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7 authors, including:

Gennaro Alberto Stefania
Università degli Studi di Milano-Bicocca

Marco Rotiroti
Università degli Studi di Milano-Bicocca

Letizia Fumagalli
Università degli Studi di Milano-Bicocca

Pietro Capodaglio
Environmental Protection Agency of Aosta Valley

Modeling groundwater/surface-water interactions in an Alpine valley (the Aosta Plain, NW Italy): the effect of groundwater abstraction on surface-water resources

Gennaro A. Stefania[1]*, Marco Rotiroti[1], Letizia Fumagalli[1], Fulvio Simonetto[2], Pietro Capodaglio[2], Chiara Zanotti[1], Tullia Bonomi[1]

1. Department of Earth and Environmental Sciences, University of Milano-Bicocca, Piazza della Scienza 1, 20126 Milan, Italy.

2. Regional Environmental Protection Agency - Aosta Valley Region, Loc. Grande Charrière 44, 11020 St. Christophe (AO), Italy.

*corresponding author g.stefania1@campus.unimib.it

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Abstract

A groundwater flow model of the Alpine valley aquifer in the Aosta Plain (NW Italy) showed that well pumping can induce river streamflow depletions as a function of well location. Analysis of the water budget showed that ~80 % of the water pumped during two years by a selected well in the downstream area comes from the baseflow of the main river discharge. Alluvial aquifers hosted in Alpine valleys fall within a particular hydrogeological context where groundwater/surface-water relationships change from upstream to downstream as well as seasonally. A transient groundwater model using MODFLOW2005 and the Streamflow-Routing (SFR2) Package is here presented, aimed at investigating water exchanges between the main regional river (Dora Baltea River, a left-hand tributary of the Po River), its tributaries and the underlying shallow aquifer, which is affected by seasonal oscillations. The three-dimensional distribution of the hydraulic conductivity of the aquifer was obtained by means of a specific coding system within the database TANGRAM. Both head and flux targets were used to perform the model calibration using PEST. Results showed that the fluctuations of the water table play an important role in groundwater/surface-water interconnections. In upstream areas, groundwater is recharged by water leaking through the riverbed and the well abstraction component of the water budget changes as a function of the hydraulic conditions of the aquifer. In downstream areas, groundwater is drained by the river and most of the water pumped by wells comes from the base flow component of the river discharge.

1. INTRODUCTION

The interaction between groundwater and surface water (e.g., river, stream, lake, wetland, spring, etc.) is an important aspect of the water cycle since it can influence the utilization of water resources by humans. As remarked by Winter (1998), "*surface water commonly is hydraulically connected to ground water, but the interactions are difficult to observe and measure*". Along its path, a stream can be gaining, losing or both. Furthermore, along a reach, interactions between surface and groundwater bodies could change during the seasons (Alley et al. 1999). When groundwater and surface water systems are interconnected, the groundwater discharge may be an important component of the total flow in the surface water network and vice-versa. Another key aspect is the role of well withdrawal, influencing both the groundwater and surface water balance. In such interconnected systems, well pumping can have a strong influence on the amount of groundwater discharged to surface water bodies (Barlow and Leake, 2012; Sophocleous 2002). Well pumping induces some deformations on the water table that could affect the streamflow, especially in the dry season when the streamflow is mainly constituted by base flow. It follows that the identification of the impacts of well withdrawals on groundwater, surface water and their interactions should be of great importance (Alley 2007; Zhou 2009; Feinstein 2012). Furthermore, Sanz et al (2011) highlights the importance of taking into account the storage capability of the aquifer when an analysis of the impact of the well pumping on groundwater/surface-water interactions is carried out in non-steady-state systems. This aspect gains greater importance if the analysis considers the measured discharges of both wells and rivers.

In order to quantitatively evaluate groundwater/surface-water interactions, three-dimensional numerical modelling is often implemented. MODFLOW is one of the most employed numerical codes in hydrogeology studies to simulate the flow field of groundwater (Harbaugh 2005). This code is organized in various packages, and it is able to simulate the different elements constituting the hydrogeological system (e.g., river, drain, well, no-flow zone, etc.). In particular, the Streamflow Routing (SFR2) Package (Niswonger and Prudic 2005) was especially designed to simulate surface water and its exchanges with groundwater. The coupling of MODFLOW-2005 and SFR2 allows taking into account the possibility that groundwater and surface water might be directly connected through the streambed or separated by an unsaturated zone (Feinstein 2012). The SFR2 package has the following advantages with respect to the simpler and more commonly used River (RIV) Package (McDonald and Harbaugh, 1988): a) stream stage is computed, so that the case in which diversions and pumping make a stream go dry can be simulated, as opposed to RIV which keeps the river stage fixed, b) gaining and losing reaches respond to water-table oscillations and c) it is possible to add surface-water diversions and impose the river streamflow in the system to provide more information on conjunctive use of surface water and groundwater.

This work deals with the analysis of groundwater/surface-water interactions in an Alpine valley in NW Italy, more precisely in the Aosta Plain, located in the middle of the Aosta Valley Region. In this area, groundwater is the main source for public drinking water supply and industry. The investigated area contains an alluvial aquifer, commonly found in Alpine valleys, that has important relationship with the hydrographic network. The studied area is crossed by the Dora Baltea River (main river of the Region, a left-hand tributary of the Po River) which flows from west to east and has 16 tributaries mostly oriented perpendicular to its path. Previous studies found that the interaction between groundwater and surface water changes along the Aosta Plain. In particular, the Dora Baltea River changes from losing to gaining from upstream to downstream within the study area. (P.I.A.H.V.A. 1996; Triganon et al. 2003; Bonomi et al. 2013, 2015).

The main aim of this work is to evaluate the impact of well pumping on groundwater and surface water resources in the study area, where the main river is changing from losing (upstream) to gaining (downstream). Other related purposes are: a) to reproduce the wide seasonal fluctuations of the water table and to better understand the dynamics of groundwater/surface-water interactions b) to improve the knowledge of the Aosta Plain aquifer.

In order to achieve these goals, a numerical groundwater model, both in steady-state and transient conditions, was implemented using MODFLOW 2005 (Harbaugh 2005) and the SFR2 Package (Niswonger and Prudic, 2010) for simulating the Dora Baltea River and its tributaries. More specifically, the investigative approach included: a) the definition of the conceptual model of the Aosta Plain aquifer; b) the reconstruction of the 3D distribution of hydraulic conductivity by interpolation of coded data (Bonomi et al. 2009, 2013, 2014, 2015; Perego et al. 2014; Rotiroti et al. 2015; Stefania et al. 2015) extracted from the TANGRAM© database (Bonomi et al. 2014); c) the quantification of groundwater/surface-water interactions and the effect of well withdrawal on water resources by the interpretation of MODFLOW mass balance.

An innovative aspect of the present work is the application of the advanced package SFR2, that was previously used in a few studies (Feinstein et al. 2010; Masterson and Granato 2013; Leaf et al. 2015). Moreover, to the best of the authors' knowledge, no previous studies have used the SFR2 package to evaluate the effects of well pumping on groundwater/surface water interactions in complex hydrological systems such as the Alpine valleys.

2. MATERIALS and METHODS

2.1 Study area

The study area is a portion of the Aosta Plain, located in the Alpine region of Aosta Valley in northern Italy (Fig. 1). It stretches west-east for about 13.5 km, reaching a width of 2.5 km in the central part, near the town of Aosta, and its elevation ranges from 700 to 530 m above sea level (a.s.l). It lies in the mountain catchment area of the Dora Baltea River flowing from west to east (Fig. 1), with 16 tributary streams, the largest of which is the Buthier Creek. The study area is approximately 60 km^2, i.e. less than 2% of the whole Aosta Valley region, but includes 66,000 inhabitants (52% of the entire regional population) with a population density up to about 1,600 per km^2 in the town of Aosta. The water supply in the area is mostly from groundwater.

Temperature and precipitation, measured at various monitoring stations in the Aosta Valley, were analyzed to obtain the average monthly values related to different elevation zones of the region: north-west (2,336 m a.s.l.), north-east (1,979 m a.s.l.), South (1,951 m a.s.l.) and the Aosta Plain (581 m a.s.l.) (Fig. 2). Although the study area is limited to the Aosta Plain, the precipitation in the surrounding catchment areas is also of interest because it falls mostly on impermeable upland surface and then circulates to the plain as infiltration through alluvial debris and fans at the edge of the valley or as spring flow. The measurements reflect climatic conditions at different elevations; the maximum temperatures are measured in the Aosta Plain (annual average: 12.8°C), the minimum ones in the west area (annual average: 1.7°C). The minimum monthly temperature is in January or February, the maximum in August.

