low permeability, and thin intercalated sand layers, we also assume that no pumping will occur in this layer.

In equation (8), if the hydraulic conductivity of the aquitard *k'=0*, then (*k'/b'*)(*h'-h*)=*0*, and the aquifer is confined. In equation (9), if there is no infiltration, $I_w=0$, we have a decrease of aquitard water level as well as the piezometric level in the main aquifer (if *Q=const.)*

The finite difference method of approximating the solution of differential equations (8) and (9) is used. The mathematical model covers an area of 1410 km^2 discrete nodal areas, networks of rectangles, squares and polygons 0.04–41 km^2 (Fig. 8).

4.1. Boundary condition analysis and model calibration

The first three years of pumping are the most interesting in the whole process of recharge identification. In this period, pumping rate was gradually increasing, and at the same time a decrease in piezometric level occurred. Three cases have been analysed through calibration:

a) confined aquifer (*k'=0*; – impermeable covering bed)
b) semi confined aquifer (*k'>0*; $I_w=0$ – permanent drainage of covering aquitard)
c) semi confined aquifer (*k'>0*; $I_w = I_w(x,y,t)$ – calibrated solution)

The water level behaviour in all these cases is presented for four testing holes, and piezometers; PZ-4 between pumping wells 14 and 15; PZ-8 about 4400 m west of the well field, PZ-12 about 3300 m west of the well field and PB-14 300 m east of well W-6 (Fig. 1). Divergence of the first two curves confirms leakage from the covering aquitard bed after at least a few months. In contrast, after a longer period of a few years, recharge of the aquitard becomes rather obvious (Fig. 9).

The effect of the value and the distribution of hydraulic conductivity and thickness of the aquifer in al three cases are

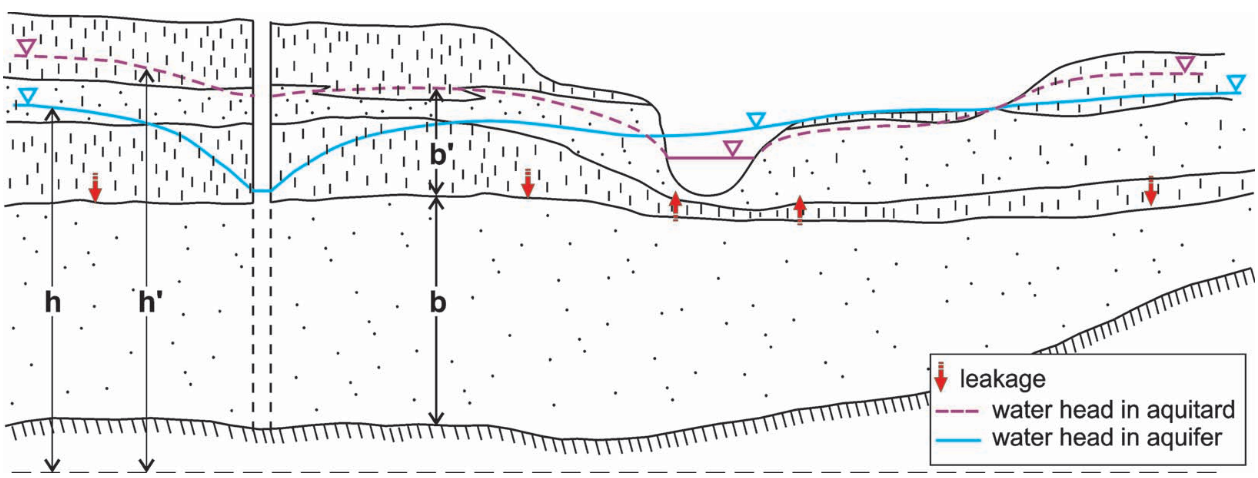

Figure 7: Sketch of the hydraulic schematisation of the aquifer system.

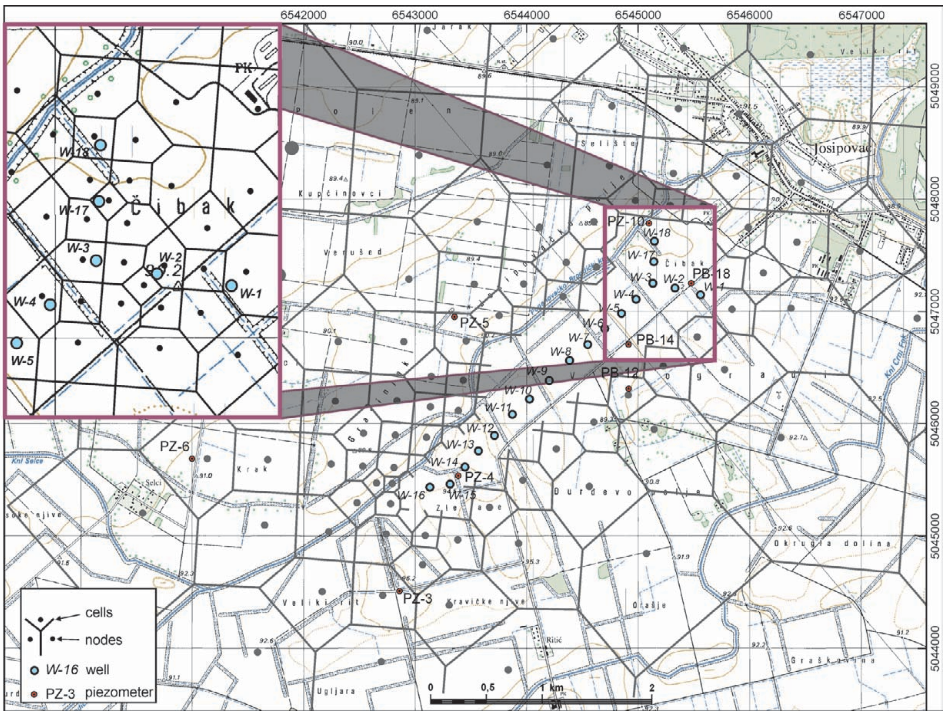

Figure 8: Graded network superimposed on the arrangement of pumping wells and piezometers.

only predominant during the first couple of days of pumping, when considerably more than 50% of drawdown in the exploitation wells takes place. Later, the effect of drainage from the covering aquitard deposit slowly prevails, together with aquitard recharge through rainfall infiltration. Model calibration is therefore implemented for two discrete time periods. The first step in model calibration simulated the behaviour of the water level during the first two days of pumping, during which process the hydraulic conductivity and the aquifer storage capacity were checked in the well field area considering the values determined at the test pumping.

