Evapotranspiration enhancement drives the European water-budget deficit during multi-year droughts

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Abstract.

   In a warming climate, periods with below-than-average precipitation will increase in frequency and intensity. During such periods, known as meteorological droughts, sparse but consistent pieces of evidence show that the decline in annual runoff may be proportionally larger than the corresponding decline in precipitation (e.g., -40% vs. -20%). Reasons behind this exacerbation

5 of runoff deficit during dry periods remain largely unknown, which challenges generalization at larger scales (i.e., beyond the single catchment), as well as the predictability of when this exacerbation will occur and how intense it will be. Here, we tested the hypothesis that runoff-deficit exacerbation during droughts is a common feature of droughts across climates and is driven by evapotranspiration enhancement. We support this hypothesis by relying on multidecadal records of streamflow and precipitation for more than 200 catchments across various European climates, which distinctively show the emergence of similar periods of

10 exacerbated runoff deficit identified in previous studies, i.e., runoff deficit on the order of -20% to -40% less than what expected from precipitation deficit. The magnitude of this exacerbation is two to three times larger for basins located in dry regions than for basins in wet regions and is qualitatively correlated with an increase in annual evapotranspiration during droughts, on the order of 11% and 33% over basins characterized by energy- and water-limited evapotranspiration regimes, respectively. Thus, enhanced atmospheric and vegetation demand for moisture during dry periods induces a nonlinear and potentially hysteretic

15 precipitation-runoff relationship for low-flow regimes, which results in an unexpectedly large decrease in runoff during periods of already low water availability. Forecasting onset, magnitude, and duration of these drops in runoff availability has paramount societal implications, especially in a warming climate, given their supporting role for water, food, and energy security. The outcome that water basins are prone to this exacerbation of runoff deficit for various climates and evapotranspiration regimes, compounded by the lack of specific parametrizations of this process in the majority of hydrological and land-surface models,

20 make further understanding of its patterns of predictability an urgent priority for water-resource planning and management in a warming and drier climate.

1 Introduction

Timing and seasonality of runoff (Q) from a river basin are dictated by the interaction across incoming precipitation (P), atmospheric and vegetation water use (evapotranspiration, ET), and the variation in water stored in the basin (ΔS): $Q = P - ET - \Delta S$ (Bales et al., 2018). While changes in precipitation will ultimately affect runoff, processes driving the precipitation-runoff relationship (Saft et al., 2016b) are complicated by the nonlinear, and often delayed response of ET and ΔS (Bales et al., 2018; Avanzi et al., 2020). Depending on the direction of precipitation change, evapotranspiration-precipitation feedback mechanisms may comprise vegetation expansion and/or mortality (Senf et al., 2020; Choat et al., 2018), wildfires (Bowd et al., 2019), a shift in vegetation water-use strategies (i.e., isohydric to anisohydric prevalent species), and depletion of regolith water storage and rock moisture (McDowell et al., 2008; Hahm et al., 2020; Rungee et al., 2019; Goulden and Bales, 2019; Klos et al., 2018). The rate and distribution of regolith-storage recharge and drainage also depends on precipitation, for example in terms of enhanced soil hydrophobicity during periods with low precipitation and the related disconnection between soil and groundwater storage (Rye and Smettem, 2017). These processes are intertwined with changes in other climatic factors, such as air temperature, making water basins to profoundly co-evolve with climate (Troch et al., 2015).

The relevance of evapotranspiration and storage for catchment hydrology has been recognized for a long time (Teuling et al., 2013), but the runoff implications of their response to precipitation changes have been rarely studied and so are still poorly understood (Goulden and Bales, 2019). This knowledge gap is particularly remarkable during droughts, because it challenges forecasting of the transformation of precipitation deficit (meteorological drought) into runoff deficit (hydrologic drought) (Loon, 2015). Previous studies in this context have shown that prolonged meteorological droughts may indeed result in a larger-than-expected decrease in runoff (Saft et al., 2015; Avanzi et al., 2020; Tian et al., 2018; Mastrotheodoros et al., 2020; Tian et al., 2020; Alvarez-Garreton et al., 2021). This observation not only shows that precipitation deficit is an insufficient predictor for fully characterizing droughts (Bales et al., 2018), but also proofs that the coevolution of water basins with climate (in the form of ET and ΔS) may play a non-negligible role in driving the impact of meteorological droughts on runoff.

The ultimate cause behind this observed exacerbation of runoff deficit is still unknown, with previous work providing mixed results related either to the buffered response time of evapotranspiration to precipitation deficit (Avanzi et al., 2020) or to streamflow memory (Alvarez-Garreton et al., 2021). It is also unclear whether this exacerbation takes place only in specific climates, such as Mediterranean regions where precipitation distribution is skewed toward winter and summer is dry (Feng et al., 2019), or whether exacerbation of runoff deficit during meteorological droughts is a common feature of water basins across climates and in particular across a variety of long-term aridity indices. In this regard, previous work in non-Mediterranean regions of Europe showed the evapotranspiration amplifies the impact of summer droughts, but these studies mostly focused on storage rather than on runoff exacerbation (Teuling et al., 2013). Large-scale assessments spanning a variety of climates, hereby defined in terms of the aridity index, are still needed to gain further understanding of the runoff implications of meteorological droughts.

Further large-scale verification of the hypothesis that a basin precipitation-runoff partitioning during dry periods can significantly deviate from what was previously encountered during wet periods – and so exacerbate runoff deficit – is a key to more

robust future water management in a warming climate. This is both because mechanisms causing this shift in the water balance are not explicitly represented by the majority of hydrological models, and because aridity and drought occurrence will increase in the future (Roudier et al., 2016; Sheffield and Wood, 2008; Samaniego et al., 2018; Wehner et al., 2011; Haile et al., 2020).

