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ANALYSIS AND MODIFICATION OF METEOROLOGICAL DROUGHT INDICES IN DIVERSE HYDROCLIMATIC ENVIRONMENTS

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Abstract

The impact of climate variability on hydrological drought in Thessaly, Greece, is investigated by applying several indices computed on 12-month time scale. Six indices (i.e. SPI, SPEI, RDI, P/T, modSPEI, modRDI) are used as indicators of meteorological drought severity when the hydrological drought severity is evaluated from the monthly UTHBAL conceptual water balance model. The latter indices, the modified SPEI and RDI (modSPEI, modRDI) use the actual instead of potential evapotranspiration. Previously, in Yermasoyia watershed, Cyprus, the most appropriate method for calculating the potential evapotranspiration is evaluated for a short but even efficient dataset and the Thornwaite method is chosen. All the indices are computed for the common period 1960-2002 using monthly data from 10 watersheds. Results suggest that: (i) The indices strongly correlate with each other at most of the regions; (ii) The correlations between the SPEI and RDI time series remain high, while between the SPI and SPEI/RDI noticeably decrease in most stations characterized by high temperatures and potential evapotranspiration; (iii) Though the correlation between most of the indices is high at all stations, several differences on the severity of dry/wet events occur as well as a nonlinear relationship emerges in several locations; (iv) From the new, modified indices, the modSPEI seems to be really efficient, while modRDI shows numerous of discrepancies. In analyzing the detected discrepancies, the response of the meteorological indices is evaluated for identified drought episodes. Also, the annual mean temperature field is checked for long-term trend. The analysis shows that temperature is really slowly increased during the last decade of the time section at several stations, mainly in East Thessaly. However, it seems that the recent temperature increase can not be always responsible for the disagreement between the indices. The effectiveness of the indices in capturing the impact of a changing climate and also the importance of using the actual instead of potential evapotranspiration on drought characteristics are evaluated and discussed.

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To anyone who never gives up

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Chapter 1 . Introduction

1.1 In general

Water – the most fundamental and valuable resource of the world. Drought means lack of water; water in any status, precipitation, soil moisture and resources over an extended period of time, resulting in water storage that normally would be available in a region and to which nature and mankind have adapted over centuries.

Because of the numerous sectors that affects and also for the numerous factors that cause drought, there is not a specific definition for it. But there is a well-known classification proposed initially proposed by Dracup and his associates (Dracup et al., 1980 and integrated later by Wilhite and Glantz, 1985)). This four- category drought classification system was adopted by the American Meteorological Society (2004). Based on the nature of the water deficit, four types of droughts are defined: a) the meteorological drought which is defined as a lack of precipitation over region for a period of time, b) the hydrological drought which is related to a period with deficiency in surface and subsurface water supplies of a given water resources management system, c) the agricultural drought, which, links impacts of meteorological drought to agricultural and usually refers to a period with declining soil moisture and consequent crop failure without any reference to surface water resources, d) the socio-economic drought which is associated to the failure of water resources systems to meet the water demands and thus, associating droughts with supply of and demand for an economic good (water).

In general, engineers and scientists consider that drought is a regional phenomenon characterized by three dimensions; severity or intensity, the duration and the areal extent (*Rossi et al,* 1992). It has also the hydrological component as the most important factor in the whole procedure. It is very difficult to objectively quantify drought characteristics in terms of intensity, magnitude, duration and spatial extent due to the adversity of defining the onset, extend and end. However, it is necessary for the analysis of droughts to detect their features and to link the drought variability to climate (*Piechota and Dracup,* 1996). For this purpose, much effort has been devoted to develop techniques in order to analyze, monitor and evaluate the droughts. Among these, drought indices are the most widely used. Drought indices are employed to characterize droughts and the drought statistical properties. A lot of researchers have tried to develop robust models for the drought indices in order to enable the severity of droughts in different locations to be compared independently of local climatic characteristics. Many attempts have been made in order to produce new drought indices and improve the existing ones (*González and Valdés,* 2004; *Keyantash and Dracup,* 2004; *Wells et al.,* 2004; *Tsakiris et al.,* 2007; *Vicente-Serrano et al.,* 2010). Along the various indices proposed for the characterization of meteorological drought,, the Palmer Drought Severity Index (PDSI; *Palmer,* 1965) and the Standardized Precipitation Index (SPI; *McKee et al.,* 1993) are the most widely used for monitoring purposes.

The PDSI is based on the supply-and-demand concept of the water balance equation for a two-layer soil model using monthly precipitation and evapotranspiration (ET). The index bases upon the calculation of the moisture departure between actual precipitation and the precipitation expected to occur for the average conditions of the climate, which implies performing a monthly water balance and the calibration of local monthly coefficients for the various terms of the water balance. In addition, the calibration of the index is required at local or regional level. The PDSI values depend upon the way how the soil water balance is computed, the soil available water capacity (AWC) of the underlying layer, the ET estimation method, and on the calibration procedures adopted. However, the PDSI is conceived differently from an aridity index, which is a merit to be recognized to its author. In view of updating the calculation procedures and improving the information provided through its use, a modification for Mediterranean conditions was produced, the MedPDSI (*Pereira et al.*, 2007, *Rosa et al.,* 2010), where actual ET is used instead of potential ET. Due to computation requirements and limitations on the PDSI (*Alley,* 1984; *Karl and Knight,* 1985), usually a “standardized” and “multi-scale” drought index, like the SPI, has been preferred to facilitate the quantitative comparison of drought events at different locations and time scales. The SPI is based on precipitation field alone. It was designed to quantify the precipitation deficit for multiple time scales, which reflect the impact of drought on the availability of the different water resources. The index computation for any location is based on the long-term precipitation record accumulated over the selected time scale. The long-term record is fitted to a probability distribution, usually a Gamma distribution (*Guttman,* 1999), which is then transformed through an equal-probability transformation into a normal distribution.

Recently, concerns have been raised by some authors on the suitability of the SPI in properly detecting the impact of a changing climate on drought characteristics. Drought is a relative concept (relative to some established norm) and originates from a deficiency of precipitation over an extended period of time. However, other climatic factors such as high temperature, high wind and low relative humidity, can aggravate its severity. The SPI is based on the assumption that drought variability is mainly controlled by variations in precipitation field alone, without taking into account the temperature issue. In terms of climate change, tendencies towards increasing temperatures during the last decades have been observed in several regions around the world and it is expected that may affect the intensity and duration of drought events. For this reason, two new drought indices have been proposed: the Standardized Precipitation Evapotranspiration Index (SPEI) and the Reconnaissance Drought Index (RDI). The SPEI index (*Vicente-Serrano et al.,* 2010) is based on the original SPI calculation procedure. In this case, the input variable is the difference between precipitation and potential evapotranspiration accumulated on a selected time scale, which is a simple measure of the theoretical water surplus or deficit for the analyzed month, but it performs like an aridity index because potential evapotranspiration (PET) is used. The empirical statistical distribution of the input variable is fitted to a Log-logistic distribution and then, as in the SPI computation, transformed into a normal distribution. It is based on precipitation and potential evapotranspiration (PET) and combines the sensitivity of PDSI and to changes in evaporation demand (caused by temperature fluctuations and trends) with the simplicity of calculation and the multi-temporal nature of the SPI. According to recent studies (*Vicente-Serrano et al.,* 2010; *Potop,* 2011) the role of temperature increase on drought conditions was not recognized using the precipitation-based SPI drought index, but was identified for the drought in the decade of 2000 using the SPEI index. The RDI index (*Tsakiris and Vangelis,* 2005; *Tsakiris et al.,* 2007) has been developed to estimate a water deficit. It is based on the ratio between precipitation and potential evapotranspiration, and can be computed for different time scales. The index, in its standardized form, can be easily compared with the SPI. It is considered as an ideal index to study the effects of climate instability conditions.

In both of the indices, the key factor is the calculation of the potential evapotranpiration (PET). This goal can be achieved through a lot of different models (Hargreaves, Thornwaite, Kharrufa, Blanney, Turc, Jensen-Haise, Penman) but as did Palmer for the PDSI, the Thornthwaite method (*Thornthwaite,* 1948) is common used for large-scale drought analysis. It provides more flexibility for the calculations as less input are needed and also gives results which are really close to actual evapotranspiration (AET), especially for large time scale drought analysis. However, using potential evapotranspiration (PET) instead of actual evapotranspiration (AET) makes the index to approach an aridity index. So, it is really important a possible replacement of the potential evapotranspiration from the actual.

Many studies have been done in order to test the effectiveness of drought indices in monitoring and analyzing the regional droughts. Hayes (*Hayes et al.,* 1999) and his associates used the Standardized Precipitation Index (SPI) in order to monitor the drought of 1996 in the southern and southwestern USA. It was concluded that the SPI has the ability to detect the onset of the drought and monitor its progression even though the onset of the drought was detected at least one month earlier. Earlier, Guttman (1998) with a spectral analysis, compared historical time series of PDSI with time series of SPI and demonstrated the spatially invariance of the spectral characteristics of the SPI in contrast with those of the PDSI which were spatially variable. For this reason, it concluded that SPI is a better drought index for interregional comparison of drought events. In his study, Guttman (1998) also found that the 12-month SPI oscillations were found to be in phase with the oscillations of PDSI. Loukas et al. (2003) in an intercomparison of SPI, Zscore, RAI, the PDSI and the Palmer moisture anomaly index (Z-Index) (*Palmer,* 1965) at 28 meteorological stations in Greece found similar results. In that study, the oscillations of the SPI were found to be in phase with the oscillations of the Z-score and the RAI for the same time-scales and with the oscillations of the PDSI for time-scales larger than 6-months. Also, the SPI was used in recent regional drought analyses in the Mediterranean region using rain gauge data (*Bordi et al.,* 2001; *Lana et al.,* 2001), gridded rain gauge data (*Lloyd-Hughes and Saunders*, 2002; *Tsakiris and Vangelis,* 2004), and NCEP/NCAR reanalysis gridded precipitation data (*Bordi and Sutera*, 2001). The above studies indicate that the Mediterranean region has been afflicted by severe and more or less prolonged periods of drought in the last 50 years. Another crucial issue for the drought assessment is the relationship between the timescales of the indices and the types of drought. Vicente- Serrano and Lopez-Moreno (2005) analyzed the usefulness of different SPI timescales to monitor droughts in river discharges and reservoir storages. It was shown that the response of the river discharges to higher timescales than 3-months is very low and there must be higher timescales in order to monitor river flow droughts in the mountain hydrological system. On the other hand, the timescales of SPI, useful to analyze droughts in the reservoir storages are longer than for river discharges. Vasiliades and Loukas (2006) comparing SPI and the Palmer four indices, showed that SPI is better related to hydrological drought. However, different timescales of SPI were best correlated to hydrological drought for the study watersheds depending on their area, geophysical, and hydroclimatic characteristics.

Concerning the climate change which might have as a result an increase in the temperature and also the reliability of the indices in climate variability, Vicente-Serrano et al.(2010), compared time series of three drought indices for a set of observatories with different climate characteristics located in different parts of the world. They used the Standardized Precipitation Index (SPI), the sc-PDSI and the proposed Standardized Precipitation Evapotranspiration Index (SPEI) and they concluded that under global warming conditions only the sc-PDSI and SPEI identified an increase in drought severity associated with higher water demand due to evapotranspiration. Comparing the two indices, the SPEI is reliable for drought analysis and monitoring due to its multi-scalar character.

G.Tsakiris et al., (2007) compared the Reconnaissance Drought Index (RDI) with the Standardized Precipitation Index (SPI) and the method of deciles for the efficiency in a changing climate environment. They found out that in the majority of cases RDI responds similarly to the SPI and in accordance with the deciles, even thought it is expected to be a more sensitive index for changing climate.

Recently, Raziei et al. (2011) investigated the effect of changing climate characteristic in Iran, comparing the time series of Standardized Precipitation Index (SPI), Standardized Precipitation Evapotranspiration Index (SPEI) and the Reconnaissance Drought Index (RDI) and found out that they strongly correlate with each other when temperature is low. As the temperature and the potential evapotranspiration increased, the correlation coefficients between SPI an SPEI noticeably decreased.

1.2 The aim of this study

This study investigates the efficiency of several indices in capturing the climate variability on drought characteristics. The comparison is provided for the time period 1960–2002, in 10 watersheds distributed in the region of Thessaly. The 12-month time scale is considered as it better illustrates the long-term behavior of the indices and filtering out high frequency fluctuations. As it has been mentioned previously, a key factor for the two latter indices is the calculation of PET. As a first step, in a watershed in Cyprus with really good quality of data, different methods of PET are performed in order to choose the most accurate. The second step of the study was carried out in selected watersheds located in Thessaly region, Greece, where there is a high complexity of hydrological processes with a marked seasonal and interannual variability. Drought is a recurrent phenomenon in Greece, and especially in Thessaly which is an agricultural region. In this point, it has been made an attempt to replace the potential with the actual evapotranspiration, estimated through the water balance model in the watershed. The objective is to determine the most adequate index to monitor drought to develop a drought preparedness plan in the region.

Chapter 2 . Methodology

2.1 Calculation of Potential Evapotranspiration

Potential evaporation (PET) is one of the main inputs of hydrological models. Yet, there is limited consensus on which PET equation is most applicable in hydrological climate impact assessments. In this study six different methods to derive global scale reference PET monthly time series from data are compared There are approximately 50 methods or models available to estimate PET, but these methods or models give inconsistent values due to their different assumptions and input data requirements, or because they were often developed for specific climatic regions (*Grismer et al.,* 2002). Past studies at multiple scales have suggested that different PET methods may give significantly different results

A major finding is that for part of the investigated basins the selection of a PET method may have only a minor influence on the resulting river flow. Within the hydrological model used in this study the bias related to the PET method tends to decrease while going from PET, AET and runoff to discharge calculations. However, the performance of individual PET methods appears to be spatially variable, which stresses the necessity to select the most accurate and spatially stable PET method. Although often recommended, the Penman-Monteith method was not applied as it has a high data demand equation and also in arid regions the equation has often resulted in relatively low PET values and, consequently, led to relatively high discharge values for dry basins.

2.1.1 Temperature-based methods

Those evaporation (ET) estimation methods that require only temperature as input variable are considered as temperature-based methods. The temperature methods are some of the earliest methods for estimating ET. The relation of ET to air temperature dates back to the 1920s (see Jensen *et al*., 1990). Most temperature-based equations take the form $ET = cT^a$.