The maximum precipitation is generally recorded in April, although in the north-west watershed the wet period extends from April to June. The minimum precipitation is usually in October in the South watershed, in January in the north-east and north-west watersheds and in August in the Aosta Plain. Except for the plain area, the precipitation is in the form of snow from November until March, and the snow melt occurs from March until June.

2.2 Conceptual model

2.2.1 Hydrogeological system

The Aosta Plain is underlain by a series of fluvioglacial, lacustrine, alluvial and fan sediments, Quaternary in age, which in turn lay on a deep crystalline basement eroded by the Balteo Glacier. The aquifer is bounded to the north and south by the crystalline bedrock of the Alps (PIAHVA 1992). In this sedimentary basin, a silty-sandy deposit, never completely penetrated by wells, is located at the depth of about 50-90 m (decreasing from west to east) from the land surface and its thickness is estimated to be over 40 m (Pollicini 1994; Triganon et al. 2003; Bonomi et al. 2013, 2015; Rotiroti et al. 2015). This deposit of lacustrine origin is considered to act as a low-permeability basement boundary for the overlying aquifer used for water supply. Furthermore, two recent Electric Resistivity Tomography surveys (ERT) conducted in the towns of Aosta and Pollein during 2013 (unpublished data provided by the Regional Environmental Protection Agency of Aosta Valley (ARPA VdA)) revealed the presence of a deeper sandy-gravel aquifer (of likely glacial origin), not yet exploited, below the lacustrine aquitard. Moreover, the ERT showed that the lacustrine aquitard has some discontinuities (coarse deposits that act as higher permeability windows) that allow the deeper glacial aquifer and the overlaying exploited aquifer to be interconnected.

The exploited shallow aquifer, consisting mainly of heterogeneous alluvial deposits, ranges in thickness from 85 to 90 m in the western part to 50 m in the eastern part where the aquifer is divided into an unconfined (about 20 m thick) and a semi-confined portion (between 25 and 12 m) by a silty layer (Pollicini, 1994; Nicoud et al. 1999; Triganon et al. 2003).

2.2.2 Available hydrological and hydraulic head data

The available discharge data of the stream network between 2008-2014 were provided from the institutional monitoring network of ARPA VdA. The streamflow data were recorded by four hydrometric stations: Aymavilles (Ay - 618 m a.s.l.), Pollein (Po - 545 m a.s.l.), Nus (Nu - 534 m a.s.l.) and Roisan (Bu - 742 m a.s.l.). The first three are related to the Dora Baltea River whereas the Roisan station is related to the Buthier Creek (Fig. 1). Generally, the Dora Baltea River is characterized by two yearly streamflow peaks respectively in June and November (Fig. 3). The first one is greater than the second and represents the snowmelt and high precipitation period of the year, the second arises only from the autumnal rainfall peak. With regard to 2009, in January the average streamflow at Aymavilles (3.77 m^3/s) was higher than the one at Pollein (2.51 m^3/s) which was lower than at Nus (3.06 m^3/s). In June, the peak recorded at Pollein (129.73 m^3/s) hydrometric station was higher than both those at Aymavilles (109.15 m^3/s) and at Nus (100.80 m^3/s). As for the Butheir Creek, it had the same yearly streamflow features of the Dora Baltea, though its discharge has always been much lower than the previous one. For the surface water, the low flow period is usually in winter when the temperatures are low and the precipitation is in the form of snow.

The hydraulic-heads trend analysis was carried out in order to better understand the behavior of the aquifer. Hydraulic head data from the institutional monitoring network of ARPA VdA (Fig. 4) were available from 2000 to 2012. Special attention was given to heads data measured within the modelled area between 2009-2010 (selected period for the groundwater flow model). The hydraulic head (Fig. 4) decreases from about 582 m a.s.l. (Sa03 – upstream area) to 522 m a.s.l. (Nu05 – downstream area), suggesting a principal flow direction from west to east. Hydraulic head data show a seasonal oscillation between the yearly maximum value in summer and the minimum value in March-April. In particular, for the upstream area (Sa03) the oscillation is about 6-7 m and decreases by up to 1 m while moving to the downstream area (Qu16). This sharp reduction is related to a likely draining behavior of the Dora Baltea River on the groundwater which seems to decrease starting from Pollein hydrometric station toward the downstream area (Bonomi et al. 2015).

2.2.3 Groundwater flow

The water table, built using Ordinary Kriging interpolation of monthly hydraulic heads from piezometers (monitored from ARPA VdA) and from the Po hydrometric station, is represented by the piezometric map in Fig. 5 (January 2009).

As a consequence of the valley shape, the aquifer structure, and the relationships with surface water, the piezometric heads vary from 600 to 526 m a.s.l. from west to east along the valley. The hydraulic gradient decreases downstream: it is 1% in the western, 0.5-0.6% in the middle and 0.3-0.4% in the eastern sector of the area.

As for the relationship between surface water and groundwater, the Dora Baltea River is losing upstream of the Po hydrometric station (PIAHVA 1996; Triganon et al. 2003), but downstream of the Po hydrometric station it becomes gaining with hydraulic connection (PIAHVA 1996; Triganon et al. 2003; Bonomi et al. 2013, 2015).

The water depth decreases longitudinally from west to east and transversely from the valley slopes to the Dora Baltea River, ranging from a maximum of around 20 m and 25 m respectively in the western sector and in the alluvial fan of the Buthier, to a minimum of 10 m and 4 m respectively in eastern sector, near the Dora Baltea River. This distribution restates the relationship between groundwater and the Dora Baltea River which is losing in the western sector while it is gaining in the eastern sector, in particular from the Po hydrometric station.

With respect to the temporal groundwater oscillations, the difference between the hydraulic head of July and January 2009 shows that in the western area the combined effect of the narrow valley, snowmelt and probable recharge from the

Dora Baltea River causes water-level oscillations up to 6-8 meters, while towards the widening of the valley, lower oscillations occur. Finally, in the eastern part, the river drainage reduces groundwater head fluctuations to several meters.

2.2.4 Well pumping

The Aosta Plain aquifer is subjected to intense groundwater exploitation serving both drinking and industrial supplies. Local industries are served by 19 wells located in the central and in the eastern part of the Aosta Plain whereas the drinking water supply comprises 8 wells located in the western part of the plain (Fig. 6a). Therefore, the total number of operating wells in the area is 27. Other wells for domestic supply exist in the study area, however, their smaller discharges are negligible for the purposes of the present work.

Data on groundwater abstraction from these 27 wells were provided by ARPA VdA. These data concern the monthly average well discharge (m^3/hour) and the well working time (hours) for the period 2009-2010. The total volume of groundwater pumped was 19,656,447 m^3 in 2009 (82.87% for industrial and 17.13% for drinking uses) and 2,2024,240 m^3 in 2010 (81.35% for industrial and 18.65% for drinking uses). From January to August 2009 the total well discharge decreases from 63,000 to 44,000 m^3/d. Between September and December 2009 a peak was recorded in October (i.e. 69,000 m^3/d) due to industrial needs. Concerning the 2010, the well discharge gradually rose from January to July, then it fell in August representing the annual minimum (49,000 m^3/d). From September to December 2010, the discharge was quite stable around 60,000 m^3/d.

2.3 Numerical model

2.3.1 Previous models

Although the Aosta Plain aquifer was already modelled by three previous works (Triganon et al. 2003; Bonomi et al. 2013, 2015), some key aspects concerning groundwater/surface-water interactions remained unclear and required the development of a new improved model. A brief description of these earlier models (and their limitations) is given in the following.

The first model presented by Triganon et al. (2003) took into account the shallower part (up to 90 m) of the Quaternary sediments subdividing it into two layers. Only the Dora Baltea River was simulated using the RIV package, hence without considering its discharge. The hydraulic conductivity was discretized in four isotropic zones of piecewise constancy with the same distribution between layers. This steady state model provided a preliminary analysis on the effects of the surface water level on both the water table elevation and the water exchanged between the main river and groundwater. The coarse refinement of the grid and the hydraulic conductivity coupled to the use of the RIV package led to rough estimates for results, in particular, concerning the amount of exchanged water between the river and aquifer.

