Hydrological Sciences Journal
Publication details, including instructions for authors and subscription information:
http://www.tandfonline.com/loi/thsj20

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Published online: 17 Sep 2013.

To cite this article: Marcelo Varni, Rocío Comas, Pablo Weinzettel & Sebastián Dietrich , Hydrological Sciences Journal (2013): Application of the water table fluctuation method to characterize groundwater recharge in the Pampa plain, Argentina, Hydrological Sciences Journal, DOI: 10.1080/02626667.2013.833663

To link to this article: http://dx.doi.org/10.1080/02626667.2013.833663 http://dx.doi.org/10.1080/02626667.2013.833663

Application of the water table fluctuation method to characterize groundwater recharge in the Pampa plain, Argentina

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Received 8 September 2011; accepted 25 January 2013; open for discussion until 1 April 2014

Editor D. Koutsoyiannis

Citation Varni, M., Comas, R., Weinzettel, P., and Dietrich, S., 2013. Application of water table fluctuation method to characterize the groundwater recharge in the Pampa plain, Argentina. *Hydrological Sciences Journal*, 58 (7), 1–11.

Abstract The water table fluctuation (WTF) method is based on accepting that rises of a water table are due to recharge water reaching the groundwater. To apply the method, an estimate of the specific yield of the zone of fluctuation of the groundwater level is required. In this paper, a method for estimation of the specific yield (S_y) is proposed; it consists of a graphical procedure which relates rises in groundwater level to the precipitation from which they originated. The method presents more reliable S_y values as the number of events measured increases. Eighteen years of daily measurements were analysed to obtain a S_y value of 0.09, which was used to apply the WTF method. The obtained recharge values show consistency with values calculated by other authors for the same region.

Key words groundwater recharge; water table fluctuation method; specific yield; plain

Application de la méthode de fluctuation du niveau piézométrique pour caractériser la recharge des eaux souterraines dans la plaine de la Pampa (Argentine)

Résumé La méthode de fluctuation du niveau piézométrique (FNP) est basée sur l'hypothèse que l'augmentation du niveau piézométrique d'une nappe est dû à la recharge atteignant la nappe. Pour appliquer cette méthode, il est nécessaire d'estimer la porosité efficace dans la zone de fluctuation de la nappe. Dans cet article, nous proposons, pour estimer la porosité efficace S_y, une méthode graphique reliant l'augmentation du niveau de l'eau souterraine à la précipitation dont elle procède. Les valeurs de S_y estimées seront d'autant plus fiables que de nombreux évènements seront pris en compte. Nous avons utilisé dix-huit ans de mesures quotidiennes qui nous ont permis d'estimer à 0,09 la valeur de S_y, que nous avons utilisée dans l'application de la méthode FNP. Les valeurs de recharge obtenues sont comparables aux valeurs calculées par d'autres auteurs pour la même région.

Mots clefs recharge des eaux souterraines; méthode de fluctuation du niveau piézométrique; porosité efficace; plaine

1 INTRODUCTION

Economically, the humid pampas of Argentina (Fig. 1) is the most important Argentine territory, because its climate and soil characteristics mean it is an excellent agricultural and livestock area. In this paper, we make estimates of recharge to the central area of the River Azul aquifer, and the relationship between surface water and groundwater are analysed. Both issues are of great importance to the management of water resources in the region.

Groundwater recharge was estimated using the water table fluctuation (WTF) method at the University Campus in Azul, in central Buenos Aires Province, Argentina (Fig. 1). The WTF method (Healy and Cook 2002) is applicable to unconfined

