Impact of the dry day definition on Mediterranean extreme dry spells analysis

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ABSTRACT

To define a dry day, the most common approach is to identify a fixed threshold below which precipitation is considered equivalent to zero. This fixed threshold is usually set to account for measurements errors and also for precipitation losses due to the atmospheric evaporation demand. Yet, this threshold could vary in time according to the seasonal cycle but also in the context of long-term trends such as the increase of temperature due to climate change. In this study, we compare extreme dry spells defined either with a fixed threshold for a dry day (1 mm) or with a time-varying threshold estimated from reference evapotranspiration (ET_0) for a large data base of 160 rain gauges covering large parts of the Mediterranean basin. Results indicated positives trends in ET_0 in particular during summer months (June, July and August). However, these trends do not imply longer dry spells since the daily precipitation intensities remains higher than the increase in the evaporative demand. Results also indicated a seasonal behavior: in winter the distribution of extreme dry spells is similar when considering a fixed threshold (1 mm) or a time-varying threshold defined with ET_0. However, during summer, the extreme dry spell durations estimated with a 1 mm threshold are strongly underestimated by comparison with extreme dry spells computed with ET_0. We stress the need to account for the atmospheric evaporative demand instead of using fixed thresholds to define a dry day when analyzing dry spells, in particular with respect to agricultural impacts.

Keywords:

Extremes, dry spell, Mediterranean, reference evapotranspiration, atmospheric evaporative demand

1. INTRODUCTION

The Mediterranean region is affected by severe droughts episodes, linked to the strong inter-annual variability of precipitation patterns (Mariotti and Dell'Aquila, 2012). These droughts can impact agricultural production (Páscoa et al., 2017), and water resources (Lorenzo-Lacruz et al., 2013), in particular when occurring during the (wet) winter season (Raymond et al., 2016). In addition, several studies indicate a tendency toward a warming and drying of the Mediterranean region that could intensify in the future according to climate projections (Hoerling et al. 2012, Hertig and Tramblay 2017, Naumann et al. 2018).

There are different methods to analyze droughts, by means of drought indices (Mishra and Singh, 2010, Mukherjee et al., 2018) but also by explicitly modelling the frequency/duration of dry spells (Vicente-Serrano and Beguería-Portugués, 2003). A dry spell is meteorologically defined as a sequence of consecutive dry days with no precipitation, or precipitation below a certain threshold. Although dry spells cannot be used to determine drought severity, as a consequence of the climatological differences, they are highly useful to assess spatial differences in the drought hazard probability (Lana et al., 2006), but also to determine possible trends associated to climate change (Raymond et al., 2016). Moreover, analyses based on dry spells have usually been used for agricultural management purposes in different regions of the world (Sivakumar, 1992, Lana et al., 2006, Mathugama and Peiris, 2011, Raymond et al., 2016).

Several authors analyzed long dry spells, considering different precipitation thresholds (1 to 10 mm/day), but fixed for the whole observation period (Vicente-Serrano and Beguería-Portugués, 2003, Lana et al., 2006, Serra et al., 2016, Raymond et al., 2016, 2017, Tramblay and Hertig, 2018). About the threshold considered for a "dry" day it is usual to use values higher than zero to account for measurements errors or very little amounts of rain that are not available for plants or water resources, due to interception or/and direct evaporation (Douguedroit, 1987, Raymond et al., 2016). In a climate change context it is also used to reduce the typical "drizzle effect" of dynamical models which results in too many low precipitation amounts compared to observations. The determination of this threshold, noted Daily Rainfall Threshold (DRT), can be a key issue to relate dry spells risk to impacts in different sectors. Douguedroit (1987) defined a threshold of 1 mm of precipitation in environments with a Mediterranean climate, because below this amount the rainfall is

101 generally not absorbed by soils under conditions of high evapotranspiration. It is the most
102 widely used daily rainfall threshold (Polade et al., 2014, Raymond et al., 2016, 2017), even
103 though this arbitrary value has not been supported by any experimental study.
104
105 However, fixed thresholds are not representative of real ground conditions, since the
106 evaporation varies throughout the year and for different locations. The atmospheric
107 evaporative demand (AED) can strongly modulate the net precipitation that is available for
108 the plants, affecting water stress levels by plants and crops (Allen et al., 2015; Anderegg et al.,
109 2016; Lobell et al., 2015; Lobell and Field, 2007). It is expected that based on precipitation
110 records, dry spells of similar duration could be characterized by different water stress as a
111 function of the differences in the AED as suggested by drought indices using precipitation and
112 the AED for calculations (Beguería et al., 2014, Manning et al., 2018). AED can be calculated
113 using meteorological data from different approaches such the potential evaporation
114 (McMahon et al., 2013) or the reference evapotranspiration (ET_0) (Allen et al. 1998) but it can
115 be also measured by means of Evaporation Pans. In the Mediterranean region different studies
116 have shown an increase in the AED in recent decades (Vicente-Serrano et al., 2014) that has
117 increased drought severity (Vicente-Serrano et al., 2014a, Stagge et al., 2017). It is unclear
118 how these trends could affect extreme dry spell severity.
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120 The goal of the present study is to evaluate the influence of different daily precipitation
121 thresholds to define a dry day on the estimation of seasonal extreme dry spell hazard in the
122 Mediterranean. The novelty of the approach proposed herein, is to use the AED to identify dry
123 days prior to the analysis of extreme dry spell risk. Two thresholds to define a dry day are
124 compared: 1 mm/day, the threshold commonly used in most Mediterranean studies, and a
125 daily precipitation threshold defined by the AED, thus seasonally and temporally variable.
126 Two questions are addressed in the present work: (i) are there trends in extreme dry spell
127 length in Mediterranean region and is the trend detection influenced by the way at which dry
128 days are defined? (ii) since in most studies a distinction is made between winter and summer
129 dry spells due to their different characteristics and impacts (Raymond et al. 2018, Tramblay
130 and Hertig, 2018)- is there a different impact on the estimation of extreme dry spells in winter
131 or summer according to different daily rainfall thresholds?
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2. PRECIPITATION AND REFERENCE EVAPOTRANSPIRATION DATA

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A network of 160 stations with long daily precipitation records in the Mediterranean region is considered (see Hertig and Tramblay, 2017, Tramblay and Hertig, 2018 for more details about this dataset). Since most stations have almost complete records between 1960 and 2000, it is the period considered in the present analysis to allow a comparison between stations. The years with more than 5% missing days have been discarded from subsequent analysis. A preliminary sensitivity analysis considering different missing day ratios has shown that it does not impact the results.

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In addition to precipitation data, as a representative and spatially comparable metric of the AED, the reference evapotranspiration (ET_0) from the Climate Research Unit (CRU) dataset version 4.2 is considered (Harris et al., 2014). Several studies (McVicar et al., 2012a, 2012b, Todorovic et al., 2013, Vicente-Seranno et al., 2014b, Anabalón and Sharma, 2017) highlighted the need to consider a physically based ET_0 calculation, such as the FAO-PM, to account for possible changes in other variables than temperature in the AED and to have an accurate quantification of the climate change effect on drought (Trenberth et al., 2014). Reference evapotranspiration is defined as the rate of evapotranspiration, only influenced by the atmospheric conditions, from a clipped grass surface having 0.12 m height and bulk surface resistance equal to $70 \, \mathrm{s \, m^{-1}}$, an assumed surface albedo of 0.23 and no moisture stress. In the CRU dataset, the ET_0 is computed from a simplified version of the FAO Penman-Monteith (FAO-PM) equation (Allen et al. 1998) that uses data of air temperature, sunshine duration, vapor pressure deficit and a climatology for wind speed. The detail for the computation is given in Harris et al. (2014). By comparison, Potential Evapotranspiration (PET) is the evapotranspiration from a given crop surface, requiring the use of crop coefficients that can vary in time due to the development stage of the vegetation. The use of ET_0 allows the comparison between stations and does not require estimating local crop coefficients.

