Science of the Total Environment 440 (2012) 33–41

Contents lists available at SciVerse ScienceDirect

Science of the Total Environment

journal homepage: www.elsevier.com/locate/scitotenv

Drought and climatic change impact on streamflow in small watersheds

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HIGHLIGHTS

► Assessment of meteorological drought through the reconnaissance drought index.
► Assessment of hydrological drought through the streamflow drought index.
► Forecasting regression equations between RDI of 3, 6 and 9 months and annual SDI.
► Nomographs of streamflow change based on drought and climatic change scenarios.
► Assistance to authorities and stakeholders to mitigate drought consequences.

ARTICLE INFO

Article history:
Received 18 June 2012
Received in revised form 5 August 2012
Accepted 6 August 2012
Available online 8 September 2012

Keywords:
Drought characterisation
Climate change scenarios
Hydrological drought forecasting
Reconnaissance drought index (RDI)
Streamflow drought index (SDI)
Rainfall–runoff model

ABSTRACT

The paper presents a comprehensive, thought simple, methodology, for forecasting the annual hydrological drought, based on meteorological drought indications available early during the hydrological year. The meteorological drought of 3, 6 and 9 months is estimated using the reconnaissance drought index (RDI), whereas the annual hydrological drought is represented by the streamflow drought index (SDI). Regression equations are derived between RDI and SDI, forecasting the level of hydrological drought for the entire year in real time. Further, using a wide range of scenarios representing possible climatic changes and drought events of varying severity, nomographs are devised for estimating the annual streamflow change. The Medbasin rainfall–runoff model is used to link meteorological data to streamflow. The later approach can be useful for developing preparedness plans to combat the consequences of drought and climate change. As a case study, the area of N. Peloponnese (Greece) was selected, incorporating several small river basins.

1. Introduction

Drought is a recurrent regional multi-dimensional phenomenon affecting wide areas and a large number of people. It has been observed that droughts have dramatically increased in number and intensity over the last few decades. The consequences of droughts are also increased and according to the Commission of the European Communities the total cost of droughts in Europe for the last three decades amounts to 100 billion Euros (ComEC, 2007).

Drought as a natural phenomenon represents long periods of lower natural water availability and it is incorporated in the climate variability of an area. Nowadays, however, droughts are increasing in magnitude and frequency, affecting more people than any other natural hazard for long periods of time. This may be partially attributed to the change of climatic conditions, which is an additional threat that puts increased pressure on hydrological systems and water resources (Soo Jun et al., 2011).

Since drought is a very complex phenomenon with no clear beginning and end it has attracted the interest of numerous researchers around the world. Plethora of studies has been convened leading to comprehensive procedures for identifying, characterising and analysing droughts over selected territorial units.

Due to the difficulty of studying the various dimensions of drought phenomena simultaneously, simplifying procedures were recently proposed. Among them is the uni-dimensional approach, which considers a given territorial unit (e.g. river basin or sub-basin) and constant reference periods (e.g. 3, 6, 9 and 12 months). This simplification leads to the assessment of drought phenomenon by its intensity, which can be estimated by drought indices based on meteorological parameters (Tsakiris, 2008).

Several drought indices have been proposed in the literature during the last few decades and have been extensively used in many geographical regions of the world. Among them are the palmer drought severity index (PDSI), the standardised precipitation index (SPI), the rainfall anomaly index (RAI), the standardised anomaly index (SAI), the deciles, the percent of normal, the crop moisture index (CMI),

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http://dx.doi.org/10.1016/j.scitotenv.2012.08.035

the soil moisture drought index (SMDI) and indices based on the normalised difference vegetation index (NDVI). For a comprehensive review of these indices the reader should consult specialised reports (e.g. Tsakiris et al., 2007a; Niemeyer, 2008; Cloppet, 2011).

Using the potential evapotranspiration in addition to the precipitation as the key variables for assessing the severity of drought, a promising new drought index, the reconnaissance drought index (RDI), was proposed (Tsakiris and Vangelis, 2005; Tsakiris et al., 2007b). RDI has been used in several regions and is gaining ground, mainly due to its low data requirements and its high sensitivity and resilience (Asadi Zarch et al., 2011; Khalili et al., 2011; Elagib and Elhag, 2011). The assessment of severity of drought in this paper is based on this index, which is presented briefly below.

The most crucial aspect of drought is the significant decrease of streamflow (Allen et al., 2011) and the lower water storage in reservoirs, conventionally referred to as "hydrological drought". As in the case of drought indices mentioned above, which are more related to the meteorological aspects of the phenomenon, there are also drought indices representing the hydrological drought. A dimensionless powerful hydrological drought index named streamflow drought index (SDI; Nalbantis, 2008; Nalbantis and Tsakiris, 2009) seems to be a rational representative measure of hydrological drought.

The objective of this paper is to devise a simple procedure by which early meteorological drought indications recorded by RDI are used for forecasting the level of hydrological drought of the entire year represented by SDI. Further, the estimation of the reduction of streamflow for a wide range of climatic scenarios is attempted. As a case study, for illustrating the proposed methodology, the area of Northern Peloponnese (Greece) was selected. Several climatic change and drought scenarios were formulated, based on the historical timeseries of the meteorological parameters (precipitation, potential evapotranspiration), that were used to assess the impact on streamflow and on the hydrological droughts in several watersheds of the area.

