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ORIGINAL ARTICLE

Evaluation of water resources in a high-mountain basin in *Serra da Estrela*, Central Portugal, using a semi-distributed hydrological model

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Abstract High-mountain basins provide a source of valuable water resources. This paper presents hydrological models for the evaluation of water resources in the high-mountain Zêzere river basin in *Serra da Estrela*, Central Portugal. Models are solved with VISUAL BALAN v2.0, a code which performs daily water balances in the root zone, the unsaturated zone and the aquifer and requires a small number of parameters. A lumped hydrological model fails to fit measured stream flows. Its limitations are overcome by considering the dependence of the temperature and precipitation data with elevation and the spatial variability in hydrogeomorphological variables with nine sub-basins of uniform parameters. Model parameters are calibrated by fitting stream flow measurements in the Zêzere river. Computed stream flows are highly sensitive to soil thickness, whereas computed groundwater recharge is most sensitive to the interflow and percolation recession coefficients. Interflow is the main component of total runoff, ranging from 41 to 55% of annual precipitation. High interflows are favored by the steep relief of the basin, by the presence of a high permeability soil overlying the fractured low permeability granitic bedrock and by the extensive subhorizontal fracturing at shallow depths. Mean annual groundwater recharge ranges from 11 to 15% of annual precipitation. It has a significant uncertainty due to uncertainties in soil parameters. This methodology proves to be useful to handle the research difficulties regarding a complex mountain basin in a context of data scarcity.

Keywords High-mountain hydrology · Groundwater recharge · Water balance · Semi-distributed hydrological model · *Serra da Estrela*

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Introduction

Mountain areas are the source of many large river systems. They usually represent some of the blackest "black boxes" in the hydrological cycle. The UNESCO International Hydrological Programme envisions high mountains as country's water towers and considers the hydrology of mountain areas as one of the most essential water-related research issues (Aureli 2002). Although mountainous areas occupy a significant portion of the Earth land surface, water resources and groundwater systems in high-mountain regions are generally poorly understood (Forster and Smith 1988).

The seasonality and spatial variability of local groundwaters and the complex role of soil, geomorphology, geology, climate, land use and human activities in mountain areas are difficult to model even when relevant data are available (Chalise 1997; Marques et al. 2001, 2003; Gurtz et al. 2003; Ofterdinger et al. 2004; Carvalho et al. 2005).

The water table in crystalline and other lower permeability regions is often relatively close to the land surface, even below high ridges (Tiedeman et al. 1998). High-relief and high water table elevations suggest that significant gravity-driven regional flow could be present in mountainous terrain. However, regional flow may be limited by other factors such as the incision of mountain valleys and very low permeability at depth within the mountain mass (Gleeson and Manning 2008).

The appropriate estimation of groundwater recharge requires the knowledge of the geologic, geomorphologic and climatic conditions (Castany 1975; Custodio and Llamas 1996; Fitts 2002). Quantifying interflow is one of the main challenges of hydrological models of high-mountain environments. Interflow is commonly the main runoff component in high-mountain basins (Gurtz et al. 2003; Wu and Xu 2005) because the steep relief shortens the water path in the unsaturated zone. The presence of high permeability soil overlying fractured low permeability bedrock also contributes to significant interflow (Fetter 2001). Extensive subhorizontal discontinuities are common in rock units exposed close to the topographic surface. These fractures tend to be more abundant and opened within the first meters of the top layer, therefore increasing the horizontal permeability in this part of the vadose zone. Understanding the hydrologic characteristics of subhorizontal discontinuity systems is critical in vadose zone studies in crystalline-rock settings (Hsieh 1998; Shapiro 2001). Yet, the current knowledge of interflow in mountain hydrogeological systems is still insufficient.

The mountainous region of *Serra da Estrela* in Central Portugal has unique geological, climatic and geomorphological characteristics. In spite of the high quality and economic value of its water resources, the hydrologic data from previous studies are scarce. This paper illustrates how, under this scenario, the use of appropriate hydrological models provides a way to understand and quantify subsurface and deep groundwater flow and recharge in high-mountain basins.

Lumped and semi-distributed hydrological models for the evaluation of water resources in a high-mountain basin in *Serra da Estrela* are presented. Models are solved with VISUAL BALAN v2.0, a code which performs daily water balances in the root zone, the unsaturated zone and the aquifer. This code has shown to be particularly suited for a complex high-mountain hydrological system because it requires very few input data and yet provides sufficiently good results. The main geological and hydrological features of the study area are described first. Then, hydrological models are presented. Model construction and calibration are described. Finally, model results and their relevance for high-mountain hydrology are discussed.

Hydrogeology of *Serra da Estrela*

The River Zêzere Drainage Basin Upstream of Manteigas (ZBUM) is located in the *Serra da Estrela* region (40°19′N, 7°37′W), Central Portugal (Fig. 1). *Serra da Estrela* Mountain is located at the Central-Iberian Zone of the Iberian Massif (Ribeiro et al. 1990, 2007). The study area is mainly composed of Variscan granitoids and Precambrian–Cambrian metasedimentary rocks as well as alluvial and quaternary glacial deposits (Figs. 1, 2). The main regional deep crustal structure is the Bragança–Vilariça–Manteigas fault zone (BVMFZ), with a general trend of NNE–SSW (Cabral 1989). This megastructure is part of a late-Variscan fault system that was reactivated by the alpine compressive tectonics and originated the uplift of the mountain as a horst in a pop-up structure (Ribeiro et al. 1990). Geotectonic and geomorphological features play a major role on hydrological processes and features, such as infiltration, interflow and groundwater recharge, groundwater flow paths, hydrogeochemistry, and determine the type of rock porosity (porous vs. fractured).

Serra da Estrela is part of the Iberian Central Cordillera, an ENE–WSW mountain range that crosses the Iberian Peninsula. This region shows distinctive climatic and geomorphological characteristics which play an important role on local surface and groundwater resources. The Zêzere river drainage basin upstream of Manteigas has a surface area of 28 km^2. Its elevation ranges from 875 m a.s.l. at the stream flow gauge station in Manteigas to 1,993 m a.s.l. at the summit (see Fig. 3).

The relief of the study region consists mainly of the following two major plateaus which are separated by the NNE–SSW valley of the Zêzere river (Figs. 2, 3): (a) the western Torre–Penhas Douradas plateau (1,450–1,993 m a.s.l.) and (b) the eastern Alto da Pedrice-Curral do Vento plateau (1,450–1,760 m a.s.l.). Late Pleistocene glacial landforms and deposits are common at the upper Zêzere catchment (Daveau et al. 1997; Vieira 2004). The thermal water occurrences are associated with the main fault zone (e.g., Carvalho 2006; Espinha Marques et al. 2006a).

