Assessing spatiotemporal variability of drought trend in Iran using RDI index

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Abstract

Drought is one of the natural phenomena which occurs in all climates in different parts of the world. Iran is located in the dry belt of the world. The increase of desertification, drought reoccurrence, and destruction by human in this geographical region needs more studies on spatial and temporal trend of rainfall. In this study, trends of climatic drought during 1975-76/ 2004-05 in seasonal and yearly time scales were evaluated at 50 synoptic stations in Iran using a drought index, RDI (Reconnaissance Drought Index), and Mann-Kendall non-parametric test, at 90 and 95% confidence levels. Results showed that among studied stations, in which RDI was calculated for, 89% had an ascending trend, and results of Mann-Kendall non-parametric test on annual values of RDI showed that among 21 stations, 76% of them had a negative trend and 24% had a significant positive trend. Based on results of this study, there exists an increasing drought trend at all-time scales in Iran.

Keywords

1. Introduction

Drought is a recurring phenomenon that affects a wide variety of ranges, making it difficult to develop a single definition of it. It is often generally defined as a temporary meteorological event that stems from the lack of precipitation over an extended period of time compared with some long-term average condition (e.g. precipitation) (Morid et al., 2006). According to a water-resource-oriented definition which considered the water requirements related to biological, economic, and social characteristics of a region, drought refers to a random condition of severe reduction of water supply availability (compared to normal value), extending along a significant period of time over a large region (Rossi, 2000). It has been pointed out that the criteria to define severe reduction, significant period (duration), and large region are affected by subjectivity, as they stem from the demand level as well as perception of negative impacts of the water deficits.

Drought indices are important elements of drought monitoring and assessment since they simplify complex interrelationships between many climate and climate-related parameters which are applied in some studies for climate change (Dastorani et al., 2011) and its effect on water resources (Asadi Zarch et al., 2011; Kousari et al., 2014). Indices make it easier to communicate information about climate anomalies to diverse user audiences and allow scientists to assess quantitatively climate anomalies in terms of their intensity, duration, frequency, and spatial extent (Wilhite and Glantz, 1985). This allows the analysis of the

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historical occurrence of droughts and estimation of recurrence probability. This information is very useful for planning and designing applications of water resources development schemes, related to various uses and environments.

Drought severity is conventionally assessed by drought indices. Several drought indices with varying complexity have been used in many geographical areas. Recently, a powerful drought index, the Reconnaissance Drought Index (RDI), has gained wide acceptance mainly in the arid and semiarid climatic regions (Vangelis et al., 2013).

McKee et al. (1993) developed the Standardized Precipitation Index (SPI) to address the variety of time scales on which precipitation deficits/surpluses can affect different aspects of the hydrologic cycle. A relatively short-lived precipitation deficit during the growing season may have barely noticeable effect on Groundwater levels, but could, at the same time, be detrimental to crop growth. SPI is based on precipitation data summed at any time scale (usually between 1-month and 24-month sums depending on the area of interest) and fitted to a statistical distribution.

Logan et al. (2010) examined the changes in precipitation using the Standardized Precipitation Index (SPI) on a regional scale over a portion of the Central United States. The linear trend of SPI values over the time period was calculated and analyzed, showing many areas of increasing wetness throughout the area, with drying in isolated regions of West and North.

Vangelis et al. (2013) compared the results of RDI for various reference periods using some popular empirical potential evapotranspiration (PET) methods with minimum data requirements. The selected methods were: Hargreaves, Thornthwaite, Blaney–Criddle, and FAO Penman–Monteith (only temperature). Results showed that no significant influence on RDI was detected by using the selected PET methods. This supports the notion that RDI is a robust drought index, not dependent upon the PET calculation method.

Among the indices presented in the world, two indices have more acceptability (Richard and Heim, 2002; Hayes, 2004; Tsakiris and Vangelis, 2005): the Palmer Severity Drought Index (PSDI) (Palmer, 1965; Guttman et al., 1992) and Standardized Precipitation Index (SPI) (McKee et al., 1995; Agnew, 2000). Palmer index uses precipitation, evapotranspiration, and soil moisture conditions as key components. Although it is useful for evaluating drought, it is not sensitive enough for monitoring drought parameters and complex relations which are necessary for its calculation. While SPI index has less complexity and it can easily be applied to any location, its main disadvantage is using one parameter (rain) for the description of water shortage (Tsakiris and Vangelis, 2005).

Thornthwaite (1947) believes that drought cannot be defined in a region only with shortage of rainfall but evaporation also should be considered. Therefore, in this study, new drought index, RDI, is used because this index uses precipitation and potential evapotranspiration for calculation of different severities of drought, so it has more accurate scientific basis than other indices which use only rainfall as input data (Tsakiris and Vangelis, 2005). So, this study tries to show the trends of climatic drought in Iran using RDI and other indices during 1975-76 / 2004-05 in seasonal and yearly time scales.

