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Palmer drought severity index as soil moisture indicator: physical interpretation, statistical behaviour and relation to global climate

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Abstract

Palmer Drought Severity Index (PDSI) series, based on monthly homogenised temperature and precipitation data, are analysed for the 1901–1999 period at three stations in West-Hungary, i.e. in an objectively separating region of the country, concerning spatial variations of the monthly PDSI fluctuations. All displayed results represent computations by the Thorthwaite-type potential evapotranspiration. Some comparison with those index-series, computed by the Blaney–Criddle method is given in the Discussion. Series of PDSI exhibit strong correlation with series of two independent soil moisture estimations. Having the regression coefficient standardised by standard deviation of the soil-moisture, we obtain similar coefficients during the year (allowing for only 10–30% difference). This means, PDSI can be considered as a soil-moisture indicator. It is shown that the monthly standard deviation of PDSI exhibit small variation, ca. 10%, with a minimum in the summer period. Distribution of monthly PDSI can be considered as Gaussian, according to the Kolmogorov–Smirnoff test, whereas according to the χ^2-test this is true for more than 2/3 of the cases. Exceptions all fall in the second half of the year. Finally, multi-annual relation of PDSI to the global temperature trends are analysed using the method of "slices" (Mika, 1988), dividing the local and global values into uniform time sequences, the so called time-slices and calculating regression coefficients between the local PDSI and two hemispherical temperature variables. One of the latter is the hemispherical mean, the other is the continent–ocean air temperature contrast. This correlation is always negative and frequently significant, which means that in the 20th century local soil moisture conditions became drier parallel to the hemispherical changes.

Keywords: Soil moisture; Palmer-index; Climate change; Method of slices

1. Introduction

Water budget of a region is determined by both hydrological and thermal processes, which govern the income and outcome components of the balance, respectively. These characteristics are often represented by so called "drought indices", (e.g. Wilhite and Glantz, 1985) to manifest the actual hydrological conditions in a single number. According to their characteristics, drought indices can be sorted as: precipitation indices, water budget indices, soil moisture indices, hydrological and various aridity indices, all of them meant to indicate from different points of view moisture conditions, deficiency or surplus of water, for a given area. One of the most common indices is the Palmer Drought Severity Index (PDSI), introduced by Palmer (1965) to simulate moisture content of the soil month by month and to compare its monthly anomalies at regions having

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totally different climate and seasons. PDSI indicates the severity of a wet or dry spell: the greater the absolute value the more severe the dry or the wet period. Another widely used index is the Standard Precipitation Index and the reader can find broad literature even about comparison of the two indices in selected regions (e.g. Guttmann, 1998; Domonkos et al., 2001).

Distribution of world agriculture shows adaptation to the present-day climate patterns, but this situation could change due to the likely global warming (IPCC, 2001). So, a hot topic has become the identification of the water-scarcity periods, their severity, duration and statistical behaviour. Sometimes it is stated that the most dangerous consequence of the global warming is not the change in the averages but the overall increase of extreme events. Among the extreme meteorological events, droughts are possibly the most slowly developing ones, that often have the longest duration, and at the moment the least predictability among all atmospheric hazards. In large areas of relative depressions, where local flow is blocked in wet years, we also experience lengthy inundation and formation of accumulation of salts after the water evaporates. This artificially modified landscape has its special water cycle with altered processes of water retention. These processes (not properly known, in many respects) should be described in the conditions of the continuous rise of human impact on natural environment (especially water resources) and the global climatic change, showing unpleasant trends for both thermal and hydrological characteristics.

The major hydrologic hazards (e.g. flood and drought) are consequences of precipitation extremes, which is likely the most sensitive and most unpredictable component of climate at the temperate latitudes both at inter-annual and also, at inter-decadal time scales. Hence, objective of the present paper is to analyse how the local soil moisture content, expressed in terms of the Palmer Drought Severity Index, had been varied in the 20th Century in a relatively well watered region of Hungary in Central Europe.

The paper is structured as follows: Section 2 specifies basic concepts and main steps of the PDSI computation and also two independent methods of soil moisture estimation, based on local meteorological parameters, as well. Section 3 introduces the selected region and its objectively expressed relation to a larger part of the Carpathian Basin in Central Europe. The applied statistical methods are introduced by Section 4. According to the particular objectives of the paper, they include correlation and regression between PDSI and soil moisture estimations, statistical tests of distribution normality and the so-called method of slices to quantify regression between local indices and hemispherical temperature variables. The results are demonstrated in Section 5, which is followed by a Discussion in Section 6.