The medium hydraulic conductivity of the captured aquifer is from $1.4*10^{-4}$ to $2.0*10^{-4}$ m/s, and the vertical hydraulic conductivity of the covering leaky deposits is from 10^{-8} to 10^{-7} m/s. Aquifer storage ranges from $5*10^{-4}$ to $2*10^{-3}$, and the average effective porosity of the aquitard deposit is from 0.03 to 0.16.

In the second phase, simulation of the fluctuation of water level in the first 3.5 years of pumping of the Vinogradi well field was conducted (Fig. 9). In this period of water level fluctuation, the predominant influence comes from the porosity of the covering leaky layers and the water stored therein, together with the capacity of some wells and primarily the whole well field, and from groundwater recharge by rainfall infiltration into the aquitard deposit. Rates of 80–170 mm total annual infiltration to the aquitard water table have been identified for this period (Urumović et al., 1996). This represents 12–19% of the total annual precipitation, and these values compare well with agricultural investigation of the water balance at the adjacent Bizovac experimental field (Tadić et al., 1994) and statistical correlation of precipitation and water level fluctuations (Urumović, 1982).

4.2. Model validation

Numerical model validation of the aquifer system was primarily initiated by the occurrences of significantly lower than speculated falling groundwater levels in some wells, (Fig. 14) which could have been interpreted in different ways. An additional incentive for the validation was the data gathered during the 20 months interruption of pumping due to reconstruction of the pipelines between the well field and the water processing facility. During that time, the water level in the wells rose to a height only 0.5 m lower than the level registered before the beginning of pumping (Table 2), and in some distant piezometers it rose above the initial height (Table 1). These observations, as well as a lot of data gathered during groundwater monitoring at the Vinogradi well field proved challenging for the exact mathematical verification of possible causes of the registered state of groundwater before and after the interruption of pumping. The research was primarily focussed on groundwater recharge of the deep aquifer layers, and therefore as part of the validation of previously calibrated models, attention was on the manifestations of groundwater level fluctuation in pumped wells and observed piezometers during the 10-year period analysed.

Prior analysis of boundary conditions illustrated the aquifer inertia in the process of drainage of the covering aquitard deposit (Fig. 9). This is expressed by a strong reduction in drawdown in the captured aquifer as a consequence of water leaking from the aquitard deposit, and in the case when there is no recharge by infiltration of rainfall, (Fig. 9, case k'>0, I_w=0). In such conditions the effect of rainfall infiltration is completely suppressed by the effects of the pumping regime of some wells and the whole well field. The effects

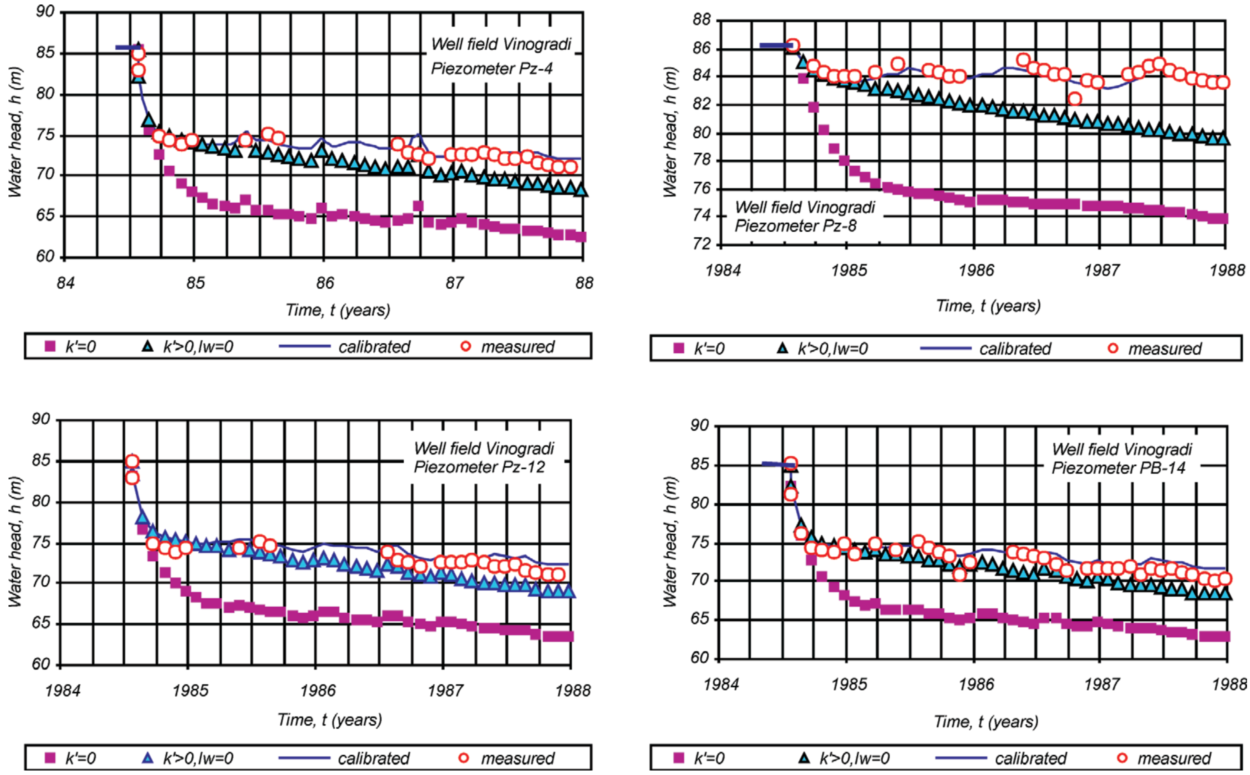

Figure 9: Results of simulated groundwater levels in the model calibration process.

of recharge are clearly visible only in very distant observed piezometers where vertical water balance components become dominant and the groundwater fluctuation primarily expresses the local manifestation of the vertical water balance. For long term, regionally significant trends mean annual rainfall has a dominant effect and its influence is often carried over into the next year. In this case it turned out to be a good outcome, as there were intervals of poor water level measurement in the 10-year observation span, especially during the war in 1991 and 1992.