60   Here, we tested the hypothesis that runoff-deficit exacerbation compared to precipitation deficit during droughts is in fact a common feature of water basins across environments characterized by different long-term aridity indices and evapotranspiration regimes (that is, water- or energy-limited regimes). We did so by using long-term observations from 210 basins across different European climates (area from 200 to 50,000 km^2). We calculated if basins show a shift in the water balance, and thus a runoff deficit comparatively larger than expected, by fitting a multivariate regression across annual cumulative streamflow, basin-wide annual precipitation, and a categorical variable denoting drought and non- drought years. The basins analysed, located from

65     $-10°$ to $+25°E$ and from $+35$ to $+70$ °N, have experienced at least one multi-year drought episode over the period 1979-2016 (see Figure 2).

2   Data and methods

2.1   Precipitation, temperature, evapotranspiration and soil data

The present analysis relied on two high-quality precipitation data sets: one having a comparatively high spatial resolution (i.e.,

70   0.25 °) and using ground-observations (the European Climate Assessment dataset project E-OBS, Haylock et al. (2008)), and a reanalysis dataset (ERA5, Hersbach et al. (2020), spatial resolution 36 km see below). The gridded-precipitation dataset E-OBS is derived through interpolation of the ECA&D (European Climate Assessment & Data) station data (Haylock et al., 2008). The station dataset comprises a network of 2,316 stations, with the highest density of station in northern and central Europe and lower density in the Mediterranean, northern Scandinavia and eastern Europe. E-OBS was used to analyse the

75   precipitation-runoff relationship for each basin, while ERA5 was used for the drought characterization. The motivations of the use of two different precipitation data sets are further explained in Section 2.5.

    From ERA5 we also extracted actual and potential evapotranspiration. ERA5 (Hersbach et al., 2020) is the latest climate reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), providing hourly data on many atmospheric, land-surface and sea-state parameters together with estimates of uncertainty. Evapotranspiration from ERA5

80   reanalysis was used because its relatively high quality (Martens et al., 2020) especially over Europe where a substantially large volume of observations is ingested.

    Precipitation, actual and potential evapotranspiration as well as air temperature variables used in this study are characterized by a spatial resolution of 36 km and monthly temporal resolution. ERA5 is available from the Copernicus Climate Change service (https://climate.copernicus.eu/climate-reanalysis, last access: 24 April 2020).

85   Both precipitation and evapotranspiration data were extracted for each basin by selecting pixels falling within the catchment boundaries and then averaged to provide basin-averaged annual precipitation and evapotraspiration time series since 1979. E-OBS and ERA5 precipitation and evapotranspiration variables were accumulated over the yearly time scale, while monthly temperature data were average out to obtain mean yearly temperature.

Catchment-average soil properties (rooting depth and total available water content) were obtained from the European Soil
90 Database Derived Data product (Hiederer and Hiederer, 2013; European Commission. Joint Research Centre. Institute for
Environment and Sustainability., 2013) for each basin.

2.2 Runoff data

Daily streamflow records for the 1980-2015 period over Europe were obtained by merging the Global Runoff Data Base
(GRDC); the European Water Archive (EWA); the Italian ISPRA HIS national database; the Portuguese national database; and
95 the Spanish national database (see data availability note for more information).

We considered the following processing steps:

1. From an initial number of more than 3,900 stations, 1,043 stations were retained by excluding (via visual inspection)
    those with evident dubious patterns due to human regulations (such as constant flows), inhomogeneity, problems in low
    flow range, missing values for a long period of time (> 2 year) (as suggested in Kundzewicz and Robson, 2004), or an
100 observation period below 20 years. Although care was taken in identifying these issues, some human-induced alterations
    are likely to be still present in these time-series. Nevertheless, a certain degree of disturbance can be tolerated (Murphy
    et al., 2013), considering also the annual granularity of our analyses.

2. Gauged stations for catchments with an area larger than 50,000 km^2 were excluded by the analysis because human
    disturbance is unavoidable at that scale (Piniewski et al., 2018).

105 3. Discharge time series were partially gap-filled via linear interpolation for a maximum time window of 5 days.

4. Only years where the number of observations were available for more than 315 days were retained.

2.3 Study area

The study area was initially composed of 1,043 basins, with an area ranging in size from 200 to 50,000 km^2. Basins are scattered
across Europe, over a longitude varying from -10 to 25°E and a latitude from 35 to 70°N (see Figure 1). The considered region
110 is characterized by a complex topography, with the Alpine and Pyrenees mountain chains crossing the continent from west
to east. Hilly plateaus gently slope towards the Great European Plain, a low, flat region extending from the Atlantic coast of
France to the Urals, crossed by many rivers and with densely populated cities. The climate is humid continental in central
and eastern Europe (with cold summers) and Mediterranean in southern Europe (with dry summers and humid winters). Mean
annual precipitation across Europe ranges from about 300 to 4000 mm yr^{-1}, depending on the location. The north Atlantic
115 coast of Spain, the Alps and Balkan countries generally receive higher precipitation amounts. Flood occurrence ranges from
spring to summer, moving from northeastern Europe towards the Alps, whereas the Mediterranean region and western Europe
are mainly subject to winter floods (Berghuijs et al., 2019).

2.4 Multi-year drought definition

Multi-year droughts were identified based on the precipitation deficit. The reason for using precipitation to characterize the multi-year drought period is that we are interested in analyzing the runoff response, therefore it was not used to define the drought (Saft et al., 2016b). In particular, we used the indications of Saft et al. (2016b) to define a multi-year drought periods, but we relied on the Standardized Precipitation Index (SPI, McKee et al., 1993) rather than on precipitation anomalies. The following procedure was adopted:

1. Calculation of the SPI, by fitting annual precipitation accumulations with a variety of distribution (i.e., normal, lognormal, exponential, generalized extreme, Fisk, Weibull, Gamma) and selecting the one that best fit data (i.e., having the maximum pvalue obtained with the Kolmogorov-Smirnov test). SPI was calculated both on the mean annual precipitation and on precipitation smoothed with a three-year moving window. Smoothing was applied to avoid single wet years to interrupt a long and significantly dry period.