2. Data of PDSI and soil moisture

The basic data set used in the present paper contains 99 years long monthly series of air temperature and atmospheric precipitation at three meteorological stations in West Hungary, presented in Fig. 1.

Since both temperature and precipitation data sets used in our examinations exhibit several discovered inhomogenities, (Szentimrey, 1999) the homogenised series are used and compared in the analysis. The MASH (Multiple Analysis for Homogenisation) method, developed in the Hungarian Meteorological Service (Szentimrey, 1999) was performed for this purpose. It is a relative homogeneity test procedure that does not assume the reference series are homogeneous. Possible break points and shifts can be detected and adjusted through mutual comparisons of series within the same climatic area.

2.1. Computation of the PDSI

In the present study, drought is considered as a meteorological anomaly characterised by a prolonged and abnormal moisture deficiency. In order to determine onset and severity of meteorological droughts, Palmer's Drought Severity Index (PDSI) is evaluated (Palmer, 1965; Alley, 1984; Karl, 1986; Dalezios et al., 1991). Monthly PDSI values have been calculated for 10 stations of Hungary, using the homogenised temperature and precipitation data (see above), for the 1901–1999 period.

In general, several methods can be used to calculate the potential evapotranspiration, a key variable of the water balance and also of the PDSI computation procedure. Palmer (1965) applied the Thornthwaite (1948) formula which is rather a climatological characteristic, while later the Blaney–Criddle method provided better estimations (Alley, 1984) especially for vegetation-specific alternatives. In our analysis, the climatic approach by Thornthwaite is applied, and some comparison to the PDSI series obtained by applying the Blaney–Criddle method (maize plant, more specifically) are presented.

Detailed procedure of computing the PDSI is described in the above sources. The procedure considers monthly precipitation, evapotranspiration and soil moisture conditions, and these meteorological variables determine hydrological and agricultural drought. PDSI is standardised for different regions and time periods, which is useful in common assessment for a wide area with different climate. Basic concepts and steps of computation are as follows:

Step 1: *Hydrological Accounting*. Computation of PDSI begins with a climatic water balance using series of monthly precipitation and temperature records. An empirical procedure is used to account for soil moisture storage by dividing the soil into two arbitrary layers.

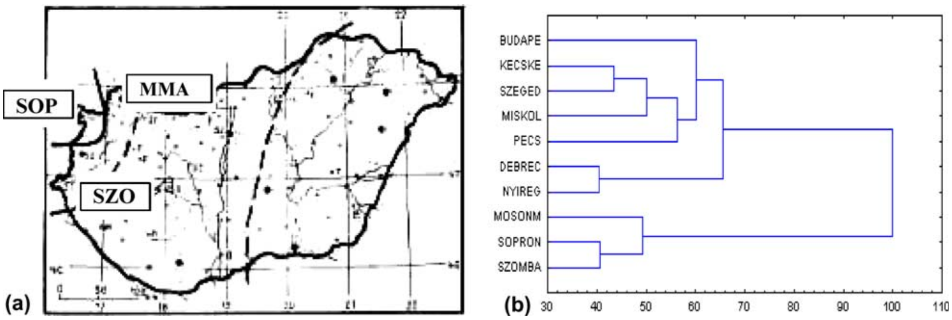

Fig. 1. The three stations of north-west Hungary, selected with respect to objective regional classification from ten stations. (a) results of factor analysis and (b) results of cluster analysis. MMA—Mosonmagyaróvár (47°53′ N, 17°16′ E; 122m above the sea level), SOP—Sopron (47°41′ N, 16°35′ E, 230m a.s.l.), SZO – Szombathely (47°15′ N, 16° 35′ E, 218m. a.s.l.) Examples for June month, 1901–1999.

The upper layer is assumed to contain 25mm of available moisture at field capacity. The loss from the underlying layer depends on the initial moisture content, as well as on the computed *Potential Evapotranspiration* (PE) and *the Available Water Capacity* (AWC) of the soil system. In the present calculations of PDSI, AWC values of 100mm are used for all stations, even with, possibly, different soil types. Runoff is assumed to occur, if and only if, both layers reach their combined moisture capacity, AWC. In addition to PE, three more potential terms are used and defined as follows: *Potential Recharge* is the amount of moisture required to bring the soil to its water holding capacity. *Potential Loss* is the amount of moisture that could be lost from the soil by evapotranspiration during a zero precipitation period. *Potential Runoff* is defined as the difference between precipitation and Potential Recharge.

