Estimation of regional evapotranspiration in the extended Salado Basin (Argentina) from satellite gravity measurements

Andrés Cesanelli · Luis Guarracino

Abstract In this study, regional evapotranspiration is estimated in a wide flatland area that includes Salado River Basin and four tributary basins by using gravity measurements of the space mission Gravity Recovery and Climate Experiment (GRACE). Monthly estimates of large-scale variations in the land-water storage are obtained from the satellite data. Evapotranspiration is computed with the water-balance equation using the GRACE land-water solutions, rainfall data from the Global Precipitation Climatology Center and runoff values obtained as 5% of the precipitation. GRACE-derived evapotranspiration values are consistent with the different climatic scenarios observed, and they satisfactorily agree with estimates provided by a global hydrological model. The overall results show that the method used is a valid tool for characterizing the evapotranspiration in the Argentine Pampas and that it can be used to detect and examine changes in the evapotranspiration pattern associated with the occurrence of extreme climatic events. This study illustrates the ability of GRACE to analyze and predict evapotranspiration and other processes on a regional scale in a flatland area.

Keywords Evapotranspiration · Hydrogeophysics · GRACE · Remote sensing · Argentina

Introduction

Salado River is the major watercourse in Buenos Aires province and receives inflows of surface water and groundwater from a wide flatland area of the Argentine Pampas. The main activities in this region are agriculture and cattle breeding, which strongly contribute to the national economy. However, the alternation of periods with excess and deficit of precipitation leads to the potential occurrence of flood and drought events that affect both the environment and human activities (Scarpati et al. 2002; Herzer 2003). Due to the very low surface slopes and the poorly developed drainage system, water is mainly discharged by evapotranspiration (ET; Varni and Usunoff 1999). This process plays a key role in the hydrologic behavior of the region since it regulates the development of surface waterlogging (Vázquez et al. 2007). Despite all the efforts made to characterize the regional ET in this area, the estimation of ET on large scales is still a difficult task, mainly due to the lack of available hydrological and meteorological data.

Satellite remote sensing tools has been acknowledged as being very useful for improving the modeling of hydrological processes on a large scale (Brunner et al. 2007). In particular, these tools can provide information to estimate ET over large areas (Couralt et al. 2005). For example, ET can be obtained by using the energy balance equation considering surface temperature measurements from satellite images. Other methods take advantage of the relationship between ET and vegetation indices such as the normalized difference vegetation index (NDVI), which describes plant growth and coverage. These kinds of techniques have been implemented in different places in the Argentine Pampas to derive models that compute ET (Di Bella et al. 2000; Rivas and Caselles 2004). Unfortunately, these models cannot be applied to the entire region because they require a location-specific calibration. However, a new method to obtain regional ET has been made available since the launch of the Gravity Recovery and Climate Experiment (GRACE) space mission in March 2002. GRACE provides gravity measurements that can be used to quantify the changes in the land-water storage on a large scale (Wahr et al. 1998). ET values are determined by using the water-balance equation, which combines the GRACE-derived land-water solutions and information on precipitation and runoff. Using this method, it is possible to obtain estimates comparable to the ET rates provided by different global models for large drainage basins (> 5×10^5 km^2) distributed over the globe (Rodell et al. 2004a; Ramillien et

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al. 2006; Bonorina and Ramillien 2008). Although water storage is a key component of the hydrological cycle, until a short time ago its estimation was made essentially from modeling due to the lack of adequate observation networks. GRACE means an important advance towards characterizing water balance on large scales since it is the first remote sensing tool that allows for obtaining measurements of the water storage with global coverage. The aim of this study is to use this new method to estimate the monthly ET in a region that includes the Salado River Basin and other four tributary basins. This region, hereinafter referred to as the *extended Salado Basin*, is characterized by its relatively flat surface and covers approximately 3×10^5 km^2.

The GRACE satellite mission is jointly managed by the National Aeronautics and Space Administration (NASA) and the Deutsches Zentrum für Luft und Raumfahrt (DLR). Its main objective is to measure spatio-temporal variations of the terrestrial gravity field. GRACE determines the gravity potential with high precision by using a pair of identical satellites moving in a nearly polar orbit at an initial altitude of 485 km (Tapley et al. 2004). The gravity potential is described by a spherical harmonic expansion whose coefficients (Stokes coefficients) are computed from the perturbations observed in the satellite movement. At annual and sub-annual time scales, the gravity variations are mainly caused by the mass redistribution inside the fluid envelopes of the Earth (atmosphere, oceans, continental reservoirs). After removing atmospheric and oceanic effects from the satellite data, it is possible to compute the land-water-storage variations with a monthly resolution. These variations are usually expressed in terms of equivalent water thicknesses and can be used to analyze the land-water dynamic. Several researchers have shown the utility of GRACE measurements in hydrology and climatology studies (Andersen et al. 2005; Syed et al. 2005; Rodell et al. 2007; Morishita and Heki 2008; Yirdaw et al. 2008).

In this study, temporal changes in water storage are quantified using the method described by Wahr et al. (1998) and the filtering technique proposed by Swenson and Wahr (2006). The filtering method is used to eliminate systematic errors in the short wavelength components (i.e. high-degree Stokes coefficients) of the gravity field. Its use yields reliable estimates of water-storage variations in regions with a similar size as that of the extended Salado Basin (Swenson et al. 2006, 2008). Using the available Stokes coefficients, monthly water-storage variations are computed for the period May 2002–October 2009. These values show a similar pattern to that of the water-table variations observed at different wells in a 2-year period. Rainfall data from the Global Precipitation Climatology Center (GPCC)—based on measurements of a large number of rain gauges (Rudolf et al. 1994)—has been used to obtain the regional ET. Since the surface of the region is approximately flat, runoff is very low and thus it is modeled as a small fraction of the rainfall using a linear function. The ET values obtained show a temporal variation consistent with the different climatic scenarios observed. These values are compared to the ET estimates of the Global Land Data Assimilation System (GLDAS), which are generated from the combination of land-surface models and climate forcing (Rodell et al. 2004b). In addition, this global model is used to analyze the validity of the model considered for runoff due to the lack of in situ measurements. The satisfactory agreement found between the time series of ET obtained with GRACE and GLDAS shows that the applied method is a valid tool for estimating ET in the Argentine Pampas.