2. To reduce the blurring effect of the moving window, the exact end date of the dry period was determined through analysis of the unsmoothed SPI data from the last negative three-year anomaly. The end year was set as the last year of this three-year period unless:

   - there was a year with a positive SPI >0.15, in which case the end year was set to the year prior to that year; or
   - if the last two years had slightly positive SPI (but each < 0.15), the end year was set to the first year of positive anomaly;

3. The first year of the drought remained the start of the first three-year negative SPI.

4. To ensure that the dry periods were sufficiently long and severe, we only used dry periods with the following characteristics: i) length over three years; ii) mean dry period anomaly < -0.8.

By defining drought in this way we ended up with 210 basins out of 1043 having experienced at least one multi-year drought episode over the available period of record. Although relaxing the procedure for the multi-year drought definition would have brought to a larger sample of basins, we preferred to maintain this approach to have consistent results with previous studies (Saft et al., 2016b) and because doing so guaranteed that the period analysed coincided with a period of a severe precipitation deficit. The above procedure resulted in a satisfactory multi-year drought definition (see Figure 2) that was validated with data found in the literature (Parry et al., 2012) with a minimum of three years to a maximum value of eight years for few basins (median duration of four years).

2.5 Shift in the precipitation-runoff relationship

We detected shifts in the precipitation–runoff relationship by fitting a multivariate regression across annual cumulative streamflow (target variable), basin-wide annual precipitation, and a categorical variable denoting drought and non- drought years

(Avanzi et al., 2020; Saft et al., 2016b):

$$Q_{BC} = b_0 + b_1 I + b_2 P + \epsilon \tag{1}$$

150 where I is a categorical drought variable (1 for years characterized by multi-drought and 0 otherwise, b_0 , b_1 , and b_2 are regression coefficients, ϵ is noise, and Q_{BC} is annual streamflow transformed according to a Box–Cox transformation following the arguments in (Avanzi et al., 2020):

$$Q_{BC} = \frac{Q^\lambda - 1}{\lambda} \tag{2}$$

where λ has been estimated from data to ensure linearity and heteroscedasticity (i.e., the λ that maximizes the log-likelihood

155 function, Box and Cox (1964)). A parameter b_1 different from zero (p-value <0.05) indicates a shift of the precipitation–runoff relationship during multi-year droughts. Following Avanzi et al. (2020), the statistical significance of coefficient b_1 during droughts was assessed based on whether the signs of the confidence bounds agreed (significance level α=5%, Kottegoda and Rosso (2008)).

The relative magnitude of the shift in precipitation vs. runoff (M_Q) for each basin was calculated by using the approach

160 suggested in Saft et al. (2016a):

$$M_Q = \frac{Q_{dry,P_I} - Q_{dry,P}}{Q_{dry,P}} \tag{3}$$

where Q_{dry,P_I} is the (predicted annual) runoff for a representative precipitation during dry periods according to the shifted precipitation–runoff relationship (1, $I = 1$), while $Q_{dry,P}$ is the full-natural flow for the same precipitation according to the non-shifted relationship (Eq. 1, $I = 0$). We assumed as representative annual precipitation the mean between average and

165 minimum annual precipitation across the entire period of record.

In this study, I in Eq. 1 was estimated by using SPI calculated based on ERA5 precipitation (I=1 during multi-year drought and I=0 for the other years), while the annual precipitation P was calculated based on E-OBS precipitation dataset. There are three reasons for that: i) we wanted to maintain as much as possible the independence between the drought definition (i.e., I) and the annual precipitation (P) in Equation 1, as to avoid influences on the fitting of Eq. 1. ii) we wanted to have consistent

170 ERA5-based drought definition evapotranspiration anomalies (both coming from the same dataset), and, iii) we wanted to rely on a higher spatial resolution product (i.e., E-OBS) for relating precipitation and runoff within the basin. However, both ERA5 and E-OBS precipitation are characterized by a relatively high accuracy over Europe (Massari et al., 2020) and interchanging them guaranteed very similar results (not shown here).

3 Results and discussion

175 ### 3.1 Multi-year droughts in Europe and water-budget-deficit exacerbation

During the last five decades, Europe has experienced various multi-year drought episodes (Parry et al., 2012; Spinoni et al., 2015; Hanel et al., 2018), which have been less studied but are as relevant as those that have impacted other world's regions,

such as Australia, California, or South America (Dijk et al., 2013; Griffin and Anchukaitis, 2014; Garreaud et al., 2017). For instance, the 1995-1997 multi-year drought impacted almost all Central and North Europe, but unlike episodes prior to 1979

180 (which were not taken into consideration here), it had a limited initial spatial extent and coherence on a regional basis, with a late exacerbation in terms of severity and extent by 1997 (Parry et al., 2012). The 1989-1991 drought impacted Belgium, France, Luxembourg, The Netherlands, as well as Balkan countries, the Mediterranean, and the Iberian Peninsula (Spinoni et al., 2015). The 2000-2005 drought impacted Northern Italy and the Iberian Peninsula (Santos et al., 2007; Fink et al., 2004). Our drought characterization based on the Standard Precipitation Index (McKee et al., 1993, see section 2.4 for details and

185 motivation of this choice) provided consistent results with the above mentioned studies (see Figure 2).

