The impact of droughts and water management on various hydrological systems in the headwaters of the Tagus River (central Spain)

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Abstract. The influence of climate variation on the availability of water resources was analyzed in the headwaters of the Tagus River basin using two drought indices, the standardized precipitation index (SPI) and the standardized precipitation evapotranspiration index (SPEI). This basin is highly regulated and strategic, and contains two hyperannual reservoirs that are the origin of the water supply system for Mediterranean areas of southeast Spain. The indices confirmed that drought conditions have prevailed in the headwaters of the Tagus River since the 1970s. The responses in river discharge and reservoir storage were slightly higher when based on the SPEI rather than the SPI, which indicates that although precipitation had a major role in explaining temporal variability in the analyzed parameters, the influence of temperature was not negligible. Moreover, the greatest response in hydrological variables was evident over longer timescales of the climatic drought indices. Although the effect of climate variability on water resources was substantial during the analyzed period, we also showed a major change in hydrological-climatic relationships in regulated systems including reservoir storage and outflow. These were closely related to changes in external demand following commencement of the water transfer system to the Júcar and Segura basins after the 1980s. The marked reduction in water availability in the basin, which is related to more frequent droughts, contrasts with the amount of water transferred, which shows a clear upward trend associated with increasing water demand in the Mediterranean basin.

34 **Keywords:** drought, global warming, standardized precipitation index, standardized precipitation
35 evapotranspiration index, water transfer, reservoir management, Tagus River, Spain.

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39 **1. Introduction**

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41 An effect of global change on environmental conditions in the western Mediterranean region is
42 increasing uncertainty in water resource availability (García-Ruiz et al., 2009). The projections of
43 climate models are for a general decrease in precipitation and increasing temperatures in the region
44 (Giorgi and Lionelo, 2008), which may markedly reduce river flows (*e.g.*, Kilsby et al., 2007).
45 Nevertheless, there are various sources of uncertainty in climate change simulations (Räisänen,
46 2007), and difficulties are associated with establishment of direct relationships between climate
47 variability and water resources, as a consequence of the substantial influence of land cover (Llorens
48 et al., 1995; Beguería et al., 2003; García-Ruiz et al., 2008) and water management strategies
49 (López-Moreno et al., 2007) on the response of river flows to climate variability. Analysis of the
50 temporal evolution of water resources in the Mediterranean area is complex, given the large natural
51 climate variability of the region (Vicente-Serrano and Cuadrat, 2007; López-Bustins et al., 2008; De
52 Luís et al., 2009), land cover changes in headwaters (Vicente-Serrano et al., 2004; Lasanta et al.,
53 2005; MacDonald et al., 2000; Sluiter and De Jong, 2007), and the intense water regulation
54 necessary to meet high urban and agricultural demand (Batalla et al., 2004; López-Moreno et al.,
55 2009). To understand the possible consequences of climate change processes on the future
56 availability of water resources in the region, it is necessary to determine the current relationship
57 between climate variability and water resources, taking account of different hydrological
58 subsystems (*e.g.*, river discharge, reservoir storage) and the major role of dam and canal operations
59 in the regulation of river flows.

60 In the western Mediterranean region the availability of water resources is critical during certain
61 periods. River flows show strong seasonality characterized by low natural flow in summer;

62 management of reservoirs is thus focused on meeting urban and irrigation demands during this
63 season (Maneux et al., 2001; Snoussi et al., 2002). The high frequency of droughts in the area
64 makes it necessary to improve management strategies during dry periods. To this end it is important
65 to determine the empirical relationship between climatic and hydrological droughts, which will
66 enable accurate assessment of the possible impact of environmental change processes.

67 Isolating the influence of climate is difficult because the response of hydrological systems to
68 precipitation can vary markedly as a function of time (Changnon and Easterling, 1989; Elfatih et al.,
69 1999; Pandey and Ramasastri, 2001), as a result of temporal differences in the frequencies of
70 hydrological and climatic variables (Skøien et al., 2003). In a study in the central Spanish Pyrenees,
71 Vicente-Serrano and López-Moreno (2005) showed very different timescale responses to
72 precipitation accumulation between river discharge (short timescale) and reservoir storage (long
73 timescale). A similar pattern was found by Szalai and colleagues (2000) in Hungary. In Greece,
74 Vasiliades and Loukas (2009) reported different response times for soil moisture and river
75 discharge to two Palmer drought indices, with variations in soil moisture occurring at higher
76 frequency than river discharge. This was more marked for groundwater level, which responds to
77 precipitation only following long-term accumulation (Khan et al., 2008). Thus, the effect of climatic
78 droughts on groundwater shows distinct temporal inertia (Peters et al., 2005).

79 Reservoir regulation and water transfer disrupt climate–hydrology relationships, sometimes
80 dramatically, making it difficult to determine the role of precipitation variability on availability of
81 water resources. In addition, several studies have shown that recent temperature increases (Jones
82 and Moberg, 2003) are having a negative effect on the availability of water resources, as a
83 consequence of water losses caused by evapotranspiration (Nicholls, 2004; Cai and Cowan, 2008;
84 Gerten et al., 2008). To date, no studies have analyzed this issue in the western Mediterranean
85 region.

86 Few empirical analyses have related hydrological records to climate indices, or tested the usefulness
87 of various drought indices for monitoring water resources in different systems. The main objective
88 of this study was to determine the relationship between two different multi-scalar drought indices
89 (the standardized precipitation index, SPI; and the standardized precipitation evapotranspiration
90 index, SPEI) and three hydrological variables (river discharge, reservoir storage, and reservoir
91 release) in the headwaters of the Tagus River. This is a highly regulated and strategic basin that has
92 two hyperannual reservoirs (*i.e.*, where the storage capacity exceeds the annual discharge from a
93 regulated river) that are the origin of a water transfer system supplying the Mediterranean areas of
94 southeast Spain. Additional objectives of the study were i) to identify the best drought index and
95 timescale for monitoring water resources in the different subsystems, and, ii) to assess the influence
96 of water management and warming processes on temporal changes in climate-hydrological
97 relationships.