Serra da Estrela has a Mediterranean climate with dry and warm summers (Daveau et al. 1997; Vieira et al. 2003). The wet season starts in October and lasts until May. The mean annual precipitation reaches 2,500 mm in the highest areas. Precipitation is mainly controlled by the slope orientation and the altitude (Mora 2006). Total

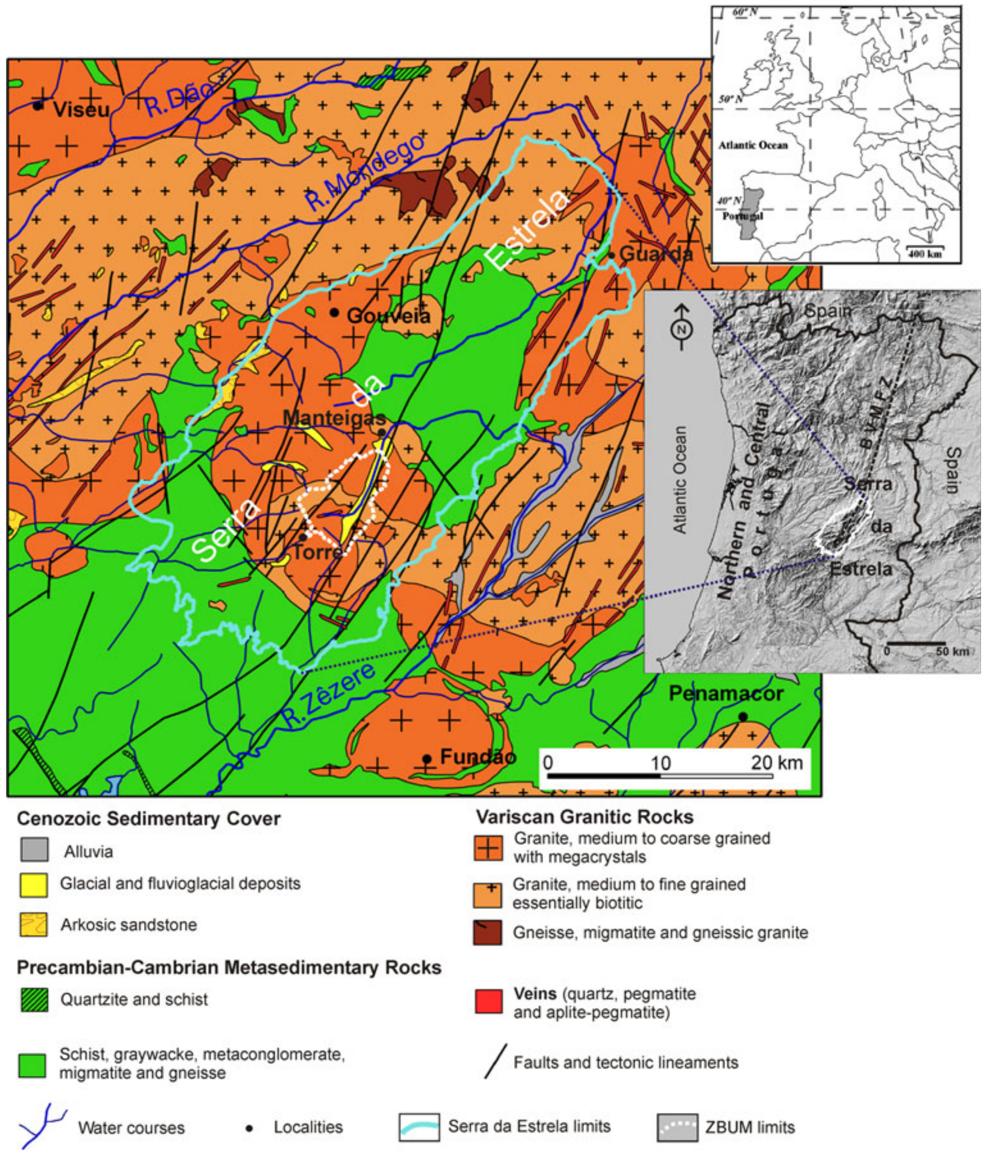

Fig. 1 Geological map of *Serra da Estrela* region (adapted from Oliveira et al. 1992). *ZBUM* River Zêzere Drainage Basin Upstream of Manteigas, *BVMFZ* Bragança–Vilariça–Manteigas fault zone

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 elevation. 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. Mean annual air temperatures are below 7°C in most of the plateau area and as low as 4°C in the vicinity of 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. These data can be adequately characterized by linear regression of P and T against elevation, z. Monthly regression lines vary throughout the year (see Tables 1, 2).

The spatial distribution of infiltration and aquifer recharge is controlled by the type of porosity of the subsurface (porous or fractured). Porous media dominate in shallow alluvium and quaternary glacial deposits as well as in the most weathered granites and metasedimentary rocks. On the other hand, fractured media prevail in poorly weathered rocks. Such media may be present very close to the surface especially on granitic outcrop-dominated areas, with thin or absent sedimentary cover or below the referred porous geologic materials. The main regional hydrogeological units correspond to the main geological units (Table 3). They include: (a) sedimentary cover of alluvium and quaternary glacial deposits; (b) metasedimentary rocks such as schists and graywackes; and (c) granitic rocks.

Fig. 2 Pictures of the Zêzere river drainage basin upstream Manteigas: (**a**) Nave de Santo António alluvia (foreground), fluvioglacial deposits (intermediate plan) and Cântaros slopes (in the background); (**b**) Cântaro Magro; (**c**) *Nardus stricta* grassland, common juniper shrubland and granitic outcrops near Torre; (**d**) Zêzere glacial valley; (**e**) Zêzere valley bottom; (**f**) Leptosol; and (**g**) snow covered slopes

The appropriate estimation of infiltration and groundwater recharge requires accounting for the spatial variability of relief, geology and climate (Castany 1975; Custodio and Llamas 1996; Fitts 2002). The ZBUM region is characterized by a strong variability of the factors which control infiltration, interflow and groundwater recharge.

The study area has distinctive geological, geomorphological and climate features which control the water availability. Water resources are of high quality and economic value and include groundwater (normal and thermomineral) and surface waters.