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Two different definitions for a dry spell are used in the present work. The first one considers a dry spell as consecutive days with precipitation below 1 mm. For the second one, the ET_0 is considered as a threshold to define a dry day when P-ET_0 <= 0. In addition, to provide a measure of rainfall intensity we computed from daily precipitation the Simple Precipitation

Intensity Index (SDII), defined as the monthly sum of precipitation during wet days divided by the number of wet days in the month (expressed as mm/day). It is an interesting metric for the present dry spells analysis, since the SDII can provide a measure of rainfall intensity that can be compared with the threshold used to define a dry day during a dry spell.

3. METHODS

3.1 Statistical tests

To test the presence of trends in the different station time series, the non-parametric Mann-Kendall (Mann, 1945) test was used. Since the presence of autocorrelation in the data could lead to an increased number of type I errors (Serinaldi et al., 2018), we used the trend-free pre-whitening method introduced by Yue et al. (2002) and modified according to Serinaldi et al. (2006). In addition, since the tests are repeated on a large ensemble of stations (160), we also implemented the false discovery rate (FDR) method of Benjamini and Hochberg (1995) to distinguish between at-site and regionally significant trends (Wilks, 2018).

To compare the different extreme dry spells distributions, computed with different definitions of a dry day, the Anderson Darling test (Scholz and Stephens, 1987, Viglione et al., 2007) is considered. The test verifies the hypothesis that two independent samples belong to the same population without specifying their common distribution function. The test statistic measures the distance between the empirical cumulative distribution functions and places more weight towards the tail of the distributions, hence making it adapted to the analysis of extreme values.

3.2 Distribution fitting

To compute the return levels for different extreme dry spell durations, there is the need to fit a distribution to the samples. No single distribution is commonly applied to extreme dry spell lengths and also we define differently dry spells than previously (Vicente-Serrano and Beguería-Portugués, 2003, Lana et al., 2006, Serra et al., 2016). Thus, the GEV, Gamma and Log normal distribution are first compared to represent extreme dry spells, using the maximum likelihood estimation method. A split-sample procedure has been implemented to validate the choice of the distribution. The same procedure as described in Zkhiri et al. (2017) or Renard et al., (2013) is retained, based on a bootstrap cross-validation. The relative average

203 root mean square error (RRMSE) for the validation samples is used as an evaluation metric to
204 select the best distribution. The best distribution retained is then used to compute extreme dry
205 spell quantiles computed with different precipitation thresholds for a dry day.

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207 **3.3 Definition of the seasons**

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209 The Mediterranean regions are classified as Csa and Csb climate types in the Köppen
210 classification (Peel et al., 2007), defined as climates with a precipitation deficit during
211 summer months (when the sub-tropical high pressure belt moves northward and prevent
212 moisture advection from westerlies). The Mediterranean climate is then characterized by two
213 contrasted seasons: A summer (dry) season from around April to September and an extended
214 winter season (wet) from October to March, with most of the precipitation occurring during
215 this period. Yet the transitional months could vary depending on the location and one single
216 definition of the Mediterranean seasons is probably not appropriate due to strong North/South
217 and West/East variations on the beginning/finishing dates for the season of precipitation
218 deficit. This has been highlighted be the recent study of Raymond et al. (2018). Reiser and
219 Kutiel (2009) previously observed different lengths for the wet season, (on 40 stations) with
220 less than 6 months in the south and up to 10 months in the North. Thus, in the present study
221 we choose to define the season lengths for each station according to an objective criterion,
222 being the precipitation deficit in summer (ie. the months when P-ET_0 = 0 are defined as the
223 summer season). Then a clustering approach (Ward, 1963) is used to group stations with a
224 similar seasonality. The optimal number of clusters is estimated with the gap statistic
225 (Tibshirani et al. 2001) and silhouette plot (Kaufman and Rousseeuw, 1990).

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227 **4. RESULTS**

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229 **4.1 Climatic trends**

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231 There are increasing trends in ET_0 in West/Central Mediterranean stations, mostly during
232 summer months and to a lesser extent in March for the Iberian Peninsula (Figure 1). These
233 monthly trends imply an increase of ET_0 at the annual scale for these stations (Spain, South
234 France, Italy, East Algeria and Tunisia). When tested on the annual total ET_0, the trends are
235 regionally significant at 67 stations, located in south France, Spain, Middle East, Tunisia and
236 Algeria, Italy and the Adriatic. With both thresholds 1mm or Et_0 to define a dry days (named

thereafter respectively S1 and SET_0), there is an increase in the frequency of dry days in February and March, centered on the stations in Spain, Portugal and South France (Figure 2 and Figure 3). The spatial patterns of detected trends are similar with the two thresholds but the increase is more pronounced, with more regionally significant trends, when using ET_0 as threshold for dry days. Yet, the increase in ET_0 during summer months does not imply an increase in the frequency of dry days during this season when considering ET_0 to define a dry day. On the contrary, in March the increase in Et_0 in the Western Mediterranean is accompanied by an increased frequency of dry days. The monthly ET_0 during winter months lies in the interval 0.5 to 2 mm for all stations, when for the summer daily ET_0 ranges between 3 mm and 7 mm/day.

Additionally, we tested the trends for the Simple Daily Intensity Index (SDII). The results indicate a decrease of SDII for a few stations, in particular in February in South France, but overall these trends are not regionally significant. An interesting feature is illustrated in Figures 4 and 5: the ratio between ET_0 and the SDII during June, July and August show a remarkable North/South difference: In the south the average precipitation amounts during summer stay below evapotranspiration during rainfall events. During the summer months there is also a large variability and the ration is often exceeding 1. This implies that, on average, precipitation events will not be able to end a succession of dry days and this characteristic favor very long dry spells during summer. On the opposite, in the north the average precipitation during an event stay above ET_0.

4.2 Seasonal comparison of extreme dry spells

As mentioned in the previous section and in section 3.3, there is a different seasonal behavior of dry spells between winter and summer months. In addition, several studies have shown that long dry spells during the winter season may have more severe consequences than those occurring during summer. This justifies a seasonal analysis of the extreme dry spells defined according to different dry day definitions. Nevertheless, prior to a seasonal comparison, a classification of stations according to monthly net precipitation (P- ET_0) has been performed as explained in section 3.3. The classification shows a marked distinction between two clusters as shown in Figure 6, very similar to the spatial patterns of Figure 5, with northern stations (approximately north of 40°N) with a precipitation deficit from April to September and southern stations with a precipitation deficit from March to October.

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272 Then for each season and each year, the maximum dry spell lengths have been extracted at the

273 different stations according to two thresholds for a dry day: 1 mm and ET_0 (thereafter the

274 extreme dry spells derived from the two thresholds are noted S1 and SET_0). Then, the

275 Anderson-Darling test has been applied between summer and winter maxima. For S1, the test

276 rejects the null hypothesis at the 5% significance level for 135 stations. The remaining 25

277 stations where the winter and summer distributions are found similar are located in northern

278 Mediterranean countries such as France (including Perpignan, Nîmes, Orange), Spain

279 (Huesca, Valencia, Soria, Vallalolid), Italy (Ferrara, Genoa), Croatia (Gospic, Zavizan). For

280 SET_0, the test rejects the null hypothesis for 155 stations (except Mantova, Verona, Reijka,

281 Milan, Mons). This indicates that the majority of stations the winter and summer distributions

282 of extreme dry spell are different whatever the threshold considered for a dry day. Indeed, the

283 extreme dry spells tends to be longer in summer than in winter for all stations and this feature

284 accentuates with increased aridity. This result justifies the need to perform a seasonal analysis

285 when considering extreme dry spells risk.

