The Guarani Aquifer System: estimation of recharge along the Uruguay–Brazil border

Andrea A. Gómez · Leticia B. Rodríguez · Luis S. Vives

Abstract The cities of Rivera and Santana do Livramento are located on the outcropping area of the sandstone Guarani Aquifer on the Brazil–Uruguay border, where the aquifer is being increasingly exploited. Therefore, recharge estimates are needed to address sustainability. First, a conceptual model of the area was developed. A multilayer, heterogeneous and anisotropic groundwater-flow model was built to validate the conceptual model and to estimate recharge. A field campaign was conducted to collect water samples and monitor water levels used for model calibration. Field data revealed that there exists vertical gradients between confining basalts and underlying sandstones, suggesting basalts could indirectly recharge sandstone in fractured areas. Simulated downward flow between them was a small amount within the global water budget. Calibrated recharge rates over basalts and over outcropping sandstones were 1.3 and 8.1% of mean annual precipitation, respectively. A big portion of sandstone recharge would be drained by streams. The application of a water balance yielded a recharge of 8.5% of average annual precipitation. The numerical model and the water balance yielded similar recharge values consistent with determinations from previous authors in the area and other regions of the aquifer, providing an upper bound for recharge in this transboundary aquifer.

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Introduction

Campana (2005) defined that "transboundary ground water refers to a continuous ground water reservoir (generally an aquifer) that underlies or whose water flows beneath two or more political jurisdictions and can be exploited by each jurisdiction". Jurisdictions can be either different nations or different states (or provinces) within a nation. While debates regarding management of transboundary river basins have been taking place for many years, transboundary aquifers have received more recent attention. Wolf and Giordano (2002) reported that more than 3,600 water-related treaties have been signed by countries sharing some of the 263 international river basins between the years 805 and 1984. In contrast, agreements or treaties on international groundwater resources date back only to the last 50 years, with none of them developed in South America (Jarvis et al. 2005).

Over the past two decades, the scientific community started developing an increasing interest in transboundary groundwaters. As a result of preliminary meetings, a program for an international initiative on internationally shared/transboundary aquifer-resources management (ISARM/TARM) was established (Puri et al. 2001). Recently, the World-wide Hydrogeological Mapping Assessment Program (WHYMAP), launched in 1999 under the sponsorship of many organizations, published a world inventory of transboundary aquifer systems (TAS; BGR/UNESCO 1999).

One of the TAS mapped is the Guaraní Aquifer System (GAS), the largest in South America and one of the largest in the world. The GAS covers an area of approximately 1.2 million km^2 (Fig. 1), with an estimated volume of water of 40,000 km^3 (Araújo et al. 1999). It extends under the territory of four countries: 840,000 km^2 in Brazil, 255,000 km^2 in Argentina, 71,700 km^2 in Paraguay and 58,500 km^2 in Uruguay. More than 20 million people live in the area where the aquifer is being increasingly exploited and constitutes an important source of freshwater for urban supply as well as for industrial and agricultural use.

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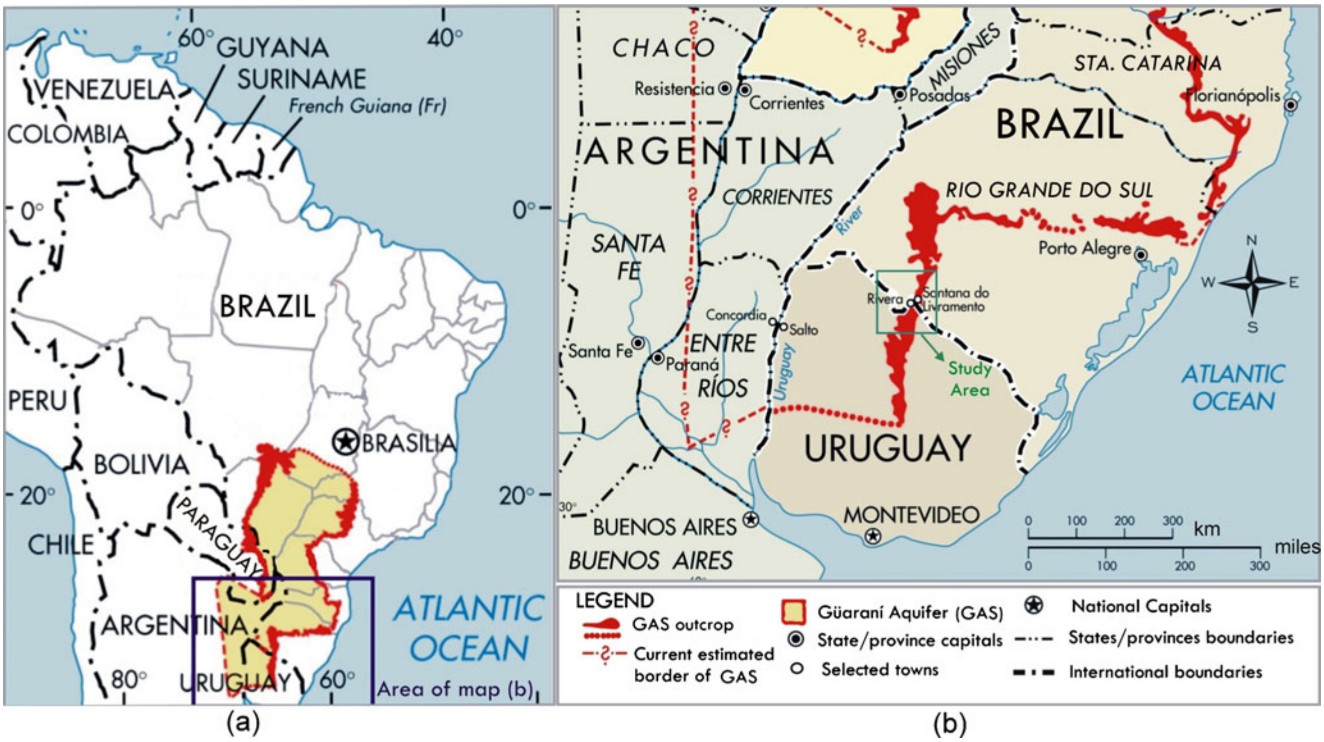

Fig. 1 **a** Location of Guaraní Aquifer System (GAS) in South America. *Red dashed line* indicates approximate aquifer western–southern boundary. **b** Location of study area within the southern portion of the GAS

A multidisciplinary, multinational work team recently concluded the Guaraní Aquifer System Project, hereafter GASP, developed thanks to the sponsorship of the Organization of American States, the World Bank and other cooperating agencies. The long-term objective of the GASP was to achieve the management and sustainable use of the aquifer. The work presented here was part of a local-scale modeling effort developed within GASP to study particular areas intensively exploited, characterized by different hydrogeological, social, political and economic aspects.

The aquifer is contained within sandstones of Jurassic-Cretaceous origin, overlain by a thick mantle of basalts in the central portions of the aquifer (Araújo et al. 1999). Water quality and depth to groundwater vary regionally. Due to its economic and social strategic relevance, it is the subject of increasing research efforts. In spite of those efforts, the knowledge and understanding of the system functioning is still limited.

Sandstones crop out along aquifer edges, deepening toward the center of the basin (see Fig. 1), where they can reach a maximum thickness of some 600 m and depths of 2,200 m. Outcropping areas are supposed local recharge or discharge zones, depending on piezometric conditions. The outcropping area located in northern Uruguay and southern Brazil, around the cities of Rivera (Uruguay) and Santana do Livramento (Brazil), is of special interest due to its population concentration, water uses, transboundary nature and recharge area. Overall, the GAS contains an enormous volume of water. However, it is not well known, so that it is hard to assess the impact of exploitation and the magnitude of mass-balance components, especially recharge. This is a sensitive issue because the aquifer is shared by four countries. The situation in the Rivera/Santana conglomerate is a local scale example, with its peculiarities, of the regional situation.

Recharge estimation has been the topic of numerous studies in the last 15–20 years (Scanlon et al. 2002). Devlin and Sophocleous (2005) stated that groundwater recharge rates are not required to assess sustainable pumping; however, they are needed to address sustainability, a broader term that involves issues such as water quality, ecology and human and environmental welfare. In tune with the long-term goal of the GASP and the statement by Devlin and Sophocleous (2005), this work presents the results of research aimed at estimating recharge rates in the Rivera-Santana area.

De Vries and Simmers (2002) reported that regional recharge can be reasonably estimated applying methods including regional-flux determination by isotope dating, Darcian flow modeling, chloride mass-balance calculations, and direct measurement of spring discharge or base flow, among others. Choosing an appropriate technique depends on the study objectives, the precision of the sought results, the working time/space scales and background information on recharge (Lerner et al. 1990; Scanlon et al. 2002). Given the uncertainties and factors involved in the estimation of recharge, it is desirable to apply several methods to restrict the resulting rates (Rushton and Ward 1979). Data availability in the study area was determinant for selecting applicable methods. In this work, a numerical model and a water balance were

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implemented to obtain two independent recharge values. Comparison with previous determinations in the area and other regions of the aquifer were performed to assess the consistency of the results.

Study area

The study area covers approximately 750 km^2 around the neighboring cities of Rivera (Uruguay) and Santana do Livramento (Brazil; Figs. 1 and 2). It is located on the southern Brazil/northern Uruguay border, at 30º53'40"S latitude and 55º32'17"W longitude. The landscape is characterized by smooth topography with some steep slopes, with maximum elevations reaching 400 m above sea level (m.a.s.l.) in the western part, disrupted by the transition between basalts and the adjacent outcropping sandstones, noticeable in the field or in topographic maps by the presence of steep slopes representing successive lava flows. The minimum elevation is around 130 m.a.s.l. in the east of the area. The elevation dataset generated by the Shuttle Radar Topography Mission carried out by the National Aeronautics and Space Administration (NASA) and the National Geospatial-Intelligence Agency from USA was used to generate the digital terrain model (DTM) of the area shown in Fig. 2, where the aforementioned geoforms can be identified.

Only one weather station, Rivera, is available within the area. Based on the time series 1960–1997, the mean annual precipitation is 1,639 mm (see Fig. 3 for the annual precipitation distribution in Uruguay). The mean annual temperature is around 17.5°C (DNM 2006). The relative humidity of the air oscillates, on average, between 72 and 77%.