2.1.1.1 Thornthwaite method.

A widely used method for estimating potential evapotranspiration was derived by Thornthwaite (1948) who correlated mean monthly temperature with evapotranspiration as determined from water balance for valleys where sufficient moisture water was available to maintain active transpiration. In order to clarify the existing method, the computational steps of Thornthwaite equation are discussed.

The annual value of the heat index I is calculated by summing monthly indices over a 12-month period. The monthly indices are obtained from the equations

$$i = \left(\frac{T_\alpha}{5}\right)^{1.51} \quad (2.1)$$

and

$$I = \sum_{j=1}^{12} i_j \quad (2.2)$$

in which I is the annual heat index, i is the monthly heat index for the month j (which is zero when the mean monthly temperature is 0 ◦C or less), T_a is the mean monthly air temperature (◦C) and j is the number of months (1–12).
The Thornthwaite general equation calculates unadjusted monthly values of potential evapotranspiration, ET' (in mm), based on a standard month of 30 days, 12 h of sunlight/day

$$ET' = C\left(\frac{10T_a}{I}\right)^a \quad (2.3)$$

in which $C = 16$ (a constant) and $\alpha = 67{,}5 \times 10 - 8\mathrm{I}3 - 77{,}1 \times 10 - 6\mathrm{I}2 + 0{,}0179\mathrm{I} + 0{,}492$.

The value of the exponent a in the preceding equation varies from zero to 4,25 (e.g. Jain and Sinai, 1985), the annual heat index varies from zero to 160, and $ET'$is zero for temperature below 0 ◦C.

The unadjusted monthly evapotranspiration values ET' are adjusted depending on the number of days N in a month (1 ≤ N ≤ 31) and the duration of average monthly or daily daylight d (in hours), which is a function of season and latitude.

$$ET = ET'\left(\frac{d}{12}\right)\left(\frac{N}{30}\right) \quad (2.4)$$

in which ET is the adjusted monthly potential evapotranspiration (mm), d is the duration of average monthly daylight (hr); and N is the number of days in a given month, (1–31 days).

Thornthwaite's equation has been widely criticized for its empirical nature but is widely used. Because Thornthwaite's method of estimating ET can be computed using only temperature, it has been one of the most misused empirical equations in arid and semi-arid irrigated areas where the requirement has not been maintained (*Thornthwaite and Mather*, 1955).

2.1.1.2 Blanney–Criddle method.

The Blanney and Criddle (1959) procedure for estimating ET is well known in the western USA and has been used extensively elsewhere also (Singh, 1989). The usual form of the Blaney–Criddle equation converted to metric units is written as

$$ET = kp(0.46T_a + 8.13) \quad (2.5)$$

where ET is evapotranspiration from the reference crop (in mm) for the period in which p is expressed, T_a is mean temperature in °C, p is percentage of total daytime hours for the period used (daily or monthly) out of total daytime hours of the year (365×12), and k is a monthly consumptive use coefficient, depending on vegetation type, location and season. According to Blaney and Criddle (1959) for the growing season (May to October) k varies from 0,5 for orange tree to 0,5 for dense natural vegetation. In this study, an average value of 0,85 will be used for the preliminary comparison.

2.1.1.3 Kharrufa method

Kharrufa (1985) derived an equation through correlation of $ET\ /\ p$ and T in the form of

$$ET = 0{,}34pT_{\alpha}^{1{,}3} \quad (2.6)$$

where ET is the Kharrufa potential evapotranspiration (in mm/month) and T_a and p have the same definitions as given earlier.

2.1.1.4 Hargreaves method.

Hargreaves and Samani (1982, 1985) proposed several improvements for the Hargreaves (1975) equation for estimating grass-related reference ET. Because solar radiation data frequently are not available, Hargreaves and Samani (1982, 1985) recommended estimating R_s from extraterrestrial radiation, R_A, and the difference between mean monthly maximum and minimum temperatures, TD (in °C). The resulting form of the equation is

$$ET = 0.0023 R_A TD^{1/2}(T_a + 17{,}8) \qquad (2.7)$$

The extraterrestrial radiation, R_A, is expressed in equivalent evaporation units. For a given latitude and day R_A is obtained from tables or may be calculated using a set of equations (Jensen et al., 1990). The only variable for a given location and time period is air temperature. Therefore, the Hargreaves method has become a temperature-based method.

2.1.2 <u>Solar and Net Radiation-based methods</u>

Numerous studies have shown that net radiation accounts for most of the variability in evapotranspiration when soil water and vegetative cover are not limiting (*Jensen*, 1967).

2.1.2.1 Turc model

The formula by Turc (1954) reads

$$ET_p = \frac{P+80}{\sqrt{1+\left(\frac{P+45}{L^{Tc}}\right)^2}} \qquad (2.8)$$

where ET_p is the 10-day potential evapotranspiration (mm), P the 10-day precipitation (mm), L^{Tc} the evaporative demand of the atmosphere, calculated as

$$L^{Tc} = \frac{(T_a+2)\sqrt{R_s}}{11{,}1} \qquad (2.9)$$

in which T_a is the average air temperature at 2 m (^oC) and R_s the incoming short-wave radiation (W/m^2).

2.1.2.2 Jensen-Haise model

The Jensen-Haise (1963) formula, with adjusted units, reads

$$ET_p = (0{,}025T_\alpha + 0{,}08)\frac{R_s}{28{,}6} \tag{2.10}$$

where ET_p is the potential evapotranspiration rate (mm/d), R_s and T_α have the same definitions as given earlier.

2.2 Drought assessment

The aim of this study is to evaluate the efficiency of certain meteorological drought indices to monitor hydrological droughts in river discharges and soil moisture in selected watersheds located in the region of Thessaly, Greece. Six indices (i.e. SPI, SPEI, RDI, P/T, modSPEI, modRDI) were used as indicators of meteorological drought severity. The hydrological drought severity was evaluated from the standardized outputs of the monthly UTHBAL conceptual water balance model. The use of simulated discharges was mandatory due to the discontinuous discharge observations. Standardized simulated river discharges were compared with the drought indices to demonstrate the efficacy of drought indices in different climates to monitor hydrological drought. In order to have more adequate results two new indices are introduced using the actual evapotranspiration (AET) instead the potential evapotranspiration (PET) in the calculations. For comparison, the Pearson correlation coefficient was used and the characteristics (i.e. severity and duration) of the identified drought episodes were evaluated and compared. In the next paragraphs the UTHBAL model and the drought indices are presented.

2.2.1 Hydrological drought assessment

Most of the observed discharge time series were intermittent. For this reason, a lumped conceptual water balance model was used to reconstruct and extent the observed runoff data and to produce soil moisture time series. Lumped water balance models has been developed at various time scales (e.g. hourly, daily, monthly and yearly) and at varying degrees of complexity. Monthly water balance models were first developed in the 1940s and have since been adopted, modified, and applied to a wide spectrum of hydrological problems. Recently, these hydrological models have been employed to explore the impact of climatic change. They also have been utilized for long-range streamflow forecasting. Although such applications may use hourly or daily models, these models are, however, more data intensive and have more parameters than the monthly models. A complete review of water balance model applications could be found in Xu and Singh (1998).

In this study the monthly conceptual water balance UTHBAL model (*Loukas et al.*, 2007) has been used. The water balance model allocates the watershed runoff into three components, the surface runoff, the interflow runoff and the baseflow runoff using a soil moisture mechanism with the first priority of the balancing being the fulfillment of actual evapotranspiration. The model separates the total precipitation into rainfall and snowfall, because the correct division of precipitation is essential for accurate runoff simulation. The rain- snow percentage is estimated using a logistic relationship based on mean monthly temperature and the snowmelt is calculated estimated using the simple degree-day method (*WMO*, 1986). The UTHBAL model requires monthly values of mean temperature, precipitation, and potential evapotranspiration and produces values for actual evapotranspiration, soil moisture, groundwater and surface runoff. The input timeseries were estimated using the methods presented in the previous paragraphs. The UTHBAL model has six parameters to be optimized in order to estimate watershed runoff. The optimization was performed using the Generalised Reduced Gradient Algorithm and the Nash-Sutcliffe Model Efficiency was used as the objective function. The UTHBAL model was calibrated with the available observed runoff data to extend, reconstruct and produce runoff and soil moisture timeseries for the period of analysis (1960–2002) and for all study watersheds.

The produced synthetic runoff time series were used for the estimation of hydrological drought. McKee et al. (1993) select the Gamma distribution for fitting monthly precipitation data series, and suggest that the procedure can be applied to other variables relevant to drought, e.g., streamflow or reservoir contents. In pursuing this suggestion for model-based runoff, we note that distributions other than the Gamma may be more appropriate, depending on the runoff variable's sample characteristics (especially skew and kurtosis), which vary greatly with geographic location and degree of temporal aggregation. For example, Shukla and Wood (2008) showed that the two-parameter log normal (LN) distribution provides a better fit perform better for low runoff values for the Feather River basin, at California, USA. López-Moreno et al. (2009) chose the Pearson type III distribution for modeling monthly streamflow in the Tagus basin based on L moments ratios diagrams. In this study, skewness coefficients of simulated discharge values were estimated for all study watersheds for the period of analysis on monthly basis. The skewness test of normality (*Snedecor and Cochran*, 1980) and the Filliben probability plot correlation test (*Filliben,* 1975) were both applied at the 10% significance level for each month (24 tests for each watershed). The results show that for all watersheds and almost all months (except the summer months) are statistically significant at the 10% significance level. To remove skewness, the Gamma transformation was used since it is capable to deal with non-linear data over widely varying hydro-climatic regimes

The transformed runoff values are then standardized to translate into the standard normal distribution as function of *Z*, where *Z* is the variable in normalizing process, using:

$$Z_{WBI} = \frac{Y-\bar{\bar{Y}}}{\sigma_Y} \tag{2.11}$$

where, Z_{WBI} are the values of the standardized time series, $\bar{Y}$ is the mean value of the transformed time series and σ_Y is the standard deviation of the transformed time series. The transformation and standardization have been performed on monthly basis, meaning that the monthly values have been analyzed individually and separately for each month of the year. The transformed and standardized runoff time series were used as indicators of hydrological drought severity and were compared with the other drought indices timeseries estimated by basin-wide meteorological data

This normalization approach is different from the threshold level approach for streamflow deficits (*Tallaksen and van Lanen* 2004) and applied in several studies (e.g. *Hisdal and Tallakse*n, 2003; *Tsakiris et al.,* 2007; *Pandey et al.,* 2008; *Wu et al.,* 2008). Once standardized the strength of the anomaly is classified as set out in Table 2.1. This table also contains the corresponding probabilities of occurrence of each severity level arising naturally from the Normal probability density function. Thus, at a given location for an individual month, moderate dry periods ($Z_{WBI} \leq -1$) have an occurrence probability of 15,9%, whereas, extreme dry periods ($Z_{WBI} \leq -2$) have an event probability of 2,3%. Extreme values in the Z_{WBI} will, by definition, occur with the same frequency at all locations. Negative Z_{WBI} values indicate droughts and positive Z_{WBI} values denote wet weather conditions (*Table 2.1*). Magnitude and duration of hydrological drought, as well as the probability of emergence from drought, are determined from the Z_{WBI} index. One major advantage of the Z_{WBI} is its spatial invariance across different climate regions.

Drought index value	Category	Probability (%)
2.00 or more	Extremely wet	2.3
1.50 to 1.99	Severely wet	4.4
1.00 to 1.49	Moderately wet	9.2
-0.99 to 0.99	Near normal	68.2
-1.49 to -1.00	Moderately dry	9.2
-1.99 to -1.50	Severely dry	4.4
-2 or less	Extremely dry	2.3

Table 2.1: *Drought classification by drought index values and corresponding event probabilities.*

2.2.2 <u>Meteorological drought assessment</u>

Many indices have been used for the identification of more than one type of drought (*Tate and Gustard*, 2000; *Keyantash and Dracup*, 2002) and their categorization may not be appropriate, although it is widely used (*Wilhite and Glantz*, 1985; *AMS*, 2004). In this study six drought indices were used for drought assesment. The first is the Standardized Precipitation Index proposed by McKee and his associates (1993) and the second is the Standardized Precipitation-Evapotranspiration Index recently proposed by Vicente-Serrano et al. (2010). Both indices were calculated for 12- month time scale for the historical period. Precipitation is the main variable explaining the frequency, duration and severity of droughts (*Chang and Cleopa*, 1991; *Heim*, 2002). However, recent studies have shown that the effect of temperature (nevertheless of evapotranspiration also) is significant (*Hu and Willson*, 2000). It has been showed that evaporation and transpiration can consume up to 80% of rainfall, and found that the efficiency of drying due to temperature anomalies is as high as that due to rainfall shortage. Moreover, Syed et al. (2008) showed that precipitation dominates terrestrial water storage variation in the tropics, but evapotranspiration explains the variability at middle latitudes. In addition, studies have shown that anomalous high temperatures related to warming processes have in recent years exacerbated the impact of climatic droughts on water resources (*Nicholls*, 2004; *Cai and Cowan,* 2008).

2.2.2.1 Standardized Precipitation Index

The Standardized Precipitation Index (SPI) has been developed by McKee and his associates (1993) and used to define and monitor droughts. Among others, the US Colorado Climate Center, the US Western Regional Climate Center, and the US National Drought Mitigation Center use the SPI to monitor drought in the United States. SPI can be calculated for multiple time-scales. This is very important because the timescale over which precipitation deficits accumulate functionally separates different types of drought (*McKee et al.*, 1995) and, therefore, allows to quantify the natural lags between precipitation and other water usable sources such as river discharge, soil moisture and reservoir storage. Recent studies have used SPI as indicator of hydrological and water resources variables, like soil moisture, surface runoff and reservoir storage (*Loukas and Vasiliades*, 2005; *Vicente-Serrano and Lopez- Moreno,* 2005).

The SPI is calculated by adjusting the precipitation series to a given probability distribution. Initially, the Gamma distribution was used to calculate the SPI (McKee et al., 1993). Computation of the SPI involves fitting a Gamma probability density function to a given frequency distribution of precipitation totals for a station, area or a watershed. The alpha and beta parameters of the Gamma probability density function are estimated for each watershed, for 12- month scale. The Gamma distribution is defined by its probability density function:

$$g(x) = \frac{1}{\beta^{\alpha}\Gamma(\alpha)} x^{\alpha-1} e^{-x/\beta} \text{ for } x>0 \qquad (2.12)$$

where α, β >0 are the shape and scale parameters respectively, x >0 is the precipitation amount and $\Gamma(\alpha)$ is the gamma function. The unbiased Probability Weighted Moments are used to estimate α and β.