Bonomi et al. (2013) presented a steady state model, which considered the whole hydrographic network of the Aosta Plain simulated by the SFR2 Package and MODFLOW2000. That work focused on the definition of a preliminary conceptual model of the aquifer in order to improve the general knowledge of the system. However, the use of a steady state condition prevented the understanding of the temporal variations on groundwater/surface-water interactions.

Bonomi et al. 2015 improved their previous work implementing a transient model based on monthly recharge values with the aim of simulating the seasonal oscillation of the water table, however no evaluations of the water budget and exchanged fluxes were done.

In order to overcome the limitations of the above-mentioned previous models, the present work implements the following new aspects: (1) a new reconstruction of the system geometry that incorporates the deeper glacial aquifer which is likely interconnected with the shallower units as evidenced by new available data (i.e. ERT surveys and new well logs, see section 2.2.1); (2) an improved three-dimensional reconstruction of the hydraulic conductivity distribution (see the Electronic Supplementary Material - ESM for details) in order to include the new available well logs; (3) an automatic calibration, using PEST, for estimation of model parameters.

Nevertheless, the detailed calculation of monthly recharge inputs made by Bonomi et al. (2015) is retained in the present work. These recharge values were calculated as the water average surplus obtained by the Thornthwaite-Mather's hydrological balance model starting from the average monthly precipitation and temperature data (available only for 2009 and 2010 and calculated for the different regional watersheds; Fig. 2).

2.3.2. Hydraulic conductivity reconstruction

In order to reconstruct the spatial distribution of hydraulic conductivity, the TANGRAM© database was used (Tangram, 2016; Bonomi et al. 2014). This database is meant to manage administrative, borehole construction, stratigraphic and piezometric information. TANGRAM© has an internal coding system that allows the standardization of stratigraphic information and the extraction of stratigraphic data in terms of hydraulic conductivity (k), according to depth, measured at the meter scale in the present work. The hydraulic conductivity values of 176 stratigraphic logs were extracted from TANGRAM© in order to obtain a XYZk point dataset, where X and Y are coordinates in the UTM system, Z is the elevation (m a.s.l.) and kis the hydraulic conductivity value (in terms of ln(k)). The values of hydraulic parameters are defined in relation to textural percentages (Bonomi et al. 2002; Bonomi et al. 2009), by means of a conductivity data conversion table integrated into the TANGRAM© database. The distribution of hydraulic conductivity was assumed to be lognormal (Martin and Frind 1998) and a well-known method (Dagan and Lessoff 2007; Dagan 2012) for calculating the weighted standard mean (Sanchez-Vila et al. 1995) of hydraulic conductivity values attributed to the individual lithologies was followed (ranging from 436 m/d for gravel to 0.864 m/d for silty-clay; for more details on the interpolation see the ESM).

For this study, the hydraulic conductivity values assigned to the conversion table in TANGRAM© were previously calibrated by trial and error on the basis of nine pumping tests reported in literature (Pollicini 1994; De Luca et. al 2004). The calibration was focused on the minimization of the difference between the ln(k) value calculated by TANGRAM© and the ln(k) value from the literature.

The coded lithologs were imported into a 3D grid and interpolated by geostatistical approach in GOCAD®. The GOCAD® code (Geological Object Computer Aided Design, Paradigm 2008) was used to build the computational grid, then apply the geostatistic technique of Ordinary Kriging, based on the assumption of stationary spatial conditions with respect to the statistical distribution of the subsurface stratigraphic data. For more details on the interpolation see the Interpolation Grid section of the ESM.

2.3.3 Model design

This work is the evolution of specific existing models (Bonomi et al. 2013, 2015). Indeed, data about hydrogeological balance (e.g. recharge, pumping rate, river parameters) derive from these previous works. They have been integrated with new information allowing researchers to improve the knowledge of the investigated aquifer.

The model grid has 3,183,300 cells divided in 243 rows and 655 columns with a uniform cell spacing of 20x20 m, covering an area of ~60 km^2 (active cell ~25 km^2). The vertical discretization of the grid reflects the stratigraphic grid used for the interpolation of the hydrogeological parameters. It consists in twenty layers: eighteen of them have a constant thickness, whereas two (second and third layers) are variable (Min 0.8 m; 25th percentile 5.6 m; median 10.4 m; 75th percentile 20.2 m; Max 72.8 m). The first sixteen layers were used to model the exploited aquifer, whereas the last four layers (90 m thick) were used to consider the silty layer (lacustrine aquitard of the main shallow aquifer; see section 2.2.1) and the underlying coarse (gravelly-sand) deposits recorded by ERT. The refinement of the vertical spacing helps to capture hydrogeologic trends and allows the model to closely match the intervals over which pumping wells are screened, thereby increasing the accuracy of the model in simulating withdrawals. Furthermore, the low thickness of the upper layers has allowed an accurate positioning of the riverbed, improving the simulation of the rivers and groundwater interaction.

The complex three-dimensional reconstruction of hydraulic conductivity (Fig. S1 of the ESM) was subsequently discretized in ten zones of piecewise constancy by means of the analysis of the frequency distribution of the obtained hydraulic conductivity. As for the last four layers, two additional zones were defined: in the central part was assigned a low value (0.0864 m/d) in order to reproduce the silty-clay lacustrine layer, whereas at the edge a high value (69.12 m/d), typical of the sandy deposit, was assigned. The groundwater flow equation of a transient model is complicated by another term representing the aquifer storage capability (i.e. specific storage (Ss) and a specific yield (Sy)). As for the storage, the discretization in zones was obtained from the hydraulic conductivity distribution, whereas the values were selected from literature. In particular, the assigned values on each zone varied between 0.05 and 0.28 for Sy, and between $5x10^{-6}$ and $1.5x10^{-4}$ for Ss (Domenico and Mifflin 1965). The obtained conductivity/porosity zones and their values became the basis for the flow model.

The used code was MODFLOW 2005 (Harbaugh 2005), visualized through the GroundwaterVistas6 interface (Rumbaugh and Rumbaugh 2004).

The model was solved both in steady-state and transient condition with GMG package and considering the rewetting capability. The steady-state configuration was referred to January 2009 in order to reproduce an average condition of the aquifer between 2000 and 2012 (Fig. 4). The transient analysis was limited to a two-year period (2009-2010) using 24 stress periods (one per month). This approach allowed both evaluation of the non-steady-state model response to variable stresses (Taviani and Henriksen 2015) and evaluation of how the change of the water balance during seasons takes into account the storage component. The choice of the selected simulated period was mainly related to the availability of measured data on well abstractions (see section 2.24.). This is a key aspect since the main goal of this work is to understand the effects of well pumping on the relationship between groundwater and surface water. The use of measured discharges rather than a speculative estimation of their values is preferred since it decreases the uncertainty and increases the accuracy of model results.

The streamflow data of the Dora Baltea River and Buthier Creek in addition to the piezometric and precipitation data referred to 2009-2010 were used both in the steady state model (January 2009) and in the transient one (2009-2010).

In order to evaluate the impact of wells on both rivers and groundwater, simulations with different configurations of the pumping were conducted (Feinstein et al. 2010; Barlow and Leake 2012). The transient solution with 27 active wells (baseline scenario) was compared with two new scenarios. Each new scenario differs from the baseline scenario in the discharge value of only one well: scenario A, which aims to evaluate the effects of well pumping in the upstream area where the Dora Baltea River is losing, considers an increase of 1,000 m^3/d for the discharge of the well W_US (see location in Fig. 9); scenario B, concerning the effects of pumping in the downstream area where the Dora Baltea River is draining, considers an increase of 1,000 m^3/d for the discharge of the well W_DS (see location in Fig. 9). Results of the different scenarios where compared through the analysis of the MODFLOW mass balance output file (i.e. *.lst file).