98

99 **2. Study area**

100

101 Water resources are intensively regulated in the *Alto Tajo* region, which comprises the headwaters
102 of the Tagus River, between the Iberian Range and the Plateau of Castille. The river basin is in a
103 mountainous area ranging in altitude from 600 m at the Bolarque reservoir to 1,935 m at the eastern
104 extremity of the basin, and has a surface area of 7,417 km². The relief is dominated by moorlands,
105 the main structural unit, which descend gradually over a limestone base from the foot of the
106 Orihuela and Albarracín ranges. Below 1,000 m a marl substrate is evident, and the valley opens up
107 and provides a suitable area for the retention of water.

108 The spatial distribution of precipitation follows a topographic pattern. Mean annual precipitation
109 decreases from 1,000 mm in the Albarracín Range (east) to about 500 mm in low-altitude areas.
110 Precipitation is strongly seasonal, with a peak in May and another in November; the area is also
111 subject to the low summer rainfall typical of the Mediterranean region. Temperatures are

112 characterized by extremes and oscillations given the continental nature of the climate. The
113 maximum daily oscillation, 29.4°C, was recorded in 1981. The mean annual temperature is less than
114 10°C for most of the basin, but during winter the average temperature drops to 1–3°C, making it one
115 of the coldest regions on the Iberian Peninsula. In the hottest month (July) the average temperature
116 is 21.9 °C.

117 The area has a dense hydrological network divided into two main components, the Tagus and
118 Guadiela rivers, whose waters are collected and regulated downstream (Figure 1). The water is
119 controlled to supply various water uses in the region, and to transfer water to the Mediterranean
120 coastlands. The basin has two large reservoirs, the Entrepeñas and Buendía dams. The Entrepeñas
121 has a capacity of 802 hm³ and the wall height of the dam is 87 m; it was built in 1956 to collect the
122 waters of the Tagus River. The Buendía dam stores the waters of the Guadiela, Mayor, and
123 Guadamejud rivers. The dam was built in 1957, the wall height of the dam is 78 m, and has a
124 maximum storage capacity of 1,638 hm³. Both reservoirs act as a single storage unit, as they are
125 connected by a tunnel. The Bolarque Dam is located downstream of the discharge of the two main
126 dams, at the convergence of the Tagus and Guadiela rivers. The dam was built in 1910, but
127 underwent several modifications that were not finalized until 1951. The main function of the dam is
128 distribution of the water collected upstream in the Entrepeñas and Buendía dams; some is sent to the
129 Tajo-Segura water transfer and the Bolarque-Jarama irrigation supply systems, and the remainder
130 flows a few kilometers down the main stream of the Tagus River to the Zorita reservoir, which was
131 built in 1947 for hydroelectric production.

132 Two principal land cover types exist in the study area: Above 1000 m the predominant land cover
133 are the forests, composed mainly by conifers, which occupy 42.5% of the study area. Dry-farmed
134 cropland covers 43.9% of the surface, mainly in the flat areas. Other important land cover types in
135 the study area are thicket (9.9%) and grasslands (2.3%). Irrigated lands, urban areas and agro-
136 forestry occupy percentages lower than 1% of the study area.

137 The area is highly strategic in terms of water resources, as it provides water to the Tajo-Segura
138 Transfer. The diversion of water to the Mediterranean region involves political decisions at a
139 national level, and has been the focus of territorial conflicts, especially during drought periods
140 (Garrote et al., 2007).

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143 **3. Dataset and methodology**

144

145 *3.1. Climate data*

146

147 Data on monthly precipitation, and maximum and minimum temperatures between 1961 and 2006
148 were provided by the Spanish Meteorological Agency (AEMET). This consisted of 64 monthly
149 temperature series and 147 monthly precipitation series within the study area and neighboring
150 zones. The availability and quality of the series were variable; both precipitation and temperature
151 series contained numerous discontinuities, and long continuous series were very rare. Therefore, for
152 further analysis we selected only those series with < 10% gaps (9 precipitation series and 3
153 temperature series; see Fig. 1). Gaps were filled using multiple linear regressions.

154 Among the different existing techniques to detect and adjust the temporal non-homogeneities of
155 climate series, (see reviews in Peterson et al., 1998 and Beaulieu et al., 2007), we have used the
156 standard normal homogeneity test (SNHT; Alexandersson, 1986), which is widely used to
157 homogenize temperature and precipitation series in different regions (e.g. Alexandersson and
158 Moberg, 1997; Begert and Schlegel, 2005). Following Peterson and Easterling (1994), a relative
159 homogenization procedure was applied using reference series incorporating the three most
160 correlated series for precipitation and the two most correlated for temperature. The ANCLIM
161 program (Stepánek, 2004) was used to perform the homogenization process. Using the precipitation
162 and temperature series for the different stations we created a precipitation and mean temperature
163 regional series using the weighted average of the monthly records for each station (Vicente-Serrano

164 and López-Moreno, 2005). The weight of each series was determined from the surface area
165 represented by each station, calculated using Thiessen polygons (Jones and Hulme, 1996).

166
167 *3.2. Hydrological records*
168

169 Hydrological data for 1961–2006 were provided by the Tagus Water Authority (*Confederación*
170 *Hidrográfica del Tago*). Four parameters were considered in the analysis:

- 171 i) the monthly river discharges at three gauging stations (Fig. 1) that had no data gaps for
172 the 1961–2006 period;
- 173 ii) inflows to the interconnected Entrepeñas and Buendía reservoir system;
- 174 iii) the total storage registered in both reservoirs; and
- 175 iv) the net outflows to the Tagus River below the Bolarque and Zorita reservoirs, located
176 immediately downstream of the Entrepeñas and Buendía reservoir system.