The Gamma distribution is not defined for x=0 and a precipitation distribution may contain zeros. In this study a "naïve" method has been applied. According to this method the null precipitation is substituted with a small amount of precipitation, for example 0,1 mm. This substitution does not affect the distribution of precipitation and circumvent the problem. The error introduced by this method depends on the number of months with null precipitation and it is usually evident for the 1-month precipitation.

The estimated parameters are then used to find the cumulative probability, H(x), of an observed precipitation event for the given month and timescale for the station in question. The cumulative probability, H(x), is then transformed to the standard normal random variable z with mean equal to zero and variance of one, which is the value of the SPI. Once standardized the strength of the anomaly is classified as set out in Table 2.2. This table also contains the corresponding probabilities of occurrence of each severity arising naturally from the Normal probability density function. Thus, at a given location for an individual month, moderate dry periods (SPI ≤ -1) have an occurrence probability of 15.9%, whereas extreme dry periods (SPI ≤ -2) have an event probability of 2.3%. Extreme values in the SPI will, by definition, occur with the same frequency at all locations. Negative SPI values indicate droughts and positive SPI values denote wet weather conditions (*Table 2.2*).

In this study the observed and areal averaged monthly precipitation were used for the estimation of the monthly SPI for the watersheds for 12- month time scales for the historical period (1960-2002). The SPI time series were analyzed in order to identify the frequency of droughts events (moderate, severe and extreme) in the whole time series. Using the threshold level method, the average deficit, maximum severity and maximum duration were identified.

SPI value	Category	Probability (%)
2.00 or more	Extremely wet	2.3
1.50 to 1.99	Severely wet	4.4
1.00 to 1.49	Moderately wet	9.2
-0.99 to 0.99	Near normal	68.2
-1.49 to -1.00	Moderately dry	9.2
-1.99 to -1.50	Severely dry	4.4
-2 or less	Extremely dry	2.3

Table 2.2: *Drought classification by SPI values and corresponding event probabilities.*

2.2.2.2 Standardized Precipitation-Evapotranspiration Index

Recently, Vicente-Serrano et al. (2010) proposed a new multi-scalar drought index. The Standardized Precipitation Evapotranspiration Index (SPEI). It is based on precipitation and temperature data, and has the advantage of combining a multi-scalar character with the capacity to include the effects of temperature variability on drought assessment. The calculation procedure of SPEI is based on this of the original SPI. The SPI is calculated using monthly (or weekly) precipitation as the input data. The SPEI uses the monthly (or weekly) difference between precipitation and Potential Evapotranspiration (PET). This represents a simple climatic water balance (*Thornthwaite,* 1948) which is calculated at different time scales to obtain the SPEI. The first step, calculation of the PET, is difficult because of the involvement of numerous parameters including surface temperature, air humidity, soil incoming radiation, water vapor pressure and ground-atmosphere latent and sensible heat fluxes (*Allen et al.,* 1998). Different methods have been proposed to indirectly estimate the PET from meteorological parameters measured at weather stations. According to data availability, such methods include physically based methods (e.g. the Penman–Monteith method; PM) and models based on empirical relationships, where PET is calculated with fewer data requirements. Although some methods in general provide better results than others for PET quantification (*Droogers and Allen,* 2002), the purpose of including PET in the drought index calculation is to obtain a relative temporal estimation, and therefore the method used to calculate the PET is not critical.

Mavromatis (2007) recently showed that the use of simple or complex methods to calculate the PET provide similar results when a drought index such as the PDSI is calculated. Therefore, a simplier but also accurate approach to calculate PET (*Thornthwaite,* 1948) can be followed, which has the advantage of only requiring data on monthly mean temperature. With a value of PET, the difference between the precipitation (P) and PET for the month i is calculated according to:

$$D_i = P_i - PET_i \tag{2.13}$$

which provides a simple measure of the water surplus or deficit for the analyzed month. The calculated D_i values are aggregated at different time scales, following the same procedure as that for the SPI:

$$D_n^k = \sum_{i=0}^{k-1} P_{n-i} - PET_{n-i} \tag{2.14}$$

Where k (months) is the timescale of the aggregation and n is the calculation month. In quantifying the SPEI, a three parameter distribution is needed to used, since in two parameter distributions the variable x (precipitation) has a lower boundary of zero ($0>x<\infty$), whereas in three parameter distributions x can take values in the range ($\gamma>x<\infty$, where γ is the parameter of origin of the distribution), consequently, x can have negative values, which are common in D series. To model D_i values at different time scales the probability density function of a three parameter Log-logistic distribution are used:

$$f(x) = \frac{\beta}{\alpha}\left(\frac{x-\gamma}{\alpha}\right)^{\beta-1}\left(1+\left(\frac{x-\gamma}{\alpha}\right)^{\beta}\right)^{-2} \tag{2.15}$$

where α, β and γ are scale, shape and origin parameters, respectively, for D values in the range ($\gamma>D<\infty$). Parameters of the Log-logistic distribution can be obtained following different procedures. Among them, the unbiased Probability Weighted Moments can be used and therefore the parameters of the Pearson III distribution can be obtained following Singh et al. (1993):

$$\beta = \frac{2w_1 - w_0}{6w_1 - w_0 - 6w_2} \tag{2.16}$$

$$\alpha = \frac{(w_0 - 2w_1)\beta}{\Gamma(1+1/\beta)\Gamma(1-1/\beta)} \tag{2.17}$$

$$\gamma = w_0 - \alpha\Gamma(1 + 1/\beta)\Gamma(1 - 1/\beta) \tag{2.18}$$

where $\Gamma(\beta)$ is the gamma function of β. The Probability Weighted Moments of order s are calculated as:

$$w_s = \frac{1}{N}\sum_{i=1}^{N}(1 - F_i)^s D_i \tag{2.19}$$

where D_i is the time series of precipitation and F_i is a frequency estimator calculated following the approach of Hosking (1990):

$$F_i = \frac{i - 0.35}{N} \tag{2.20}$$

where i is the range of observations arranged in increasing order, and N is the number of data points.

The Log-logistic distribution adopted for standardizing the D series for all time scales is given by:

$$F(x) = \left[1 + \left(\frac{\alpha}{x-\gamma}\right)^{\beta}\right]^{-1} \tag{2.21}$$

$F(x)$ value is then transformed to a normal variable by means of the following approximation (*Abramowitz and Stegun,* 1965):

$$SPEI = W - \frac{C_0 + C_1 W + C_2 W^2}{1 + d_1 + d_2 W^2 + d_3 W^3} \tag{2.22}$$

where C_0, C_1, C_2, d_1, d_2, d_3 are similar constants as for SPI and W is probability-weighted moments:

$$W = \sqrt{-2\ln(P)} \quad \text{for } P \leq 0.5 \tag{2.23}$$

where P is the probability of exceeding a determined D value, P =1-F(x). If P > 0.5, P is replaced by $1-P$ and the sign of the resultant SPEI is reversed. The constants are: $C_0,$ = 2.515517, $C_1,$ = 0.802853, $C_2,$ = 0.010328, $d_1,$ = 1.432788, $d_2,$ = 0.189269,

d_3 = 0.001308. The average value of SPEI is 0, and the standard deviation is 1. The SPEI is a standardized variable, and it can therefore be compared with other SPEI values over time and space. An SPEI of 0 indicates a value corresponding to 50% of the cumulative probability of *D*, according to a Log-logistic distribution. identified.

2.2.2.3 The Reconnaissance Drought Index (RDI)

After a systematic study of the various indices applied to identify and assess the meteorological drought severity it was concluded that although all these indices were useful, none of them seemed to attract universal applicability.

A new reconnaissance drought identification and assessment index was first presented in the coordinating meeting of MEDROPLAN (*Tsakiris,* 2004), while, a more comprehensive description was presented in other publications (*Tsakiris and Vangelis,* 2005; *Tsakiris et al.*, 2006).

The index, which is referred to as the Reconnaissance Drought Index, RDI, may be calculated by the following expressions: For illustrative purposes the yearly expressions are presented first. The first expression, called the initial value of RDI (α_o), is presented in aggregated form using a monthly time step and may be calculated for each month of the hydrological year or a complete year. The α_o is usually calculated for the ith year in annual basis using the following equation:

$$\alpha_o{}^{(i)} = \sum_{j=1}^{12} P_{ij} / \sum_{j=1}^{12} PET_{ij} \, , i = 1(1)N \; and \; j = 1(1)12 \qquad (2.24)$$

in which P_{ij} and PET_{ij} are the precipitation and potential evapotranspiration of the j_{th} month of the i_{th} year, starting usually from October as it is customary for Mediterranean countries and *N* is the total number of years of the available data.

A second expression, the Normalized RDI (RDI_n) is computed using the following equation for each year, in which it is evident that the parameter αo is the arithmetic mean of αo values calculated for the *N* years of data.

$$RDI_n^{(i)} = \frac{a_o^{(i)}}{\bar{a}_o} - 1 \tag{2.25}$$

The third expression, the Standardized RDI (RDI_{st}), is computed following a similar procedure to the one that is used for the calculation of the SPI: The expression for the Standardized RDI is:

$$RDI_n^{(i)} = \frac{y_k^{(i)} - \bar{y}_k}{\hat{\sigma}_{y_k}} \tag{2.26}$$

in which $y_k^{(i)}$is the $\ln(a_o^{(i)})$, $\bar{y}_k$ is its arithmetic mean and $\hat{\sigma}_{y_k}$ is its standard deviation. It is noted that the above expression is based on the assumption that the α_o values follow a lognormal distribution. The Standardized RDI behaves similar to the SPI and so is the interpretation of results. Therefore, the RDIst can be compared to the same thresholds as the SPI.

The choice of the lognormal distribution is not constraining, so in this study case it has been assumed that it follows a gamma. In all examples analyzed during the establishment of the RDI, the goodness-of-fit tests confirmed that apart from the lognormal distribution, gamma distribution also fits the data satisfactorily .

It should be emphasized that the RDI is based both on precipitation and on potential evapotranspiration. The mean initial index $(\bar{a}_o)$ represents the normal climatic conditions of the area and is equal to the Aridity Index as was proposed by the FAO.

Among others, some of the advantages of the RDI are as follows:

- It is physically sound, since it calculates the aggregated deficit between precipitation and the evaporative demand of the atmosphere.
- It can be calculated for any period of time (e.g., 1 month, 2 months etc).
- The calculation always leads to a meaningful figure.
- It can be effectively associated with agricultural drought.
- It is directly linked to the climatic conditions of the region, since for the yearly value it can be compared with the FAO Aridity Index.
- It can be used under “climate instability” conditions, for examining the significance of various changes of climatic factors related to water scarcity.

From the above advantages, it can be concluded that the RDI is an ideal index for the reconnaissance assessment of drought severity for general use, giving comparable results within a large geographical area, such as the Mediterranean.

It should be mentioned that usually droughts in the Mediterranean are accompanied by high temperatures, which lead to higher evapotranspiration rates. Evidence for this has been produced from simultaneous monthly data of precipitation and evapotranspiration in many Greek watersheds. From the analyzed cases, it seems that about 90% of them comply with the previous statement (*Tsakiris and Vangelis,* 2005). Therefore, the RDI is expected to be more sensitive index than those related only to precipitation, such as the SPI.

The RDI can be calculated for any period of time from 1 month to the entire year, even starting from a month different than October, which is customary for the Mediterranean. Very significant results can be derived if the period of analysis coincides with the growing season of the main crops of the area under study or other periods related to sensitive stages of crop growth. Then, the RDI can be associated successfully with the expected loss in rainfed crop production, which in turn is linked to the anticipated hazard in the agricultural sector due to drought occurrence.

As it was shown from previous studies, precipitation (and therefore the SPI) was not successfully correlated to agricultural production (*Tsakiris and Vangelis,* 2005). However, the inclusion of potential evapotranspiration (PET) in the calculation of the RDI enhances its validity in studies aiming at risk assessment in agriculture caused by drought occurrence.

Likewise, PET may be a representative quantity of the consumption in various sectors apart from agriculture. Water demand is increasing in general in case of higher temperatures. Therefore, the RDI could be modified to be used in the future as an indicator for the drought risk assessment related to the various sectors of water use.

2.2.2.4 The ratio between precipitation and temperature

The ratio between the precipitation and the temperature of the dataset has been chosen as an index to test how it gets influenced from the climate variability and also how it performs for the drought assessment, compared with the other drought indices.

The SPI is calculated by adjusting the precipitation series to a given probability distribution. Initially, the Gamma distribution was used to calculate the SPI (*McKee et al.,* 1993). Computation of the SPI involves fitting a Gamma probability density function to a given frequency distribution of precipitation totals for a station, area or a watershed. The alpha and beta parameters of the Gamma probability density function are estimated for each watershed, for 12- month scale. The Gamma distribution is defined by its probability density function:

$$g(x) = \frac{1}{\beta^{\alpha}\Gamma(\alpha)} x^{\alpha-1} e^{-x/\beta} \quad \text{for } x>0 \qquad (2.27)$$

where α, β >0 are the shape and scale parameters respectively, x >0 is the precipitation amount and $\Gamma(\alpha)$ is the gamma function. The unbiased Probability Weighted Moments are used to estimate α and β.

The Gamma distribution is not defined for x=0 and a precipitation distribution may contain zeros. In this study a "naïve" method has been applied. According to this method the null precipitation is substituted with a small amount of precipitation, for example 0,1 mm. This substitution does not affect the distribution of precipitation and circumvent the problem. The error introduced by this method depends on the number of months with null precipitation and it is usually evident for the 1-month precipitation.

The estimated parameters are then used to find the cumulative probability, H(x), of an observed precipitation event for the given month and timescale for the station in question. The cumulative probability, H(x), is then transformed to the standard normal random variable z with mean equal to zero and variance of one, which is the value of the index (P/T). Once standardized the strength of the anomaly is classified as set out in Table 2.1. This index has the same methodology and also the same interpretation as the SPI index.