2. Study area

Iran, with about 1,648,000 km^2, is located in the southwest of Asia and lies approximately between 25N and 40N in latitude and between 44E and 64E in longitude. Iran's important mountains are Alborz and Zagros. Alborz and Zagros Chains extend in northwest-northeast and northwest- southeast respectively. They play an important role in the non-uniform spatial and temporal distribution of precipitation across the country. The country's climate is mainly arid or semi-arid, except for the northern coastal areas and western parts of Iran. The climate is extremely continental with hot and dry summers and cold winters particularly in inland areas. Apart from the coastal regions, the temperature in Iran is extremely continental with relatively large annual range about 22°C to 26°C.

The rainy period in most parts of the country is from November to May followed by a dry period between May and October with rare precipitation. The average annual rainfall of the country is about 240 mm with maximum amounts in the Caspian Sea plains, Alborz, and Zagros slopes with more than 1,800 and 480 mm, respectively (IRIMO, 2010). At the central and eastern inland plains, ranges of precipitation decreases to less than 100 mm annually depending

on the location. From the synoptic aspects, the climate of most parts of Iran is dominated by subtropical high in most times of the year. This phenomenon causes hot and dry climate in summer. Rainfall in country is produced by Mediterranean synoptic systems, which move eastward along with westerly winds in cold season.

In this study, considering various criteria in selection of stations with long-term data, less incomplete data, and suitable distribution, among meteorological data of 50 synoptic stations of Iran, monthly rainfall, monthly mean, minimum, and maximum temperature were used for calculation of potential evapotranspiration using Hargreaves-Samani method (Hargreaves and Samani, 1982) and RDI. A 30-year time series of data (1975-76/ 2004-05) was selected and since drought has direct impact on the hydrological cycle, water year (October-September) was determined as the base of calculations. First, using RDI, various drought severities were calculated at annual and seasonal time steps, then by the use of Mann-Kendall non parametric test, trend analysis of time changes was performed. In addition to original data, 3-year moving average of data was calculated and evaluated by Mann-Kendall test, too. Figure 1 shows the position of selected stations for the study of drought in Iran.
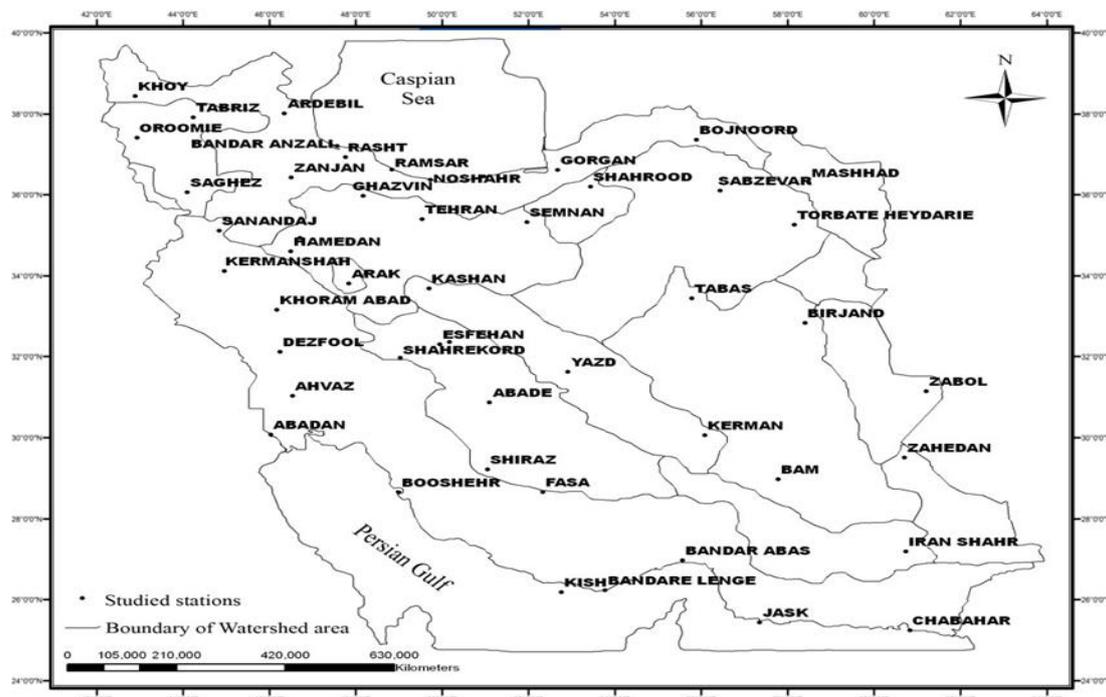

Fig. 1. Position of stations at study area

3. Materials and Methods

3.1. Reconnaissance Drought Index (RDI)

RDI can be characterized as a general meteorological index for drought assessment (Tsakiris and Vangelis, 2005; Tsakiris et al., 2007). The RDI can be expressed with three forms: the initial value α_k, the normalized RDI (RDI_n) and the standardized RDI (RDI_{st}). This paper focuses on the α_k and RDI_{st}.