The generic ZBUM weathering profile, from top to bottom, includes: (1) a sandy soil cover with large content of organic matter; (2) a sandy–clayey material derived from prolonged in situ decomposition of the granitic bedrock, which is absent above 1,600 m a.s.l. due to the Late Pleistocene glacial erosion (Daveau 1971); (3) a small-scale fissuring layer, generally characterized by dense subhorizontal fissures in the first few meters and a depth-decreasing density of subhorizontal and subvertical discontinuities; and (4) an unweathered basement which is locally permeable only where tectonic fractures are present.

The hydraulic conductivity varies significantly across the profile. According to Espinha Marques et al. (2007), the field-saturated hydraulic conductivity of the soil cover is around 1.9×10^{-5} m/s, while a recent yet unpublished study points out to estimates of 1.4×10^{-5} m/s for the highly weathered granite. Estimates of hydraulic properties of layers 3 and 4 are not available for the study area. A reference transmissivity value of 1.7 m^2/day is presented by Carvalho et al. (2005) for granitic rocks of North and Central Portugal.

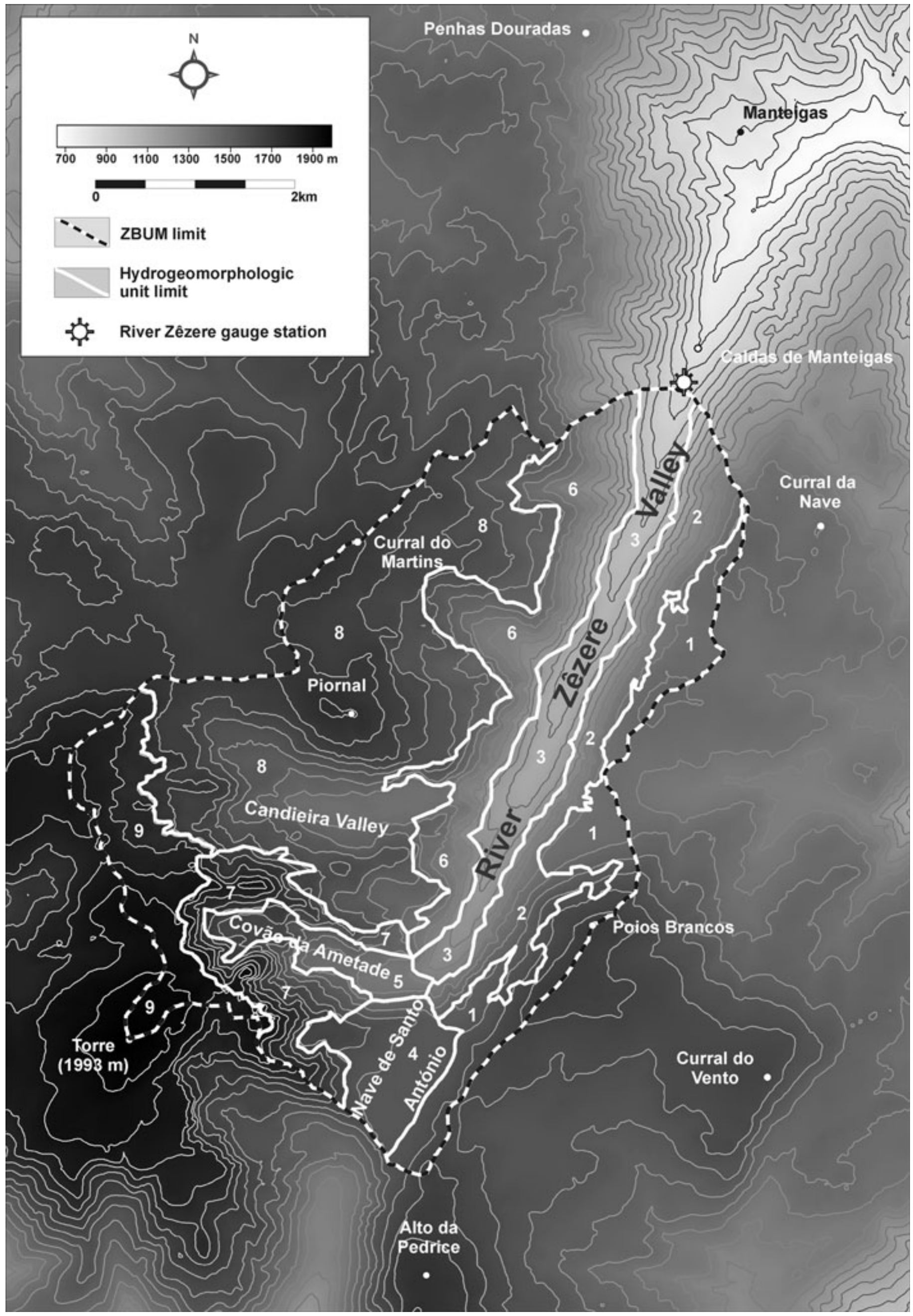

Fig. 3 Hypsometric features of the Zêzere river drainage basin upstream Manteigas. Hydrogeomorphologic units are also shown: *1* Eastern plateau; *2* Zêzere valley Eastern slopes; *3* Lower Zêzere valley floor; *4* Nave de Santo António col; *5* Upper Zêzere valley floor; *6* Zêzere valley Western slopes;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

The spatial variability of the weathering profile originates four different types of vadose zone structures (Espinha Marques et al. 2007). The first type is composed of a single and poorly weathered granitic layer with very thin or absent soil cover. It is present in granitic outcrop areas (Fig. 2) of plateau and slope hydrogeomorphologic units (Fig. 3). Water circulates in fractures. It is related to the soil hydrologic group D (see Table 4). The second type has a soil layer typically less than 0.5-m thick overlying a continuous and hard granitic layer. It is present in plateaus especially above 1,600 m a.s.l. and in slopes. Water may flow through pores and fractures. It corresponds to lithic and umbric leptosols (Fig. 2) related to soil hydrologic group D. The third type of vadose zone structure is composed of a soil layer frequently between 0.5- and 1.0-m thick overlying an intensely weathered granite layer and/or a slope deposit. It is present in lower altitude areas of slopes and plateaus where chemical weathering processes are more active, or along tectonized zones. Water may flow through pores and fractures. It corresponds to leptic umbrisols with a C horizon composed of weathered granite and/or slope deposits. These soils are included in hydrologic group C. The fourth type is composed of a soil layer frequently over 1-m thick overlying a glacial deposit. It is