2.2.2.5 The modified drought indices using AET instead of PET

The calculation of the PET, which is required for computation of both RDI and SPEI, has to reflect climatic conditions and not vegetation/crop potential demand like when using a reference ET equation. Regardless the accuracy of the method of calculation for potential evapotranspiration (PET), the result may be not reliable for drought assessment as it is not the actual evapotranspiration (AET). Using the water balance model in each watershed, we estimated the actual evapotranspiration (AET) and in the two indices the factor PET has been replaced. So, for the new two indices (modSPEI, modRDI), the new variables are the following:

$$D_i' = P_i - AET_i \quad (2.28)$$

$$\alpha_o'^{(i)} = \sum_{j=1}^{12} P_{ij} / \sum_{j=1}^{12} AET_{ij}, i = 1(1)N \text{ and } j = 1(1)12 \quad (2.29)$$

The SPI, SPEI, RDI, P/T, modSPEI, and modRDI time series were calculated for the observed and stochastic time series for the historical period for 12- month time scale. For the estimation of the SPEI the time series of PET, which were calculated with the method of Thornthwaite from the observed and synthetic time series of temperature, after a test in a different watershed with a lot of methods. Furthermore the average deficit, maximum severity and maximum duration were identified using the threshold level method.

2.3 The threshold level method

As it has been mentioned previously, drought is a regional phenomenon characterized by three dimensions; severity or intensity, the duration and the areal extent (Rossi et al, 1992). There are two main methods to select and characterize deficits, namely the threshold level method and the sequent peak algorithm. The threshold level method which was initially named method of crossing theory (*Tallaksen,* 2000), is the most frequently applied quantitative definition of a drought is based on defining a threshold, Q_0, below which the river flow is considered as a drought (also referred to as a low flow spell in the literature). It is also referred to as run sum analysis because it generally study runs below or above a given threshold. The method is relevant for storage/yield analysis and is associated with hydrological design and operation of reservoir storage systems. Important areas of application are hydropower and water management, water supply systems and irrigation schemes.

The threshold level Q_0 is also referred to as the truncation level and is used to define whether the flow in a river is in deficit. The deficit starts when the flow goes below the threshold and ends as soon as the flow returns above the threshold. Thus, the beginning and the end of a deficit can be defined. In addition the following deficit characteristics can be defined:

- The duration, which is the period of time where the flow is below the threshold level and is also referred to as drought duration, low-flow spell or run length (d_i);
- The volume or cumulative severity, which is also referred to as drought volume or run sum (v_i);
- The intensity, which is also referred to as deficit or drought severity/ magnitude, (m_i) is the ratio between deficit volume and deficit duration;
- The minimum value of each deficit event (Q_{min});
- The end of occurrence, for example, the starting date, the mean of the onset and termination, or the date of minimum value.

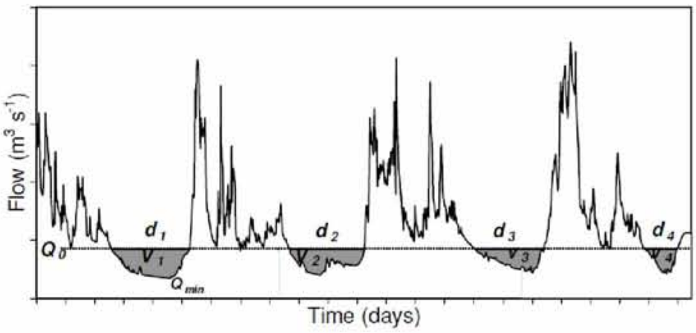

Figure 2.1: *Definition of low flow and drought characteristics (modified from Tallaksen, 2000).*

Based on the time series of the deficit characteristics, it is possible to determine indices, such as the average deficit duration or average deficit volume. The threshold might be chosen in a number of ways and the choice is amongst other a function of the type of water deficit to be studied (*Dracup et al.,* 1980). The choice is influenced by the purpose and region of the study and the data availability.

In each watershed and for each method for drought assessment, the method of threshold has been applied in order to identify the average deficit, maximum severity and maximum duration for the observed time series of all the indices as well as for the synthetic time series of precipitation, for 12- month time scale for the base period (1960-2002). The threshold level was chosen to have the value of -1, since the indices values bellow -1 (*Table 2.1,2*) identify drought event.

Chapter 3 . Study Area and Database

3.1 Yermasoyia

The first study basin is the Yermasoyia watershed, which is located in the southern side of mountain Troodos of Cyprus, roughly 5 Km north of Limassol city (*Fig. 3.1*). The watershed area is 156.7 Km^2 and its elevation ranges from 70 m up to 1400 m. Most of the area is covered by typical Mediterranean type forest and sparse vegetation. A reservoir with storage capacity of 13,6 million m^3 was constructed downstream the mouth of the watershed in 1969, for irrigation and municipal water supply purposes. The climate of the area is of Mediterranean maritime climate with mild winters and hot and dry summers. The precipitation is usually generated by frontal weather systems moving eastwards. The average basin wide annual precipitation is 640 mm, ranging from 450 mm at the low elevations up to 850 mm at the upper parts of the watershed. The mean annual runoff of Yermasoyia river is about 150 mm, and 65% of it is allocated during winter months. The river is usually dry during summer months *(Loukas et al.,* 2003). The peak flows are observed in winter months and produced by rainfall events. Good quality daily precipitation from three meteorological stations located at 70 m, 100 m, and 995m of elevation are used.

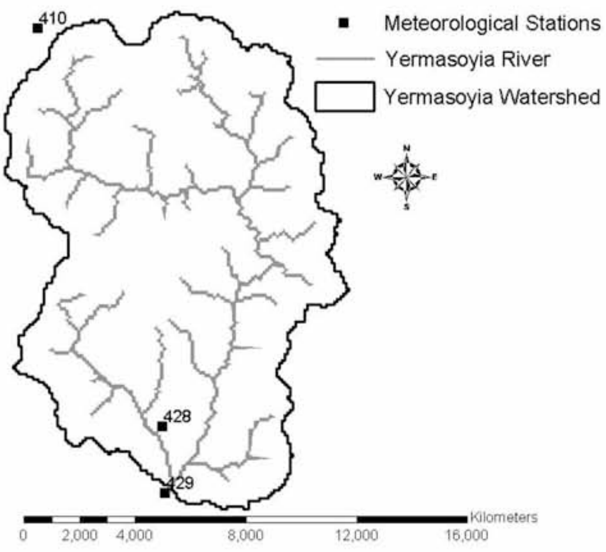

Figure 3.1: *The Yermasoyia watershed and the monitoring stations within the watershed*

Data of maximum and minimum temperature measured at the low elevation station are used in this study. Streamflow is measured at the mouth of the watershed (*Fig.3.1*). In total, eleven years of reliable data (1986-1997) were available from the Yermasoyia watershed. This watershed was chosen in order to check the most reliable model for the calculation of Potential Evapotranspiration (PET). Even thought the dataset is really short (only eleven years, shorter than thirty years a well defined dataset) the validity of this is a really important factor for choosing the most appropriate method of analysis.

3.2 Thessaly

3.2.1 Description

The second study area, Thessaly, is a plain region surrounded by high mountains. Thessaly plain is one of the most productive agricultural regions of Greece. The main crops cultivated in the plain area are cotton, wheat and maize whereas apple, apricot, cherry, olive trees and grapes are cultivated at the foothills of the eastern mountains. Pinios River and its tributaries traverse the plain area, and the basin total drainage area is about 9.500 km^2. The waters of the Pinios River are used primarily for irrigation. Climate is continental at the western and central side of Thessaly and Mediterranean at the eastern side. Winters are cold and wet and summers are hot and dry with a large temperature difference between the two seasons. Mean annual precipitation over the Thessaly region is about 700 mm, varies from about 450 mm at the central plain area to more than 1850 mm at the western mountain peaks and it is distributed unevenly in space and time. Generally, rainfall is rare from June to August. The mountain areas receive significant amounts of snow during the winter months and transient snowpacks develop.

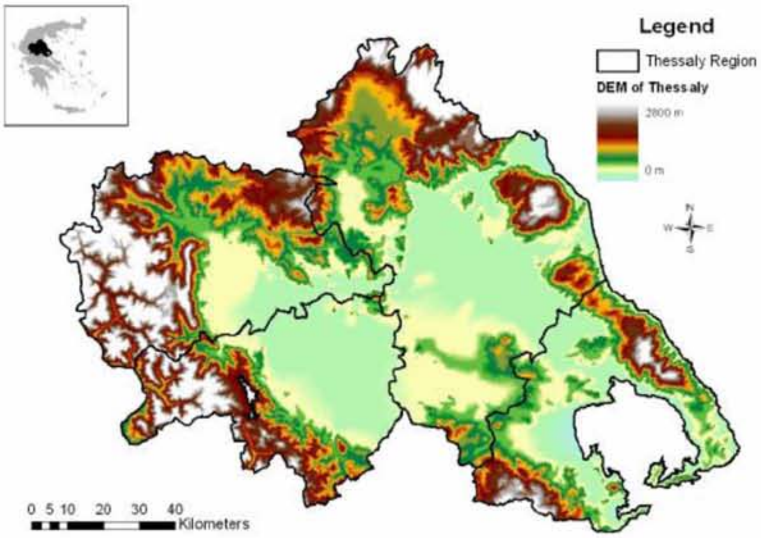

Figure 3.2: *Digital Elevation model of the region of Thessaly, Greece.*

Greece, and especially Thessaly, experienced severe, extreme and persistent droughts during the periods from the mid to late 1970s, from the late 1980s to the early 1990s and the first years of 2000s. These three drought periods were quite remarkable and affected large areas. The first drought episode (1976–1977) affected southern and western Europe, the second drought episode (1988–1991) affected the whole Mediterranean region with an estimated economic cost lager than €2.1 billion, whereas the third drought episode (2000–2001) affected central Europe and the Balkans with total damage of €0.5 billion (*European Environmental Agency,* 2004).

During these three periods the monthly and annual precipitation was significantly below normal in Thessaly. Especially, the hydrological years 1976–1977 and 1989 1990 are the first and second driest hydrological years on record for annual precipitation was significantly below normal in Thessaly. Especially, the hydrological years 1976–1977 and 1989–1990 are the first and second driest hydrological years on record for Thessaly, respectively (*Loukas et al.,* 2004). The annual precipitation for the 1976–1977 year was 467 mm and for the year 1989–1990 was 521 mm. The driest January and February and the second driest March in record occurred during the hydrological year 1989–1990. The prolonged and significant decrease of monthly and annual precipitation has a dramatic impact on water resources of the region. Usually, the dry periods are accompanied with high temperatures, which lead to higher evapotranspiration rates and dry soils. These conditions inversely affect both the natural vegetation and the agriculture of the region as well as the available storage of the reservoirs. Severe and extremely dry conditions result in irrigation cutbacks, overexploitation of groundwater and significant losses of crop yields. For example, during the dry hydrological year 1989–1990, the water supply for irrigation from the N. Plastiras reservoir, the only large reservoir in operation in the area at that time, was cut-off by more than 70% and the irrigated areas with surface water resources were reduced by 90% (*Loukas et al.,* 2004). In general, Thessaly is region with a lot of interest because it combines extremely wet and extremely dry areas, with really steep seasonal temperature changes, variable geomorphology and high demands in irrigation (and evapotranspiration) because of agriculture.

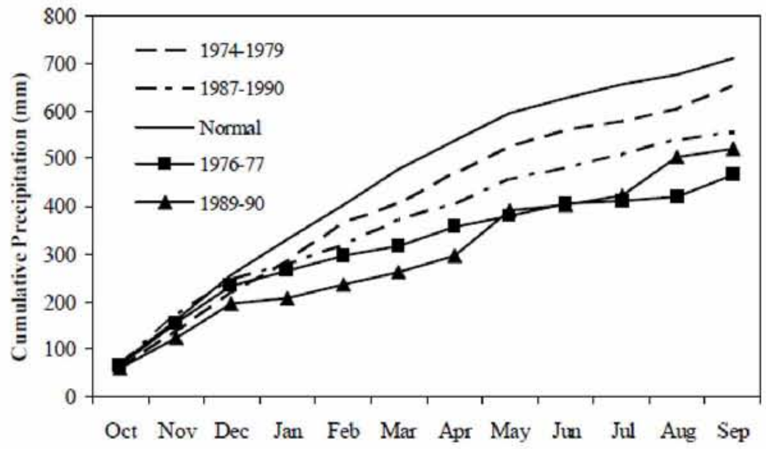

Figure 3.3: *Cumulative areal precipitation for selected dry years and periods.*

3.2.2 <u>Database</u>

Ten small, medium and large watersheds with areas ranging from 133 to 6591 km^2 located in the region of Thessaly, and most of them being tributaries of the Pinios River, have been selected to demonstrate the efficacy of drought indices in different climates to monitor hydrological drought and also to check the importance of using the actual evapotranspiration (AET) in the procedure. Processed monthly precipitation data from a large network of precipitation for the period October 1960 to September 2002 were used. However, many hydrological applications require the average depth of rainfall occurring over an area which can then be compared directly with runoff from that area. Hence, average areal rainfall is required. The areal precipitation has been estimated using the methods of Thiessen polygon and the precipitation gradient method. Similarly, for the estimation of the temperature of each watershed, the temperature gradient was used, using monthly data from the available meteorological stations for the base period.