2.3.4 Boundary hydraulic conditions

The boundary conditions were defined on the basis of the hydrogeological configuration of the Aosta Plain and the piezometric trend of the period 2009-2010. The conditions imposed are: 1) No Flow, Neumann-type boundary conditions, at the northern and southern edges of the plain, at the boundary between fan deposits and the slope. In depth, the limit to no-flow defines the geological shape of the valley, cut in the western part and progressively widening towards the east (Bonomi et al. 2014). 2) General Head Boundary (Cauchy-type boundary conditions), at the western and eastern edges of the model, with the aim to simulate groundwater flow into and out of the valley. GHBs have been reconstructed, one for each time step. Their position derived from piezometric maps related to January and July 2009.The assigned head values to each time step were taken from well piezometric data measured from 2009 to 2010. 3) Stream (Cauchy-type boundary conditions), to match the hydrographic network and to simulate the interactions between the surface and groundwater systems. Considering the importance of groundwater/surface-water exchange to the study overall, more details about this boundary condition are discussed in the next paragraph.

2.3.5 Hydrographic system modelling

The hydrographic network was simulated using SFR2. This package routes stream flow along the hydrographic network and calculates the exchange between the surface and groundwater systems. Moreover, it allows researchers to simulate the induced water flow from the river to the aquifer by means of the wells pumping up to the maximum discharge available into river, after which the cells become dry. The SFR2 selected option allows stream stage to be calculated based on specified overland flow rates at upstream locations, the calculated groundwater exchange for each model cell occupied by a stream, Manning's equation relating channel roughness to streamflow, and a channel geometry assumed to be rectangular.

An outline of the hydrographic network in a series of reaches and segments, ranked in a strict sequence according to tributary order, allows the modelling of complex hydrographic networks. In particular, a reach corresponds to a single cell of the model grid, while a segment is made of one or more reaches with characteristics similar to one another (Prudic 2004). A sequence of 32 segments (1,908 reaches) represents the hydrographic network of the study area, each segment corresponding to a tributary of the Dora Baltea or to a section of the Dora Baltea between two confluence points (Fig. 6).

For each reach the following information was specified: stage of stream, streambed elevation, width of stream, length of stream, thickness of streambed, streambed hydraulic conductivity, streambed slope, streambed roughness, flow entering segment (only for the boundary segments).

Streambed roughness was imposed at 0.05 (-), intended to represent river beds with pebbles and some rocks (Barnes 1967). Streambed slope was calculated between adjacent cells, as a function of the inclination of the streambed. The values obtained for the Dora Baltea River are between 0.1% and 1.5%, but for the tributaries the slope can be as high as 40%.

An inflow (Q) to the segments at the model upstream boundaries was applied to the first cell of each segment and for each stress period, to allow the simulation of flow rate propagation in the river (Prudic 2004). The estimated average monthly flow values referred to 2009 and 2010 were assigned to the first reach of SFR2 related to the Dora Baltea River (Aymavilles hydrometric station, estimated net of diversion) and the Buthier Creek, whereas for the remaining streams it was estimated on the basis of information gathered and occasional ARPA VdA measurements.

2.3.6 Model recharge and discharge

As remarked by Bonomi et al. (2015), the recharge of the studied aquifer is closely related to the snowmelt (occurring from April to July) which, in turn, is related to both the quantity of snow occurring in winter and the increase of the temperature during following seasons. The recharge was imposed through 4 zones and using monthly time steps (Bonomi et al 2015). The minimum imposed value was 0.099 mm per month in the central part of the plain in January 2010, whereas the maximum value was 2292.14 mm per month per day in the southern-west of the plain in May 2009.

Another important stress and element of the hydrologic budget is well withdrawal. Measured monthly well withdrawals were assigned to the 27 operating wells (see section 2.2.4) depending on their own flow rate, ranging between 0.26 and 18,000 m^3/d. The screen lengths and depths were inserted for each well, and they are distributed between 20 m and 80 m in depth.

2.3.7 Calibration and Target

Both hydraulic head target and flux target were used to calibrate the model. The head targets were compiled from the piezometric monitoring network of ARPA VdA (Fig. 1) on a monthly basis or quarterly data. It was considered inappropriate to define an allowable error in the head target comparison values (Sonnenborg et al. 2003) both because the data are the actual monthly measurements (not derived from the mean of data referred to longer periods) and because it is considered reliable (ARPA VdA monitoring). Local errors might still be related to piezometric measurements influenced by the pumping of the surrounding wells. The number of hydraulic head observations was 25 for the steady-state condition and 348 for transient simulation heterogeneously distributed among 37 head target.

The flux target (ARPA VdA monitoring) is located in the central area of the model (Po hydrometric station - Fig. 1). The flux target reported 24 values defined as the monthly average of the daily streamflow recorded by Po hydrometric station between 2009 and 2010. All targets are shown in Fig. S3 of the ESM.

The model calibration was done on the steady state model using PEST (Doherty 2008, Doherty and Hunt 2010). A preliminary sensitivity analysis on hydraulic parameters (kx=kv, kz=kx/10) and SFR2 conductance was performed on the steady-state model. All of those most sensitive parameters (>1%) were selected for the calibration process (Hill 1998). A reasonable upper and lower bound was applied to the initial value of each selected parameter (±50% of the starting value for *k* and 0.0001 to 5 (m/d) for the stream bed conductivity). The obtained calibrated model was employed as a starting condition for the transient model. To improve the performance of the transient model and to better reproduces the strong oscillation of the water table, a second calibration was carried out by PEST. A new sensitivity analysis on storage values (Ss and Sy) and hydraulic conductivity (kx=ky) was done (Fig. S4 of the ESM). Parameters with sensitivity more than 1% of the greatest value of the composite sensitivity were calibrated applying reasonable upper and lower bounds (Table 1).

3. RESULTS AND DISCUSSION

3.1 Steady state simulation (January 2009)

The choice of January is connected to the flow rate of the surface system, which is representative of the low-flow period (Feinstein et al. 2010); in fact as shown before, this is the month when the water table is lowest and the runoff from stormflow is largely absent. Furthermore, January 2009 represents an average condition of the aquifer during the last decade.

Fig. S5 of the ESM shows the distribution of the calibrated value of the hydraulic conductivity within layer 4. The values of this parameter are in agreement with previous studies on the Aosta Plain (Triganon et al. 2003; Bonomi et al. 2013, 2015; Stefania et al. 2015) ranging between 0.0864 m/d to 572 m/d.

The Dora Baltea River conductance was calibrated starting from a hydraulic conductivity of 1 m/d. As a result of the calibration, it changes among 0.15 m/d in the western part to 0.01 m/d in the central part and it increases to 4 m/d in the eastern part (Hatch et al. 2010) (Fig. 6 a). The simulated head, which takes into account the wells pumping and the exchange between rivers and groundwater, produced the piezometric configuration shown in Fig. S3 of the ESM. It represents, with spacing of 1 m, a variation of head from 580 m a.s.l. in the west to 526 m a.s.l. in the east, and a flow direction west-east. It closely reproduces the variations of the gradient, which increases from the west towards the Aosta Plain, decreases near the areas with high hydraulic conductivity around the fan of the Buthier and increases again towards the downstream sector.

The solution has the following statistics: the residuals ranged from -0.96 m to +0.94 m, with a Residual Mean of -0.02 m, Absolut Residual Mean of 0.40 m, and a Scaled Root Mean Square value of 1%, and the sum of square residuals (the square root of the average of the squared error, Anderson and Woessner 1992) was 5.63.

The simulated flux at the Po hydrometric station (2.76 m^3/s) appears in good agreement with the observed one (2.55 m^3/s), considering the observed Dora Baltea River discharge and all of the factors that cause it, among which: 1) the diversion near Aymavilles hydrometric station, 2) tributaries discharge contribution, 3) exchange between surface water and groundwater.

With regard to the mass balance (which has an error less than 0.001%), approximately 21,521 m^3/d are conveyed into the model due to the recharge contribution, and about 62,849 m^3/d exit due to 27 wells. In the present configuration, the surface water system recharges the aquifer at a rate of 132,534 m^3/d (IN from stream to aquifer), and drains the aquifer at a rate of about 90,130 m^3/d after the Po hydrometric station (OUT to the stream).