Step 2: *Climatic Coefficients*. This is accomplished by simulating the water balance for the period of available weather records. Monthly coefficients are computed as proportions between climatic averages of actual vs. potential values of evaporation, recharge, runoff and loss, respectively.

Step 3: *CAFEC Values*. The derived coefficients are used to determine the amount of precipitation (I) required for the Climatically Appropriate For Existing Conditions (CAFEC), i.e. “normal” weather during each individual month.

Step 4: *Moisture Anomaly Index*. Difference between the actual and CAFEC precipitation is an indicator of water deficiency or surplus in that month and station, expressed as $D = P–I$. These departures are converted into indices of moisture anomaly as $Z = K(j)D$, where $K(j)$ is a weighting factor, also accounting for spatial variability of the departures (D).

Step 5: *Drought Severity*. In the final step the Z-index time series are analyzed to develop criteria for the beginning and ending of drought periods and an empirical formula for determining drought severity, such as:

$$X_j = 0.897X_{j-1} + Z_j/3,$$

where Z_j is the moisture anomaly index and X_j is the PDSI for the j-th month. The equation indicates that PDSI of a given month strongly depends on its value in the previous months and on the moisture anomaly of the actual month. It causes strong autocorrelation of PDSI.

In general, monthly PDSI time series range between −9 and +9, specifically, severe and extreme conditions are characterized by absolute values greater than 4 and 6, respectively. These thresholds may vary among the geographic regions of the world, whereas the original attribution considered ±4 to be the extremity threshold. Furthermore, drought events occur in the case of negative PDSI values while positive values imply wet conditions.

Compared to other traditional drought indices, PDSI can demonstrate several advantages: It is able to simulate moisture content of the soil month by month, and it is suitable to compare the severity of drought events at regions having rather different climate and seasons.

Table 1 indicates the original verbal classification of PDSI indices still introduced by Palmer (1965), and also an overall statistics from the three investigated stations of North-West Hungary (see Section 3). The table indicates that only 2.6% of all cases (stations x months) exhibit high (wet) extremes and 3.6% could be sorted as dry extreme.

2.2. Computed soil moisture content

Besides the PDSI, two further data sets estimating the soil moisture content were used. Both sets use combinations of the daily meteorological elements, influencing the water balance of the surface. The data set by Dunkel (1994) was used between 1901 and 1990, while that of Bussay (Lambert et al., 1993) was applied between 1901 and 1988. Monthly correlation coefficients between the monthly values of the two series are between 0.7 and 0.9 in the most part of the year, while they are only between 0.4 and 0.7 from May to September. These parallel series were used to certify the PDSI as a quantitative

Table 1
Interpretation of the PDSI-values and frequency distribution summarised for the three investigated stations in north-west Hungary (1901–1999)

PDSI	Category	Number of cases	Relative freqency
6.00–6.99	Extremely wet	1	0.0%
5.00–5.99		28	0.8%
4.00–4.99		65	1.8%
3.00–3.99	Very wet	159	4.5%
2.00–2.99	Moderately wet	355	10.0%
1.00–1.99	Slightly wet	467	13.1%
0.50–0.99	Incipient wet spell	270[a]	7.6%[a]
0.49 to −0.49	Near normal	461	12.9%
−0.50 to −0.99	Incipient drought	303[a]	8.5%[a]
−1.00 to −1.99	Mild drought	659	18.5%
−2.00 to −2.99	Moderate drought	415	11.6%
−3.00 to −3.99	Severe drought	251	7.0%
−4.00 to **−4.99**	Extreme drought	106	3.0%
−5.00 to **−5.99**		11	0.3%
−6.00 to **−6.99**		12	0.3%
−7.00 to **−7.99**		1	0.0%
Total		3564	100.0%

[a] Halved interval width, according to Palmer (1965).

indicator of the soil moisture content, and to establish the conversion of the PDSI into the soil moisture, as a physical parameter.

The basis of both calculations is the water balance equation. The composite loss of water to the air from all surfaces is the evapotranspiration (ET). If there is no lack of water and the rate of evapotranspiration determined only by the atmospheric conditions, we speak about potential evapotranspiration (PE). Under favourable conditions, the cloud droplets fall to the surface as precipitation (P). Over land areas, where P is greater than ET and the excess, called runoff (R) occurs. Under certain circumstances a part of the excess water infiltrates to the deeper soil layers. Infiltration (F) is not easily determined, so for practical purposes it is better to consider a column which extends from the surface to a depth where significant vertical exchanges are already absent. In general, the form of the water balance, including also the net change in soil moisture content (ΔS) is given by:

