15 Reconstruction of Water Level Time Series in an Aquifer Using Geostatistical Technique

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INTRODUCTION

Water level is the only parameter that is measured from the aquifer directly and also depicts the dynamics of the aquifer. It is measured normally in static condition but in spite of all precautions taken while measuring the water level there are data gaps. In aquifer modelling, particularly in the calibration of the model in transient condition, a sufficiently long time series of water level is required on all the wells that are used as control points. So to deal with two different situations that are to have a sufficiently long time series of water level at any well, geostatistics was applied to fill the data gap, at the time when the water level could not be measured. With this we obtained a mixed time series of water level at any well containing measured water levels without any error term and the estimated water level with estimation errors at uniform frequency.

A geostatistical technique from the theory of regionalized variables was first documented by Matheron (1963). The theory of regionalized variables very well applied to groundwater hydrology are by Delhomme (1978), Gambolati and Volpi (1979), Marsily (1986), Isaaks and Srivastava (1989) and Kitanidis (1997) and many others. This theory of regionalized variables has been applied here in analyzing time variant parameter, the water level data. As this parameter is very important in understanding the dynamics as well as stress-strain behaviour of the aquifer, it is of utmost importance that it should be available in adequate interval of time. This was used during calibration of the aquifer model.

In the present study, water level data from IFP wells (Fig. 1) was used from November 2000 to August 2003. The water level was monitored from

25 wells every month during the first week in one day. Due to unavoidable circumstances few wells were not monitored during some months. By this process after three years there were data gaps for some of the wells during various months (Fig. 2). But the water level had to be known for those wells for all months to have a complete time series of water level. The data gaps from January 2000 to October 2000 were completed by geostatistical estimation using unbiased kriging estimation variance (Best Linear Unbiased Estimator BLUE, Kitanidis, 1997). After the estimation of the water level was made along with variance of estimation, it was plotted with complete data set along with estimation variance and the water level now available from January 2000 to August 2003.

STUDY AREA

The study area, Maheshwaram watershed, falls between the geographical coordinate 17° 06' 20" to 17° 11' 00" North latitude and 78° 24' 30" to 78° 29' 00" East longitude and is situated in Ranga Reddy district at about 30 km south of Hyderabad, Andhra Pradesh, India and covering an area of 55 km^2, is a typical watershed in a granitic terrain with outcrops, structural features and fissure systems (Fig. 1). The average annual rainfall of the area

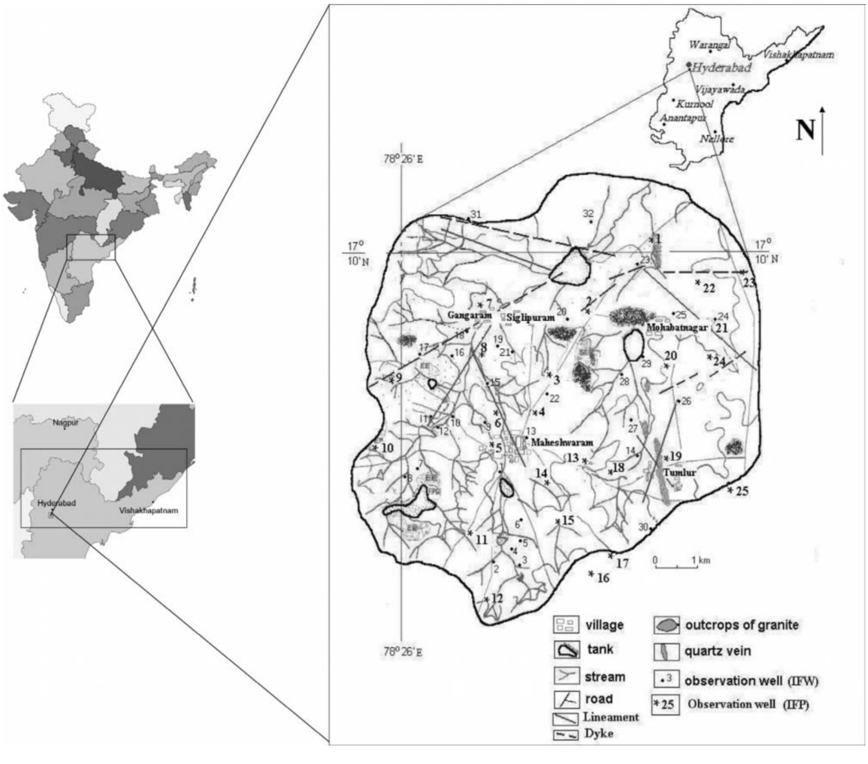

Figure 1. Study area and location of wells.