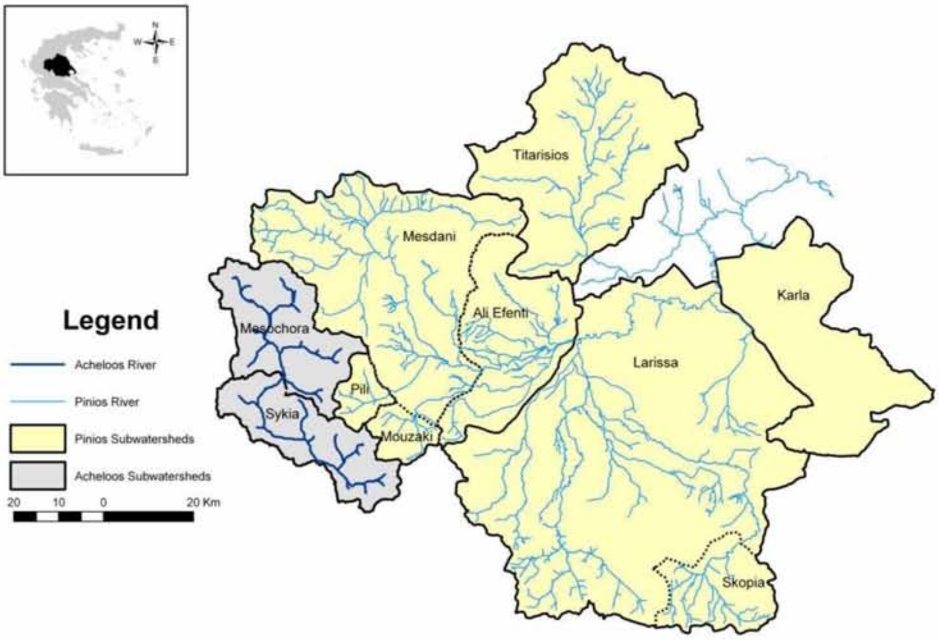

Figure 3.4: *The Watersheds with the river tributaries*

	Area (km^2)	Elevation range (m)	Mean elevation (m)	Mean annual precipitation (mm)	Mean annual runoff (mm)	Runoff coefficient
Mesochora	615	700-2300	1400	1860	1150	0,62
Sykia	1155	500-2300	1288	1875	1290	0,69
Mouzaki	145	200-2000	838	1450	830	0,57
Pili	133	300-1800	949	1820	1130	0,62
Tirtarisios	1520	100-2800	695	534	113	0,21
Ali Efenti	2869	150-2000	555	890	420	0,47
Mesdani	2055	180-2000	660	985	600	0,61
Skopia	410	300-1600	657	594	126	0,21
Karla	1171	50-1900	230	560	52	0,09
Larissa	6591	70-2000	451	720	290	0,40

Table 3.1: *The characteristics of the watersheds*

3.2.2.1 Estimation of areal precipitation

The widely-used method for estimation of the areal precipitation is the Thiessen method (1911). The Thiessen polygon method accounts for the variability in spatial distribution of gauges and the consequent variable area which each gauge represents. The areas representing each gauge are defined by drawing lines between adjacent stations on a map. The perpendicular bisectors of these lines form a pattern of polygons (Thiessen polygons) with one station in each polygon. Stations outside the basin boundary should be included in the analysis as they may have polygons which extend into the basin area. The area of a polygon for an individual station as a proportion of the total basin area represents the Thiessen weight for that station. Areal rainfall is thus estimated by first multiplying individual station totals by their Thiessen weights and then summing the weighted totals as follows:

$$Po = \sum \frac{(Ai \times Pi)}{A} = \sum \left(\frac{Ai \times Pi}{A} \right) \tag{3.1}$$

where A_i is the area of Thiessen polygon for station i, A the total area under consideration, P_i is the monthly precipitation for station i and P_0 the areal precipitation. In order to estimate the areal precipitation with this method, precipitation data from more than 50 raingages were used. Stations outside the watershed included in the analysis since they affected the basin area.

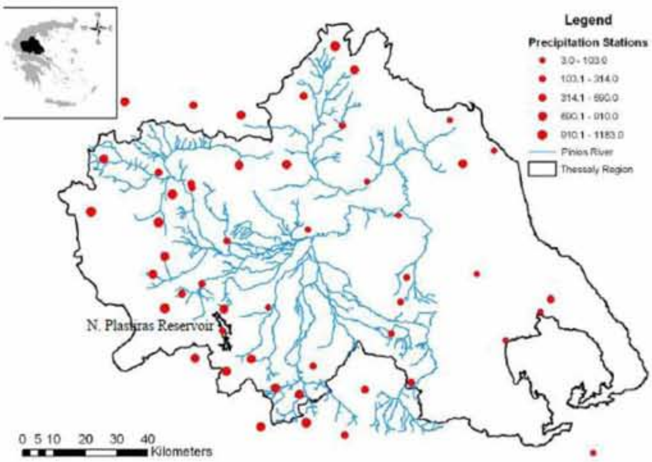

***Figure 3.5**: Location and elevation of the precipitation stations used in this study.*

This method is not ideal for mountainous areas where orographic effects are significant or where rain gauges are predominantly located at lower elevations of the basin. Therefore the precipitation gradient method was used. The method is based on the assumption that rainfall increases as the elevation increases. It uses a linear relationship between elevation and precipitation and since a satisfactory correlation exists, it can be used in order to estimate rainfall at any elevation

The first step was to find the linear relationship between the elevation of each station and the mean annual precipitation of each station. The coefficient of determination was not satisfied using the data from the stations that used for the Thiessen method. Hence, more than 50 precipitation stations were included in the linear relationship.

Thereafter, the mean annual precipitation of the watershed was estimated by the following equation:

$$P_k = P_{T(k)} - \frac{(Z_T - Z)b}{100} \tag{3.2}$$

where P_k is the mean annual areal precipitation of the watershed, $P_{T(k)}$ is the mean annual precipitation estimated with the Thiessen method, Z_T is the weighted mean elevation of the watershed estimated by Thiessen, Z is the elevation of the watershed estimated with the method of gradients and b is the slope of the linear relationship *Fig. 3.6*). The mean monthly precipitation is estimated by:

$$P_i^k = \frac{P_k P_{T(i)}^k}{P_{T(k)}} \tag{3.3}$$

where P_i^k is the monthly precipitation at the *i* month and *k* year, P_k is the mean annual areal precipitation of the watershed, $P_{T(i)}^k$ is the monthly precipitation at the *i* month and *k* year estimated with the Thiessen method and , $P_{T(k)}$ is the mean annual precipitation estimated with the Thiessen method.

With the above method the mean monthly areal precipitation over each watershed was estimated (*Fig. 3.7*).

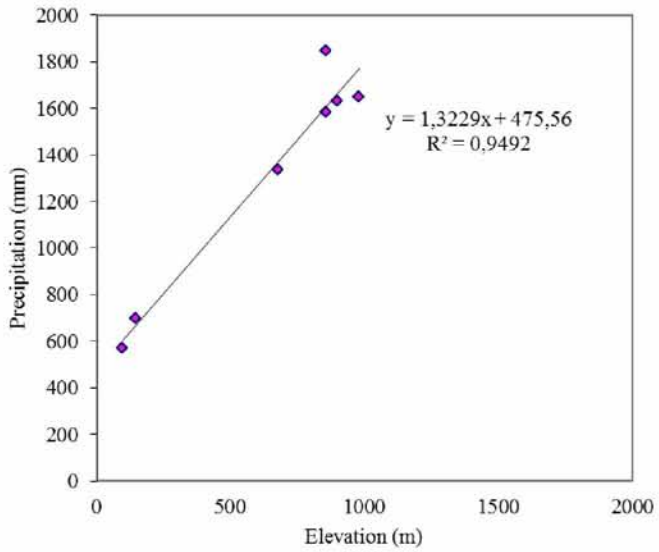

Figure 3.6: *Precipitation gradient of Pili watershed.*

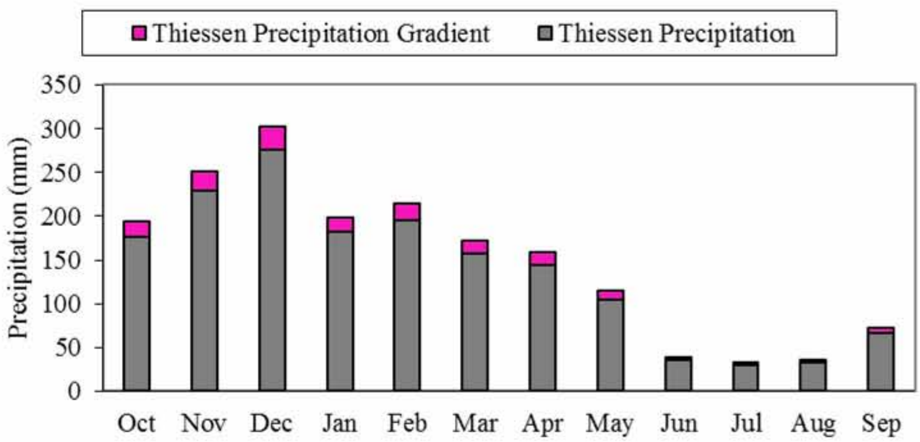

Figure 3.7: *Monthly areal precipitation estimated with the method of Thiessen and the precipitation gradient method for the period (1960-2002) for Pili watershed.*

3.2.2.2 Estimation of areal temperature

The temperature gradient method was used for estimation of the areal temperature. According to this method the temperature decreases as the elevation increases every 100m. In order to estimate the mean annual and monthly temperature, the linear relationship between the elevation of each station and the mean annual temperature of each station should be found (*Fig.3.8*).

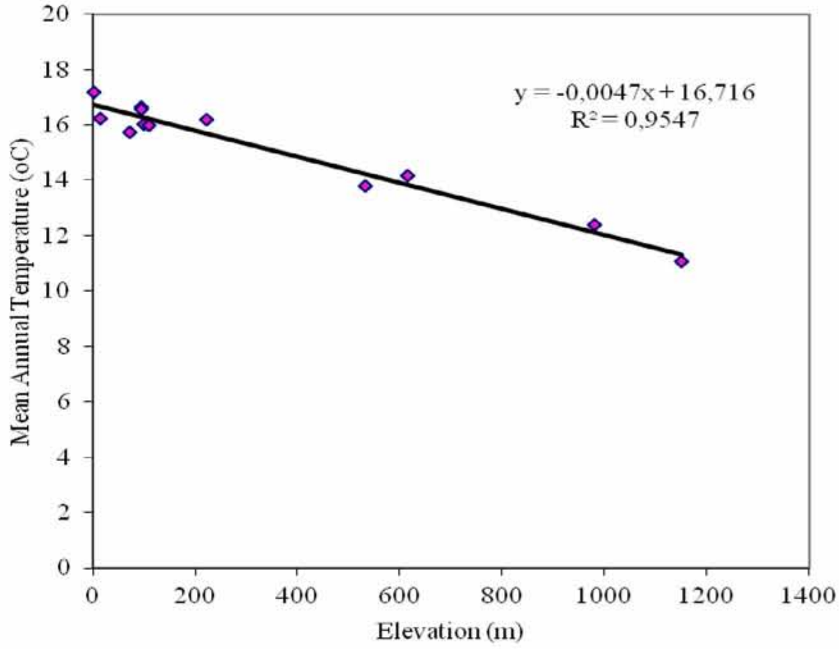

Figure 3.8: *Precipitation gradient of Pili watershed.*

The mean annual and monthly temperature according to the temperature gradient method is estimated using temperature data from a base station. Therefore, the temperature can be defined as:

$$T_k = T_{T(k)} - \frac{(Z_b - Z)b}{100} \qquad (3.4)$$

$$T_i^k = \frac{T_k {T_{T(i)}}^k}{T_{T(k)}} \qquad (3.5)$$

where $T_{T(k)}$ is the mean annual temperature at the base station at year *k*, T_k the mean annual temperature of the watershed at year *k*, T_i^k the mean monthly temperature of the watershed at month *i* and year *k*, ${T_{T(i)}}^k$ the mean monthly temperature of the base station at month *i* and year *k*, *b* is the slope of the linear regression, Z_b is the elevation of the base station and Z is the elevation of the watershed. Figure 3.9 indicates the areal temperature estimated with the temperature gradient method as well as the temperature of the base station.

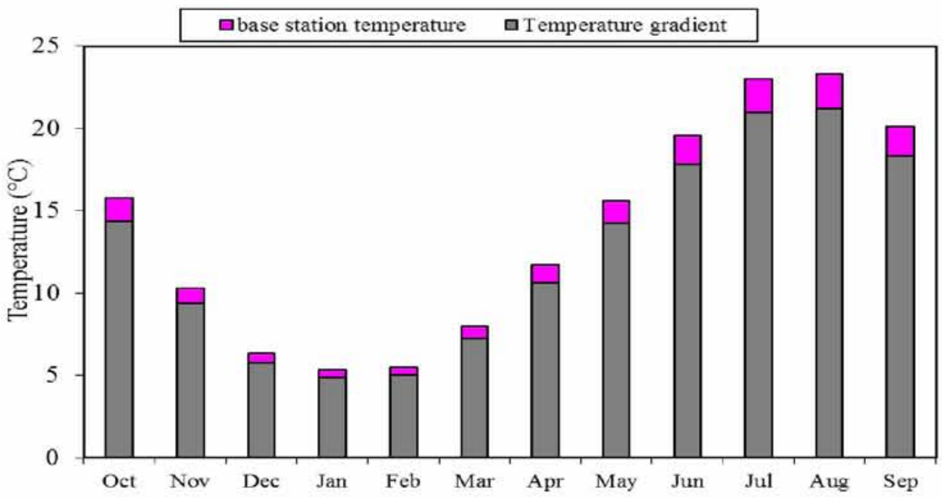

Figure 3.9: *Monthly areal temperature of the base station and Pili watershed estimated with the temperature gradient method for the period (1960-2002).*

3.2.2.3 Estimation of Potential Evapotranspiration.

A lot of different methods have been proposed to estimate the PET from meteorological parameters. After research in another watershed where the dataset was accurate and valid, Thornthwaite (1948) has been chosen. In this test, the methods Hargreaves, Thornwaite, Kharrufa, Blanney, Turc, Jensen-Haise were applied in the dataset and for each of the time series of PET, the original indices were calculated and then the threshold level method applied. For different levels of thresholds the results of the indices compared to each other and also with the results using the actual evapotranspiration. The Thornwaite method had almost the most "stable" results and also high correlation between the indices. Using only monthly mean temperature, it is one of the simpliest methods which is also considered to give the most close to actual evapotranspiration (AET) results. Following this method monthly PET (mm) is obtained by:

$$PET = 16L_d\left(\frac{10T_a}{I}\right)^a \tag{3.6}$$

where T is the monthly mean temperature in °C; I is a heat index, which is calculated as the sum of 12 monthly index values i, the latter being derived from mean monthly temperature using the formula:

$$i = \left(\frac{T}{5}\right)^{1.514} \tag{3.7}$$

a is a coefficient depending on I: $a = 6.75E^{-7}I^3 - 7.71E^{-5}I^2 + 1.79E^{-2}I + 0.492$ and $L_d$is a correction coefficient computed as a function of the latitude and month.

Therefore using the monthly temperature estimated with the temperature gradient method, monthly PET was calculated for each of the watersheds with the method of Thornwaite.