The drainage network can be easily identified from the DTM (Fig. 2). Most streams in the area carry rapid flows during the rainy season, originating in high elevations, where they traverse along valleys carved in basalt terrain. As they descend they interact with groundwater in the sandstones. The most important water course is the Cuñapirú Creek, located along the southeastern border of the study area. Its sub-watershed occupies all the outcropping area in the Rivera County. This creek joins the Tacuarembó River outside the study area. There are no gauging stations within the study region, either in Uruguay or Brazil. The headwaters of Cuñapirú Creek and Tacuarembó River are shown in Fig. 2. Located approximately 75 km south of the southern boundary of the study area is the Manuel Díaz gauging station on the Tacuarembó River, which drains 5,500 km^2 (Fig. 3). The mean discharge of Tacuarembó River is 22 m^3/s and the mean annual precipitation nearby is 1,450 mm (DNM 2006). Therefore, the ratio between mean annual discharge and mean annual precipitation yields a runoff coefficient of 8.7%.

From a regional point of view, the GAS is formed by a collection of consolidated and fractured geologic units with structural control. Shallow cracks and fractures are associated with basalts overlying the sandstone Guaraní Aquifer. Deep fractures and faults associated with vertical

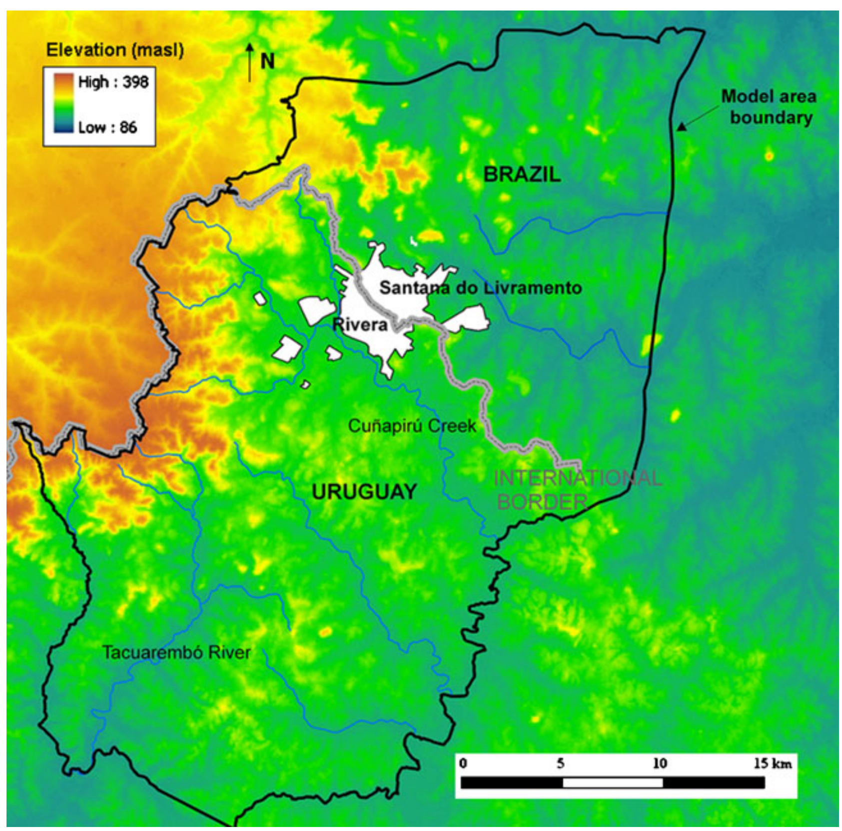

Fig. 2 Digital terrain model (DTM), main cities, numerical model boundary, and main water courses in the study area

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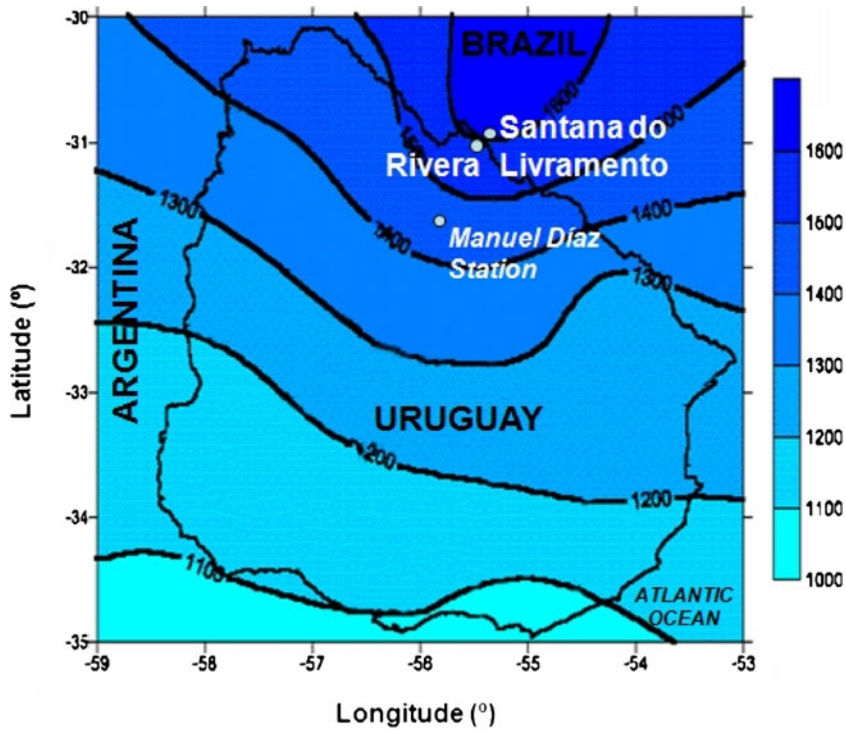
Fig. 3 Mean annual precipitation distribution of Uruguay (in mm/year). Source: National Direction of Meteorology (Uruguay; DNM 2006)



and horizontal movements may control groundwater flow at regional scale. A more detailed regional description of the complex geological characteristics of the GAS can be found in Rebouças (1976), Araújo et al. (1999) and Kittl Tage (2000), among others.

On a local scale, the study area belongs to the "Cuenca Norte Uruguaya", an interior cratonic basin composed by depositional events from the Devonian to the late Cretaceous (Oleaga 2002). The Jurassic-Cretaceous sequence is represented by the Cuchilla Ombú, Tacuarembó and Rivera Formations (see Fig. 4 for their identification on a stratigraphic profile). These formations, composed of sediments of fluvial and eolean origin, form the GAS, which crops out at the surface along the eastern boundary of the aquifer, and hence the study area, deepening in the E–W direction as it enters the Argentinean territory. Basalts of the Arapey formation, known as Serra Geral in Brazil, are not currently considered part of the GAS; however, they may be hydraulically connected to underlying sandstones through fractures; this hypothesis is reflected in the conceptual model that supports the numerical model presented in the following sections. Quaternary sediments constitute the uppermost unit of the lithologic profile, though its areal extent is limited to small areas along streams and rarely described in stratigraphic profiles from wells.

Direct recharge to the GAS originates in precipitation while an unknown quantity of water from overlying fractured basalts may potentially contribute to indirect recharge (Silva Busso 1999; Oleaga 2002). This recharge, for which the magnitude is still unknown, would occur mainly in places where the basalt is not thick (near sandstone outcroping areas) and its fractures would be well connected between each other, reaching the underlying GAS (Rosa Filho et al. 2003). More details regarding the hydrogeology of the area are included in the following section.

Conceptual model and aquifer parameters

The first step within the modeling process was the development of a sound conceptual model representing the aquifer system structure and its hydraulic behavior. Stratigraphic profiles from 141 boreholes (70% of which were no more than 20 m deep), geologic maps, hydrogeochemical data and water levels measured at 69 wells were analyzed and integrated into the conceptual model (Gómez 2007; Rodríguez et al. 2008).

Sandstones of the Tacuarembó/Rivera Formations, basalts of the Arapey Formation and Quaternary sediments aligned with streams are the predominant geologic units surrounding both cities. The thickness of basalts ranges from 0 to 500 m, while sandstone layer thickness ranges from 0 to 200 m.

Between August and December 2005, a field campaign was conducted to measure static water levels and sample wells for hydrochemical analyses. These data, completed with water levels surveyed in rural areas by Collazo (2006), allowed the authors to postulate a conceptual model and to define the flow system. Differences in water levels were identified in the field (Rodríguez et al. 2008), which could trigger vertical flows. One level was within basalts, above water levels in shallow sandstones, with water-level differences of up to 40 m. Basalt data are scarce and this could well be a perched-aquifer situation.

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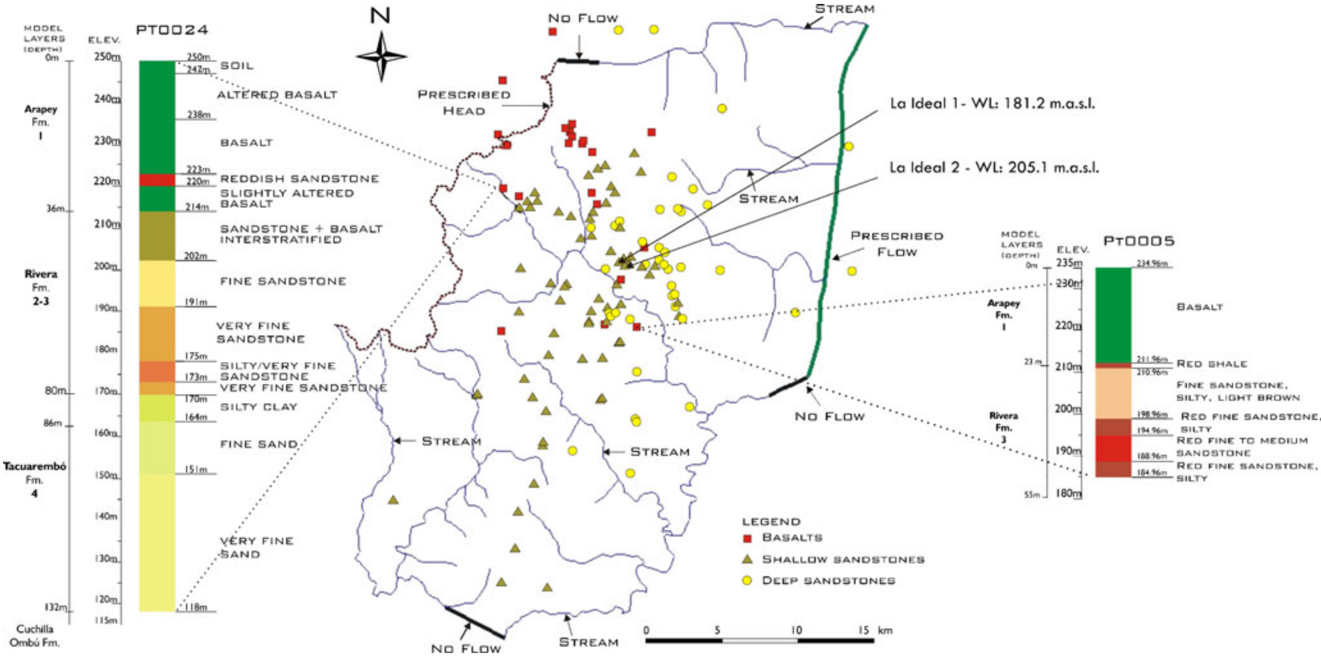

