REFERENCE EVAPOTRANSPIRATION VARIABILITY AND TRENDS IN SPAIN, 1961–2011

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Abstract. In this study we analyzed the spatial distribution, temporal variability and trends in reference evapotranspiration (ET_0) in Spain from 1961 to 2011. Twelve methods were analyzed to quantify ET_0 from quality controlled and homogeneous series of various meteorological variables measured at 46 meteorological stations. Some of the models used are temperature based (e.g., Thornthwaite, Hargreaves, Linacre), whereas others are more complex and require more meteorological variables for calculation (e.g., Priestley–Taylor, Papadakis, FAO–Blaney–Criddle). The Penman–Monteith equation was used as a reference to quantify ET_0, and for comparison amongst the other methods applied in the study. No major differences in the spatial distribution of the average ET_0 was evident among the various methods. At annual and seasonal scales some of the ET_0 methods requiring only temperature data for calculation provided better results than more complex methods requiring more variables. Among them the Hargreaves (HG) equation provided the best results, at both the annual and seasonal scales. The analysis of the temporal variability and trends in the magnitude of ET_0 indicated that all methods show a marked increase in ET_0 at the seasonal and annual time scales. Nevertheless, results obtained suggested substantial uncertainties among the methods assessed to determine ET_0 changes, due to differences in temporal variability of the resulting time series, but mainly for the differences in the magnitude of change of ET_0 and its spatial distribution. This suggests that ET_0 trends obtained by means of methods that only require temperature data for ET_0 calculations should be evaluated carefully under the current global warming scenario.

Key words: reference evapotranspiration, Penman-Monteith, climate change, global warming, Mediterranean region, drought

1. Introduction

Evapotranspiration (ET) is an essential component of both climate and hydrological cycles, and has significant agricultural, ecological and hydrological implications. ET uses approximately three fifths of the available annual solar radiation globally received at the Earth's surface (Wang and Dickinson, 2012; Wild et al., 2013). In addition to the energy balance, ET is also a major component of the water cycle, as it accounts for approximately two thirds of the precipitation falling on land (Baumgarter and Reichel, 1975). ET is important in several atmospheric processes, as it determines the supply of water to the atmosphere from the oceans and terrestrial areas. It affects the magnitude and spatial distribution of global temperature and pressure fields (Shukla and Mintz, 1982), and it may affect the occurrence of heat waves (Seneviratne et al., 2006) and precipitation processes (Zveryaev and Allan, 2010).

The concepts of actual evaporation (ET_a) and reference evaporation (ET_0) are defined as follows: the ET_a is the quantity of water that is transferred as water vapour to the atmosphere from an evaporating surface (Wiesner, 1970) under real conditions (e.g. water availability, vegetation type, physiological mechanisms, climate), whereas ET_0 represents the atmospheric evaporative demand of a reference surface (generally a grass crop having specific characteristics), and it is assumed that water supply from the land is unlimited (Allen et al., 1998). The only factors affecting ET_0 are climatic parameters, given some reference crop and associated parameters, e.g., albedo and vegetation height. Consequently, ET_0 is a climatic parameter and can be computed from weather data. ET_0 expresses the evaporating power of the atmosphere at a specific location and time of the year and it allows for spatial and temporal comparisons, independently of different land cover types and temporal coverage changes (Katerji and Rana, 2011). ET_a will be less than or equal to ET_0, but never greater. Equally, ET_0 cannot be measured directly using meteorological instruments, as it depends on a number of meteorological variables (net radiation, air temperature, surface pressure, wind speed and relative humidity).

In recent decades paradoxical processes have been detected related to the evolution of the atmospheric evaporative demand (AED). Despite the observed recent climate warming, a general decrease in pan evaporation has been reported (Peterson et al., 1995; Roderick and Farquar, 2004), which could be explained by decreased solar radiation (e.g., Matsoukas et al., 2011; Roderick and Farquar, 2002) and/or wind speed decrease (McVicar et al., 2012). Nevertheless, Brutsaert and Parlange (1998) offered theoretical explanations why a trend of decrease in pan evaporation is not necessarily an indication of decreasing ET_0 and ET_a. Moreover, recent studies have suggested major limitations in the use of pan ET measurements to assess current AED trends (Fu et al., 2009; Abtew et al., 2011).

ET_0 is currently considered to be a reliable parameter for assessing long-term trends of the AED (Katerji and Rana, 2011), as it only depends on the meteorological conditions, has a clear physical meaning, and the meteorological variables necessary to calculate ET_0 are available worldwide and have been measured for many years. Although ET_0 may not correspond to accurate ET_a estimates, which depend largely on water availability, soil characteristics and vegetation properties, assessing ET_0 trends is of great interest because it is a measure of aridity conditions and crop requirements, and has major implications for land desertification and food production.

Various studies have analyzed global ET_0 trends based on interpolated gridded datasets (e.g. Dai, 2010; Sheffield et al., 2012) and reanalysis data (Matsoukas et al., 2011), but the results have differed markedly, depending on the datasets and methods used to estimate ET_0. Regional and local studies based on observational datasets have shown a variety of results in different regions of the world. In some cases the trends in ET_0 have been negative, including for the Yangtze River (Xu et al., 2006), the Yellow River (Ma et al., 2012) and the Tibetan plateau (Zhang et al., 2007) in China. Other studies have shown positive trends in ET_0, including in central India (Darshana et al., 2012), Iran (Kousari and Ahani, 2012; Tabari et al., 2012) and Florida (Abtew et al., 2011). Moreover, in some areas (e.g. Australia) there has been large spatial variability in the evolution of ET_0 during recent decades (Donohue et al., 2010).

One of the most important areas worldwide in relation to the impact of climate change processes is the Mediterranean region, because of its high spatial and temporal variability in precipitation (Lionello, 2012). Various empirical studies have shown that water availability has decreased in this area in recent decades (García-Ruiz et al., 2011). Hypotheses to explain this decrease are related to land cover changes and human management, but also climate change processes to which ET is strongly connected.

Although there is a number of agronomic studies estimating the atmospheric evaporative demand (AED) with the purpose of improving the selection of more appropriate crops and irrigation practices (i.e., water saving) in the Mediterranean region, some of them using evaporation observations from lysimeters for validation (e.g., Steduto et al., 2003; Lorite et al., 2012), there are very few studies that have analysed temporal variability and trends of ET_0 in the last decades. Among these, Espadafor et al. (2011) analyzed ET_0 trends from 1960 to 2005 at eight stations in southern Spain, and showed a general increase in ET_0. Papaioaunou et al. (2011) showed a general increase in ET_0 in Greece since the early 1980s, mainly driven by the evolution of global radiation, whereas Platineau et al. (2012) used the same calculation method to show a general increase in ET_0 in Romania, resulting from an increase in temperature. Palumbo et al. (2011) analyzed the trends in ET_0 in southern Italy; they found an increase of 14 mm/decade between 1957 and 2008, which has increased the water requirements of the main cultivated crops by 7 mm/decade. Vergni and Todisco (2011) analyzed the evolution of ET_0 in central Italy, and found a dominant positive trend between 1951 and 2008. In the studies noted above, ET_0 was calculated using a variety of formulae, which makes it difficult to compare the magnitudes of change reported, and to assess the robustness of the observed trends. Moreover, some of the studies are applying empirical methods to estimate ET_0 only using temperature records. Limitations of the use of this type of formulation are obvious in climate change studies since an increase in temperature will translate to increased AED (Roderick et al., 2009), when this is a synthesis of two (radiative and aerodynamic) components not only determined by the evolution of temperature but also of changes in solar radiation, wind speed and relative humidity (Penman, 1948). For these reasons, studies that compare the reliability of temperature-based methods and robust physical estimates based on both radiative and aerodynamic components to estimate the AED evolution are high priority in this region.

In this study we analyzed trends in ET_0 in Spain from 1960 to 2011. Some of the methods for calculating ET_0 were based on temperature records, while others involved several meteorological variables (e.g. relative humidity, wind speed, radiation). The objectives were: i) to compare average estimates of ET_0 obtained using the various methods; ii) to determine the magnitude and spatial patterns of ET_0 variability; and iii) to evaluate the reliability of the different methods for assessing ET_0 trends. Overall, this is the first study covering the complete Spanish territory and, to our knowledge, including a complete comparison of methods based on quality controlled and homogenised datasets of different climate variables across the Mediterranean basin.

2. Methods

2.1. ET_0 methods

The International Commission for Irrigation (ICID), the Food and Agriculture Organization of the United Nations (FAO), and the American Society of Civil Engineers (ASCE) have adopted the Penman-Monteith (PM) method (Allen et al., 1998) as the standard method for computing ET_0 from climate data. The PM method is widely used because it is predominantly a physically-based approach that can be used globally, and has been widely tested using lysimeter data obtained under a broad range of climate conditions (e.g. Itenfisu et al., 2000).

The main drawback of the PM method is the relatively large amount of data involved, as it requires data on solar radiation, temperature, wind speed and relative humidity. For this reason, numerous other methods have been developed to calculate ET_0 using less data. In this study we used the PM method as a reference, and 11 other methods commonly used worldwide that require much less information. Some of them are recommended when there is low availability of data (e.g., Hargreaves; Allen et al., 1998) whereas others are of high use for agricultural purposes and irrigation management (e.g., Blaney-Criddle, Priestley-Taylor). They do not cover the complete methods existing to obtain ET0, but they are a representative sample and it included the most used methods. We distinguished between the temperature-based methods and those requiring additional meteorological variables.

2.1.1. The reference FAO-56 Penman-Monteith (PM; Allen et al., 1998) equation

The FAO PM method was developed by defining the reference crop as a hypothetical crop with an assumed height of 0.12 m, a surface resistance of 70 s m^{-1} and an albedo of 0.23. This closely approximates the evaporation expected from an extensive surface of actively growing and adequately watered green grass of uniform height (Allen et al., 1998), and is defined by the equation:

$$ET_0 = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T + 273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)}$$

where ET_0 is the reference evapotranspiration (mm day^{-1}), R_n is the net radiation at the crop surface (MJm^{-2} day^{-1}), G is the soil heat flux density (MJ m^{-2} day^{-1}), T is the mean air temperature at a height of 2 m (°C), u_2 is the wind speed at 2 m height (m s^{-1}), e_s is the saturation vapor pressure (kPa), e_a is the actual vapor pressure (kPa), e_s-e_a is the saturation vapor pressure deficit (kPa), Δ is the slope vapor pressure curve (dependent on temperature) (kPa $°C^{-1}$) and γ is the psychrometric constant (kPa $°C^{-1}$).

2.1.2. Methods based on temperature data

2.1.2.1. The Thornthwaite equation (TH; Thornthwaite, 1948)

This is one of the simplest and most widely used approaches to calculation of ET_0, and only requires monthly mean temperature. The ET_0 (mm month^{-1}) is obtained using the equation:

$$ET_0 = 16K\left(\frac{10T}{I}\right)^m$$

where I is a heat index (calculated as the sum of 12 monthly index values i, which is derived from mean monthly temperature as $i = \left(\frac{T}{5}\right)^{1.514}$), s a coefficient depending on I ($m = 6.75E^{-7}I^3 - 7.71E^{-5}I^2 + 1.79E^{-2}I + 0.492$), and K is a correction coefficient computed as a function of the latitude and month ($K = \left(\frac{N}{12}\right)\left(\frac{NDM}{30}\right)$), where NDM is the number of days of the month and N is the total daytime hours for the month.