286

287 Finally, the same Anderson-Darling test has been applied for a given season between extreme

288 dry spells computed with the threshold 1 mm (S1) and extreme dry spells computed with ET_0

289 (SET_0). As shown in Figure 7, there are strong differences in summer when extreme dry spells

290 are computed with the dry day threshold 1 mm or ET_0. For most stations, the two distributions

291 are significantly different at the 5% level. On the opposite, for winter it can be assumed that

292 extreme dry spells computed with 1mm or the ET_0 are stemming from the same distribution.

293 This is due to the fact that during winter the AED is low and close to the value 1mm.

294

295 **4.3 Return levels of extreme dry spells**

296

297 Prior to the fitting of statistical distributions, there is the need to verify the hypothesis of

298 stationarity. Overall, there are not significant trends in extreme dry spells duration, neither for

299 winter or summer, with the threshold 1 mm or ET_0 to define dry days. This finding is quite

300 surprising since there is an increase of ET_0 in summer and one would expect an increase in

301 dry spells when considering ET_0 as daily rainfall threshold. As elements of explanations, it

302 was shown before that the increase of ET_0 is focused only in the months of June, July and

303 August (see Figure 1). Furthermore, two extreme cases are exemplified here, Montpellier in

304 the North (783mm/year on average) and Gafsa in the South (168mm/year). In Figure 8 the

305   daily rainfall for a random year (1998) is plotted together with Et_0 at the beginning of the time

306   period (1960), in 1998 and for the end of the time period (2000). In Gafsa or Montpellier, the

307   increase of ET_0 in summer is not high enough to exceed daily events of intense precipitation

308   (often thunderstorms). In the south, the ET_0 is already higher than most of precipitation events

309   (e.g. Figure 5), except for a few high-intensity events above ET_0. Still, the increase in ET_0

310   does not impact the longest dry spells sequences as indicated by the trend analysis.

311

312   The GEV, Log Normal and Gamma distributions have been compared to fit extreme dry

313   spells. The results are illustrated in Figure 9 for 10 stations located in different regions having

314   long records and very little or no missing data over their full records. For both S1 and S ET_0,

315   the Gamma distribution outperforms the GEV or Log-Normal since it provides lower mean

316   RRMSE values in validation results on independent samples. Quantiles corresponding to a 20-

317   year return period have been computed from a Gamma distribution for each station and each

318   season, according to the two different thresholds for dry days. A relative difference between

319   the two quantiles has been computed, taking the S1 quantile as reference, since it is up to now

320   the most widely used approach to estimate dry spell durations. Results, shown in Figure 10,

321   indicate a strong underestimation of extreme dry spells during summer when using the fixed

322   threshold of 1 mm. This underestimation is on average -29%, but only 4% in winter. This

323   result question the use of a fixed threshold of 1 mm during summer, since it is not

324   representative of the real amount of water available on the ground due to evaporation. On the

325   contrary, focusing on winter only with a fixed threshold 1 mm does not induce strong

326   uncertainties due to the low AED during this season.

327

328   **5. Discussion**

329

330   The results obtained in the present work indicate the need to consider AED in particular

331   during summer months to define a dry day, which is probably more realistic than with a fixed

332   threshold of 1 mm. In more arid environments than the Mediterranean region, such as the

333   Middle East and North Africa regions, it would mean that the analysis of dry spell could be

334   strongly impacted whether the AED is taken into account or not. It implies that it necessary to

335   re-define appropriate thresholds to define dry days according to different regions. By

336   comparison with other drought indices, such as the SPI or SPEI that are averaged on a

337   monthly basis for different time horizons (Mukherjee et al., 2018), the explicit consideration

338   of extreme dry spells could be an interesting way of relating dry spells to impacts. Indeed, dry

spell durations computed with dry day thresholds representative of real climate conditions could be directly related to plants phenology to study the drought impacts for different agricultural productions. This new definition of dry spells, considering a time varying threshold based on AED, is a departure from the classical viewpoint of a meteorological drought index since it tries to relate the atmospheric and ground conditions to assess the amount of water that is actually available for plants or water use. In that sense, it relates to the SPEI but tailored at the scale of individual dry spell events.

The results of the present study rely on the estimation of AED with reference evapotranspiration. Still more reliable than ET_0 estimates from temperature only, the FAO-PM equation may not be fully representative of the AED at the different locations considered. McMahon et al. (2013) provided a synthesis of the uncertainties related to the estimation of the AED: data limitations such as wind or humidity, which are not always available for all gauging stations, but also the fact that reference evapotranspiration rely on an hypothetical grass surface that may not be representative of the real land cover at the different stations during the different seasons of the year. Indeed, it is possible to derive the potential evapotranspiration from reference evapotranspiration using crop coefficients representative of the real ground conditions. These changes in land cover could modulate the AED between different locations. As an alternative, it could be possible to use actual evapotranspiration but since it cannot be measured (at least for large areas) this would require the use of land surface modelling. However, there are differences in actual evapotranspiration computed from different land surface models, due to different parametrization, climate forcing, and representation of the semi-arid surface processes (Quintana-Seguí et al., 2019). Finally, it must be stressed that the estimation of AED in the Mediterranean for a long term perspective and climate change impact studies, must face several sources of uncertainties, such as land cover changes, forest fires that could induce drastic changes in surface processes, water soil conditions influenced by human activity and irrigation, among others.

6. Conclusions

In this study, the extreme dry spells defined either with a fixed dry-day threshold (1 mm/day) or with a time-varying threshold estimated from reference evapotranspiration (ET_0) have been compared for a large data base of 160 rain gauges covering the whole Mediterranean basin. An increase in ET_0 is found for summer months (JJA) mainly in the central/western parts of

373 the Mediterranean basin. The reported trends for summer are consistent with previous studies

374 in Spain, driven by a decrease in relative humidity and an increase of maximum temperature

375 (Vicente-Serrano et al., 2014a, 2014b). Also increases in the number of dry days are found for

376 February and March at a large number of stations, either with 1mm or ET_0 to define a dry day.

377 However, no trends are detected for extreme dry spell lengths when using both thresholds to

378 define a dry day. The distributions of extreme dry spells have been found to be different for

379 winter and summer, with much longer extreme dry spells during summer. Also, for many

380 locations a stronger variability in winter extreme dry spells became apparent. These results

381 highlight the need of a seasonal analysis to avoid the misestimating the extreme dry spells

382 risk. Despite the climatic trends on precipitation and evapotranspiration, there are no

383 significant trends in seasonal extreme dry spells risk in most areas. The frequency analysis of

384 seasonal extreme dry spells reveals that using a fixed threshold set to 1 mm implies an

385 underestimation of extreme dry spells risk by comparison to a time-varying threshold

386 representing evapotranspiration during the extended summer season. The time-varying

387 thresholds appear a more relevant choice representative of real atmospheric conditions, but

388 this needs to be further confirmed by relating extreme dry spells computed with this new

389 approach and drought impacts in different sectors (agriculture, vegetation, etc.). As a

390 conclusion, we stress the need to account for the atmospheric water demand when analyzing

391 dry spells in particular if the goal is to relate them with agricultural impacts.

392

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394

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401 **References**

402

403 Allen, R.G., Pereira, L.S., Raes, D., Smith, M.: Crop evapotranspiration, guidelines for

404 computing crop water requirements, Irrigation and drain, Paper No 56 FAO, Rome, Italy,

405 2008.

406

407   Allen, C. D., Breshears, D. D., McDowell, N. G.: On underestimation of global vulnerability

408   to tree mortality and forest die-off from hotter drought in the Anthropocene, Ecosphere, 6(8),

409   doi:10.1890/ES15-00203.1, 2015.

