

Climate change impact on groundwater resources of a hard rock mountain region (Serra da Estrela, Central Portugal)

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Abstract Water resources in mountain regions are at risk due to demographic and economic growth and climate change. This paper presents an evaluation of the impacts of climate change on the groundwater resources of Serra da Estrela Mountain (Portugal). The changes in the water resources in the last 30 years of the twenty-first century have been evaluated with respect to the hydrometeorological conditions of the control period 1975–2005. The predictions for the period 2069–2099 were made by using the Representative Concentration Pathways RCP4.5 and RCP8.5, which are two climate scenarios of the EURO-CORDEX project. The climate scenarios RCP4.5 and RCP8.5 cover a reasonably wide range of possible future trends. The impacts of the climate change have been assessed by using the simulated daily temperature and precipitation values from the climatic models and by solving the daily hydrological water balance model with the code VISUAL-BALAN. The mean annual temperature for the RCP4.5 and RCP8.5 scenarios will increase 3.1 and 5.4 °C, respectively. The increase of temperature in the winter will reduce snow precipitation and favor the melting of the snow cover in the highest sub-basins. The mean

annual precipitation will decrease from 8% (RCP4.5 scenario) to 15% (RCP8.5 scenario). The mean annual snow precipitation will decrease drastically from 54 to 84%. The mean interflow and aquifer recharge will decrease from 12 to 22%. The mean streamflow will decrease from 10 to 18%. The largest decrease in monthly streamflows will occur from March to May due to the decrease of rainfall and snow precipitation. Monthly streamflows during the snow melting season will decrease from 37 to 45%. It can be concluded that the water resources in the Serra da Estrela mountain basin will be very vulnerable to the predicted changes in precipitation and temperature.

Keywords Climate change · Water balance · Serra da Estrela · Groundwater recharge · Mountain hydrogeology · Hard rocks · VISUAL-BALAN

Introduction

Mountains are the source of a significant part of the Earth's liquid freshwater. Mountain regions play an essential role in providing high-quality water to the population in many countries. Mountain water resources are fundamental for millions of people throughout the world and, therefore, achieving a sustainable balance between supply and demand is essential for human wellbeing. The study of mountain water resources has gained recently more scientific and political attention (Viviroli and Weingartner 2008).

Water supply is one of the most important ecosystem services provided by mountains. This supply is very sensitive to changes in environmental drivers such as climate and land use. On the other hand, water demand is globally increasing due to rapidly altering water consumption

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patterns related to demographic and economic growth (Jong 2015). The importance of mountains in freshwater supply is a result of its specific hydrologic functioning (Messerli et al. 2004; Vergara et al. 2011). Rain and snow precipitation in high areas is often significantly larger than in the surrounding lowlands due to the orographic effect. Mountains also contribute to control the distribution of water resources in space and time, with important socio-economic benefits for the populations living downstream (Viviroli et al. 2007). Water stored in mountains provides surface and groundwater flows which may reach great distances and often represent the major part of the freshwater resources of the catchment, especially in arid regions.

Hydrological modeling of mountain areas is particularly difficult due to the complex spatial distribution of the factors controlling precipitation and air temperature (Espinha Marques et al. 2011; Krogh et al. 2014; Samper et al. 2015a). Therefore, scientific studies of mountain water resources are fundamental to overcome this knowledge gap and achieve a sustainable management of water resources in a context of climate change and prevent human conflicts in transboundary mountain catchments.

High altitude territories are particularly sensitive to climate change and, consequently, important modifications in basin scale hydrological processes such as snow–rainfall ratios, evapotranspiration, infiltration, aquifer recharge and runoff patterns should be expected (Provenzale and Palazzi 2015; Jong 2015; Lutz et al. 2016).

The most widely used climatic experiments in Europe include: (1) PRUDENCE (Christensen et al. 2007; AEMet 2008); (2) ENSEMBLES (Van der Linden and Mitchell 2009); and the most recent EURO-CORDEX (<http://www.euro-cordex.net/>). PRUDENCE and ENSEMBLES simulations for Europe have spatial resolutions of 50 and 25 km, respectively. The EURO-CORDEX simulations provide regional climate projections for Europe at 50 and 12.5 km resolution, thereby improving the resolution of the former projects PRUDENCE and ENSEMBLES (Jacob et al. 2014). The climatic simulations carried out in the projects PRUDENCE and ENSEMBLES consider several emission scenarios from the Special Report on Emission Scenarios (SRES) (Nakicenovic et al. 2000). PRUDENCE simulations (AEMet 2008) use SRES scenarios A2 and B2, which cover a reasonable range of conditions. The A2 scenario describes a very heterogeneous world with preservation of local identities. It also considers that the global population will continuously increase and that the economic development will be regionally oriented and technological change will be slower than in other scenarios (Nakicenovic et al. 2000). The B2 scenario foresees a world with continuously increasing population at a rate lower than A2 and intermediate levels of economic

development (Nakicenovic et al. 2000). ENSEMBLES simulations consider only the A1B SRES emission scenario, which lies between A2 and B2 scenarios. SRES scenarios of PRUDENCE and ENSEMBLES explicitly specify socioeconomic scenarios. EURO-CORDEX simulations use the Representative Concentration Pathways (RCPs), which take a different approach. RCPs were defined for the Fifth Assessment Report (5AR) of the IPCC (Moss et al. 2010) and do not specify socioeconomic scenarios. RCPs consider different pathways to target radiative forcing at the end of the twenty-first century (Jacob et al. 2014). RCP4.5 and RCP8.5 scenarios assume that the increase in radiative forcing at the year 2100 relative to pre-industrial conditions will be 4.5 and 8.5 W/m², respectively (Thomson et al. 2011). The RCP8.5 scenario is characterized by greenhouse gas emissions that increase over time and is representative of high greenhouse gas concentration levels. The scenario described by the RCP8.5 can be considered as a non-mitigation situation. The RCP4.5 is a more moderate scenario where total radiative forcing is stabilized before the year 2100 by employing a range of technologies and strategies for reducing greenhouse gas emissions (Clarke et al. 2007). RCP4.5 can be considered as a weak climate change mitigation scenario. Rogelj et al. (2012) presented a comparison of the SRES and RCP scenarios and found that the A1B scenario (used in ENSEMBLES) leads to a global mean temperature increase from 2.8 and 4.2 °C, which lies within the values of RCP4.5 and RCP8.5 scenarios (Jacob et al. 2014).

There are several methods to evaluate the impacts of the climate change on water resources. They usually involve comparing the results of the hydrological model in the control period and in the prediction period. One common problem that the modeller usually confronts is that the results (or simulations) of the climate model do not reproduce correctly the observed climate normals in the historical period. The biases of the simulations also affect the predictions for the future climate scenarios (Wilby et al. 2000; Crane et al. 2002). One method to overcome this limitation is to evaluate the climate change impacts by running the hydrological model with the data computed by the climatic model for both the control and the prediction periods and comparing the results. The drawback of this approach is that the absolute results of the hydrological model are of little interest. The impacts of the climate change are calculated from the relative differences between the results of the hydrological model in the control and prediction periods (Akhtar et al. 2008; CEDEX 2011). A different method consists of using the observed meteorological data for the control period and the results of the climate model in the prediction period after correcting the biases. The biases of the climate normals in the control period are calculated by using the available observed and

simulated meteorological data. The biases in the control period can be calculated by computing monthly linear regressions between observed and modeled values of precipitation and temperature (Wood et al. 2002; Vidal and Wade 2008). Then, the biases of the climate normals in the prediction period are removed by applying the calculated monthly linear regressions. Afterwards, the monthly values in the prediction period are transformed into daily time-series by using a weather generator (Álvarez et al. 2009). This methodology was applied by Álvarez et al. (2009), Stigter et al. (2012) and Samper et al. (2015b) to some Spanish basins. The direct use of the climatic model results has the advantage of preserving the day-to-day variability of the climatic variables simulated by the climatic models (Hay et al. 2002; Wood et al. 2004; CEDEX 2011) and this leads to a much more accurate evaluation of the water balance components when using daily time steps as the hydrological code VISUAL-BALAN does (Stigter et al. 2012; Pisani et al. 2013).