2.1.2.2. Blaney-Criddle equation (BC; Blaney and Criddle, 1950)

Blaney and Criddle (1950) developed a temperature-based equation for agricultural purposes. In this method ET_0 (mm day^{-1}) is calculated using the equation:

$$ET_0 = p(0.46T + 8.13)K$$

where p is the percentage of total daytime hours in the month in relation to the total daytime hours in the year, and K is a coefficient that ranges from 0.15 and 1.44 depending on the cultivation type and the region. For this study an average value of 0.85 was selected, following Xu and Singh (2002).

2.1.2.3. The Linacre equation (LIN; Linacre, 1977)

Linacre simplified the Penman equation in relation to a vegetation surface that has an albedo of 25% and is well provided with water. In this method ET_0 (mm day^{-1}) is calculated using the equation:

$$ET_0 = \frac{500Tm/(100 - A) + 15(0.0023h + 0.37T + 0.53R + 0.35R_{an} - 10.9)}{80 - T}$$

where $Tm = T + 0.006h$, h is the elevation above sea level (m), A is the latitude in degrees, R is the difference between the maximum and minimum temperatures (monthly averages; °C) and R_{an} is the difference between the average mean temperature of the warmest and coldest months.

2.1.2.4. The Hargreaves equation (HG; Hargreaves and Samani, 1985)

This method only requires information on the maximum and minimum temperatures, and extraterrestrial solar radiation. The ET_0 (mm day^{-1}) is calculated using the equation:

$$ET_0 = 0.0023 R_a R^{0.5}(T + 17.8)$$

where R is defined in the Linacre equation, and R_a is the extraterrestrial solar radiation (mm day^{-1})

2.1.2.5. The Kharrufa equation (KH, Kharrufa, 1985)

Kharrufa (1985) derived an equation through correlation of ET_0/p and T. In this method ET_0 (mm month^{-1}) is calculated using the equation:

$$ET_0 = 0.34pT^{1.3}$$

2.1.2.6. The modified Hargreaves equation (HG-PP; Droogers and Allen, 2002)

Droogers and Allen (2002) modified the original Hargreaves equation by including a rainfall term, on the assumption that monthly precipitation can represent relative levels of humidity. The ET_0 (mm day^{-1}) is calculated using the equation:

$$ET_0 = 0.0013 R_a (T + 17.0)(R - 0.0123P)^{0.76}$$

where P is the monthly total precipitation in mm.

2.1.3. Methods requiring more meteorological variables

2.1.3.1. The Turc equation (T; Turc, 1955)

Turc (1955) proposed an empirical relationship in which ET_0 is calculated using the relative humidity, the average temperature, and the solar radiation. ET_0 (mm month^{-1}) is function of the average relative humidity. If the monthly average relative humidity is > 50%, $ET_0 = 0.40\ [T/(T+15)]\ (23.884R_S + 50)$. If the monthly average relative humidity is < 50%, $ET_0 = 0.40\ [T/(T+15)]\ (23.884R_S + 50)[1+(50–RH)/70]$. In these equations R_s is the solar radiation (MJ m^{-2} day^{-1}) and *RH* is the mean relative humidity (%).

2.1.3.2. The Papadakis equation (P; Papadakis, 1966)

Papadakis used saturation vapor pressure corresponding to monthly temperatures to estimate ET_0 (mm $month^{-1}$) using the equation:

$$ET_0 = 5.625\ [e_s(T_{max}) - e(T_d)]$$

where $e_s(T_{max})$ is the saturation water pressure corresponding to average maximum temperature (kPa), and $e(T_d)$ is the saturation water pressure corresponding to the dewpoint temperature (kPa).

2.1.3.3. The Priestley-Taylor equation (PT; Priestley and Taylor, 1972)

Priestley and Taylor (1972) used an equation derived from the combination method of Penman, in which the aerodynamic term is replaced by a coefficient (α). The ET_0 (mm day^{-1}) is calculated using the equation:

$$ET_0 = \alpha \left[\frac{\Delta}{\Delta + \gamma}\right] R_n$$

where a standard value for α (1.26) is used.

2.1.3.4. The FAO-Blaney-Criddle equation (FAO-BC; Doorenbos and Pruitt, 1977)

Doorenbos and Pruitt (1975) made an important modification of the Blaney-Criddle method, which includes the influence of radiation, wind speed and relative humidity. The equation is derived from a calibration using lysimeter measurements. The ET_0 (mm day^{-1}) is calculated (Frevert et al., 1981) using the equations:

$$ET_0 = a_b + b_b f,$$

$f = p(0.46T + 8.13)$,

$a_b = 0.0043\ RH_{min} - \frac{n}{N} - 1.41$ and

$$b_b = 0.81917 - 0.0040922\ RH_{min} + 1.0705\frac{n}{N} + 0.065649\ u_2 - 0.0059684\ RH_{min}\frac{n}{N} - 0.000597\ RH_{min}u_2$$

where RH_{min} is the minimum relative humidity (monthly average) (%) and *n* is the observed number of sun hours (monthly average; hours).

2.1.3.5. The Radiation method (R; Doorenbos and Pruitt, 1977)

This is similar to the Priestley-Taylor method, but based on surface solar radiation rather than net radiation. The equation proposed by Doorenbos and Pruit (1977) is:

$$ET_0 = a + b\left[\frac{\Delta}{\Delta + \gamma}\right]R_s$$

where R_s is the solar radiation (mm day^{-1}). The coefficients *a* and *b* can be obtained according to Frevert et al. (1982), where a = –0.3 and $b = 1.0656 - 0.0012795RH + 0.044953u_2 - 0.00020033RHu_2 - 0.000031508RH^2 - 0.0011026u_2^2$.

2.2. Datasets

In applying the various ET_0 equations we used data for variables measured at numerous meteorological stations. Allen et al. (1998; Chapter 3 of the FAO-56 publication) detailed the variables required to calculate ET_0 using the PM equation. These include: i) monthly average maximum and minimum air temperatures (°C); ii) monthly average actual vapor pressure (e_a; kPa); iii) average monthly net radiation (MJ m^{-2} day^{-1}); and iv) monthly average wind speed (m s^{-1}) measured 2 m above ground level. Among these e_a is not measured at meteorological stations, but can be calculated from relative humidity and temperature (Allen et al., 1998). In addition, the monthly average net solar radiation is not commonly available from meteorological stations, and generally few and only short time series of surface solar radiation are available in Spain (Sanchez-Lorenzo et al., 2013). However, this parameter is commonly estimated from the monthly averages of daily sunshine hours, measured using sunshine duration recorders (e.g. the Campbell-Stokes recorder) given close agreement between sunshine duration and surface shortwave radiation (Long et al., 2010). Figure 1 provides an example showing the relationship between monthly average global solar radiation (Sanchez-Lorenzo et al., 2013) and daily average duration of sunshine hours (Sanchez-Lorenzo et al., 2007) for seven stations in Spain from 1980 to 2010. Close agreement between the two variables is evident (Pearson's $r = 0.89$) and provided a high degree of reliability in determining R_n and R_s from time series of the duration of daily sunshine.

The necessary parameters: soil heat flux density (G), extraterrestrial radiation (R_a), net and surface solar radiation (R_n and R_s, respectively), the psychrometric constant, the mean saturation pressure (e_s), the slope of the saturation vapor pressure curve (Δ) and wind speed at the standard height of 2 m above ground, were obtained according to Allen et al. (1998) using maximum and minimum temperature, sunshine duration, wind speed, relative humidity, and surface atmospheric pressure. Precipitation was also included to enable application of the modified Hargreaves equation.

Only the first order meteorological stations (approximately 100) of the weather observation network of the Spanish State Meteorological Agency (AEMET) measure all the variables necessary to calculate ET_0 using the equations described above, but these contain all the historical records needed to determine recent trends. Using these records, Sanchez-Lorenzo et al. (2007) created a homogeneous dataset of sunshine duration for the Iberian Peninsula since the beginning of the 20th century. González-Hidalgo et al. (2011) developed a dense and homogeneous precipitation dataset for Spain. Vicente-Serrano et al. (2014) obtained 50 homogeneous time series of relative humidity in Spain. To obtain specific humidity they also obtained quality controlled and homogeneous series of maximum and minimum temperature and surface pressure. Finally, Azorin-Molina et al. (2013) have recently developed a homogeneous dataset of wind speed at 10 m height for the entire Iberian Peninsula and the Balearic Islands. We used these datasets, updated to 2011, as they are the most reliable corresponding to the various meteorological variables needed to calculate ET_0 series for Spain using the 12 equations noted above.

A total of 46 stations are available for continental Spain and the city of Melilla, in northern Africa (Figure 2). From the homogeneous series of temperature, precipitation, pressure, wind speed, sunshine duration and relative humidity, we computed a single regional series for mainland Spain following Jones and Hulme (1996).

2.3. Validation statistics and trend analysis

Using the time series of ET_0 derived from the 12 ET_0 equations we determined the seasonal (winter: December–February; spring: March–May; summer (June–August; autumn; September–November) and annual ET_0 averages. As the PM equation provided the most reliable estimates of ET_0, we used the PM values as a reference against which to compare the spatial and temporal estimates obtained using the other methods, despite the limitations associated with the large number of variables involved in its calculation. For this comparison we used various error/accuracy statistics (Willmott, 1982) including: the coefficient of determination (R^2); the mean bias error (MBE); the mean absolute difference (MAD), which is a measure of the average difference of the ET_0 estimations; and the agreement index (D; Willmott, 1981). D is a standardized measure of the degree of model prediction error and varies between 0 and 1. A value of 1 indicates a perfect match, and 0 indicates no agreement at all. It overcomes some disadvantages of the abovementioned measures since it scales with the magnitude of variables and enables spatial and seasonal comparison of ET_0 values, independent of differences in the ET_0 magnitude and range for each month. Table 1 provides the formulations of error measures used in this study.

To analyze changes in ET_0 we used the nonparametric coefficient (Mann-Kendall tau) that measure the degree to which a trend is consistently increasing or decreasing. To assess the magnitude of change we used a regression analysis between the series of time (independent variable) and the ET_0

series (dependent variable). The slope of the regression line indicated the change (ET_0 change per year), with greater slope values indicating greater change.

3. Results

3.1. Average values

The average annual and seasonal ET_0 values show variability among the 46 stations independent of season, but differences are also evident among the ET_0 methods (Fig. 3). For example, the HG-PP, LIN, KH, FAO-BC and P equations indicated greater ET_0 spatial variability relative to the other methods. Seasonal differences were apparent, with the LIN and BC methods showing the greatest overestimation of ET_0 for winter and autumn compared with the PM method, whereas for spring and summer the FAO-BC equation showed the greatest overestimates. The THO and PT methods tended to underestimate ET_0 during the various seasons. At the annual scale the HG and HG-PP methods tended to provide the most similar estimates of ET_0 to those obtained using the PM method.