177 Among the used hydrological parameters, only the river discharges and inflows to the reservoirs
178 system are natural non-human managed resources. Nevertheless, the analysis of the hydrological
179 response to climate can not be restricted to the natural non-human affected systems. In addition to
180 different human interferences, in the Tagus basin climate variability also controls reservoir storages
181 and releases downstream the dams (López-Moreno et al., 2007). Thus, assessing the connection
182 between climate variability and human inference is of great interest, as most of the water supplied
183 to urban settlements, agriculture and industry depends from highly regulated fluvial stretches.

184
185 *3.3. Hydrological drought index calculation*
186

187 The various approaches to analyzing hydrological droughts are commonly based on discharge
188 thresholds (Tallaksen et al., 1997; Fleig et al., 2006). This enables the identification of low
189 discharge periods, but does not take into account the seasonality of discharges, which usually leads
190 to naturally low summer flows being classified as low flow periods. This is a particular problem in
191 highly seasonal regimes, including the Mediterranean rivers. We quantified hydrological drought

192 conditions by relating monthly discharge anomalies to average conditions, following Dracup et al.
193 (1980). For this purpose we followed the approach commonly used for climatic drought index
194 calculations: namely the corresponding number of standard deviations relative to average values.
195 This approach has been used by Zaidman et al. (2001), Shukla and Wood (2008) and López-
196 Moreno et al., (2009), among others.

197 As hydrological records are highly biased and do not commonly follow a normal distribution, it was
198 necessary to standardize the probability distribution of the hydrological records. A comparison was
199 made among several skewed probability distributions, based on an L moment ratio diagram
200 (Greenwood et al., 1979; Hosking, 1990). The resulting plot enabled comparison of empirical
201 parameters obtained from inflows, reservoir storages and releases, with parametric theoretical
202 distributions. The L-moments ratio diagram (Fig. 2) shows the empirical values of L-skewness and
203 L-kurtosis for monthly series of river flows, inflows to the Entrepeñas–Buendía system, reservoir
204 storages, and outflows to the Tagus River, along with theoretical curves for some parametric
205 distributions. López-Moreno et al. (2009) showed the effectiveness of the Pearson III distribution
206 for obtaining a hydrological drought index in the lower part of the Tagus basin. Zaidmann et al.
207 (2001) used a lognormal distribution in north and central Europe. For the United States of America,
208 Shukla and Wood (2008) showed a generally good performance with the 2-parameters gamma and
209 lognormal distributions, but also showed that the 3-parameters lognormal and the generalized
210 extreme value distributions were also applicable over widely varying hydroclimatic regimes. In the
211 headwaters of the Tagus River, we showed large differences among the variables analyzed. River
212 flows and inflows to the Entrepeñas and Buendía system showed high dispersion but, in general the
213 Pearson III and the lognormal distribution were suitable for modeling the data. High dispersion was
214 also found for the monthly series of outflows as a function of the monthly series. After several
215 attempts we selected the Pearson III distribution to obtain an outflow drought index. The reservoir
216 storages did not show dispersion, showing the different monthly series an adjustment to the

217 Generalized Pareto distribution, which was used to calculate a reservoir drought index. The
218 confirmation step, using the Kolmogorov-Smirnov test, allowed selection of the Generalized Pareto
219 distribution for the different monthly series of reservoir storages, and the Pearson III distribution for
220 inflows and outflows.

221 The L-moments procedure was used to obtain the parameters of the Pearson III and Generalized
222 Pareto distributions. The L-coefficients of skewness and kurtosis, τ_3 and τ_4 respectively, were
223 calculated as follows:

$$224 \quad \tau_3 = \frac{\lambda_3}{\lambda_2}$$

$$225 \quad \tau_4 = \frac{\lambda_4}{\lambda_2}$$

226 where λ_2 , λ_3 and λ_4 are the L-moments of the precipitation series. These were obtained from
227 probability-weighted moments (PWMs), using the formulae:

$$228 \quad \lambda_1 = \alpha_0$$

$$229 \quad \lambda_2 = \alpha_0 - 2\alpha_1$$

$$230 \quad \lambda_3 = \alpha_0 - 6\alpha_1 + 6\alpha_2$$

$$231 \quad \lambda_4 = \alpha_0 - 12\alpha_1 + 30\alpha_2 - 20\alpha_3$$

232 The PWMs of order s were calculated as:

$$233 \quad \alpha_s = \frac{1}{N} \sum_{i=1}^N (1 - F_i)^s x_i$$

234 where x_i is the data from a given precipitation series and F_i is the frequency estimator. F_i was
235 calculated following the approach of Hosking (1990):

$$236 \quad F_i = \frac{i - 0.35}{N}$$

237 where i is the range of observations arranged in increasing order, and N is the number of data
 238 points. According to the Pearson III distribution, the probability distribution function of x is given
 239 by:

$$240 \quad F(x) = \frac{1}{\alpha \Gamma(\beta)} \int_{\gamma}^x \left(\frac{x-\gamma}{\alpha} \right)^{\beta-1} e^{-\left(\frac{x-\gamma}{\alpha} \right)} dx$$

241 The probability distribution function of x , according to the Generalized Pareto distribution, is given
 242 by

$$243 \quad F(x) = 1 - \left[1 - \frac{\kappa}{\alpha} (x - \varepsilon) \right]^{1/\kappa}$$