A lot of studies on the impacts of climate change on water resources have been published in recent years, many of them about European mountain regions, especially in the Alps and Pyrenees (López-Moreno et al. 2008, 2014; Nogués-Bravo et al. 2008; Viviroli et al. 2011; Beniston and Stoffel 2014; Gobiet et al. 2014). Two good examples of these are the so-called CIRCLE (Climate Impact Research Coordination for a Larger Europe) on “Climate change impacts and response options in mountainous areas” funded by the 6th EU Framework Programme and the project ACQWA [Assessing Climate impacts on the Quantity and quality of Water (Beniston and Stoffel 2014)] which was funded by the 7th EU Framework Programme under the subject “Climate change impacts on vulnerable mountain regions”. Nevertheless, water resources studies at the scale of small basins are scarce (Viviroli et al. 2011). Meixner et al. (2016) show that one of the greatest sources of uncertainty in the evaluation of the impacts of climate change on groundwater recharge in North America is the limited quantity of “studies quantitatively coupling climate projections to recharge estimation methods by using detailed, process-based numerical models”. This is also true for mountain areas in Europe.

Most of the studies in European mountains show that the climate change will affect the quantity, the timing and possibly the quality of the water resources (Stewart 2008; Beniston and Stoffel 2014; Mora Alonso-Muñoyerro et al. 2016). López-Moreno et al. (2008) analyzed the environmental changes observed on hydrology during the 20th century in the Pyrenees and possible future trends. These authors found that the future equilibrium between resources and water demand will be threatened in the Pyrenees as well as in many other Mediterranean mountain areas.

Mountain snowpack and spring runoff are key components of surface water resources in mountain regions.

Stewart (2008) showed that warmer temperatures at mountain areas had decreased snowpack and caused the earlier melt of snow, especially at mid-altitude zones which are above or near freezing temperatures during the winter. However, local climatic conditions may not be consistent with the general findings and thereby there is a research need for good quality meteorological data from monitoring stations located at mid and high altitudes (Stewart 2008).

Gobiet et al. (2014) presented a review of the state-of-knowledge about climate change in the Alps during the twenty-first century. Gobiet et al. (2014) concluded that warmer temperatures will probably be associated with changes in the seasonality of precipitation, global radiation and relative humidity as well as with more intense precipitation events. Also, Gobiet et al. (2014) predict that snow cover is expected to decrease significantly at altitudes below 2000 m.

It is well known that future climate simulations contain uncertainties which often may be significant. Viviroli et al. (2011) noted that, besides the uncertainties associated to the climate models, the climate of European mountains such as the Pyrenees and the Alps located at transition zones between Mediterranean and Atlantic conditions show a high variability due to the strong altitudinal gradients and exposure to solar radiation. Notwithstanding the uncertainties, most of the published studies on European mountains point to the following trends during the twenty-first century: (1) rainfall and snow precipitation will decrease, especially in spring; (2) runoff will decrease in the spring and the summer and will increase in the fall and the late winter in some mountain areas; (3) snow melt will take place earlier and its contribution to the runoff may decrease significantly; and (4) water resources will decrease.

Here, we present an update of the hydrological model of the Serra da Estrela hard rock mountain basin and a detailed evaluation of the climate change impacts on the water resources for the last 30 years of the twenty-first century by using up-to-date climatic simulations in Europe. The paper starts with a description of the Serra da Estrela Basin. Then, the hydrological water balance model is presented. The selected climate change scenarios are described afterwards. After that, the effects of climate change on the water resources are presented. The paper ends with the main conclusions.

Description of the study area

Serra da Estrela (40°15–38'N; 7°18–47'W) is located in the Western part of the Central Iberian mountain range (see Fig. 1). The main crustal deep structure is the NNE-SSW Bragança-Vilariça-Manteigas fault zone (Ribeiro et al. 2007).