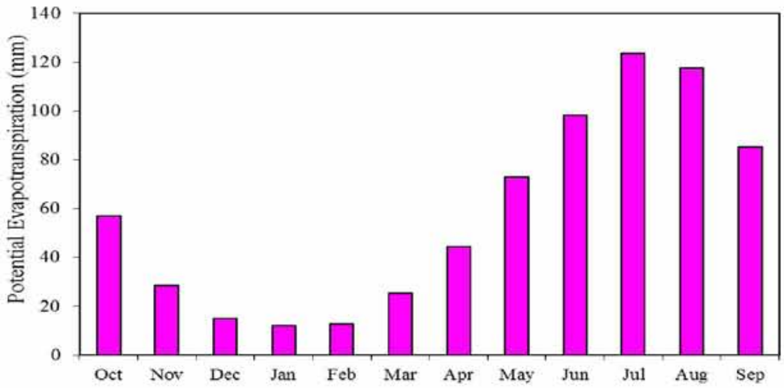

Figure 3.10: *Monthly Potential Evapotranspiration in Pili watershed estimated with the method of Thornthwaite for the period (1960-2002).*

Chapter 4 . Application - Results

4.1 Results from Yermasoyia

Firstly, the method of calculation for PET had to be determined. For this reason, the dataset of Yermasoyia, Cyprus was the ideal one as it has really good quality of data. In this test, the methods Hargreaves, Thornwaite, Kharrufa, Blanney, Turc, Jensen-Haise were applied in the dataset and for each of the time series of PET, the original indices were calculated and then the threshold level method applied. For different levels of thresholds the results of the indices compared to each other and also with the results using the actual evapotranspiration (*Table 4.1*)

Hargreaves	Thornwaite	Kharrufa	Blanney	Turc	Jensen-Haise
0,410	0,481	0,088	0,496	0,408	0,129

Table 4.1: *The correlation coefficients between actual and potential evapotranspiration with different methods of estimation*

As it is obvious from the correlation matrix, the most adequate methods seem to be the Hargreaves, Blanney-Criddle and Thornwaite. In general those three methods are commonly recommended. As it is well- known, Hargreaves and Blanney-Criddle methods perform really well in humid regions where the results they give are close to the lysimeter. Also, Turc (which performs better for annual data) and Jensen-Haise equations generally underestimate ET, during spring, and overestimate it during summer, because T, is given too much weight and R, too little. At last, Kharrufa method sometimes gives a lot of bias in several of cases.

Comparing between Thornwaite and Blanney-Criddle, the results of the level threshold method are almost the same (*Fig.4.2*). Mimikou et. a1 (1991) used the Blanney-Criddle method for estimating potential evapotranspiration in combination with a monthly water balance for basins in the central mountainous regions in Greece. But it has to be mentioned that the regions were humid and small.

Thornwaite				Blanney- Criddle		
Volume	Duration	Date	TLM - RDI	Volume	Duration	Date
0,44	1	1-Jan-90		0,46	1	1-Jan-90
0,61	3	1-Dec-90		0,66	4	1-Dec-90
3,18	9	1-Oct-91		3,32	9	1-Oct-91
0,13	1	1-Dec-93		0,12	1	1-Dec-93
0,73	4	1-Feb-96		0,73	4	1-Feb-96

Thornwaite				Blanney- Criddle		
Volume	Duration	Date	TLM - SPEI	Volume	Duration	Date
0,44	1	1-Jan-90		0,45	1	1-Jan-90
0,57	3	1-Dec-90		0,62	3	1-Dec-90
3,22	9	1-Oct-91		3,34	9	1-Oct-91
0,13	1	1-Dec-93		0,12	1	1-Dec-93
0,79	4	1-Feb-96		0,75	4	1-Feb-96

Table 4.2: *The results of the Threshold Level Method with different methods of calculation of PET*

The Thornwaite method had almost the most adequate results and also resulted high correlation between the indices. Also, as it has been mentioned previously, Thornwaite results PET which behaves the same way with AET especially for annual time scale.

4.2 Results from Thessaly

In the frame of this study, the SPI, SPEI, RDI, P/T, modSPEI and modRDI on 12-month time scale have been computed for all of the 10 watersheds as the hydrological Z_{WBI} index. Since the time scale has been determined 12 –months, the Z_{WBI} does not perform so well. This might happen because this time scale is not representative for this index. Nevertheless, this index can be used for comparison with the meteorological indices. For illustrative purposes, only three of the watersheds with completely different climate conditions are presented here: Sykia, Tirtarisios, Karla (wet, semi-arid, arid) (*Fig 4.1*). As it has been illustrated, even though there are huge differences in the precipitation amount, the amount of the potential evapotranspiration is almost the same at the three watersheds. For sure, there is a temperature increase but it does not trigger an enormous increase in potential evapotranspiration. For sure, this has to do with the vegetation of the region and also the river discharges.

Figure 4.1: *The hydrological cycle components for each month for (a) Sykia, (b) Tirtarisios and (c) Karla watershed for the whole time period (1960- 2002)*

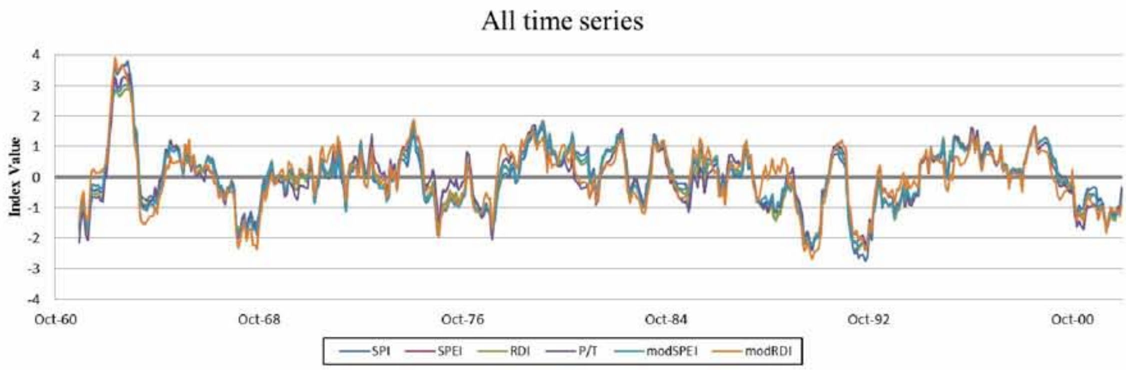

Figure 4.2: *The whole time series of the meteorological indices for Sykia watershed*

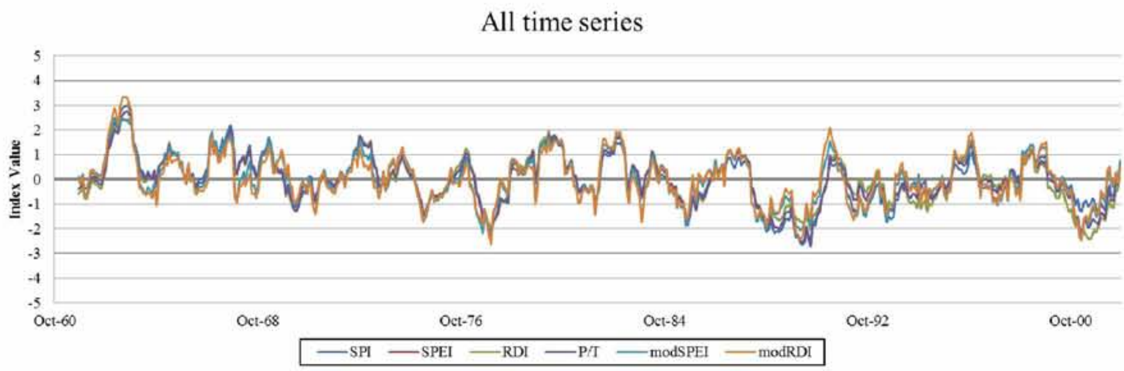

Figure 4.3: *The whole time series of the meteorological indices for Tirtarisios watershed*

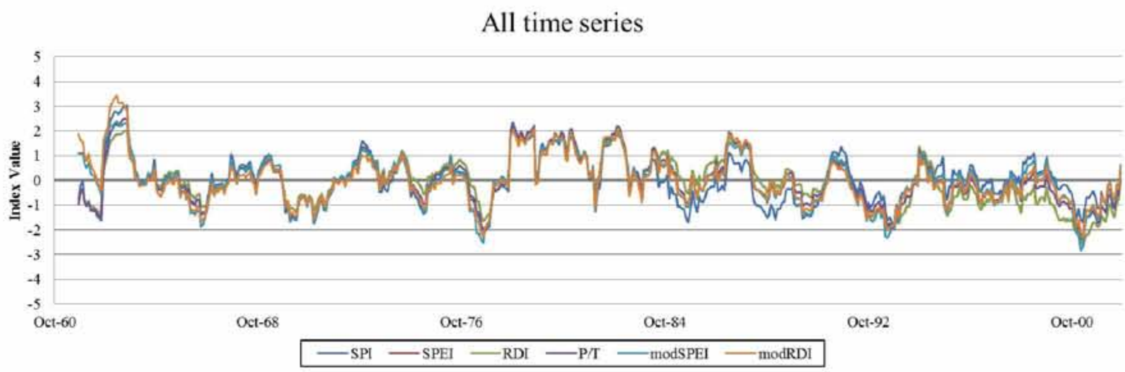

Figure 4.4: *The whole time series of the meteorological indices for Karla watershed*

From a quick view at the plots (*Fig.4.2-4*), results suggest that the time behaviors of the meteorological indices are in close agreement in all the watersheds. This is clearly evident from the correlation matrices where the correlation coefficients varied between 0,86 and 0,99 (*Table 4.1-3*) apart from Karla watershed where the temperatures are high and evapotranspiration exceeds the precipitation, where there are discrepancies between (SPI) and (SPEI), (RDI). Even though the correlation between the indices remains strong, it is much weaker as the temperature increases. It is not so clear how the system responses between an arid and semi- arid region (Tirtarisios and Karla). It is not evident if the temperature is the main factor that triggers the indices of the features like the difference (P-PET) or the ratio (P/PET).

	SPI_12	**SPEI_12**	**RDI_12**	**P/T_12**	**modSPEI_12**	**modRDI_12**
SPI_12	1					
SPEI_12	0,984	1				
RDI_12	0,939	0,973	1			
P/T_12	0,940	0,972	0,998	1		
modSPEI_12	0,987	0,992	0,952	0,952	1	
modRDI_12	0,902	0,909	0,914	0,919	0,938	1

Table 4.3: *The correlation matrix of the indices for Sykia watershed*

	SPI_12	**SPEI_12**	**RDI_12**	**P/T_12**	**modSPEI_12**	**modRDI_12**
SPI_12	1					
SPEI_12	0,929	1				
RDI_12	0,964	0,989	1			
P/T_12	0,973	0,984	0,998	1		
modSPEI_12	0,921	0,906	0,919	0,926	1	
modRDI_12	0,872	0,859	0,875	0,882	0,981	1

Table 4.4: *The correlation matrix of the indices for Tirtarisios watershed*

	SPI_12	SPEI_12	RDI_12	P/T_12	modSPEI_12	modRDI_12
SPI_12	1					
SPEI_12	0,793	1				
RDI_12	0,879	0,984	1			
P/T_12	0,918	0,961	0,993	1		
modSPEI_12	0,911	0,880	0,925	0,945	1	
modRDI_12	0,887	0,862	0,909	0,930	0,979	1

Table 4.5: *The correlation matrix of the indices for Karla watershed*

Although, the good results, the application of the threshold level method present some really interesting points for further analysis(*Table 4.4-6*). There are differencies at the magnitude and also the severity of the event. Also, there is not a clear view for the temperature impact in this behavior since One point that it is really remarkable is the fact that for all the indices and all the water sheds the percentage of the drought period is around 16%, which is equal to the corresponding event probability (*Table 2.2*).

	Max.Deficit	Total vol.	Duration	%Drought Days
SPI_12	15,51	41,24	69	13,69
SPEI_12	11,67	39,39	76	15,08
RDI_12	8,88	38,19	73	14,48
P/T_12	8,90	37,48	69	13,69
modSPEI_12	12,08	39,13	78	15,48
modRDI_12	13,23	46,04	72	14,29

Table 4.6: *The threshold level method for Sykia watershed*

	Max.Deficit	Total vol.	Duration	%Drought Days
SPI_12	28,15	42,15	73	14,48
SPEI_12	16,66	39,55	78	15,48
RDI_12	23,00	45,94	72	14,29
P/T_12	23,50	44,99	73	14,48
modSPEI_12	7,86	37,52	74	14,68
modRDI_12	7,62	33,66	68	13,49

Table 4.7: *The threshold level method for Tirtarisios watershed*

	Max.Deficit	Total vol.	Duration	%Drought Days
SPI_12	5,81	27,30	77	15,28
SPEI_12	20,53	37,55	78	15,48
RDI_12	12,47	28,57	73	14,48
P/T_12	10,17	27,47	67	13,29
modSPEI_12	10,67	39,92	74	14,68
modRDI_12	7,74	27,34	67	13,29

Table 4.8: *The threshold level method for Karla watershed*

The temporal evolution of the drought indices also showed that significant drought episodes have been occurred in the hydrological years Oct 1976- Sep 1978, Oct1987- Sep 1990 and Oct 2000- Sep 2002 for all study watersheds. For this reason, it is really important to insert into these periods and investigate more the evolution of the indices (*Fig.4.5-7*).

With this further analysis several discrepancies are noticeable on the magnitude of several dry/wet events, especially in the warmer regions (Tirtarisios and Karla). This finding suggests that, as expected, correlation is not a good metric for properly comparing drought indexes. Thus, the indices differ in representing the severity and the time behavior of dry/wet events especially for the drier climates (Karla). As it is clearly evident from the plots, the indices perform better for the more wet climates (Sykia) and there are some discrepancies for the drier. For the wet watershed, all the indices follow the same behavior with a remarkable correlation between the (SPEI) and (RDI) index. A departure from the other indices shows the modified RDI (modRDI). For a wet watershed the difference between the actual and the potential evapotranspiration is not so high so the differences between the (RDI) and the (modRDI) should not be so important. A moderate behavior is seen in Tirtarisio, a semi-dry watershed and finally in Karla , the driest, there are a lot of discrepancies (*Fig.4.8-10*). In Karla, where PET largely exceeds P noticeable differences between the (SPI) and and (SPEI) /(RDI) are expected.