Fig. 4 Location of wells for which logs were available for examination, superimposed onto the boundary of the study/modeled area (*symbols* at each location indicate the outcropping material at the well site). Stratigraphic profiles of two wells (*PT0024* and *PT0005*) show geologic formations. Locations of wells (*La ideal 1* and *La ideal 2*) are also shown. *WL* water level. Elevations are in m above sea level

Even so, one can hypothesize that a large enough hydraulic gradient exists so as to produce downward flows through preferential flow paths associated with fractures. This is also supported by the hydrochemistry (Rodríguez et al. 2008). During the field campaign, a surface lineaments reconnaissance was performed within Uruguayan territory. With the aid of satellite images, 366 shallow lineaments were mapped, characterized by a mean length of 340 m and a standard deviation of 300 m (Gómez 2007). Even though the connectivity of this family of fissures is yet to be explored, the hypothesis of basalts/GAS hydraulic connectivity is feasible.

Underlying the basalt and outcropping to the east, a shallow, permeable sandstone layer, hydraulically connected to streams that cut across the landscape, was identified. This layer disappears in Brazilian territory.

A deep, sandstone aquifer located in very permeable sectors of the GAS was present, with hydraulic parameters and water yield varying locally. Within Uruguay, groundwater from the shallow aquifer would discharge into the Cuñapirú Creek. Water levels in the shallow sandstone aquifer were always higher than in the deep sandstone aquifer supporting the hypothesis of downward flows between these two layers. A maximum water level difference of 24 m between shallow and deep sandstone was measured at two wells that are close to each other and tapping the two formations (wells La ideal 1 and La ideal 2; see Fig. 4). A high-clay-content interbedded layer was detected in a few wells in the central area of the study region. The well log for PT0005 (see Fig. 4) shows a sequence of finer sandstones that hydrogeologists have associated with aquitard conditions that would locally affect groundwater flow.

The thick, deep aquifer accommodates most of the water-supply wells for both cities. Flow direction is mainly towards Brazil, locally disrupted by two incipient cones of depression in urban areas, one in Uruguay, one in Brazil. The hydraulic gradient near the international border is 3.5×10^{-4}. Close to the northwestern border of the study area, an easterly flow is present, with a hydraulic gradient slightly higher (5×10^{-4}). Close to the area affected by pumping, a higher hydraulic gradient was found.

Consequently, the proposed conceptual model for the aquifer system, from top to bottom, is:

1. Upper aquifer: contained within altered basalts areas
2. Lower, multilayer sandstone Guaraní Aquifer with the following layers

   a. Shallow Guaraní Aquifer that sustains small and sparse domestic wells
   b. Aquitard, composed of high clay-content sandstones
   c. Deep Guaraní Aquifer that sustains high pumping-rate wells, exploited by water-supply companies

Figure 5 shows the reconstructed stratigraphy and conceptual model. The three-dimensional (3D) stratigraphic model was built with the Groundwater Modelling System (GMS V 6.0 2006).

Pumping tests previously conducted near the cities of Rivera and Santana do Livramento were re-interpreted to estimate the hydraulic conductivity (K) of the different formations. Data from a total of 24 wells, 17 from Rivera and 7 from Santana do Livramento, were analyzed. Most of the tests were performed at the time of the well

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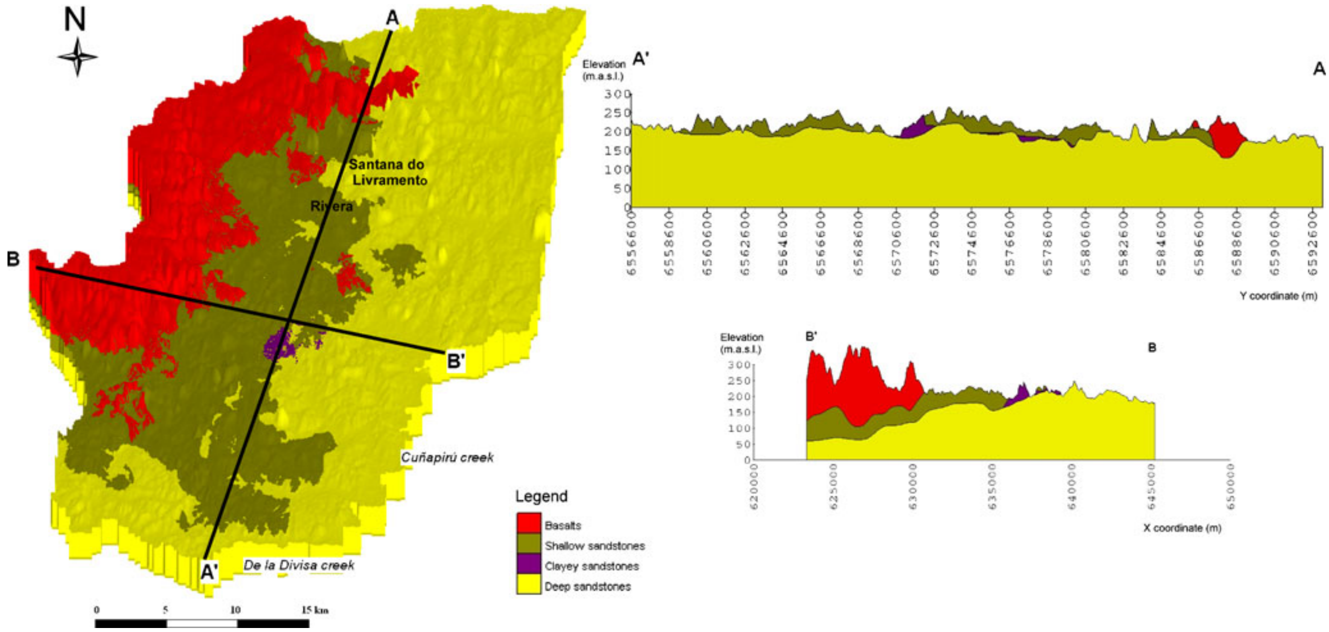

Fig. 5 Three-dimensional stratigraphic model of the aquifer system. Characteristic profiles in NE–SW and W–E directions

construction; wells were partially penetrating and screen-length data was readily available. Aquifer thickness at the sites was unknown. Transmissivity (T) was estimated through the Theis method, and K was calculated by dividing T by the screen length at each well. It is acknowledged that this procedure strictly applies to fully penetrating wells, which causes the flow in the aquifer to be strictly horizontal. Hantush (1964) recommended reaching a certain distance between the pumping well and the observation well in order to neglect the effect of a partially penetrating well. Nonetheless, he stressed that, if both pumping and observation wells are partially penetrating, the drawdown formula becomes quite complex. Neuman (1974) also suggested conditions to overcome such an effect that would make it possible to use the Theis curve. The absence of data regarding observation wells distances and aquifer thickness complicated the analysis and leads one to make simplifying assumptions that may yield overestimated aquifer parameters. For this reason, estimated values were compared to K values obtained in other parts of the aquifer. Table 1 contains estimated K values while Fig. 6 shows the location of pumping test wells used for the analysis. Notice that all data are highly concentrated. For the deep aquifer, K ranged from 0.12 to 5.76 m/day, with a mean of 1.5 m/day. The high-end value seems high for sandstone aquifers; however, Araújo et al. (1999) reported an average K as high as 8.64 m/day for the GAS, while Sracek and Hirata (2002) published K values ranging from 2.07 to 64.8 m/day. For the shallow sandstone a single value of 0.48 m/day was found.

All K values were in accordance with K values reported for these materials published by Custodio and Llamas (1983), towards the upper end of the interval reported by Freeze and Cherry (1979) and towards the lower end of the range published for the entire GAS. No pumping tests were available for basalts at the study site. Hydraulic conductivity obtained at other intensively studied basalt areas were used as reference, being cautious due to geologic particularities of each site. Nimmo et al. (2004)

Table 1 Estimated hydraulic conductivities (K) values from pumping tests. *D* deep sandstone; *S* shallow sandstone; *B* basalt; *ND* not determined

Borehole	K (m/day)	Layer
PT0005	0.72	D
PT0021	1.97	ND
Registro 3	1.15	D
AC2	0.48	D
La ideal	0.48	S
La ideal	1.44	D
10.4.031	3.12	B
10.4.033	0.17	D
10.4.003	1.15	ND
10.4.005	5.28	ND
10.4.008	2.16	D
10.4.011	0.31	D
PT0007	5.76	D
PT0019	19.92	D
Vertedero	1.68	D
10.4.025	0.34	D
Kennedy 2	3.36	D
AL1	0.55	D
10.4.034	2.98	D
Wilson 3	0.77	D
AC4	0.12	D
Prado 1	8.88	D
10.4.007	1.13	ND
10.4.030	1.13	D