Table 1 Coefficients a and b of the regression lines, $y = ax + b$, used for mean monthly precipitation, y (mm), as a function of altitude, x (m)

Month	a	b	R^2
January	0.142	77.9	0.63
February	0.102	94.84	0.60
March	0.098	67.78	0.71
April	0.075	49.65	0.76
May	0.077	36.34	0.76
June	0.460	22.04	0.75
July	0.009	7.97	0.75
August	0.012	8.64	0.83
September	0.036	23.87	0.92
October	0.11	41.98	0.79
November	0.13	70.97	0.72
December	0.12	86.81	0.67
Annual	0.99	542.22	0.73

Table 2 Coefficients a and b of the regression lines, $y = ax + b$, used for mean monthly temperature (°C) as a function of altitude, x (m)

Month	a	b	R^2
January	-5×10^{-3}	9.6	0.99
February	-5×10^{-3}	10.9	0.99
March	-5×10^{-3}	12.8	0.99
April	-6×10^{-3}	14.8	0.98
May	-6×10^{-3}	17.91	0.98
June	-6×10^{-3}	21.77	0.98
July	-5×10^{-3}	24.15	0.90
August	-5×10^{-3}	24.46	0.92
September	-5×10^{-3}	21.87	0.96
October	-5×10^{-3}	17.51	0.98
November	-5×10^{-3}	12.57	0.99
December	-4×10^{-3}	9.51	0.98
Annual	-5.6×10^{-3}	16.4	0.98

present in the base of slopes as well as in col and valley floor hydrogeomorphological units. Water flows through pores. Skeletic and humic umbrisols (A, B or C hydrologic groups) and subdominant umbric fluvisols (C or D hydrologic groups) are the prevailing soils.

The presence of a layer of low vertical hydraulic conductivity, of subhorizontal fracturing or steep relief are factors that, alone or combined, create conditions favorable to interflow in the first three types of vadose zone structure, i.e., in most of the studied area.

The hydrogeologic conceptual model of the ZBUM area (e.g., Espinha Marques et al. 2006a) was delineated in accordance with the results of a multidisciplinary study comprising geology, geomorphology, climatology, hydrogeology, hydrogeochemistry, isotope hydrology, hydropedology and geophysics. This model considers three main types of aquifers: (i) shallow unconfined aquifers, hydraulically connected to the vadose zone; (ii) shallow semi-confined aquifers; and (iii) a deep thermomineral aquifer. Waters from aquifer types 'i' and 'ii' have TDS ≈ 40 mg/L, pH ≈ 6 and temperature ≈ 10°C, whereas thermomineral waters have TDS ≈ 160 mg/L, pH ≈ 9.5 and temperature ≈ 42°C. The recharge of shallow aquifers takes place mostly in the plateaus; an additional part of the recharge may occur in the slopes of the Zêzere valley and its tributaries (Figs. 2, 3). The discharge areas are located in the Zêzere and Candeeira valley bottoms and in the Nave de Santo António col (Figs. 2, 3). As pointed out by the morphostructural and isotopic data (e.g., Espinha Marques et al. 2006a; Marques et al. 2008), discharge areas of shallow aquifers may act simultaneously as recharge areas of the deep thermomineral aquifer since they correspond to the most permeable zones of the granitic massif which are associated to deep tectonic structures. The recharge of the thermomineral aquifer is not well known though possibly is not large because of the low permeability of the granites.

Hydrological model

The hydrological model of ZBUM basin was solved with VISUAL BALAN, a lumped hydrological code especially designed to estimate water resources (Samper et al. 1999). VISUAL BALAN solves the water balance equations in the root zone of 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). Balance components are computed on a daily basis in a sequential manner (Fig. 4). The balance equation in the soil is given by

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

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.

Soil thickness is one of the most important model parameters. For a given value of porosity, usually derived from field tests, the usable water content is the product of soil thickness times the difference between field capacity and wilting point. Therefore, one can choose to calibrate the soil thickness while the rest of the parameters of the soil are fixed at values consistent with field test data.

Table 3 Main hydrogeological features from *Serra da Estrela* region

Regional hydrogeological units	Hydrogeological units	Hydrogeological features: Connectivity to the drainage network: With	Without	Possible	Type of flow: Porous media	Fractured media	Weathering: Low thickness[a]	High thickness[b]	Clayey	Sandy	More suitable exploitation structures: Dug-wells, galleries and springs	Boreholes
Sedimentary cover	Glacial deposits	●			●		n.a.	n.a.	n.a.	n.a.	●	
	Alluvium deposits	●			●		n.a.	n.a.	n.a.	n.a.	●	
Metasedimentary rocks	Schists, graywackes			●	●	●	●		●			●
Granitic rocks	Granitoids			●	●	●	●	●		●	●	●

● symbol represents the choice of one or more from the available options for each hydrogeological feature

n.a. not applicable

[a] Less than 5 m

[b] More than 5 m

Interception can be calculated using either Horton or Singh methods (Samper et al. 1999). Calculations of snow accumulation, runoff and snowmelt contribution to overland flow are based on the conceptual model of the SWMM code (Huber and Dickinson 1992). Infiltration can be calculated using the curve number method of the Soil Conservation Service (USDA 1986) or with a Horton-type equation according to which infiltration capacity, CIM, is a quadratic function of water content. This function is defined in terms of the infiltration capacities at wilting point, CIM0, and field capacity, CIM1. Overland flow is calculated as the excess of precipitation over infiltration capacity. Water infiltrated in the soil may (1) return to the atmosphere as evapotranspiration, ET; (2) flow downwards into the underlying unsaturated zone as potential recharge; (3) contribute to the soil water storage. Potential evapotranspiration, PET, can be provided directly by the user or calculated with one of the following methods: Thornthwaite, Blanney-Criddle, Makkink, Penman, Turc and Hargreaves (Samper et al. 1999). Actual ET is calculated from PET with the Penman–Grindley method.

VISUAL BALAN accounts for preferential flow, R_d, through fissures, cracks or macro-pores of the soil by assuming that (1) R_d is a fraction of the water available at the ground surface for infiltration, CKRD ($0 < \text{CKRD} < 1$); (2) preferential flow occurs when the water deficit is above a threshold, FRD (see Samper et al. 1999). In addition, the potential recharge contains a Darcian drainage term which depends on soil field capacity and hydraulic conductivity. The total drainage of the soil is the main input to the underlying unsaturated zone where water may flow horizontally as interflow or percolate vertically as groundwater recharge. Percolation is computed from Darcy's law by assuming the formation of perched aquifers during groundwater recharge episodes.