410

411   Anderegg, W. R. L., Klein, T., Bartlett, M., Sack, L., Pellegrini, A. F. A., Choat, B., Jansen, S.:

412   Meta-analysis reveals that hydraulic traits explain cross-species patterns of drought-induced

413   tree mortality across the globe, Proc. Natl. Acad. Sci. U. S. A., 113(18), 5024–5029,

414   doi:10.1073/pnas.1525678113, 2016.

415

416   Anabalón, A., Sharma, A.: On the divergence of potential and actual evapotranspiration

417   trends: An assessment across alternate global datasets, Earth's Future, 5, 905-917.

418   doi:10.1002/2016EF000499, 2017.

419

420   Beguería, S., Vicente-Serrano, S. M., Reig, F., Latorre, B.: Standardized precipitation

421   evapotranspiration index (SPEI) revisited: parameter fitting, evapotranspiration models, tools,

422   datasets and drought monitoring, Int. J. Climatol., 34, 3001–3023, 2014.

423

424   Benjamini, Y., Hochberg, Y.: Controlling the false discovery rate: A practical and

425   powerful approach to multiple testing, J. Roy. Stat. Soc. B, 57, 289–300, 1995.

426

427   Douguedroit, A.: The variations of dry spells in marseilles from 1865 to 1984, J. Climatol., 7,

428   541–551, 1987.

429

430   Greve, P., Orlowsky, B., Mueller, B., Sheffield, J., Reichstein, M., Seneviratne, S. I.: Global

431   assessment of trends in wetting and drying over land, Nat. Geosci., 7(10), 716–721,

432   doi:10.1038/NGEO2247, 2014.

433

434   Harris, I., Jones, P., Osborn, T., Lister, D.: Updated high-resolution grids of monthly climatic

435   observations – the CRU TS3.10 Dataset, Int. J. Climatol., 34: 623-642. doi:10.1002/joc.3711,

436   2014.

437

438   Haylock, M.R., Hofstra, N., Klein Tank, A.M.G. , Klok, E.J., Jones P.D., New, M.: A

439   European daily high-resolution gridded dataset of surface temperature and precipitation, J.

440   Geophys. Res (Atmospheres), 113, D20119, doi:10.1029/2008JD10201, 2008.

441

442   Hertig, E., Tramblay, Y.: Regional downscaling of Mediterranean droughts under past and

443   future climatic conditions, Global and Planetary Change, 151, 36-48, 2017.

444

445   Hoerling, M., Eischeid, J., Perlwitz, J., Quan, X., Zhang, T., Pegion, P.: On the increased

446   frequency of Mediterranean drought, J. Clim., 25, 2146–2161, 2012.

447

448   Kaufman L., Rousseeuw P.J.: Finding Groups in Data: An Introduction to Cluster Analysis,

449   Hoboken, NJ: John Wiley & Sons, New York., 1990.

450

451   Lana, X., Martínez, M. D., Burgueño, A., Serra, C., Martín-Vide, J. and Gómez, L.:

452   Distributions of long dry spells in the iberian peninsula, years 1951–1990, Int. J. Climatol.,

453   26: 1999–2021. doi:10.1002/joc.1354, 2006.

454

455   Lobell, D. B. and Field, C. B.: Global scale climate–crop yield relationships and the impacts

456   of recent warming, Environ. Res. Lett., 2(1), 014002, doi:10.1088/1748-9326/2/1/014002,

457   2007.

458

459   Lobell, D. B., Hammer, G. L., Chenu, K., Zheng, B., Mclean, G. and Chapman, S. C.: The

460   shifting influence of drought and heat stress for crops in northeast Australia, Glob. Chang.

461   Biol., 21(11), 4115–4127, doi:10.1111/gcb.13022, 2015.

462

463   Lorenzo-Lacruz, J., Vicente-Serrano, S. M., González-Hidalgo, J. C., López-Moreno, J. I. and

464   Cortesi, N.: Hydrological drought response to meteorological drought in the Iberian

465   Peninsula, Clim. Res., 58(2), doi:10.3354/cr01177, 2013.

466

467   Naumann, G., Alfieri, L., Wyser, K., Mentaschi, L., Betts, R. A., Carrao, H., et al.: Global

468   changes in drought conditions under different levels of warming. Geophysical Research

469   Letters, 45, 3285–3296. https://doi.org/10.1002/2017GL076521, 2018.

470

471   Mann, H. B.: Nonparametric tests against trend, Econometrica, 13, 245– 259, 1945.

472

473 Manning, C., Widmann, M., Bevacqua, E., Van Loon, A.F., Maraun, D., and Vrac, M.: Soil

474 Moisture Drought in Europe: A Compound Event of Precipitation and Potential

475 Evapotranspiration on Multiple Time Scales, J. Hydrometeor., 19, 1255–1271, 20180

476

477 Mariotti, A., Dell'Aquila, A.: Decadal climate variability in the Mediterranean region: roles of

478 large-scale forcings and regional processes, Clim. Dyn. 38, 1129–1145, 2012.

479

480 McMahon, T. A., Peel, M. C., Lowe, L., Srikanthan, R., and McVicar, T. R.: Estimating

481 actual, potential, reference crop and pan evaporation using standard meteorological data: a

482 pragmatic synthesis, Hydrol. Earth Syst. Sci., 17, 1331-1363, https://doi.org/10.5194/hess-17-

483 1331-2013, 2013.

484

485 McVicar, T. R., Roderick, M. L., Donohue, R. J., Li, L.T., VanNiel, T. G., Thomas, A.,

486 Grieser, J., Jhajharia, D., Himri, Y., Ma-howald, N.M., Mescherskaya, A.V., Kruger, A.C.,

487 Rehman,S., Dinpashoh, Y.: Global review and synthesis of trends in observed terrestrial

488 near-surface wind speeds: Implications for evaporation, J. Hydrol., 416–417, 182–205,

489 2012a.

490

491 McVicar, T.R., Roderick, M. L., Donohue, R.J., Van Niel, T.G.: Less bluster ahead?

492 Overlooked ecohydrological implications of global trends of terrestrial near-surface wind

493 speeds, Ecohydrology, 5, 381–388, 2012b.

494

495 Mishra, A.K., Singh, V.P.: A review of drought concepts. J. Hydrol. 391, 202–216, 2010.

496

497 Mathugama, S. C., Peiris, T. S. G.: Critical Evaluation of Dry Spell Research, Int. J. Basic

498 Appl. Sci., 11, 153–160, 2011.

499

500 Mukherjee, S., Mishra, A. Trenberth, K. E.: Climate Change and Drought: a Perspective on

501 Drought Indices, Curr. Clim. Chang. Reports, 4(2), 145–163, doi:10.1007/s40641-018-0098-

502 x, 2018.

503

504 Páscoa, P., Gouveia, C. M., Russo, A. Trigo, R.M.: The role of drought on wheat yield

505 interannual variability in the Iberian Peninsula from 1929 to 2012, Int. J. Biometeorol., 61(3),

506 439–451, doi:10.1007/s00484-016-1224-x, 2017.

507

508 Peel, M. C., Finlayson, B. L., McMahon, T.A.: Updated world map of the Köppen-Geiger

509 climate classification, Hydrol. Earth Syst. Sci., 11, 1633-1644, https://doi.org/10.5194/hess-

510 11-1633-2007, 2007.

511

512 Quintana-Seguí, P., Barella-Ortiz, A., Regueiro-Sanfiz, S. and Miguez-Macho G.: The Utility

513 of Land-Surface Model Simulations to Provide Drought Information in a Water Management

514 Context Using Global and Local Forcing Datasets, Water Resources Management,

515 https://doi.org/10.1007/s11269-018-2160-9, 2019.