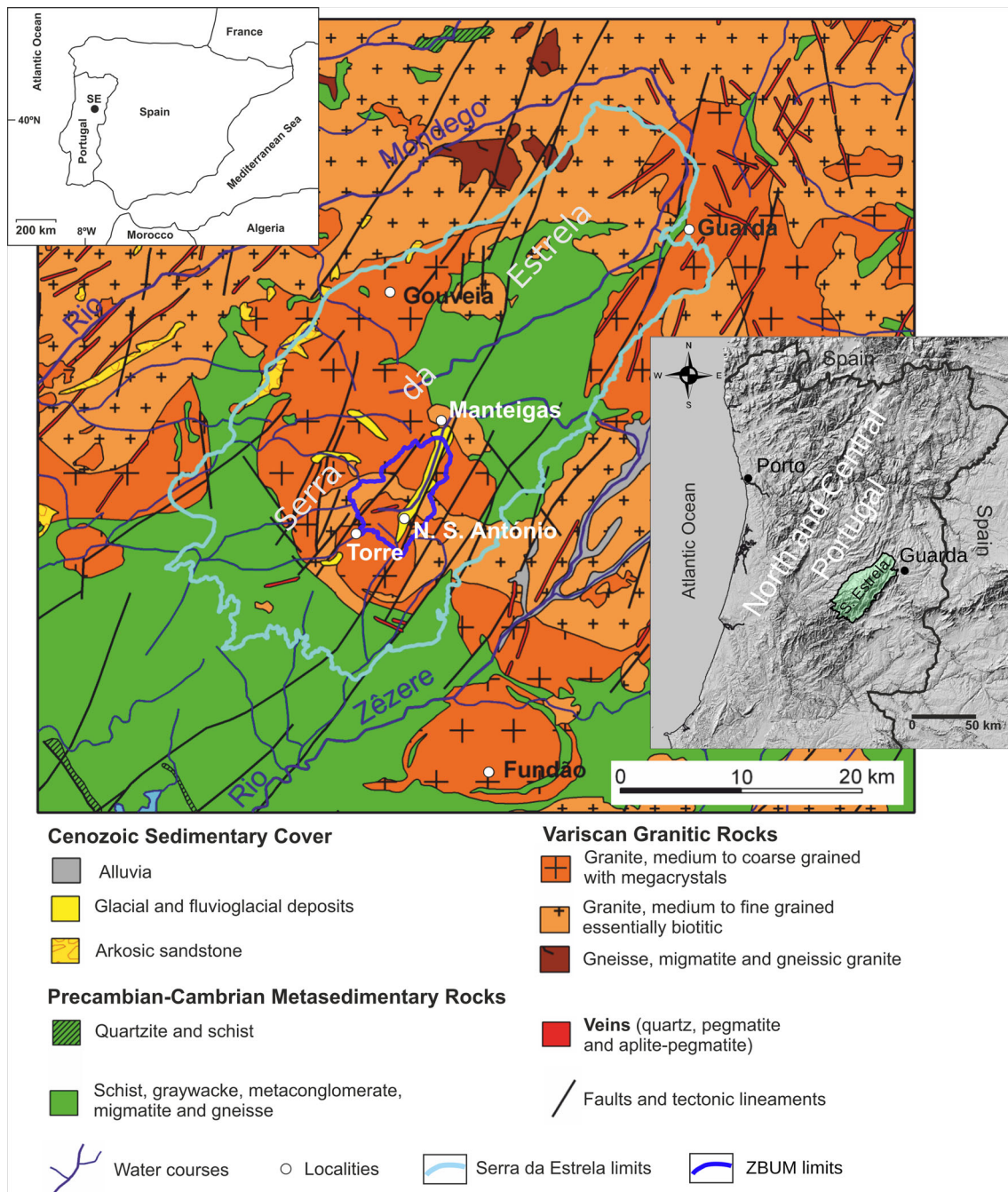


Fig. 1 Geological map of Serra da Estrela region (adapted from Oliveira et al. 1992). ZBUM River Zêzere Drainage Basin Upstream of Manteigas, BVMFZ Bragança–Vilarica–Manteigas fault zone

The study region is the River Zêzere Drainage Basin Upstream of Manteigas (ZBUM) which has an area of around 28 km² and a mean altitude of 1505 m a.s.l. (Fig. 2). The streamflow gauge station at Manteigas is the lowest point (875 m a.s.l.) and the Torre summit is the highest point (1993 m a.s.l.). Other geomorphological features are: a mean slope of 20°; a perimeter of 24 km; a shape index (basin area/length of the basin's longest axis)

of 0.39; a Gravelius compactness index ($0.28 \times \text{basin perimeter}/\text{basin area}^{1/2}$) of 1.31.

The regional geomorphology is also characterized by two major plateaus, separated by the NNE-SSW U-shaped glacial valley of the Zêzere River, as well as by other Late Pleistocene glacial landforms and deposits originated during the Last Glacial Maximum (Daveau et al. 1997; Vieira 2008).

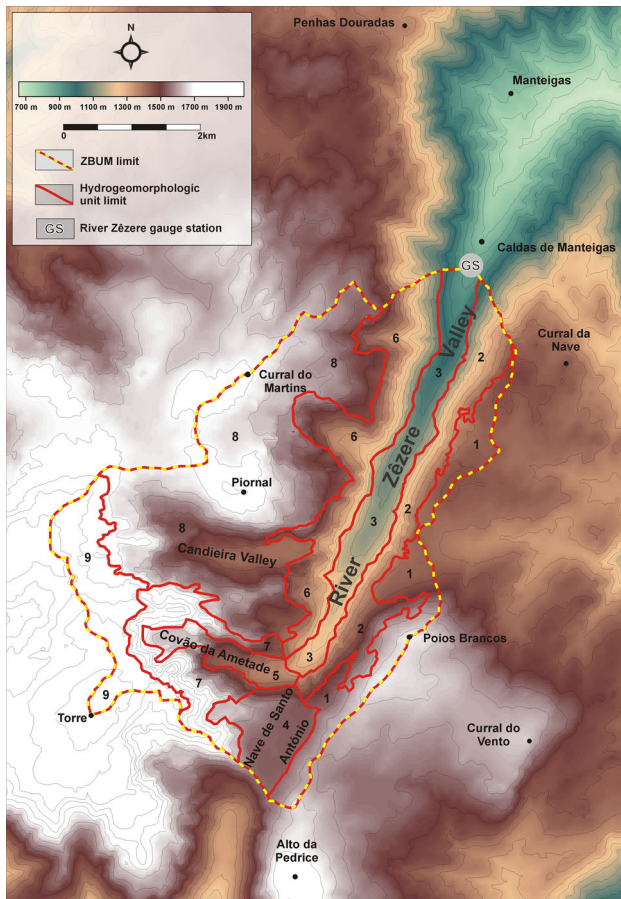


Fig. 2 Hypsometric features of the Zêzere river drainage basin upstream Manteigas. Hydrogeomorphic units are also shown: (1) Eastern plateau; (2) Zêzere valley eastern slopes; (3) Lower Zêzere valley floor; (4) Nave de Santo António; (5) Upper Zêzere valley floor; (6) Zêzere valley western slopes; (7) Cântaros slopes; (8) Lower western plateau; and (9) Upper western plateau (modified after Espinha Marques et al. 2006)

The following hydrogeological units were defined in the study area (Figs. 1, 3): (1) sedimentary cover; (2) metasedimentary rocks; and (3) granitic rocks. The underground phase of the local water cycle is controlled by the geological and tectonic conditions which determine infiltration, aquifer recharge, type of flow medium (porous vs. fractured), type of groundwater flow paths and hydro-geochemistry (Espinha Marques et al. 2013).

The Köppen–Geiger climate classification is Csb, which is typical of Northwestern Iberia (Kottek et al. 2006; Peel et al. 2007), and corresponds to warm temperate, with dry and warm summers. Yet, the mountain massif is located very close to the transition to the Csa climatic subtype (warm temperate, with dry and hot summers) which is dominant to the South of Serra da Estrela. Consequently, the Serra da Estrela climatic conditions are complex due to the combined Mediterranean and Atlantic influences.

Precipitation is mainly controlled by slope orientation and altitude (Mora 2006). Total precipitation in the Western side of the mountain is smaller than that at the Eastern part, even though the number of days with rainfall in the Western side is larger than on the Eastern part. In general, precipitation increases with altitude. However, precipitation at a local scale shows complex patterns due to the complex behavior of the air mass fluxes, air divergence and convergence mechanisms which are controlled by the mountain morphology. The mean annual precipitation is around 2500 mm in the highest areas. Rain and snow precipitation is mainly dependent on slope orientation regarding the North Atlantic air circulation patterns and altitude (Daveau et al. 1997; Mora 2010). The mean annual air temperatures are below 7 °C in most of the plateau area and as low as 4 °C near the summit.

Available data on snow precipitation and depths are scarce and of poor quality (Mora and Vieira 2004). Monthly precipitation P , and temperature T , data from 1953 to 1983 are available at the meteorological stations of Gouveia, Seia, Vale de Rossim, Valhelhas, Covilhã, Celorico da Beira, Fornos de Algodres, Penhas Douradas, Lagoa Comprida, Penhas da Saúde and Fundão.

Hydrological water balance model

The numerical model of the hydro-meteorological water balance of ZBUM was solved with the code VISUAL BALAN v2.0 (e.g., Samper et al. 1999, 2007, 2015a; Espinha Marques et al. 2011), which is based on a semi-distributed model that performs daily water balances in the soil, the underlying unsaturated zone and the aquifer.

VISUAL BALAN evolved from earlier versions of the code which had the generic name of BALAN (Samper and García-Vera 1992; Samper et al. 1999). The hydrological components are evaluated daily in a sequential manner. The balance equation in the soil is given by:

$$P + D - I_n - O_f - AET - R_p = \Delta\theta \quad (1)$$

where P is precipitation, D is irrigation water, I_n is interception, O_f is overland flow, AET is actual evapotranspiration, R_p is potential recharge (which coincides with groundwater recharge if there is no interflow) and $\Delta\theta$ is the change in soil water content.