Groundwater recharge is the main input to the aquifer. Water balance in the aquifer in VISUAL BALAN can be calculated with a lumped model (single cell) or with a 1D transient distributed model consisting of N interconnected cells. Fluxes across cells are computed 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 a basin is computed as the sum of overland flow, interflow and groundwater discharge. A detailed account of the methods and parameters of BALAN and VISUAL BALAN can be found in Samper et al. (1999) and Castañeda and García-Vera (2008).

VISUAL BALAN requires a small number of parameters and incorporates friendly interfaces for data input and output processing. Model parameters are calibrated by

Table 4 Features of hydrogeomorphological units

Hydrogeomorphological units	Mean altitude (m)	Percent of the basin area	Dominant lithology	Dominant soil profile	Soil units[1]	Soil hydrologic groups[2]	Vegetation[3]
1. Eastern plateau	1,514	8.85	Granite	AC, AR	i. Umbrisol, ii. Rock outcrops and Leptosol	C, D	a, b, c, d
2. Zêzere valley Eastern slopes	1,337	10.38	Granite	AC, AR	i. Umbrisol, ii. Rock outcrops and Leptosol	C, D	a, b, c, d
3. Lower Zêzere valley floor	1,113	9.44	Fluvioglacial deposit	AC	i. Umbrisol, ii. Umbrisol and Fluvisol	A, B, C, D	e, c, a, d
4. Nave de Santo António col	1,557	4.36	Fluvioglacial deposit	AC	i. Umbrisol, ii. Fluvisol	A, B, C, D	f, d, c
5. Upper Zêzere valley floor	1,511	2.90	Granite/Fluvioglacial deposit	AR, AC	i. Rock outcrops, ii. Leptosol, Fluvisol, iii. Umbrisol	D, C	f, c, a, b, d, e, f, g
6. Zêzere valley Western slopes	1,354	14.73	Granite	AR, AC	i. Rock outcrops, iia. Leptosol (higher zone), iib. Umbrisol (lower zone)	D, B, A, C	c, d, e, g, a
7. Cântaros slopes	1,704	8.57	Granite	AR	i. Rock outcrops, iii. Leptosol	D	d
8. Lower Western plateau	1,596	33.76	Granite	AR	i. Rock outcrops, ii. Leptosol, iii. Umbrisol	D, C	d, c, f, h
9. Upper Western plateau	1,857	7.01	Granite	AR	i. Rock outcrops, ii. Leptosol, iii. Umbrisol.	D	f, h

(1) Dominant (i), subdominant (ii), possible (iii)

(2) USSCS (1964), Langan and Lammers (1991), and Boulding (1993)

(3) Maritime pine woodland (a), *Quercus pyrenaica* forest (b), *Genista florida* and *Cytisus* sp.pl. scrubland (c), heathland (d), meso-hygrophilous grassland (e), *Nardus stricta* grassland (f), meso-xerophilous grassland (g), common juniper shrubland (h)

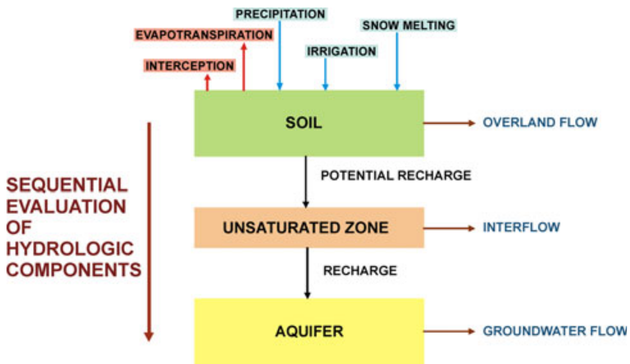

Fig. 4 Hydrological conceptual model used in VISUAL BALAN (adapted after Samper et al. 1999)

comparing computed stream flows and piezometric heads to measured data. The code has the capability of automatically estimating model parameters by minimizing a least-squares objective function with a Powell's multidimensional method (Samper et al. 1999). VISUAL BALAN and its previous version, BALAN, have been used to assess surface and groundwater resources (San Román et al. 2005), to estimate groundwater recharge in different hydrogeological environments (Samper et al. 1991, 1999, 2007; Samper and Carrera 1995; Samper and García-Vera 1997, 2004; Heredia and Murillo 2002; Aliaga et al. 2004; García-Vera and Arqued 2004; Espinha Marques et al. 2006b), to study the hydrology of playa lakes (Castañeda and García-Vera 2008) and to evaluate groundwater seepage into tunnels (Pisani 2008).

VISUAL BALAN is based on a lumped model which may not provide appropriate results in complex or large basins. Such limitation can be overcome by defining sub-basins within the basin to account for the spatial variability in meteorological and hydrological data. In this way, a semi-distributed model was built. The interactions between different sub-basins are disregarded. Therefore, water balances in each sub-basin are computed independently from those of neighbor sub-basins. The total stream flow of the basin is the sum of the total runoff of all the contributing sub-basins. This type of approach is similar to that used by other codes such as SWAT (Neitsch et al. 2002).

Hydrological model of *Serra da Estrela*

Lumped model

Mean daily air temperature and precipitation data for the basin were derived from the meteorological station of Penhas Douradas (Fig. 3) for the hydrological years from 1986–1987 to 1994–1995. This station was selected because its altitude (1,380 m) is closer to the mean altitude of ZBUM (1,505 m) than that of the alternative station of Manteigas (815 m). Daily stream flow measurements for the same period were available at the river Zêzere gauge station (Fig. 3), which operated until 1996.

The hydrology of the River Zêzere Drainage Basin Upstream of Manteigas has been modeled first with a lumped model based on available field lithology, structural geology, geomorphology, hydrogeochemistry, soil, land cover and climatology features and published data from other nearby mountain areas. This model revealed to be unrealistic because (1) it fails to account for the high spatial variability of basin features and (2) the mean yearly precipitation recorded in Penhas Douradas for the studied period, 1,406 mm, is much smaller than the normal precipitation for the period 1951–1980, which is 1,799 mm (INMG 1991), and also smaller than the measured average yearly flow at the river Zêzere gauge station, 1,601 mm. On the other hand, the mean yearly precipitation in the Manteigas station for the studied period is 1,570 mm, which is closer to the normal value, 1,668 mm. The poor quality of the lumped model has been clearly confirmed by its calculated mean annual stream flow of 998 mm which is much smaller than the value recorded at the Zêzere gauge station (see Fig. 5).