516

517 Raymond, F., Ullmann, A., Camberlin, P., Drobinski, P., Chateau Smith C., 2016. Extreme dry

518 spell detection and climatology over the Mediterranean Basin during the wet season,

519 Geophys. Res. Lett., 43, 7196–7204, doi:10.1002/2016GL069758.

520

521 Raymond F., Ullmann A., Camberlin P., Oueslati B., Drobinsky P.: Atmospheric conditions

522 and weather regimes associated with extreme winter dry spells over the Mediterranean basin,

523 Climate Dynamics 50, 4437-4453. doi:10.1007/s00382-017-3884-6, 2018.

524

525 Reiser H., Kutiel H.: Rainfall uncertainty in the Mediterranean: definition of the daily rainfall

526 threshold (DRT) and the rainy season length (RSL), Theor. Appl. Climatol. 97: 151–162,

527 2009.

528

529 Renard, B., Kochanek, K., Lang, M., Garavaglia, F., Paquet, E., Neppel, L., Najib, K.,

530 Carreau, J., Arnaud, P., Aubert, Y., Borchi, F., Soubeyroux, J.M., Jourdain S., Veysseire J.M.,

531 Sauquet E., Cipriani, T., Auffray, A.: Data-based comparison of frequency analysis methods: a

532 general framework, Water Resour Res 49:825–843, 2013.

533

534 Serra C., Lana X., Burgueno A., Martinez M.D.: Partial duration series distributions of the

535 European dry spell lengths for the second half of the twentieth century, Theorical and Applied

536 Climatology 123, 63-81, 2016.

537

538 Scholz, F. W., Stephens, M. A.: K-Sample Anderson–Darling Tests, Journal of the American

539 Statistical Association, 82:399, 918-924, 1987.

540

541 Serinaldi, F., Kilsby, C.: The importance of prewhitening in change point analysis under

542 persistence, Stochastic Environmental Research and Risk Assessment, 30(2), 763-777, 2016.

543

544 Serinaldi, F., Kilsby, C.G., Lombardo, F.: Untenable non-stationarity: An assessment

545 of the fitness for purpose of trend tests in hydrology, Adv. Water Resour., 111,

546 132–155, https://doi.org/10.1016/j.advwatres.2017.10.015, 2018.

547

548 Sivakumar, M.V.K.: Empirical analysis of dry spells for agricultural applications in West

549 Africa. J Climate 5:532–540, 1992.

550

551 Stagge, J.H., Kingston, L. M. Tallaksen, Hannah, D.M.: Observed drought indices show

552 increasing divergence across Europe. Sci. Rep., 7, 4045,https://doi.org/10.1038/s41598-017-

553 14283-2, 2017.

554

555 Tibshirani, R., G. Walther, Hastie T.: Estimating the number of clusters in a data set via the

556 gap statistic, Journal of the Royal Statistical Society: Series B., 63(2), 411–423, 2001.

557

558 Todorovic, M, Karic, B, Pereira, L.S.: Reference evapotranspiration estimate with limited

559 weather data across a range of Mediterranean climates, J. Hydrol., 481, 166–176, 2013.

560

561 Tramblay, Y., Hertig, E.: Modelling extreme dry spells in the Mediterranean region in

562 connection with atmospheric circulation. Atmospheric Research, 202, 40-48, 2018.

563

564 Trenberth, K. E., Dai, A., van der Schrier, G., Jones, P. D., Barichivich, J., Briffa, K. R. and

565 Sheffield, J.: Global warming and changes in drought, Nat. Clim. Chang., 4, 17,

566 http://dx.doi.org/10.1038/nclimate2067, 2014.

567

568 Vicente-Serrano, S. M., Beguería-Portugués, S.: Estimating extreme dry-spell risk in the

569 middle Ebro valley (northeastern Spain): a comparative analysis of partial duration series with

570 a general Pareto distribution and annual maxima series with a Gumbel distribution, Int. J.

571 Climatol., 23: 1103–1118, 2003.

572

573 Vicente-Serrano, S. M., C. Azorin-Molina, A. Sanchez-Lorenzo, J. Revuelto, E.

574 Morán-Tejeda, J. I. López-Moreno, Espejo F.: Sensitivity of reference evapotranspiration to

575   changes in meteorological parameters in Spain (1961–2011), Water Resour. Res., 50, 8458–

576   8480, doi: 10.1002/2014WR015427, 2014a.

577

578   Vicente-Serrano, S.M., Azorin-Molina, C., Sanchez-Lorenzo, A., Revuelto, J., López-Moreno,

579   J.I., González-Hidalgo, J.C., Espejo, F.: Reference evapotranspiration variability and trends in

580   Spain, 1961–2011. Global and Planetary Change, 121, 26–40, 2014b.

581

582   Vicente-Serrano, S.M., Lopez-Moreno, J.I., Beguería, S., Lorenzo-Lacruz, J., Sanchez-

583   Lorenzo, A., García-Ruiz, J.M., Azorin-Molina, C., Tejeda-Moran, E., Revuelto, J., Trigo, R.,

584   Coelho, F., Espejo, F.: Evidence of increasing drought severity caused by temperature rise in

585   Southern Europe. Environ. Res. Lett., 9 (4), 044001. http://dx.doi.org/10.1088/1748-

586   9326/9/4/044001, 2014c.

587

588   Vicente-Serrano, S.M., Van der Schrier, G., Beguería, S., Azorin-Molina, C., Lopez-Moreno,

589   J.I.: Contribution of precipitation and reference evapotranspiration to drought indices under

590   different climates, J. Hydrol., 526, 42-54, 2015.

591

592   Vicente-Serrano, S.M., D.G. Miralles, F. Domínguez-Castro, C. Azorin-Molina, A. El

593   Kenawy, T.R. McVicar, M. Tomás-Burguera, S. Beguería, M. Maneta, Peña-Gallardo, M.:

594   Global Assessment of the Standardized Evapotranspiration Deficit Index (SEDI) for Drought

595   Analysis and Monitoring, J. Climate, 31, 5371–5393, 2018.

596

597   Ward, J. H.: Hierarchical Grouping to Optimize an Objective Function, Journal of the

598   American Statistical Association, 58 (301): 236–244, 1963.

599

600   Wilks, D.S.: The Stippling Shows Statistically Significant Grid Points: How Research

601   Results are Routinely Overstated and Overinterpreted, and What to Do about

602   It, Bulletin of the American Meteorological Society, 97(12), 2263-2273, 2016.

603

604   Zkhiri, W., Tramblay, Y., Hanich, L., Berjamy, B.: Regional flood frequency analysis in the

605   high-atlas mountainous catchments of Morocco, Natural Hazards, 86(2), 953-967.

606   http://dx.doi.org/10.1007/s11069-016-2723-0, 2017.