The potential recharge, R_p , is the main input to the underlying unsaturated zone where water may flow horizontally as interflow or percolate vertically as groundwater recharge. Potential recharge may flow horizontally and discharge into the atmosphere as interflow if the topography permits so. The rest of the potential recharge percolates downwards to the aquifer and becomes the aquifer recharge. The conceptual model implemented in VISUAL-



Fig. 3 Some aspects of the study area: **a** snow covered granitic peaks; **b** Zêzere river valley bottom; **c** Nave de Santo António alluvial (foreground), fluvioglacial deposits (intermediate plan) and granitic

slopes (in the background); **d** fluvioglacial deposits at Manteigas village; **e** granite showing sub-horizontal fracturing near the Torre summit; **f** glacial valley of Zêzere river

BALAN assumes that R_p , while descending towards the aquifer, may form perched aquifers over low-permeability layers, especially during groundwater recharge episodes (Fig. 4).

The daily values of interflow, Q_h , and percolation, Q_p , in mm/day are calculated from the following expressions (Samper et al. 1999):

$$Q_h = \alpha_h V \quad (2)$$

$$Q_p = K_v + \alpha_p V \quad (3)$$

where V (mm) is the water content of the unsaturated zone expressed as equivalent water height (volume per unit surface area), K_v (mm/day) is the saturated vertical hydraulic conductivity of the low-permeability layer and α_h (day^{-1}) and α_p (day^{-1}) are discharge or recession coefficients of interflow and recharge, respectively.

The interflow recession coefficient, α_h , is related to the horizontal hydraulic conductivity, K_h , the porosity of the unsaturated zone, Φ , the mean hill slope, i , and the distance between the top and the bottom of the hill, L , through the following expression (Kirkby 1978; Samper et al. 1999):

$$\alpha_h = \frac{2K_h i}{L\Phi}. \quad (4)$$

The percolation Q_p in Eq. 3 is derived by applying Darcy's law between points A and B (Fig. 4) and by assuming that the effective vertical hydraulic conductivity is that of the low-permeability layer, K_v :

$$Q_p = K_v \frac{(y_h + p + b)}{p + b}, \quad (5)$$

where y_h is the water head in the perched aquifer, b is the thickness of the low-permeability layer and p is the

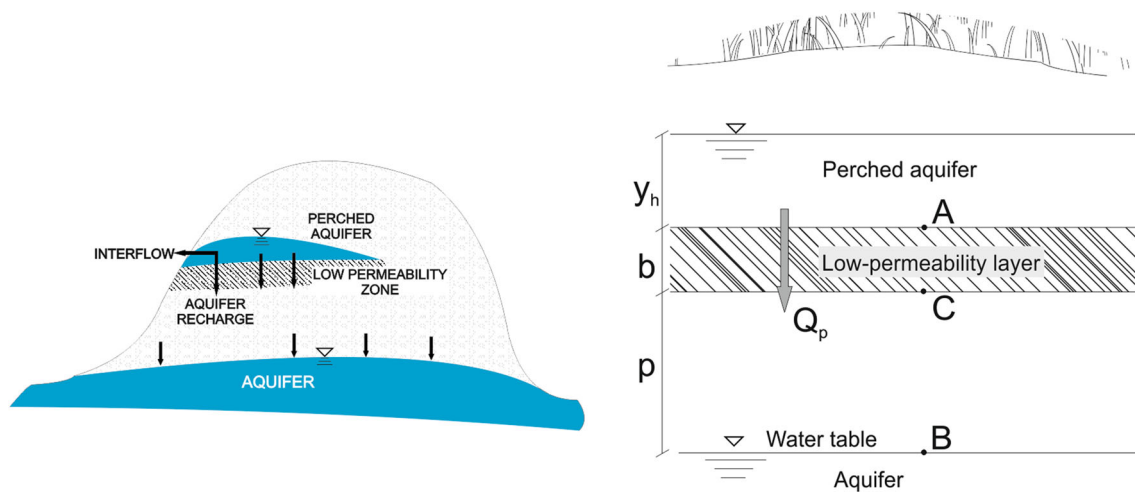


Fig. 4 Conceptual model of interflow with a perched aquifer in the unsaturated zone (*left*) and detail of the calculation of the percolation based on Darcy's Law (*right*) (Samper et al. 2015a)

distance between the bottom of the low-permeability layer and the water table (Fig. 4). If $b \ll p$ then:

$$Q_p = K_v \frac{(y_h + p)}{p}. \quad (6)$$

The water content in the unsaturated zone V and the saturated thickness of the perched aquifer y_h are related through:

$$y_h = \frac{V}{\Phi}. \quad (7)$$

Substituting Eq. 7 into Eq. 6, one obtains Eq. 3, where the recession coefficient of the percolation, α_p , is related to K_v , p and Φ through:

$$\alpha_p = \frac{K_v}{p\Phi}. \quad (8)$$

The numerical model implemented in VISUAL-BALAN to compute interflow and the aquifer recharge leads to an excellent fit between computed and measured streamflows. Interflow is often the largest component of the total runoff in mountain basins (Hewlett 1961; Eckhardt et al. 2002; Sophocleous 2002; Becker 2005; Samper et al. 2015a). According to Xu et al. (2013), near-surface lateral flow in the unsaturated soil layer is a critical hydrological process for runoff generation and failing to consider it may result in the overestimation of baseflow.

The water balance in the aquifer can be calculated with VISUAL BALAN with a lumped model (single cell) or with a 1-D transient distributed model consisting of N interconnected cells. Fluxes across cells are computed by using an explicit finite difference method. Natural groundwater discharge occurs at springs, rivers or other water bodies. Changes in the water stored in the aquifer (ΔV_a) per unit surface area in the single cell model are

related to changes in piezometric heads (Δh) through $\Delta V_a = S\Delta h$, where S is the storage coefficient. The total outflow from the basin is computed as the sum of overland flow, interflow and groundwater discharge. A detailed account of the methods and parameters of VISUAL BALAN can be found in Samper et al. (1999) and Castañeda and García-Vera (2008).

VISUAL BALAN was first applied in the study area as a lumped model for the hydrological years from 1986–87 to 1994–95 by using mean daily air temperature and daily precipitation values from the meteorological station of Penhas Douradas which is situated at 1380 m a.s.l., close to the mean altitude of ZBUM (1505 m a.s.l.). The Zêzere river gauge station, that operated until 1996, provided daily streamflow data from 1986–1987 to 1994–1995. The results from this modelling step were not satisfactory due to two main factors: (1) the high spatial variability of its lithology, landforms, climate, soil type and land cover of the basin and (2) the mean annual precipitation from Penhas Douradas for this period (1406 mm) is much lower than the value from the climate normals (1799 mm)—INMG (1991).

The limitations of the lumped model were overcome with a semi-distributed hydrological model which considers 9 sub-basins as described by Espinha Marques (2007)—Fig. 2.

Monthly precipitation P , and temperature T , data from 1953 to 1983 can be adequately characterized by linear regression of P and T against altitude x , through the equation $y = a + bx$, where y is the mean monthly precipitation (or temperature). Monthly regression lines vary throughout the year, so the intercept a and the slope b of the regression lines were calculated for P and T for each month (Espinha Marques et al. 2006, 2011).