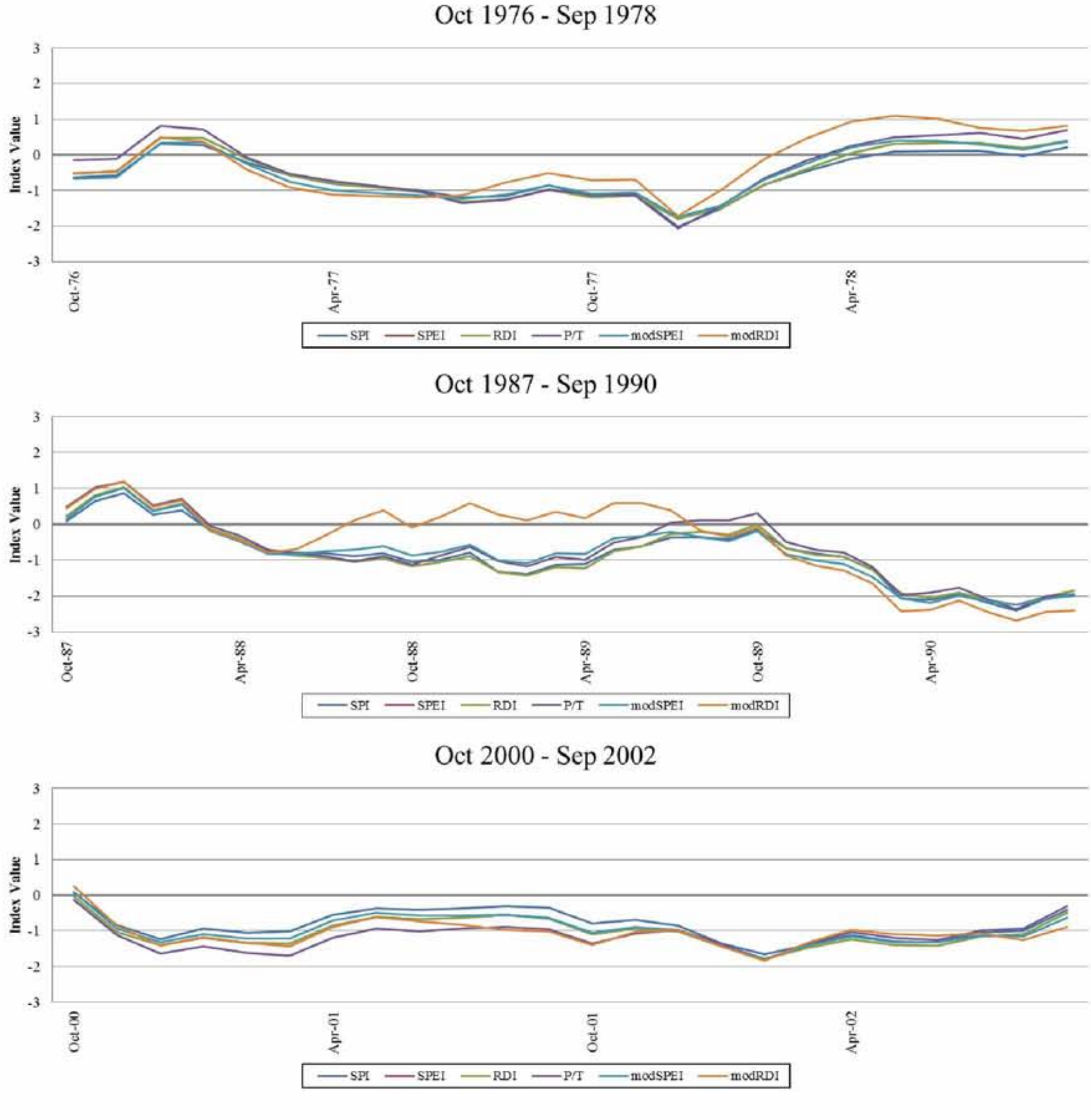

Figure 4.5: *Response of the meteorological indices for indentified drought episodes in Sykia watershed*

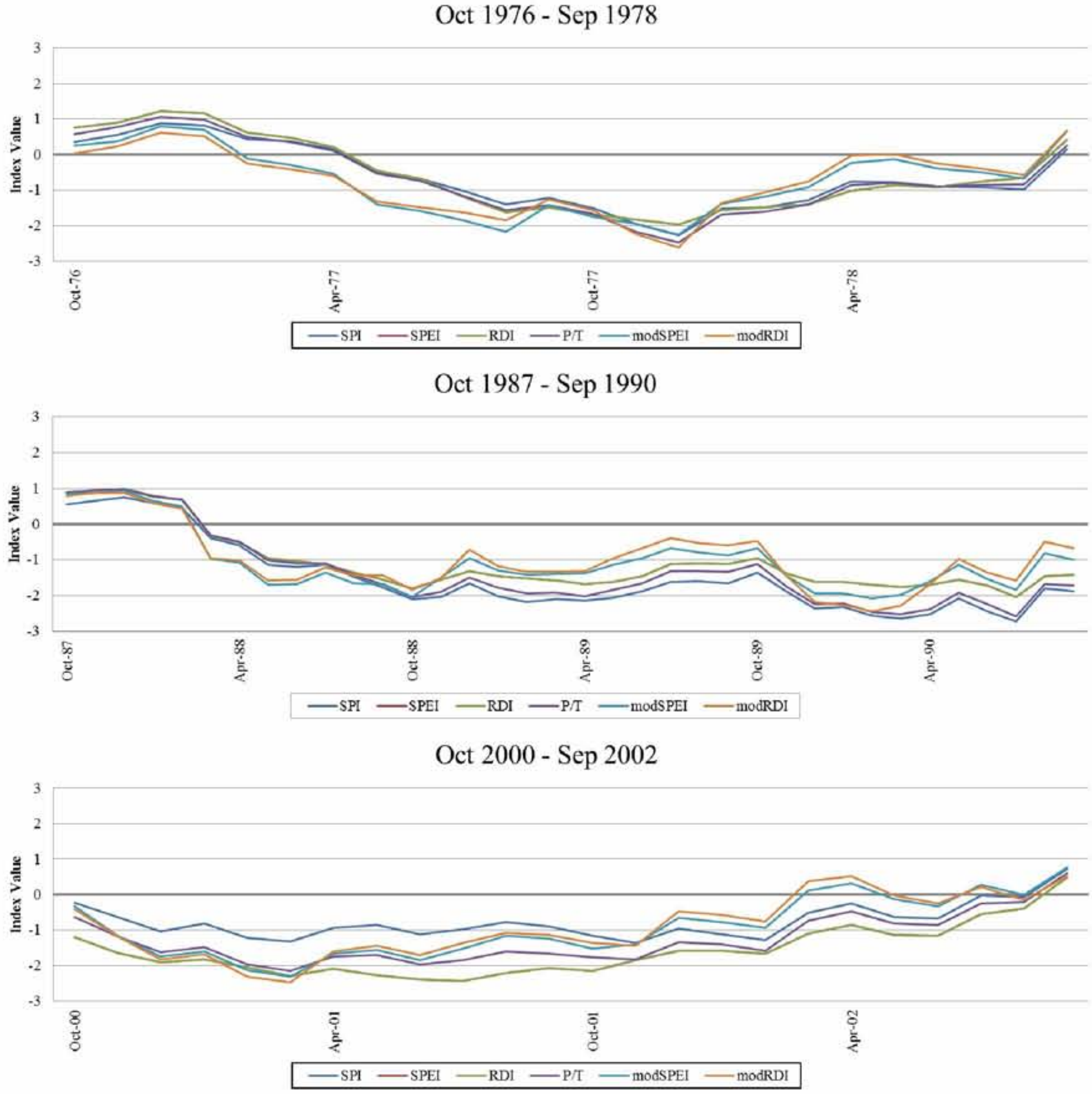

Figure 4.6: *Response of the meteorological indices for indentified drought episodes in Tirtarisios watershed*

Having the three watersheds as origin, in order to determine the reasons there are those discrepancies, it is of interest to show the time behavior of the basic variables. These are the precipitation (P), the temperature (T), the potential evapotranspiration (PET), the actual evapotranspiration (AET), the differences (P-PET), (P-AET) and the ratios (P/T), (P/PET), (P/AET) (*Fig.4.8-10*). All these variables are the main elements in order to compute the meteorological indices and apart from that, they are important hydrological components. There is interaction between them through the hydrological cycle and through the environment, the climate.

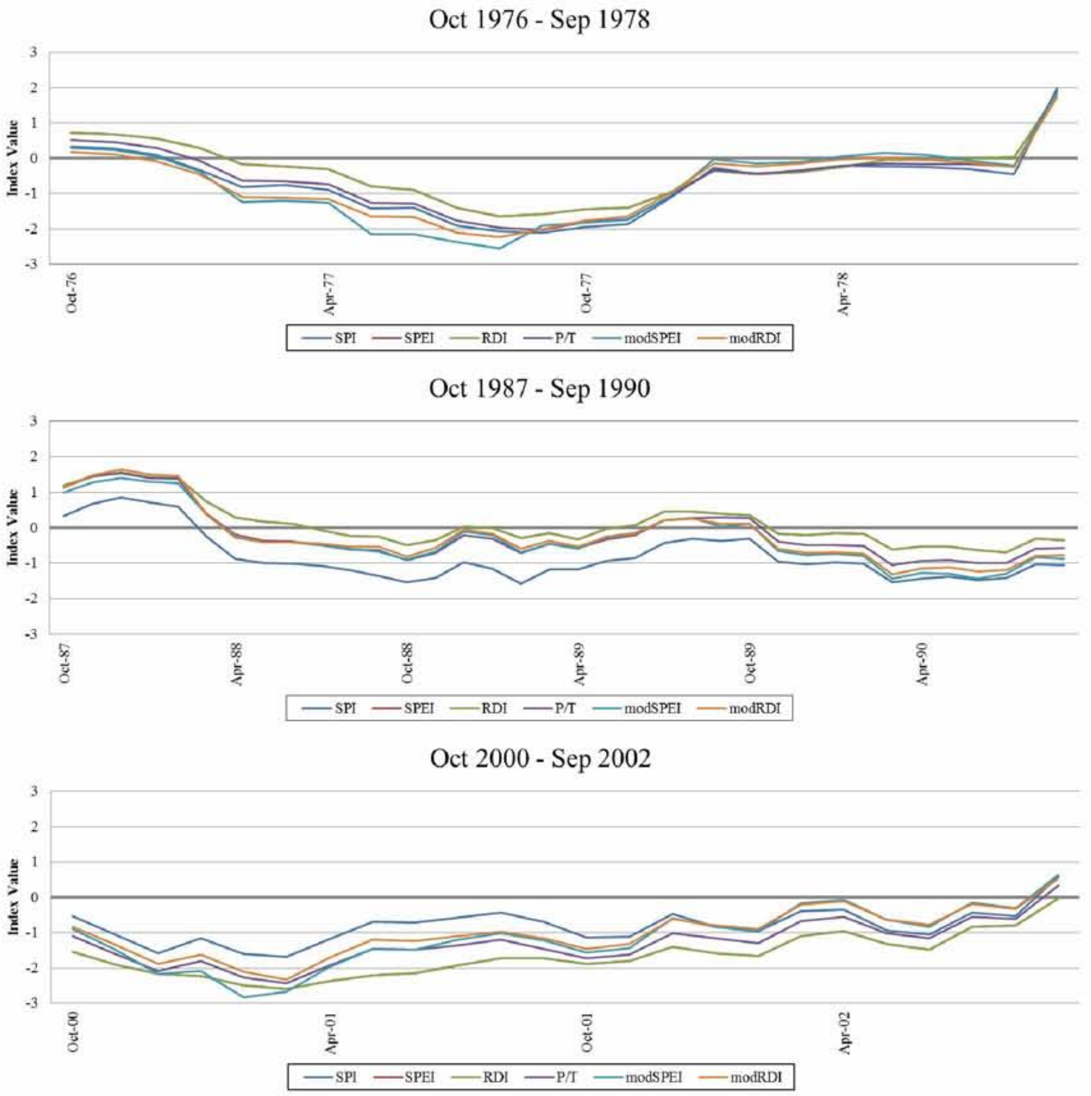

Figure 4.7: *Response of the meteorological indices for indentified drought episodes in Karla watershed*

Nevertheless the differences between the indices for different climates, From the figures it is evident that the ratios (P/PET) and (P/AET) follow the behavior of (P/T) which also logic since PET, calculated with Thornwaite formula, is analogous to the temperature T. Furthermore, the time series of (AET) and (PET) have almost the same behavior, the drier the climate is, the higher the difference between (AET) and (PET) is (*Karla, Fig. 4.10*). As it has been mentioned before, (AET) and (PET) have almost the same behavior, since (AET) as an actual, hydrological component shows a lot of fluctuations.

DEFICITS						
	SPI_12	SPEI_12	RDI_12	P/T_12	modSPEI_12	modRDI_12
Oct 76- Sep 78	1,77	2,17	2,11	2,30	1,81	1,30
Oct 87- Sep 90	10,48	10,17	9,50	9,22	9,79	13,23
Oct 00- Sep 02	2,44	4,35	5,79	5,13	3,57	3,95
DURATION						
	SPI_12	SPEI_12	RDI_12	P/T_12	modSPEI_12	modRDI_12
Oct 76- Sep 78	4	7	6	7	9	5
Oct 87- Sep 90	15	17	16	14	14	12
Oct 00- Sep 02	10	15	17	15	13	14

Table 4.9: *The deficits and the duration for identified drought episodes in Karla after performing the threshold level method (threshold=-1)*

DEFICITS						
	SPI_12	SPEI_12	RDI_12	P/T_12	modSPEI_12	modRDI_12
Oct 76- Sep 78	4,60	5,18	6,07	6,10	6,91	6,34
Oct 87- Sep 90	28,15	13,68	23,00	23,50	8,07	8,41
Oct 00- Sep 02	1,62	16,95	12,46	10,84	7,86	7,62
DURATION						
	SPI_12	SPEI_12	RDI_12	P/T_12	modSPEI_12	modRDI_12
Oct 76- Sep 78	9	10	9	9	10	10
Oct 87- Sep 90	31	29	31	31	14	12
Oct 00- Sep 02	8	22	16	16	13	13

Table 4.10: *The deficits and the duration for identified drought episodes in Karla after performing the threshold level method (threshold=-1)*

DEFICITS						
	SPI_12	SPEI_12	RDI_12	P/T_12	modSPEI_12	modRDI_12
Oct 76- Sep 78	5,81	2,48	3,81	4,87	8,42	6,41
Oct 87- Sep 90	5,03	____	____	____	1,74	1,03
Oct 00- Sep 02	2,58	21,33	12,93	10,35	9,58	6,45
DURATION						
	SPI_12	SPEI_12	RDI_12	P/T_12	modSPEI_12	modRDI_12
Oct 76- Sep 78	8	5	8	8	11	10
Oct 87- Sep 90	19	0	0	0	5	5
Oct 00- Sep 02	9	28	26	21	13	12

Table 4.11: *The deficits and the duration for identified drought episodes in Karla after performing the threshold level method (threshold=-1)*