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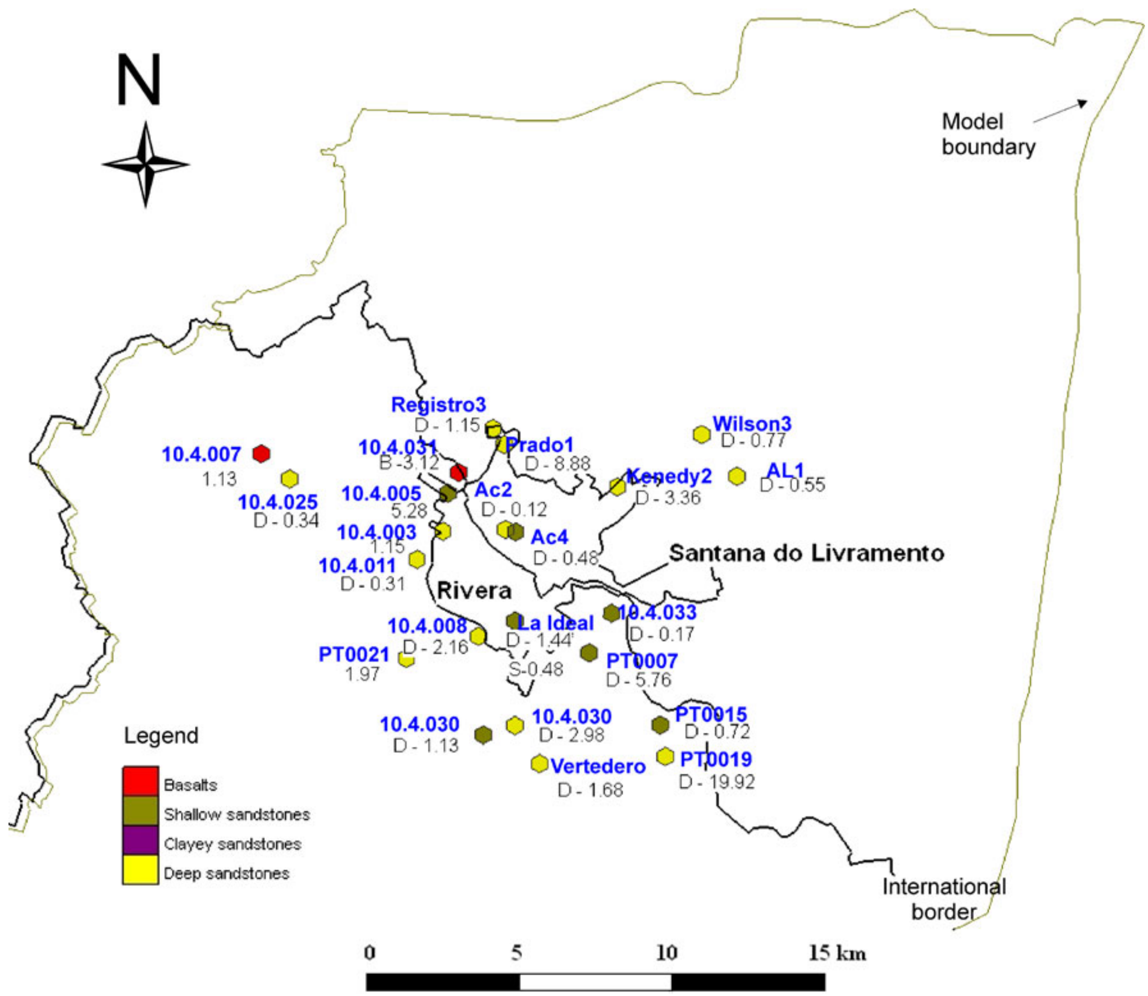

Fig. 6 Location of pumping test wells. Highlighted in *gray* is the hydraulic conductivity value (in m/day) estimated at each site. The letter *D* indicates that the well belongs to deep sandstones, *S* to shallow sandstones and *B* to basalts. If *no letter* is indicated, the well could not be associated to a particular material/layer

used several approaches to determine basalts hydraulic properties, from single-well aquifer tests, laboratory measurements, large-scale infiltration tests, to inverse modeling, and forward modeling. The testing site for their work was the Idaho National Engineering and Environmental Laboratory Vadose Zone (INEEL) in the USA. Based on estimates from single-well aquifer tests of 114 wells at and near INEEL, they found saturated K ranging from 9.6×10^{-6} to 9,760 m/day. They reported that the largest values correspond to fractured basalt and near-vent volcanic deposits while the smallest correspond to dikes, dense basalt, and altered basalt. About two thirds of these estimates are>25.9 m/day, and about one third are> 302 m/day. For permeable basalts, Domenico and Schwartz (1990) provide K values between 0.034 and 1730 m/day.

Most of the K values for Serra Geral basalts result from calibration of groundwater flow models. For fractured basalt in Paraguay, Vassolo (2007) estimated horizontal K (K_h)=2 m/day and vertical K (K_v)=3×10^{-3} m/day. In Brazil, Rebouças (1988) provided a range for K between 8.6×10^{-2} and 86 m/day. For altered basalts K_h=0.2 and K_v=0.8 m/day were estimated through a groundwater model calibration, defining $K_h/K_v<1$ due to fracturing (Heine 2008). Fracturing can create preferential flow paths defining a secondary K bigger than that of the rock matrix. Fernandes and Rudolph (2001) stated that in Serra Geral, apertures in the order of 1–2 mm are enough to increase basalts transmissivity considerably.

Methods

Data availability in the study area was important for selecting applicable methods to estimate recharge. A numerical model developed with MODFLOW (McDonald and Harbaugh 1988) was used to validate the postulated conceptual model and calibrate recharge rates and other aquifer parameters to match current piezometric conditions. The water balance EASY-BAL (Vazquez-Suñé and Castro 2002) was implemented to obtain a second recharge rate estimate in accordance with recommendations regarding the use of multiple methodologies. Results of the water balance are presented first.

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Water balance model

Simple water balance models have been extensively used for estimating groundwater recharge *R* (see, for instance, Finch 1998). The most common way to estimate *R* with this method is the residual approach in which all variables are calculated/measured separately, except *R* (Scanlon et al. 2002). In this work, EASY-BAL, a sequence of simple equations built on a spread-sheet environment applied to monthly conditions (monthly serial water balance), was used. EASY-BAL computes potential evapotranspiration (ETP) using the Thornthwaite method (Thornthwaite and Mather 1955), transforms ETP into real evapotranspiration (RET) based upon water availability in the soil profile, field capacity, wilting point, soil thickness and precipitation. A critical parameter to the model is the runoff threshold (RT), which defines the upper bound for infiltrated water for each month, i.e. RT controls the maximum amount of monthly precipitation available for infiltration. Depending on the relative magnitude of RET and precipitation, calculations may result in a water deficit or water excess; in the latter case recharge will be estimated. A complete explanation of model variables and equations can be found in Vazquez-Suñé and Castro (2002). A summary of EASY-BAL is given in at the end (see Appendix).

The series 1960–1997 was used for monthly meteorological data (temperature and precipitation) corresponding to the Rivera Station, obtained from the National Direction of Meteorology of Uruguay. In the EASY-BAL spreadsheet the user has to define RT, the initial reserve and the useful reserve. In this work, the RT was defined as 100 mm, the initial reserve was set at 3.4 mm and the useful reserve was set at 100 mm.

Calculated monthly recharge was annualized to render values comparable to other authors' estimates. Average annual recharge for the period 1960–1997 was 139.5 mm/year (8.5% of mean annual precipitation, which is 1,639 mm/year) for a value of the monthly RT of 100 mm (named RT_{100} hereafter). Figure 7 shows normalized annual precipitation *P*, annual recharge *R* and real evapotranspiration ETR. The effect of evapotranspiration over the magnitude of recharge is noticeable. For instance, in 1967, the precipitation was close to average while RET was well below average producing a mean annual recharge above average. The same situation occurs in 1985 when RET suffered a significant reduction, and in 1997. Other cases are observed in 1969 and 1979 when, in spite of below-average ETR, a low precipitation produces a small annual recharge.

Because the magnitude of the initial reserve was unknown, several water-balance calculations were performed starting from different initial conditions to analyze the sensitivity of the results during the first few months. Those would be candidates to show some effect, if any. It was found that the effect of the initial condition quickly dissipated after the first month. For instance, an initial reserve value of 24 mm, i.e. seven times higher than the calibrated value, caused a 50% increment on RET for the first month and 0% change in following months. On an annual basis, RET for the first year increased 3%. Recharge for the first month was not affected. The useful reserve constitutes an upper bound for the initial reserve. When the initial reserve was set equal to 100 mm, RET

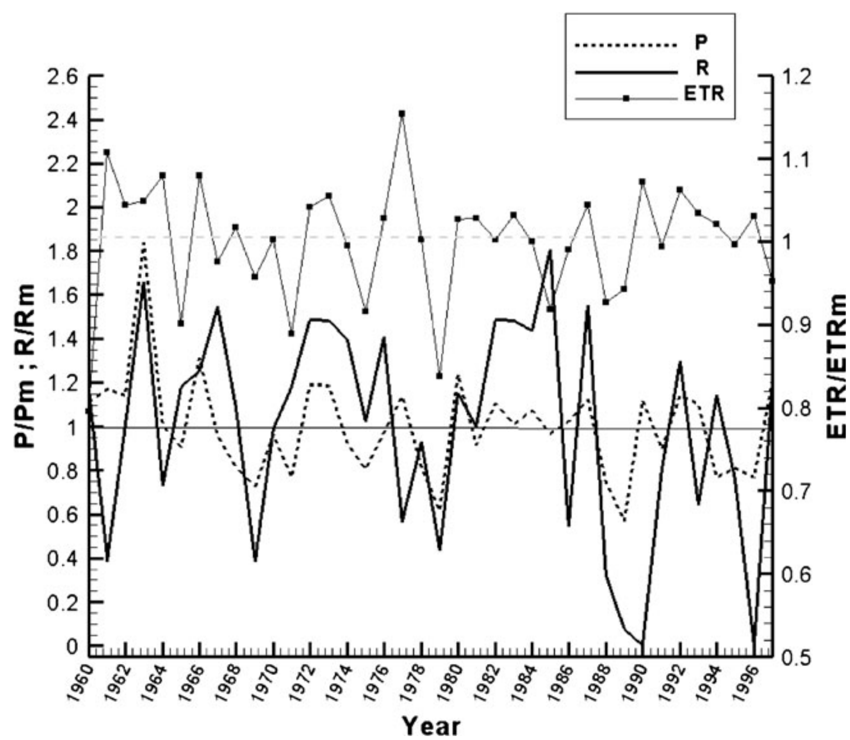

Fig. 7 Annual precipitation (*P*), real evapotranspiration (*ETR*) and recharge (*R*) at Rivera Station (Uruguay). The subscript *m* means mean value of each variable used for normalization

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for the first month increased as much as 240%, which means a 14.3% increment on annual RET for 1960. Recharge was not affected. In conclusion, effects were constrained to the first month; hence, the first year of the time series was discarded from the analysis of results.