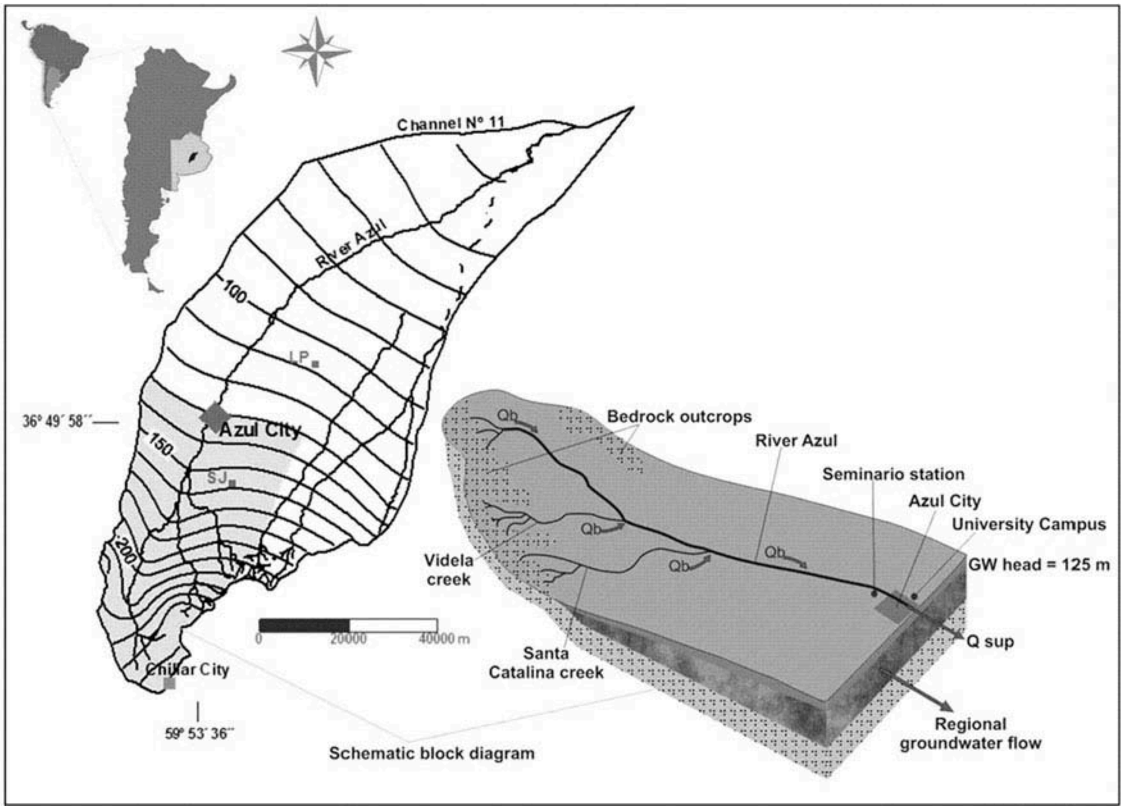

Fig. 1 Left: location map with equipotential lines (m a.s.l.) and new stations with data loggers (SJ and LP). Right: schematic view of the study area (Q_b: baseflow; Q_{sup}: fluvial flow).

aquifers. The method is simple and easy to apply (Delin *et al*. 2007) and based on accepting that rises of the water table are due to recharge water reaching the water table. The equation applied is (Healy and Cook 2002):

$$R = \Delta S_{GW} = S_y \frac{dh}{dt} = S_y \frac{\Delta h}{\Delta t} \quad (1)$$

where R is the groundwater recharge, ΔS_{GW} the groundwater storage variation, h is the water level and t the time.

To apply the WTF method, an estimation of the specific yield (S_y) at the depth of water table fluctuation is required. The S_y is defined as (Healy 2010):

$$S_y = \varphi - S_r \quad (2)$$

where φ is porosity and S_r is specific retention. Methods commonly used to determine S_y are laboratory methods, aquifer tests, water-budget methods and water table response to recharge. Laboratory methods usually determine S_y by measurement of porosity and specific retention, and application of equation (2). A large variability exists in both laboratory- and field-determined S_y values. Aquifer tests provide measurements of S_y integrated over relatively large areas, depending on the distance between observation wells and pumping well. The water-budget method, proposed by Walton (1970), requires a water balance at the basin scale, preferably during winter (when evapotranspiration is minimal) and the soil is near saturation. The balance can be solved for the change in groundwater storage and, knowing the water table variation, S_y can be estimated. The water budget equation can be written as:

$$P = \mathrm{ET} + \Delta S + Q_{off} \quad (3)$$

where P is the precipitation, ET is the evapotranspiration, ΔS the change in water storage (in surface reservoirs, the unsaturated zone and the saturated zone) and Q_{off} is surface plus subsurface water flow out of the basin. The change in groundwater storage can be written as a function of S_y (see equation (1)) and then, rearranging equation (3), S_y can be obtained. The method may involve large errors, since the variation of groundwater storage must be estimated from the calculation of the difference of other terms of the water budget involving much larger water volumes. However, this approach allows an average value for a large area rather than a local value to be obtained.

The response of water table to recharge method consists of calculating the ratio of water table rise to total rainfall (Risser *et al*. 2005b) for all the registered events in the analysed area. The height of water table rise measured after a rainfall event provides an estimate of the amount of open pore space available in the unsaturated zone (i.e. S_y). The method is valid for a shallow water table.

Crosbie *et al*. (2005) tested different methods to estimate S_y at two sites, determining the water retention curve in the laboratory, through the analysis of the response of the water table and by pumping tests. The results, after testing several methods, show that the rainfall–water table response seems to provide the best estimates.

Therefore, in this work, a method that uses field measurements is proposed (rainfall and water level variations), which is a variant of the water table response to recharge events approach. The method is a graphical procedure that provides more accurate results the greater the number of events measured. As there are extensive data records for the central River Azul aquifer, a substantial number of water table rise events can be analysed and so the S_y determination is less uncertain. Moreover, the records are especially suitable for this analysis given the fast response of the water table during recharge events. For these reasons, the graphical procedure proposed in this article was chosen for estimation of the S_y.

As stated above, the recharge was estimated by the WTF method and applied to 18 years of daily water level data in central Buenos Aires Province, Argentina. As the method has a good temporal discrimination of water level rise, monthly and annual recharge were obtained. The recharge estimates were related, through regression equations, with other variables of the water balance of the basin, such as precipitation, baseflow and water table level.

2 STUDY AREA

The study basin consists of a low hilly area to the south (upper area), where the River Azul headwaters occur, and a large plain (depressed area) to the north. The River Azul drains to the River Salado from the northeast of the basin. The transition area between the hilly and the depressed landforms has a gently rolling relief. Surface slopes average 5% for the upper area, 0.05–0.1% for the lower area and 0.2% for the middle area.

The basin is underlain by Quaternary-age sediments. These are silts, sandy silts, and clayey silts overlain by fine sands and silts of aeolian origin, reworked by fluvial processes (Fidalgo *et al*. 1975). Bedrock outcrops occur in the upper area and consist of metamorphic rocks, granite, tonalite, migmatite and quartzite (González Bonorino *et al.* 1956). These rocks form the lower boundary of the aquifer and occur at a depth of 120 m in Azul city where the water table measurements used in this work were made (36° 46′ S, 59° 53′ W) (Fig. 1). The soil in the area of the measured well is a petrocalcic Paleudol according to the US Soil Taxonomy (Soil Survey Division Staff 1993).