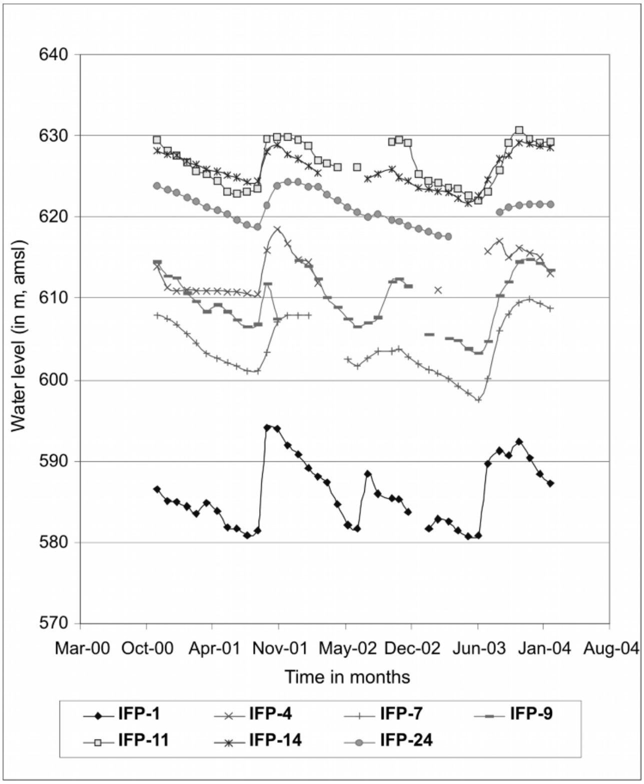

Figure 2. Hydrographs of observation wells showing data gap.

is 750 mm. The area lies at an elevation of 600 – 670 m above mean sea level. The area consists of hard crystalline rocks mainly pink and grey granites of Archaean age with variable density of fractures within the subsurface. Major part of the basin consists of biotite granite and some part of it consists of leucogranite. Foliation of the granites is traced on outcrops of granite. Granites with pegmatite show higher density of fractures/fissures. The area consists of dendritic drainage pattern. Major part of the basin is covered with pediplain having shallow weathering. Soil consisting of clay loam, red loam and sandy loam with variable thicknesses (0-1 m) form the top layer (Hashimi and Engerrand, 1999 and Krishnamurthy et al., 2000). Moderate thicknesses of weathered rocks are present and thus a two-tier system viz., weathered and fractured aquifers co-exist in the entire area. Due to over-exploitation of the groundwater resources, the water level has gone down and presently the groundwater occurrence is mainly in the fissured rock under unconfined conditions (Subrahmanyam et al., 2000; Maréchal et al., 2004).

Hydrogeology

Mostly horizontal fractures were prominent but vertical fractures are also contributing as a conduit for groundwater flow. Permeability in the fissured layer is greater than the laminated layer. In general the horizontal permeability is greater than the vertical permeability. Groundwater occurs mostly in fractured zone with water table varying between 11 and 20 m below ground level. The groundwater is mostly tapped by means of bore-wells in the fractured rock and the yield ranges between 1000-5000 gph (Subrahmanyam et al., 2000). High yields of some of the bore-wells are due to encountering of water bearing fractures at depths. Four sets of joints were mapped. The most prominent joint set strikes in N-S direction based on rose diagram plot. The strike varies between N10° E to N10° W. The joints are either vertical or dip very steeply towards west with angles 70° to 75°. The second set of joints strikes E-W with steep dips. The strike of this joint set varies between N80°W and N80°E. The third set of joints strike NE-SW direction with steep dips of 60°-75° towards SE. The fourth set of joints is more or less parallel to the surface of the topography. All these joints and its system help in the percolation of rainwater from the surface to the groundwater table. They also help to act as conduits for the transmission of groundwater. There are two sets of dykes and the strike of the first set is 90°-110° and the second set strikes 50° to 60°. There are joints and fractures in the dykes also but the surfaces are curved. These fracture also act as conduits for the transmission of surface water to the groundwater. The area is also traversed by lineaments, identified through the study of aerial photographs and land-sat imageries. The high yielding wells are aligned along these lineaments.