The initial value (α_k) is presented in an aggregated form, using a monthly time step and may be calculated on monthly, seasonal (3-month, 4-month, etc.), or annual basis. The α_k for the year i and a time basis k (months) is calculated as:

$$a_k^{(i)} = \frac{\sum_{j=1}^{k} P_{ij}}{\sum_{j=1}^{k} PET_{ij}} \qquad i = 1 \ to \ N \tag{1}$$

where P_{ij} and PET_{ij} are the precipitation and potential evapotranspiration of month j of the year i respectively, starting usually from October which is customary for Mediterranean countries and N is the number of years with available data.

The initial formulation of RDI_{st} (Tsakiris and Vangelis, 2005) used with the assumption that α_k values follow the lognormal distribution and RDI_{st} is calculated as:

$$RDI_{st}^{(i)} = \frac{y^{(i)} - \bar{y}}{\bar{\sigma}_y} \tag{2}$$

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

Based on the studies and analysis of various data from several locations and different time scales (3, 6, 9 and 12 months), it was concluded that αk values satisfy both lognormal and gamma distributions in almost all of the locations and time scales, but in most of the cases, gamma distribution was better. Therefore, calculation of RDI_{st} in many cases could be performed in a better way by fitting the gamma probability density function (pdf) to the given frequency distribution of α_k, as below. This approach also solves the problem of calculating RDI_{st} for small time steps, such as monthly, which may include zero-precipitation values ($\alpha_k = 0$), where Eq. (2) cannot be applied. Gamma distribution is defined by its frequency or probability density function (Tsakiris et al., 2008):

$$g(x) = \frac{1}{\beta^\gamma \Gamma(\gamma)} x^{\gamma-1} e^{-x/\beta}, for \rightarrow x > 0 \tag{3}$$

where γ and β show the shape and scale parameters respectively, x represents the precipitation amount and $\Gamma(\gamma)$ is the gamma function. Parameters γ and β of gamma pdf were estimated for every station and every time scale of interest (1, 3, 6, 9, 12 months, etc.). Maximum likelihood estimations of γ and β are:

$$\gamma = \frac{1}{4A}\left(1 + \sqrt{1 + \frac{4A}{3}}\right) \qquad \beta = \frac{\bar{x}}{\gamma} \qquad A = \ln(\bar{x}) - \frac{\Sigma \ln(x)}{n} \tag{4}$$

where n is the number of observations.

Then, obtained parameters were used to find the cumulative probability of α_k for a given year of location. Since, gamma function is undefined for x = 0 and precipitation distribution may contain zero, the cumulative probability becomes:

$$H(x) = q + (1-q)G(x) \tag{5}$$

where q is the probability of zero precipitation and G(x) is cumulative probability of incomplete gamma function. If m is the number of zeros in α_k time series, q can be estimated by m/n. Then, cumulative probability H(x) would be transformed to the standard normal random variable z with mean zero and variance of one (Abramowitz and Stegun, 1965), which is the value of RDI_{st}. Standardized RDI behaves in a similar manner as SPI (McKee et al., 1993) and so is the interpretation of results. Therefore, RDI_{st} values can be compared to the same thresholds as SPI (Table 1).

Table 1. Drought classification thresholds

RDI_{st} or SPI value	Category
2.00 or more	Extremely wet
1.50 to 1.99	Severely wet
1.00 to 1.49	Moderately wet
0 to 0.99	Normal condition - wet
0 to -0.99	Normal condition - dry
-1.00 to -1.49	Moderate drought
-1.50 to -1.99	Severe drought
-2 or less	Extreme drought

3.2. Mann-Kendall non-parametric test

In 1945, Mann originally used this test, and in 1962, Kendall subsequently derived the test statistic distribution (Serrano et al., 1999) and this test is used widely in trend analysis of hydrologic and meteorological series (Lettenmaier et al., 1994). This test allows inquiring the presence of a tendency of long period in rainfall data, without any assumption about its distributional properties. Moreover, the non-parametric methods are less influenced by the presence of outliers in data compared to other methods (Turgay and Ercan, 2005).

In trend test, the null hypothesis H_0 is that there is no trend in the data group which the dataset is drawn; hypothesis H_1 is that there is a trend in the analyzed records. Mann-Kendall test was applied to the monthly dataset. The test statistic, Kendall's S, is calculated as:

$$S = \sum_{i=1}^{n-1} \sum_{j=i+1}^{n} Sgn(X_j - X_i) \tag{6}$$

where X is the data values at times i and j, n represents the length of the data set and

$$Sgn(X_j - X_i) = \begin{cases} 1 \rightarrow if \rightarrow (X_j - X_i) > 0 \\ 0 \rightarrow if \rightarrow (X_j - X_i) = 0 \\ -1 \rightarrow if \rightarrow (X_j - X_i) < 0 \end{cases} \tag{7}$$