During these periods of severe precipitation deficit, 69 out of the considered 210 basins with at least one multi-year drought (i.e., 33%) showed a statistically significant shift in the water balance (i.e., a negative shift in the precipitation-runoff relationship, see Figure 3). This means that these 69/210 basins experienced statistically significant less runoff than would be expected based solely on the historical functional dependency of runoff with precipitation. This so-called negative shift is in contrast

190 with experiencing no shift or a positive shift, where the runoff deficit during droughts would be equal to or smaller than that expected based on the precipitation deficit, respectively. By way of examples, a shift in the precipitation-runoff relationship of -30% during one year belonging to a multi-year meteorological drought with mean annual precipitation equal to the long-term 10th percentile (which corresponds to a Standardized Precipitation Index of about -1.6 and thus to a severe drought, see McKee et al. (1993)) means that a basin of 1400 km^2 (75th percentile of the areas of the considered basins) will see an addi-

195 tional reduction in runoff volume of 43 Mm^3 compared to what one would predict solely based on precipitation deficit. This reduction in runoff is equivalent to the annual renewable freshwater resource for 10k people, considering that that the annual renewable freshwater resources averaged over the total European population for the period 1990-2017 reached 4,560 m^3 per person (noa). Although this exacerbation could also take place during shorter dry periods (Avanzi et al., 2020), we focused here on multi-year droughts because of their relevance from a water-management standpoint. Also, we expect precipitation deficit

200 to be particularly intense, sustained, and prolonged during multi-year droughts, which facilitates the quantification of any shift in the precipitation-runoff relationship and so the testing of our research hypothesis.

The shift magnitude ranges from about -85% to -12% (-28% in median, see Figure 3b), consistent and even larger that what found in previous works (Avanzi et al., 2020; Saft et al., 2016a; Tian et al., 2018, 2020). Only two basins showed a statistically significant positive shift (i.e., 3% of the sample). A similar result was also found by Saft et al. and no clear explanation was

205 provided. The reasons behind it could be due to the quality of the data used, the screening process, the uncertainty in the multi-year drought definition and in the fitting of the precipitation-runoff relationship. However, given the relatively high number of basins used here, the fact that only 2 basins show statistically positive shift is an index of the high control of the experiment and the high quality data used.

3.2 Evapotranspiration enhancement, catchment aridity, and water budget-deficit exacerbation

210 The distribution of basins with a statistically significant shift shows no obvious pattern of variability with the aridity index (see Figure 3c). Note that we assume the aridity index, calculated as the ratio between precipitation and potential evapotranspiration,

a proxy of the climate Unesco (1979). The only exception is that no shift was observed at very high latitudes (>65 °N), where winters are comparatively cold and the evapotranspiration regime is strongly energy-limited.

The magnitude of runoff-deficit exacerbation during droughts is strongly related to mean annual runoff, being larger for 215 drier basins (Figure 4). This outcome qualitatively agrees with earlier findings related to the pre-drought aridity index being an important predictor of shifts in the precipitation-runoff relationship in Australia (Saft et al., 2016b). Runoff exacerbation occurs in both rainfall- and cryosphere-dominated basins as defined by the month of maximum daily discharge (see again Figure 4). Exacerbation occurs both in energy- and water-limited regimes, as delimited using a standard Budyko framework (Budyko and Miller, 1974; Maurer et al., 2021), Figure 4b. This demonstrates that catchments may experience a shift in the 220 precipitation-runoff relationship and so an exacerbation of runoff deficit during droughts regardless of the predominant local climate (i.e., as defined by their long-term aridity index). Indeed, we found a statistically significant shift for 25% of the basins within the water-limited domain and for 35% of the basins in the energy-limited one (Figure 4b), including snow-dominated basins characterized by annual-runoff peak during late spring and summer. Nonetheless, drier catchments experience a larger runoff reduction during multi-year droughts than wetter catchments: shift magnitude asymptotically tends to -20% for wet 225 catchments, while drier basins reach shift magnitudes as large as -80% (Figure 4).

Given the annual water balance ($Q = P - ET - \Delta S$), we explain this relationship between shift magnitude and aridity with the potentially enhanced contribution of evapotranspiration to the annual water budget, particularly for water-limited regimes during droughts. In basins located over water-limited regimes, atmospheric demand for moisture is generally well above the available water storage needed to support evapotranspiration, so that the latter will have a significant impact on 230 already low runoff, especially at the beginning of a multi-year drought when water storage is comparatively large. In energy-limited environments, instead, evapotranspiration is mainly controlled by the available energy and may play a minor role in the annual allocation of incoming precipitation (Seneviratne et al., 2010).

The distribution of actual-evapotranspiration anomalies does show enhanced evapotranspiration during multi-year droughts compared to the remainder of the years, for both catchments located over energy- and water limited regimes (Figure 5). 235 Catchments located in a water-limited regime show a larger increase compared to those located in the energy-limited one (33 % vs 12%). A two-sample Kolmogorov-Smirnov test carried out between the distributions of evapotranspiration anomalies during droughts vs. during non-drought years confirms that the anomaly during droughts is statistically different (p<0.01) from that during non-drought years. This anomaly is generally larger for basins with the largest shifts (in absolute values, see Figure S2 in the supplementary material). Note that in Figure S2 we divided basins between those located above and those 240 located below 50 °N, because we did not observe basins with a positive anomaly in evapotranspiration below than 30% in the water-limited domain (we assumed that northern basins are mainly energy limited).

This regime of enhanced evapotranspiration during droughts was previously suggested by Teuling et al. (2013) and points to generally warmer conditions during droughts leading to additional demand for moisture, as also suggested by Mastrotheodoros et al. (2020). Here, we further expanded these findings by showing that evapotranspiration anomaly exacerbates runoff deficit 245 in the form of shifts in the precipitation-runoff relationship beyond Alpine regions and across various climates. A similar result

(not shown) was also found by calculating evapotranspiration as $ET = P - Q$ and thus neglecting the contribution of the change in storage, as in Teuling et al. (2013).