Among the various methods the spatial patterns of the annual ET_0 average values showed clear differences along a north–south gradient (Fig. 4). Although the spatial patterns were similar (higher values in the south and southeast of the Iberian Peninsula, and lower values in the north) the magnitudes varied. Values (spatially and in magnitude) based on the HG method were in agreement with those of the PM method. In Appendices, the various error/accuracy statistics used to compare the ET_0 averages based on the PM and other equations are showed in Table A.1.

3.2. Temporal variability

Some ET_0 methods (THO, BC, PT and R) were characterized by low temporal variability and low relative differences in ET_0 among years (Fig. 5). In contrast, other methods (LIN, FAO-BC and P) showed marked interannual variability. The PM method provided intermediate temporal variability that was similar to that observed for the HG, HG-PP and KH methods. Independent of the method there was a large increase in ET_0 at the annual scale over continental Spain. The HG method showed the closest agreement with the PM ET_0 in terms of temporal evolution, and particularly following 1990 was very similar in both the temporal variability of ET_0 and its magnitude. The differences in ET_0 variability among the methods may be important at the seasonal scale (Fig. 6). In winter the low temporal variability in ET_0 based on the PM method was similar to that observed using the R and PT methods. Other methods including FAO-BC and LIN showed marked interannual variability, and the BC method provided the highest ET_0 values. There was no general increase in ET_0 during winter, independent of the method used. Interannual variability in spring was much greater, with the highest ET_0 values being associated with the FAO-BC method. The evolution of ET_0 as measured by the PM method was very similar to that found for the HG, HG-PP and PT methods. As for winter the THO method produced the lowest values and showed much less temporal variability than the other methods. The highest ET_0 rates were found in summer, although some methods (FAO-BC, LIN, HG-PP and P) showed much greater temporal variability than the PM method. In summer the PM method shows the greatest increase in ET_0. From 1960 to 1990 the HG method showed similar values to those derived from the PM method, but produced lower values for the period 1990–2011. For autumn, most of the methods showed higher ET_0 values than the PM method. Thus, in autumn a general increase in ET_0 was found using most of the equations, but was much less than was observed for summer.

Based on the coefficients of determination obtained for each of the 46 stations, the methods that require more variables in their calculation tended to show higher R^2 values than the temperature-based models (Fig. 7). This pattern was observed at both the seasonal and annual scales. Thus, the FAO-BC and R methods showed very high coefficients and small differences among observatories in spring and summer. The temperature-based methods tended to show greater variability in the R^2 coefficients among stations than did the methods requiring a greater number of meteorological variables; the exception was the PT method, which also showed marked differences among observatories. There were no clear spatial patterns in the spatial distribution of R^2 values obtained from the annual series, but the methods that provided the highest average R^2 values tended to show higher coefficients for most observatories (Figure 8). The same pattern was identified for those methods showing low coefficient values. Exceptions were the HG and HG-PP methods, for which higher coefficient values were found for central Spain relative to observatories located in southern and northern regions. Nevertheless, although the methods based on a greater number of meteorological variables tended to be more accurate in reproducing the temporal variability in ET_0 derived using the PM method, they did not always accurately reproduce the magnitude of ET_0. For this reason the D index (see Section 2.3) provided a more reliable comparison among methods. A box plot of D values, enabling comparison of the seasonal and annual PM ET_0 series with series obtained using the other methods showed no clear differences between the methods based on temperature alone and those involving other meteorological variables (Fig. 9). For winter the D values tended to be low for the various observatories, and consequently there was no method better able to reproduce both the temporal variability and magnitude of the ET_0 values obtained using the PM method. In contrast, for spring and summer the temperature-based methods tended to produce higher D values (with the exception of the TH method) than the other methods. Thus, for both these seasons the HG method produced slightly higher D values than the other methods. At the annual scale the HG and T methods also produced higher D values. Therefore, in terms of the efficiency of reproducing the ET_0 variability found using the PM equation, the number of variables needed in the calculation of ET_0 was not the determining factor. Thus, with the exception of the TH method, simple equations including the KH method, which only depends on mean temperature, provided better agreement coefficients than other more complex methods (e.g. the FAO-BC and R methods). At the annual scale there were no marked spatial patterns in the distribution of D values (Fig. 10), suggesting there were no regions for which one method better reproduced the temporal variability in ET_0 based on the PM method.

3.3. Long-term trends

In Appendices, the number of observatories with positive and negative trends in annual and seasonal ET_0 between 1961 and 2011 is showed in Table A.2. Although the various methods were in general agreement in indicating a dominant positive trend in ET_0 in Spain, the magnitude of change differed markedly among the methods. Analysis based on data for each of the meteorological observatories indicated marked differences between the magnitude of change based on the PM method and the other methods used in the study (Fig. 11). It was evident that, relative to the results obtained using the PM method, methods requiring additional variables were not clearly advantageous compared with methods based on temperature alone for assessing ET_0 trends. Thus, the box plots show that the method showing the best agreement with the PM method in one season could have the least agreement in a different season (e.g. the TH method for summer and autumn).

The magnitude of change based on the PM method did not show clear spatial patterns: with the exception of some observatories in the southeast, the main increase in ET_0 occurred in the northeast area of the Iberian Peninsula (Fig. 12). Changes in magnitude were much lower based on the TH, HG, BC, PT, T and R methods. The HG-PP method indicated a similar pattern to the PM method for northeast Spain, but for other areas it tended to underestimate the magnitude of change. The FAO-BC method provided a more similar spatial pattern to the PM method, but tended to overestimate the increase in ET_0 in the northeast, while the LIN, KH and P methods appeared to overestimate the change in ET_0 over most of mainland Spain. In average, the PM method indicated an increase of 24.5 mm decade^{-1}, with the greatest increase occurring in summer (12 mm decade^{-1}), although there were significant increases in the other seasons (Table 3). The other methods also showed positive changes, but the magnitudes differed markedly from those derived using the PM method. The temperature-based methods varied substantially, with the TH, HG and BC methods underestimating ET_0 changes at both the annual and seasonal scales relative to the PM method. In contrast, the LIN and KH methods overestimated the magnitude of ET_0 changes. Methods using more variables than temperature alone also showed differences from the PM method. Thus, the PT and R methods clearly underestimated the increase in ET_0 at both the seasonal and annual scales relative to the PM values, and the changes based on the T method were also smaller. In contrast, the P method substantially overestimated trends in ET_0, while the FAO-BC method provided the most accurate values in relation to the PM results.

The various ET_0 methods show inability to reproduce the spatial patterns in the magnitude of change in ET_0 using the PM method (Fig. 13); at both the annual and seasonal scales there was very little agreement with the latter method. The various ET_0 equations based only on temperature data failed to reproduce the patterns in the magnitude of ET_0 change across Spain. In addition, the methods requiring more variables for calculation differed markedly. For example, the spatial pattern obtained using the PT method showed no agreement with the PM method, whereas the FAO-BC and R methods showed a degree of agreement. At the seasonal scale the pattern was quite similar. Temperature based-methods tended to show worse results than the methods involving more variables, mainly during summer months.

4. Discussion and conclusions

In this study we estimated the magnitude and temporal evolution of ET_0 in Spain between 1961 and 2011 using a high quality dataset of diverse meteorological variables. Using the Penman-Monteith (PM) method as a reference, we compared the reliability of a range of other methods to quantify ET_0. We showed that these provided reasonable estimates with respect to the spatial patterns of average ET_0. For the annual and seasonal averages some of the ET_0 methods requiring only temperature data for calculation provided more agreement with the PM than more complex methods requiring more variables. Nevertheless, although all methods are reproducing the geographic gradients of ET_0, the differences in average magnitude can be important, even among methods that only use temperature in calculation (e.g. Thornthwaite and Linacre). It means that not only the available meteorological records are relevant in ET_0 calculations but also the calculation algorithms are also largely determining noticeable differences.

Among the various methods the Hargreaves (HG) equation provided the best results, at both the annual and seasonal scales. This equation has been suggested to be the best alternative where data are scarce (e.g. Droogers and Allen, 2002; Martínez-Cob, 2002; Hargreaves and Allen, 2003 Alexandris et al., 2008). Therefore, to determine average ET_0 values in Spain when data availability is limited we recommend use of the HG equation. There were no significant spatial differences in the performance of this equation in either humid (northern) or dry (southeast) climatic areas in Spain.

We also showed a general positive increase in ET_0 using the various methods. Thus, most of the 46 meteorological stations analyzed showed positive and statistically significant trends in ET_0. This is consistent with other studies in the Mediterranean region based on observational datasets (Chaouche et al., 2010; Espadafor et al., 2011; Papaioaunou et al., 2011; Polumbo et al. 2011; Vergni and Todisco, 2011; Kitsara et al., 2012). The magnitude of ET_0 change in Spain at the annual scale found using the PM equation (24.4 mm decade^{-1}) was similar to that reported for Greece between 1983 and 2001 (Papaioaunou et al., 2011), and southeast France between 1970 and 2006 (16–40 mm decade^{-1}; Chaouche et al., 2010), and was very similar to that reported by Espadafor et al. (2011) for nine stations in southern Spain between 1960 and 2005 (24 mm decade^{-1}).

We particularly note that the observed trends since the 1960s are the first to have been determined using high quality and homogeneous datasets of the variables used in Spain. Moreover, the patterns observed are consistent with observations in other Mediterranean regions (e.g. Brunetti et al., 2009; Papaioaunou et al., 2011), which implies that evaporative demand by the atmosphere may be increasing in the Mediterranean region, associated with the evolution of the meteorological variables involved; this is likely to increase aridity in the region. Vicente-Serrano et al. (2014) have showed that changes in ET_0 in Spain may be mainly determined by the evolution of relative humidity and maximum temperature. The decrease in relative humidity would have enhanced the increase in maximum temperature since the 1960s, particularly during the summer months. This would explain that the PM equation, which includes a complete evaluation of the aerodynamic component (based on relative humidity, air temperature and wind speed) shows the highest magnitude increase in ET_0 among the analysed equations.

Methods having limited data requirements (temperature based-methods) were found to be highly reliable in reproducing average ET_0 values and its general increase in Spain. Nevertheless, analysis of the temporal variability and trends in the magnitude of ET_0 suggested substantial uncertainties among the methods assessed, given the different climate variables involved in the calculations. Although temporal variability in the ET_0 series found using temperature-based methods (particularly the HG and HG-PP methods) was similar to that based on the PM method, reproducing the magnitude of change in ET_0 was more problematic. Thus, with very few exceptions the methods did not adequately reproduce the spatial patterns of change observed using the PM equation. Temporal changes in ET_0 are not only driven by temperature rise, and the differences in evolution of the other factors that determine the radiative (i.e., solar radiation) and aerodynamic components (i.e., air temperature, relative humidity and wind speed) (e.g., McVicar et al., 2012) would introduce differences among methods and spatially.