Material and methods

In this section, the study area as well as the method and data used for estimating regional ET are presented. The water-balance equation and the expressions required to estimate water-storage variations are briefly described. Further details of the quantification of water storage from GRACE data can be found in Wahr et al. (1998) and Swenson and Wahr (2006).

Study area

The extended Salado Basin comprises a wide region extending over the provinces of Buenos Aires, La Pampa, San Luis, Córdoba and Santa Fe. This drainage basin includes five major sub-basins: Salado River Basin, Southern Channels Region, Las Encadenadas Lakes System, Arreic Region and Quinto River Basin. The boundaries of the five sub-basins are defined in the Digital Atlas of the Hydrologic Surface Resources of Argentina (SRHN and INA 2002) and shown in Fig. 1. Most of the region is a large plain with very low surface slopes, ranging mainly between 10^{-3} and 10^{-4} (Sala et al. 1983; Tanco and Kruse 2001). The climate is temperate with the largest percentage of annual rainfall during the summer (Herzer 2003; Viglizzo et al. 2009). Rainfall rates are approximately 1,100 mm/year on the coast of La Plata River and decline from NE to SW (Kruse and Laurencena 2005).

From a geological point of view, the extended Salado Basin is a sedimentary basin where Cretaceous, Tertiary and Quaternary sediments are superimposed over the Precambrian crystalline basement (Santa Cruz and Silva Busso 1999). The uppermost two layers, namely Pampeano and Puelche formations, are the main source of groundwater exploitation in the region (Carbó et al. 2009). The Pampeano (Quaternary-age sediments) is composed of clayey and sandy silts (loess) and ranges in thickness between 20 and 120 m. On the other hand, the Puelche is a semiconfined aquifer composed of Tertiary-age sands which ranges in thickness from 10 to 25 m. On a regional scale, both formations constitute a multilayered aquifer mainly recharged by rainfall. Water tables are very shallow (generally less than 3-5 m deep) and groundwater flow is relatively small due to the poor hydraulic

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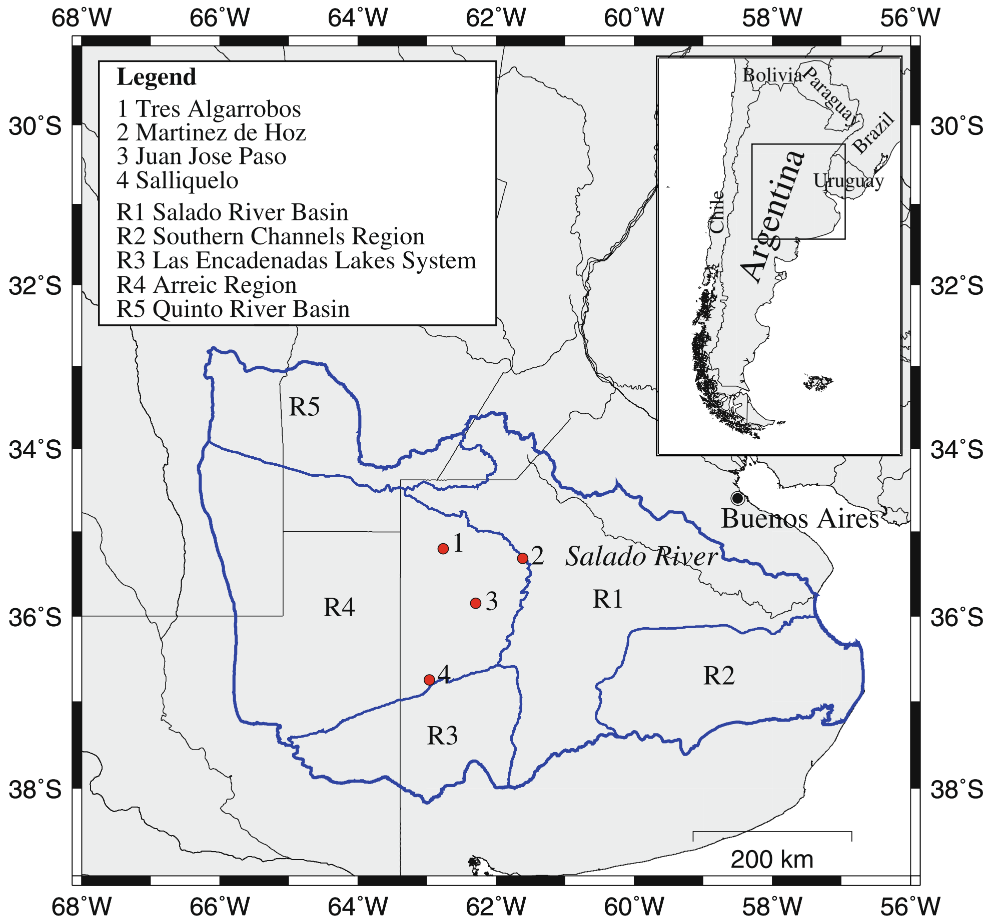

Fig. 1 The extended Salado Basin and the five major sub-basins (*blue lines*). *Red circles* indicate the location of the wells where water-table records are available

conductivity and the very gentle regional gradients (Aradas et al. 2002). As a result of the low regional slopes and the disintegrated drainage system, the surface is characterized by the presence of shallow lagoons and wetlands, while surface runoff is usually poor. Although local slopes can favour significant water flows, large-scale runoff rates are small (Viglizzo et al. 2009). Thus, infiltration towards deep aquifers plays an important role in the hydrological balance of the basin, mainly during periods of water excess (Carol et al. 2010).