The Mann-Kendall test has two parameters that are important for trend detection. These parameters are significance level that indicates the test strength, and slope magnitude estimate that indicates direction as well as magnitude of trend. Under the null hypothesis that y_i is independent and randomly ordered, the statistic S is approximately distributed normally when n≥8, with mean equal to zero and variance as follows:

$$V(s) = \frac{n(n-1)(2n+5) - \sum_{i=1}^{m} t(t-1)(2t+5)}{18} \tag{8}$$

where t is frequency of data with equal values and m is number of series with at least a repetitive data. Standardized test statistic Z, computed by:

$$Z = \begin{cases} \dfrac{S-1}{\sqrt{Var(S)}} \rightarrow S > 0 \\ \\ 0 \longrightarrow S = 0 \\ \\ \dfrac{S+1}{\sqrt{Var(S)}} \rightarrow S < 0 \end{cases} \tag{9}$$

follows a standard normal distribution. In this analysis, confidence level at 90 and 95% were considered.

4. Results and Discussion

First, various drought severities were calculated for various stations of Iran at annual and seasonal time steps. Figures 2 to 5 show the different values for the stations of Rasht and Yazd as two examples of wet and dry climates of Iran. According to these figures, because Yazd station in the last quarter of water year (July-September) in most years, had no rainfall, RDI values in this period are not logical and calculation of trend is not significant.
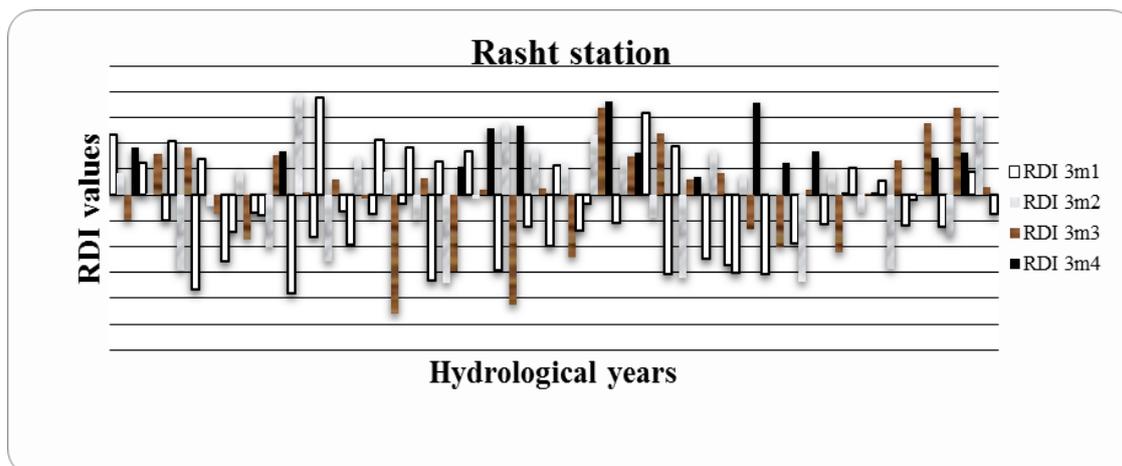
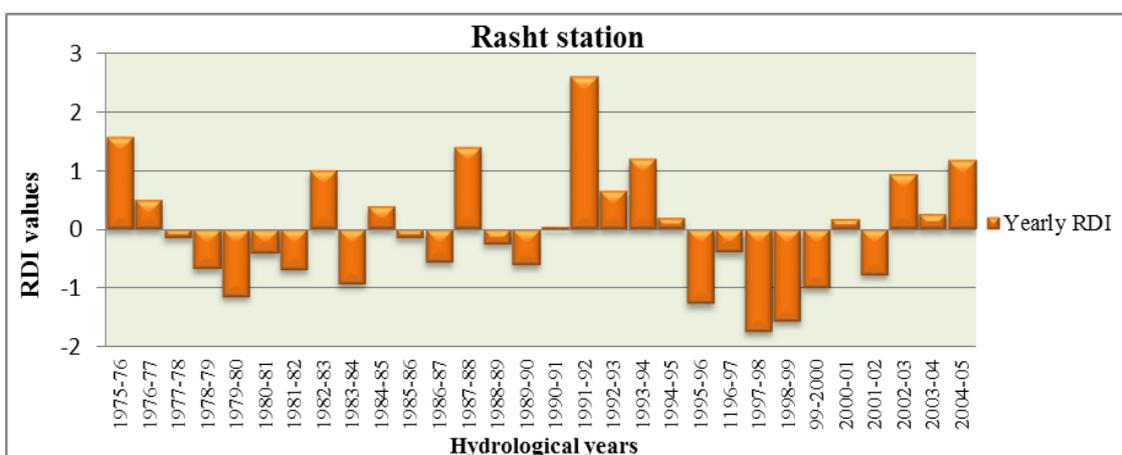**Fig. 2. Seasonal RDI values of Rasht station**