The monthly precipitation of the m th month at the centroid of a sub-basin, P_C^m , located at an altitude z_C is calculated from the measured monthly precipitation at a reference meteorological station, P_{ref}^m located at an altitude z_{ref} with the following equation:

$$P_C^m = P_{ref}^m + b_P^m(z_C - z_{ref}), \quad (9)$$

where b_P^m is the slope of the linear regression equation for monthly precipitation versus altitude. Similarly, the mean monthly temperature of the m th month at the centroid of a sub-basin, T_C^m , is calculated from the measured mean monthly temperature at a reference meteorological station, T_{ref}^m by:

$$T_C^m = T_{ref}^m + b_T^m(z_C - z_{ref}), \quad (10)$$

where b_T^m is the slope of the linear regression equation for mean monthly temperature versus altitude. The daily precipitation at the centroid of a sub-basin in the i th day of the m th month, P_C^i , is assumed to be directly proportional to the measured daily precipitation at the reference station, P_{ref}^i , with a scaling factor equal to the ratio of the monthly precipitations at the centroid and the reference station. P_C^i is calculated according to:

$$P_C^i = P_{ref}^i \frac{P_C^m}{P_{ref}^m} = P_{ref}^i \frac{P_{ref}^m + b_P^m(z_C - z_{ref})}{P_{ref}^m}. \quad (11)$$

The daily temperature at the centroid of a sub-basin in the i th day of the m th month, T_C^i , is calculated from the measured daily temperature at the reference station, T_{ref}^i , by using the temperature/altitude regression equation (Eq. 10):

$$T_C^i = T_{ref}^i + b_T^m(z_C - z_{ref}). \quad (12)$$

The calculated mean annual precipitation for the ZBUM is equal to 2268 mm. This value is similar to those reported previously by Daveau et al. (1997).

The concentration time of the basin is 55 min. Therefore, instantaneous streamflow measurements performed once a day at the same hour may under or overestimate mean daily streamflows, especially in days of heavy precipitation events. This problem was overcome by calibrating model parameters to fit measured monthly streamflows while ensuring a global coherence of the mean annual values of actual and potential evapotranspiration and groundwater recharge with the values reported by others for this study area in previous studies (e.g., Mendes and Bettencourt 1980; Carvalho et al. 2000). The achieved fit between measured and computed flows is excellent (Espinha Marques et al. 2011, 2013). Calibrated model parameters are listed in Espinha Marques et al. (2011).

To achieve a systematic and objective fit to measured streamflow data, the calibration was performed by minimizing the following least-squares objective function, O_1 ,

$$O_1 = \frac{1}{M} \sum_{i=1}^M \left[\log \left(\frac{F_c^i}{F_m^i} \right) \right]^2, \quad (13)$$

where F_c^i and F_m^i are computed and measured streamflows at month i , respectively, and M is the number of monthly streamflow data. The procedure to minimize the objective function consisted in changing one parameter at a time, in an interval of $\pm 20\%$ of its value, except for the soil vertical permeability which was changed by an order of magnitude. The value for which the objective function was smaller was the base for the next iteration, which started after finding the values of the rest of the parameters that lead to a decrease in the objective function. The optimum of the objective function was found after five iterations. The initial value of O_1 obtained with the trial-and-error calibration is 0.495 while its final value was reduced to 0.387. Parameters controlling preferential flow as well as the interflow and groundwater recession coefficients are similar in both calibrations, meaning that these parameters have been calibrated with small uncertainty. On the other hand, there are important changes in the parameters CIM0 and CIM1 which control overland flow and attain values ranging from 41 to 85 mm. Similarly, the percolation recession coefficient ranges from 0.038 to 0.06/day.

The hydrological model of Espinha Marques et al. (2011, 2013) has been updated by extending the simulation period from 9 hydrological years (1986–1995) to 35 years (1980–2015) by using daily data of precipitation and temperature from the Manteigas and Penhas Douradas meteorological stations. Unfortunately, the Manteigas gauging station stopped operating in 1995 and therefore there are no available streamflow data since then.

The goodness of the model fit to the measured streamflow data has been reassessed by computing the Root Mean Squared Error (RMSE) and the Nash–Sutcliffe efficiency with logarithmic values (Krause et al. 2005). The Nash–Sutcliffe efficiency with logarithmic values, ELN, is defined as follows:

$$ELN = 1 - \frac{\sum_{i=1}^N (\text{Ln}O_i - \text{Ln}C_i)^2}{\sum_{i=1}^N (\text{Ln}O_i - \overline{\text{Ln}O})^2}, \quad (14)$$

where $\text{Ln}O_i$ and $\text{Ln}C_i$ are the natural logarithms of the monthly measured (O_i) and calculated (C_i) streamflows, respectively, in the i th month, and N is the number of months with measured data between the hydrological years 1986–1987 and 1994–1995. The use of logarithms helps to reduce the weight of large values of streamflows in ELN.

Table 1 shows the values of ELN and RMSE for the model with the observed meteorological data and with the meteorological data of the climate model for the same historical period. As can be seen in Table 1, the fit of the model with historical (observed) meteorological data is

Table 1 Goodness of the fit of the water balance model for the period 1986–1995 computed with the historical meteorological data and with the meteorological data calculated by the climatic model

Objective function	Model with historical meteorological data		Model with meteorological data calculated by the climate model	
	Monthly series	Annual series	Monthly series	Annual series
RMSE (hm ³)	1.623	2.672	4.238	15.952
ELn (–)	0.854	0.959	0.589	0.181

much better than with data generated by the climate model for the same period, even though the mean annual results of the water balance model are very similar for both data sets. This discrepancy is further analyzed in the next section.

Climate change

The analysis of the climate change impacts relies on assumptions about the future evolution of the environmental and socio-economic conditions. These assumptions are usually grouped or linked together in what are known as projections or scenarios. Each climatic scenario is obtained from the combination of a CO₂ emission scenario, a driving Global Circulation Model (GCM) and a Regional Climate Model (RCM). The climate scenarios for this case study were firstly selected to match the following criteria: (1) good quality data covering the study zone; (2) data covering the period 1960–2100; (3) data with a good spatial resolution; and (4) balanced emission scenarios that span a reasonably wide range of predictions, from the most pessimistic to the most optimistic.

The impacts of the climate change on the water balance in the Serra da Estrela were computed with the most recent results of the European climatic simulations, the EURO-CORDEX project. Climate scenarios RCP4.5 and RCP8.5 were selected because they cover a representative range of future climatic conditions.

Fifty-eight RCMs were registered in the EURO-CORDEX project by the end of the year 2016. Of all the available RCMs, the Hadley Centre's HadGEM2-ES was selected for this work. The HadGEM2-ES model was developed by the British Meteorological Office to run the major scenarios for IPCC 5AR (Moss et al. 2010) and can be considered the Hadley Center's standard climate model. The Hadley Center's models have been also widely used in Portugal (Brandão 2006; Miranda et al. 2006) and were used in the 2nd phase of the SIAM report for calculating the impacts of the climate change at the scale of the country (Santos et al. 2002).

The results of the RCM for the historical period are available in the CORDEX site for the period 1950–2005. Daily values of temperature (*T*) and precipitation (*P*) from the EURO-CORDEX site for the control period,

1975–2005, and the prediction period, 2069–2099 were downloaded from the CORDEX site. The reference period, also called the control period, is the period for which historical data are available and the comparison with the RCM results can be done. The last 30 years of the historical period (1975–2005) were selected for this study to overlap as much as possible with the observed meteorological data available from 1980 to 2015 in the Serra da Estrela. The data of the RCM's cell which best fits the study zone were adopted for this study.