The mean annual precipitation in Azul is 908 mm (1901–2009). The maximum monthly mean precipitation (137 mm) occurs in March, while the minimum (43 mm) occurs in June. The mean annual temperature is 14.5°C, with a maximum monthly mean in January of 21.4°C and lowest monthly mean in July, 7.7°C (1966–2009).

Figure 1 shows schematically the water supply area of Azul city: the rock outcrop areas and the deepening of the basement to the north, ending in a transverse groundwater flow section that corresponds approximately to the 125 m a.s.l. equipotential line. The arrows represent water drained by the stream (Q_{sup}) and by the aquifer (regional groundwater flow). The recharge to the phreatic aquifer that occurs in the area contributes to the streamflow through the baseflow (Q_b) (upper groundwater flow), to the regional groundwater flow (deeper groundwater flow), to direct evapotranspiration losses (variable in space and time) and pumping extractions. Naturally, Q_{sup} includes the contributions of pure surface water plus Q_b. The block diagram indicates the relationship between the bottom of the aquifer and water table depth (120 m and 5 m b.g.l., respectively). Under these conditions, the annual water table variation affects mainly the baseflow contribution, and the regional groundwater flow remains approximately constant.

The daily measurement of the water table allows the analysis of aquifer recharge events. The campus of Azul University is located in a rural area near Azul city, where there is no pumping, no nearby industries and land use surrounding the campus is pasture where horses graze. Wells that extract groundwater for drinking water supply are located over 800 m away, on the other side of the River Azul, which limits the area of influence of the abstraction wells. The campus is supplied with water from there as the distribution network crosses the river. Sewage from the campus is also integrated with the city network, and is treated and discharged into the river downstream of the city. Therefore, it can be argued that the studied well is not influenced by extraction or discharge of human origin. Pumping extractions can be grouped into those concentrated in Azul and those conducted in rural areas. In Azul, about 12 hm^3/year are extracted, while there are no pumping records for existing wells in rural areas. The great majority of the agriculture in the basin is rainfed due to the relatively even rain distribution. However, irrigation, mainly of potatoes and corn, takes place in some parts to ensure high crop yields. Despite its recent introduction, most water extracted in rural areas is used for irrigation; it is all obtained by groundwater extraction. The few surface streams are insufficient to satisfy the irrigation demand, even for adjoining lands. The irrigation needs were estimated, assuming an annual value of 80 mm/year (the irrigation is complementary) and an irrigated area of approx. 10 000 ha (the two values are upper limits), to be about 7% of the groundwater recharge in the Azul basin above Azul city. In the event of irrigation water coming from sources outside the basin, a method such as that proposed by Efstratiadis *et al*. (2008) could be used. The studied well is located far from irrigation wells and so there should be no influence on the levels analysed.

3 METHOD

The groundwater levels analysed (Fig. 2) comprise a record of 18 years of daily levels registered in a shallow well (7.7 m deep) on the Azul University Campus at the centre of the Azul River basin (Fig. 1). The levels were measured daily on five days per week (Monday to Friday), and always at the same time. If a recharge event was taking place (rising levels), levels were also recorded at the weekends. The levels were measured with a Seba water level contact dipmeter KLL (Seba Hydrometrie GmbH). During the period, the water table depth ranged between 0.63 and 6.25 m b.g.l. (Fig. 2).

The WTF method is based on the recharge effect, i.e. the rise of the water table due to previous rainfall. Careful analysis of the records enables the exclusion of variations in the water table due to fluctuations in climate and anthropogenic activities (pumping, irrigation, land use, etc.) (Healy and Cook 2002).

The WTF method was used in this study because frequent and accurate water level measurements were available in a well that is in an area unaffected by human activities, and because the method is very useful for detecting temporal recharge variations. Data to allow application of another method of recharge estimation for the 18 years were not available. Data were available that allowed use of alternative methods for shorter periods, and the results were compared for the common time periods.

The main challenge of the WTF method involves the estimation of the specific yield (S_y) of the aquifer at the depth of the zone of water table fluctuation.

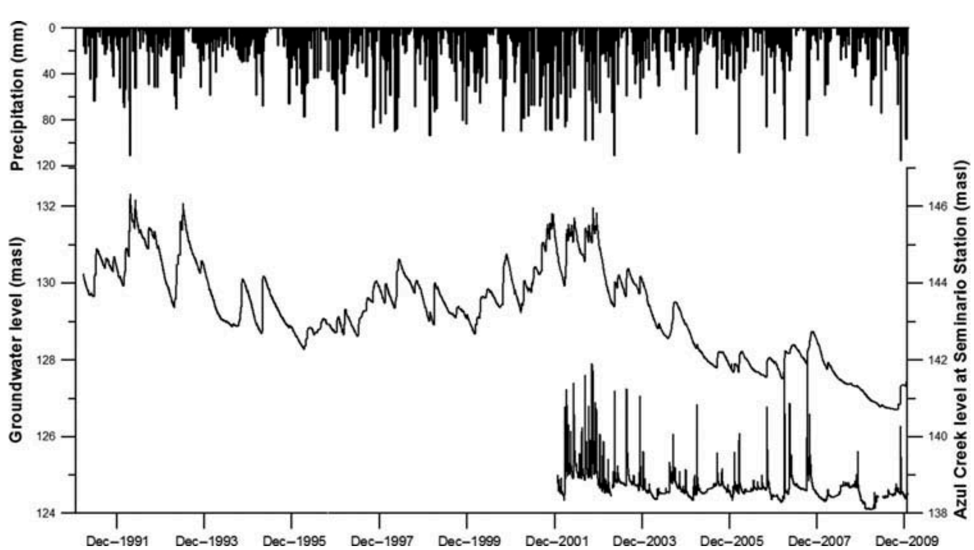

Fig. 2 Precipitation and groundwater level at the University Campus and River Azul level at Seminario Station.