The useful reserve parameter would represent the soil capacity to retain water. It would depend on various variables such as root depth, effective soil porosity, among others. Root depth and depth to groundwater vary greatly throughout the study area, characterized by several land uses. Considering a porosity equal to 0.1 and a soil column of 1 m, a 100 mm useful reserve results. This crude estimate is considered a lower bound for the study area. Consequently, recharge calculations based on the water balance would be somehow overestimated. A higher useful reserve would cause more water to be retained in the soil profile and less water available for recharge. A 50% increment on the useful reserve value produced a 30% decrease in mean annual recharge.

Runoff threshold was estimated from the runoff coefficient obtained for the Tacuarembó River basin and the precipitation at Rivera Station. Note that approximately 550 km^2 of the study area are contained within that basin occupying the highest slopes and elevations. Slopes can reach as much as 20%. Also note that the mean annual precipitation at Rivera is 13% higher that at Manuel Díaz, the site where streamflow and precipitation data outside the model area were available. Based on this information, a higher runoff coefficient of 25% was estimated. Applying that coefficient to the mean monthly precipitation of 134 mm for the calculation series, yields 100 mm for the monthly threshold runoff value.

Water balance models are known to be sensitive to land surface parameters (Finch 1998). According to its authors, EASY-BAL is more sensitive to the parameter RT; therefore, the water-balance sensitivity to this parameter was also explored. Figure 8 shows the response of the relative recharge R_r to changes in RT relative to the value RT_{100}. The variable R_r was defined as the normalized relative recharge equal to (R_i/R_{100}–1), where R_{100} is the recharge obtained for RT_{100} and R_i the recharge corresponding to RT_i, respectively. It can be seen that a 50% reduction in RT causes a 100% reduction of R_r, while a 50% increase in RT causes a 59% increase in R_r. Higher values of RT were not explored because they were considered physically infeasible. Physically, this asymmetric behavior could be explained considering the soil capacity to infiltrate water. When RT increases, i.e. for the case $RT/RT_{100}>1$, there is less runoff water and more water available for the soil profile and potentially for recharge, so water availability is less limiting. When RT decreases, i.e. $RT/RT_{100}<1$, there is more runoff water and less water available for the soil profile to satisfy ETP and any water deficit in the profile; therefore, recharge may be reduced.

Groundwater flow model

Model set up

The numerical model selected for this work was MODFLOW 2000 (Harbaugh 2005). A multi-layer, heterogeneous and anisotropic model was built. Due to the lack of water-level data encompassing long enough time periods and extensive areas, the model was run considering the steady state in equilibrium. Rabelo and Wendland (2009) used a similar approach in the Guaraní Aquifer System in the state of Sao Paulo, Brazil.

The conceptual model of the study area was transformed to a numerical groundwater model within the user interface GMS V 6.0. Individual coverages made up of points, arcs, and polygons were developed within GMS to represent components of the system such as wells, streams and recharge areas, respectively. The first task was the buildup of the three-dimensional (3D) reconstruction of the aquifer system, based upon borehole data and the layering identified on the vertical conceptualization shown in Fig. 4. The resulting construction is presented in Fig. 5. MODFLOW relies on alternative approaches to formulate the internal flow terms, i.e. flows between adjacent finite-difference

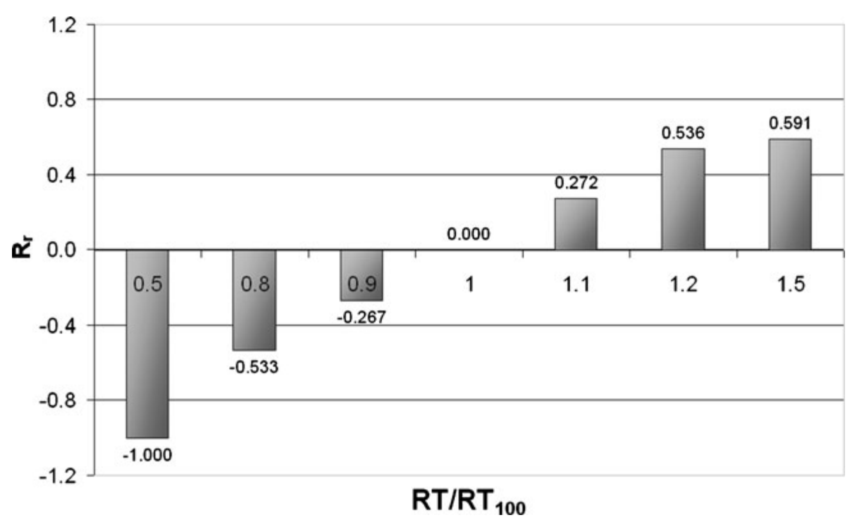

Fig. 8 Sensitivity of water-balance-calculated relative recharge (R_r) to the parameter threshold runoff *RT*

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cells/blocks: the Block-Centered Flow (BCF) package and the Layer-Property Flow (LPF) package. The LPF package supports two types of aquifer layers: confined and convertible. A confined layer is one in which transmissivity, computed from hydraulic conductivity and cell elevation, is constant throughout the simulation. A convertible layer is one in which transmissivity varies based on head throughout the simulation. The stratigraphic model shown in Fig. 5 was later imported into MODFLOW to replicate the layers within the LPF module. A more detailed explanation regarding the LPF package can be found in Harbaugh (2005).

GMS tools were used for automated grid generation and MODFLOW packages input files construction, based upon the stratigraphic model. The finite difference grid consisted on 135 rows and 156 columns, with a regular cell size of 250×250 m. Vertically, the model contained four layers coincident with the aquifer units identified in the conceptual model. As shown in Fig. 9, a different number of active cells was used in each layer to represent their distinct aerial extent. Layer 1 (basalts) had 2,049 active cells, layer 2 (shallow sandstone) 5,666 active cells, layer 3 (sandstone + embedded aquitard) 5,666 active cells, and layer 4 (deep sandstone) 12,476 active cells.

GMS contains a suite of tools for interpolating and manipulating layer elevation data. Cell size in the z direction for each layer equaled the thickness of the corresponding aquifer unit (Fig. 5). Ranges for vertical cell sizes were 3–259, 3–52, 4.4–65.1, and 45.9–240 m, for layers 1, 2, 3 and 4, respectively. The thickness of each unit was obtained from stratigraphic profiles, supported with ancillary data such as geologic maps and geophysical information. Stratigraphic profiles were transformed into the 3D representation known as "solid" or subsurface model shown in Fig. 5 using graphical tools available in GMS with linear interpolation. Then the "Solids-to-MODFLOW" command within GMS was used to automatically define the elevation arrays in MODFLOW.

The definition of an appropriate set of boundary conditions (BC) for each layer was part of the calibration process considering the actual magnitude and direction of flows across model boundaries. Combinations of no-flow, prescribed flow, prescribed head, and river boundaries were set for all layers, as can be seen in Fig. 9. No-flow boundaries were assumed for layer 1. Boundary conditions for layer 2 included a no-flow boundary reach in the north, a prescribed head condition along the west boundary coincident with the confinement area caused by the Serra Geral formation, and a river-type condition along the southwestern portion simulating the presence of the Aurora Creek in outcropping sandstones. In layer 3, boundary conditions were similar to those of layer 2, except that the river condition was replaced by a no-flow boundary. For layer 4, the border within Brazilian territory was particularly critical due to scarce information regarding the aquifer head distribution. In that case, the following strategy was applied: given few water levels available, and knowing it is an outflow boundary, first a prescribed head was defined according to those levels and topographic elevations. Once the magnitude of the outflow was obtained, that was imposed to replace the prescribed head boundary. The starting value for prescribed head in layers 2, 3 and 4 was defined based on water levels surveyed during the field campaign and topographic elevations, adjusted during calibration.

MODFLOW allows simulation of river–aquifer interactions by means of add-on packages. Streams within the model area and along portions of the model boundary, shown in Fig. 10, were simulated with the River (RIV) package, which considers constant river heads and no variation in river flows (McDonald and Harbaugh 1988). The package does not simulate surface-water flow in the river, only the river–aquifer seepage. River seepage is independently simulated for each river reach and added up to compute river–aquifer interaction fluxes. River–aquifer seepage magnitude is proportional to the hydraulic head gradient between the water level in the stream, i.e. stage, and the groundwater level in the adjacent aquifer. The conductance of the streambed material is the proportionality coefficient between the gradient and the seepage magnitude.

Simulated streams intercepted the corresponding uppermost model layer. No data existed for parameters related to the surface-water system; water level in the streams was first assigned the topographic elevation and later adjusted during the calibration process. A uniform conductance value of 1 m^2/day/m was defined and its uncertainty explored through sensitivity analysis. Considering that most of the streams are intermittent and the lack of field data, those values were deemed a good first approximation to represent base flows instead of fast surface flows.

Pumping was simulated in layers 2, 3 and 4, and distributed with 17, 7 and 106 wells and a total pumping rate of 1.06×10^3, 2.68×10^3 m^3 and 69.3×10^3 m^3/day, respectively. These rates were estimated from information provided by local water companies and additional rates from a small number of domestic wells. The pumping layer was assigned based upon the well-screen depth data, comparing this depth with the layer elevation on the 3D stratigraphic model. The estimation also considered the percentage of urban and rural areas, and return flows due to losses along the water supply system. The distribution of pumping wells is shown in Fig. 9.