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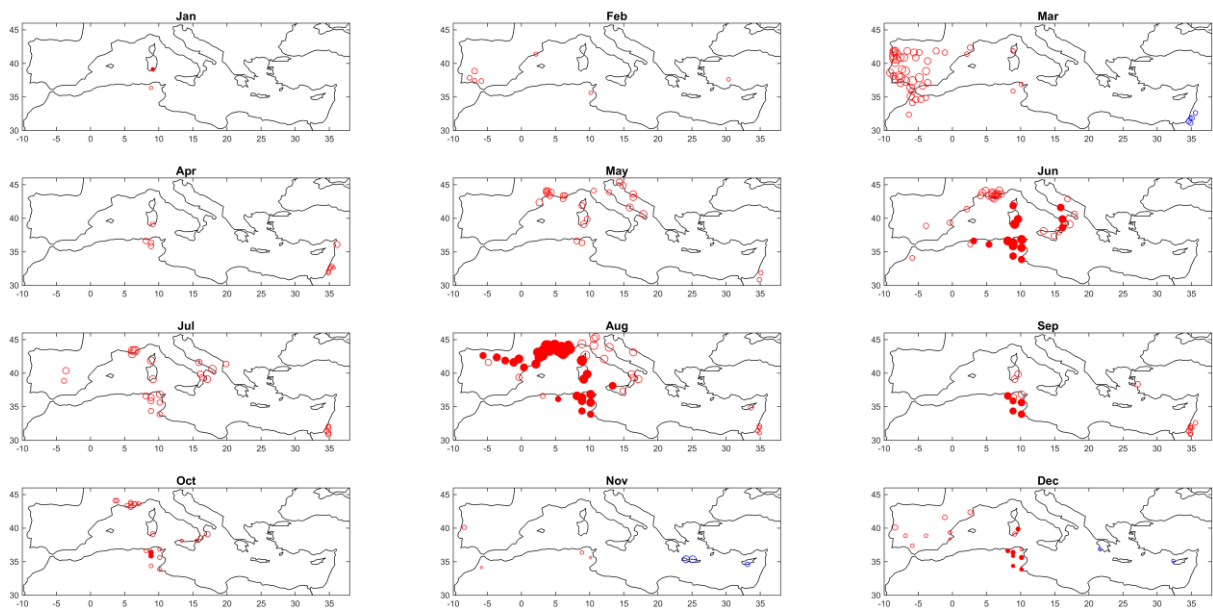
Figure 1: Significant trends (5% level) in monthly ET_0. The size of the circles indicate the magnitude of the trends (red = increasing, blue = decreasing) and the filled circles denote regional significant trends

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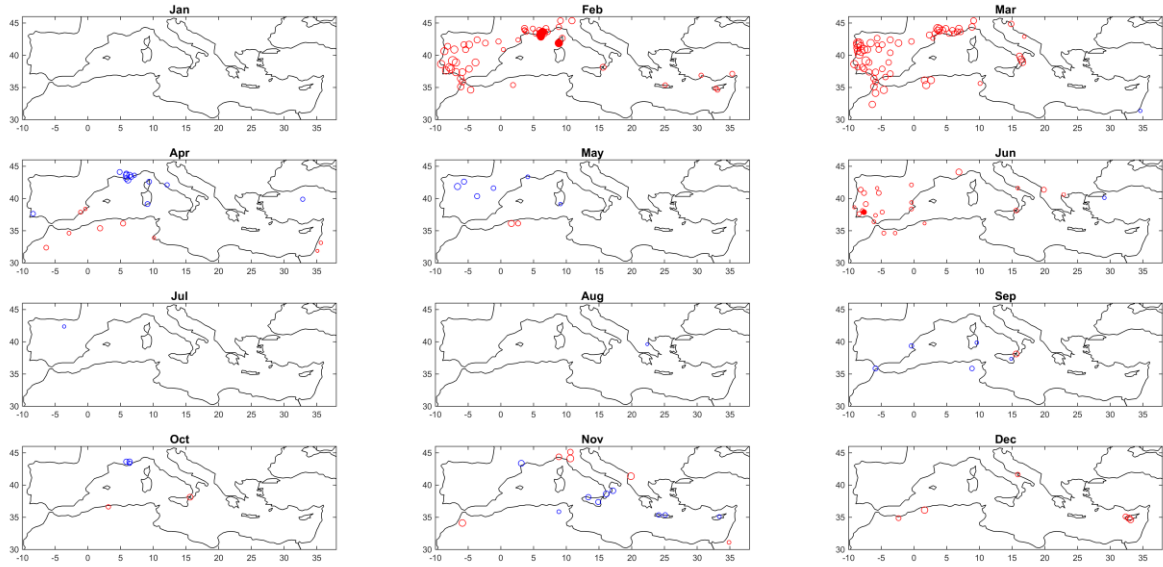
630   Figure 2: Significant trends (5% level) in the frequency of dry days when considering the

631   1mm threshold to define a dry day. Same as Fig.1 for the display

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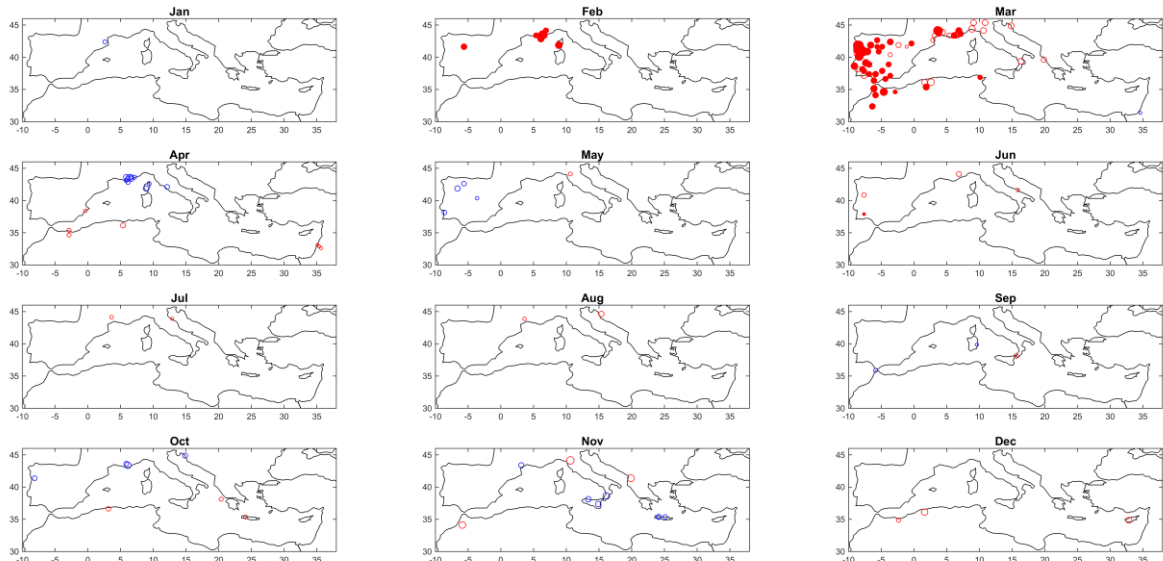
635   Figure 3: Significant trends (5% level) in the frequency of dry days, considering that P-ET$_0$ =