In the central area of the Aosta Plain, around the fan of the Buthier, the simulated head shows values higher and lower than the observations (Fig. S3 of the ESM). These discrepancies may be associated to different aspects: a) an in-depth analysis is necessary for a better description of heterogeneities distribution which regulate groundwater flow on a local scale, b) the attribution of monthly average pumping rate values in the model, whereas the real hourly flow rate might have influenced the measured water levels, generating noise in the targets which is not reproducible by the simulation, c) the real head target values represent monthly average levels and they may not be perfectly correlated with monthly pumping. In the eastern part, the simulated head suffers from an overestimation. This result suggests that as to the downstream area, probably, further geological investigations should be considered, to correctly understand how the water leaves the modelled area through the narrowing of the valley edges. In order to evaluate the impact of well pumping on the water table and stream depletion, a simulation without pumping was done, then it was compared with the calibrated solution. The simulation without pumping shows a considerable rise of the water table especially in the central part of the area where the wells' withdrawal is bigger. In this area, beneath of the Aosta town, the water table was pushed down more than 5 m due to well pumping (Fig. 6b). Table 2 shows the changes induced by pumping on water balance. Overall, the pumping limits the amount of the water gained by the Dora Baltea River (~34,400 m^3/d) and increases its leakage (~16,300 m^3/d). This result highlights that surface water is an important part of the hydraulic budget of the Aosta Plain. The groundwater/surface-water exchanges (average flux m^3/d) along the Dora Baltea River are shown graphically to emphasize their spatial variability for both pumping and no pumping scenarios (Fig. 6). The drawdown induced by the wells' withdrawal not only affects the shape of the water table, in fact, it also induces changes on the surface/groundwater exchange flux with a slight increase of the stream leakage by segment 11 and about one order of magnitude from the segment 17. Segment 18 shows the greatest variation with a sharp decrease of the water drainage (e.g. from 490 to 21 m^3/d) as a result of the groundwater depletion. Also segments 20, 22 and 24 decrease their drainage flux, however, less than the previous one. The largest difference of water exchange recorded in segment 18 is related to the downstream moving of the transition point from losing to gaining behaviour of Dora Baltea River. This transition point is located within segment 18 in both scenarios, however, its movement downstream in the pumping scenario, due to water-table lowering, leads to a decrease of the water drained by the river. In the central area (close to segment 13 and 15), the water-table elevation never exceeds the riverbed elevation thus? preventing any variation of water exchange between the river and aquifer. The well pumping has no influence on the tributary streams (Table 2) because a) the great unsaturated thickness (around 30 m on average and up to 96 m) beneath tributary streams prevents any direct interaction between the tributary water and groundwater; b) the greater slope of the riverbed up to

0.45% (Bonomi et al. 2013) increases the water flowing velocity while, on the other hand, decreasing the possibility of its infiltration toward groundwater; c) the available discharges of the tributary streams are about two orders of magnitude lower than the main river. Furthermore, some of the tributary streams (i.e. segments 2-6-8-14-25-27-29) have an impermeable riverbed that prevents any water exchanges.

3.2 Transient simulation (January 2009 - December 2010)

Models with two different sets of hydraulic parameters were compared (Fig. 7). The new set of parameters obtained by the second analysis of sensitivity (Fig. S4 of the ESM) and subsequent recalibration (Tab. 1), as defined in Section '*Calibration and target'*, better reproduces the hydraulic head trend both upstream and downstream of Aosta town (Fig. 7). The spatial and temporal behavior of residuals between observed and simulated values highlights that the first solution with respect to the recalibrated solution, overestimates the hydraulic head, especially during the low flow periods, inducing a water accumulation in the downstream area.

The analysis of sensitivity has shown that the water accumulation was dependent on both storage and conductivity parameters. Another important aspect was that the newest calibrated values remained consistent with those that were initially derived from the TANGRAM database (see section 2.3.2), however a slight increase of the values (except to sy_3) was observed. With regard to the reconstruction of the streamflow of the Dora Baltea River, the modelled discharge (Fig. 7) compared with the measured data was in good agreement, even if it has not correctly reproduced the fall in discharge peaks. However, the modelled streamflow seems to be acceptable taking into account that the discharge of tributaries has been estimated and the hydraulic conductivity is a challenging parameter to be estimated (Hatch et al. 2010). Comparing the simulated water table with the bottom elevation of the Dora Baltea River, it becomes clear how the piezometric oscillation affects the Aosta Plain aquifer during the seasons and how it changes their relationship along the investigated area (Fig. S6 of the ESM). Upstream of Aosta town, the distance between the water table beneath the streambed and the stream bottom decreases steeply, with occasionally sharp reduction dependent on streambed jumps. Close to Aosta town, the water table falls, probably due to the wells' withdrawal with a variable magnitude which changes seasonally. Another important outcome is that some of the Dora Baltea reaches located upstream of Aosta town change from losing to gaining and vice-versa. In fact, upstream of Aosta (between 5.5 and 7.2 km) during the summer of 2009, when the water table was the highest, it was above the streambed, whereas close to Aosta, this relation was reversed by pumping, making the Dora Baltea River losing. This behaviour is reasonable because during June 2009 the well pumping was lower and the recharge was higher inducing a great rise of the water table. Downstream of Aosta town, the Dora Baltea River becomes draining; moreover the reach where the river changes its behavior is moved upstream by about 1 km, compared to June 2010. The water budget of the Aosta Plain aquifer was obtained from the transient simulation (Fig. 8): it is characterized by two peaks of inflow due to recharge during the spring 2009-2010.

The hydrography network is both an inflow and an outflow: the latter increases when the water table is higher (summer), confirming that the water table oscillation is reduced by the river drainage especially in the downstream area. The greater river inflow occurs during the periods of most infiltration, confirming that snowmelt plays a key role in the recharge of the groundwater system. As for the storage, it appears an important element of the groundwater budget because it acts as a tank. In fact, the aquifer is recharged during the water-available periods (SINK element) and becomes a source of water during the low flow period, typically in winter. In October 2009 there is the highest well withdrawal (about 69,000 m^3/d) which induced the most important outlet from the storage. These results have highlighted that the Aosta Plain aquifer storage capability is a key aspect of the groundwater system which together with recharge and surface water seepage/drainage, contributes to the complex dynamics that occur on the studied aquifer, allowing fluctuations of the water table more than 5 m in amplitude.

3.3 Impact of wells on the groundwater and river budget

The first important feature which emerges and confirms the goodness of the obtained groundwater model is that it is able to give back the imposed increase on the well discharge without producing mass balance errors. The obtained results highlight two different behaviors of the aquifer: in downstream area (i.e. Scenario B) the water to the well comes mainly from the river, whereas for the upstream area (i.e. Scenario A) different elements of the water budget actively contribute to the well discharge. In particular, as for the well in the downstream area (W_DS in Fig. 9b), about 80% of the extracted water, for each considered period, would have reached the river (stream diverted) adding itself to the baseflow component of the river flow, but it is instead pumped out by the well. The fact that the impact of W_DS well on the budget is constant over time strengthens the hypothesis that, in the downstream area, the river is hydraulically connected with the aquifer and it acts on the water table, regulating its seasonal oscillations and the exchanged fluxes. As for the well in the upstream area (W_US in well (Fig. 9a), the pumped water comes from different budget items (i.e. aquifer storage, river, upstream groundwater inflow simulated by the GHB). In particular, the contributions of the storage and the river (e.g. stream induced = water induced through the riverbed) change over time as a function of the hydraulic condition of the aquifer and river. In fact, when the water table is the lowest (e.g. February 2009) then the well pumping is not able to abstract water from the river because the flow through the unsaturated zone is likely low, consequently the water is mainly produced from the storage (Sanz et al. 2011). Instead, when the water table is higher (e.g. June 2009), then the well water is mainly from the river (stream induced). For the two other periods referred to W_US, the well impact on the water balance is similar between the lower and higher period because the water table elevation changes less than the previous periods (Fig. S6 of the ESM). Nevertheless, for the higher period (i.e. June 2010), the amount of induced water from the river is more than the amount induced during the lower period (i.e.

January 2010), confirming that the water table oscillation plays a key role in the water exchanges between the groundwater and surface water.