Computation of regional ET using the water-balance equation

The instantaneous conservation of water mass in each point of a given watershed is stated to derive an expression of the water-balance equation:

$$\frac{\partial S(t)}{\partial t} = P(t) - R(t) - ET(t) \qquad (1)$$

where t is the time, S is the water volume in the storage, and P, R and ET are water fluxes representing precipitation, runoff and evapotranspiration, respectively. The water budget on a regional scale is obtained by integrating Eq. (1) in time and space:

$$\int_A \int_{t_i}^{t_{i+1}} \left[\frac{\partial S(t)}{\partial t}\right] dtdA = \int_A \int_{t_i}^{t_{i+1}} [P(t) - R(t) - ET(t)]\, dtdA \qquad (2)$$

where A is the basin area and t_i and t_{i+1} are, respectively, the beginning and end of the period considered for the water balance. In this study, ET is computed on a monthly basis and thus t_i and t_{i+1} are the beginning of month 'i' and month 'i'+1, respectively.

After dividing both sides of Eq. (2) by A and solving time integrals, the basin water balance is expressed as follows:

$$\Delta S^{i,A} = P^{i,A} - R^{i,A} - ET^{i,A} \qquad (3)$$

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where

$$\Delta S^{\mathrm{i,A}} = \frac{1}{A}\int_A \Delta S^{\mathrm{i}} dA \tag{4}$$

$$P^{\mathrm{i,A}} = \frac{1}{A}\int_A P^{\mathrm{i}} dA \tag{5}$$

$$R^{\mathrm{i,A}} = \frac{1}{A}\int_A R^{\mathrm{i}} dA \tag{6}$$

$$ET^{\mathrm{i,A}} = \frac{1}{A}\int_A ET^{\mathrm{i}} dA \tag{7}$$

where, at each point of the basin, P^{i}, R^{i} and ET^{i} are the monthly values of precipitation, runoff and evapotranspiration, and

$$\Delta S^{\mathrm{i}} = S(t_{\mathrm{i+1}}) - S(t_{\mathrm{i}}) \tag{8}$$

is the water-storage variation during month i. Note that each term in Eq. (3) is a water volume expressed in equivalent water thickness. According to Eqs. (4)–(7), if the basin area is approximated as the sum of N_{A} surface elements of size ΔA, these volumes can be estimated with the following expression:

$$F^{\mathrm{i,A}} = \frac{1}{A}\sum_{\mathrm{j=1}}^{N_{\mathrm{A}}} F^{\mathrm{i}}(\varphi_{\mathrm{j}}, \lambda_{\mathrm{j}})\Delta A_{\mathrm{j}} \tag{9}$$

where F is either P, R, ET or ΔS, and

$$\Delta A_{\mathrm{j}} = a_{\mathrm{j}}^2\, \Delta\lambda\, \Delta\varphi \cos(\varphi_{\mathrm{j}}) \tag{10}$$

where φ_{j}, λ_{j} and a_{j} are the latitude, longitude and distance from the origin of coordinates at the center of the surface element 'j', while $\Delta\varphi$ and $\Delta\lambda$ are the dimensions of the surface element in latitude and longitude (generally $\Delta\varphi = \Delta\lambda$). It is necessary to consider the curvature of the Earth in the estimation of ΔA since the basin has a large area. Therefore, parameter a_{j} is obtained in terms of latitude φ_{j} with the following expression (Hoffman-Wellenhof and Moritz 2005):

$$a_{\mathrm{j}} = a_{\mathrm{e}}\left(1 - \alpha \sin^2(\varphi_{\mathrm{j}})\right) \tag{11}$$

where a_{e} is the equatorial radius of the Earth (6,378,137 m) and a is the flattening of the reference ellipsoid (3.35281×10^{-3}).

Although the information available to quantify the runoff in the extended Salado Basin is poor, on regional scales this process is known to be usually less significant than water-storage variations (Carol et al. 2010). Moreover, long-term water budgets show that runoff represents only a small fraction of the mean annual rainfall (Varni and Usunoff 1999). Different studies indicate that this fraction is approximately 5% (Sala et al. 1982; Cacik et al. 2000; Auge 2001; Kruse and Laurencena 2005). Then, the following linear model is used to compute $R^{\mathrm{i,A}}$ (Kadioğlu and Şen 2001):

$$R^{\mathrm{i,A}} = k_{\mathrm{R}} P^{\mathrm{i,A}} \tag{12}$$

where k_{R} is a coefficient ranging from 0 to 1. In this study, a constant value of 0.05 is assumed for this coefficient. Temporal variations of k_{R} are not considered since they may introduce minimal changes in the monthly water balance. Nevertheless, given that these small variations are not explicitly quantified, they must be considered a source of uncertainty in the estimation of ET.