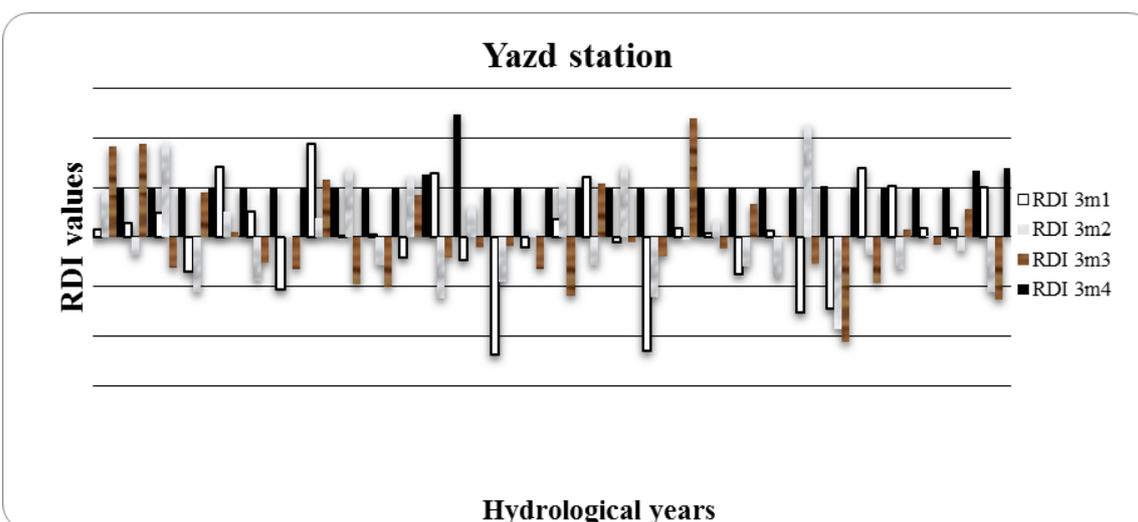**Fig. 3. Annual RDI values of Rasht station**

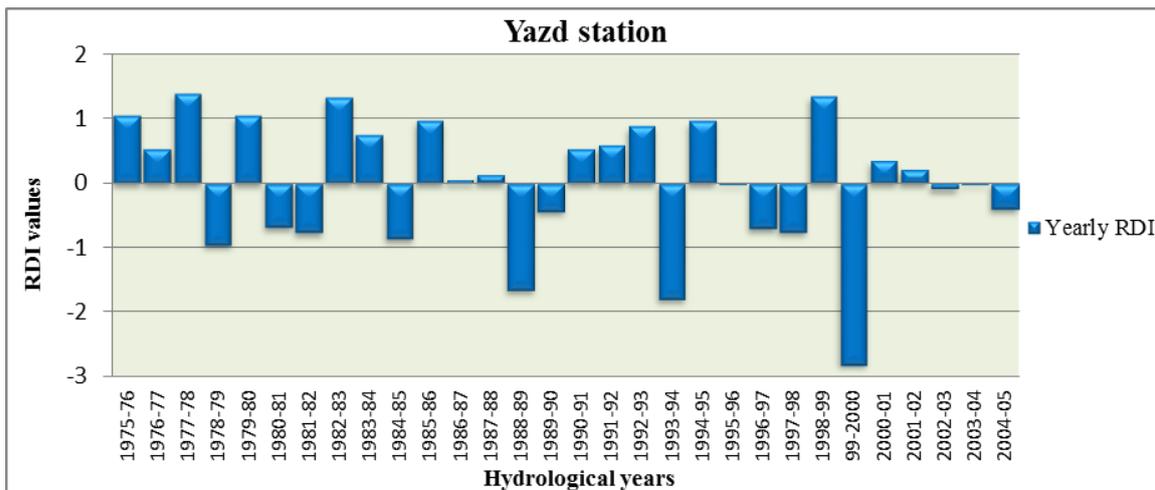**Fig. 4. Seasonal RDI values of Yazd station**

Fig. 5. Annual RDI values of Yazd station

In evaluation of drought trend in autumn (at 90% level of confidence), according to Figure 6, half of the stations which are scattered all over Iran have trend in RDI, which, except for Noshahr station, the rest of them have a negative trend because of increased aridity.