The main idea behind validation of the mathematical model of the aquifer system was measurement of the water level in piezometers with screens installed in most of the aquifer layers of the captured aquifer (Fig. 10 to 13). The accuracy of simulated water level fluctuations for the aquifer under groundwater recharge of the source aquitard layer, where infiltration of 48 to 180 mm annually was equally distributed in 12 monthly increments, has been checked. The best match of the measured data and the simulated curve (Fig. 10 to 13) occurs where constant recharge of aquitard water is from 8 mm per month, or 96 mm/year.

It can therefore be concluded that the water level fluctuation in the area influenced by the well field is primarily conditioned by the pumping regime (Fig.15), and the long term trend is driven by the capacity of the average recharge of the aquifer system.

The results of identical simulations for some of the pumping wells are shown in Figs. 14 & 15. At the start, these graphs show a good match between the calculated water levels and the control measurements. In the later period there is a divergent sequence of data. Water levels in pumping wells start to diverge from the simulated sequence as soon as the pumping is stopped, when the water level rises up to that of the simulated curve. After pumping of all wells ceased with disconnection from the water supply system, the water level suddenly rises to the approximate level in the nearby piezometers and gradually reaches the initial level recorded before the start of pumping (Fig. 10–15 , Table, 1–2). The simulated 10-year period finished at the end of May 1994, 13 months after pumping was stopped.

The average annual rainfall recorded for the period at a rain station near Osijek, 10 kilometres east of the well field was 620 mm per annum, excluding 1991 and 1992 when data was incomplete (Table 3). This amount is around 5% below the 30-year average. As the average rainfall amount gradually rises westwards (with an average gradient of round 2 mm/km), it can be estimated that the amount of calculated water recharge in the aquitard deposit by rainfall infiltration was around 14%–15% of the average annual rainfall. This amount is well within the range from earlier prognosis and research of this area. Considering that the validation was made on the basis of data on the effects of a relatively strong long term pumping regime, there are important factors that give these values regional significance.

5. THE IMPACT OF AQUIFER ANISOTROPY ON GROUNDWATER LEVELS AND QUALITY

The mechanism of groundwater behaviour within the aquifer system was recorded in nested piezometers that all had

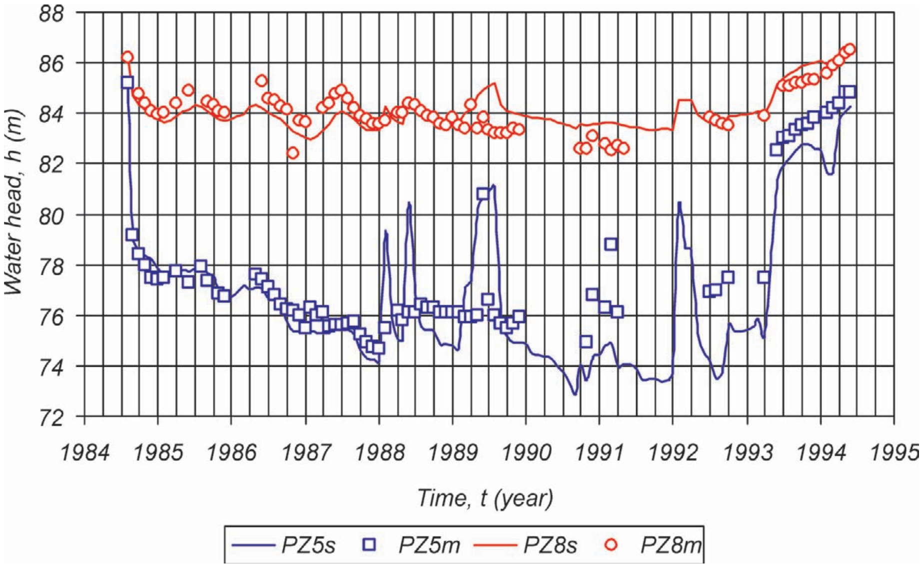

Figure 10: Simulated (s) and measured (m) groundwater heads in piezometers west of the well field.

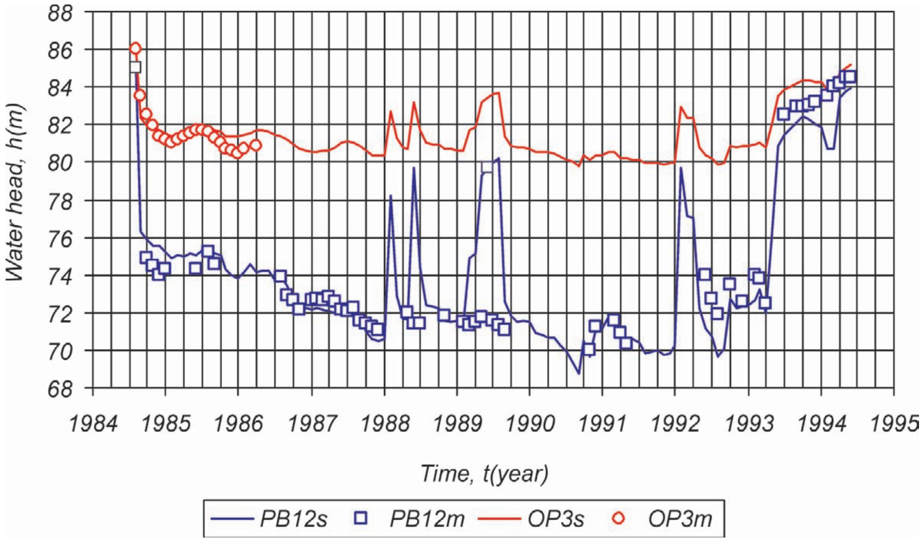

Figure 11: Simulated (s) and measured (m) groundwater heads in piezometers east of the well field.