Finally, combining Eqs. (3) and (12) yields the following expression to compute the regional ET:

$$ET^{\mathrm{i,A}} = P^{\mathrm{i,A}}(1 - k_{\mathrm{R}}) - \Delta S^{\mathrm{i,A}} \tag{13}$$

Considering that precipitation is easily obtained from a global dataset (e.g. GPCC), GRACE provides a method to solve Eq. (13) since $\Delta S^{\mathrm{i,A}}$ can be estimated from the gravity measurements. GRACE-derived values are a measure of the vertically integrated mass variations within surface reservoirs, the unsaturated zone, aquifers, snow and ice packs, and the biomass. As the extended Salado Basin has no snow or ice reservoirs, the water-mass variations obtained with GRACE are mainly due to fluctuations of the surface waters, groundwater and soil moisture. The method used to compute these variations is described in the next sub-section.

Computation of water-storage variations from GRACE

The temporal pattern of water storage is determined from the changes in the Stokes coefficients. Local equivalent water heights which represent anomalies (i.e. deviations

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from a reference value) of the water storage are obtained by using these coefficients. For a given month i, the water-storage anomaly at a point of latitude φ and longitude λ is (Wahr et al. 1998):

$$H^{\mathrm{i}}(\varphi,\lambda) = \frac{GM}{a_{\mathrm{e}}\gamma(\varphi)} \times \sum_{\mathrm{n}=1}^{N_{\max}}\sum_{\mathrm{m}=1}^{\mathrm{n}} P_{\mathrm{nm}}(\cos\theta)\left(\Delta C^{\mathrm{i}}_{\mathrm{nm}}\cos(\mathrm{m}\lambda) + \Delta S^{\mathrm{i}}_{\mathrm{nm}}\sin(\mathrm{m}\lambda)\right) \tag{14}$$

where

$$\begin{Bmatrix} \Delta C^{\mathrm{i}}_{\mathrm{nm}} \\ \Delta S^{\mathrm{i}}_{\mathrm{nm}} \end{Bmatrix} = \frac{\rho_{\mathrm{e}}(2n+1)}{3\rho_{\mathrm{w}}(1+k_{\mathrm{n}})} \begin{Bmatrix} C^{\mathrm{i}}_{\mathrm{nm}} - \overline{C}_{\mathrm{nm}} \\ S^{\mathrm{i}}_{\mathrm{nm}} - \overline{S}_{\mathrm{nm}} \end{Bmatrix} \tag{15}$$

where $\theta = 90^{\circ} - \varphi$ is the colatitude, GM is the product of the universal gravitational constant and the mass of the Earth (3.986005×10^{14} m^3 s^{-2}), γ is the normal gravity, P_{nm} are the normalized associated Legendre functions of degree 'n' and order 'm' ($N_{\max}$ is the maximum degree), ($C^{\mathrm{i}}_{\mathrm{nm}}$,$S^{\mathrm{i}}_{\mathrm{nm}}$) and ($\overline{C}_{\mathrm{nm}}$,$\overline{S}_{\mathrm{nm}}$) are, respectively, the Stokes coefficients of degree n and order m for gravity field of month i and for mean gravity field (obtained as the time average of the available coefficients), ρ_{e} is the average density of Earth (5,517 kg/m^3), ρ_{w} is the density of water (1,000 kg/m^3), and k_{n} is the load Love number of degree n. Love numbers have been introduced to consider the Earth's elastic compensation generated by surface loads due to mass redistribution. In this study, the k_{n} values are obtained from Wahr et al. (1998), which are based on the structural parameters of the preliminary reference Earth model (PREM) of Dziewonski and Anderson (1981).

The Stokes coefficients provided by GRACE have errors that produce a short wavelength noise (i.e. long, linear features called stripes) on the global maps of water-storage anomaly. To attenuate this noise, the coefficients obtained with Eq. (15) are filtered using the method proposed by Swenson and Wahr (2006). These authors have found that the presence of stripes is related to certain correlation between the coefficients, and designed a spectral filter that removes most of the correlated errors without a significant degradation of the geophysical signal. A Gaussian filter, which eliminates residual errors by smoothing high-degree coefficients, is then used to complete the post-processing of the data. The smoothing radius of the Gaussian filter must be selected according to the size of the study area to minimize the effect of unwanted signals from nearby regions (Rodell et al. 2004a). In their work, Swenson et al. (2006) have suggested using a smoothing radius of 300 km when the area is approximately 2.8×10^5 km^2. However, as solutions obtained with this radius contain significant striping in the extended Salado Basin, a 400 km radius is considered for higher smoothing and efficient noise reduction.

Regional water-storage variations are obtained from anomaly values computed with Eq. (14). Given that the difference between S and H is a constant value, the following equation can be derived from Eq. (8):

$$\Delta S^{\mathrm{i}} = H(t_{\mathrm{i+1}}) - H(t_{\mathrm{i}}) \tag{16}$$

where the right-hand side is the difference between the values of the water-storage anomaly at the beginning of months i and i+1. Since GRACE estimates of water storage are not instantaneous, but rather monthly solutions, the value of H at time t_{i} can be approximated by averaging the estimates obtained for months i-1 and i (Ramillien et al. 2006):

$$H(t_{\mathrm{i}}) \approx \frac{1}{2}\left(H^{\mathrm{i}-1} + H^{\mathrm{i}}\right) \tag{17}$$

Finally, the subtitution of Eqs. (16) and (17) in Eq. (4) yields the following expression to compute the water-storage variation in the basin:

$$\Delta S^{\mathrm{i,A}} = \frac{1}{2}\left(H^{\mathrm{i+1,A}} - H^{\mathrm{i-1,A}}\right) \tag{18}$$

where

$$H^{\mathrm{i,A}} = \frac{1}{A}\int_A H^{\mathrm{i}}dA \tag{19}$$

is the water-storage anomaly spatially integrated in the basin surface. Note that Eq. (19) is analogous to Eqs. (4)–(7) and thus $H^{\mathrm{i,A}}$ is computed using the expression proposed in Eq. (9).