In winter, 44% of the stations had negative trend in RDI the same as the autumn distribution across the country. In all stations, trend and severity of droughts during the 30-year time series of data was increasing (Fig. 7).
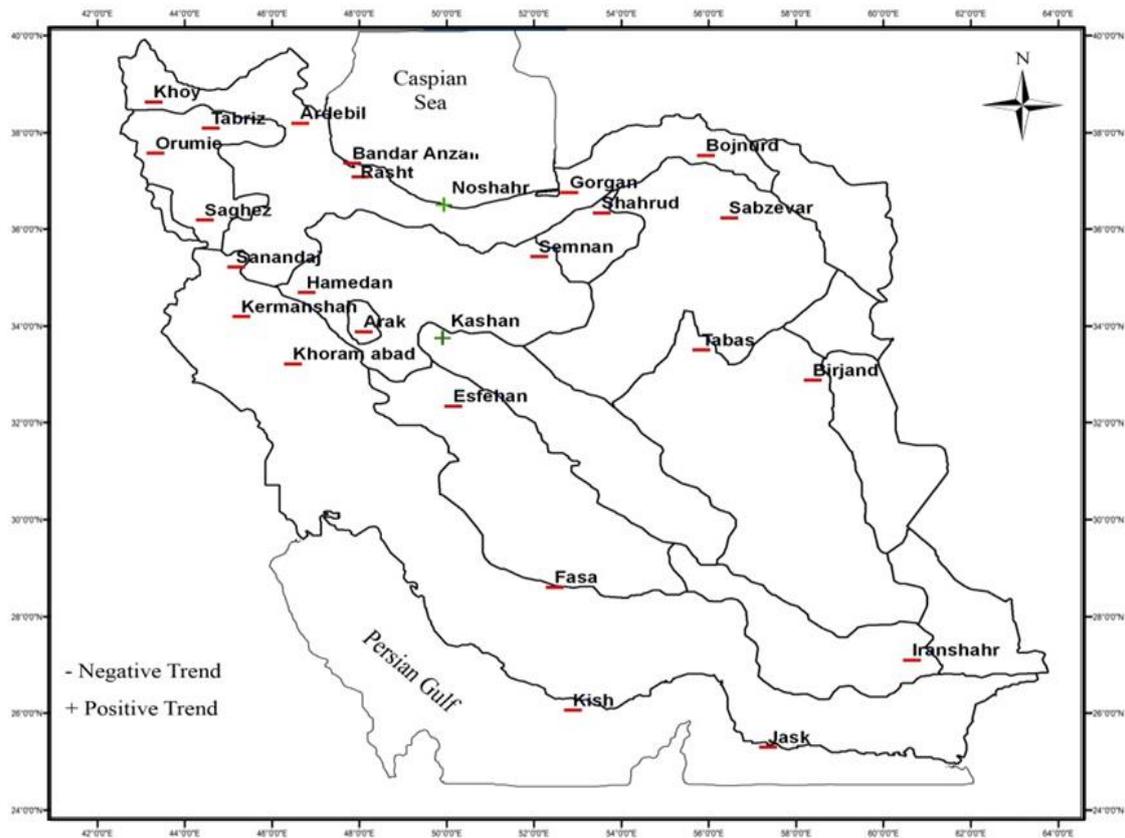

Fig. 6. The spatial distribution of RDI index trend in autumn in Iran
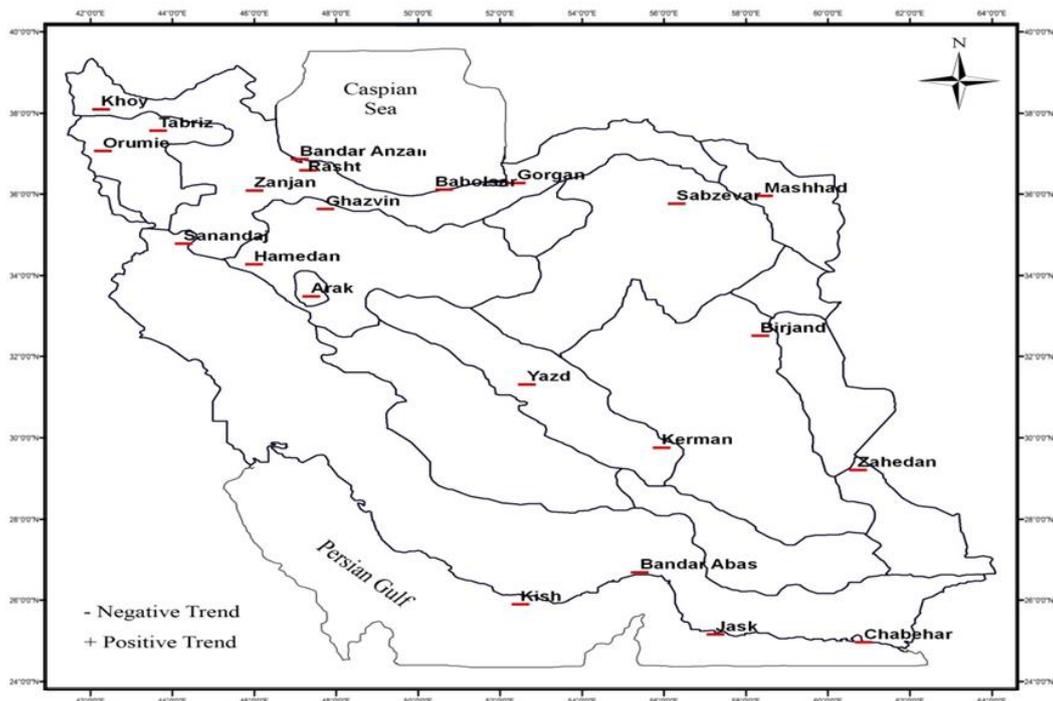
Fig. 7. Spatial distribution of RDI index trend in winter in Iran