1-D Variogram of Water Level in Time

In the present study using the water level data from 25 experimental observation wells, the 1-D variogram analysis using Ahmed (1995, 2001) was carried out individually for each well to obtain the water level variability in time. The variogram was calculated and fitted with the theoretical variogram for all the monitoring wells (IFP's).

$$\gamma(\underline{d}) = \frac{1}{2N_{\mathrm{d}}} \sum_{i=1}^{N_{\mathrm{d}}} [h(t_{\mathrm{i}} + \hat{d}) - h(t_{\mathrm{i}})]^2 \tag{1}$$

where

$$d - \Delta d \le \hat{d} \le d + \Delta d \tag{2}$$

with

$$d = \frac{1}{N_{\mathrm{d}}} \sum_{i=1}^{N_{\mathrm{d}}} \hat{d}_{\mathrm{i}} \tag{3}$$

where $h(t)$ is water level, d is the initially chosen time interval between two measurement periods; t and $(t + d)$ with Δd as tolerance on time interval. d is actual time for the corresponding calculated variogram. N_d is the number of pairs for a particular time interval. The additional equation 3 avoids the rounding-off error of pre-decided time intervals (only multiples of the initial time is taken in conventional cases). It is very important to account for every term carefully while calculating variograms.

WATER LEVEL ESTIMATION IN TIME

Method of ordinary kriging was used for geostatistical estimation by Kitanidis (1997). Given the monsoon climate in the area and if there is no heavy pumping, the water level time series could be assumed stationary. The estimation were made at the month(s) for each well where the water level was not monitored due to some unavoidable cause e.g., due to the pumping in the nearby well. Thus the water level was estimated using the existing observed data and the corresponding best fitted variogram's model parameters. The estimation was made by ordinary kriging using equations 4-7 which gives the estimated water level along with the variance of the estimation error.

METHODOLOGY APPLIED

The theory of regionalized variable (Matheron, 1971) was used to analyze the water level data. The Ordinary Kriging equation used are as follows:

$$h^*(t_o) = \sum_{i=1}^{n} \lambda_i h(t_i) \tag{4}$$

$$\sum_{j=1}^{n} \lambda_j \gamma(t_i, t_j) + \mu = \gamma(t_i, t_o), \; i = 1, \ldots\ldots\ldots n \tag{5}$$

$$\sum_{i=1}^{n} \lambda_i = 1 \tag{6}$$

$$\sigma^2_{(t_o)} = \sum_{i=1}^{n} \lambda_i \gamma(t_i, t_o) + \mu \tag{7}$$

where $h^*(t_o)$ is the estimated value of water level at the point t_o and t_o is the particular time. The λ, γ, σ^2_o and μ are the kriging weight, variogram, kriging variance and Lagrange multiplier respectively.

RESULTS AND DISCUSSIONS

The estimated water level was combined with the existing (observed) water level data and the plot was made to view the trend of the water level fluctuations. The observed and the estimated water level were plotted along with the standard deviation value at the estimated time. As an example two cases are shown (Figs 3 and 4). In the beginning the water level data was not available during the month of January 2000 to October 2000 for all the 25 wells as the wells were newly drilled and the regular monitoring was started from November 2000 and thus the estimated water level for this period shows higher standard deviation compared with the in-between estimated water level. In-between estimated water level for various wells for certain months shows good agreement with the observed water level and they follow the trend of the observed water levels. During aquifer modeling the output of the model is water level which has to be matched with the observed water levels and any deviation in the water level trend or abnormal behaviour of water level fluctuation can be justified by calibrating the model. These results were very useful during calibration of the numerical aquifer model in transient state with the updated data till July 2003 as the calibration