Semi-distributed model

Several sub-basins were defined to overcome the limitations of the lumped hydrological model of the ZBUM area. These units were defined in three stages. First, units in the ZBUM basin were defined based on landforms: plateaus, slopes, valley bottoms and cols. Then, these units were refined to account for the boundaries of granitic rocks and fluvioglacial deposits. Some units were subdivided based on their elevation into upper and lower units. Valley slopes were subdivided according to their position into Eastern and Western units. Finally, the following nine hydrogeomorphologic units were defined (Fig. 3): (1) Eastern plateau; (2) Zêzere valley Eastern slopes; (3) Lower Zêzere valley floor; (4) Nave de Santo António col; (5) Upper Zêzere valley floor; (6) Zêzere valley Western slopes; (7) Cântaros slopes; (8) Lower Western plateau; and (9) Upper Western plateau. The main features of hydrogeomorphological units were derived from Agroconsultores and Geometral (2004), Espinha Marques et al. (2006b) and Espinha Marques (2007) and are listed in Table 4. They include mean altitude, lithology, type of soil and dominant vegetation.

The model was also improved by adopting the Manteigas station as the reference station for which meteorological data were ascribed to each sub-basin. A virtual meteorological station was considered at the centroid of each sub-basin to which meteorological data were assigned using the monthly temperature and precipitation regression lines of Tables 1 and 2. The resulting mean annual

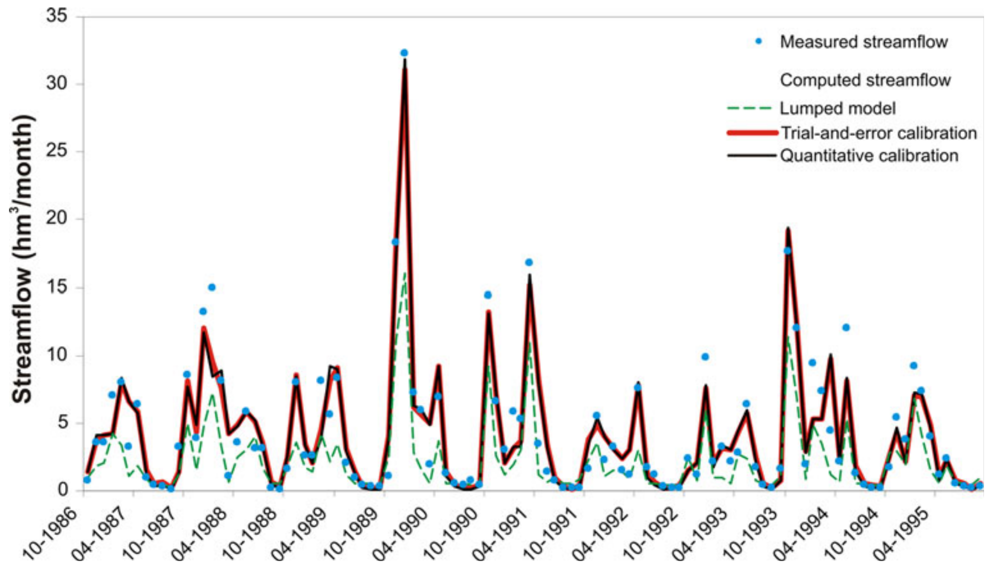

Fig. 5 Measured and computed monthly stream flows at the Zêzere river gauge station with the trial-and-error and the quantitative calibration methods

precipitation for the entire basin is equal to 2,336 mm which is similar to the value reported in previous studies by Daveau et al. (1977, 1997).

The model was calibrated by trial-and-error fit of measured mean monthly stream flows. The use of daily values for calibration has proven to be inadequate because these consist of instant measurements performed once a day at the same hour.

The concentration time of the basin is 55 min. Instantaneous stream flow measurements may underestimate or overestimate mean daily stream flows, especially in days of heavy precipitation events. This problem was overcome by calibrating model parameters by trial-and-error to fit measured monthly stream flows while ensuring a global coherence of the mean yearly 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). Calibrated model parameters are listed in Table 5. The fit between measured and computed flows achieved in this calibration is very good (see Fig. 5). Mean annual results of the water balance obtained by trial-and-error fitting are listed in Table 6.

Calibration revealed that the model is especially sensitive to (1) the usable soil storage (difference between field capacity and permanent wilting point); (2) the soil thickness and (3) the interflow recession coefficient. The Horton method for calculating overland flow and infiltration provides better results than those obtained with the curve number method.

Quantitatively calibrated semi-distributed model

To achieve a more systematic and objective fit to measured stream flow data, a second calibration stage was performed which consisted on minimizing the following least-squares objective function, O_1:

$$O_1 = \frac{1}{M}\sum_{i=1}^{M}\frac{\left(F_c^i - F_m^i\right)^2}{\left(F_m^i\right)^2} \quad (2)$$

where F_c^i and F_m^i are computed and measured stream flows at month i, respectively, and M is the number of monthly stream flow data. The procedure to minimize the objective function consisted in changing one parameter at a time, in an interval of ±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 the objective function obtained with the trial-and-error calibration is 0.495 while its final value was reduced to 0.387. Parameter values obtained in this second calibration are listed in Table 5. 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 differences 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^{-1}. Calibration was performed also using the following logarithmic least-squares objective function, O_2:

$$O_2 = \frac{1}{M}\sum_{i=1}^{M}\left[\log\left(\frac{F_c^i}{F_m^i}\right)\right]^2 \quad (3)$$

The value of function O_2 decreased from 10.07 in the first calibration to 8.88 in the second stage. Both