Aerial recharge was defined in two different areas, one over basalts coincident with the active grid area of layer 1 (Fig. 9a) and the other over outcropping sandstones, at the uppermost cell whether it belonged to layer 2 or to layer 4 (Fig. 9b and d). Initial recharge rates were defined as 10% of mean annual precipitation at the site and later adjusted during the calibration process. No direct recharge was applied to layer 3 as this layer is completely covered by layer 2 having both the same X,Y dimensions (see Fig. 9b and c)

Calibration and results

A trial-and-error calibration was performed to validate the conceptual model and determine an independent estimate of recharge rate, under the premise that simulated groundwater levels closely match field observations. The model

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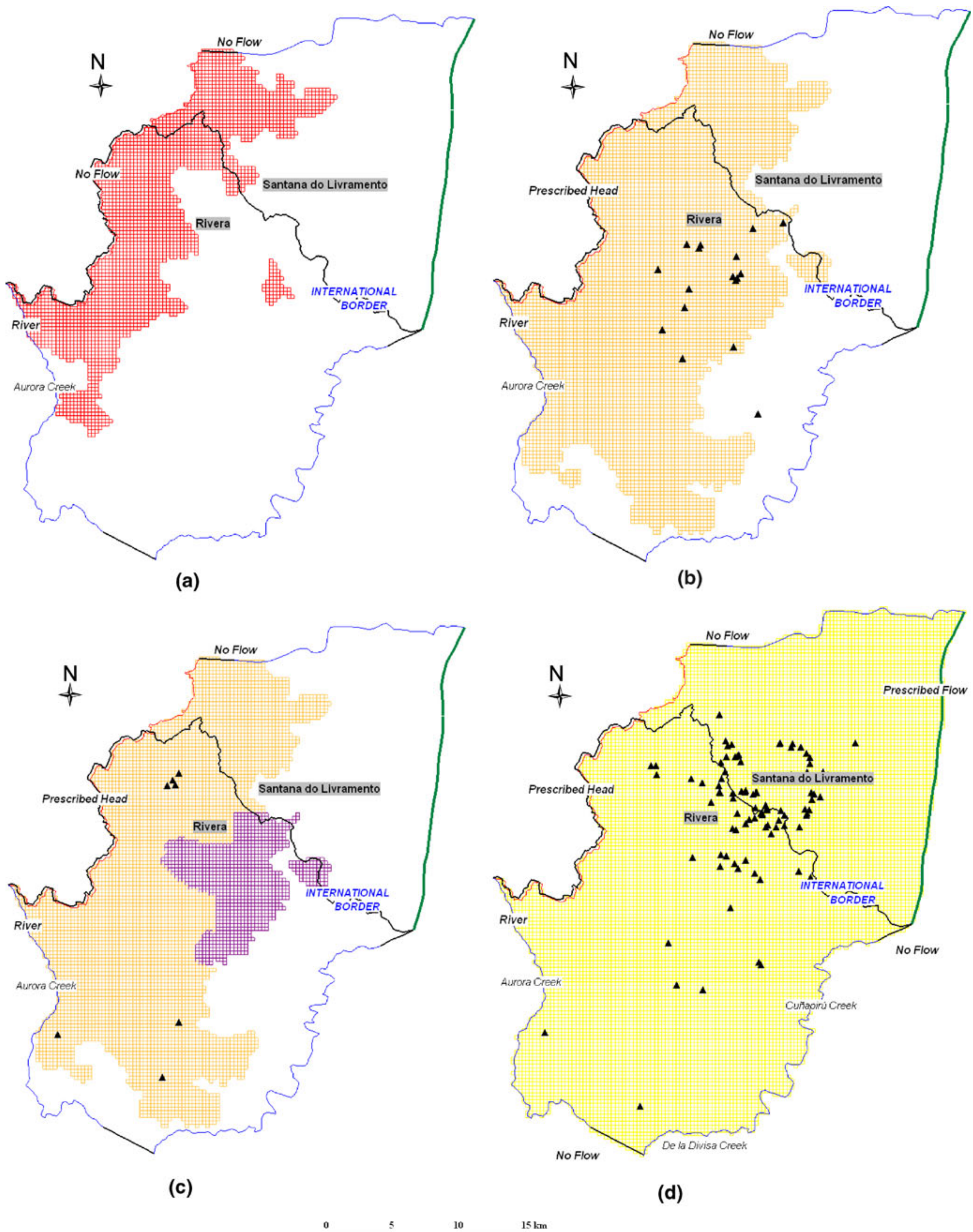

Fig. 9 Finite difference grid, pumping wells (*triangles*) and boundary conditions used in each layer. **a** Layer 1, basalts; **b** layer 2, shallow aquifer; **c** layer 3, shallow aquifer + aquitard (*violet cells*); **d** layer 4, deep aquifer

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was run to reproduce current abstraction-influenced groundwater flow conditions under a constant pumping scenario. Boundary conditions, recharge rates, stream/aquifer conductances and hydraulic conductivities were adjusted during the calibration process.

MODFLOW is not intended for use in fractured media but, in this case, the likely presence of connected fractures within basalts was simulated with an anisotropy ratio K_h/K_v=0.1, i.e. a vertical hydraulic conductivity value 10 times higher than horizontal hydraulic conductivity. Heine (2008) used a similar approach to model Serra Geral basalts in southern Brazil. Table 2 includes calibrated hydraulic conductivities for all layers. Calibrated K for sandstones were consistent with aquifer test values and K values reported in other regions of the GAS.

Water levels in the streams resulted in a simulated water depth of 0 m in elevated areas of the streams and 0.5 m in the rest of the drainage network. An uniform calibrated conductance, equal to 1 m/day, was obtained. Conductance values were in the same order as *K* for sandstone because a big portion of the study area is contained within the areas where slopes are high. Therefore, streams may not have low-conductance-streambed sediments typical of lower terrain where smaller slopes causes sedimentation processes from flash flows.

A total of 23 and 39 static levels for layers 2 and 4, respectively, were used for the calibration process (Fig. 10). Permissions from water companies were requested in order to temporarily shut down selected wells so as to obtain close-to-static conditions. In spite of this precaution, the reliability of some field data was questionable as wells are not cased in many cases, resulting in an integrated reading rather than an unique water level that could be easily associated with a particular layer for use in model calibration. The scatter plot of the goodness of fit for layer 4 is presented in Fig. 11. The 95% confidence interval, the regression line and the correlation coefficient were included. The mean error (ME) between computed and measured heads for layer 4 was −2.45 m. The mean absolute error (MAE) and the root mean square error (RMSE) were 4.66 and 6.3 m, respectively. The maximum positive error was 11.3 m at a well located to the west (the one shown in Fig. 4), the maximum negative error was –16 m. It is worth noting that 7 out of the 39 observation points resulted with errors well beyond calibration standards. If those points are not considered in the analysis, calibration results improved considerably: ME=–0.96 m; MAE=3.02 m and RMSE = 3.77 m. Factors that may have contributed to those outliers include non-static conditions at the time of surveying, wells not cased, and readings representing multiple aquifers; therefore, a well assigned to layer 4 may belong to another layer. Given a 73.5 m difference between maximum and minimum observed water levels, a normalized RMSE of 5% renders the model calibration within standards. The lowest head values on the bottom left of Fig. 11 were closely reproduced by the model, corresponding to an incipient cone of depression located on the Brazilian side.

Table 2 Calibrated hydraulic conductivities (*K*). K_h horizontal *K*; K_v vertical *K*; A_v vertical anisotropy

Layer	Aquifer	K_h (m/day)	$A_v=K_h/K_v$
1	Basalt	0.168	0.1
2	Shallow sandstone	1.5	500
3	Aquitard	0.09	1,000
3	Shallow sandstone	1.5	500
4	Deep sandstone	0.4–5.0	10

The classical approach of groundwater model calibration or inversion used to predict recharge rates from information on water levels, hydraulic conductivity and other parameters may lead to non-unique modelling results (Scanlon et al. 2002). Hydraulic conductivity and recharge rates are often highly correlated; consequently calibration based only on water level data is limited to estimating the ratio of recharge to hydraulic conductivity. Hence, as Scanlon et al. (2002) stated, the reliability of recharge estimates depends on the accuracy of the hydraulic conductivity data. Due to a reasonable number of pumping test data and to the fact that hydraulic conductivity range does not vary more than one order of magnitude in the study area, it was assumed that hydraulic conductivities are rather representative and recharge rates were the main calibration parameter. Besides, a comparison was made between *K* values for sandstones in the study area and in other areas of the aquifer in order to build confidence on model results. Calibrated recharge rate over basalts resulted in a 1.3% of mean annual precipitation; recharge rate over sandstones resulted in 8.1%.

The water budget for the entire model is summarized in Table 3 and Fig. 12, which should be assessed qualitatively rather than quantitatively due to limited field data to quantify budget terms independently. In the table, top and bottom flows are explicitly reported; in the figure, net vertical flow between layers is presented. According to observed water levels, there could be downward flow between layers. The simulated direction of net flow terms between layers driven by vertical hydraulic gradients reproduces this situation. Those flows are indicated as net downward flow (NDF) in Fig. 12 and differs from recharge *R* in the sense that the former represents an exchangeable flux between layers, i.e. an internal flux or indirect recharge, while *R* represents direct recharge from precipitation.

Simulated stream/aquifer flow (STR) totaled 222.5×10^3 m^3/day. Compared to other terms of the mass balance, it amounts to just 2.57 m^3/s from the aquifer system to streams, mainly concentrated in outcropping areas of the fourth layer, which contains Cuñapirú Creek, one of the main permanent streams in the study area.