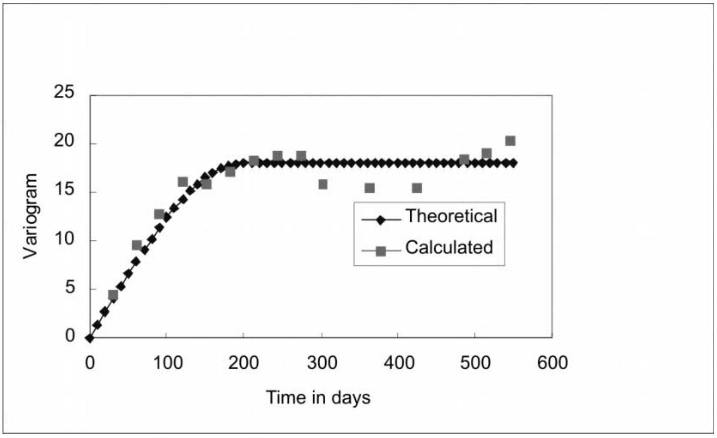

Figure 3a. 1-D variogram of water level for well IFP-1.

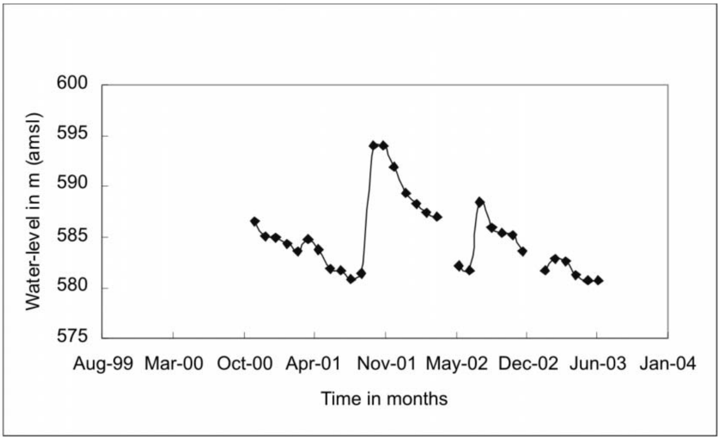

Figure 3b. Observed water level showing data gap for well IFP-1.

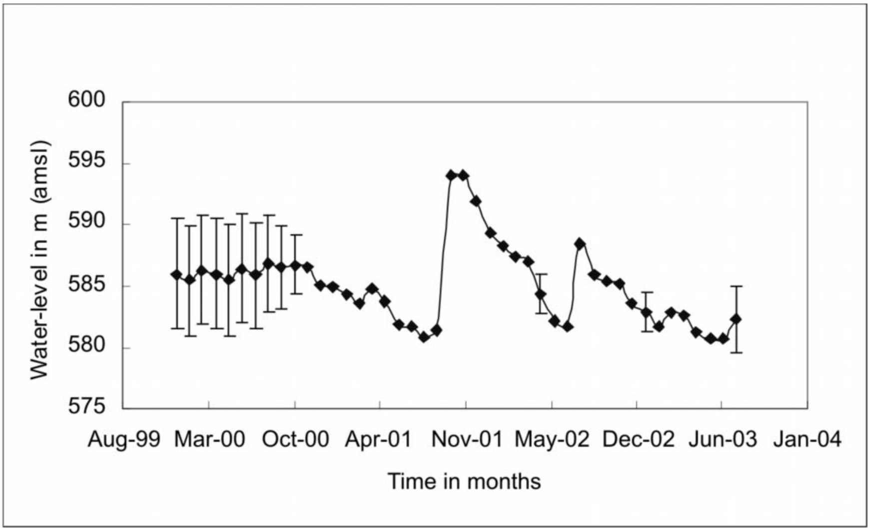

Figure 3c. Observed and estimated water level with standard deviation for well IFP-1.