The future climate change scenarios of this study consider that the Serra da Estrela basin will continue to be a protected natural area and so no significant change in the use of soil will occur.

Table 1 shows the values of RMSE and Nash–Sutcliffe efficiency with logarithmic values (ELn) for the water balance model with observed meteorological data and with meteorological data simulated by the climatic model. The fit of the hydrological model to the observed streamflows is much better for the historical meteorological data than for the data calculated by the climate model, for both the monthly and annual time series (Table 1). While this is usual, it is remarkable that the ELn of the fit to the annual streamflows of the hydrological model with the meteorological data simulated by the climatic model (0.181) is much smaller than the ELn of the monthly series (0.589). The water balance model with the meteorological data simulated by the climatic model underestimates the large monthly streamflows. However, this happens only a few months each year. Thus, the underestimation of streamflows affects almost all the values of the annual series, but only a few (in relative terms) of the monthly series. Consequently, the use of ELn with logarithmic values reduces the weight of the discrepancies in the large values of the monthly series more than in the annual time series.

The expected changes of precipitation and temperature for the period 2069–2099 have been calculated with respect to the results of the climatic model for the control period 1975–2005. Table 2 shows the mean annual values of precipitation and temperature for the control period (1975–2005) and the expected changes for the last 30 years of the twenty-first century for RCP4.5 and RCP8.5 scenarios.

It is expected that by the end of the twenty-first century the mean annual temperature will increase 3.1 °C for the

Table 2 Mean annual precipitation and temperature for the control (1975–2005) and prediction (2069–2099) periods and expected changes for the RCP4.5 and RCP8.5 scenarios

	Control 1975–2005	RCP4.5 2069–2099	RCP8.5 2069–2099
Temperature			
Annual mean (°C)	11.4	14.5	16.8
Change (°C)	–	+3.1	+5.4
Precipitation			
Annual mean (mm)	894	715	585
Change (mm)	–	–179	–309
Change (%)	–	–20	–35

RCP4.5 scenario and 5.4 °C for the RCP8.5 scenario. The mean annual precipitation in the period 2069–2099 will decrease 20% for the RCP4.5 scenario and 35% for the RCP8.5 scenario compared to the climate normals of the control period.

Figure 5 shows the mean monthly precipitations and temperatures for the control (1975–2005), RCP4.5 (2069–2099) and RCP8.5 (2069–2099) scenarios. The monthly temperatures will increase throughout the year in both scenarios (Fig. 5). The increase ranges from 2.0 °C in January to 4.3 °C in September for the RCP4.5 scenario and from 3.8 °C in March to 7.4 °C in September for the RCP8.5 scenario. The mean monthly precipitations will

decrease throughout the year. A large decrease of the precipitation is predicted in both scenarios. The maximum decrease of the monthly precipitation will be 54% for the RCP4.5 scenario (in August) and 73% (in September) for the RCP8.5. However, the predictions of the two scenarios for the winter show differences. While it is expected that the February monthly precipitation will increase 22% in the RCP4.5 scenario, the February precipitation will decrease 7% in the RCP8.5 scenario. The increase in precipitation in the winters is also predicted by other climate models for the Northwest of the Iberian Peninsula (Santos et al. 2002; Xunta de Galicia 2009; Stigter et al. 2012).

Effects of climate change on water resources

The biases of the meteorological data in the control period were not removed and the climate change impacts were evaluated by comparing the results of the hydrological model with meteorological data from the climate model in the control and in the prediction periods.

The simulated daily P and T series for the prediction period were used to calculate the daily values of P and T at the centroids of the sub-basins of the hydrological water balance model by using a procedure similar to that used for the extrapolation of the measured meteorological data in the historical period. The daily precipitation at the centroid

Fig. 5 Mean monthly precipitation and temperature for the control period (1975–2005) and the prediction period (2069–2099) for the RCP4.5 and RCP8.5 scenarios

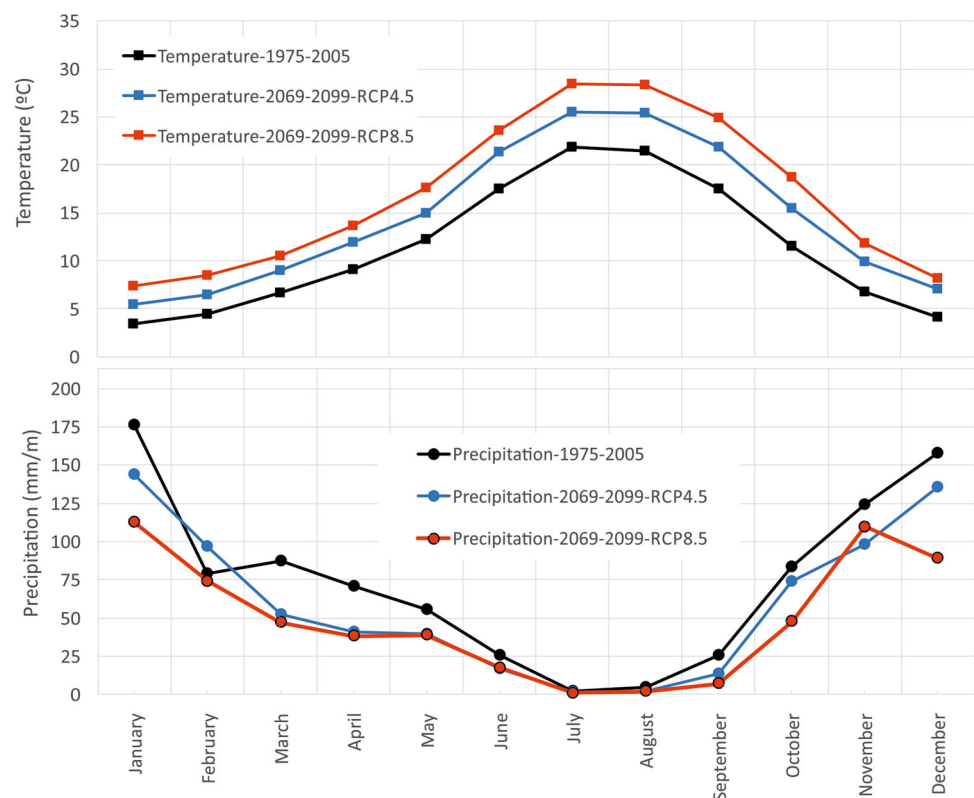


Table 3 Mean annual values of the temperature and the components of the water balance in the Serra da Estrela basin computed with historical data (1980–2015) and with the results of the climate model for the control (1975–2005) and the prediction periods (2069–2099)

	Model with observed meteorological data Historical 1980–2015	Model with meteorological data calculated with the RCM		
		Control 1975–2005	RCP4.5 2069–2099	RCP8.5 2069–2099
Temperature (°C)	9.5	9.5	12.6	14.9
Precipitation (mm/year)	2268	2265	2083	1933
Snow precipitation (mm/year)	361	396	181	65
Interception (mm/year)	390	390	358	332
Potential recharge (mm/year)	1520	1504	1320	1180
Direct potential recharge (mm/year)	23	28	37	49
Potential evapotranspiration (mm/year)	612	613	722	826
Actual evapotranspiration (mm/year)	340	324	326	326
Surface runoff (mm/year)	28	59	84	97
Interflow (mm/year)	1271	1258	1101	984
Recharge/groundwater flow (mm/year)	249	247	219	197
Streamflow (mm/year)	1549	1565	1404	1278

of a sub-basin in the i th day of the m th month, P_C^i , for the prediction period was calculated from:

$$P_C^i = P_{\text{cell}}^i \frac{P_{\text{cell}}^m + b_P^m(z_C - z_{\text{cell}}^P)}{P_{\text{cell}}^m}, \quad (15)$$

where P_{cell}^i is the daily simulated precipitation at the cell of the RCM in which the study area is located, P_{cell}^m is the monthly simulated precipitation and z_{cell}^P is the virtual altitude of the cell.