Data

Water-storage variations are computed using Level 2 GRACE data (Release 04) determined by the Center for Space Research (CSR), University of Texas. Non-hydrological gravitational contributions (i.e. atmospheric and ocean circulation, and solid Earth tides) were removed from these coefficients (Bettadpur 2007). The available data consist of 89 sets of coefficients up to degree and order $N_{\max}$=60. These data comprise the period between April 2002 and November 2009, except 3 months: June 2002, July 2002 and June 2003.

The values of monthly precipitation are estimated using rainfall data provided by the GPCC. These data are computed from rain gauge observations and are available on regular grids at monthly resolution. The products generated by the GPCC have been checked using a high-level quality control process in order to be used in climate-related research and monitoring activities (Rudolf and

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Schneider 2005). Monthly 1.0º×1.0º grids for the period 2002–2009 are especially used in this study.

Uncertainties

The uncertainties in monthly estimates of ET are computed by error propagation through Eq. (13). According to the expressions presented in Rodell et al. (2004a), the relative uncertainty in a monthly ET value is:

$$\upsilon_{\mathrm{ET}} = \frac{\sqrt{\left(\upsilon_{\mathrm{P}}^{2} + \upsilon_{\mathrm{k_R}}^{2}(k_{\mathrm{R}})^{2}\right)(P^{\mathrm{i,A}})^{2} + \upsilon_{\Delta\mathrm{S}}^{2}(\Delta S^{\mathrm{i,A}})^{2}}}{P^{\mathrm{i,A}}(1-k_{\mathrm{R}}) - \Delta S^{\mathrm{i,A}}} \tag{20}$$

where υ_{P}, $\upsilon_{\mathrm{k_R}}$ and $\upsilon_{\Delta\mathrm{S}}$ are the relative uncertainties in the monthly precipitation, runoff coefficient and GRACE land-water-storage variation, respectively. Given υ_{ET}, the 95% confidence limits on ET are computed as$\pm\upsilon_{\mathrm{ET}}\, ET^{\mathrm{i,A}}$.

The term $\upsilon_{\Delta\mathrm{S}}\ \Delta S^{\mathrm{i,A}}$ is the absolute error of a GRACE estimate of monthly water-storage variation. This value is obtained multiplying the error in water-storage anomaly by $\sqrt{2}$ to account for month-to-month variations. Different studies have shown that the mean standard error in the monthly H estimates can be considered approximately between 11 and 18 mm (Swenson et al. 2006; Strassberg et al. 2007; Swenson et al. 2008). A value of 15.0 mm is chosen here for the error in $H^{\mathrm{i,A}}$, and therefore $\upsilon_{\Delta\mathrm{S}}\ \Delta S^{\mathrm{i,A}}$ is assumed to be 21.2 mm. Relative uncertainties of GPCC data in the basin are not well characterized at present. Mota (2003) found differences between GPCC data and rain gauge measurements ranging from 1 to 10% in a large region of Argentina. Although GPCC data are produced considering local precipitation information, rainfall values from 20–26 rain gauges of the National Meteorological Service (A. Gómez, Meteorological Data Center – National Meteorological Service of Argentina, personal communication, 2010) located in the basin are used to evaluate the performance of the database. These measurements are averaged to obtain monthly rainfall values for the period 2002–2009. The root mean square error (RMS) between both GPCC and observed data is 7.5 mm/month (90 mm/year), which represents approximately 10% of the annual rainfall. Therefore, a value of 0.1 is chosen for υ_{P}. On the other hand, a realistic value for the relative uncertainty on simulated estimates of runoff can be 30% (Ramillien et al. 2006). However, this parameter must be larger to include the errors associated with the temporal variability of k_{R} in the estimation of the ET error, and thus a value of 0.5 is assumed for $\upsilon_{\mathrm{k_R}}$. Note that even though the selected $\upsilon_{\mathrm{k_R}}$ is large, the uncertainty in runoff can be considered within the range of error of both rainfall and GRACE estimates.

Results

In this section, the expressions and data presented previously are used to estimate the regional evapotranspiration in the extended Salado Basin. The time series of $P^{\mathrm{i,A}}$ values obtained from GPCC data are illustrated in Fig. 2, which shows the seasonal pattern of the rainfall, with the largest values (greater than 100 mm) occurring mainly during the summer. The annual accumulated values of precipitation, which show a maximum (≈1,060 mm) in 2002 and two minimums (≈715 mm) in 2008 and 2009, are presented in Table 1. To quantify the water-storage variations in the basin, the procedure described in Computation of water-storage variations from GRACE is applied. Using the available Stokes coefficients, monthly 1.0º×1.0º grids of water-storage anomaly are computed for the period April 2002–November 2009. When data are not available, monthly land-water solutions are interpolated based on values corresponding both to the previous and following months (Ramillien et al. 2006). The time series of $H^{\mathrm{i,A}}$ and $\Delta S^{\mathrm{i,A}}$ values obtained are shown in Fig. 2. During most of the study period, $H^{\mathrm{i,A}}$ ranges between −50 and 50 mm. The largest deviations from the mean water-storage status ($H^{\mathrm{i,A}}$=0) occur in 2002 and