To estimate S_y, a graphical procedure was proposed: a graph of rainfall values is plotted against water table rises. The inverse of the slope of a line drawn through the origin to just above all of the measured points gives the maximum value of the specific yield. This S_y estimation method was chosen because it provides a robust S_y value due to the long data record (18 years). This method is equivalent to calculating the precipitation–groundwater rise ratios of all measured events and selecting the smallest value, but the graphical method has the advantage of clearly displaying whether there are several points near the minimum value of S_y, or if there is one point far from the rest that needs to be revised, etc.

Methodologies such as Sophocleous's (1991), that combine a Darcian soil water balance with the WTF method, could have improved the estimation, but unfortunately there is no historical information (neutron probe and tensiometer data) to do so. The total annual declines in groundwater level were calculated and the relationship between these and the mean annual groundwater level were analysed. Relationships between recharge and other basin budget variables were also analysed.

As background, the following studies can be mentioned: Nichols and Verry (1981) analysed the relationships between recharge, precipitation and river runoff; they found that recharge varies linearly with precipitation and can explain up to 93–96% of the variations in recharge and river runoff by applying multiple regression using variables such as seasonal precipitation, summer and autumn precipitation of the previous year, and non-winter temperature as independent variables. The annual recharge values are related to the annual precipitation and the mean annual depth of the water table through simple and multiple correlations. Examples of similar relationships can be found in: Rabinowitz *et al.* (1977) (recharge–precipitation); Sophocleous (1992) (recharge with: annual precipitation, spring precipitation, depth to water table, soil profile water storage, number of spring precipitation days, number of precipitation days during the year and spring precipitation rate); Taylor and Howard (1995) (recharge–precipitation); Sami and Hughes (1996) (recharge–precipitation); and Scanlon *et al.* (2006) (recharge–precipitation and irrigation recharge–precipitation plus irrigation).

The relationship between groundwater recharge and nearby stream baseflow is described in this paper. Baseflow is sometimes used as an approximation of recharge when underflow (the flow of groundwater beneath and so bypassing a stream), evapotranspiration from riparian vegetation, and other losses of groundwater from the watershed are thought to be minimal (Combalicer *et al*. 2008, Healy 2010). The major assumptions in using baseflow for estimating recharge are that baseflow equals groundwater discharge and that groundwater discharge is approximately equal to recharge (Risser *et al*. 2005a). In the Azul basin, the underflow is not negligible and, therefore, the baseflow cannot be equated to recharge. Thus, we are interested in exploring other relationships between recharge and baseflow. The baseflow is obtained from hydrographs measured at the Seminario station (Fig. 1), located a few kilometres upstream of Azul city, using graphical hydrograph separation (Linsley *et al*. 1982).

4 RESULTS AND DISCUSSION

4.1 Specific yield estimation

Figure 3 shows the relationship between rainfall and groundwater level rise for individual recharge events. The drawn line corresponds to $S_y = 0.09$. This number is a maximum value because, if all the rainfall–recharge events (dots in the graph) occurred when the soil and vadose zone profile were not at field capacity, this value could be lower. Likewise, the fraction of rainfall that produces runoff, which has not been taken into account, has the same bearing. Weinzettel *et al.* (2005) obtained a value of 0.07 at the Azul University Campus by considering the water level response towards selected rainfall events.

The S_y value obtained was compared with the results of a new method based on the correlation

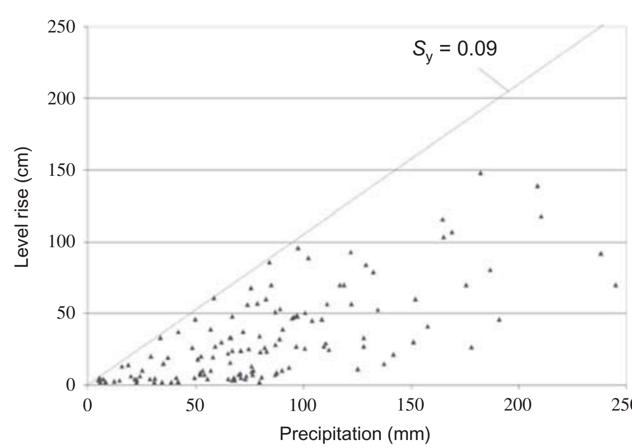

Fig. 3 Determination of specific yield using the envelope straight line of the precipitation–groundwater rise points of each registered rainfall–recharge event.

between satellite measurements from the GRACE (Gravity Recovery and Climate Experiment) mission and water table variations. To estimate the specific yield, a theoretical linear relationship between satellite data and water table fluctuation is defined; the available data verified this relationship and, thus, the specific yield could be estimated by a simple linear regression model. The technique was tested in Azul using a 2-year-long series of data. The estimated value of S_y is approximately 0.09 (Guarracino *et al.* 2011).