Very few studies have compared the performance of various methods for assessing changes in ET_0. At a global scale Dai (2011) and Sheffield (2012) produced contradictory results for how ET_0 is changing, based on analyses using the Thornthwaite (THO) and PM methods, respectively. Donohue et al. (2010) analyzed recent changes in ET_0 in Australia, using five different formulae. They reported very diverse spatial and temporal changes based on the various methods, and indicated that those methods based only on temperature variables (e.g. THO) tended to underestimate ET_0 changes, both positive and negative. In Spain, the methods that best reproduced the PM-based average magnitude, temporal variability and general positive trends in ET_0, including the HG, HG-PP and Turc (T) equations, are not suitable for identifying the magnitude of changes in ET_0 in Spain, and failed to reproduce its spatial patterns. The more complex methods did not provide better results; this highlights the difficulty of quantifying ET_0 changes using simple methods involving few variables.

The results of our study in Spain are in general agreement with current hypothesis and observations that suggest a general increase in atmospheric evaporative demand at the global scale (Brustaert and Parlange, 1998; Brustaert, 2006), and are consistent with continental water balance studies (e.g. Walter et al., 2004). Nevertheless, these patterns do not imply greater water supply to the atmosphere because ET_a is largely controlled by the available soil moisture. Droughts have increased in Spain during recent decades as a consequence of a reduction in precipitation (Vicente-Serrano, 2013). Under this scenario the observed ET_0 increase would not favor higher ET_a rates (Vicente-Serrano et al., 2013), but rather an increase in climate aridity because the soil water availability cannot supply the increased atmospheric demand. Thus, this relationship between ET_0, ETa and aridity was conceptually enunciated by Budyko (1974). His model has been validated by a number of studies (e.g., Van der Velde et al., 2013; Xu et al., 2014) that showed that the relationship between the evaporative index (ETa/Precipitation) describes a potential relationship with the dryness index (ET_0/Precipitation) and determined how water-limited or energy-limited are the different world environments.

In conclusion, our results along with other studies cited above suggest recommending the use of the HG equation to estimate the average ET_0 when only temperature data is available, which can be useful for agronomic and environmental purposes. Nevertheless, the differences found between the ET_0 estimations by means of PM and the rest of the methods in relation to the ET_0 temporal variability, the magnitude of the ET_0 changes and its spatial variability prevent for the recommendation of any alternative method to the PM equation when few data is available. Consequently, there is a need of evaluating trends based on methods that only require limited data for ET_0 calculations and developing higher quality series of relative humidity, wind speed and sunshine duration, in order to apply the robust Penman-Monteith equation. This may relevant for climate change studies, which are trying to determine ET_0 trends under current warming scenario.

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Table 1: Studies published in the last 15 years analysing ET0 trend across the Mediterranean region. * indicates the use of a modified method. The evolution correspond to: +, positive trend, -, negative trend, o, no trend.

Study	**Place**	**Method**	**Period**	**Evolution**
Pavanelli and Capra (2014)	Central Italy	Hargreaves	1926-2006	-
García-Garízabal et al. (2014)	North Spain	Hargreaves	1971-2000	+
Capra et al. (2013)	South Italy	Hargreaves*	1921-2007	-
Croitoru et al. (2013)	Romania	PM	1961-2007	+
Ugarkovic and Kelava (2013)	Croatia	Blaney-Criddle	1950-2010	+
Pravalie (2013)	South Romania	Thornthwaite	1961-2009	+
Kitsara et al. (2013)	Central Greece	Hargreaves*	1951-2001	+
Paltineanu et al. (2012)	South Romania	PM	2000-2007	+
Espadafor et al. (2011)	South Spain	PM	1960-2005	+
Moratiel et al. (2011)	Central Spain	PM	1980-2009	+
Papaioannou et al. (2011)	Greece	PM	1979-1999	+
Mavromatis and Stathis (2011)	Greece	Thornthwaite	1961-2006	+
Polumbo et al. (2011)	South Italy	Hargreaves	1957-2008	+
Vergni and Todisco (2011)	Central Italy	PM*	1951-2008	o
Matzneller et al. (2010)	North Italy	Hargreaves	1952-2007	+
Chaouche et al. (2010)	South France	PM	1970-2006	+
Kafle and Bruins (2009)	Israel	Thornthwaite	1970-2002	o
Zanchettin et al. (2008)	North Italy	Thornthwaite	1820-2002	+
Ozdogan and Salvucci, (2004)	South Turkey	PM	1979-2001	-
Moonen et al. (2002)	Northeast Italy	Hargreaves	1880-2000	-
Cohen et al. (2002)	Israel	PM	1964-1998	o

Table 2: Error measures used in this study

	Definitions: *N*: number of observations, *O*: Observed value, $\overline{O}$: mean of observed values, *P*: Predicted value, $P_i^{'} = P_i - \overline{O}$, $O_i^{'} = O_i - \overline{O}$
MBE (Mean bias error)	$MBE = N^{-1}\sum_{i=1}^{N}(P_i - O_i)$
MAD (Mean absolute difference)	$MAE = N^{-1}\sum_{i=1}^{N}\left\|P_i - O_i\right\|$
D	$D = 1 - \frac{\sum_{i=1}^{N}(P_i - O_i)^2}{\sum_{i=1}^{N}\left(\left\|P_i^{'}\right\| + \left\|O_i^{'}\right\|\right)^2}$

Table 3: Annual and seasonal magnitudes of change in ET_0 (mm decade^{-1}) based on the 12 methods for the regional series for mainland Spain.

	Annual	**Winter**	**Spring**	**Summer**	**Autumn**
Penman-Montheith	24.5	1.8	7.3	12	3.5
Thornthwaite	14.3	0	3.5	9.8	1
Hargreaves	15.1	1.9	5.6	6.4	1.3
Hargreaves-pp.	19.2	2.8	7.1	8.2	1.4
Linacre	42.8	7.2	12.4	16.7	7.2
Blaney-Criddle	12.3	1.5	4	5	1.9
Kharrufa	31.6	3	9.7	14.4	4.8
Priestley-Taylor	6.1	0.7	3	2.2	0.5
FAO-Blaney-Criddle	29.7	3.8	9.4	12.8	4
Turc	18.6	2.3	5.4	9.1	2.1
Papadakis	37.3	3.6	8.9	19.7	5.2
Radiation	13.4	1	4.1	6.4	2

Figure legends

Figure 1: Relationship between monthly average daily sunshine duration (hours) and monthly average global radiation (W m^{-2}), measured in seven stations in Spain (see Sanchez-Lorenzo et al., 2007, 2013) between 1980 and 2010.

Figure 2. Spatial distribution of the 46 meteorological stations used to calculate ET_0 in Spain. The polygons represent the weighting of each station in calculation of the regional series for Spain.

Figure 3: Box plot showing the annual and seasonal average ET_0 corresponding to the 46 meteorological stations used in the study.

Figure 4: Annual average ET_0 (mm) determined using the 12 equations for the 51 years of the study (1961–2011).

Figure 5. Evolution of annual ET_0 (mm) from the regional series for mainland Spain, determined using the 12 equations for the 51 years of the study (1961–2011).

Figure 6. Evolution of seasonal ET_0 from the regional series for mainland Spain, determined using the 12 equations for the 51 years of the study (1961–2011).

Figure 7: Box plot showing the R^2 coefficients between the annual and seasonal PM ET_0 series and the series of the other 11 methods for the 46 meteorological stations for the 51 years of the study (1961–2011).

Figure 8: Spatial distribution of the R^2 coefficients between the annual and seasonal PM ET_0 series and the series of the other 11 methods for the 46 meteorological stations for the 51 years of the study (1961–2011).

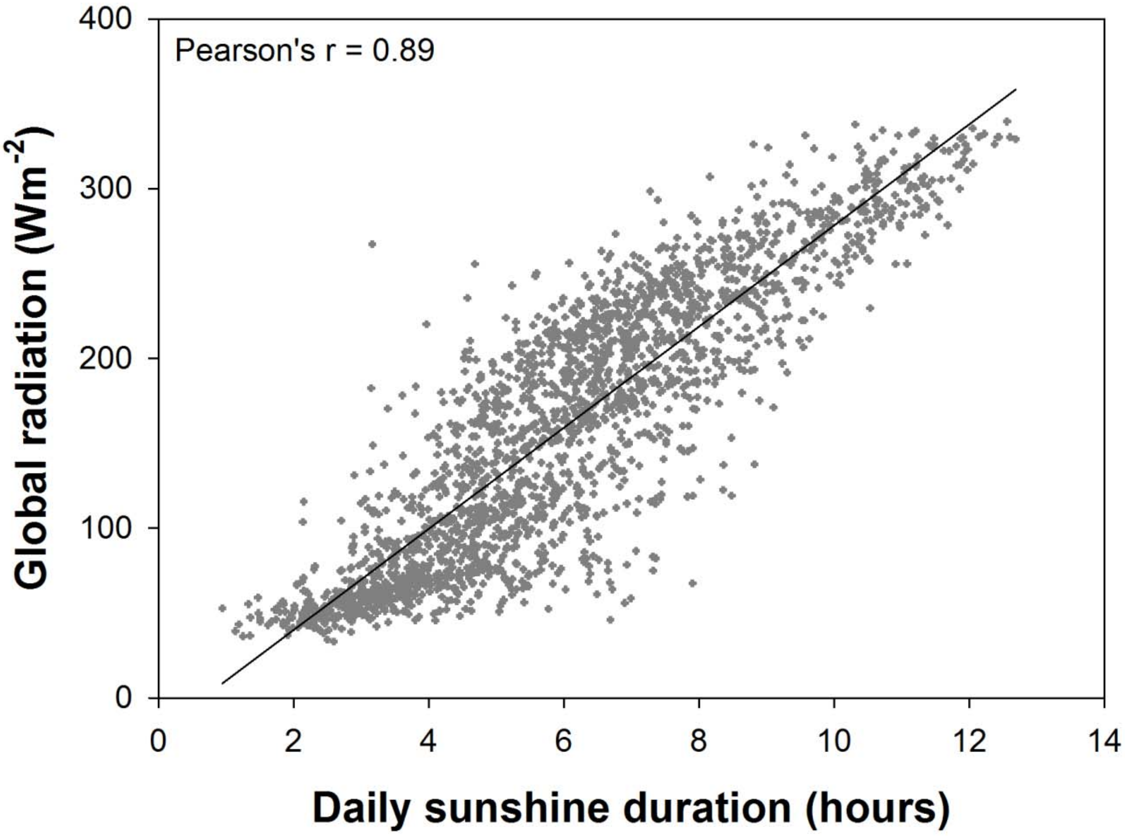
Figure 9: Box plot showing the Willmott's D statistics between the annual and seasonal PM ET_0 series and the series of the other 11 methods for the 46 meteorological stations for the 51 years of the study (1961–2011).

Figure 10: Spatial distribution of the Willmott's D statistic between the annual and seasonal PM ET_0 series and the series of the other 11 methods for the 46 meteorological stations for the 51 years of the study (1961–2011).