244 The parameters of the Pearson III distribution were obtained following Hosking (1990):

245 If $\tau_3 \geq 1/3$ then $\tau_m = 1 - \tau_3$, and β can be obtained using the expression:

$$246 \quad \beta = \frac{(0.36067\tau_m - 0.5967\tau_m^2 + 0.25361\tau_m^3)}{(1 - 2.78861\tau_m + 2.56096\tau_m^2 - 0.77045\tau_m^3)}$$

247 If $\tau_3 < 1/3$ then $\tau_m = 3\pi\tau_3^2$, then β can be obtained using the expression:

$$248 \quad \beta = \frac{(1 + 0.2906\tau_m)}{(\tau_m + 0.1882\tau_m^2 + 0.0442\tau_m^3)}$$

$$249 \quad \alpha = \sqrt{\pi}\lambda_2 \frac{\Gamma(\beta)}{\Gamma(\beta + 1/2)}$$

$$250 \quad \gamma = \lambda_1 - \alpha\beta$$

251 The parameters of the Generalized Pareto distribution were also obtained according to Hosking
 252 (1990) using L-moments:

$$253 \quad \alpha = \lambda_1 \left(\frac{1}{\tau_2} - 1 \right)$$

$$254 \quad \kappa = \frac{1}{\tau_2} - 2$$

$$255 \quad \varepsilon = \lambda_1 - \lambda_1(2 + \kappa)$$

256 The $F(x)$ values obtained were converted to z-standardized values. For example, following the
257 classical approximation of Abramowitz and Stegun (1965):

$$258 \quad z = W - \frac{C_0 + C_1 W + C_2 W^2}{1 + d_1 W + d_2 W^2 + d_3 W^3},$$

259 where

$$260 \quad W = \sqrt{-2 \ln(P)} \text{ for } P \leq 0.5,$$

261 where P is the probability of exceeding a determined D value, $P = 1 - F(x)$. If $P > 0.5$, P is replaced
262 by $1 - P$, the sign of the resultant z-value is reversed. The constants are: $C_0 = 2.515517$, $C_1 =$
263 0.802853 , $C_2 = 0.010328$, $d_1 = 1.432788$, $d_2 = 0.189269$, $d_3 = 0.001308$. The average value of the
264 hydrological index is 0, and the standard deviation is 1.

265 Figure 3 shows the evolution of the three hydrological drought indices, for which marked
266 differences were evident. The inflows to the Entrepeñas–Buendía system had higher temporal
267 frequency than the reservoir storages. The outflows behaved very differently before and after the
268 1980s, with positive values dominating prior to this period, and very negative outflow values
269 dominating in the following two decades.

270 271 3.4. Calculation of climatic drought indices 272

273 From regional series of monthly precipitation and temperature we obtained two multi-scalar climate
274 drought indices: the standard precipitation index (SPI) (McKee et al., 1993) and the standardized
275 precipitation evapotranspiration index (SPEI) (Vicente-Serrano et al., 2009). It is commonly
276 assumed that drought is a multi-scalar phenomenon (McKee et al., 1995). This is very important for
277 drought quantification and monitoring, as the time scale over which precipitation deficits
278 accumulate functionally separates different types of drought, and enables quantification of the
279 natural lags between precipitation and other usable water sources, such as river discharges and
280 water storages (Changnon and Easterling, 1989; Elfatih et al., 1999; Pandey and Ramasastri, 2001);
281 there are with several empirical lines of evidence for this lag (Szalai et al., 2000; Sims et al., 2002;

282 Vicente-Serrano and López-Moreno, 2005; Patel et al., 2007; Khan et al., 2008). For this reason we
283 considered a diversity of time scales for both drought indices (1–48 months). The justification for
284 using two different drought indices is the fact that the SPI only accounts for precipitation effects,
285 whereas the SPEI accounts for inputs (precipitation) and outputs (evapotranspiration) to the system.
286 Although precipitation is the main variable explaining the frequency, duration and severity of
287 droughts (Chang and Cleopa, 1991; Heim, 2002), recent studies have shown that the effect of
288 temperature (or evapotranspiration) is significant (Hu and Willson, 2000), particularly under global
289 warming scenarios (Dubrovsky et al., 2008). Abramopoulos et al. (1988) showed that evaporation
290 and transpiration can consume up to 80% of rainfall, and found that the efficiency of drying due to
291 temperature anomalies is as high as that due to rainfall shortage. Syed et al. (2008) showed that
292 precipitation dominates terrestrial water storage variation in the tropics, but evapotranspiration
293 explains the variability at middle latitudes. In addition, studies have shown that anomalous high
294 temperatures related to warming processes have in recent years exacerbated the impact of climatic
295 droughts on water resources (Nicholls, 2004; Cai and Cowan, 2008).

296 The SPI is calculated by adjusting the precipitation series to a given probability distribution.
297 Initially, the Gamma distribution was used to calculate the SPI (McKee et al., 1993), but the
298 Pearson III distribution is more robust, given its three parameters (Vicente-Serrano, 2006). The
299 complete formulation of the SPI following the Pearson III distribution and the L-moments method
300 for calculating parameters is described in Vicente-Serrano (2006), and López-Moreno and Vicente-
301 Serrano (2008). A C++ program developed by the Spanish Scientific Research Council (CSIC)
302 (<http://digital.csic.es/handle/10261/10006>) was used for this purpose.

303 The SPEI is based on a monthly climatic water balance (precipitation minus potential
304 evapotranspiration), which is adjusted using a 3-parameter log-logistic distribution to take into
305 account common negative values. The values are accumulated to different time scales, following a
306 similar approach to that for the SPI, and converted to standard deviations with respect to average

307 values. The complete methodology is described by Vicente-Serrano et al. (2009); a C++ program,
308 also developed by the CSIC (<http://digital.csic.es/handle/10261/10002>), was used for calculations.