The distribution of evapotranspiration anomalies in Figure 5 shows a larger spread during droughts than during non-drought years. We attributed this increased variability in evapotranspiration during droughts to the regulation operated by energy (that
250 is, vapor pressure deficit) and available water (that is, storage) during these water-scarce periods. Figure 6a and b shows two such examples, which also iterate how a positive actual evapotranspiration anomaly is intimately coupled with runoff exacerbation (precipitation-runoff relationships for these two basins are shown in Figure S1). Figure 6a shows a multi-year drought period in the northern UK (1989-1994); this drought was characterized by both negative precipitation anomalies (-95% on average) and a positive anomaly in potential evapotranspiration (+79% on average). The result of this dry and
255 warm period was a positive actual evapotranspiration anomaly (+18%) and a markedly negative runoff anomaly (-106%). This situation significantly differs from 1996, a single dry year with i) much less precipitation than observed during many of the multi-year-drought years (e.g., 1990, 537 mm/y vs 368 mm/y) and, importantly, ii) a substantially lower potential evapotranspiration anomaly (-127%) denoting a much colder year with respect to 1989-1994. This cold-dry 1996 resulted in a negative actual evapotranspiration anomaly (-228%), which translated into a much smaller runoff deficit than the multi-year
260 drought (-79%, as opposed to -118 % in 1990). This demonstrated that in such energy-limited environments, the emergence of an enhanced-evapotranspiration regime during droughts is regulated by the available energy: if this is not sufficient, then actual evapotranspiration will not increase.

Similar conclusions can be drawn for the basins located in central Spain (Figure 6b), with some notable differences in this water-limited region. The multi-year drought period 1991-1995 in this area was characterized by a close-to-zero anomaly in
265 potential evapotranspiration (-2% on average) and a below-than-average precipitation (-98 %). This dry-mild period signifi-cantly differs from another single-dry and warm year, 2012 (+255 % of potential evapotranspiration and -183% precipitation). Despite the much warmer and drier 2012, we observed a relatively larger runoff deficit during the multi-year drought period (-99% on average, 25.9 mm/year) than in 2012 (-44%, 46 mm/year). Differently from the basin located in the northern UK (i.e., in an energy-limited region), the emergence of an enhanced-evapotranspiration regime in a water-limited region is much more
270 complex and regulated by both energy and available water storage (that can even result from carryover from previous years). Here, demand for moisture may also trigger plant-stomata closure thus reducing transpiration. Therefore, in water-limited regimes the year-to-year comparison of runoff deficit and evapotranspiration anomaly is not straightforward and can be further complicated by the precipitation variability typical of Mediterranean regions (Seager et al., 2019) . In any case, if storage is not sufficient, and/or other feedback mechanisms like stomata closure occur, then actual evapotranspiration will not increase and
275 runoff may be substantially higher than in relatively wetter periods.

As basin storage (i.e., ΔS) plays an important, but frequently neglected role in modulating runoff deficit via sustaining evapotranspiration during multi-year droughts (Van Loon and Laaha, 2015), we compared the average rooting depth and the total available water content (TAWC) distribution for basins characterized by significant versus non-significant shifts (see Figure 7 in the supplementary information). Results highlight that basins showing a significant shift in the precipitation-runoff
280 relation are characterized by a different distribution of rooting depth and TAWC from basins showing no-significant shifts (two-

sample Kolmogorov-Smirnov test with p-value<0.05). Because basins with a statistically significant shift show both a slightly deeper rooting depth and a larger TAWC, these findings tally with the enhanced ET anomaly for shifting basins in Figure 5, because a deeper rooting depth may provide access to deeper storage during water stress and so sustain evapotranspiration even during dry periods. Nonetheless, these findings are only of qualitative nature, given that distributions in Figure 7 overlap.

285 4 Implications and concluding remarks

In this study, we showed that exacerbation of runoff deficit compared to precipitation during droughts is a common feature of water basins across contrasting evapotranspiration regimes and aridity indices. This leads us to accept our initial hypothesis.

Runoff exacerbation is related to an increase in evapotranspiration occurring under two defined and concurrent preconditions: i) water storage can support ET during the drought period, and ii) there is a sufficient vapor-pressure deficit (mainly driven by

290 the temperature increase) to generate evapotranspiration (this is always true over water-limited regimes, while it may not happen over energy limited regimes, hence larger shifts in drier regions). When both circumstances are verified, then the catchment water balance shifts toward a new regime in which ET proportionally weights more than during wet periods. The macroscopic, bulk effect of this regime change is the shift in precipitation-runoff relationship as observed earlier (Avanzi et al., 2020). This shift is more pronounced in drier catchments, because evapotranspiration tends to be proportionally higher as long as enough

295 water is available to sustain atmospheric and vegetation demand for moisture. It is noteworthy that these drier catchments are areas of the world where water planners and ecosystem services are already challenged by limited water resources.

These results were obtained from an empirical, strictly data-based analysis, but are in line with earlier findings (Saft et al., 2016a, 2015; Avanzi et al., 2020), as well with those inferred from blending data with mechanistic modelling across the European Alps (Mastrotheodoros et al., 2020). The fact that we provided pieces of evidence that basins may develop a new

300 hydrological regime in response to evapotranspiration enhancement during droughts across aridity indices requires an understanding of catchment behaviour that goes beyond the assumption that runoff fluxes stationarily fluctuate as a function of precipitation variability, including the need to more comprehensively acknowledge the role of evapotranspiration for the long-term streamflow patterns (Mastrotheodoros et al., 2020).