4. CONCLUSION

This work dealt with the modelling of an Alpine valley aquifer in the Aosta Valley Region, assessing the water exchanges occurring between the main regional river (i.e. Dora Baltea River) and the underlying shallow aquifer. After having defined the conceptual model of the system and reconstructed the distribution of the hydraulic parameters, a steady state model was calibrated on the low flow period and an evaluation of the pumping from the whole well system was performed. The steady state model was used to initialize a two-year transient model by which both the water table oscillation during the seasons and the impact of the pumping on the groundwater/surface water budget were evaluated. This work has improved the understanding of the Aosta Plain aquifer, in particular:

a) the reconstruction of the hydraulic parameters shows an aquifer mainly characterized by a gravelly and sandy deposit with a local silty-clay layer especially in the downstream area;

b) the main recharge of the aquifer derives from the slope region, and only secondarily from infiltration on the plain. The obtained recharge rate is typically highest during the summer as a result of high precipitation and snow melt. This explains the wide range of water table oscillation which affects the studied aquifer;

c) the calibration phase, using head and flux targets, has demonstrated the importance of the three-dimensional reconstruction of the system heterogeneity for the model response; moreover the calibration analysis suggested that the modification of the streambed conductivity and also a few other hydraulic parameters, especially Sy and Ss, have turned out to be influential parameters to achieving a good fit on targets;

d) the model highlighted that from west to east of the valley, the Dora Baltea River changes its relation with the aquifer, acting as a source of water in the upstream area and as a sink in the downstream area. The analysis on the whole well pumping system determined that it limits the amount of the water-gaining by the main river (~34,400 m^3/d) and increases its leakage (~16,300 m^3/d). The seasonal changes in the water table elevation play a key role in the relationship between the river and the underlying aquifer. Furthermore, it is evident that the rising of the water table, typically in summer, is a consequence of the recharge (in the form of infiltration) which during its higher discharge months is linked to snowmelt. Another key aspect is related to the storage parameters, that allow either to retain or to release the water as a function of the stress (natural and anthropogenic) to which the aquifer is subject;

e) in the downstream area, the water table is always above the riverbed surface, meaning that the river is a sink for the aquifer. In this area, about 80% of the water pumped by the selected well comes from the base flow component of the river discharge, during the whole simulated period. In the upstream area, the water table is always beneath the riverbed; however, during only one of the two simulated high flow periods, along a river segment (1.5 km length), the water table overcomes the riverbed by 0.5 m. In this area the contribution of the well pumping components of the water budget changes as a function of the hydraulic conditions of the aquifer, however the stream depletion is always less than the depletion with respect the downstream area.

In conclusion, the upstream area is a better location for a new well to be drilled, because pumping will affect the river discharge less than in the downstream area.

The present work showed that the use of SFR2 Package may be extended to a particular hydrogeological context (i.e. Alpine valley aquifers) where complex relationships between the river and groundwater could change over space and time due to natural factors (e.g. recharge) and/or human activities (e.g. well pumping). The methodology presented here allows researchers to evaluate the impact of the well pumping on the whole system in terms of both the water table lowering and depletion of exchanged water between the river and aquifer. Furthermore, it is able to assess the origin of the water pumped by wells.

This work can provide to decision-makers and stakeholders a useful tool to understand, manage and quantify the aquifer and the streamflow depletion, not as separate entities, but by considering them as an integrated system. Furthermore, it could be used to simulate the transport of dissolved pollutions, evaluating their impact both on groundwater and surface water (Bonomi et al. 2015; Stefania et al. 2015; Rotiroti et al. 2015).

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FIGURES

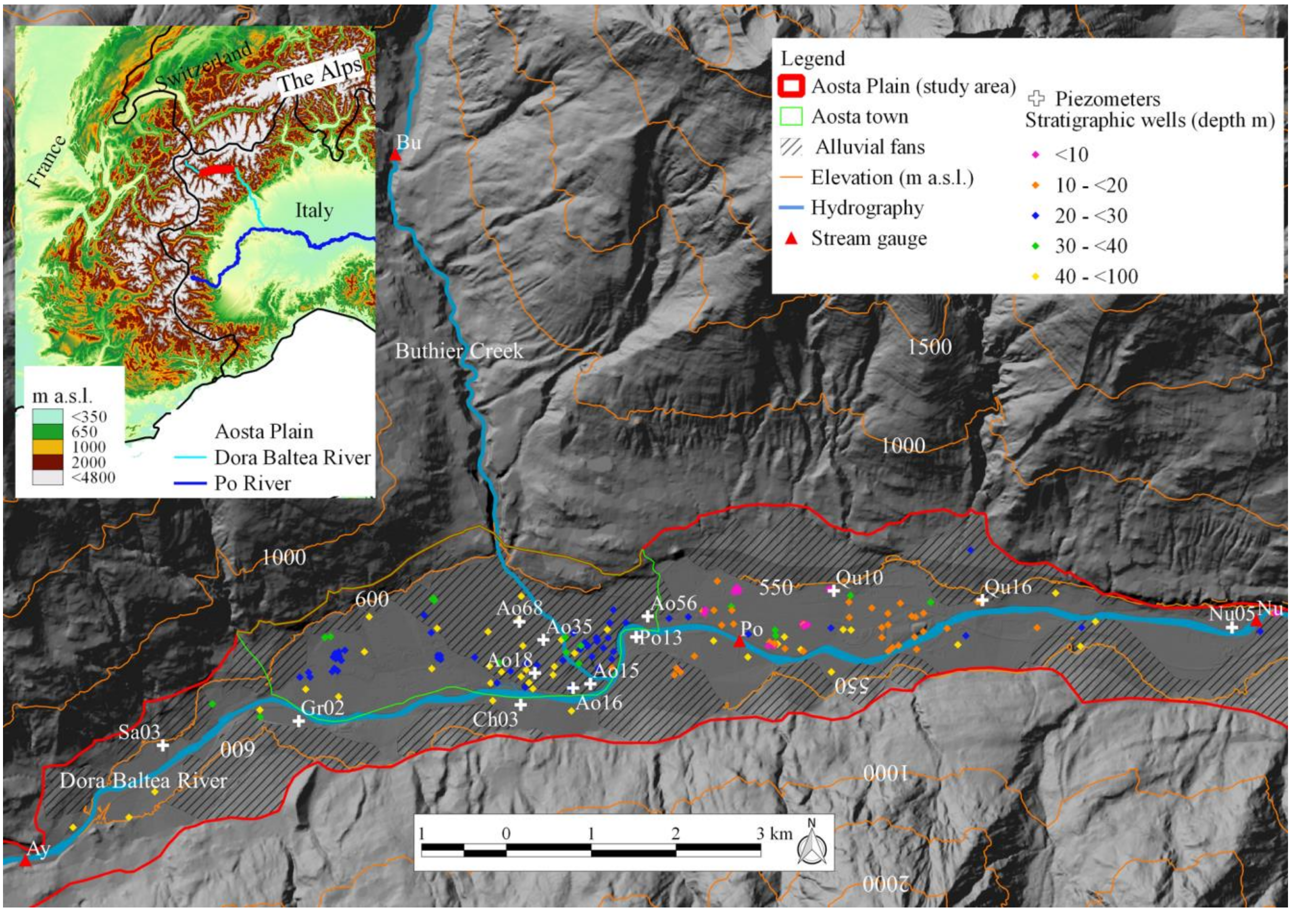


Fig. 1 - The surface water network of the Aosta Plain with locations of available hydrometric stations, piezometers for the monitoring of water-table elevation, and stratigraphic wells.

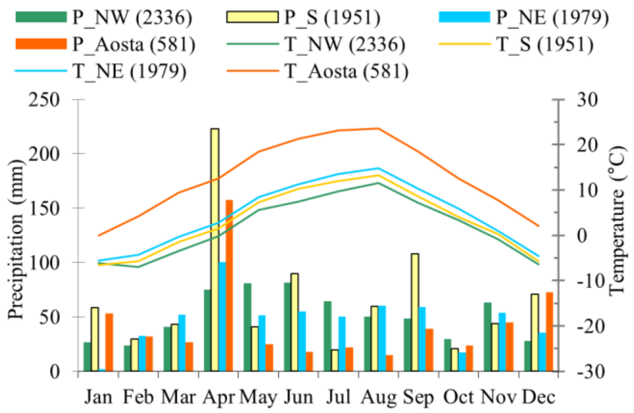


Fig. 2 - Average monthly temperature (C°) and precipitation in different sectors of the Aosta Valley Region. Number in parentheses is the topographic elevation (m a.s.l.) of the station

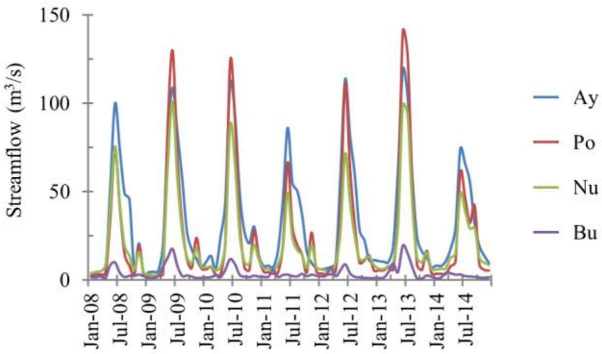


Fig. 3 - Average monthly streamflow (m^3/s) of the Dora Baltea River (Ay, Po, Nu) and Buthier Creek (Bu)