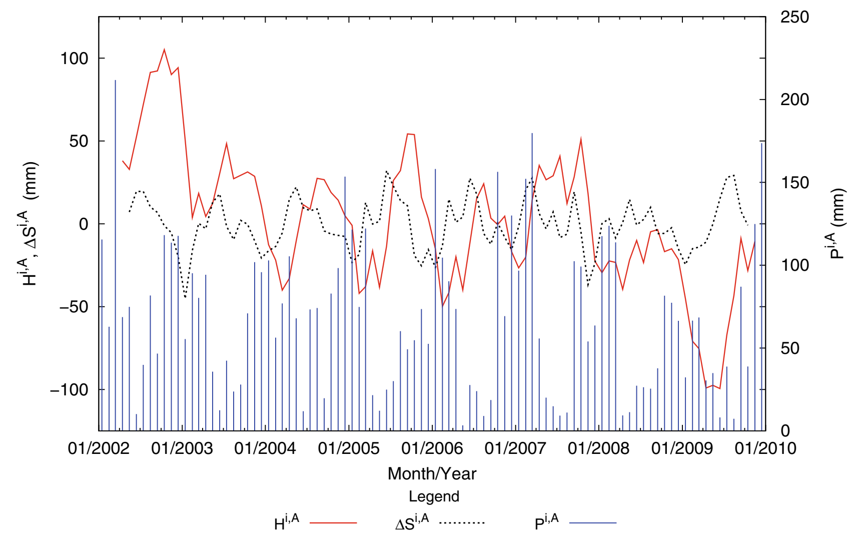

Fig. 2 Monthly values of water-storage anomaly ($H^{\mathrm{i,A}}$) and water-storage variations ($\Delta S^{\mathrm{i,A}}$) from GRACE data, and precipitation ($P^{\mathrm{i,A}}$) from GPCC data

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Table 1 Annual values of precipitation ($P^{i,A}$), runoff ($R^{i,A}$), water-storage variation ($\Delta S^{i,A}$) and evapotranspiration ($ET^{i,A}$) derived from GPCC, GLDAS and GRACE data

Year	$P^{i,A}{}_{GPCC}$ (mm)	$R^{i,A}{}_{GLDAS}$ (mm)	$\Delta S^{i,A}{}_{GRACE}$ (mm)	$ET^{i,A}{}_{GRACE}$ (mm)	$ET^{i,A}{}_{GLDAS}$ (mm)
2002[a]	(1,061.6) 602.8	(164.9) 122.0	36.9	535.7	(935.0) 560.8
2003	735.9	42.3	–73.3	772.4	743.5
2004	923.1	41.0	2.9	874.0	795.6
2005	697.4	36.9	–7.9	670.5	736.2
2006	864.7	30.8	–15.7	837.2	735.6
2007	858.9	54.6	–4.3	820.2	812.6
2008	718.2	29.5	–7.4	689.7	689.3
2009[b]	(711.4) 412.9	(20.5) 11.4	14.0	378.2	(488.2) 338.0

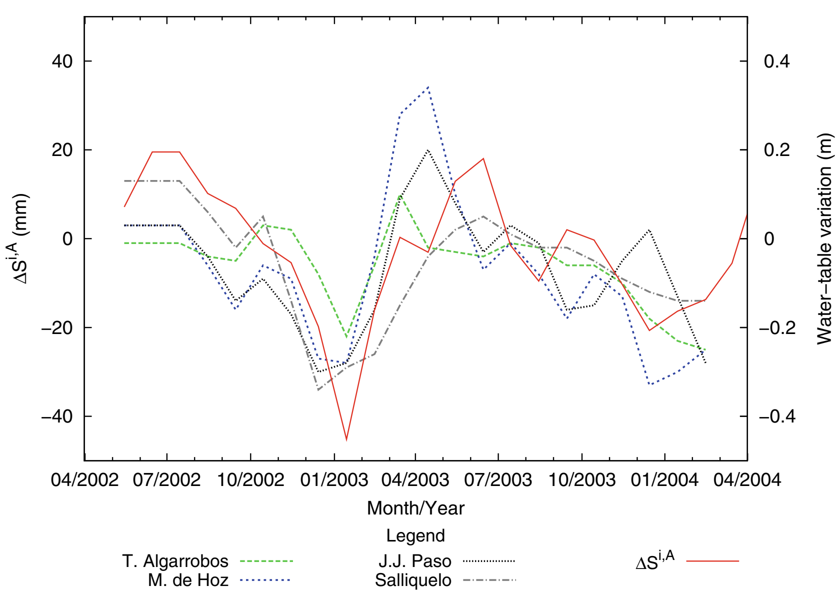
[a] For 2002, the period May to December was considered. Values in parentheses correspond to the period January to December

[b] For 2009, the period January to October was considered. Values in parentheses correspond to the period January to December

2009, when the water-storage anomaly reaches values of approximately 100 and –100 mm, respectively. These situations are associated with periods with excess and deficit of water featuring important events of flooding (Viglizzo et al. 2009) and drought (Sacchi et al. 2008). In the case of water-storage variations, monthly values generally range between –30 and 30 mm. As expected, the recharge of the water storage ($\Delta S^{i,A}>0$) is mainly observed during the winter, when water loss by ET is generally minimum. The largest water-storage variation for a single month occurs in January 2003 ($\Delta S^{i,A}=-45$ mm), after the flooding event of 2002. Long-term changes in the water storage are analyzed from the annual values of $\Delta S^{i,A}$, which are presented in Table 1. During most of the study period, these values are negative, which indicates an average depletion in the water storage (approximately 0.9% of the annual rainfall).