Achieving such a more holistic understanding of the water budget during droughts is relevant from both a scientific and an

305 operational perspective. Conceptual rainfall-runoff models are still widely used in operational practice, as well as for many scientific purposes like climate-change studies, because they are parsimonious and computationally efficient, meaning they are easy to run in real time and provide a timely response (Pagano et al., 2014). Yet, these predictive tools may be inadequate tools during periods of runoff exacerbation like those we found here across Europe. The first reason is that calibration of these models remains inevitable (Beven and Freer, 2001), due to their typically oversimplified process representations (Gupta et al., 2012),

310 the presence of data errors (Montanari and Di Baldassarre, 2013), and recurring epistemic uncertainty related to heterogeneity of hydrologic processes across the landscape. However, the available observation period is often very limited both in time and space due to the significant decline of global river discharge monitoring over the past few decades (Crochemore et al., 2020), especially for medium-to-small basins. Thus, these "historically calibrated" parameters implicitly contain an assumption of

stationarity that may bias model predictions if the calibration period is climatically different from the prediction period (Coron
315   et al., 2012; Merz et al., 2011; Vaze et al., 2010; Chiew et al., 2014; Fowler et al., 2020), as it can happen during multi-year droughts. Shifts in precipitation-runoff relationship and the associated exacerbation in runoff deficit may thus determine unreliable runoff projections and a low efficacy of water planning and management measures if calibration did not adequately include shift-inducing drought periods.

The second reason goes beyond calibration uncertainty. Indeed, previous work found that standard hydrologic models are
320   prone to drops in modeling accuracy during shifting droughts even if some of these droughts were included in the calibration window (Avanzi et al., 2020). Thus, standard hydrologic models may be exposed to inaccuracy during periods of exacerbation of runoff deficit regardless of the calibration protocol used. These drops in accuracy may thus be more related to conceptual rather than to parametric uncertainty; in other words, commonly used hydrologic models may lack the representation of specific processes involved with the exacerbation of runoff deficit during droughts as they are manifested across the water budget (Bales
325   et al., 2018). For example, many models do not include stomatal response to dry periods and hydrologic regulation of plant rooting depth (De Kauwe et al., 2015; Fan et al., 2017), as well as coevolution mechanisms such as vegetation mortality and expansion (Goulden and Bales, 2019) or soil hydrophobicity. These coevolution mechanisms affect the so-called climate elasticity of evapotranspiration (Avanzi et al., 2020), that is, the capability of ET (and indirectly runoff) to respond and adapt to changes in climate. Given the projected warming climate and aridity, and the role of ET in driving the exacerbation of
330   runoff during droughts, improving understanding of this elasticity appears an urgent task for future work. Although the study catchments were limited to Europe, the diversity of physiographic and aridity settings, as well as the number of catchments used suggest that evapotranspiration may be an important factor involved in shifting the water balance of other regions of the world.

Data availability. Data from E-OBS were obtained via https://www.ecad.eu/download/ensembles/download.php, ERA5 data were down-
335   loaded from the https://climate.copernicus.eu/climate-reanalysis, runoff data and basin areas were downloaded from the Global Runoff Data Base (GRDC); the European Water Archive (EWA); the Italian ISPRA HIS national database (http://www.hiscentral.isprambiente.gov.it/hiscentral/default.aspx); the Portuguese national database (http://snirh.pt/); and the Spanish national database (http://ceh-95flumen64.cedex.es/anuarioaforos/default.asp).

Code and data availability. All codes are available by request

340   *Author contributions.* Christian Massari and Francesco Avanzi designed, coordinated the study and made the analyses. Giulia Bruno helped in the data analysis and interpretation, Daniele Penna, Simone Gabellani helped in the result interpretation, Stefania Camici helped in the result interpretation and in the data collection. All authors contributed to the editing of the manuscript.