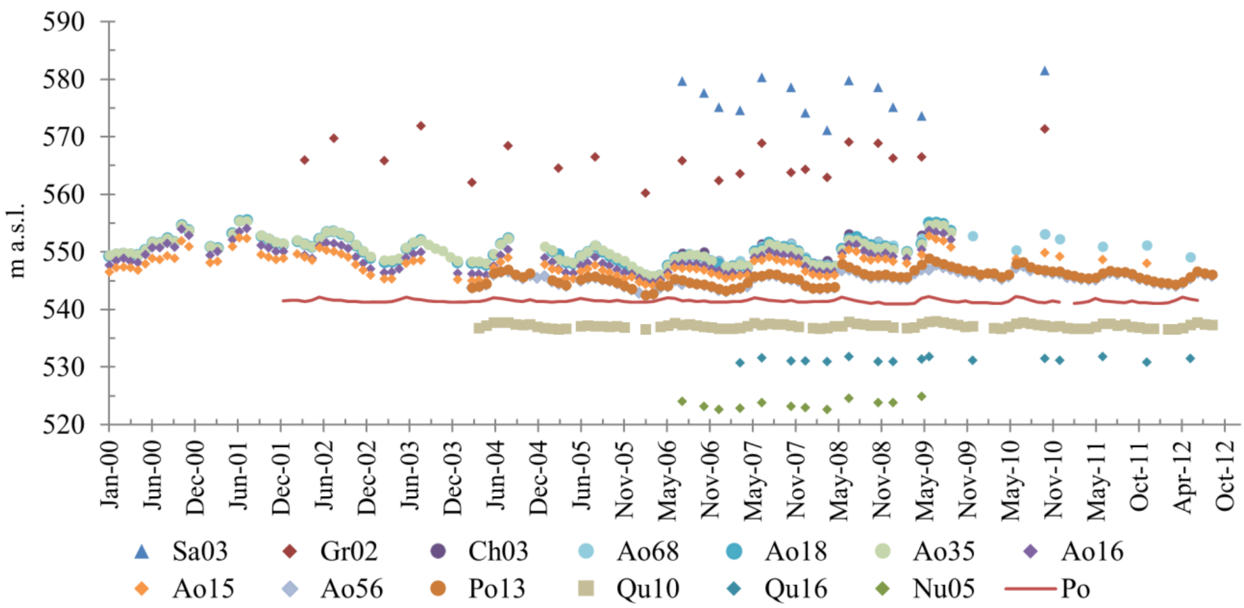


Fig. 4 - Water table elevation (m a.s.l.) measured in monitoring piezometers (see Fig. 1 for locations) and in the Dora Baltea River (Po station) between 2000 and 2012 in the Aosta Plain

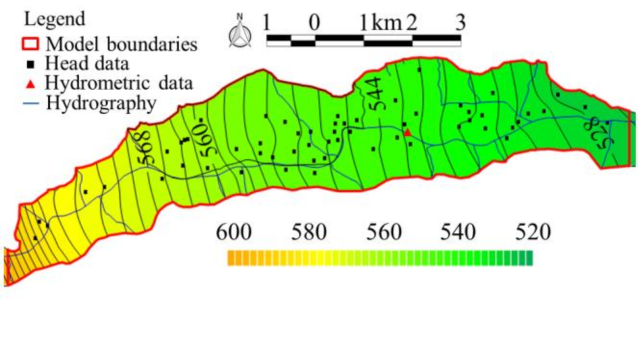


Fig. 5 - Groundwater head (January 2009) with a contour every 2 m, used as the initial condition for the steady-state model. (modified after Bonomi et al. 2013)

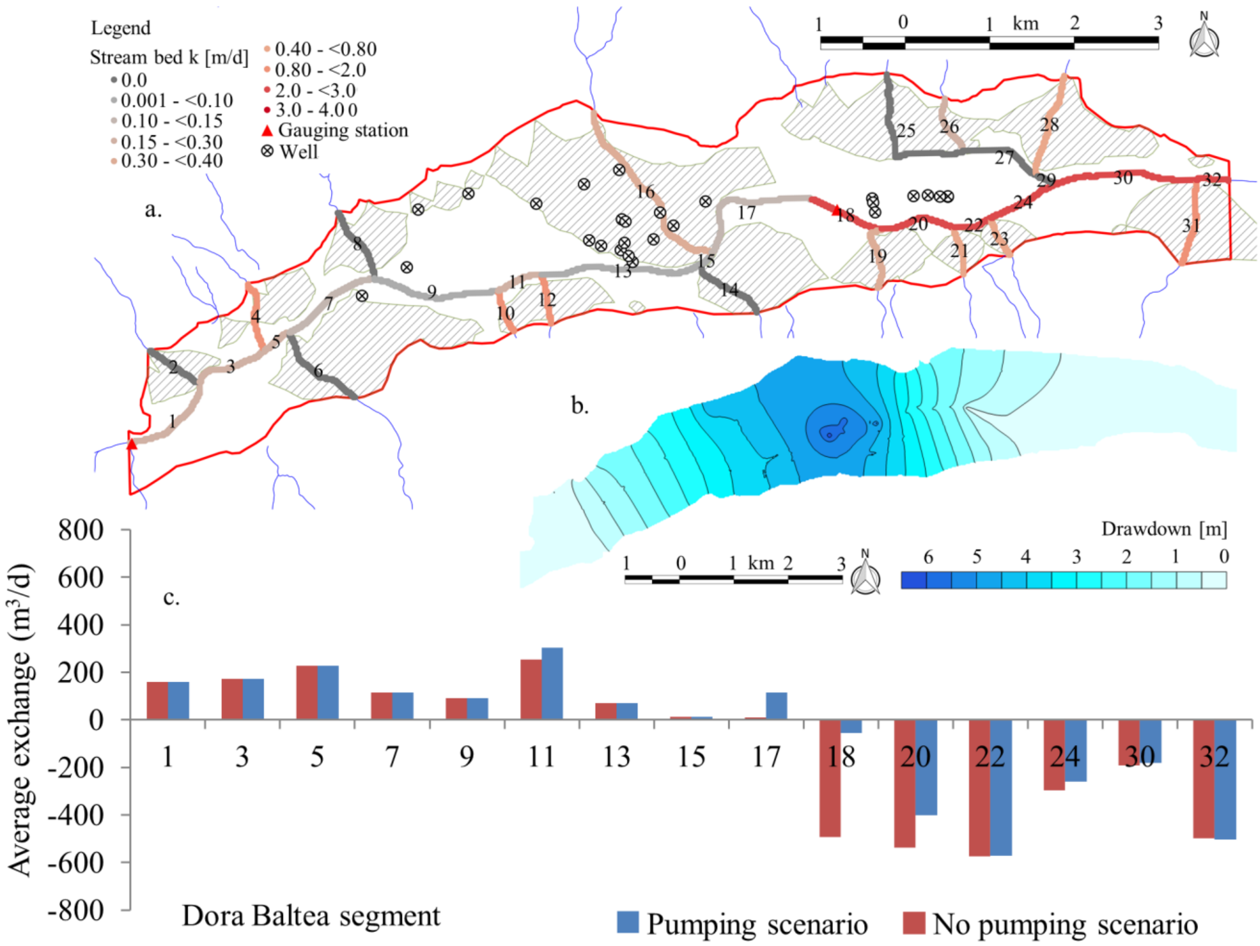
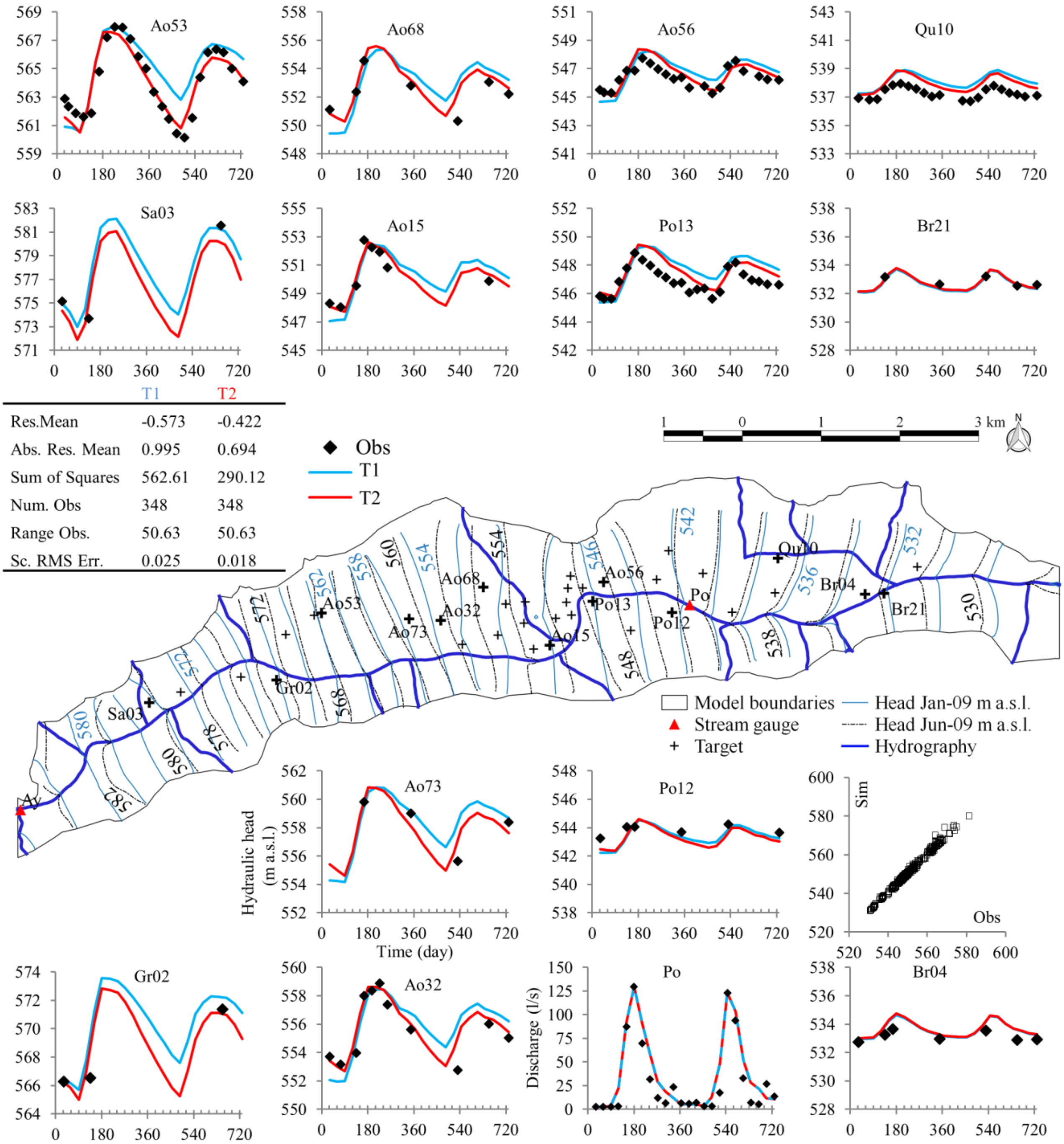