Figure 11. Box plot showing the annual and seasonal magnitude of change in ET_0 using the 12 methods for the 46 meteorological stations in Spain for the 51 years of the study (1961–2011).

Figure 12: Spatial distribution of the annual magnitude of change in ET_0 for the 46 meteorological stations in Spain for the 51 years of the study (1961–2011). The legend represents annual ET_0 changes (mm per decade^{-1}). Figure 13 shows the spatial distribution of the magnitude of change in annual ET_0 for the 46 meteorological stations.

Figure 13: Relationship between the annual and seasonal magnitudes of change in ET_0, derived using the PM method and the other 11 methods for the 46 meteorological stations for the 51 years of the study (1961–2011).



Pearson's r = 0.89
Global radiation (Wm-2)
400
300
200
100
0
0
2
4
6
8
10
12
14
Daily sunshine duration (hours)

Figure 1

Figure 2

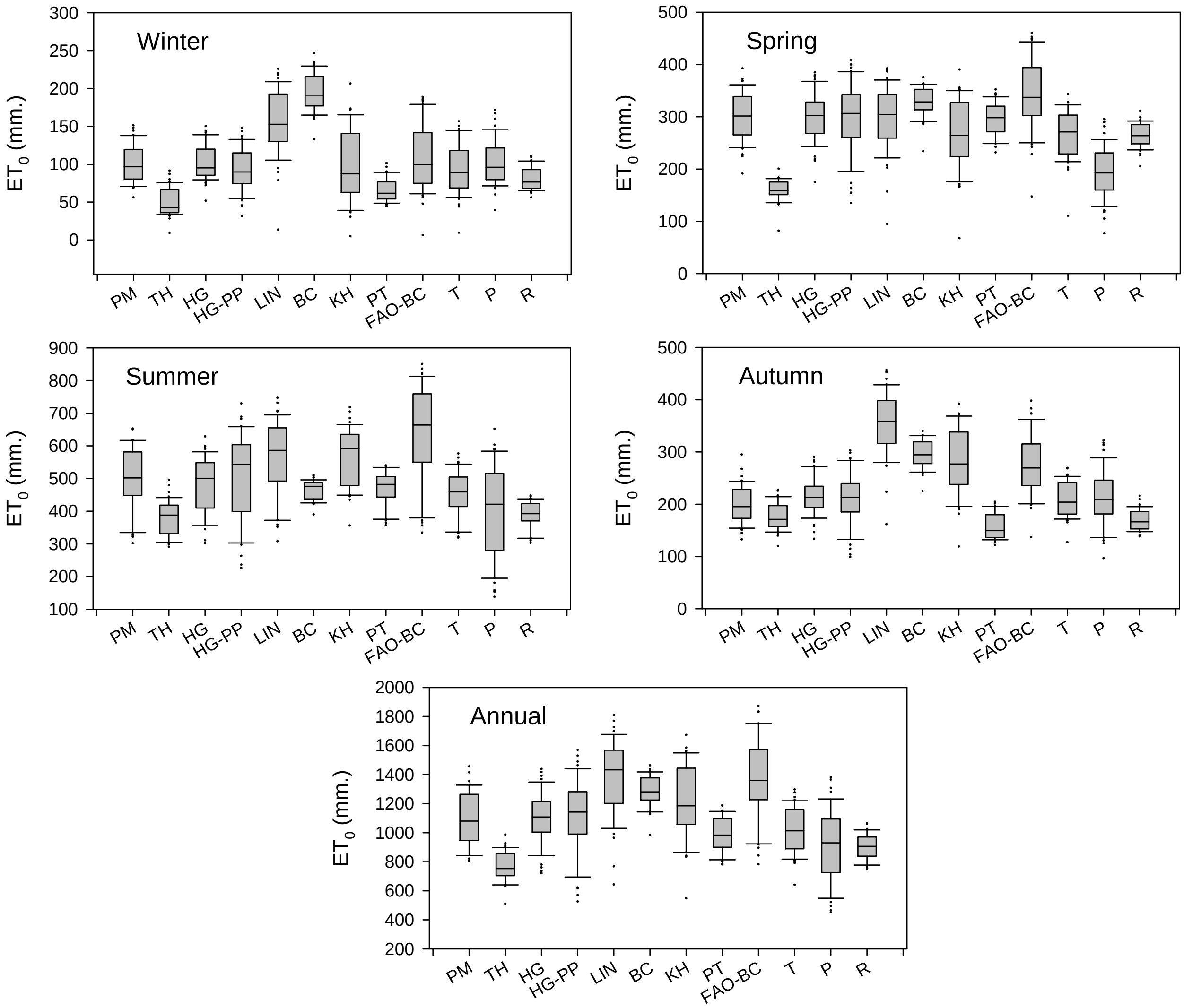

Winter
Spring
Summer
Autumn
Annual
ET0 (mm.)
PM
TH
HG
HG-PP
LIN
BC
KH
PT
FAO-BC
T
P
R

Figure 3.

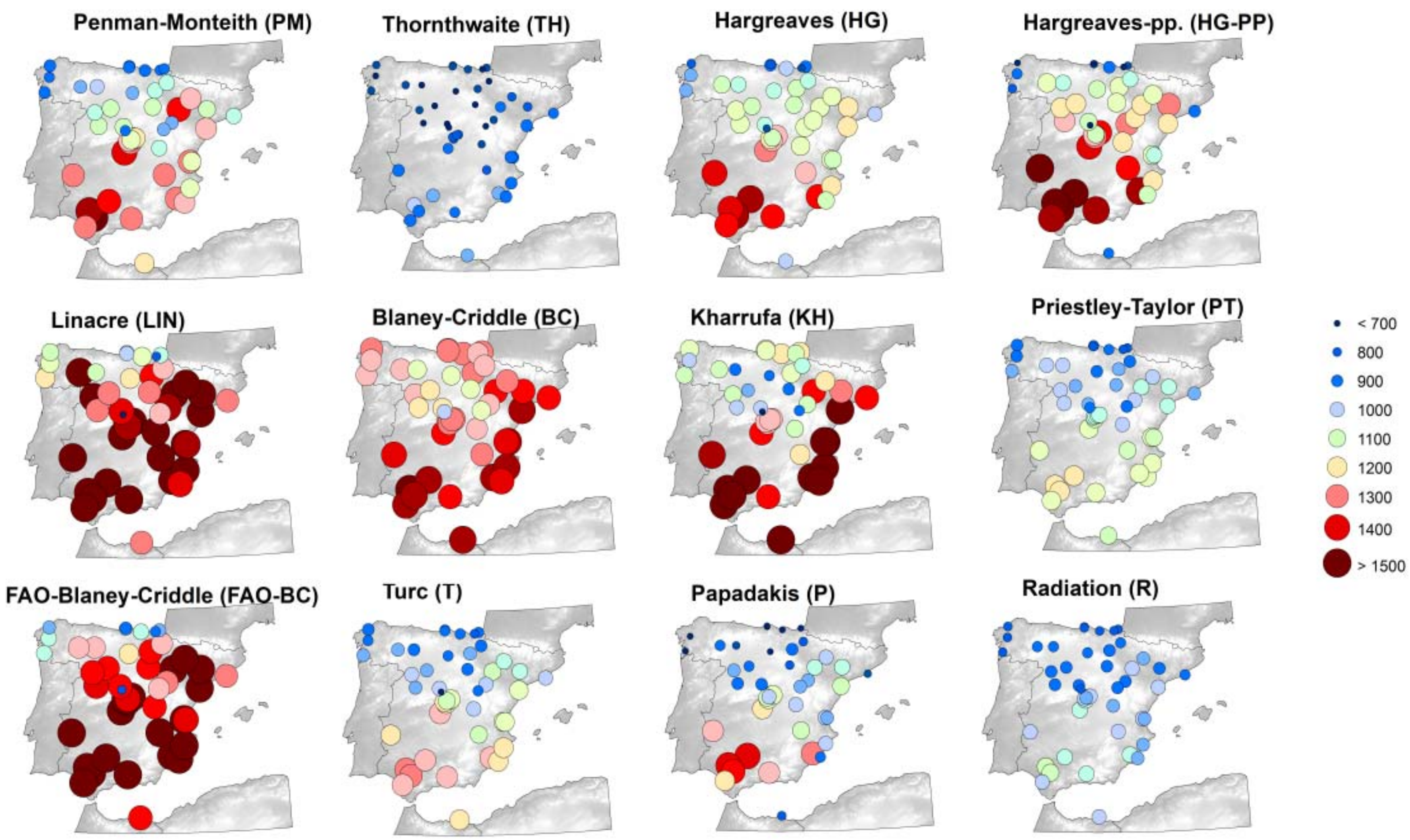

Penman-Monteith (PM)
Thornthwaite (TH)
Hargreaves (HG)
Hargreaves-pp. (HG-PP)
Linacre (LIN)
Blaney-Criddle (BC)
Kharrufa (KH)
Priestley-Taylor (PT)
FAO-Blaney-Criddle (FAO-BC)
Turc (T)
Papadakis (P)
Radiation (R)
< 700
800
900
1000
1100
1200
1300
1400
> 1500

Figure 4.