Flows across prescribed head and prescribed flow boundaries were the most uncertain terms of the water budget and should be reassessed in future model improvements. In Fig. 12, the term prescribed head (PH) is shown as a net value with the resulting flow direction, in or out of the corresponding model layer. The prescribed flow

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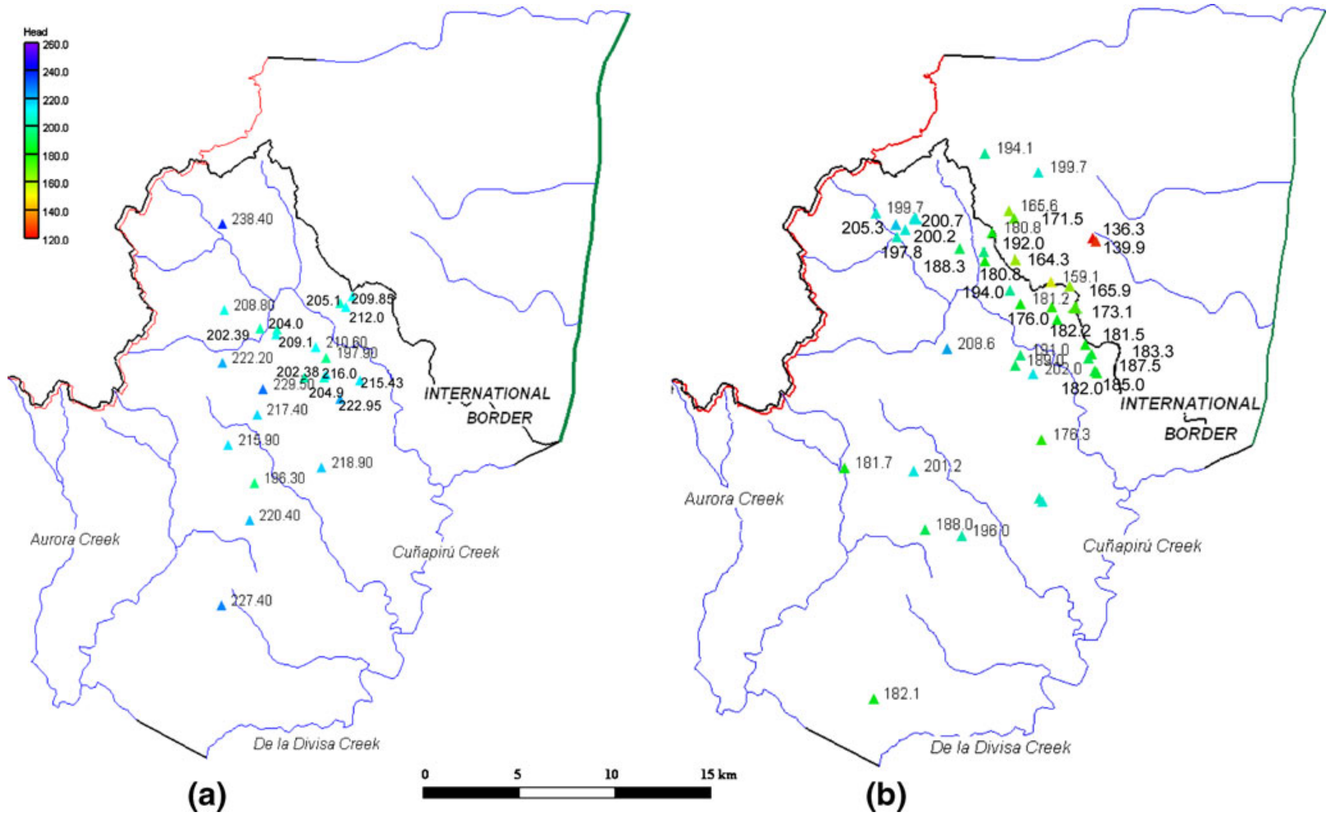

Fig. 10 Observed wells (*triangles*) and their respective piezometric head levels. **a** Shallow aquifer (model layer 2). **b** Deep aquifer (model layer 4). Head data in meters above sea level (m.a.s.l.)

boundary condition in layer 4 equaled 13.7×10^3 m^3/day, a small component within the global water budget. Total calibrated recharge resulted in 271.6×10^3 m^3/day, 8.4×10^3 m^3/day of which occurs over basalts. The results of Fig. 12 would indicate that the aquifer drains most of its recharge through streams. Recharge (represented by R in Fig. 12) was applied to the uppermost active cell in the layered system. Because some small-thickness cells in layer 2 became dry, MODFLOW applied the corresponding recharge to the cell immediately underneath (layer 3). The

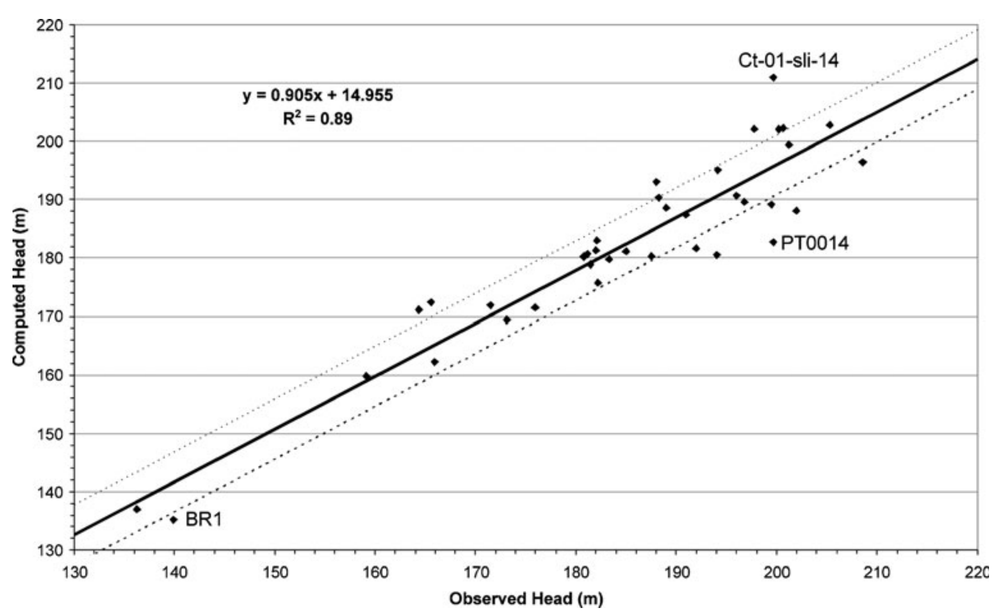

Fig. 11 Comparison between observed and computed water levels for layer 4. *Dotted line* indicates 95% confidence interval. The *three marked wells* are identified in Fig. 10

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Table 3 Groundwater model water budget. *STR* stream/aquifer flux; *R* recharge; *P* pumping; *PH* prescribed head flux; *PF* prescribed flow flux; *bottom* flux through layer bottom; *top* flux through layer top (fluxes × 10^3 m^3/day)

Layer	STR	*R*	PH	PF	Bottom	Top	Total
Input flows							
1	0	8.4	0	0	3.0	—	11.4
2	0	54.6	14.5	0	13.1	11.2	93.4
3	0	49.4	27.4	0	25.0	57.1	158.9
4	0	159.2	7.4	0	—	138.3	304.9
Total	0	271.6	49.3	0			
Output flows							
1	0.2	0	0	0	11.2	—	11.4
2	32.2	1.1	0	0	57.1	3.0	93.4
3	4.8	2.7	0	0	138.3	13.1	158.9
4	185.3	69.3	11.6	13.7	—	25.0	304.9
Total	222.5	73.1	11.6	13.7			

recharge estimated with EASY-BAL was 286.3 × 10^3 m^3/day. Although both methods were based upon very different equations and algorithms, they both yielded similar, but independently estimated recharge values. The next section presents a comparison between the recharge rates determined in the present work and previously published values.

Comparison with previous recharge estimates

Previous researchers have estimated direct recharge in various regions of the GAS using mainly water-balance models. Annual average precipitation for each site was used to compute the percentage of precipitation contributing to recharge. Table 4 compiles those estimated percentages, itemized by region.

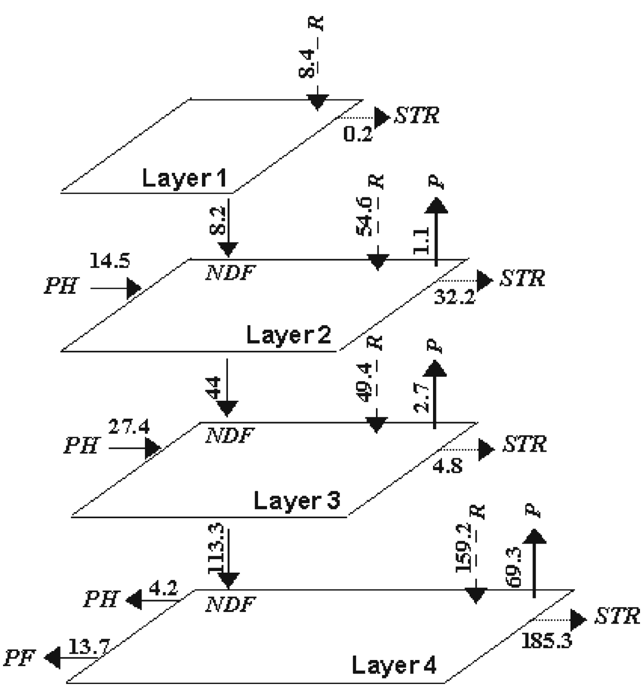

Fig. 12 Model water budget. *R* recharge; *P* pumping; *PH* prescribed head; *PF* prescribed flux; *STR* streams; *NDF* net downward flow (fluxes expressed in 10^3 m^3/day)

The water-balance analyses developed by the different authors were carried out under different hypotheses. For example, Chang (2001) took into account, not only the recharge from precipitation, but also a small amount of recharge coming from underlying formations that confine the GAS. Pacheco (2004) calculated a horizontal water balance using two flux lines and obtained the direct recharge. Collazo (2006) performed a water balance considering the soil characteristics, using the Rivera Station data and obtained an estimate of direct recharge. More recently, Wendland et al. (2007) used two methodologies; first they applied the water-level fluctuation method to obtain direct recharge and then this value was used to estimate the deep recharge through a traditional water balance.

Regarding the modeling methods, the three cited works (Vives et al. 2001; Vassolo 2007 and this work) calibrated the recharge term considering direct recharge from precipitation applied to outcropping areas. Recharge rates found during this research using two independent methods were similar though slightly higher than previous estimates for the study area. However, they were in tune with calculations in other outcropping areas of the GAS located in Paraguay and Brazil.

Expressed as percentage of precipitation, model calibrated recharge rates were approximately 2–3 times higher than water balance model recharge rates. The diversity of methods and hypotheses used would preclude any conclusive remark about the goodness of either method. However, except for the early work of Rebouças (1976) and the recent work of Collazo (2006), the rest of the values are very consistent and provide an upper bound to recharge of the GAS.

Sensitivity analysis

A sensitivity analysis was performed, aimed at evaluating the influence of selected parameters on the model response. The groundwater model showed no sensitivity to the anisotropy ratio K_h/K_v of layer 1. Neither the RMSE nor the net downward flow between layers 1 and 2

Table 4 Recharge estimates comparison

City or Region	Author	Method	% PM
Entire GAS	Vives et al. (2001)	M	< 10
Uruguay–Brazil border	Montaño and Carrión (1990)	MWB	3
	Pacheco (2004)	MWB	3.6
	Collazo (2006)	MWB	24
	This work	SWB	8.5
	This work	M	8.1
Paraguay	Vassolo (2007)	M	9.1
Sao Paulo State, Brazil	Rebouças (1976)	MWB	15
	Chang (2001)	MWB	4
	Wendland et al. (2007)	SWB/WLF	3.5

M modeling; *SWB* serial water balance; *MWB* mean water balance; *WLF* water level fluctuations; P_M mean annual precipitation

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showed measurable changes for the testing range of the parameter.