Competing interests. The authors declare no competing interests

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Figure 1. (a) Aridity index (P/PET) distribution of the basins of the study area. (b) frequency distribution of the area of the basins. (c) frequency distribution of the aridity index of the basins of the study area.
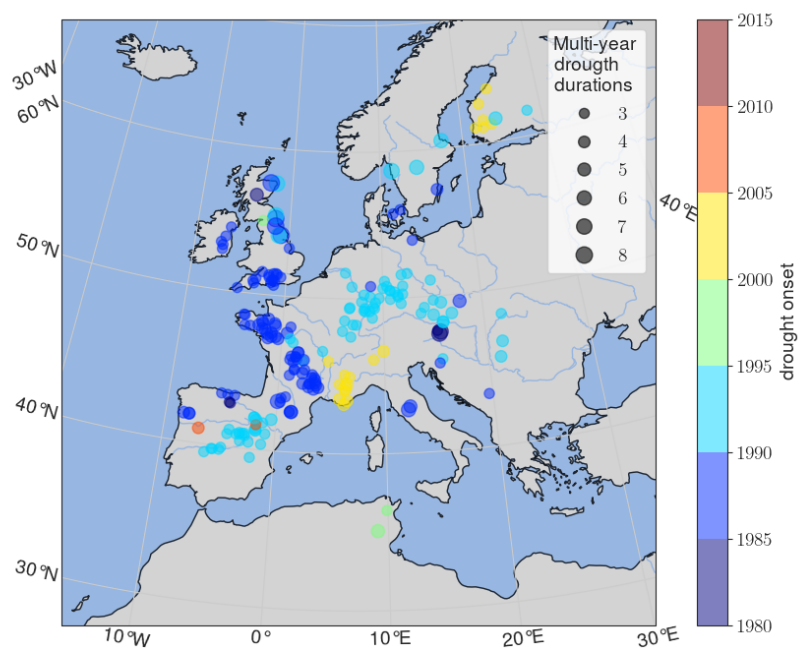
Figure 2. Multi-year drought periods observed in Europe from 1980 to 2017 with the indication of the onset and duration.
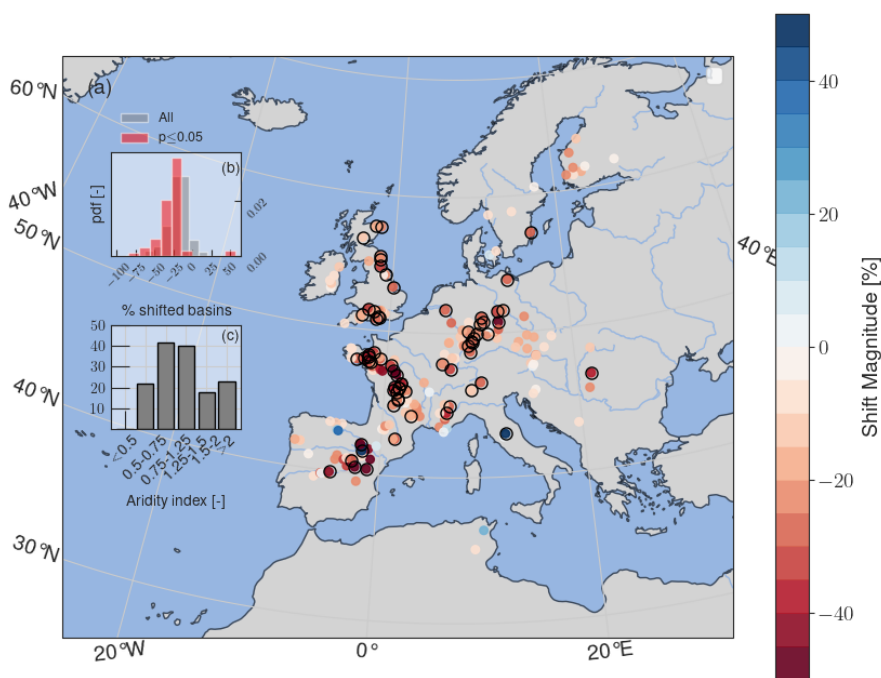
Figure 3. Spatial distribution of the magnitude of the shift in the precipitation-runoff relationship (a). The shift in the precipitation–runoff relationship was calculated by fitting a multivariate regression across annual cumulative streamflow (target variable), basin-wide annual precipitation, and a categorical variable denoting drought and non- drought years. Each dot refers to the position of the river gauge station. The basins where the shift was found statistically significant with p-value<0.05 plotted with a black edge. Darker red dots refer to catchments where a larger exacerbation of runoff deficit during multi-year drought periods was observed. (b) Probability density function (pdf) of the magnitude of shift found for the basins in the study area. (c) percentage of the basins showing statistical significant shift (p<0.05) as a function of the aridity index calculated as the ratio between ERA5 precipitation and potential evapotranspiration. The percentage has been calculated by stratifying basins for each aridity index class.
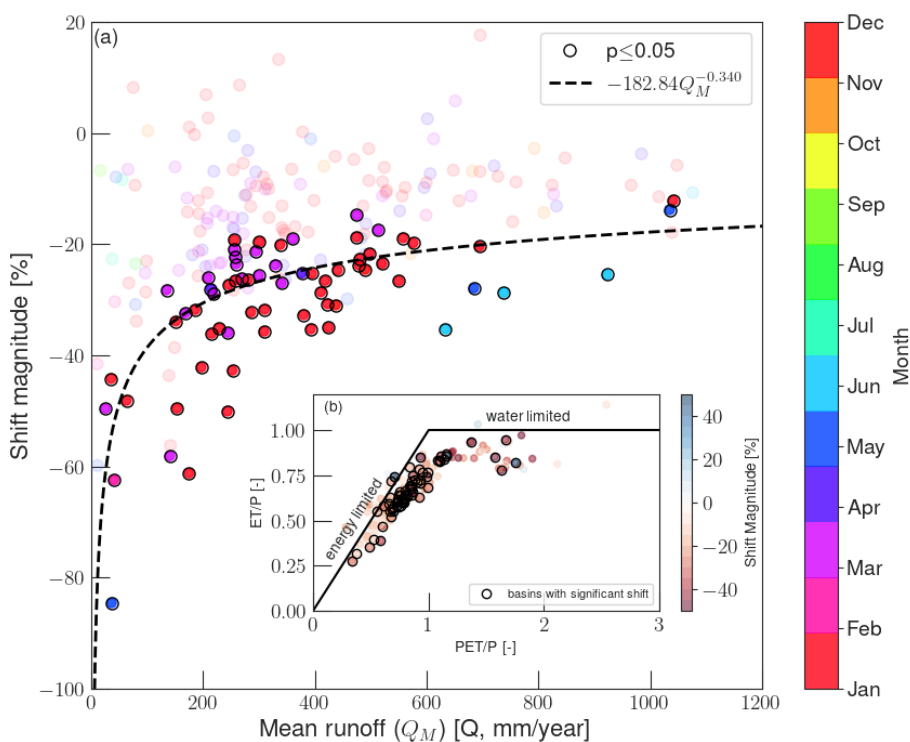
Figure 4. (a) Shift magnitude versus mean discharge grouped for maximum daily monthly discharge for all basins and for basins showing a shift (p-value<0.05 are shown with darker color and a black edge). The black dashed line curve was obtained by fitting mean runoff vs. shift magnitude of basins showing a statistically significant shift. (b) Energy- and water-limited domain of the basins of the study area as a function of the shift magnitude. Dots with black edge indicate basins showing statistically significant shift at p<0.05. PET, P and ET indicate ERA5 potential evapotranspiration, precipitation and actual evapotranspiration, respectively.
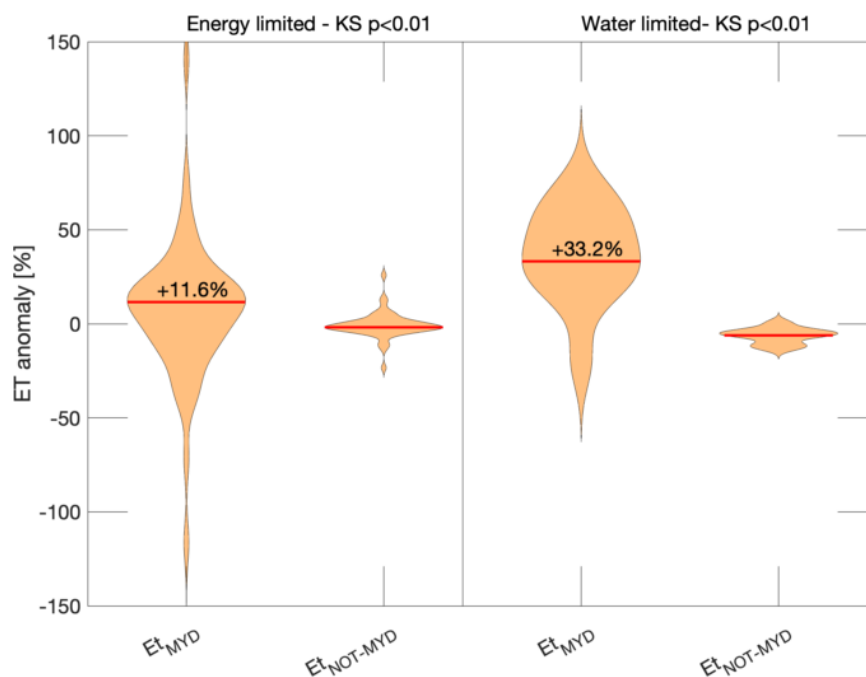
Figure 5. ERA5 evapotranspiration anomalies distribution for multi-year drought period (ET_{MYD}) versus non multi-year drought periods ($ET_{NOT-MYD}$) for basins showing statistically significant shift (p<0.05) and characterized by an energy-limited regime (left) and for basins in the water-limited regime (right). KS refer to the two- sample Kolmogorov-Smirnov test between the distribution of evapotranspiration anomalies of ET_{MYD} and $ET_{NOT-MYD}$. The red line in the violin plots refers to the median value. The anomalies are calculated as the ratio between the deviation with respect to the long-term mean and the absolute long-term standard deviation on the catchments showing statistically significant shift