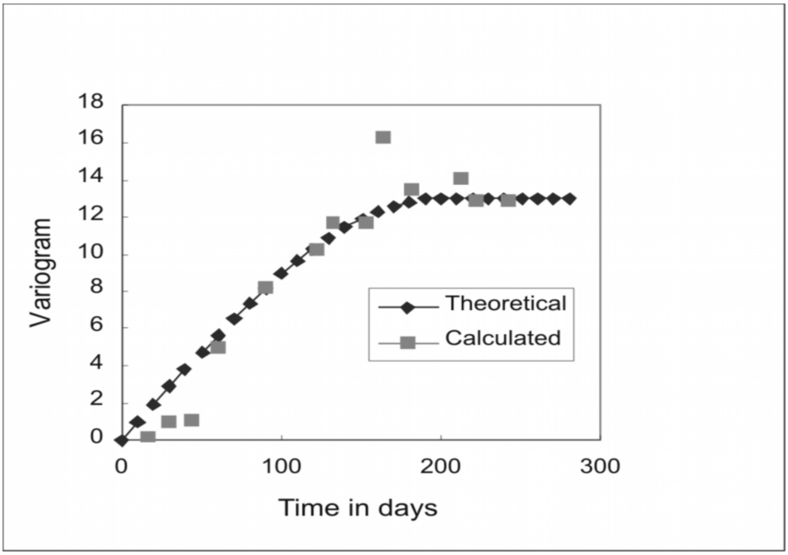

Figure 4a. 1-D variogram of water level for well IFP-9.

of the aquifer model in study area was completed up to July 2003 (Kumar, 2004). Thus these estimated water levels have helped for a reliable model

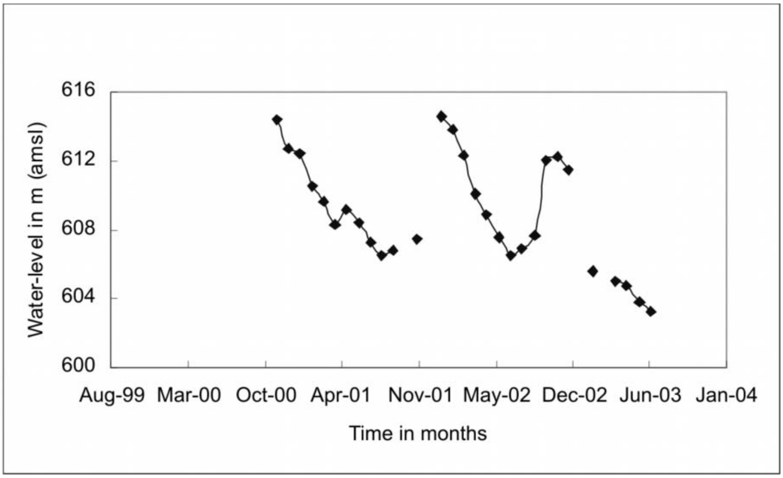

Figure 4b. Observed water level showing data gap for well IFP-9.

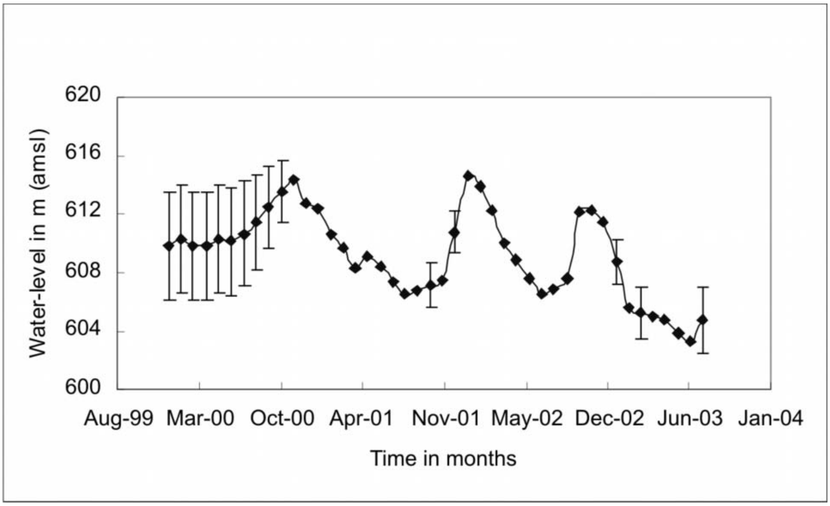

Figure 4c. Observed and estimated water level with standard deviation for well IFP-9.

calibration. Figure 5 shows the estimated water levels at three time steps well matched (very close to each other) with the calibrated water levels obtained from the aquifer model for well IFP-1. The estimation error bar helps in addition and suggests what percentage of various aquifer parameters should be increased or decreased in the aquifer model accordingly and also by keeping in mind the well condition in the field and its influence in the model. Similarly, Fig. 6 depicts that the estimated water levels are matched with the calibrated one but at one time step they are deviating. There could be many reasons for this type of situation.