Table 5 Model parameters

		Sub-basin 1	Sub-basin 2	Sub-basin 3	Sub-basin 4	Sub-basin 5	Sub-basin 6	Sub-basin 7	Sub-basin 8	Sub-basin 9
Snow precipitation	Base temperature for snow precipitation (ºC)	-	-	-	2	2	-	2	2	2
	Base melting temperature (ºC)	-	-	-	2	2	-	2	2	2
Interception (Horton's Law)	Plant type	Grassland	Pine woodland	Grassland	Grassland	Grassland	Grassland	Grassland	Grassland	Grassland
	Plant height (m)	0.3	-	0.3	0.2	0.3	0.4	0.2	0.4	0.2
	Storage capacity (mm)	0.17	0.25	0.17	0.17	0.17	0.25	0.17	0.25	0.17
	Interception coefficient	0.49	0.25	0.49	0.49	0.49	0.49	0.49	0.49	0.49
Soil	Available water content (mm)	50	30	100	100	50	50	50	50	40
	Hydraulic conductivity (m/s)	1.9×10^{-5}								
Potential recharge Preferential flow	CKRD[a]	TAE: CKRD = 0.5 // QC: CKRD = 0.32								
	FRD[a]	TAE: FRD = 0.5 // QC: FRD = 0.4								
Potential recharge Darcian flow		Conventional[a]								
Overland flow (Horton's Law)	Maximum infiltration capacity (at permanent wilting point) (mm/day)	TAE: CIM0 = 41 // QC: CIM0 = 85								
	Minimum infiltration capacity (at field capacity) (mm/day)	TAE: CIM1 = 41 // QC: CIM1 = 85								
PET		Thornthwaite's equation								
AET (modified Penman--Grindley's)	CRPG[a]	40	6	6	40	40	45	40	45	40
	CEPG[a]	1								
Unsaturated zone	Interflow discharge coefficient (day^{-1})	TAE: $\alpha_h = 0.15$ // QC: $\alpha_h = 0.18$								
	Percolation recession coefficient (day^{-1})	TAE: $\alpha_p = 0.06$ // QC: $\alpha_p = 0.038$								
Aquifer	Discharge coefficient (day^{-1})	TAE: $\alpha_{aq} = 0.03$ // QC: $\alpha_{aq} = 0.029$								

TAE trial-and-error calibration stage, *QC* quantitative calibration

[a] See Samper et al. (1999) and Castañeda and García-Vera (2008)

Table 6 Mean annual results of the water balance obtained by the trial-and-error and the quantitative calibration methods

	Trial-and-error calibration	Quantitative calibration	Fraction of mean annual precipitation (%)		
			Initial	Final	Change
Precipitation	2,336	2,336	100	100	0
Snow precipitation	373	373	16	16	0
Snow melting	334	334	14	14	0
Interception	402	402	17	17	0
PET	605	605	26	26	0
AET	325	325	14	14	0
Overland flow	310	69	13	3	−10
Potential recharge	1,308	1,541	56	66	10
Interflow	962	1,292	41	55	14
Groundwater recharge	349	252	15	11	−4
Total stream flow	1,621	1,613	69	69	0
Objective function, O_1	0.495	0.387	–	–	–
Objective function, O_2	10.07	8.88	–	–	–

All values in mm except for the objective function (see text)

objective functions in (2) and (3) lead to similar calibration results.

The mean annual values of the components of the water balance in the entire basin for the initial and final calibration stages are listed in Table 6. The results show that stream flow data can be fitted almost equally well (see Fig. 5) with two sets of parameters (see Table 5) which lead to two different water balance results (Table 6). These differences are largest for interflow and amount to 14% of P_{mean}. Differences for overland flow and groundwater recharge are equal to 10 and 4%, respectively. Results of both calibrations provide a measure of the uncertainties of runoff components. The uncertainty in mean annual groundwater recharge is large because it may range from 250 to 350 mm/year.

The largest difference between both calibrations occurs for interflow which increases from 962 to 1,292 mm/year. The increase of 330 mm/year amounts to 14% of the mean annual precipitation in the basin P_{mean}. The increase in interflow is caused by the combination of a decrease in overland flow from 310 to 69 mm/year due to the decrease in the infiltration capacity coefficients (see Table 5) and a decrease in groundwater recharge from 349 to 252 mm/year caused by the reduction of the percolation recession coefficient from 0.06 to 0.038 day^{-1}.

Sensitivity analysis

Model uncertainties caused by parameter uncertainties are evaluated by sensitivity analyses. Estimated model parameters were varied ±20% from the values obtained with the quantitatively based calibration (base run). Several values of the parameter increments were tested. It was concluded that a 20% change was appropriate for most of the parameters.

The objective function O_1 defined in Eq. 2 was used to calculate the following normalized sensitivity, S_O:

$$S_O = \frac{(O_{sens} - O_{base})}{O_{base}} \Big/ \frac{(P_{sens} - P_{base})}{P_{base}} \quad (4)$$

where O_{sens} and O_{base} are the values of the objective function for the sensitivity and base runs, respectively, and P_{sens} and P_{base} are parameter values of the sensitivity and base runs, respectively. Relative sensitivities are dimensionless and can be easily compared for different parameters which may vary over different scales. Relative sensitivities are listed in Table 7. Computed stream flows are extremely sensitive to soil thickness. Their relative sensitivities are on the order of 5, meaning that a p% change in the soil thickness leads to a 5p% change in the objective function. On the other hand, the objective function is much less sensitive to the rest of the parameters which show relative sensitivities smaller than 0.1, except for a reduction of the interflow recession coefficient and an increase in the groundwater recession coefficient which have sensitivities on the order of 0.2. Model results are not sensitive to changes of an order of magnitude in the vertical permeability of the soil with respect to the base value of 1.9×10^{-5} m/s derived from the mean of available field data (Espinha Marques 2007; Espinha Marques et al. 2007). The vertical permeability of the soil does not limit the potential recharge even for its smallest value of 1.9×10^{-6} m/s.

The relative sensitivity of groundwater recharge to changes in model parameters was evaluated by changing one-at-a-time each parameter ±20% from its base value. Relative sensitivities, S_R, were computed from:

Table 7 Relative sensitivity of the objective function, as defined in Eq. 4, to changes in model parameters

	$100(P_{sens} - P_{base})/P_{base}$	
	−20	+20
Soil thickness	5.543	4.302
Coefficient CKRD	0.039	0.052
Coefficient FRD	0.013	0.013
Maximum infiltration	0.026	0.026
Interflow recession coefficient, α_h	0.013	0.207
Percolation recession coefficient, α_p	0.052	0.013
Groundwater recession coefficient, α_{aq}	0.245	0.013

Table 8 Relative sensitivity of the groundwater recharge, as defined in Eq. 5, to changes in model parameters

	$100(P_{sens} - P_{base})/P_{base}$	
	−20	+20
Soil thickness	2.4×10^{-3}	-6.3×10^{-2}
Coefficient CKRD	5.6×10^{-3}	-5.2×10^{-2}
Coefficient FRD	5.0×10^{-3}	3.6×10^{-3}
Maximum infiltration	9.6×10^{-2}	9.3×10^{-2}
Interflow recession coefficient, α_h	−1.1	-8.2×10^{-1}
Percolation recession coefficient, α_p	7.8×10^{-1}	7.4×10^{-1}
Groundwater recession coefficient, α_{aq}	0	0

$$S_R = \frac{(R_{sens} - R_{base})}{R_{base}} \bigg/ \frac{(P_{sens} - P_{base})}{P_{base}} \qquad (5)$$

where R_{sens} and R_{base} are the recharge values of the sensitivity and base runs, respectively. Relative sensitivities are listed in Table 8. Similar to the objective function, groundwater recharge lacks sensitivity to the vertical permeability of the soil. Groundwater recharge is most sensitive to the interflow recession coefficient. The relative sensitivity is negative because recharge decreases when interflow increases. The second largest sensitivity of recharge corresponds to the percolation recession coefficient which shows relative sensitivities of 0.7. It is remarkable that the relative sensitivity of recharge to the soil thickness is so low.