Fig. 6 – (a) Segment numbers and streambed hydraulic conductivity values (m/d) of the Dora Baltea River and its tributaries, and locations of pumping wells. (b) The difference in drawdown between pumping and no pumping scenarios (m). (c) Average monthly exchange between the Dora Baltea River and the surrounding aquifer for each segment (m^3/d); blue bars represent the pumping scenario and red bars represent the no-pumping scenario.



	T1	T2
Res.Mean	-0.573	-0.422
Abs. Res. Mean	0.995	0.694
Sum of Squares	562.61	290.12
Num. Obs	348	348
Range Obs.	50.63	50.63
Sc. RMS Err.	0.025	0.018

Fig. 7 - Simulated head (m a.s.l.) and flux (m^3/s) referred to two transient solutions. Blue line (T1): solution with the model parameters from the steady state model; Red line (T2): solution with the second set of parameters from the calibrated transient model.

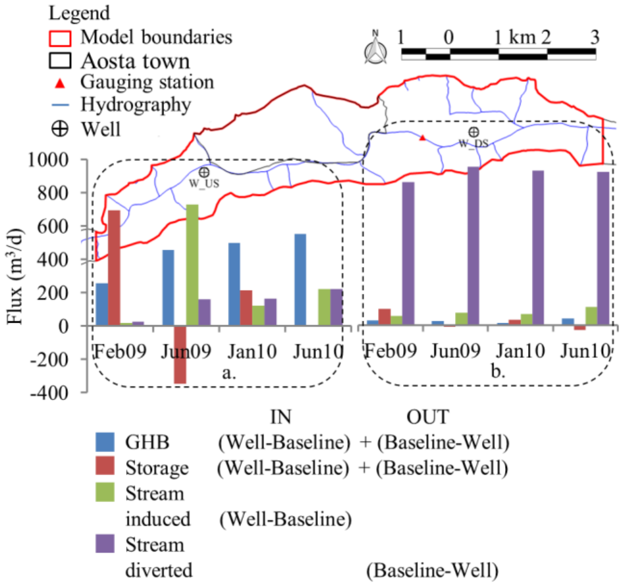


Fig. 8 - Water budget of the transient solution. Negative values (dashed line) represent water leaving the aquifer, whereas positive values (continous line) represent water entering the aquifer.

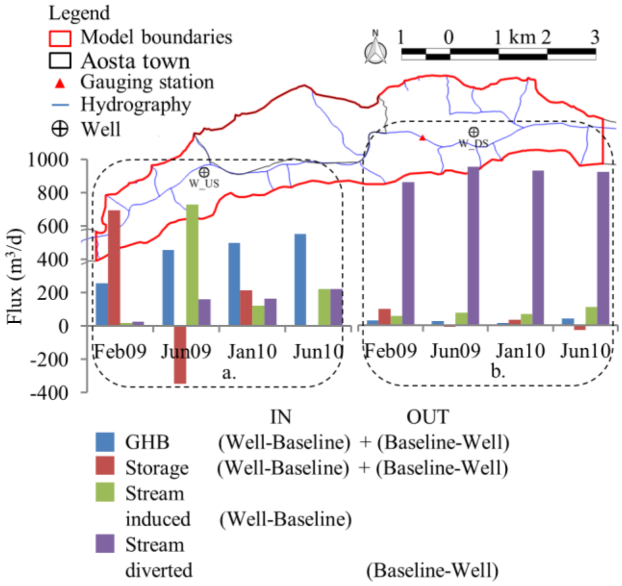


Fig. 9 - Well-pumping impact on the groundwater and surface water balance. W_US: well located in the upstream area; W_DS: well located in the downstream area. The legend shows the formulas used for the calculation of the flux variations induced by well pumping (i.e. W_US and W_DS), between the reference solution (baseline scenario) and (a) scenario A and (b) scenario B.

TABLE

Table 1 – Applied bounds and final values of the selected parameters (k - m/d, ss - 1/m) by sensitivity analysis

		initial	lower	upper	final
kx_2	gravel	571	143	1000	517.943
kx_3	sandy gravel	127	32	222	197.34
kx_4	sand (semi_conf Po)	100	25	175	162.65
kx_7	sandy gravel (fan)	164	41	289	253.25
kx_11	sand (unexp_acq)	69	17	121	101.11
kx_12	silty clay (deep aquitard)	42	11	74	37.04
Ss_1	silty clay (deep aquitard)	$5x10^{-4}$	$2.5x10^{-4}$	$7.5x10^{-4}$	$6.03x10^{-4}$
Sy_3	sandy gravel	0.2	0.01	0.28	0.17

Table 2 – Water balance (m^3/d) for the pumping and no-pumping scenarios.

	Pumping (P)		No Pumping (NP)		Pumping vs No Pamping	
Terms of the water balance	IN	OUT	IN	OUT	IN_P-IN_NP	OUT_NP-OUT_P
General Head Boundary	30282	31356	19579	32759	10703	1403
Recharge	21521		21524			
Well withdrawal		62849				
Dora Baltea River	82231	90130	66017	124544	16213	34414
Tributaries	50304		50185		119	
Total	184337	184335	157305	157303	27035	35817