According to Figure 8, in spring, the numbers of stations which have positive trends in RDI are more than other seasons. In this season, among 15 stations which had trend in RDI, 40% of them had positive trends that were located in central parts and southwest of the country. The remaining 60% of the stations had negative trends, which, except for Oroomieh, Saghez, and Birjand stations, the rest of them are located in the south.
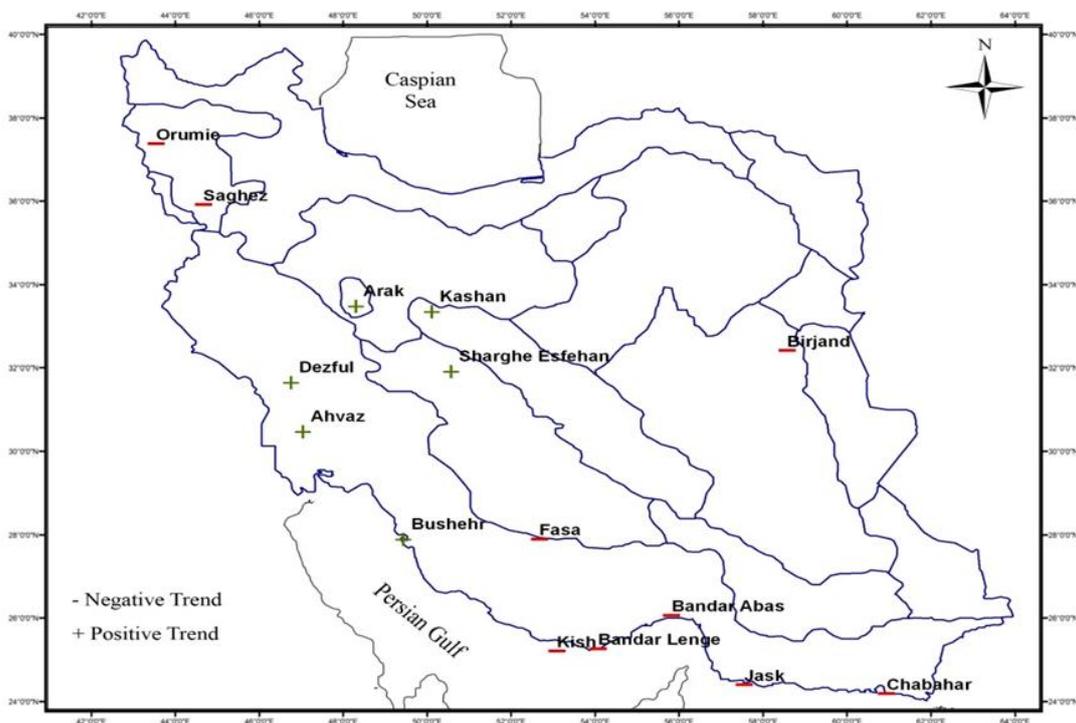

Fig. 8. Spatial distribution of RDI index trend in spring in Iran

Figure 9 shows the spatial distribution of RDI in the summer at 90% level of confidence during 1975-76/ 2004-05. Because in the summer only the northern stations have rainfall, so calculation of trend in RDI for Gorgan, Rasht, Anzali, and Ardebil stations alone are significant and due to the lack of rainfall in rest of the stations, calculation of RDI is difficult, therefore, the analysis of trend is not logical.

Results of Mann-Kendall non-parametric test on annual values of RDI showed that among 21 stations, 76% of them had a negative trend and 24% had a significant positive trend. So, most of the stations had a growing trend concerning increasing aridity (Fig. 10). Most of the stations with negative trends are located in the north, northwest, and southeast of Iran.
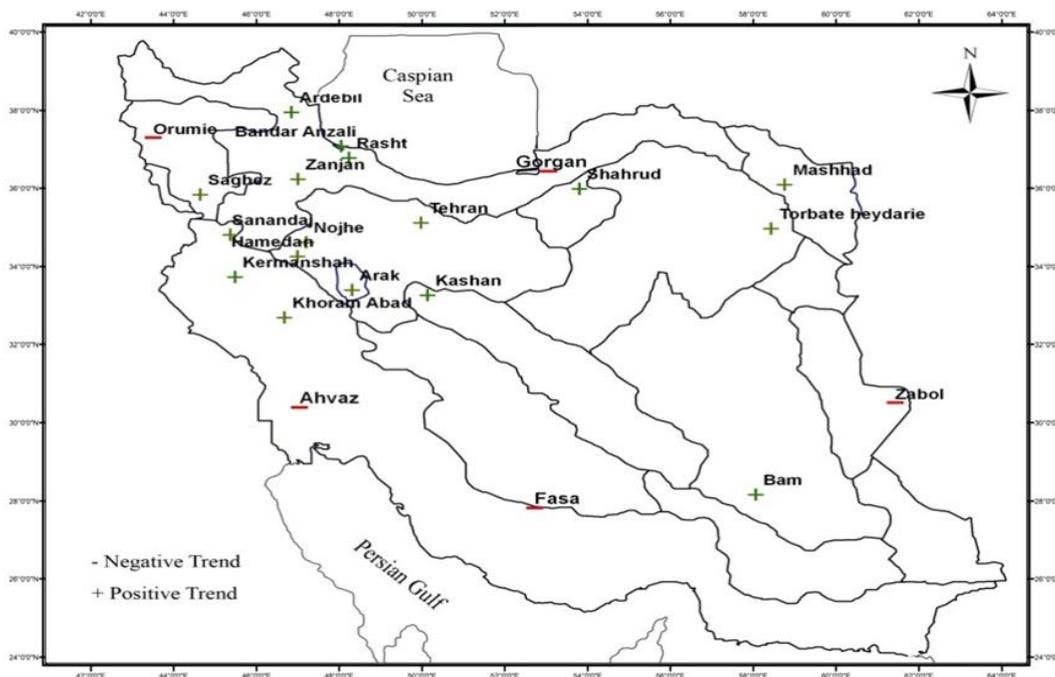