These values are within the range of S_y for silty sediments (loess-like) (see Fetter 2001 p.79 Table 3.5, Healy 2010 p.125 Table 6.2).

4.2 Annual recharge

The following is an estimate of groundwater recharge considering the S_y values of 0.07 (or 7%) and 0.09 (or 9%) mentioned in the previous section. The annual recharge can be obtained from the annual cumulated individual rises of groundwater level. Thus, Fig. 4 is obtained by plotting annual recharge ($S_y = 0.09$) *versus* the corresponding rainfall with an adjustment of $r^2 = 0.69$ and the Nash-Sutcliffe efficiency measure (Nash and Sutcliffe 1970) NSEM = 0.69. The intersection of the adjusted function with the abscissa axis indicates the precipitation for which there would be no recharge, i.e. about 480 mm. This value is only indicative of a general trend, since recharge depends greatly on rainfall distribution through time, but it does provide a reference value for the area.

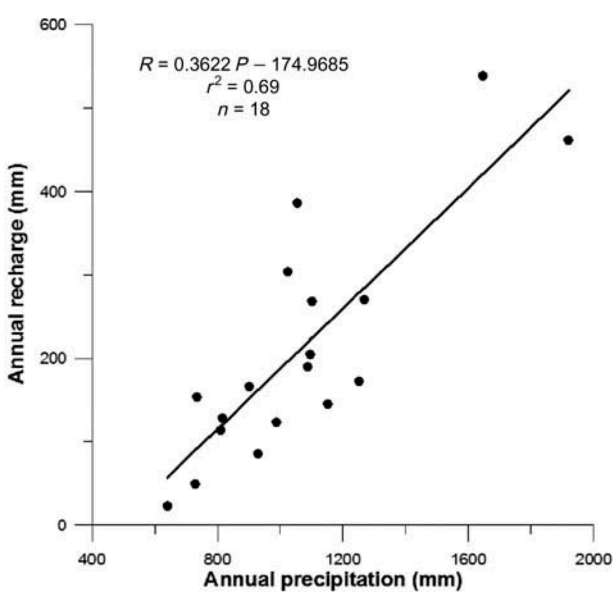

Fig. 4 Relationship between recharge ($S_y = 0.09$) and annual precipitation (1992–2009).

The mean annual rainfall for the period is 1064 mm, with a coefficient of variation (CV) of 0.30. The average annual net rise is 2.337 m (minimum 0.26 m and maximum 5.99 m) with a CV of 0.66, which is considerably larger than the CV of the precipitation. The greatest annual recharge occurred in 2002 with 539 mm, which was 33% of the annual rainfall, although the highest percentage recharge period was in 1992, with 37% of rainfall (386 mm of 1054 mm of P). The minimum recharge occurred in 2008 with 23 mm, i.e. 4% of 639 mm of precipitation, which is also the minimum of the period. This year is regarded as extremely dry for the region (Table 1). The mean annual recharge for the 18-year period was 210 mm, or 18% of precipitation ($S_y = 0.09$), while for $S_y = 0.07$ the recharge reached 164 mm (14% of P), which is connected to a high annual variability in comparison with the precipitation that generates it.

Figure 5 shows the annual variation of rainfall and recharge for $S_y = 0.07$ and $S_y = 0.09$. The recharge follows rainfall variations, but with smaller amounts: the peaks are less marked because the years with heavy rainfall tend to increase surface runoff.

Different investigators have estimated recharge to the same aquifer in this region: Varni (2005) states that in the central basin, where the well analysed in this work is located, recharge was estimated as "*between 10 and 15% for the chloride mass balance*", "*16.9% (years 1992 to 2001 with $S_y = 7\%$) for the water table fluctuation method*" and "*20% (years 1991 to 2000) for the soil water balance*". Using a Thornthwaite-Mather balance with average monthly precipitation and temperature, other estimates of groundwater recharge are 13.5% of mean precipitation (Sala *et al.* 1987), 16.4% (Nagy and Auge 1992) and 10.3% (Kruse 1992). These values are consistent with those found in this work.

Recharge information was gathered by Weinzettel and Usunoff (2005) in an experimental plot located approx. 100 m from the well analysed in this work. Recharge was estimated by applying the chloride mass balance in the unsaturated zone with suction cups and in the upper zone of the aquifer, from February 1998 to the end of November 2009. Weinzettel and Usunoff obtained a recharge of 232 mm for the chloride concentration samples of the unsaturated zone and 337 mm for the aquifer samples (evidence of preferential flow). Applying the WTF method to the record of groundwater levels of the well of this work for the same period, we obtained a recharge of 309 mm with $S_y = 0.09$ and a recharge of 240 mm with $S_y = 0.07$. These values are set between

Table 1 Annual precipitation (mm), mean depth below measuring point (m), annual level variation (m), net annual level rise (used to calculate recharge) (m), net annual level decline (m) and recharges (%*P*) for $S_y = 0.09$ and $S_y = 0.07$.