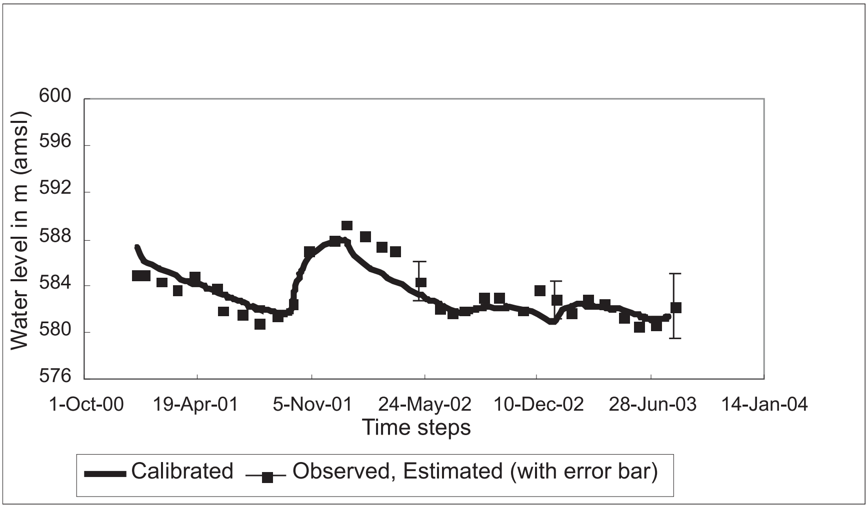

Figure 5. Calibrated, observed and estimated water levels with estimation error bar for well IFP-1.

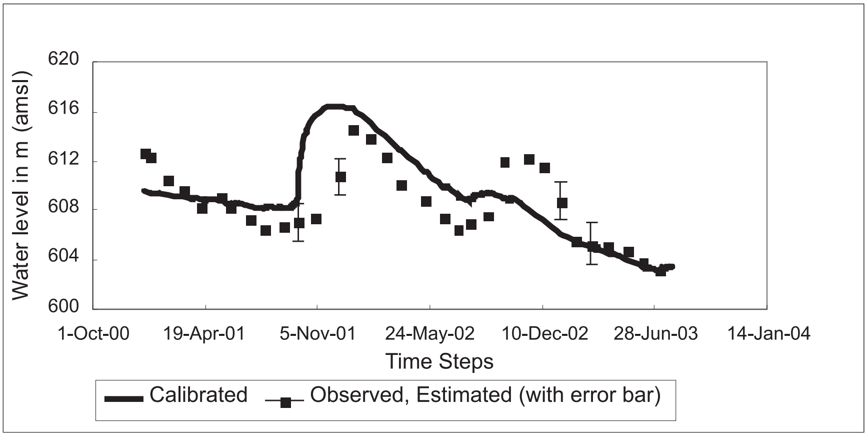

Figure 6. Calibrated, observed and estimated water levels with estimation error bar for well IFP-9.

CONCLUSIONS

Reconstructing of the time series of water levels in all the control wells has added value to the existing data sets. These estimated water levels value are plotted along with the variance of the estimation error (in the form of $\pm 2\sigma$, σ being the standard deviation of the estimated value) both in the form of upper and lower bound and with the observed water levels without error bar in the series as hydrographs. This study also helped or justified in an unbiased model calibration as the simulated hydraulic head obtained from aquifer model could be verified against the value or a range that was absent otherwise. The plot of the error permits to check that the simulated heads are within the permissible limits. The water levels being non-stationary in nature, in general, are complicated applying geostatistical estimation in space. However, this problem is avoided in 1-D time domain as the same parameter shows recurring nature and until and unless extensive recharge or pumping conditions occur, the water level time series are stationary and provide a bounded variogram. Also the aquifer modelling provides water level as output at any time intervals often very close and theoretically the time interval should be kept close. However, in practice it is just not possible to monitor water levels at close intervals in the field and thus the present study provides a solution to match the two conditions.

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