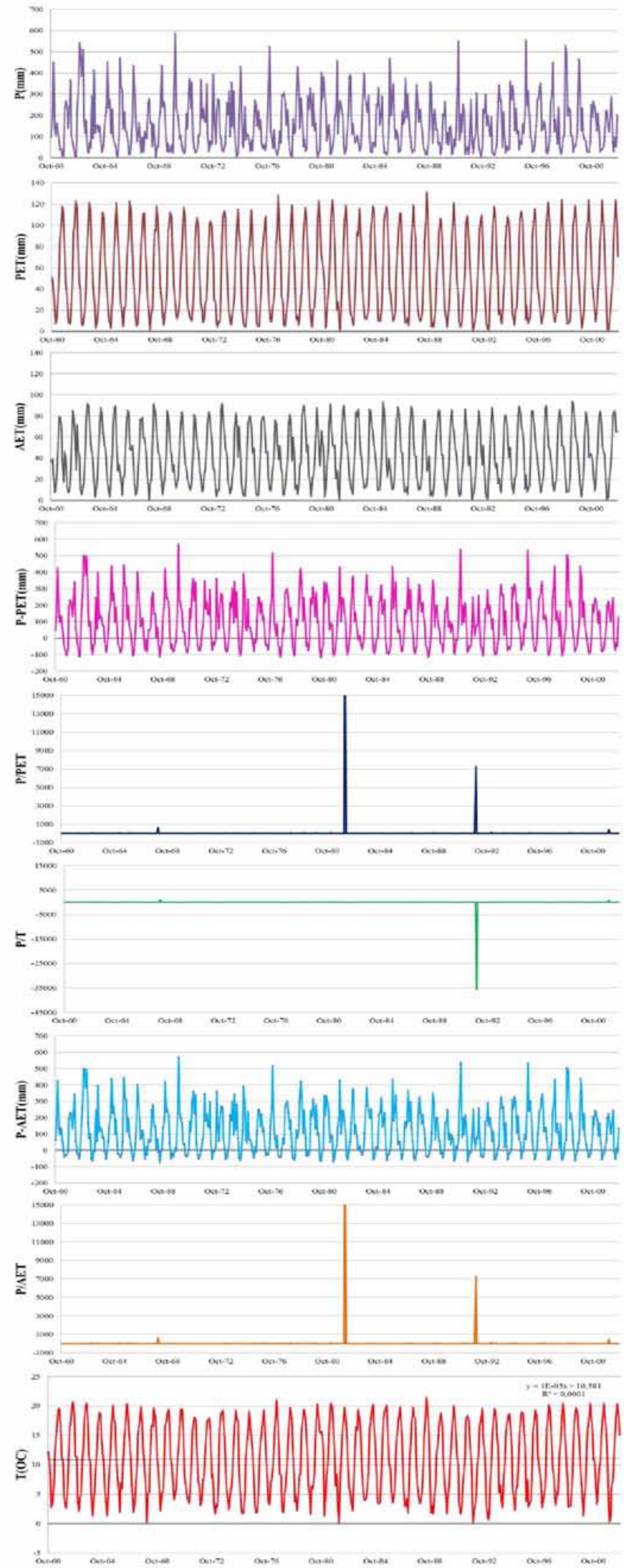

Figure 4.8: *The basic variables of the meteorological indices for Sykia watershed*

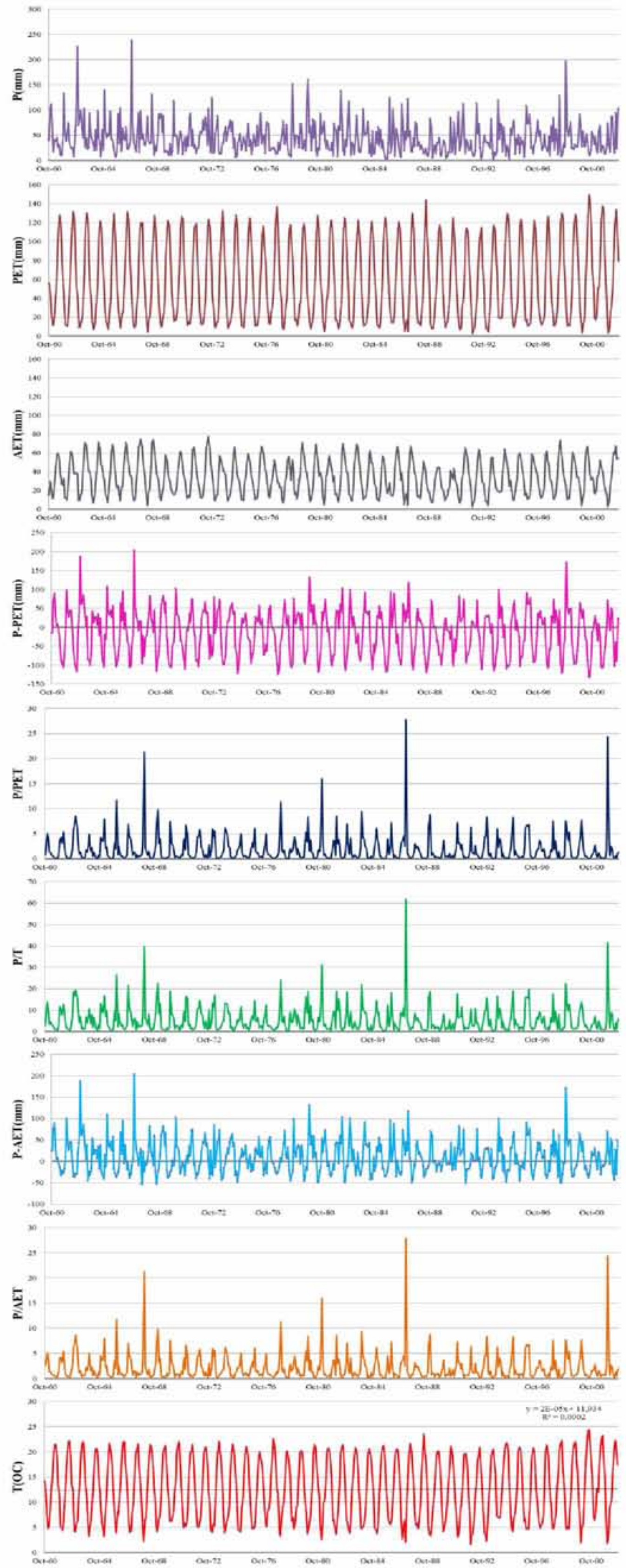

Figure 4.9: *The basic variables of the meteorological indices for Tirtarisio watershed*

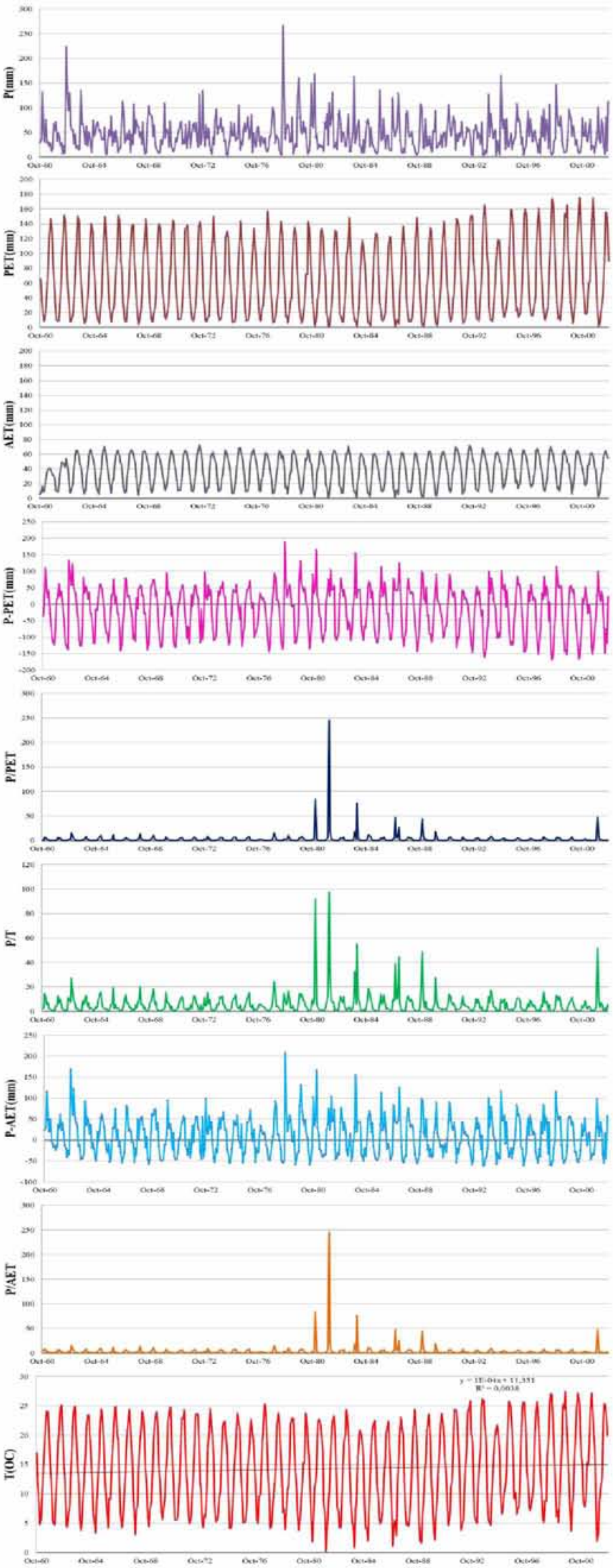

Figure 4.10: *The basic variables of the meteorological indices for Karla watershed*

As can be noted, the nonlinear behavior involves mainly the severe and extreme dry (wet) classes and it may be related to the effect of potential evapotranspiration when large precipitation deficits (surplus) occur. During severe/extreme drought conditions, particularly high evapotranspiration values provide remarkable departures between the two indexes leading to an overestimation of drought events by the RDI. On the contrary, severe/extreme wet events result to be underestimated by the RDI compared with the SPI. This feature appears more evident in arid and semi- arid regions in East Thessaly characterized by low precipitation and high potential evapotranspiration.

Other variables that are illustrated are the differences (P-PET) and (P-AET). These differences follow more or less the same behavior as the precipitation (P). The variables are in accordance and the result they give seems to be logical.

One point is the fact that for cooler climates as Sykia watershed, the lower temperatures are, the higher the ratios become (basically the (P/T) and also (P/PET), and (P/AET)). This results really steep changes in the behavior of the time series and also some disaccording since the evapotranspiration can not be zero since the 12-month moving average has been applied in the whole time series. As it has been shown also, in the frame of climate change it is thought to be really interesting and useful to check possible changes in the temperature which may affect the indices. As it is illustrated an extremely low tendency towards higher temperatures is observed in all representative watersheds. And especially in the warmer ones. This tendency does not affect the relationship between the indices but it has to be a factor for future investigation due to the risk of climate change.

At this stage of the analysis, the standardized index that is more suitable for monitoring the drought remains an open question because of the underlying assumption concerning the most appropriate basic variable to be used for drought assessment that still nowadays is debated. Furthermore, results indicate that indices based on the ratio (P/PET), (P/AET) and mainly the difference (P-PET), (P-AET) are very much influenced by the local balance between precipitation and potential evapotranspiration representing aridity, thus referring more to the impacts of droughts than to the severity of droughts, despite both are interrelated. Results open another question: to know if the behaviour of those indices would change if actual ET was used instead of PET as it is searched with the PDSI and particularly the MedPDSI indices.

Chapter 5 . Conclusions

The study provides an application of several drought indices (SPI, SPEI, RDI, P/T) in Thessaly using monthly data from 10 watersheds distributed in the region for the period 1960-2002. This study analyzes the efficacy of different meteorological drought indices to monitor drought events and also the impact of changing climate on drought characteristics. It is also checked the behavior of the original indices after the use of actual ET instead of PET. In general this study provides an inter-comparison of these indices and a first attempt to understand the observed discrepancies.

Firstly, it was investigated the role of the model for the computation of PET. It was well known that the method of analysis is not crucial especially for annual (12-month time scale) analysis but it is also confirmed after performing several of methods in good quality dataset. The differences were really small thus, as usual the Thornwaite method was chosen. The Thornwaite method gives accurate results after easy computations and really reliable comparing with AET especially in annual basis.

After this, SPI, SPEI, RDI, P/T, modSPEI and modRDI were computed for 12- month time scale and the threshold level method was applied (threshold=-1). Correlation matrices between the indices suggest that there is a good agreement between the time behaviors of the two indexes all over the country even if several differences on the severity of drought events are detectable (i.e. class changes of dry/wet events). Differently, the correlation coefficients between the SPI and SPEI/RDI suggests a good agreement in the wet and less agreement in dry regions. The final result of the threshold level method for each watershed showed different results concerning the magnitude and also the duration of the event. But apart from the duration which is shorter in mountainous watersheds like Sykia and larger in mainly agricultural watersheds, there are not adequate conclusions for the sensitivity of the of the indices.

For this reason, it was investigated the response of the meteorological indices for identified drought episodes. Thessaly experienced severe droughts during the periods from the mid to late 1970s, from the late 1980s to the early 1990s and the first years of 2000s. The dataset was investigated further for three periods; October 1976-September 1978, October 1987- September 1990 and October 2000- September 2002. From this detailed analysis, it was more evident the better performance of the indices in wet climates. Also, it was more evident the unstable performance of (modRDI) and also the strong correlation between (SPEI) and (RDI).

From an analysis of the basic variables of the indices (the precipitation (P), the temperature (T), the potential evapotranspiration (PET), the actual evapotranspiration (AET), the differences (P-PET), (P-AET) and the ratios (P/T), (P/PET), (P/AET)) the elements seemed to be in total accordance. It can be noted that the differences between the indices were controlled (by definition) by the imbalance between precipitation and potential evapotranspiration. In particular, when potential evapotranspiration greatly exceeds precipitation (regions with high temperatures like the arid and semiarid regions, the difference between the SPI and other indices becomes remarkable (differences in severity and time behavior of events).

In order to have the clearest view if the events, the possible increasing trends in the annual mean temperature has been investigated. Results showed that most of the stations, were characterized by a tiny, long-term trend towards higher temperatures, which for sure does not affect the drought characteristics but for sure has to be more investigated in the frame of the climate change.

Finally, at this stage of the analysis two key- factors remain under investigation. The first issue is the determination of the basic variable to properly assess drought conditions; this question involves the definition of drought and the identification of the most important factors affecting the severity of an event. At this purpose a clearer conceptual distinction between drought and aridity is required. For this reason the replacement of PET with AET seems appropriate. The second key-factor is the revision of the already known drought indices for applications in different climates. This problem is related to the presence of long-term trends in the variables of interest that affect the stationarity of the coefficients of the theoretical probability distributions underlying the definitions of the indices. Both issues should be topics of future investigations.

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