In recent years, many studies have shown the utility of GRACE for monitoring groundwater-storage changes (Yeh et al. 2006; Strassberg et al. 2007; Swenson et al. 2008; Rodell et al. 2009). Aquifer recharge can be estimated from the GRACE-derived water-storage variations using ancillary information (e.g. surface water and soil moisture changes). Unfortunately, hydrological monitoring networks are virtually nonexistent in the Argentine Pampas and thus there are few opportunities to make this type of analysis. Nevertheless, since water tables are usually very shallow, the surface and subsurface water storage can be treated as a hydrological unit (Tanco and Kruse 2001). Moreover, as a consequence of the low regional slopes, groundwater variations directly affect the entire hydrologic system (Kruse and Laurencena 2005). A set of well records covering a 2-year period (April 2002–March 2004; Kruse et al. 2007) is used to illustrate the relationship between water-table fluctuations and the changes in the water storage. Monthly water-table variations are estimated from these data using an expression similar to that of Eq. (18). Figure 3 shows the $\Delta S^{i,A}$ values and the water-table variations in four wells in Tres Algarrobos, Martínez de Hoz, Juan José Paso and Salliqueló. The location of the wells in the central region of the basin is indicated in Fig. 1. During the spring and summer, water tables deepen and produce a great depletion in the water storage ($\Delta S^{i,A}<0$). This is clearly visible during the summer of 2003, when the largest declines in water

Fig. 3 Monthly values of water-storage variations ($\Delta S^{i,A}$) and water-table variations in four wells in the central region of the extended Salado Basin



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tables coincide with the minimum in $\Delta S^{i,A}$. On the contrary, during the autumn and winter, water-table rise is simultaneous with the increase in the water-storage mass. Note that this comparison is qualitative because the spatial representativeness of each time series is different (regional scale for GRACE-derived values, local scale for well data). However, the pattern of water-storage variation obtained from the satellite measurements is consistent with the groundwater variations for the 2-year period considered.

Regional ET is estimated by using the monthly values of precipitation and water-storage variation. The time series of $ET^{i,A}$ values and associated uncertainties are shown in Fig. 4, while the annual accumulated values are presented in Table 1. The figure shows a seasonal variation in the ET with maximum rates during the summer (wet season), ranging between 110 and 150 mm/month (≈4–5 mm/day) for almost all the study period. The highest ET value for a single month, approximately 180 mm (6 mm/day), is observed in January 2006. During the winter (dry season) the ET is minimum and rates are generally lower than 30 mm/month (1 mm/day). The RMS of the uncertainties in the ET values computed using Eq. (20) is 22.6 mm/month (0.75 mm/day), which is in the same order of the ET error reported by Rodell et al. (2004a), Ramillien et al. (2006) and Bonorina and Ramillien (2008). The comparison between the time series of $ET^{i,A}$ and $P^{i,A}$ shows that ET represents a 95.9% of the precipitation. This result is similar to that obtained by Varni and Usunoff (1999) in the central region of Buenos Aires province using numerical simulation tools.

Based on the analysis of Figs. 2 and 4, some specific features of the temporal pattern obtained for the ET can be related to the different climatic scenarios observed in the study region. During 2002, precipitation is above normal due to a warm phase of El Niño (Waple and Lawrimore 2003). The heavy rainfalls lead to a significant rise in water tables and the waterlogging of large areas (Kruse et al. 2007; Viglizzo et al. 2009). Despite the increase of water mass in the basin, ET does not exceed the rainfall during most of 2002. On the contrary, in 2003 rainfalls are low during several months, and ET is higher than precipitation as a result of the increased water availability on the surface (Viglizzo et al. 2009). In 2004, temperatures and rainfalls show normal values for the region (Levinson 2005). However, due to the occurrence of significant rainfall events during the winter, ET (≈60 mm/month) reaches the highest rates for a dry season of the study period. The year 2005 is characterized by a strong negative anomaly in the annual precipitation (Shein 2006). As a result of the rainfall deficit, ET shows low rates (< 45 mm/month) during 6 months. Between the end of 2005 and the beginning of 2006, the storage presents a loss of approximately 100 mm of equivalent water in 5 months, while ET increases. In particular, the maximum value of ET in January 2006, is related to the occurrence of intense rainfalls in this month (≈160 mm). During the autumn and winter of 2006, mean temperatures are slightly below normal (Arguez 2007). Due to the low temperatures, ET presents very small rates (< 30 mm/month) from May to September. There are heavy rainfalls (≈785 mm) between October 2006 and March 2007. Although a large ET variability is observed, the values are high during this rainy period. In the autumn of 2007, the ET decreases and remains low until September due to the extremely cold temperatures registered in Argentina (Levinson and Lawrimore 2008). By the end of 2007, La Niña conditions affect the climate of the region and a prolonged period of intense water deficit begins (Peterson and Baringer 2009). During most of 2008, average temperatures are above normal and rainfall rates show strong negative anomalies. Due to the severe drought event in the region (Sacchi et al. 2008, Arndt et al. 2010), ET shows very low values during 2008 and 2009. In particular, in January 2009 ET (≈55 mm) reaches the lowest value for a wet season of the period 2002-2009.