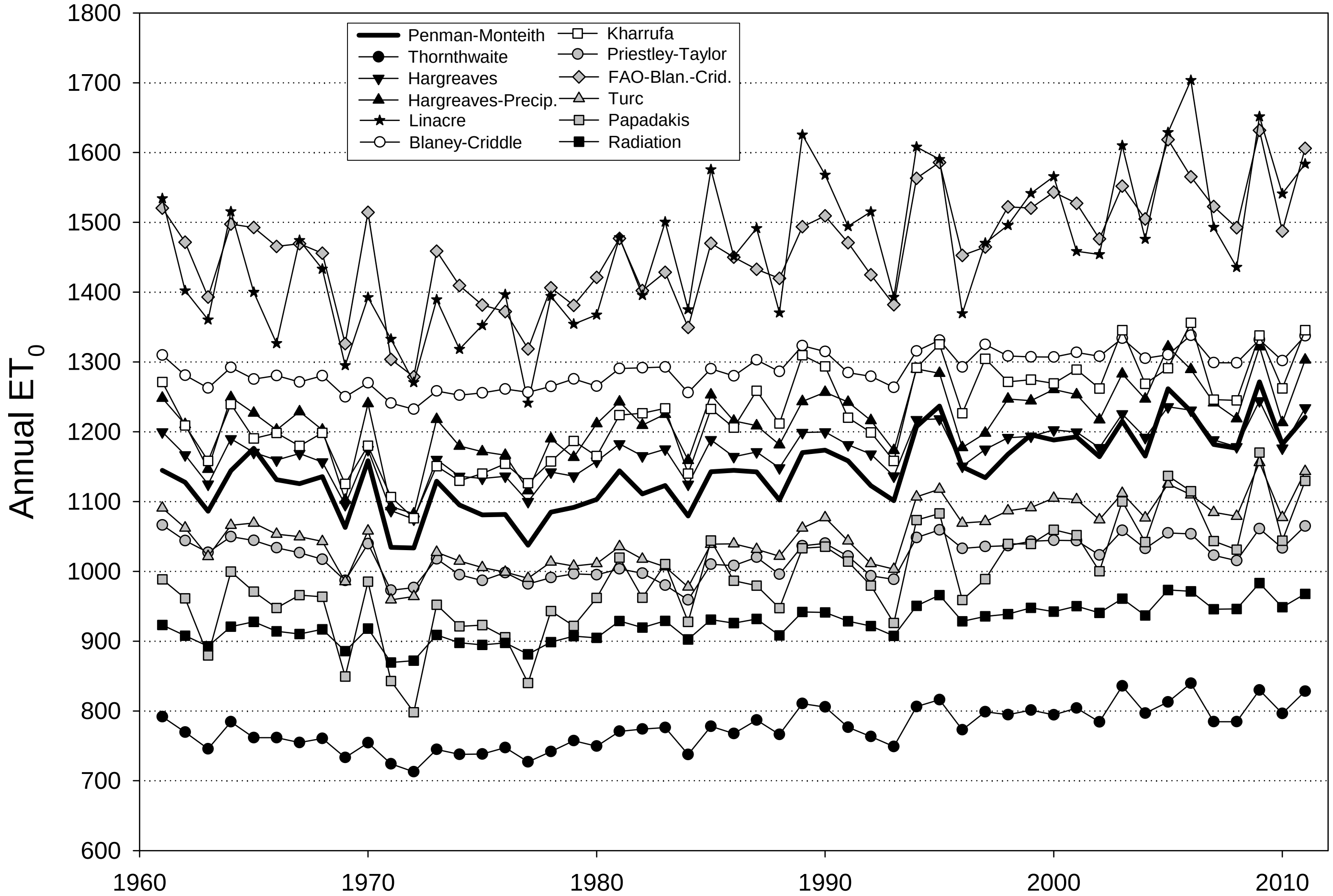

Penman-Monteith
Thornthwaite
Hargreaves
Hargreaves-Precip.
Linacre
Blaney-Criddle
Kharrufa
Priestley-Taylor
FAO-Blan.-Crid.
Turc
Papadakis
Radiation
Annual ET_0
1800
1700
1600
1500
1400
1300
1200
1100
1000
900
800
700
600
1960
1970
1980
1990
2000
2010

Figure 5.

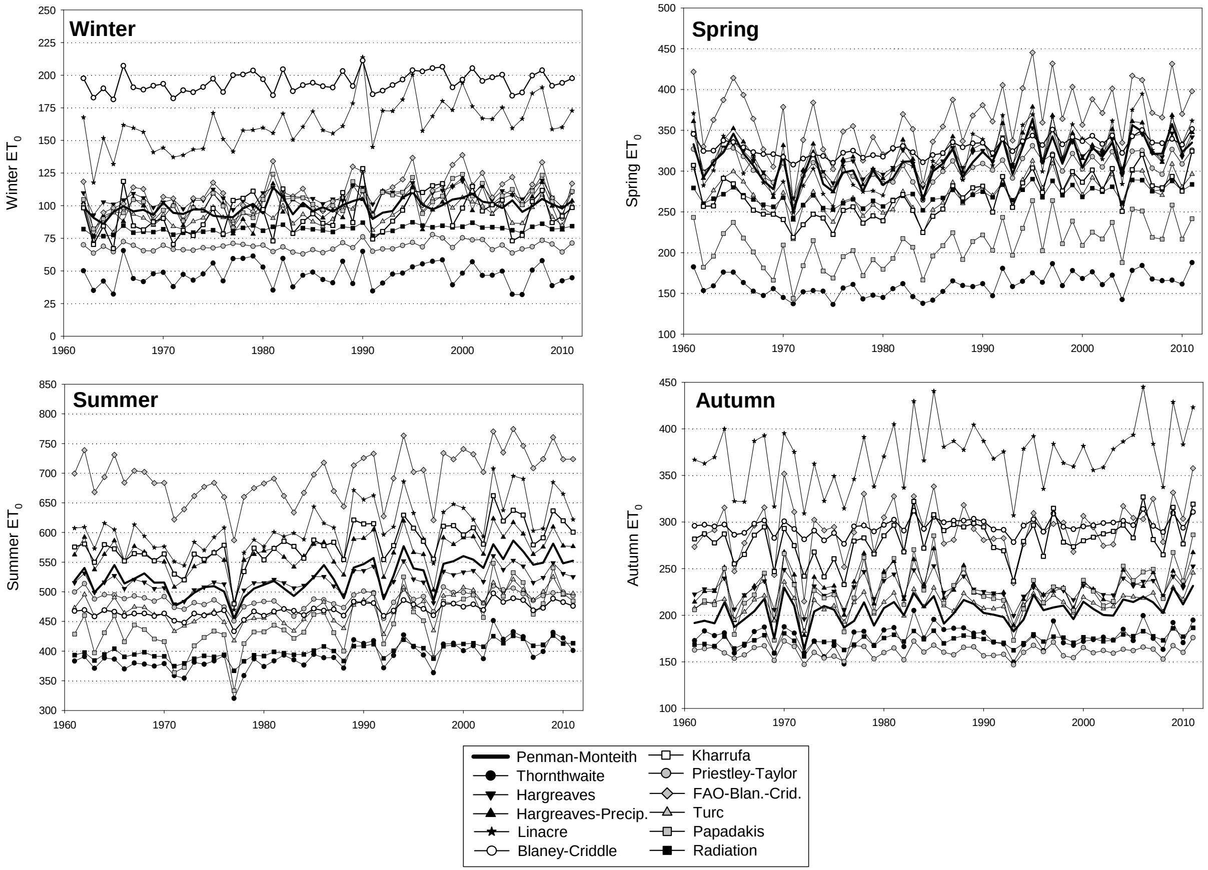
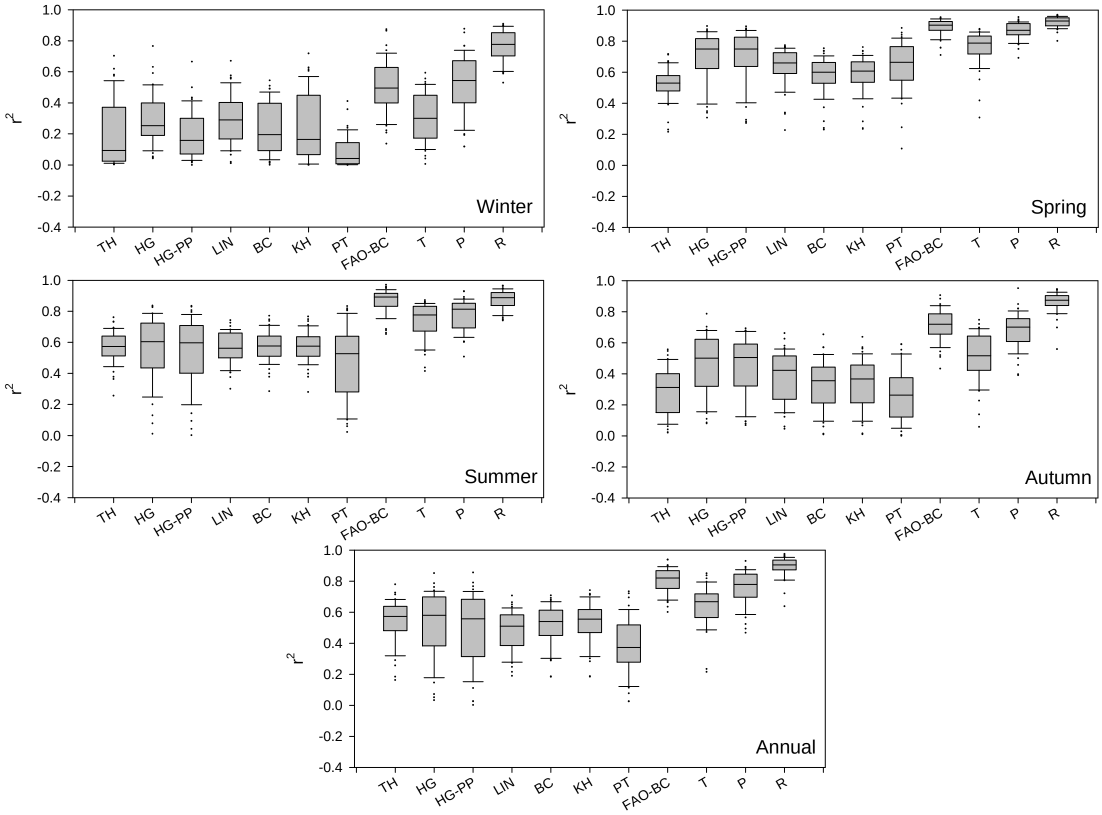
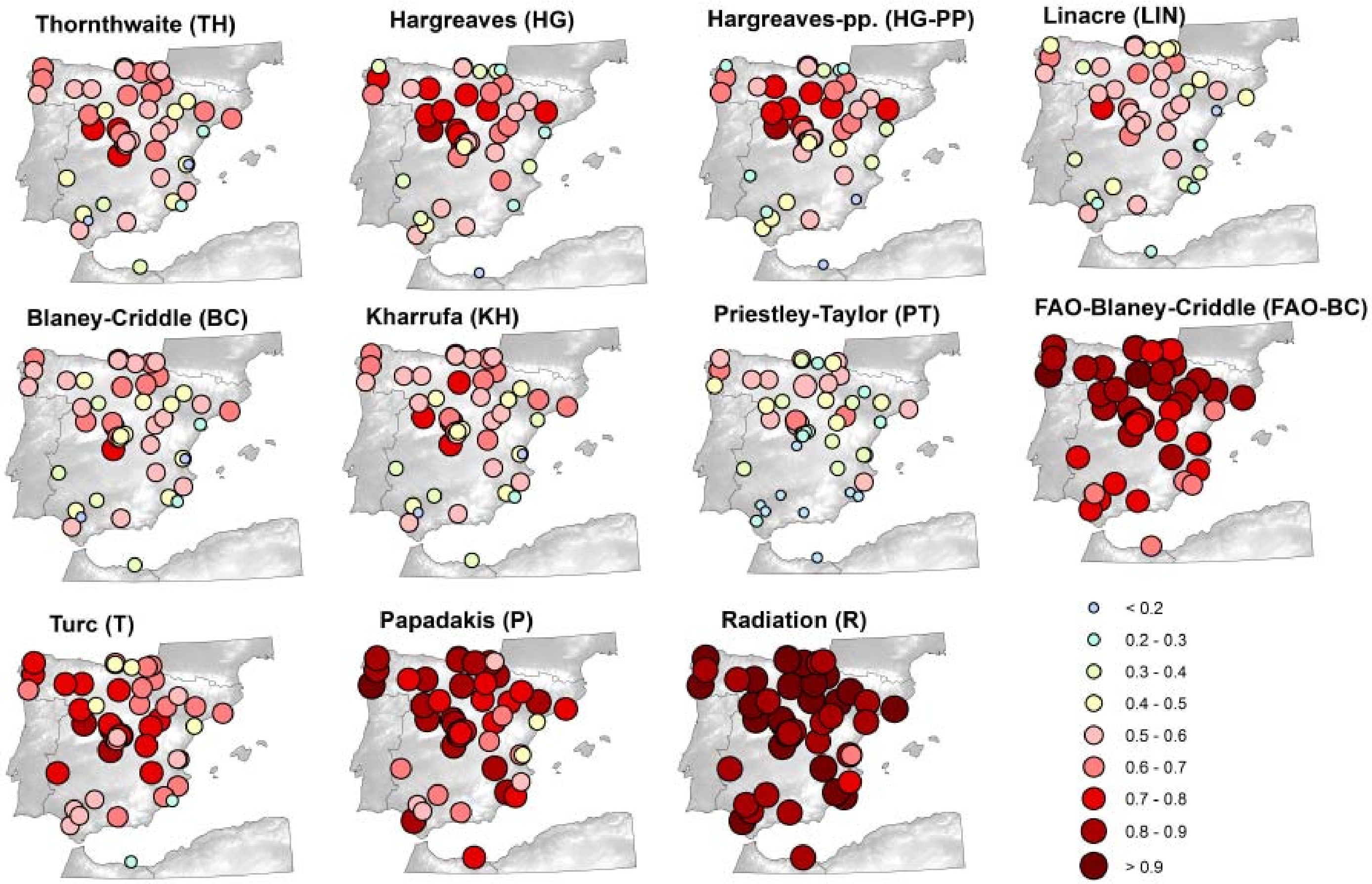