Year	*P* (mm)	Mean depth (m)	Δh (m)	Net rise (m)	Net decline (m)	*R* (%*P*)	
						$S_y = 0.09$	$S_y = 0.07$
1992	1054	1.89	0.175	4.290	4.115	37	28
1993	1024	2.42	−0.230	3.375	3.605	30	23
1994	816	3.6	−0.585	1.430	2.015	16	12
1995	736	3.66	−0.790	1.710	2.500	21	16
1996	1151	4.19	0.350	1.615	1.265	13	10
1997	1102	3.67	0.552	2.987	2.435	24	19
1998	1087	2.65	−0.342	2.109	2.451	17	14
1999	1253	3.6	−0.340	1.919	2.259	14	11
2000	1268	3.3	0.860	3.000	2.140	21	17
2001	1918	2.56	0.930	5.125	4.195	24	19
2002	1646	1.92	−0.060	5.990	6.050	33	25
2003	1097	2.97	−0.840	2.280	3.120	19	15
2004	809	3.83	−1.040	1.270	2.310	14	11
2005	730	4.82	−1.070	0.550	1.620	7	5
2006	987	5.09	0.170	1.370	1.200	12	10
2007	901	4.73	0.390	1.840	1.450	18	14
2008	639	5.38	−1.130	0.260	1.390	4	3
2009	930	6.02	0.290	0.950	0.660	9	7
Mean	1064	3.683	−0.151	2.337	2.488	18	14
SD	319.42	1.19	0.660	1.55	1.33	9	7
CV	0.30	0.32	−4.380	0.66	0.53	0.47	0.47

Note: SD: standard deviation.

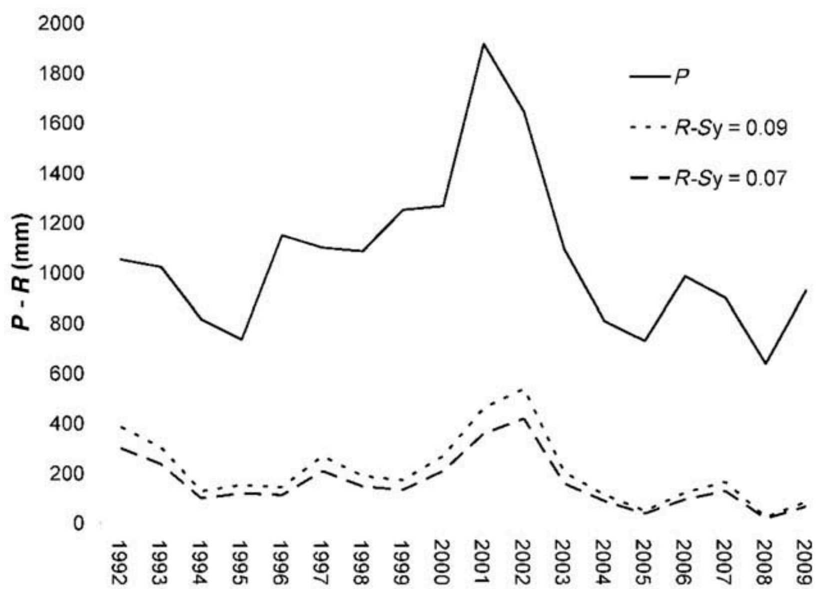

Fig. 5 Annual variations in precipitation and groundwater recharge (1992–2009).

the two estimations obtained in this work. Weinzettel *et al.* (2005) found a recharge value of 11.8% of *P* (229.8 mm) by applying a soil water balance, and a value of 11.1% of *P* (216.5 mm) using the zero flux plane method at a plane located at 120 cm below the surface at the same site during the period October 1998–September 2000. The recharge value obtained for the same period with the WTF method in the well analysed in our work is 244 mm or 12.6% of *P*. Thus, the three methodologies produce very similar results.

Limited spatial representativeness is associated with the WTF method (Delin *et al.* 2007), although the great length of the records of the well analysed here make its analysis worthwhile. However, we believe that, in a flat homogeneous area with the characteristics of the hydrological cycle of the above-mentioned plains, we can assume a bigger spatial representativeness. To verify such spatial representativeness of the water table variation in the analysed well, its monthly measured rises were correlated on a daily basis with those of two shallow wells located in the basin in which groundwater data loggers were installed about three years ago (LP and SJ in Fig. 1). A value of $r^2 = 0.79$ was obtained for the LP well (29 months) and $r^2 = 0.64$ for SJ (36 months). It was thus accepted that the analysis is representative at least for the central area of the basin.

4.3 Monthly recharge

Figure 6 shows the distribution of the mean monthly rainfall and recharge for the 1992–2009 period; the recharge distribution is bimodal with the main recharge periods during autumn and spring. At first glance, the distribution is similar to that of rainfall, March having the largest amounts of rainfall and recharge, except for the warmest months of the year (November, December, January and February), during which the greater influence of evapotranspiration reduces the recharge values to 3.7% and 5.1% of *P* in January and December, respectively. However, during August, June and March, there are high percentage recharge values (38.0%, 37.3% and 34.7% of *P*, respectively), although the actual amounts (in mm) differ greatly, since June recharge values are only 37 mm and March recharge values are 143 mm. Thus, although the mean values of annual recharge reach 18% of *P*, the monthly recharge distribution shows significant variation. The magnitude of the percentage recharge in winter and early spring is due to the usually greater water storage in the soil resulting from the combination of autumn rainfall and low winter evapotranspiration; with high soil moisture contents, even small rainfall amounts can produce groundwater recharge.

Thus, the inter-annual variation of monthly recharge values is also important, because the CV exceeds 100% for every month; the maximum values occurring in December, January and February are: 180%, 240% and 180%, respectively. Of course, these values are not surprising given the high variability of rainfall during the summer period.