309 310 *3.5. Statistical analysis*

311
312 Hydrological and climatic drought indices were related using the parametric Pearson coefficient of
313 correlation to measure the degree of association between hydrological and climatic droughts. The
314 different time scales of the SPI and the SPEI were included in the analysis to determine the best
315 time scale and climatic drought index for explaining hydrological variability. Analyses were based
316 on complete continuous series, but also separately for the 12 monthly series of standardized inflows,
317 reservoir storages and outflows.

318 To determine possible changes in the relationship between climatic and hydrological droughts we
319 undertook the same analyses using moving window correlations (15 years). This enabled
320 assessment of whether climatic–hydrological drought relationships are stable or not, and possible
321 interferences in terms of the management of water resources. The non-parametric Spearman’s rho
322 correlation coefficient (Siegel and Castelan, 1988) was applied to a moving window Pearson R
323 correlation series to detect significant trends in the climate–hydrological drought relationships in
324 various months of the year.

325 A flow chart summarizing all steps of the methodology applied is shown in Fig. 4.

326 327 **4. Results**

328 329 *4.1. Evolution of climatic droughts*

330
331 Figure 5 shows evolution of the SPI and the SPEI over 3, 12, 24, and 48 month intervals from 1961
332 to 2006. Short timescales showed a high temporal frequency of dry and moist periods. With
333 increasing timescales, drought and moist periods showed a lower temporal frequency and a longer
334 duration. Two contrasting periods were evident between 1961 and 2006 for both the SPI and the
335 SPEI. Wet conditions dominated during the 1960s and 1970s, whereas persistent drought conditions

occurred from 1980 to 2006, and were particularly severe during the period 1990-2005. Both indices showed a similar time evolution, with no notable differences. For example, at the 24-month timescale both indices demonstrated four major dry periods: from 1975 to 1976, in the first half of the 1980s, most of the 1990s, and from 2005 until the end of the analysis period. However, the average duration of droughts determined by the SPEI was longer than that identified by the SPI. This occurred at all analyzed timescales; at the timescale of 3 months the average duration of dry periods was 3.8 months according to the SPI, whereas the SPEI indicated an average duration of 4.1 months; at the scale of 12 months, the average durations were 10.7 and 13.0 months for the SPI and the SPEI, respectively; at the scale of 24 months the mean duration was 17.4 months for the SPI and 17.6 months for the SPEI. The longest average duration of dry periods (21.1 months) was registered by the SPEI at the 48-month scale, whereas the SPI showed a mean maximum duration of 19.6 months. The SPEI also identified a greater severity of droughts in the 1990s and 2000s than did the SPI; this was related to the very warm temperatures during those decades. Thus, the evolution of the drought indices is related to the observed changes of precipitation and temperature. The temperature shows a positive trend (0.23 °C per decade between 1961 and 2006) and precipitation a negative trend (-71 mm per decade between 1961 and 2006). Both trends are statistically significant (Rho-Spearman test, $p < 0.05$).

4.2. Response of hydrological subsystems to climatic droughts

Figure 6 shows the Pearson R correlation coefficients between the series of indices for inflow, reservoir storage, and outflow, and the series of the SPI and the SPEI at different timescales (1–48 months). Correlations were positive for the various timescales, and the curves for the SPI and the SPEI were of similar shape. In general, climate variability explained a high proportion of variability in hydrological variables. Different responses were identified for inflow, reservoir storage, and outflow. The greatest correlations with inflow were found at timescales of 4-12 months, for both the SPI and the SPEI. Similar behavior was found in analyses at the three gauging stations upstream of

363 the Entrepeñas and Buendía reservoirs (not shown). The maximum correlation was recorded for the
364 SPEI at the 8-month timescale ($R=0.76$). It is notable that positive and high correlations were also
365 found at long timescales. Reservoir storage and outflow showed a different behavior than did
366 inflow. The greatest correlations were found for longer timescales and reservoir storage (a
367 maximum of $R=0.87$ with the 33-month SPEI), as a consequence of the hyperannual characteristics
368 of the Entrepeñas-Buendía reservoir system. Thus, climatic conditions in the previous 3-4 years
369 were the most significant variable accounting for water quantities stored in reservoirs. Correlations
370 for outflow from the system were slightly lower, with an even longer temporal inertia (a maximum
371 of $R=0.76$ for the 48-month SPEI).

372 This analysis showed that the reservoir system markedly changed the river regime, with inflow
373 determined by relatively high frequencies of climate variability, and outflows by large scale
374 frequencies, which were greatly influenced by dam operations. It is noteworthy that for both
375 reservoir storage and outflow, correlation was slightly higher with the SPEI rather than the SPI,
376 indicating that a combined effect of precipitation and evapotranspiration better explained variability
377 of water resources than did precipitation alone.

378 Figure 7 shows the evolution of hydrological and climatic drought indices, with respect to each
379 index and the timescale at which the highest correlation was found. For inflow there was a strong
380 correlation between climatic and hydrological drought indices. Between 1960 and 1980 moist
381 conditions dominated, explaining positive anomalies for inflows. In contrast, between 1980 and
382 2006 the climatic drought periods showed a significant reduction in inflow, with the exception of a
383 few positive peaks (*e.g.*, 1988, 1997, 1998), when there was agreement between positive climatic
384 and hydrologic drought indices.

385 Reservoir storage also showed temporal variation closely related to evolution of the climatic
386 drought index, with variability in climatic droughts explaining most anomalies in water resources in
387 the reservoir system. Positive anomalies in reservoir storage were recorded during the moist period

388 (1961-1980), whereas dominant negative anomalies in storage occurred after this time, coinciding
389 with negative SPEI values. Outflow from the system showed less relationship to the evolution of
390 climatic droughts. This was expected, as river management affects flows downstream of dams, but
391 in the study area after 1985 water transfer to Mediterranean basins also influenced outflow, which
392 caused a sudden drop in the influence of climatic control on releases to the Tagus River downstream
393 of the reservoir system.