The sensitivity of model results to recharge rates over basalts and outcropping sandstone was explored. Higher and lower values than the calibrated rate were defined. The model response was measured as a percent change with respect to the calibrated value for both, the vertical downward flow between layers 1 and 2, and the RMSE for layer 4 (Fig. 13a). A 50% reduction on the basalt recharge rate produced a 43% decrease in downward flow, a 10% reduction on the recharge rate produced a 5% reduction in downward flow, while a 10% increase produced an equal increment in downward flow. Note that even though percent changes are high, the magnitude of flux between layer 1 and 2, i.e. indirect recharge to GAS, was small compared to direct recharge from precipitation in outcropping areas. The RMSE for layer 4 remained almost unaffected.

Recharge over sandstones had a significant influence on flux terms and RMSE (Fig. 13b). The figure shows fluxes between sandstone layers reduced when the recharge rate decreased. Streams in layer 4 were most affected because, as seen in the water budget analysis, they drain a big portion of the aquifer recharge. The RMSE for layer 4 increased substantially in response to a 50% reduction in sandstone recharge rate. Changes in response to smaller changes on recharge rates stayed within 10%.

Water levels and global STR flux manifested some effect in response to changes in stream conductance (Fig. 13c). Only smaller-than-calibrated conductance values were tested. RMSE was not significantly affected until the conductance value was reduced one order of magnitude with respect to its calibrated value. STR fluxes showed low sensitivity to this parameter, with percent changes below 5% for all tested cases.

Conclusions

The Guarani Aquifer System (GAS), one of the largest transboundary aquifers of the world, is being increasingly exploited for freshwater supply, and industrial and agricultural uses. Therefore, groundwater recharge rates are needed to address sustainability. This is especially pressing at local scale sites where pumping is concentrated.

The objective of this study was to validate the postulated conceptual model and obtain an estimate of recharge rates within the flow system for the Guarani Aquifer in the Rivera-Santana transboundary area. The conceptual model, delineated from background information and field data collected during this study, resulted in a multiaquifer system composed of several units with significant water level differences between them, favoring the hypothesis of vertical, downward flows.

A multilayer, heterogeneous and anisotropic groundwater flow model, under steady state in equilibrium, was built to fulfill the objective. A monthly water balance was also applied to independently estimate recharge rates. Numerical model-calibrated parameters were coherent with previous estimates obtained from the literature for like-aquifers from studies conducted in other parts of the GAS, and from field data. In the absence of flow data of any kind in the study area, recharge rates were calibrated from information on water levels, hydraulic conductivity and other parameters. It is known that this approach may lead to non-unique modelling results. Hydraulic conductivity and recharge rates are often highly correlated; consequently calibration based only on water level data

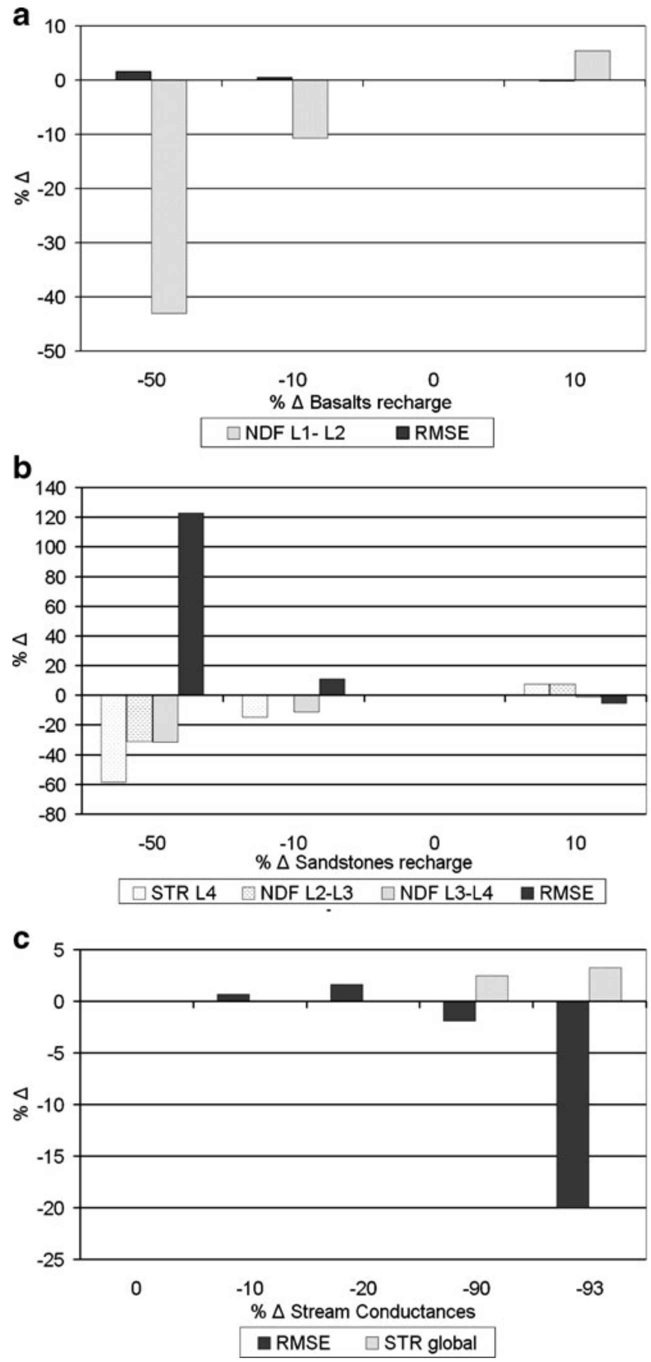

Fig. 13 Numerical model sensitivity analysis for the parameters. **a** Recharge over basalts; **b** recharge over sandstones and **c** stream conductances. *L* layer; *RMSE* root mean square error; *NDF* net downward flow; *STR* stream/aquifer fluxes

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is limited to estimating the ratio of recharge to hydraulic conductivity. Consequently, the reliability of recharge estimates depends on the accuracy of the hydraulic conductivity data. Given a reasonable number of pumping test data and field hydraulic conductivity values, constrained to a small-to-medium range available in the study area, it was assumed that hydraulic conductivities were rather representative and recharge rates were the main calibration parameter. Therefore, in spite of the uncertainties derived from the boundary conditions, the calibration was considered acceptable and within standard practice. The RMSE was close to 10%, highly influenced by a handful of wells for which the observed level may be considered questionable. Some of them were deep wells tapping several formations, probably having integrated water levels because wells are not cased. The RMSE without those outliers was less than 5% resulting in a calibration within standard practice.

The calibrated groundwater model revealed that simulated indirect recharge to sandstones represented in the model by the net downward flow from basalts would have a small magnitude in comparison with direct recharge from precipitation. A big portion of sandstone recharge would be drained by streams. The recharge rates over sandstones was 8.1% of the mean annual precipitation, which agree with the findings of similar studies conducted in the same region and in other parts of the same aquifer. The calibrated rate over basalts was 1.3% of the mean annual precipitation.

Measured in terms of changes in water level errors, layer-to-layer fluxes and stream–aquifer flux, model results were most sensitive to changes in recharge rates over basalts and sandstones. Even though the model constitutes a step toward improving the knowledge of the GAS in the area, it is recommended to verify the results with additional field data, mainly regarding boundary fluxes and baseflows on simulated streams. Converting the model to transient flow conditions would also improve recharge estimates. Unfortunately, that can not be implemented until time-varying water levels are monitored on a periodic basis.

On the other hand, a water balance was performed with monthly meteorological data from the Rivera Station. Average annual recharge for the period 1960–1997 was 8.5% of average annual precipitation. Model parameters such as useful reserve and runoff threshold were estimated based on physical characteristics of the basin, soil properties and streamflow measured outside the model area. Uncertainty regarding runoff threshold was evaluated through sensitivity analysis, showing that water-balance computations are affected by this parameter.

Even though the numerical model and the water balance are based upon very different equations and algorithms, they both yielded similar recharge values. A comparison with previous determinations in the area and other regions of the aquifer was performed to assess the consistency of the results. Recharge rates determined during this research were comparable to previous estimates obtained with the same methodologies providing an upper bound for recharge in this transboundary aquifer.

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Appendix

EASY-BAL equations

Variables definition

- I: monthly heat index
- L: annual heat index
- a: empiric coefficient
- e: daily mean potential evapotranspiration
- K: radiation coefficient
- PET: monthly potential evapotranspiration
- P: precipitation
- AR: available water from precipitation
- RU: monthly useful reserve
- RT: runoff threshold
- Initial reserve: depends on the soil type
- Useful reserve: depends on the soil type
- RET: real evapotranspiration

Thornthwaite equation

$$I = \left(\frac{T_{\mathrm{i}}}{5}\right)^{1,514},\ T_{\mathrm{i}} = \text{mean monthly temperature}$$

$$l = \sum_{\mathrm{n}=1}^{12} I_{\mathrm{n}}, n = \text{month of the year}$$

$$a = (0.000000675.l^3) - (0.0000771.l^2) + (0.01972.l) + 0.492l39$$

$$e = 16\left(10\frac{T^i}{l_{\mathrm{i}}}\right)^a$$

$$K = \left(\tfrac{N}{12}\right)\left(\tfrac{d}{30}\right),\ N = \text{daily sun hours for each month;}$$
$$d = \text{number of days in each month, considering inclusive leap years.}$$

$$\mathrm{PET} = \mathrm{K.e}$$

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Spreadsheet algorithms

1. If $P \geq$ RT, AR = RT–PET, otherwise AR = P–PET
2. If (AR + initial reserve) > useful reserve, RU = useful reserve; otherwise if (AR + initial reserve) < 0, RU = 0, otherwise RU = (AR + initial reserve)
3. Monthly water deficit/excess = AR + RU of the previous month − RU
4. If (PET + deficit/excess) > PET, RET = PET, otherwise RET = (PET + deficit/excess)
5. Recharge: If deficit/excess ≤ 0, recharge = 0, otherwise if deficit/excess > useful reserve, recharge = deficit/excess – useful reserve, otherwise recharge = deficit/excess

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