636   0 is a dry day. Same as Fig.1 for the display

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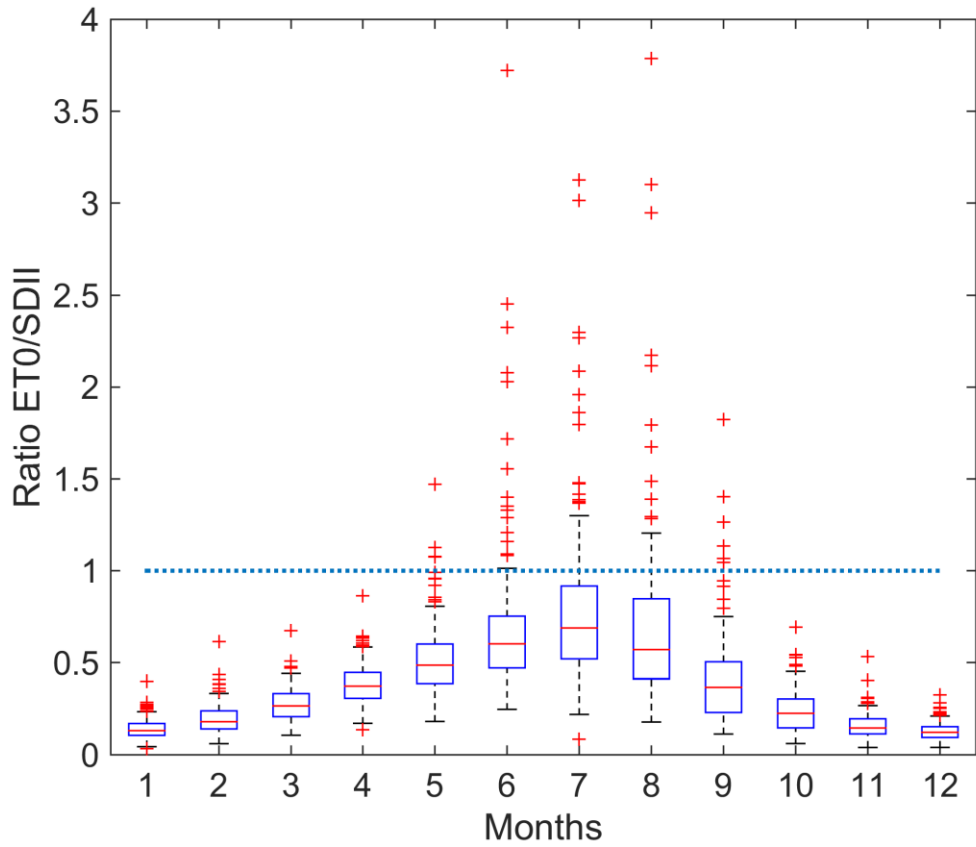
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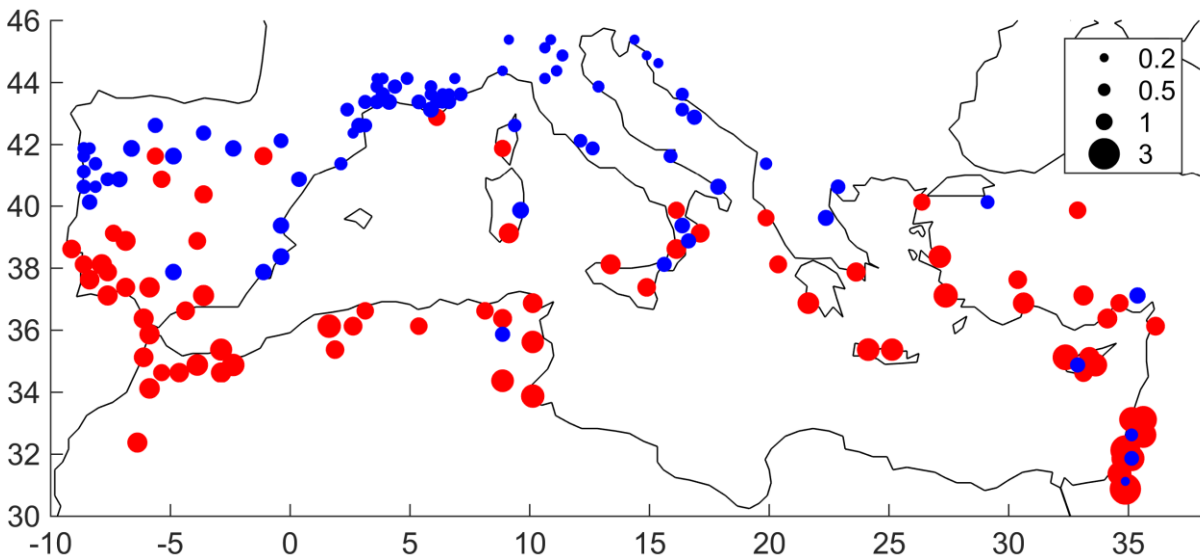
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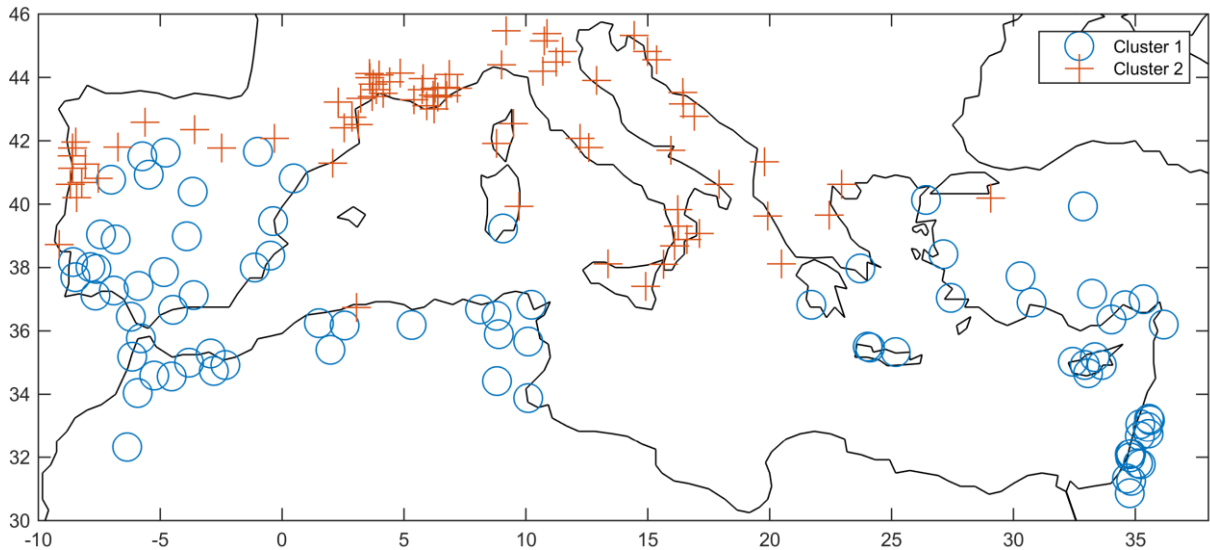
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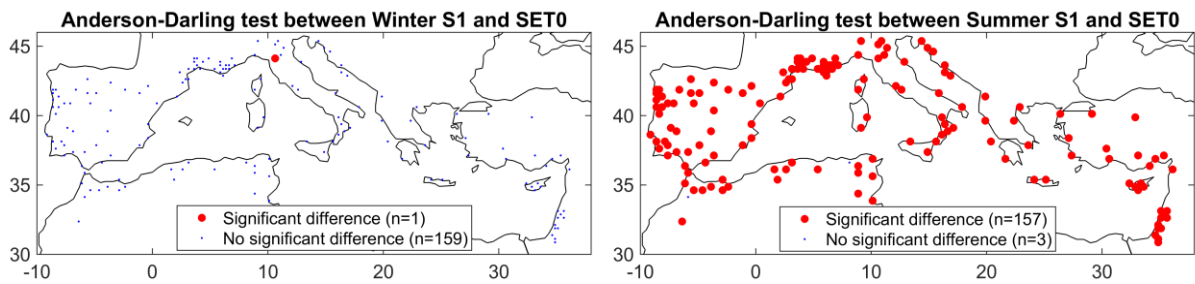