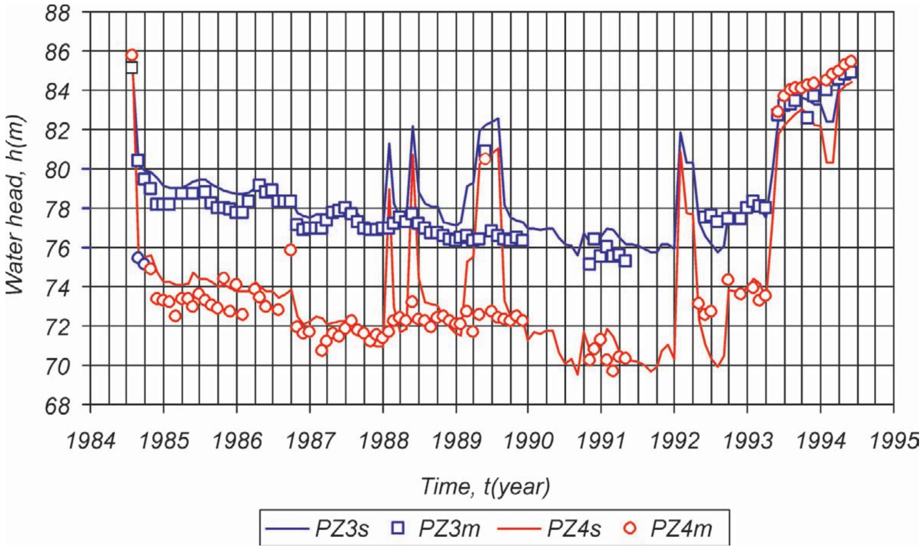

Figure 12: Simulated (s) and measured (m) groundwater heads in piezometers south of the well field.

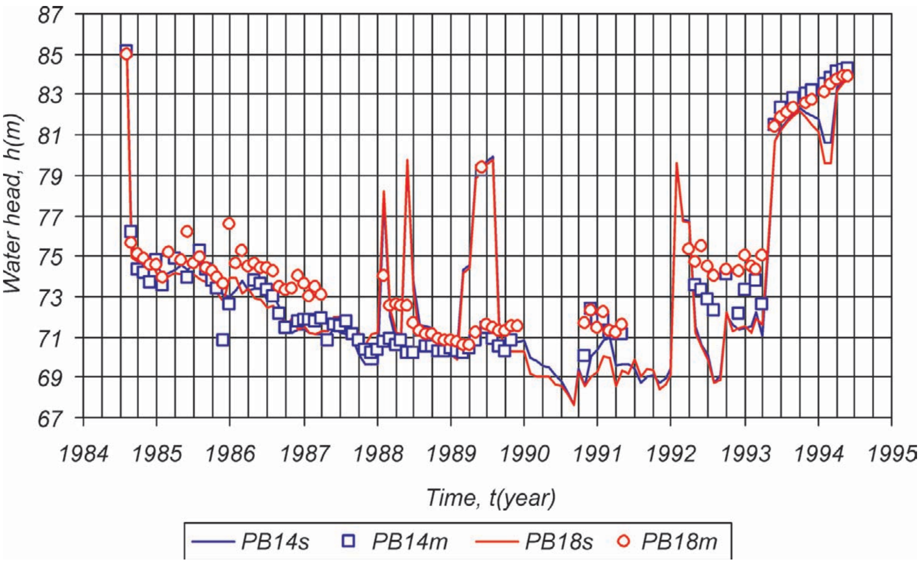

Figure 13: Simulated (s) and measured (m) groundwater heads in piezometers in the north part of the well field.

at least two piezometers, one with screens in all the major layers of the captured aquifer (PZ-4 and PZ-3, Fig 17), and the other with an isolated screen at 24–27 m in the aquitard deposit (PZ-4A and PZ-3A, Fig 17). Data regarding groundwater levels in these piezometers were used in the process of calibration and validation of the mathematical model. Nested piezometers that also measure groundwater level in specific sand layers, were placed near some of the pumping wells, in order to note the effects of heterogeneity and anisotropy on the groundwater level. Therefore the nested piezometer PZ-4, had an additional screen at 165–168 m for PZ-4/1, 102–105 m for PZ-4/2, 69–72 m for PZ-4/3 and 54–57 m for PZ-4/4. The groundwater level in these is a little different at times of inactivity, but during pumping, it was only at the same level for the first 30 months (till the end of February 1987, Fig. 16). After this period, there was a strong divergence in groundwater level in some piezometers. In a few the groundwater level dropped significantly (PZ-4/1), whereas it rose in others (PZ-4/2, 4/4), so that the maximum difference in groundwater level in 1991 was round 15 m (Fig. 16). At the same time the nearby wells W-14 and W-15 (Fig. 14) experienced a sudden drop in water level during pumping and when they were closed off, the water level rose to the pre-pumping height. This drop in well capacity is a