394 As expected, the relationship between climatic and hydrological droughts changed markedly on a
395 monthly basis. Figure 8 shows Pearson's R correlations for the z-standardized inflow and the 1- to
396 48-month SPI and SPEI. The patterns for the SPI and SPEI were quite similar. Very high
397 correlations (>0.9) were found between hydrological and climatic droughts from January to March
398 for timescales between 3 and 5 months. In contrast, correlations during summer months were very
399 low at the shortest timescales, whereas inflow was more associated with drought indices at longer
400 timescales.

401 Figure 9 shows monthly correlations between z-standardized reservoir storage and the various
402 timescales for the SPI and the SPEI. As was expected from the hyperannual character of the
403 reservoir system, there were fewer monthly differences in this system than were found for inflow.
404 For most months the SPEI showed better correlations than did the SPI, and the strongest
405 correlations were found for timescales between 25 and 45 months. Outflow also showed stronger
406 correlations with the SPEI than with the SPI, few differences between months (Figure 10), and had
407 a similar pattern to that for reservoir storage.

408 409 *4.3. Temporal changes in the hydrological response to climatic droughts*

410 Figure 11 shows moving window correlations (15 years) for the SPI and the SPEI at various
411 timescales, and the z-standardized inflow. There were few changes during the analyzed period,
412 although after 1985 the correlations were slightly lower. Figures 12 and 13 compare SPI and SPEI

413 data with reservoir storage and output, and reveal more important changes than were found for
414 inflow. For reservoir storage, correlations before 1975 were not significant at most timescales
415 considered. After 1975 the relationship between climatic and hydrologic conditions increased
416 noticeably, and, after 1990, the correlations were slightly stronger for the SPI than the SPEI. The
417 opposite was observed for outflow, where correlations were higher with the SPEI than the SPI.
418 Nevertheless, the period of significant correlations was very short (between 1975 and 1985),
419 showing that after the water transfer system was initiated the effect of climate variability in
420 explaining release to the Tagus River practically disappeared.

421 Figure 14a shows river flow in the Tagus River basin upstream and downstream of the Zorita
422 reservoir. Two highly contrasting flow periods are evident, in 1961-1980 and 1980-2006. As a
423 consequence of reservoir management, the temporal variability of flow in the Tagus River between
424 1961 and 1980, recorded downstream of the Entrepeñas-Buendía system, was less than the natural
425 variability (Figure 14b). However, some high flows ($>200 \text{ Hm}^3$) were maintained and the frequency
426 of low flows was reduced, relative to upstream reservoirs. After 1980 there was a dramatic change
427 in the river regime. Between 1980 and 1995 high flows declined upstream of the Entrepeñas-
428 Buendía reservoirs, which could explain the reduction in high flows downstream of the reservoir
429 system. In the period 1995-2006 some high flows also occurred upstream of the reservoirs, but very
430 constant flows were recorded downstream. Thus, the total flow during the irrigation-free period
431 (November-May) was dramatically reduced in the 1995-2006 period (average 127.7 Hm^3)
432 compared with the 1985-1995 interval (average 188.1 Hm^3). Prior to initiation of the water transfer
433 system, the November-May flow had increased in the regarding 1985-1995 interval (average 500
434 Hm^3 vs. 406.8 Hm^3 , respectively). This indicates that greater inflow to the reservoirs increased the
435 flows released from the system, but there was a flow reduction to the Tagus River downstream of
436 the water transfer point, with increased flow being diverted to the water transfer system. This can be
437 explained by increased demand for water both in the Mediterranean basin and for irrigation in the

438 Bolarque-Jarama area (Figure 14c). Thus, in the last 10 years the diversion of water to these uses
439 has exceeded the flow of the Tagus River downstream of the Zorita reservoir (annual average 647.5
440 Hm^3 vs. 317 Hm^3 , respectively).

441
442

443 **5. Discussion and conclusions**

444

445 This study evaluated the impact of climatic droughts on various hydrological systems in a highly
446 regulated basin of central Spain between 1961 and 2006, using two different drought indices, the
447 SPI and the SPEI. This is one of the first studies to explore the relationship between hydrological
448 and climatic droughts using such indices.

449 We have shown that construction of hydrological drought indices based on a diversity of variables
450 is complex because the statistical characteristics of series used can vary markedly. The procedure
451 used to obtain indices differed for river discharge and reservoir storage because of differences in the
452 statistical distributions most appropriate to the data, in this case the Pearson III distribution for river
453 discharge and the Generalized Pareto distribution for reservoir storage. This highlights the need for
454 prior testing to determine the most suitable distribution for deriving hydrological indices. This
455 conclusion is in line with the recent results of Shukla and Wood (2008) in the United States of
456 America.

457 The two climate drought indices (the SPI and the SPEI) effectively identified water deficits at
458 various timescales. Independent of timescale and the drought index used, the results indicate that
459 drought conditions have increased in the headwaters of the Tagus River since the 1970s; the
460 decades of 1980, 1990, and 2000 were dominated by drought conditions. This finding is consistent
461 with most studies in the western Mediterranean (Van der Schrier et al., 2006; López-Moreno and
462 Vicente-Serrano, 2008) and the Iberian Peninsula (Esteban-Parra et al., 1998; Vicente-Serrano and
463 Cuadrat, 2007; De Luis et al., 2009), where a general decrease in precipitation occurred during the
464 second half of the 20th century. Moreover, a marked increase in temperature was also seen in the

465 region during this period (Brunet et al., 2006). These conditions have impacted on water resources,
466 increasing evapotranspiration rates, the water deficit, and climate drought severity (*e.g.*, Hu and
467 Willson, 2000; Dubrovsky et al., 2008).