(a) North UK

(b) Central Spain
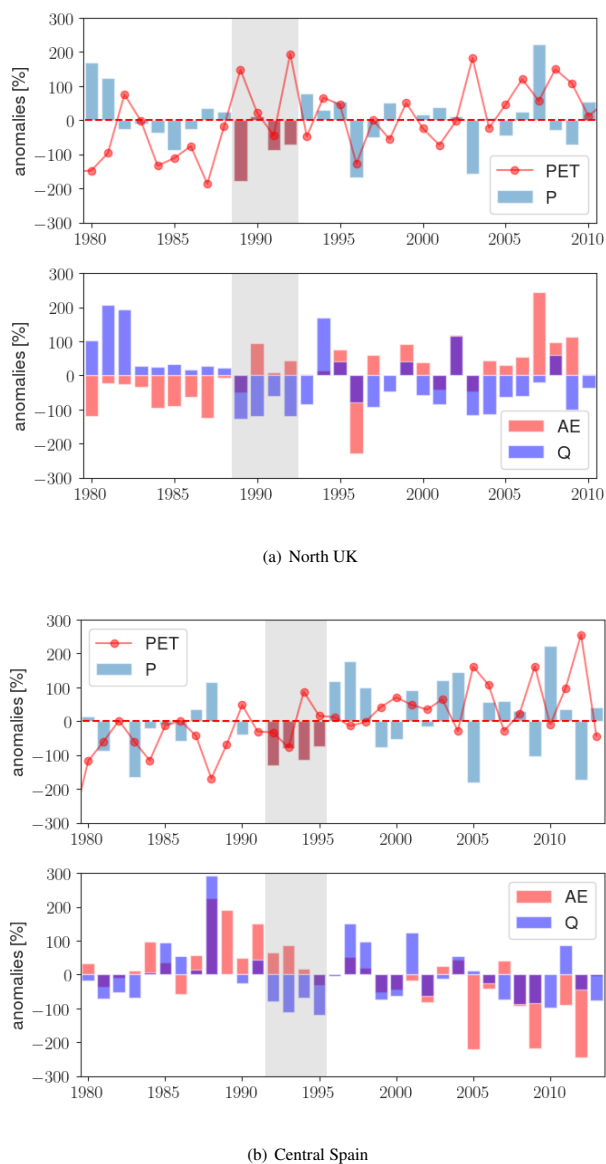
Figure 6. a) Precipitation and potential evapotranspiration long-term mean anomalies (top), actual evapotranspiration and runoff anomalies (bottom) for a basin located in northern UK (6.8°O, 53.6°N) characterized by a energy limited regime according to Figure 4b. (b): same as (a) but for Central Spain (5.4°O, 39.8°N) characterized by a water- limited-regime. Gray areas represent the identified multi-year drought periods.
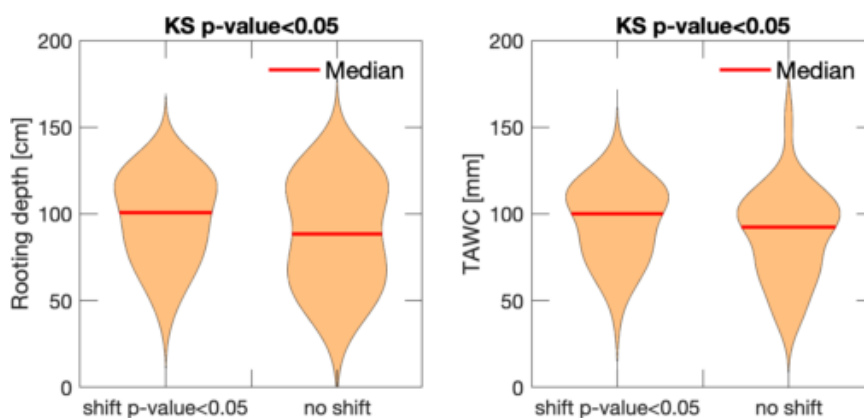
Figure 7. Average rooting depth and total available water (TAWC) content for basins characterized by a significant shift in the precipitation-runoff relation and those where shift was not significant. KS refer to the two- sample Kolmogorov-Smirnov test between the distribution of basins with shift (p-value<0.05) and and basin showing not statistically significant shift.