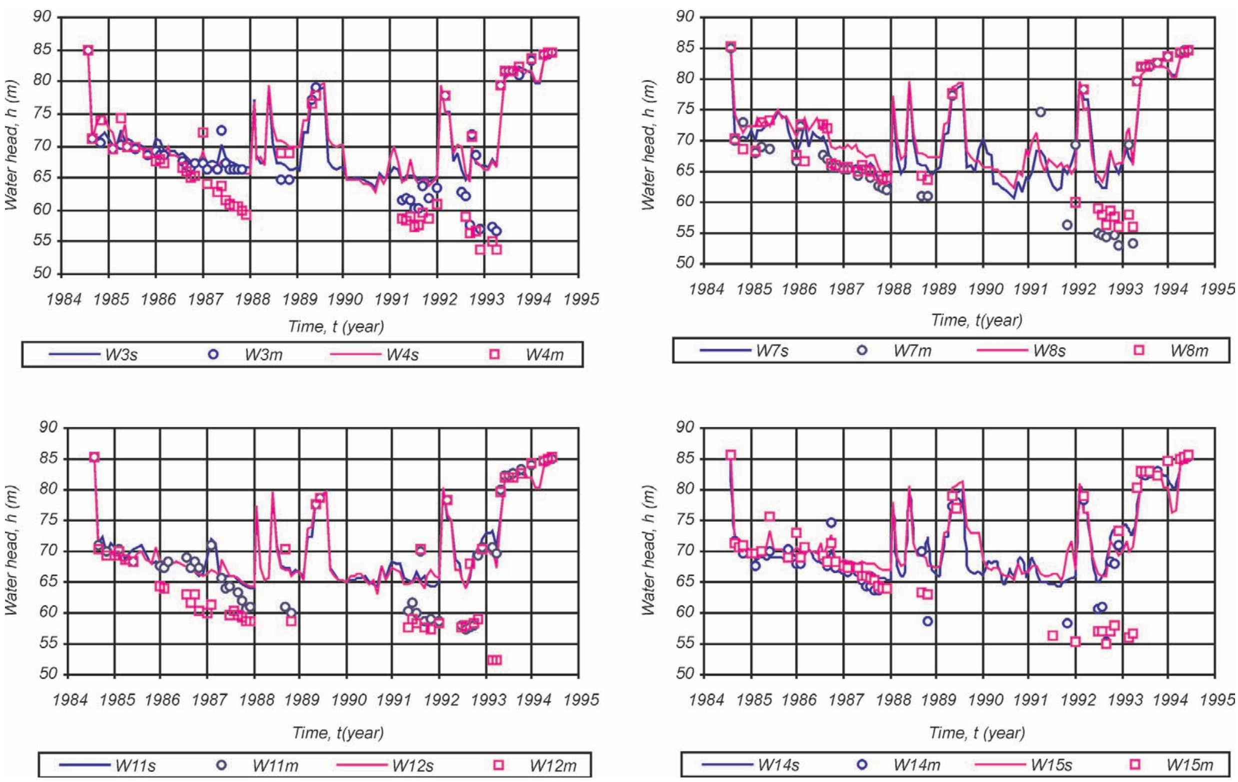

Figure 14: Simulated (s) and measured (m) groundwater heads in the pumping wells.

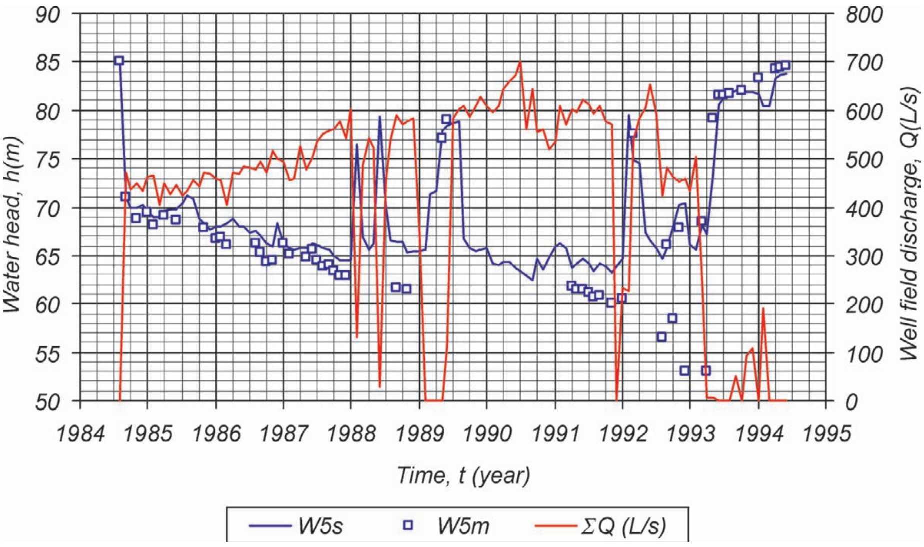

Figure 15: Total capacity (ΣQ) of the Vinogradi well field and water heads in well W-5.

consequence of the closure of some screens which happens to be a common occurrence in aquifers with higher levels of dissolved iron.

The rate of aquifer system recovery after pumping ceased between April 1993 and November 1995 is illustrated in Fig. 17. At the moment of pumping ceasation, the groundwater level in all layers of the captured aquifer rises above the groundwater level of the aquitard deposit (PZ-4A) and therefore the immediate area of well field experiences a temporary draining of the aquifer into the aquitard. Groundwa-

Table 1: Measured and simulated values of groundwater head in tested piezometers prior to and after the simulated period.

	PZ-3		PZ-4		PZ-5		PZ-6	
	s	m	s	m	s	m	s	m
Initial	85.13	85.13	85.75	85.75	85.20	85.20	85.21	85.21
End	84.69	84.90	84.43	85.40	84.26	84.80	85.56	86.00
Difference	-0.44	-0.43	-1.32	-0.35	-0.94	-0.40	-0.65	-0.21
	PZ-8		PB-12		PB-14		PB-18	
	s	m	s	m	s	m	s	m
Initial	86.20	86.20	85.05	85.05	85.13	85.13	84.95	84.95
End	86.70	86.50	83.96	84.50	83.91	84.30	83.75	83.90
Difference	+0.5	+0.30	-1.09	-0.55	-2.22	-0.83	-1.20	-1.05

Table 2: Measured and simulated value of water head in pumping wells before and at the end of simulated period.