468 The recent increase in temperature could explain why climatic droughts in the decades of 1990 and
469 2000 were more severe when analyzed by the SPEI rather than the SPI. We also showed that the
470 response to the hydrological drought was slightly higher according to SPEI data than SPI
471 information, which indicates that although precipitation plays a major role in explaining temporal
472 variability in analyzed variables, the influence of temperature is significant and is likely to increase
473 in the future, as has been indicated in numerous studies (Labat et al., 2004; Nichols, 2004;
474 Buthiyani et al., 2008; Polemio and Casarano, 2008).

475 We showed that the response of river discharge in the headwaters of the Tagus River was similar
476 when measured by either drought index. This was because of the greater sensitivity of river
477 discharge to climate variability at short timescales, and highlights the difficulty of isolating of
478 temperature variability and trends from variation in unregulated river discharge. In contrast, a
479 greater difference was found between the SPI and the SPEI when reservoir storage was examined;
480 this responds over longer timescales, and the cumulative role of temperature was evident in the
481 drought indices. In addition, various studies have shown that reservoirs are subject to evaporation
482 processes (Maingi and Marsh, 2002; Montasery and Adeloye, 2004). These factors could explain
483 why a combined precipitation and evapotranspiration drought index, such as the SPEI, is more
484 appropriate for analysis of the response of reservoir storage than an index, such as the SPI, that
485 considers precipitation alone. Moreover, water demand for crops is highly dependent on
486 temperature, and warmer conditions may enhance water consumption in the Mediterranean and
487 irrigation areas of the basin. As outflow is highly dependent on storage in the reservoir system, the
488 role of Potential Evapotranspiration is also propagated downstream, and, thus, stronger correlations
489 are found with the SPEI than the SPI.

490 It is noteworthy that the greatest responses to hydrological variables seen in the climatic drought
491 indices were noted over longer timescales in the headwaters of the Tagus basin than in other
492 regions. Vicente-Serrano and López-Moreno (2005) showed that in the central Spanish Pyrenees the
493 greatest responses were in the 2-month SPI for river discharge and the 8-month SPI for reservoir
494 storage. In Hungary, Szalai et al. (2000) also found greater responses for river discharge and
495 reservoir storage over shorter timescales than we found in the Tagus basin. The response of river
496 discharge to long timescales (24-48 months) may be attributable to the dominant limestone
497 lithology. This may favor an indirect relationship between precipitation and discharge, with
498 recharge of aquifers during precipitation periods, but release occurring slowly over long intervals.
499 The drought indices for the analyzed reservoir storage were not affected by short timescales, but a
500 strong response over longer timescales (3-4 years) was found. This is explained by the relatively
501 large inertia of inflow, in addition to the hyperannual character of the managed system.

502 There were no marked monthly differences in responses of drought indices to different timescales
503 for reservoir storage and outflow, because of the low response found at the shortest timescales.
504 Nevertheless, the indices showed large monthly differences in response to inflow to the system,
505 which were closely related to the climatic-lithological characteristics of the region. Precipitation in
506 the headwaters of the Tagus River is heavily influenced by southwestern flows associated with
507 negative phases of the North Atlantic Oscillation (López-Moreno et al., 2007), which is a major
508 influence in winter. This explains the strong correlation found between river discharge and the 2- to
509 5-month SPI and SPEI between December and March, as surface flow is dominant in these months.
510 Further, the main recharge of aquifers occurs during winter, which explains the annual or bi-annual
511 recharge seen and the greater influence of variability in summer flow on the SPI and the SPEI at the
512 longest timescales. The baseline aquifer levels are affected mainly in summer, and show slow
513 temporal inertia and a strong relationship to precipitation accumulated over long periods (Peters et
514 al., 1995); the drought index responses are also over longer timescales (Khan et al., 2008).

515 The relationship between climate and hydrological conditions was relatively stable between 1961
516 and 2006 with respect to unregulated inflow to the system (upstream of the Entrepeñas-Buendía
517 reservoirs), but slight decreases in correlation were found after the 1980s. Given the absence of
518 human interference upstream of the reservoir, this decrease was probably attributable to land cover
519 changes dominated by an increase in natural vegetation cover. In the mountainous areas of central
520 Iberian Peninsula several studies have illustrated the existing re-vegetation process as the dominant
521 land cover change (e.g., Gallart and Llorens, 2002). For example, Hill et al. (2008) showed, by
522 means of NOAA-AVHRR images, that in the headwaters of the Tagus basin the dominant land
523 cover process is a trend toward the natural vegetation recovering as a consequence of the rural
524 exodus. The increased evapotranspiration demand and water interception by vegetation, and reduced
525 the effect of climate in explaining flow variability, as has been observed in other areas of the Iberian
526 Peninsula (Beguería et al., 2003; García-Ruiz et al., 2008).

527 Nevertheless, the main changes in climatic-hydrological relationships occurred in the regulated
528 systems of reservoir storage and outflow. These were closely related to changes in external demand
529 with commencement of the water transfer system to the Júcar and Segura basins after the 1980s.
530 Although reservoir storage was closely related to low-frequency climate variability in recent
531 decades, the outflow to the Tagus River downstream of the reservoir system was not controlled by
532 climate after the water transfer system commenced. This situation leads to enormous uncertainty as
533 to the future availability of water resources. On the one hand, demand for water from the transfer
534 system has increased markedly in recent years, dramatically reducing water release to the Tagus
535 River. On the other hand, the increased frequency and severity of climatic droughts in the last
536 decades has reduced water reserves stored in the reservoirs. Despite the marked reduction of water
537 availability in the basin, there has been no reduction in the amount of water transferred; rather, there
538 has been a clear trend of increasing water demand in the Mediterranean basin. Therefore, despite
539 the onset of a period characterized by reduced availability of water resources because of more