Winter
Winter ET0
Spring
Spring ET0
Summer
Summer ET0
Autumn
Autumn ET0
Penman-Monteith
Thornthwaite
Hargreaves
Hargreaves-Precip.
Linacre
Blaney-Criddle
Kharrufa
Priestley-Taylor
FAO-Blan.-Crid.
Turc
Papadakis
Radiation

Figure 6.



Winter
Spring
Summer
Autumn
Annual
r2
1.0
0.8
0.6
0.4
0.2
0.0
-0.2
-0.4
TH
HG
HG-PP
LIN
BC
KH
PT
FAO-BC
T
P
R

Figure 7.



Thornthwaite (TH)
Hargreaves (HG)
Hargreaves-pp. (HG-PP)
Linacre (LIN)
Blaney-Criddle (BC)
Kharrufa (KH)
Priestley-Taylor (PT)
FAO-Blaney-Criddle (FAO-BC)
Turc (T)
Papadakis (P)
Radiation (R)
< 0.2
0.2 - 0.3
0.3 - 0.4
0.4 - 0.5
0.5 - 0.6
0.6 - 0.7
0.7 - 0.8
0.8 - 0.9
> 0.9

Figure 8.

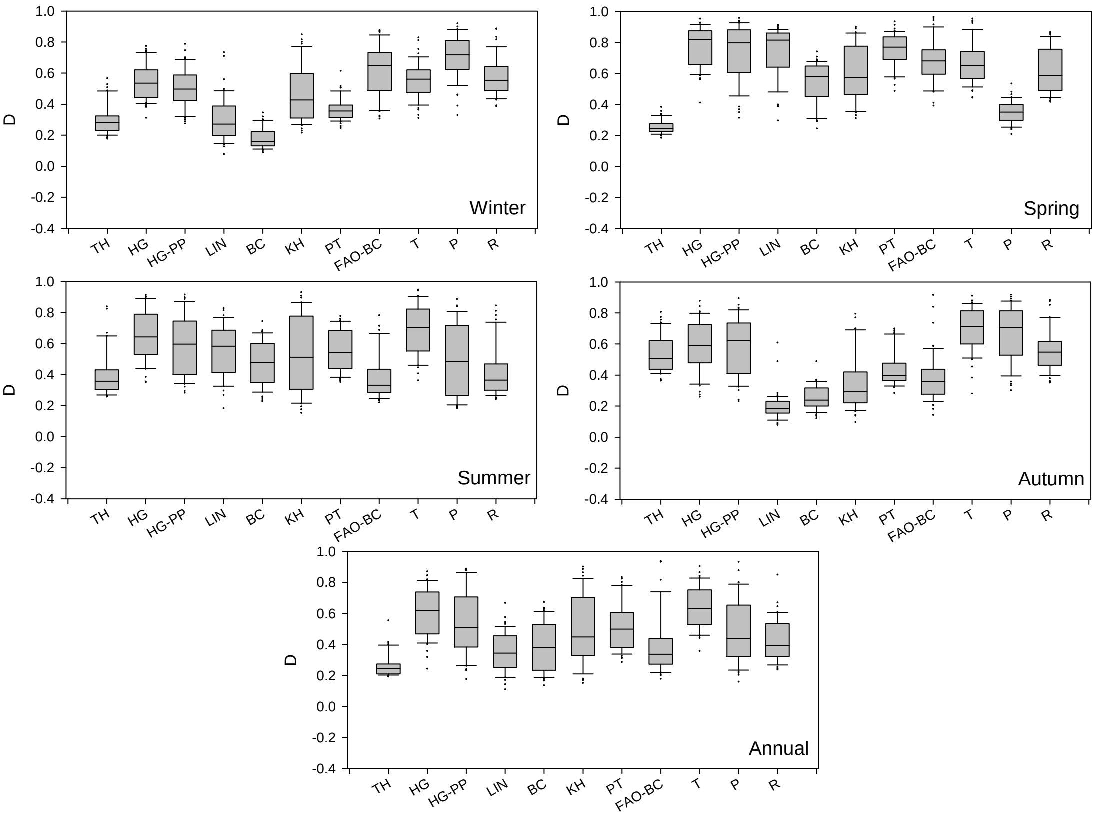

Winter
Spring
Summer
Autumn
Annual
D
1.0
0.8
0.6
0.4
0.2
0.0
-0.2
-0.4
TH
HG
HG-PP
LIN
BC
KH
PT
FAO-BC
T
P
R

Figure 9.

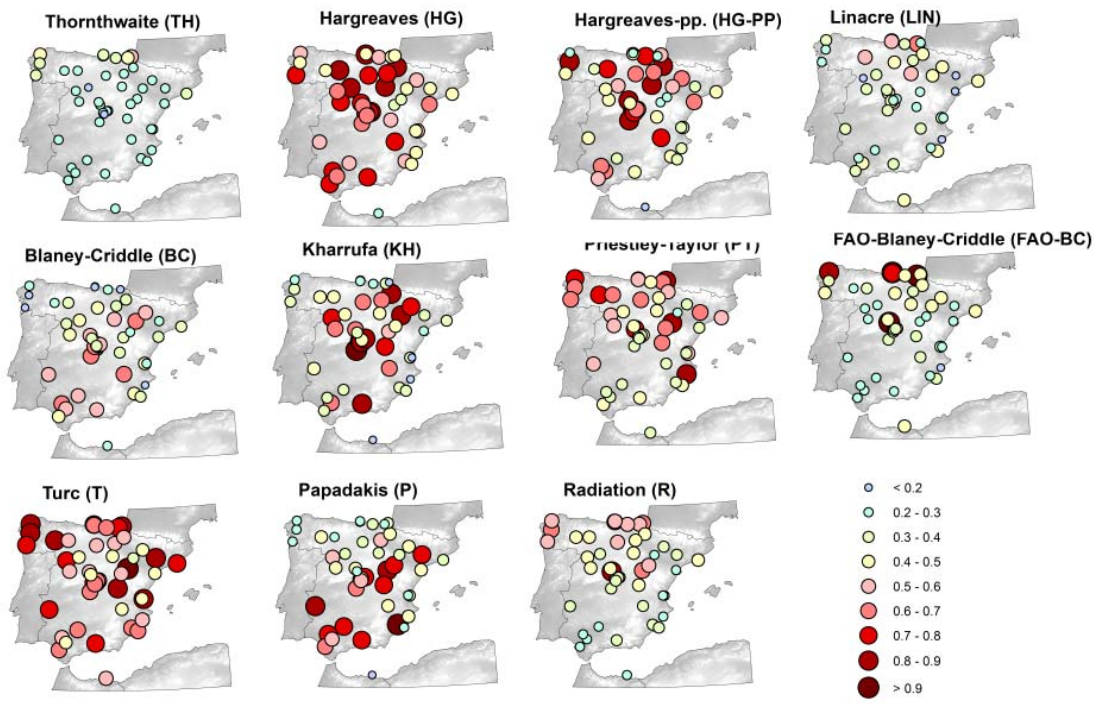

Thornthwaite (TH)
Hargreaves (HG)
Hargreaves-pp. (HG-PP)
Linacre (LIN)
Blaney-Criddle (BC)
Kharrufa (KH)
Priestley-Taylor (PT)
FAO-Blaney-Criddle (FAO-BC)
Turc (T)
Papadakis (P)
Radiation (R)
< 0.2
0.2 - 0.3
0.3 - 0.4
0.4 - 0.5
0.5 - 0.6
0.6 - 0.7
0.7 - 0.8
0.8 - 0.9
> 0.9

Figure 10.

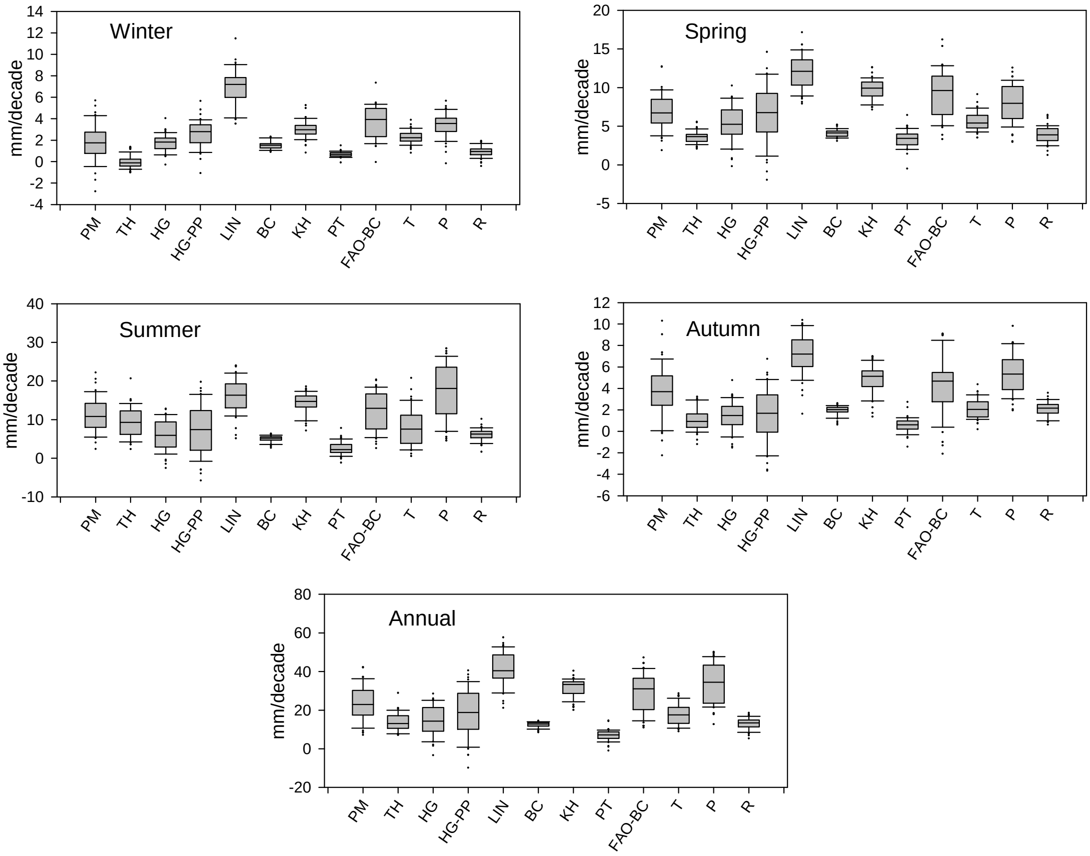

Winter
Spring
Summer
Autumn
Annual
mm/decade
PM
TH
HG
HG-PP
LIN
BC
KH
PT
FAO-BC
T
P
R

Figure 11.