The daily temperature at the centroid of a sub-basin in the i th day of the m th month, T_C^i , is calculated from the simulated daily temperature at the cell of the RCM, T_{cell}^i , by using the temperature/altitude regression equation (Eq. 12):

$$T_C^i = T_{\text{cell}}^i + b_T^m(z_C - z_{\text{cell}}^T), \quad (16)$$

where z_{cell}^T is the virtual altitude of the cell.

The values of z_{cell}^P and z_{cell}^T used to interpolate P and T data were calibrated by trial and error so that the mean annual values of P and T in the period 1975–2005 obtained from the simulated RCM data were similar to those computed with the hydrological water balance model for the 1980–2015 period with measured T and P data from the Manteigas station. The mean annual precipitation computed with the RCM simulated data is similar to the mean annual precipitation in the basin computed with the hydrological water balance model for $z_{\text{cell}}^P = 60$ m a.s.l. On the other hand, the mean annual temperature computed with the RCM simulated data is similar to the mean annual temperature in the basin computed with the hydrological water balance model for

$z_{\text{cell}}^T = 1130$ m a.s.l. This disparity in the values of z_{cell}^P and z_{cell}^T is due to the inherent uncertainties of the P and T data for the historical period. The water balance components computed with the RCM simulated P and T data with the calibrated values of z_{cell}^P and z_{cell}^T are similar to those calculated with the water balance with the observed meteorological data (see Table 3).

The impacts of climatic change expected for the last 30 years of the twenty-first century have been calculated. The changes of the water balance components were calculated with respect to the normals of the control period (1975–2005) by using the daily meteorological data simulated by the RCM HadGEM2-ES as inputs to the hydrological model, for both the control (1975–2005) and the prediction periods (2069–2099). The daily meteorological time series were only modified to take into account the vertical gradients of precipitation and temperature which were calculated and validated for the observed data in the study zone (Espinha Marques et al. 2006, 2011).

Table 3 shows the mean annual values of the temperature and the water balance components in the ZBUM calculated for the historical (1980–2015), control (1975–2005) and prediction (2069–2099) periods.

The results of the water balance model with the observed meteorological data (1980–2015) and with the RCM meteorological time series (1975–2005) are very similar. The most important difference is the mean annual surface runoff, which is equal to 28 mm/year for the model with observed data and is 59 mm/year for the model with data from the RCM (it represents 1.3 and 2.6% of the total streamflow, respectively).

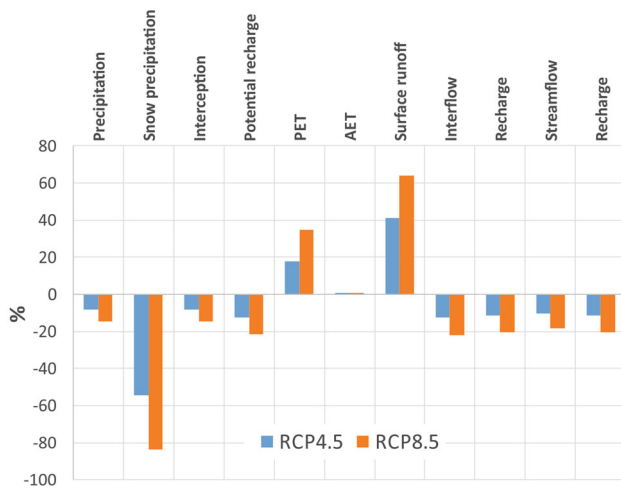


Fig. 6 Predicted changes in the mean annual water balance computed for the RCP4.5 and RCP8.5 scenarios

The temperature in the ZBUM is expected to increase significantly by the end of the twenty-first century. The mean annual temperature will increase 3.1 °C for the RCP4.5 scenario and 5.4 °C for the RCP8.5 scenario. The increase will be largest in the summer and will range from 2.0 °C in January to 4.3 °C in September for the RCP4.5 scenario and from 3.8 °C in March to 7.4 °C in September for the RCP8.5 scenario. The increase of temperature in the winter will have an important impact on snow precipitation in the highest sub-basins of the ZBUM.

Table 3 and Fig. 6 show the predicted changes in the mean annual values of the water balance components in the ZBUM for the last 30 years of the twenty-first century. Mean annual precipitation will decrease 8% in the RCP4.5 scenario and 15% in the RCP8.5 scenario. These results are coherent with those presented by López-Moreno et al. (2008), Nogués-Bravo et al. (2008) and Mora Alonso-Muñoyerro et al. (2016) for Mediterranean mountain basins. Snow precipitation will decrease from 54% (RCP4.5) to 84% (RCP8.5) and thereby the streamflow will decrease significantly during the snow melting season. However, the tendency toward earlier maximum snowmelt flows may cause unexpected runoff peaks (see López-Moreno et al. 2008; Mora Alonso-Muñoyerro et al. 2016).

The mean annual potential evapotranspiration will increase from 18 to 35% due to the temperature increase. The actual evapotranspiration, however, will remain practically unchanged due to the decrease of the soil water content.

The surface runoff will increase from 41 to 64%. The predictions for surface runoff contain large uncertainties because surface runoff is a relatively small component of the total streamflow (from 1.3 to 2.6%).

The hydrological model results show that the mean annual interflow and aquifer recharge at the end of the

twenty-first century will decrease from 12 to 22%, while the total streamflow will decrease from 10 to 18%.

Figure 7 shows the predicted relative changes in the mean monthly results of the water balance for the RCP4.5 and the RCP8.5 scenarios. Precipitation will decrease throughout the year. The maximum relative decrease of the monthly precipitation will be in July (32% for RCP4.5 and 69% for RCP8.5). The largest relative decrease of seasonal precipitation will be in spring (21% for RCP4.5 and 23% for RCP8.5). The predictions of the winter precipitation have larger uncertainties than summer precipitation. Whereas it is expected that the precipitation in February will increase 22% for the RCP4.5 scenario, a decrease of 7% is expected for the RCP8.5 scenario. Santos et al. (2002), Xunta de Galicia (2009) and Stigter et al. (2012) also predict that the winter precipitation may increase in the Northwest of the Iberian Peninsula.

Snow precipitation is expected to decrease significantly throughout the year. Whereas the largest relative decreases are expected to occur in May and October (up to 100%), the most important changes may occur in the period from November to April for which the cumulative decrease may range from 203 mm (RCP4.5) to 320 mm (RCP8.5). The streamflow during the snow melting season (in April) may decrease from 37 to 45% due to the decrease of rainfall and snow precipitation.