4.4 Calculation of annual declines in groundwater levels

The aquifer is discharging water continuously in one or more low areas (streams, wetlands, lakes, sea) having a greater or lesser difference of elevation and distance from the test well as the rise in water table occurs during recharge events which are temporally limited periods. If the groundwater level is near the ground surface, there may be discharge by evapotranspiration. The annual changes in water level can be calculated as the difference between the depth of water table at the beginning of the year and the depth at the beginning of the following year, Δh (Table 1). A positive Δh indicates a rise in level.

Figure 7 represents the analysis of the relationship between annual total declines and the depth of the mean annual groundwater level. It clearly indicates

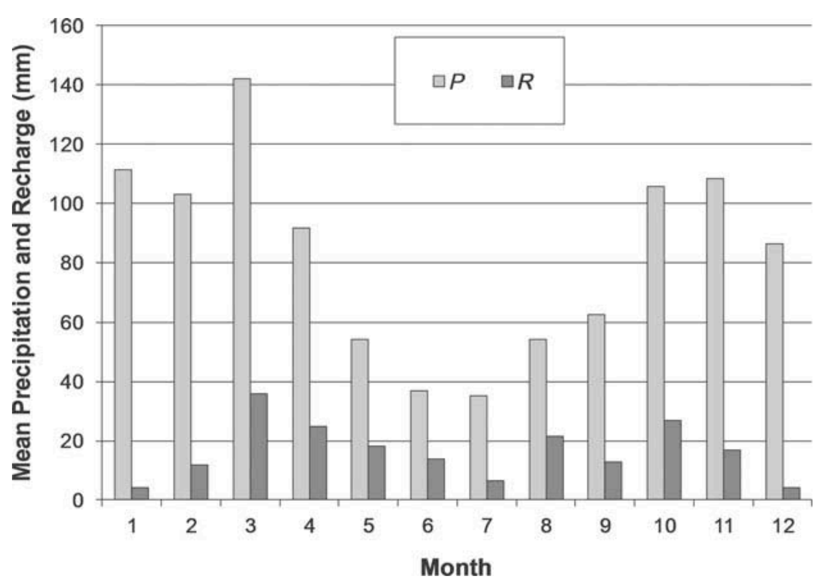

Fig. 6 Mean monthly rainfall and recharge (mm) (1992–2009).

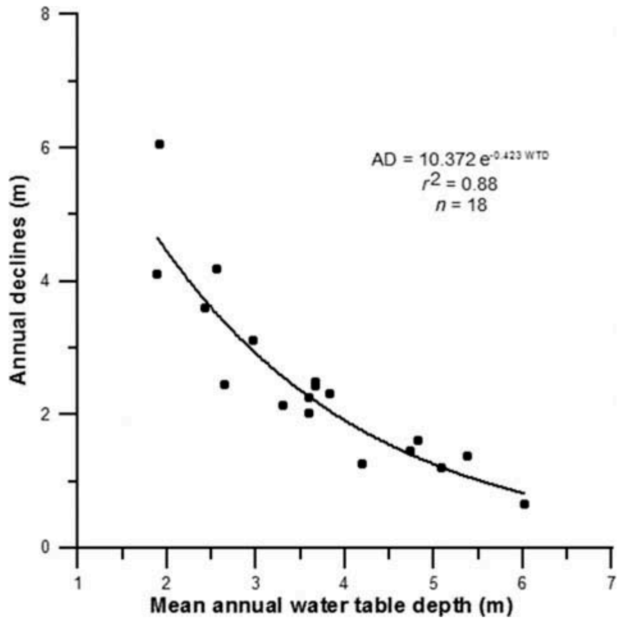

Fig. 7 Relationship between total annual groundwater level decline and water table depth in the control well.

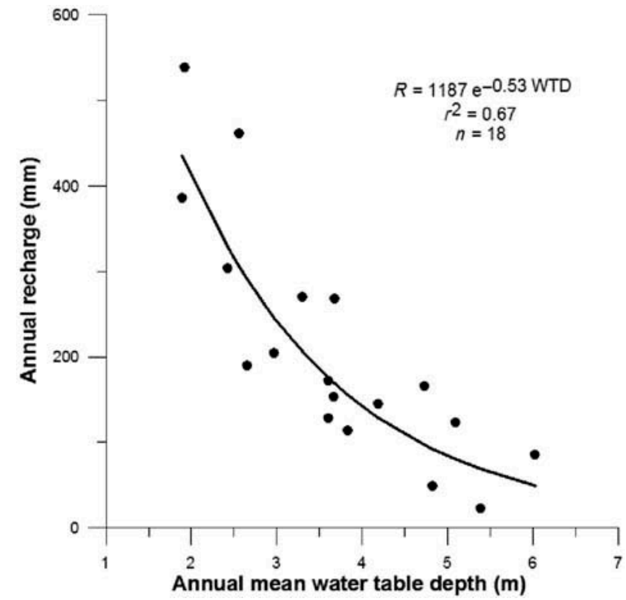

Fig. 8 Relationship between groundwater recharge and water table depth.

that annual declines are inversely proportional to the water table depth. The coefficient of determination of the relationship, r^2, is 0.88 and NSEM is 0.84, but it must not be extrapolated to other similar areas because it is heavily dependent on local discharge conditions.