Fig. 9. Spatial distribution of RDI index trend in summer in Iran
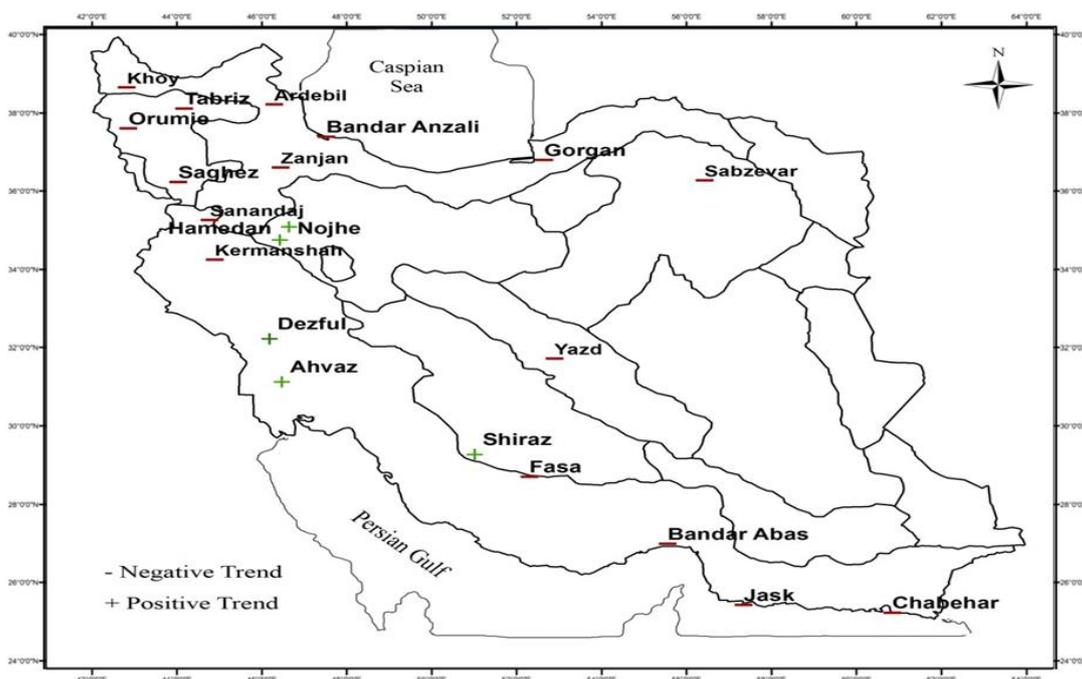

Fig. 10. Spatial distribution of annual RDI index trend in Iran

5. Conclusions

In this study, RDI trends were investigated using RDI method in seasonal and yearly time scale at 50 synoptic stations of Iran. So, RDI time series were trended by non-parametric MK test and then mapped. As the main conclusion, the frequent decreasing trends in RDI time series is a sign of climate change and riskier condition for water resources management. RDI is a simple but valuable drought index which considers both precipitation and evapotranspiration as the main input and output of the hydrological cycles, respectively. These situations are more remarkable in arid and semi-arid regions like central parts of Iran, where the lower values of precipitation are together with high demand of evapotranspiration (Kousari et al., 2014).

Generally, based on findings of this study, although the use of RDI for analyzing drought at monthly and seasonal scale cannot be properly accountable due to the lack of precipitation in most parts of the country, this problem exists not only for RDI but also for the rest of drought indices such as SPI. Moreover, because at most parts of country as a result of warm and arid climate, potential evapotranspiration in most months of the year is much more than rainfall, so the RDI which considers the role of rainfall and potential evapotranspiration will give more accurate estimation of the drought severity which many studies in various parts of Iran are in agreement with this results (for example, Asadi Zarch et al., 2011; Dastorani et al., 2011; Kousari et al., 2014).

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