The monthly surface runoff may change significantly due to the climate change (Fig. 7). However, the surface runoff is a small component of the total streamflow and therefore its predictions have large uncertainties.

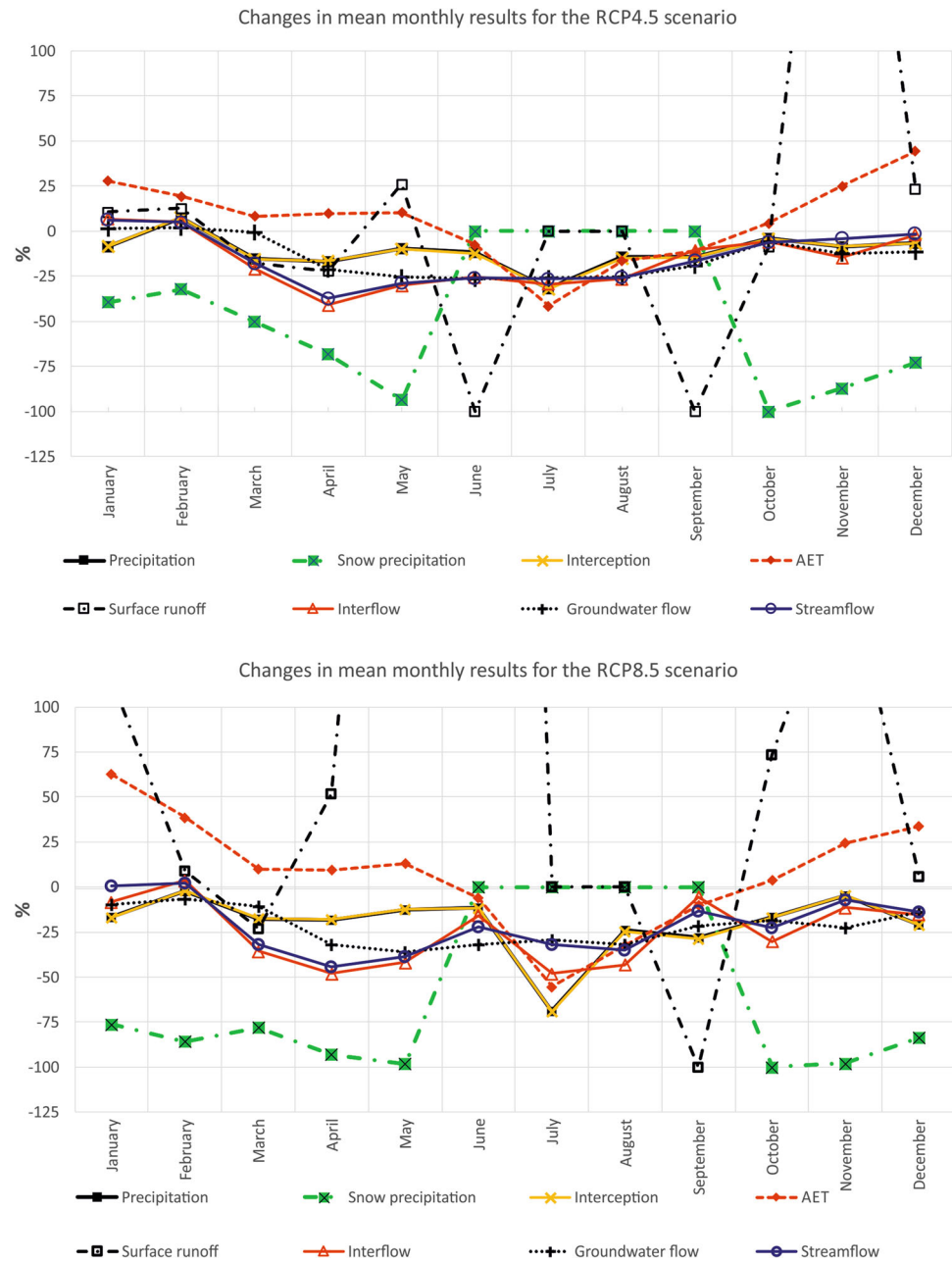
Most of the annual interflow and recharge (~96%) is evenly distributed from October to May. However, the largest decreases of the monthly interflow and recharge (in absolute terms) will be mainly from March to May. In this 3-month period, the aquifer recharge is expected to decrease 24 mm for the RCP4.5 scenario (annual decrease = 28 mm) and 33 mm for the RCP8.5 scenario (annual decrease = 50 mm).

Most of the annual streamflow occurs from November to April (80% of 1565 mm/year). The most pronounced decrease in streamflow will take place from March to May and will be mainly driven by the decrease in the precipitation and snowmelt. López-Moreno et al. (2014) present similar results, although they were computed for the mid-twenty-first century.

Conclusions

The impact of climate change on the groundwater resources of the River Zêzere Drainage Basin Upstream of Manteigas, a mountain basin located in Serra da Estrela (Central Portugal), has been evaluated. The simulated daily

Fig. 7 Predicted relative changes in the mean monthly values of the water balance components for the RCP4.5 (top) and the RCP8.5 (bottom) climate scenarios



temperature and precipitation values from the climatic models have been used directly in the hydrological modelling to preserve the day-to-day variability of the climatic variables. The changes in the water resources have been evaluated in the last 30 years of the twenty-first century for the climate scenarios RCP4.5 and RCP8.5 of the most recent European climatic simulations of the EURO-CORDEX project. The changes have been quantified in terms of the differences in the monthly and annual outputs of the water balance for the control and prediction periods.

The results of this study show that the mean annual and monthly temperatures will increase significantly by the end of the twenty-first century due to the climate change. The

mean annual temperature will increase 3.1 and 5.4 °C for the RCP4.5 and RCP8.5 scenarios, respectively. The increase will be largest in the summer. The increase of temperature in the winter will reduce snow precipitation and affect the time evolution of the melting of the snow cover in the highest sub-basins of the Serra da Estrela basin.

Precipitation will decrease throughout the year, but the relative decrease will be largest in the spring (21 and 23% for the RCP4.5 and RCP8.5 scenarios, respectively). The mean annual precipitation will decrease 8% according to the RCP4.5 scenario and 15% for the RCP8.5 scenario. The mean annual snow precipitation will decrease drastically

(54 and 84% for the RCP4.5 and RCP8.5 scenarios, respectively). The decrease will be largest from November to April. In this 7-month period, the cumulative decrease may range from 203 mm (RCP4.5) to 320 mm (RCP8.5).

The mean annual interflow and aquifer recharge in the last 30 years of the twenty-first century will decrease 12 and 22% for the RCP4.5 and RCP8.5 scenarios, respectively. The mean annual streamflow will decrease (10 and 18%) less than interflow and recharge because the surface runoff will increase slightly.

The largest decrease in mean monthly streamflows will occur from March to May due to the decrease of rainfall and snow precipitation. The mean monthly streamflow during the snow melting season in April could decrease from 37 to 45%.

It has been shown that, even by considering the uncertainties of future climatic scenarios, there is a high probability that the water resources in the Serra da Estrela mountain basin will be very vulnerable to the predicted changes in precipitation and temperature.

The results of our evaluation of the effects of the climate change in the water resources of the Serra da Estrela Basin will help to reduce the gap between Science and Policy, as highlighted by Beniston et al. (2012) and Beniston and Stoffel (2014).

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