frequent and severe droughts, water managers have given priority to water transfer for human use over maintenance of natural flow to the Tagus River, which has dramatically declined. This management practice has relied on the hyperannual character of the Entrepeñas-Buendía reservoir system, which reduces the effect of short-term drought. However, under the severe and sustained droughts recorded since the 1990s the reservoir system has not been able to maintain this management strategy. This was evident in 2005 and 2006, when sustained drought conditions resulted in the reservoir being unable to satisfy demand, and the managers were obligated to reduce flow to both the Tagus River and the water transfer system for the Mediterranean basin; this resulted in conflicts and political ramifications at the national level. Given the context of global climate change, characterized by predictions of greater severity and frequency of droughts in southern Europe (Blenkinsop and Fowler, 2007), and continuing revegetation processes, water availability in the basin is expected to decline. Thus, it will be increasingly difficult or impossible to satisfy external water demand using the current management strategy, and it is likely that new approaches will be needed.

Acknowledgements

This work has been supported by the research projects CGL2006-11619/HID, CGL2008-01189/BTE, and CGL2008-1083/CLI financed by the Spanish Commission of Science and Technology and FEDER, EUROGEOS (FP7-ENV-2008-1-226487) and ACQWA (FP7-ENV-2007-1- 212250) financed by the VII Framework Programme of the European Commission, STRIVER (Strategy and methodology for Improved IWRM—An integrated interdisciplinary assessment in four twinning river basins), financed by the VI Framework Programme of the European Commission. “Las sequías climáticas en la cuenca del Ebro y su respuesta hidrológica” Financed by “Obra Social La Caixa” and the Aragón Government and “Programa de grupos de investigación consolidados” financed by the Aragón Government.

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753 Figure 1. Location and topography of the study area, and the spatial distribution of weather and
754 gauging stations used in analysis.

755 Figure 2. L-moment ratio diagram, including monthly series of river discharges, inflows to the
756 Entrepeñas-Buendía system, reservoir storage, and outflow to the Tagus basin.

757 Figure 3. Evolution of hydrological drought indices for inflow, reservoir storage, and outflow to the
758 system.

759 Figure 4. Flow chart that shows all steps of the methodology applied.

760 Figure 5. Evolution of the 3-, 12-, 24-, and 48-month SPI and SPEI in the study area from 1961 to
761 2006.

762 Figure 6. Pearson R correlation values for the 1- to 48-month SPI and SPEI, and series of z-
763 standardized inflows, storages, and outflows.

764 Figure 7. Evolution of hydrological and climatic drought indices. The most highly correlated
765 timescale is shown for each hydrological variable.

766 Figure 8. Correlation coefficients between the z-standardized monthly inflow series and the
767 monthly SPI and SPEI values at various timescales. Significant correlations ($p<0.05$)
768 correspond to R values greater than 0.26.

769 Figure 9. Correlation coefficients between the z-standardized monthly reservoir storage series and
770 the monthly SPI and SPEI values at various timescales. Significant correlations ($p<0.05$)
771 correspond to R values greater than 0.26.

772 Figure 10. Correlation coefficients between the z-standardized monthly outflow series and the
773 monthly SPI and SPEI values at various timescales. Significant correlations ($p<0.05$)
774 correspond to R values greater than 0.26.

775 Figure 11. Moving-window Pearson R correlation coefficients between the z-standardized monthly
776 outflow series and the monthly SPI and SPEI values at various timescales. Significant
777 correlations ($p<0.05$) correspond to R values greater than 0.51.

778 Figure 12. Moving-window R-Pearson correlation coefficients between the z-standardized monthly
779 reservoir storage series and the monthly SPI and SPEI values at various timescales.
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781 Figure 13. Moving-window R-Pearson correlation coefficients between the z-standardized monthly
782 outflow series and the monthly SPI and SPEI values at various timescales. Significant
783 correlations ($p<0.05$) correspond to R values greater than 0.51.

784 Figure 14. a) Flows in the Tagus River upstream and downstream of the intake point for the water
785 transfer system, b) inflows to the Entrepeñas–Buendía system, and, c) monthly flows for
786 water transfer and irrigation.

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Figure 1. Location and topography of the study area, and the spatial distribution of weather and gauging stations used in analysis.

Figure 2. L-moment ratio diagram, including monthly series of river discharges, inflows to the Entrepeñas-Buendía system, reservoir storage, and outflow to the Tagus basin.

Figure 3. Evolution of hydrological drought indices for inflow, reservoir storage, and outflow to the system.

Figure 4. Flow chart that shows all steps of the methodology applied.

Figure 5. Evolution of the 3-, 12-, 24-, and 48-month SPI and SPEI in the study area from 1961 to 2006.

Figure 6. Pearson R correlation values for the 1- to 48-month SPI and SPEI, and series of z-standardized inflows, storages, and outflows.

Figure 7. Evolution of hydrological and climatic drought indices. The most highly correlated timescale is shown for each hydrological variable.

Figure 8. Correlation coefficients between the z-standardized monthly inflow series and the monthly SPI and SPEI values at various timescales. Significant correlations ($p<0.05$) correspond to R values greater than 0.26.

Figure 9. Correlation coefficients between the z-standardized monthly reservoir storage series and the monthly SPI and SPEI values at various timescales. Significant correlations ($p<0.05$) correspond to R values greater than 0.26.

Figure 10. Correlation coefficients between the z-standardized monthly outflow series and the monthly SPI and SPEI values at various timescales. Significant correlations ($p < 0.05$) correspond to R values greater than 0.26.

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Figure 14. a) Flows in the Tagus River upstream and downstream of the intake point for the water transfer system, b) inflows to the Entrepeñas–Buendía system, and, c) monthly flows for water transfer and irrigation.