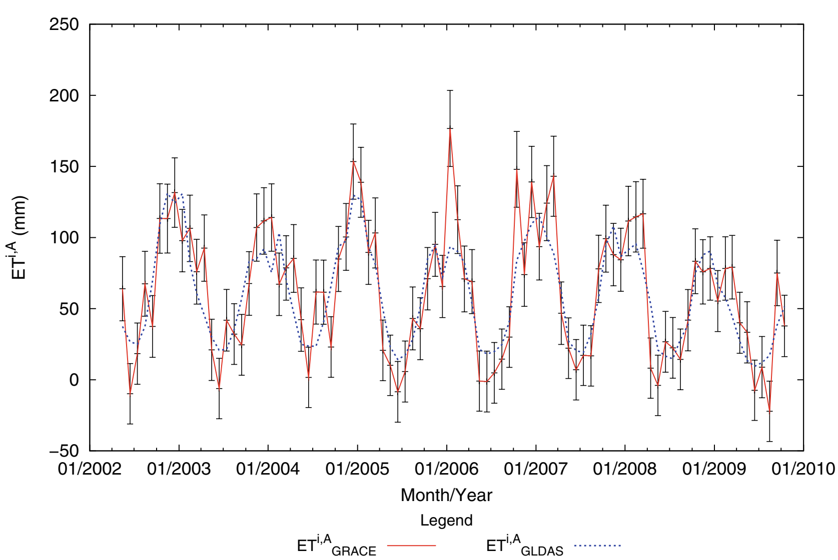

Fig. 4 Monthly values of evapotranspiration ($ET^{i,A}$) computed from GRACE and GLDAS data. *Vertical error bars* indicate the uncertainty in the GRACE-derived values

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Considering that GRACE provides measurements that are temporally and spatially homogeneous, it can be concluded that the applied method is very useful for monitoring the temporal evolution of ET and other processes on a large scale. Note that regional ET and water-storage variations are analyzed at a monthly time-scale, which allows the broadening of the understanding of the seasonal large-scale recharge pattern in the region. This type of information is important to improve the knowledge of the hydrological behaviour of the basin under different climatic conditions (Carol et al. 2010). Since the extended Salado Basin is sparsely monitored, GRACE is the only comprehensive large-scale data source on water storage (Güntner et al. 2007). Further details can be obtained if additional in situ or remotely sensed information on soil moisture and surface water is available. Nevertheless, the results obtained suggest that GRACE is a useful tool to qualitatively analyze aquifer dynamics in the region. On the other hand, the regional features of both ET and water-storage change can be used as a complement to in situ measurements in the analysis of different phenomena on a lower scale. For example, given that cultivations have a strong influence on groundwater in the basin, these estimates can be very helpful in the characterization of local variations of water tables in response to land-use changes (Viglizzo et al. 2009). Therefore, the potential of GRACE data in hydrogeological studies in the Argentine Pampas is invaluable.

Validation

In this section, the obtained results are validated using data from the global model GLDAS. This model is a project of NASA and the National Oceanic and Atmospheric Administration (NOAA) developed to support improved forecast initialization and hydrometeorological investigations (Rodell et al. 2004b). Using data assimilation techniques, GLDAS integrates surface measurements, satellite observations, and accurate land-surface models. The outputs of this model consist of fields of land-surface states and fluxes applied to water balance and energy balance. In particular, monthly 1.0º x 1.0º grids driven with the Noah land-surface model (Ek et al. 2003) for the period 2002–2009 are used.

Because of a lack of in situ observational networks for river discharge, a validation of the runoff model proposed in Eq. (12) using measured data is not possible. However, to analyze the performance of this model, the GLDAS grids of surface as well as subsurface runoff are considered. $R^{i,A}$ estimates are computed using these data, and the annual accumulated values obtained are presented in Table 1. The analysis of this table shows that the model-derived runoff does not exceed 16% of the annual rainfall, although this percentage is lower than 7% during most of the study period. From the ratio between $R^{i,A}$ and $P^{i,A}$ values, an average value of 0.06 is obtained for k_R. Therefore, runoff estimates similar to those provided by GLDAS are obtained with the proposed value for k_R (0.05).

Finally, $ET^{i,A}$ values are computed using the ET grids provided by the global model. The monthly values obtained are shown in Fig. 4, while the annual accumulated values are presented in Table 1. The ET rates computed with both GRACE and GLDAS are in good agreement. Although GLDAS estimates have smaller peak-to-peak amplitude than GRACE-derived values, the temporal pattern of both time series is similar. The differences between the two curves can be partly attributed to the fact that the Noah model does not consider variations in surface and groundwater storages (Syed et al. 2008). The RMS error between GRACE and GLDAS estimates is 25.0 mm/month (0.83 mm/day). This value is in the order of the RMS reported by Rodell et al. (2004a), Ramillien et al. (2006), and Bonorina and Ramillien (2008) when they compared the ET derived from GRACE and GLDAS in different basins over the globe. Note that the results obtained with the applied technique are based mainly on observed data. Although the use of measured runoff data would improve ET estimation, these results show that this method is a valid tool to detect and analyze changes in the ET pattern of the extended Salado Basin. However, accuracy in GRACE-derived results is expected to improve based on the development of new methods for processing satellite data.

Conclusions

In this study, regional ET rates were estimated in the extended Salado Basin from gravity measurements of the space mission GRACE. Monthly ET was determined for the period May 2002–October 2009 with the water-balance equation. The monthly water-storage variations were quantified using GRACE data, and the values obtained showed a temporal pattern consistent with the water-table variations at four wells for a 2-year period. The ET values were computed using the GRACE-derived solutions and rainfall data from the GPCC, and runoff estimates were obtained as a small percentage of the precipitation. GRACE-derived results showed that ET represented a 95.9% of the precipitation. The temporal variation of ET was in good agreement with the different climatic scenarios observed. The validity of the model considered for runoff was successfully tested by using GLDAS data. Likewise, the ET values obtained were observed to satisfactorily reproduce the predictions of the global model. The method used provides very useful information for studying the hydrologic behavior of the study region under flooding and drought conditions. The evaluation of the impact of climate variability on ET is made possible by the spatial and temporal continuity of GRACE data. On the other hand, the regional features of ET and water-storage change can be used as complementary information for studying lower-scale phenomena. The results obtained suggest that GRACE is a valuable

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contribution to improve the knowledge of the different hydrologic processes in a flatland area.

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