2. Material and methods

2.1. Drought identification

2.1.1. Reconnaissance drought index

Drought severity can be assessed through the computation of the reconnaissance drought index (RDI) and more precisely through its standardised form ($\mathrm{RDI_{st}}$). The initial value (α_k) of RDI is calculated for the i-th year in a time basis of k (months) as follows:

$$\alpha_k^{(i)} = \frac{\sum_{j=1}^{k} \mathrm{P}_{ij}}{\sum_{j=1}^{k} \mathrm{PET}_{ij}}, \; i = 1(1)N \text{ and } j = 1(1)k \tag{1}$$

in which P_{ij} and PET_{ij} are the precipitation and potential evapotranspiration of the j-th month of the i-th year and N is the total number of years of the available data.

The values of α_k follow satisfactorily both the lognormal and the gamma distributions in a wide range of locations and different reference periods, in which they were tested (Tigkas, 2008; Tsakiris et al., 2008). By assuming that the lognormal distribution is applied, the following equation can be used for the calculation of $\mathrm{RDI_{st}}$:

$$\mathrm{RDI}_{\mathrm{st}}^{(i)} = \frac{y^{(i)} - \bar{y}}{\hat{\sigma}_y} \tag{2}$$

in which $y^{(i)}$ is the $\ln(a_k^{(i)})$, $\bar{y}$ is its arithmetic mean and $\hat{\sigma}_y$ is its standard deviation.

In case the gamma distribution is applied, the $\mathrm{RDI_{st}}$ can be calculated by fitting the gamma probability density function (pdf) to the given frequency distribution of α_k (Tsakiris et al., 2008; Tigkas, 2008). For short reference periods (e.g. monthly or 3-months) which may include zero values for the cumulative precipitation, the $\mathrm{RDI_{st}}$ can be calculated based on a composite cumulative distribution function including:

a) the probability of zero precipitation, and
b) the gamma cumulative probability.

Positive values of $\mathrm{RDI_{st}}$ indicate wet periods, while negative values indicate dry periods compared with the normal conditions of the area. The severity of drought events increases when $\mathrm{RDI_{st}}$ values are getting highly negative. Drought severity can be categorised in mild, moderate, severe and extreme classes, with corresponding boundary values of $\mathrm{RDI_{st}}$ (-0.5 to -1.0), (-1.0 to -1.5), (-1.5 to -2.0) and (<-2.0), respectively. RDI is calculated for a hydrological year in 3, 6, 9 and 12 month reference periods. This implies the different nature of RDI in comparison to other drought indices, since RDI is calculated for predetermined reference periods and not as a "rolling" index of constant duration.

2.1.2. Streamflow drought index

According to Nalbantis (2008), if a timeseries of monthly streamflow volumes $Q_{i,j}$ is available, in which i denotes the hydrological year and j the month within that hydrological year ($j = 1$ for October and $j = 12$ for September), $V_{i,k}$ can be obtained based on the equation:

$$V_{i,k} = \sum_{j=1}^{3k} Q_{i,j} \quad i = 1, 2, \ldots j = 1, 2, \ldots, 12 \quad k = 1, 2, 3, 4 \tag{3}$$

in which $V_{i,k}$ is the cumulative streamflow volume for the i-th hydrological year and the k-th reference period, $k = 1$ for October–December, $k = 2$ for October–March, $k = 3$ for October–June, and $k = 4$ for October–September.

Based on the cumulative streamflow volumes $V_{i,k}$, the streamflow drought index (SDI) is defined for each reference period k of the i-th hydrological year as follows:

$$\mathrm{SDI}_{i,k} = \frac{V_{i,k} - \bar{V}_k}{s_k} \quad i = 1, 2, \ldots, \quad k = 1, 2, 3, 4 \tag{4}$$

in which $\bar{V}_k$ and s_k are respectively the mean and the standard deviation of cumulative streamflow volumes of the reference period k as these are estimated over a long period of time. In this definition the truncation level is set to $\bar{V}_k$, although other values based on rational criteria could be also used.

Generally, for small basins, streamflow may follow a skewed probability distribution which can well be approximated by the family of the gamma distribution functions. The distribution is then transformed into normal. Using the two-parameter log-normal distribution (for which the normalisation is simply reclaiming the natural logarithms of streamflow), the SDI index is defined as:

$$\mathrm{SDI}_{i,k} = \frac{y_{i,k} - \bar{y}_k}{s_{y,k}} \quad i = 1, 2, \ldots, \quad k = 1, 2, 3, 4 \tag{5}$$

in which

$$y_{i,k} = \ln\left(V_{i,k}\right), i = 1, 2, \ldots, \quad k = 1, 2, 3, 4 \tag{6}$$

are the natural logarithms of cumulative streamflow with mean $\bar{y}_k$ and standard deviation $s_{y,k}$ as these statistics are estimated over a long period of time.

According to Nalbantis and Tsakiris (2009), states (classes) of hydrological drought are defined for SDI in an identical way to those used in the meteorological drought indices SPI and RDI. Five states are considered which are denoted by an integer number ranging from 0 (non-drought) to 4 (extreme drought) and are defined through the criteria of Table 1.