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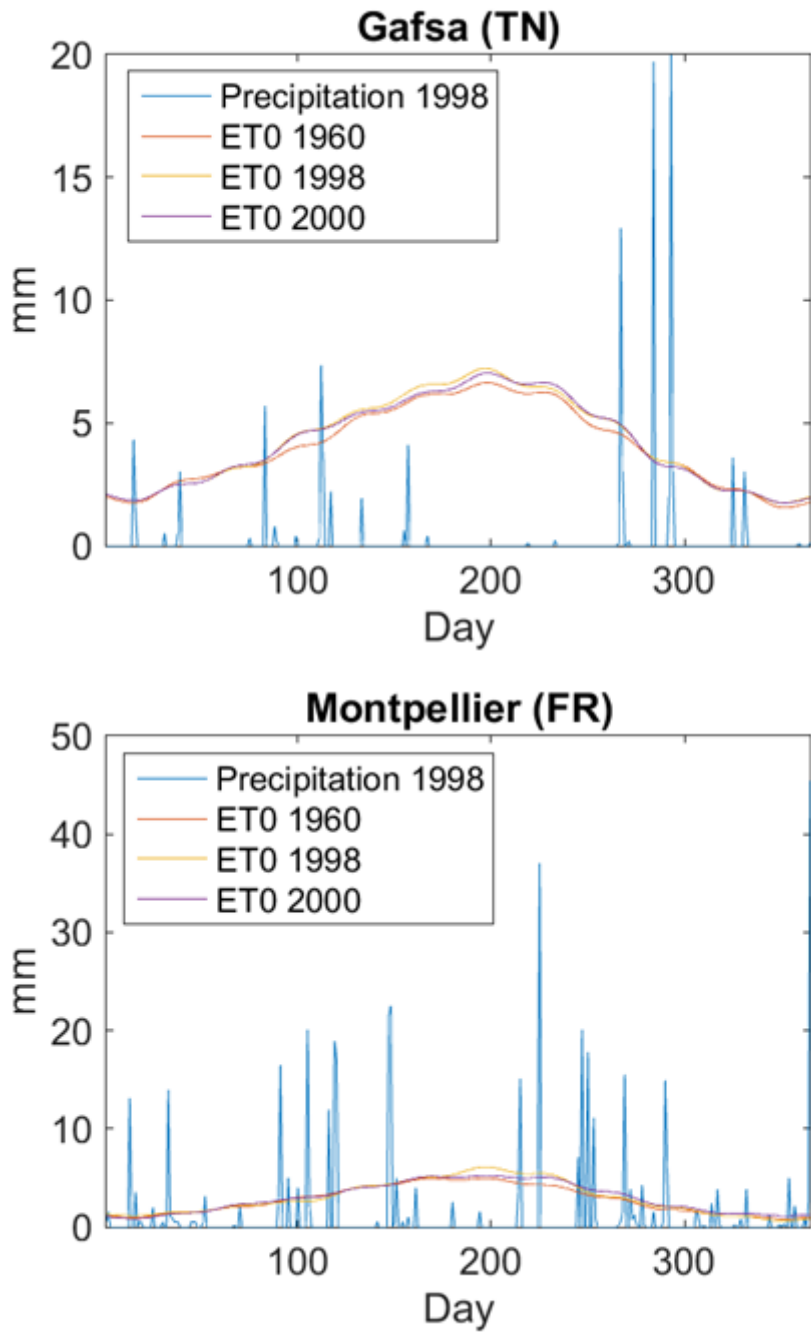
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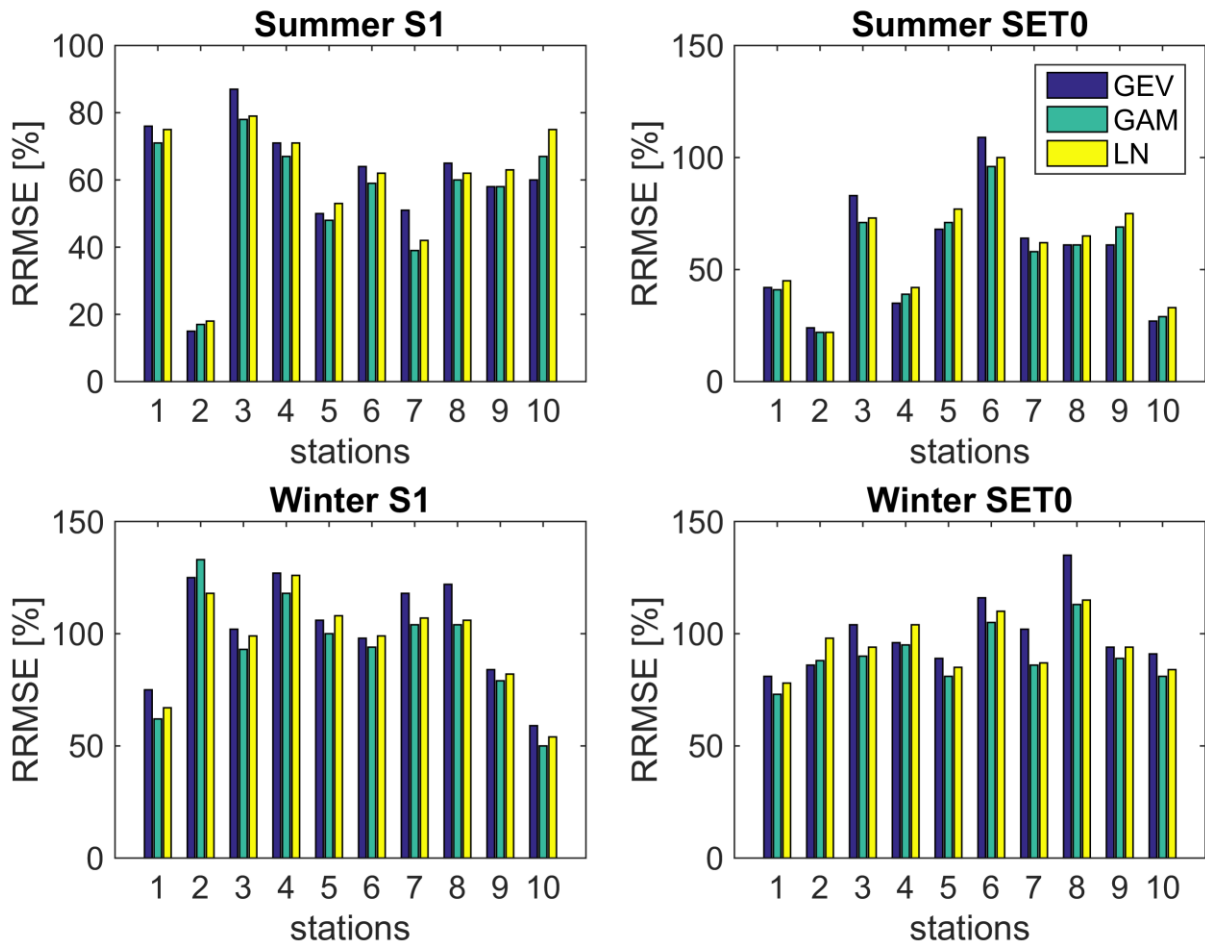
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674   Capo Bellavista (IT))

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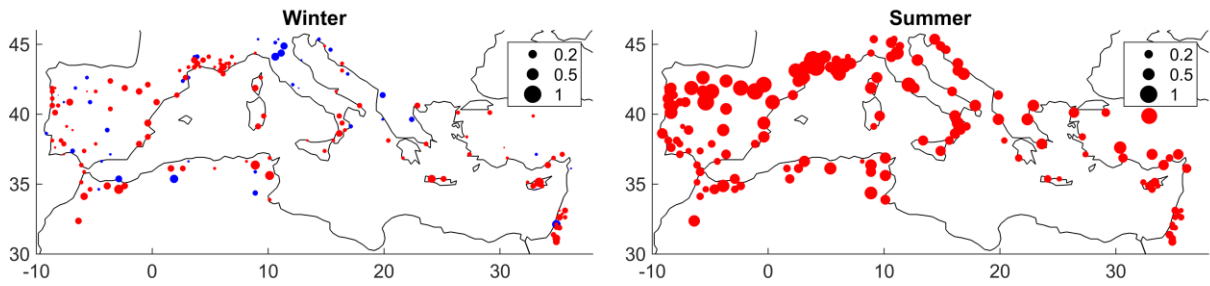
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