	W1		W2		W3		W4		W5	
	s	m	s	m	s	m	s	m	s	m
Initial	84.9	84.9	84.9	84.9	85.0	85.0	85.0	85.0	85.1	85.1
End	83.8	84.4	83.8	84.5	83.8	84.5	83.8	84.6	83.9	84.6
Difference	-1.1	-0.5	-1.1	-0.4	-1.2	-0.5	-1.2	-0.4	-1.2	-0.5
	W6		W7		W8		W9		W10	
	s	m	s	m	s	m	s	m	s	m
Initial	85.1	85.1	85.1	85.1	85.2	85.2	85.3	85.3	85.3	85.3
End	83.9	84.7	84.0	84.8	84.0	84.6	84.1	84.9	84.1	84.9
Difference	-1.2	-0.4	-1.1	-0.3	-1.2	-0.6	-1.2	-0.6	-1.2	-0.4
	W11		W12		W13		W14		W15	
	s	m	s	m	s	m	s	m	s	m
Initial	85.4	85.4	85.5	85.5	85.6	85.6	85.8	85.8	85.8	85.8
End	84.2	85.1	84.3	85.2	84.3	85.4	84.4	85.5	84.5	85.6
Difference	-1.2	-0.3	-1.2	-0.3	-1.3	-0.2	-1.4	-0.3	-1.3	-0.2

Table 3: Rainfall in Osijek during the analysed period and the 30-year average.

Year	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Σ
1980.	31.7	47.3	36.7	92.5	100	76	58	65.7	15.7	51.3	116.1	53.4	744.4
1981.	58.2	33.7	96.2	28.2	22.3	180.2	28.3	49.6	64.7	90.9	47.1	117.7	817.1
1982.	14.6	15.6	46.2	71	30.6	59.4	79.5	97.4	23.1	26.1	32.7	83.6	579.8
1983.	28.1	39.8	27.2	19.6	62.3	67.2	42.5	32.4	100	20.3	10.6	16.9	466.9
1984.	99.9	33.9	36.5	54.7	89.9	73.7	37.4	36.2	50.3	51.1	40.5	19.7	623.8
1985.	45.8	51.9	49.7	56.3	32.6	119.5	29.7	88.2	8.8	9.6	105	21.6	618.7
1986.	67.2	79.3	43.3	38.3	42.1	57.8	40.7	60.6	6.9	60.4	16.1	23.6	536.3
1987.	97.2	6.2	54.3	59.9	170.6	77.6	33.4	27.4	18.3	23.7	112.8	37.6	719
1988.	35.4	41.6	116.4	39.7	43.6	96.3	29.4	14.6	58.4	30.7	28.7	30	564.8
1989.	6.4	9.2	48.3	46.1	106	83	63.7	95.6	35.7	48.6	42.2	18	603.1
1990.	11.8	39.4	25.8	38.4	26.2	101.4	38.7	42	72.3	33.7	52.7	58.5	540.9
1991.	28.6	30	37.3	78.8	102.1	26.4	119.2	89.2	40.8				552.4
1992.			13	59	39.7	112.4	42.3	18.9	36.4	155.3	105.1	50.2	632.3
1993.	22.7	8.9	61.6	42.6	47.7	69.5	55.2	56.7	58.8	43.1	95.5	92.5	654.8
1994.	45.4	31.7	34.8	52.4	34.6	88.2	19	83.6	120.3	57.5	16.1	45.1	628.7
1995.	70.9	52.7	44.5	52.2	96.4	105.5	26.7	85.6	123.2	5.8	53.6	104.4	821.5
	41.5	32.6	48.2	51.9	65.4	87.1	46.5	59.0	52.1	44.3	54.7	48.3	631.5
Mean						1961–1990							650.4
						1971–2000							654.2

ter exchange between the main aquifer and the aquitard hides the effects of rainfall infiltration in the well field area until the end of the break in pumping which further confirms the results of the implemented simulation. These influences diminish with distance and at the location of the nested pyrometer PZ-3 the amount of rainfall infiltration half way thorugh the no pumping period exceeds the effects of water exchange with the main aquifer.

Chemical composition of groundwaters also reacts to the changes of the hydraulic resistance in the well casing. The quantity of common anions (HCO_3^-, SO_4^{2-}, Cl^-) remains almost unaltered, while the quantity of cations (Ca^{2+} and Mg^{2+}) changes in such a manner that the quantity of calcium drops, while the quantity of magnesium stays the same (Fig. 18). This could be ascribed to larger amounts of water from the deeper layers that have a higher content of sodium. Still, the most significant relative change is in the quantity of accessory components of iron and manganese (Fig. 19). The changes of these components lead to the conclusion that the influence of iron utilising bacteria on the process of closing of the well screens is faster in layers that are richer in dissolved iron and manganese, and that larger amounts of water are pumped from layers with lower amounts of these components. Therefore, the groundwater level drops rapidly in such layers, and rises in those that are less pumped. This results in a radical increase in the effect of heterogeneity and

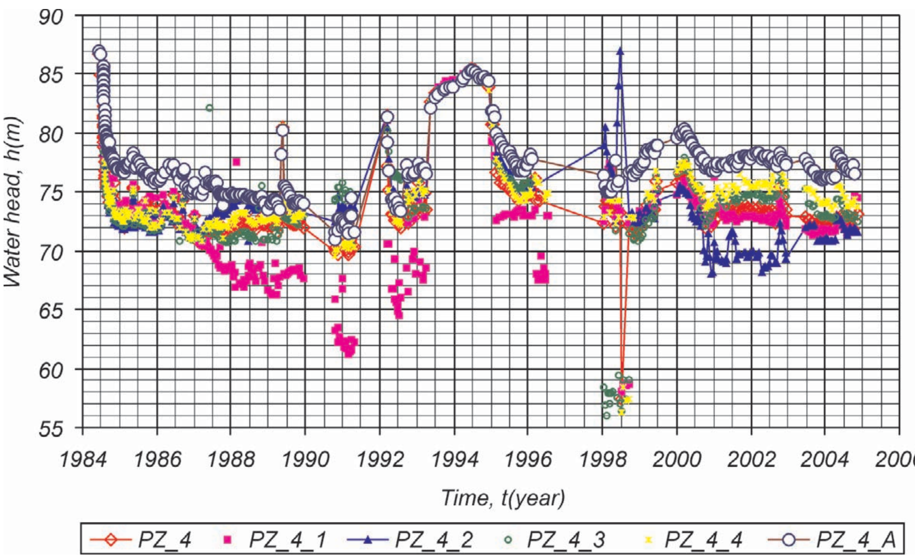

Figure 16: Hydrograph of nested piezometers PZ-4.