4.5 Groundwater recharge relationships

The higher the water level in a basin, the greater the amount of stored water in the aquifer will be. Thus, the stored water is related proportionally to the amount of recharge, when this is the only water input to the aquifer, and the recharge should have a reverse relationship with the depth of water table; Fig. 8 confirms this assumption, with $r^2 = 0.67$ and NSEM = 0.75. It should be noted that an adjustment to give a linear equation yields the same r^2 but a lower NSEM (0.67).

If recharge is dependent on annual rainfall (Fig. 4) and on the mean depth of groundwater (Fig. 8), it is reasonable to attempt to obtain a multiple correlation with these two variables (using water level, WTL, rather than groundwater depth). The equation obtained is:

$$R = 0.235P + 59.223\text{WTL} - 7695.673 \quad (4)$$

with $r^2 = 0.86$ and NSEM = 0.86, where R is the groundwater recharge (mm), P is precipitation (mm) and WTL is the mean water table (m a.s.l.), all of these representing annual values. A linear equation was adjusted because the relationships conducted separately between R, P and WTL were linear (although in the case of the relationship between R and WTL an exponential equation also fits), and because, according to the principle of parsimony, the simplest solution is always preferable.

Finally, we explored the relationship between the baseflow in a nearby stream with rainfall and the average annual groundwater level. The mean values of baseflow are available for only 8 years (Table 2). A linear equation with a good fit ($r^2 = 0.90$, NSEM = 0.83) is obtained:

$$Q_\text{b} = 0.002P + 0.389\text{WTL} - 50.013 \quad (5)$$

with Q_b in m^3/s, P in mm, and WTL in m. Therefore, if both the recharge and baseflow can be

Table 2 Mean annual baseflow registered at Seminario Station.

Year	Baseflow (m^3/s)
2002	4.06
2003	2.17
2004	1.22
2005	1.37
2006	1.27
2007	1.80
2008	1.09
2009	0.68

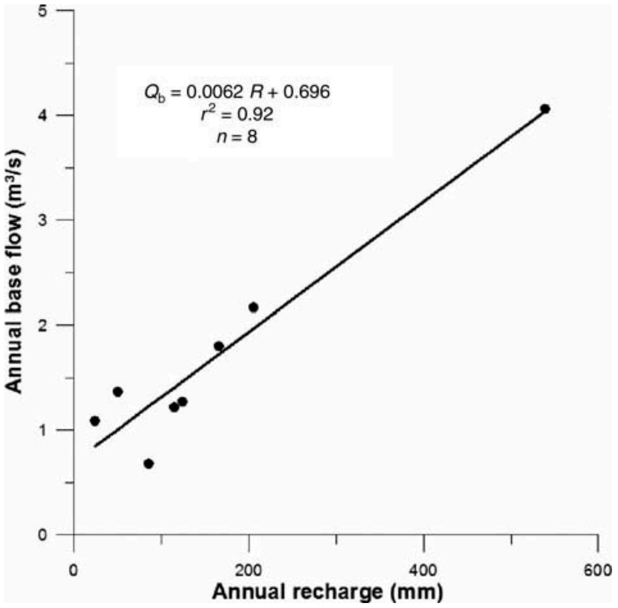

Fig. 9 Relationship between baseflow and recharge.

explained in terms of precipitation and annual average groundwater levels, it follows that there must be a good relationship between recharge and baseflow. Indeed, we found a linear equation that relates the two, backed by values of $r^2 = 0.92$ and NSEM = 0.91, whose graphic expression can be seen in Fig. 9. This last relationship expresses that the more the recharge, the higher the groundwater levels become and, given the latter, the greater the baseflow.

These relationships are useful to adjust the balance of the regional water system and will help understand the effects of an increase in exploitation, for instance. We also consider that these relationships are valid for a much wider area in the pampas plains, which has very similar weather, geomorphic and geological conditions. However, we have not been able to prove this because of lack of information for these areas.

5 CONCLUSIONS

The water table fluctuation method has proven to be effective for estimating groundwater recharge in a shallow water table aquifer in a sub-humid plain. The estimates are consistent with those found in the literature. It has also been very useful for describing the temporal variation of recharge.

The specific yield of 0.09 was estimated by analysing all recharge events measured over the period of record. A value of $S_y = 0.07$ was obtained 100 m away from the well in a plot of unsaturated zone observation.

The mean annual recharge for the period was 210 mm or 18% of precipitation ($S_y = 0.09$), while using $S_y = 0.07$, the recharge was 164 mm (14% of P). The greatest annual recharge occurred in 2002 with 539 mm (33% of annual rainfall) and the minimum recharge occurred in 2008 with 23 mm (4% P).

The relationship between annual declines of water level and mean annual water table depth, is inversely proportional, with a good fit ($r^2 = 0.88$). These declines occur due to the aquifer discharge at water courses or water bodies and because of the effect of evapotranspiration when the water table is close to the surface.

We found an equation that relates annual recharge with precipitation and mean annual water levels, with a correlation coefficient of 0.86. We also found a good relationship between mean annual baseflow with precipitation and mean annual water table, with $r^2 = 0.89$. Based on the last two equation fits, we infer a good relationship between recharge and baseflow: this was confirmed with a linear relationship ($r^2 = 0.92$).

Finally, it is important to highlight that the adjusted equations, though they are of local scope, indicate relationships between the hydrological variables of one of the most productive farming and agricultural regions of the world, with important results for irrigation development in the region and for the conservation of the environment, such as the maintenance of minimum flows in the streams of the region.

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