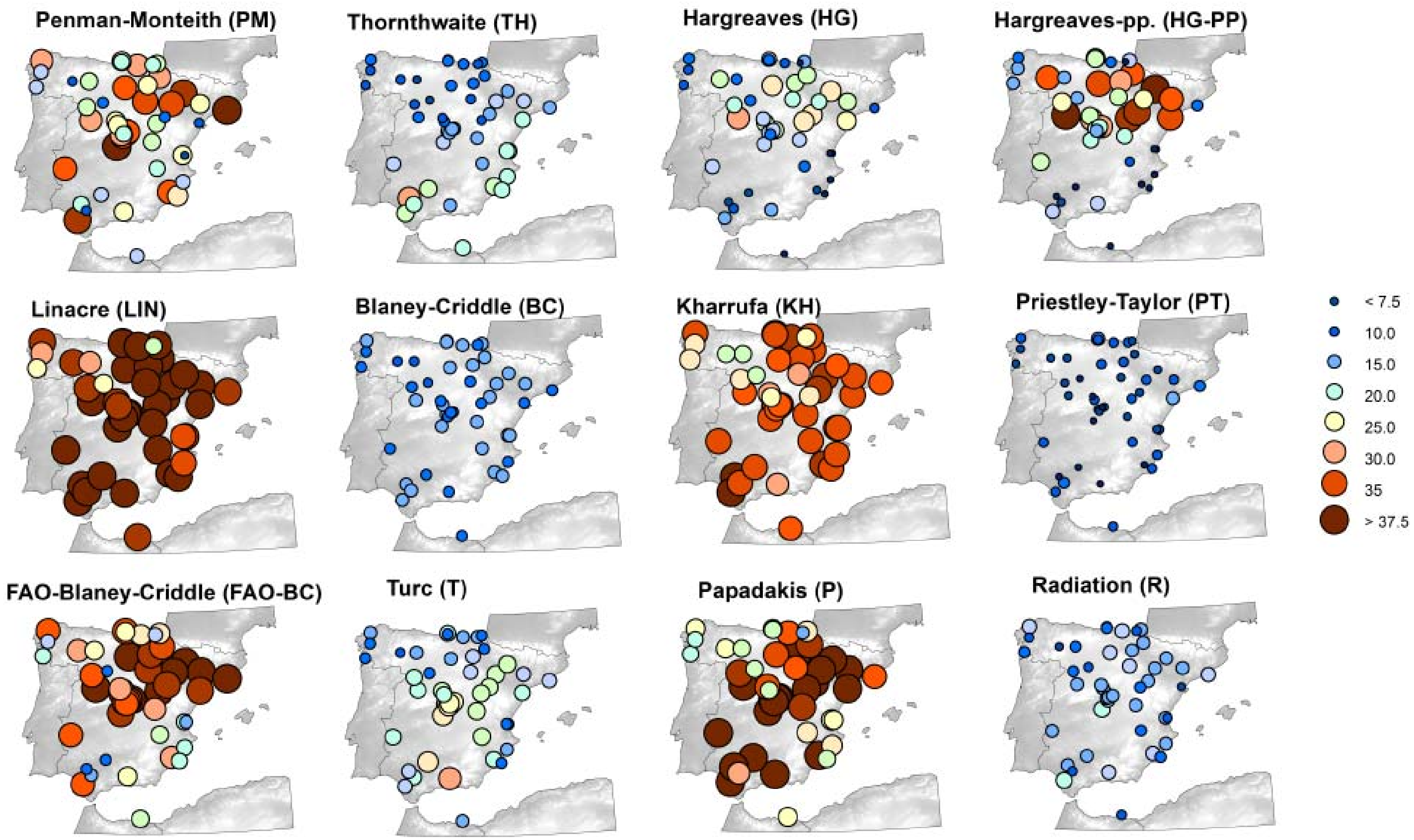

Penman-Monteith (PM)
Thornthwaite (TH)
Hargreaves (HG)
Hargreaves-pp. (HG-PP)
Linacre (LIN)
Blaney-Criddle (BC)
Kharrufa (KH)
Priestley-Taylor (PT)
FAO-Blaney-Criddle (FAO-BC)
Turc (T)
Papadakis (P)
Radiation (R)
< 7.5
10.0
15.0
20.0
25.0
30.0
35
> 37.5

Figure 12.

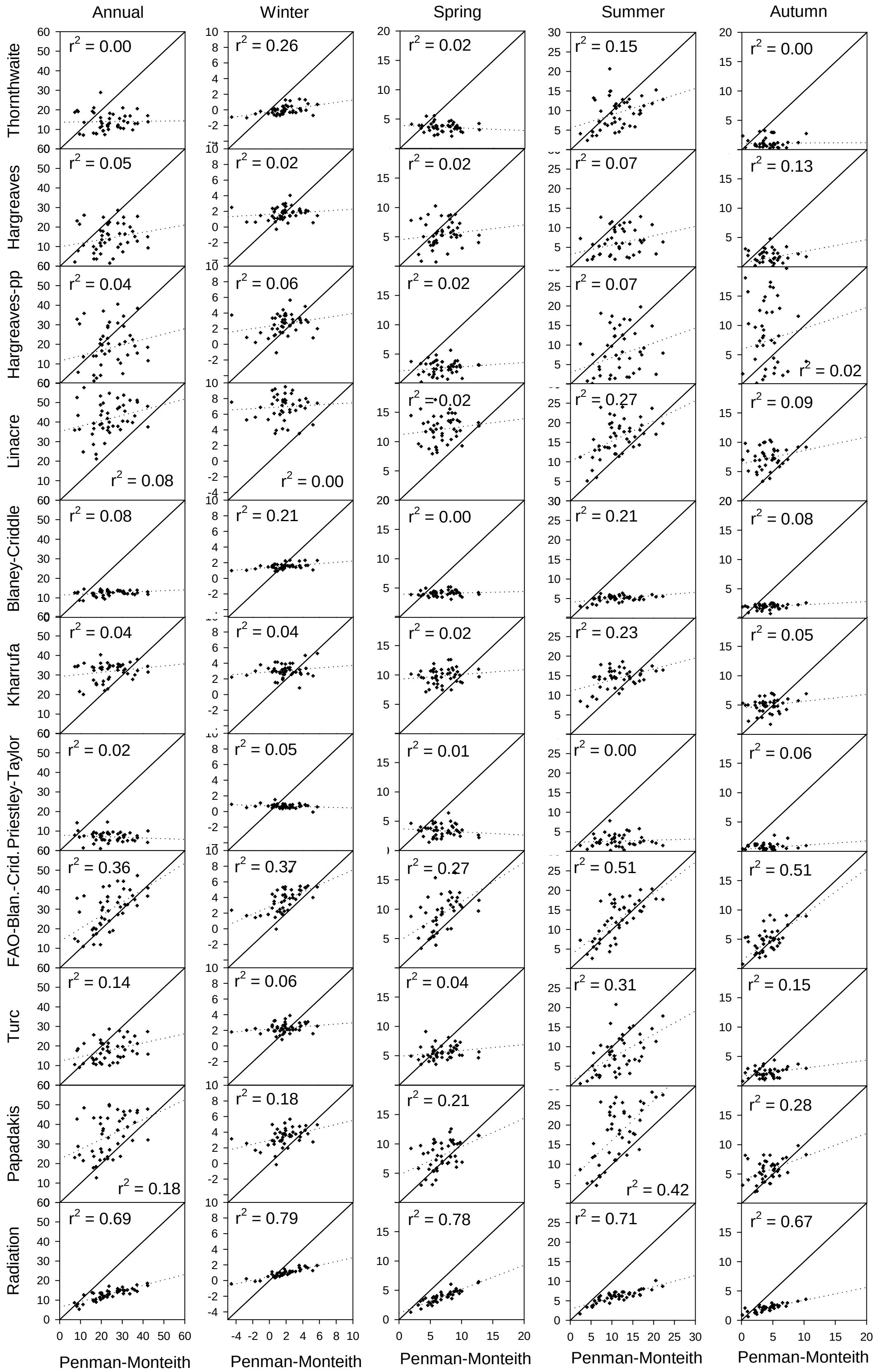

Annual
Winter
Spring
Summer
Autumn
Thornthwaite
r^2 = 0.00
r^2 = 0.26
r^2 = 0.02
r^2 = 0.15
r^2 = 0.00
Hargreaves
r^2 = 0.05
r^2 = 0.02
r^2 = 0.02
r^2 = 0.07
r^2 = 0.13
Hargreaves-pp
r^2 = 0.04
r^2 = 0.06
r^2 = 0.02
r^2 = 0.07
r^2 = 0.02
Linacre
r^2 = 0.08
r^2 = 0.00
r^2 = 0.02
r^2 = 0.27
r^2 = 0.09
Blaney-Criddle
r^2 = 0.08
r^2 = 0.21
r^2 = 0.00
r^2 = 0.21
r^2 = 0.08
Kharrufa
r^2 = 0.04
r^2 = 0.04
r^2 = 0.02
r^2 = 0.23
r^2 = 0.05
Priestley-Taylor
r^2 = 0.02
r^2 = 0.05
r^2 = 0.01
r^2 = 0.00
r^2 = 0.06
FAO-Blan.-Crid.
r^2 = 0.36
r^2 = 0.37
r^2 = 0.27
r^2 = 0.51
r^2 = 0.51
Turc
r^2 = 0.14
r^2 = 0.06
r^2 = 0.04
r^2 = 0.31
r^2 = 0.15
Papadakis
r^2 = 0.18
r^2 = 0.18
r^2 = 0.21
r^2 = 0.42
r^2 = 0.28
Radiation
r^2 = 0.69
r^2 = 0.79
r^2 = 0.78
r^2 = 0.71
r^2 = 0.67
Penman-Monteith
Penman-Monteith
Penman-Monteith
Penman-Monteith
Penman-Monteith

Figure 13.