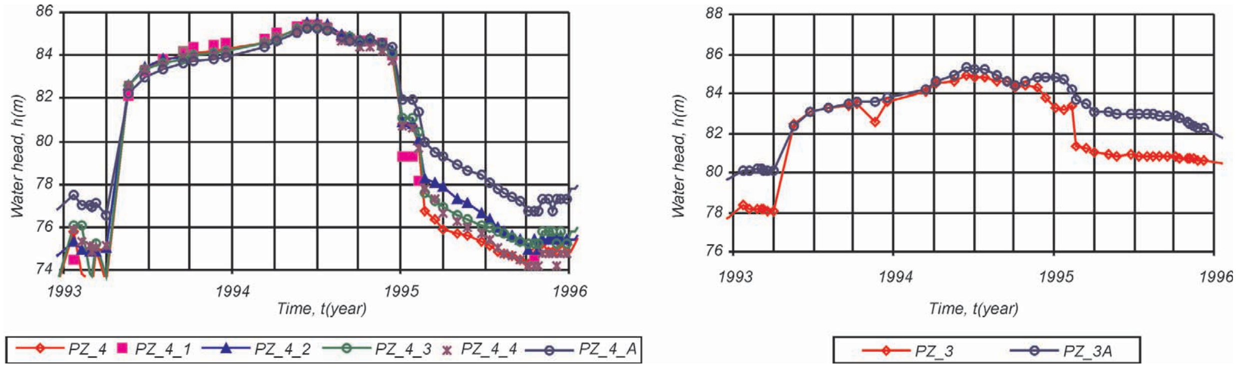

Figure 17: The relationship between the water level in the captured aquifer and the source aquitard layer during the cessation of pumping (PZ-3A and PZ-4A are piezometers with screens in the covering aquitard layer).

anisotropy of aquifer, which accentuates the importance of the casing design of wells and piezometers.

It is interesting to note that the above described effects of anisotropy disappear after the 20-month break in pumping, and reappear six moths after it is restarted (Figs. 16, 18, & 19).

6. CONCLUSION

The effect of the infiltration of rainfall on deep groundwater recharge can be precisely identified only on the basis of data from long term pumping. In spite of this, boundary conditions that enable such identification are detectable even from data collected during relatively short term experimental pumping of wells. Drainage from leaky layers, (in the case of leaky aquifers), slows down the progress of water level drawdown round a pumped well. This occurs to the extent that on the basis of the spatial pattern of drawdown progressing as a function of the logarithm of time, all the hydraulic parameters needed for long term prognosis can be determined, even though drawdown through puming is only slightly higher than 50% of the maximum drawdown. The method of this kind of interpretation of experimental pumping has been demonstrated for the Osijek, Vinogradi well field that captures a heterogeneous leaky aquifer. It is the heterogeneity of the layers themselves that cause the contrasting effects of pumping on the water level pattern within the captured aquifer and raise doubts about ground water recharge. It can be concluded that calibration and validation of the numerical mathematical model of the aquifer system have unquestionably confirmed that the condition of the aquifer, controlled by the existing monitoring, can only be performed with effective groundwater recharging. As this verification relates to a 10-year period where the average annual rainfall was 620 mm, (a quantity ~ 5% less than the 30-year average), the identified mean annual aquifer recharge of 96 mm can be regarded as a regional example of similar hydrogeologic relations and climate conditions.

Groundwater is recharged by rainfall infiltration and that process hides the effect of the particular rainfall. Those particular rainfalls, when being integrated have some effect on trends, while main divergence in groundwater levels are caused by pumping regim of the whole well field.

The change in pumped quantity of water from some layers of the aquifer after the closure of some screens resulted in a change in the composition of calcium, sodium, iron,

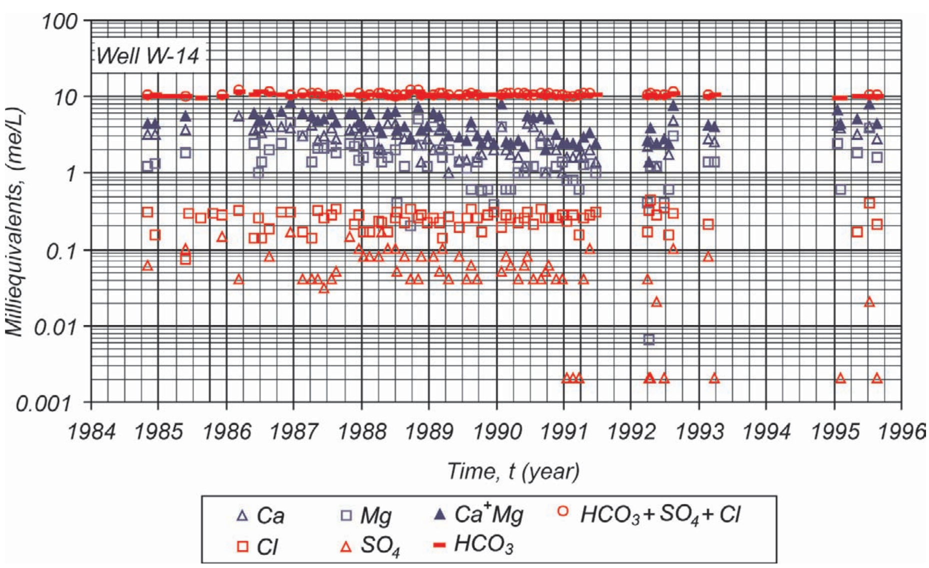

Figure 18: Primary cations and anions in water from well W-14.