Discussion of results

Mean annual results of water balance components listed in Table 6, obtained with VISUAL BALAN, are within the range of expected values for a high-mountain area in crystalline rocks. Most of the hydrological components have mean yearly values which agree with those obtained in previous studies in other high-mountain aquifer systems of the Central-Iberian range (e.g., Mendes and Bettencourt 1980; Carvalho 2006) and other parts of the world (e.g., Gurtz et al. 2003; Wu and Xu 2005).

Mean annual PET (605 mm/year) is close to the value calculated by Mendes and Bettencourt (1980) for the meteorological station of Penhas da Saúde (558 mm/year), located near the SE limit of ZBUM (1,510 m a.s.l.). However, the mean annual AET computed with VISUAL BALAN (325 mm/year) is significantly smaller than that reported by Mendes and Bettencourt (1980), 479 mm/year. The main reason for such difference is that these authors selected an unrealistic value of available water content in the soil (100 mm). In fact, soil data from the ZBUM area suggest that this parameter has a strong spatial variability. Available water content in our model is equal to 100 mm in sub-basins 3 and 4 while in the rest it varies from 30 to 50 mm (see Table 5).

Interflow is the most important component of the total runoff. It ranges from 41% of total precipitation in the initial calibration to 55% in the final calibration stage. These results are consistent with some field features observed in the ZBUM area such as the steep relief, the presence of highly permeable soil overlying fractured and much less permeable granitic bedrock and the wide presence of a layer of granitic rock showing intense subhorizontal discontinuities and/or low-angle fissure network zones (Fig. 6). The high-mountain hydrological conditions of ZBUM tend to favor interflow.

Our estimate of annual groundwater recharge is 10.8% of the mean annual precipitation. This value is consistent with those obtained in the last decades in Northern and Central Portugal with other techniques which range from 2% (Henriques 1985) to over 30% of the mean annual precipitation (Lima and Silva 2000; Oliveira 2006).

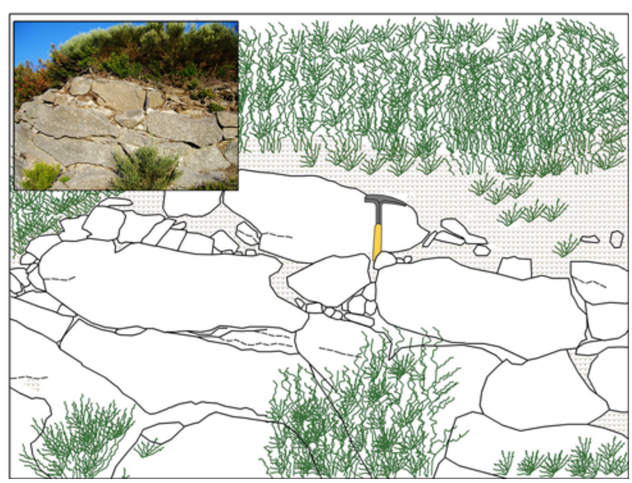

Fig. 6 Schematic illustration of the conditions that favor interflow at the Zêzere river basin: highly permeable soil overlying low permeability bedrock with subhorizontal fractures

Carvalho et al. (2000) reported recharge values in hard rock mountain basins from 6 to 18%; Carvalho and Chaminé (2001) and Carvalho (2006) reported a recharge of 10% in the Corgas Largas (Gouveia) aquifer system located in the Western part of the *Serra da Estrela* massif.

Conclusions

Water resources in the high-mountain Zêzere river basin, in *Serra da Estrela*, Central Portugal, have been evaluated with a semi-distributed hydrological model. The model has been solved with VISUAL BALAN v2.0, a code which performs daily water balances in the root zone, the unsaturated zone and the aquifer and requires a small number of parameters. Model results and the fit to measured stream flows improve when spatial variations in hydrogeomorphological variables and changes in rainfall and air temperature with elevation are taken into account by defining nine sub-basins. Daily temperature and precipitation data from Manteigas meteorological station have been extrapolated to each sub-basin using known vertical gradients of temperature and precipitation. Model parameters have been calibrated first by trial-and-error fitting stream flow measurements in the Zêzere river. The final calibration has been achieved by minimizing a least-squares objective function.

Uncertainties in model parameters and groundwater recharge have been evaluated by sensitivity analyses. Computed stream flows are extremely sensitive to soil thickness. On the other hand, they are much less sensitive to the rest of the parameters. Model results are not sensitive to changes in the vertical permeability of the soil. Computed groundwater recharge is most sensitive to the interflow and percolation recession coefficients.

Interflow is the main component of runoff. It ranges from 41 to 55% of total precipitation. Interflow in the ZBUM high-mountain area is large due to the combination of a steep relief, a high permeability soil overlying the granitic bedrock and an extensive subhorizontal fracturing.

Estimates of mean annual groundwater recharge have a significant uncertainty due to uncertainties in soil parameters and range from 250 to 350 mm/year (i.e., from 11 to 15% of annual precipitation, respectively). These estimates are coherent with values previously reported for similar basins in the Central-Iberian Zone of the Iberian Massif. Given the paramount importance of recharge rate evaluation in this area, it is recommended to perform additional studies focusing on the mechanisms and parameters of overland flow, infiltration and interflow.

This study shows that VISUAL BALAN is a useful tool for modeling the hydrology of high-mountain basins even in a situation of data shortage. The code was successful in quantifying overland flow, interflow, aquifer recharge and total stream flow.

This type of knowledge is extremely valuable for the authorities in what concerns the water resources management at a regional scale and the land management for nature protection purposes. Private water companies may also benefit since this type of information may help the planning of groundwater exploration campaigns, with clear technical and economic advantages.

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