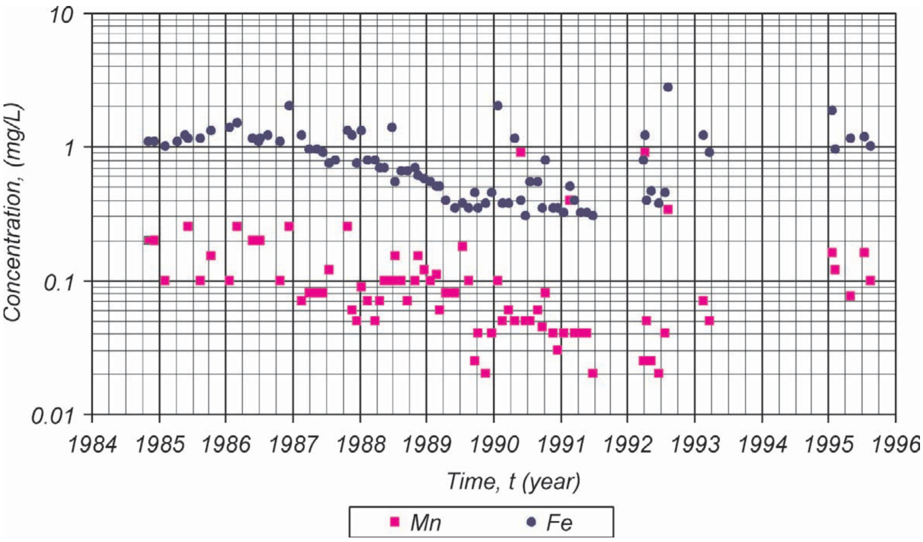

Figure 19: Amount of Fe and Mn in the water at well W-14.

manganese and possibly some other components in the water, which reflects the diversity of the sedimentary conditions of the aquifer sands and the mechanism of groundwater recharge. This is an additional reason for the construction of appropriate nested piezometers in the development of groundwater monitoring.

ACKNOWLEDGEMENT

We would like to thank the reviewers for their comments and suggestions that have improved this manuscript. This work was supported by Ministry of Science, Education and Sports of the Republic of Croatia, Croatian Waters and Vodovod Osijek.

REFERENCES

BOONSTRA, J. & RIDDER, N.A. (1981): Numerical modeling of groundwater basins.− ILRI, Pub., 29, Wageningen, 226 p.

COOPER, H.H.jr & JACOB, C.E (1946): A Generalized Graphical Method for Evaluating Formation Constants and Summarizing Well-Field History.− Trans. Amer. Geophys. Union, 27, 526−534.

DE VRIES, J.J. & SIMMERS, I. (2002): Groundwater recharge An overwiev of processes and challenges.− Hydrogeology J., 10/1, 5−17.

HANTUSH, M.S. & JACOB, C.E. (1955): Non-steady radial flow in infinite leaky aquifers.− Trans. Amer. Geophys. Union, 36, 95−100.

HANTUSH, M.S. (1956): Analysis of data from pumping tests in leaky aquifers.− Trans. Amer. Geophys. Union, 37, 702−714.

HANTUSH, M.S. (1960): Modification of the theory of leaky aquifers.− J. Geophys. Res., 65, 3713−3725.

HEALY, R.W. & COOK, P.G. (2002): Using groundwater levels to estimate recharge.− Hydrogeology J., 10/1, 91−109.

SCALON, B.R., HEALY, R.W. & COOK, P.G. (2002): Choosing appropriate techniques for quantifying groundwater recharge.− Hydrogeology J., 10/1, 18−39.

STRAFACE, S. (2009): Estimation transmissivity an storage coefficient by means of a derivative method using the early time drawdown.− Hydrogeol J., 17/7, 1679−1686.

TADIC, Z., URUMOVIC, K., TADIC, L. & GERES, D. (1994): Sustainable groundwater use. XVIIth Conference of the Danube Countries on Hydrological Forecasting and Hydrological Bases of Water Management.− Proceedings, Budapest, 493−497.

TOTH, J. (1995): Hydraulic continuity in large sedimentary basins.− Hydrogeology J., 3/4, 4−16.

URUMOVIĆ, K. (1978): Graphic method of determination of the leaky aquifer by the standard distance-slope curve. Symposium on Investigation and Economy of the Underground waters.− Working papers, B, Zagreb, 23−30.

URUMOVIĆ, K. (1982): Hidrogeološke značajke istočnog dijela dravske potoline.− Unpubl. PhD Thesis, University of Belgrad, 138 p.

URUMOVIĆ, K., HERNITZ, Z. & ŠIMON, J. (1978): O kvartarnim naslagama istočne Posavine (SR Hrvatska).− Geol. vjesnik, 30/1, Zagreb, 297−304.

URUMOVIĆ, K., HERNITZ, Z., ŠIMON, J. & VELIĆ, J. (1976): O propusnom mediju kvartarnih, te gornjo- i srednjo-pliocenskih naslaga sjeverne Hrvatske.− IV jug. simp. o hidrogeol. i inž. geol., 2, Skopje, 395−410.

URUMOVIĆ, K., HLEVNJAK, B. & GOLD, H. (1996): Replenishment of the groundwater from precipitation in Osijek area. XVIIIth Conference of the Danube Countries on Hydrological Forecasting and Hydrological Bases of Water Management.− Proceedings, A45−A50, Graz.

Manuscript received July 19, 2010
Revised manuscript accepted December 28, 2010
Available online February 22, 2011