$$P = \mathrm{ET} + R + F + \Delta S \tag{1}$$

For the calculation of the available soil moisture content, expressed in precipitation mm, the simplified form of the water balance equation was used. The greater part of the Hungary is flat and mostly during the vegetation period the percent of the runoff is very small. So the runoff was taken to be equal to zero and its value for the whole time series was neglected. We calculated only the upper one meter layer soil moisture content. According to the results of the field soil moisture measurements on bare soil and under vegetated surface, it was found that the rate of infiltration could be considerable in the of the winter during the thawing. Taking into consideration that we make a mistake the infiltration was neglected too. After the simplification for the calculation a reduced form of the equation was used. The upper one meter layer soil moisture content (SMC) in the next time unit (SMCi) will be expressed as a function of the previous soil water content:

$$\mathrm{SMC}_i = \mathrm{SMC}_{i-1} + P - \mathrm{ET} \tag{2}$$

In both cases, the time unit it is one month. The total amount of water can be stored in a soil column is field capacity (FC). The certain part of the field capacity is not available by plants. This part of the field capacity called as the wilting point (WP).The value of the field capacity depends on the type of soil. The used soil type for every stations changes between the given values according to the characterising soil type. If the calculated soil moisture content on the next time unit exceeds the field capacity, its value is taken to be equal to field capacity. The maximum value of available water content of the examined soil layer (AWC) can be expressed as AWC = FC − WP.

The two methods of soil moisture computation differ only in estimation of the ET term, but both concerning the potential evapotranspiration and also in the way of derivation of real evapotranspiration as a function of the potential one, as well.

Namely, Dunkel (1994) uses a simple empirical regression form was used (Antal, 1968). The potential evaporation is:

$$\mathrm{PE} = a \cdot [e_s(1-r)]^b \cdot (1 + d \cdot t)^c \tag{3}$$

where a, b, c and d are the empirical constants, e_s saturation water vapour pressure, r relative humidity, t mean air temperature (Antal and Kozmáné Tóth, 1980).

In case of the estimation by Bussay (Lambert et al., 1993), a bi-quadratic empirical relationship,

$$PE = A + Bt + Ct^2 + DP + EP^2 \tag{4}$$

was used, where t is the monthly mean temperature, P is the monthly precipitation sum. In Eq. (4), PE was calibrated on an other empirical formula, based on field experiments (Dunay et al., 1968) and considered to be more exact: $PE = ((100 - f)/200 - f)tn$. In this calibration formula, f is the relative humidity in percent, t is the monthly mean temperature and n is the number of days in the month.

For the other difference, i.e. determination of K coefficient in the ET = K PE relationship, the Dunkel (1994) formula applies the Budyko–Posza method was used, where $K = kw$. Here k is the plant constant, and w is an empirical variable depends on the soil moisture content. If (SMC−WP)/(FC−WP) is higher than 50%, than evapotranspiration is taken to be potential, ($w = 1$), otherwise w is equal to this proportion. The k plant constant is determined from several years' field experiments using Thornthwaite type compensation lysimeter. Using lysimeter measurements have done since late 60s at Agrometeorological Observatories in Hungary for 19 species a pentad averages of plant constants were calculated (Posza and Stollár, 1983).

In case of Bussay (Lambert et al., 1993), $K = (1 + \exp(g - hW_r)^{-1}$, where W_r is the relative soil moisture, i.e. proportion of the actual to the maximum available soil moisture contents, SMC/AWC; g and h are soil-type dependent empirical constants. For loam and sand type soils, values of g are 4.87 and 5.76, whereas for h the corresponding values are 9.37 and 11.34, respectively. In this case, as in the Dunkel‘s, one, as well, soil moisture of the previous step is considered for estimation of evaporation of the given month.

3. The selected region

Hungary is situated in the Carpathian Basin, in Central Europe. Climate of Hungary is influenced mainly by two factors: Its position in the mid-temperate climatic zone and in the middle of the Carpathian Basin. The former indicates a typical four-season temperate climate, the latter causes extremes occurring more frequently than outside the closed basin. Hungary's climate is often described as one of intersection of three dominant climatic types: the Atlantic, with moderate temperature change during the year and abundant precipitation; the east-European, with cold winters and hot, dry summers and the sub-Mediterranean, with wet, moderate winters and hot, dry summers.

In this study we selected its plain north-western part, with relatively mild climate, with three available stations Mosonmagyarovar (MMA), Sopron (SOP) and Szombathely (SZO, see Fig. 1 for the location within the country and also for the geographical co-ordinates). This part of the country is surrounded by medium hills, but it is separated from the rest of the country from meteorological point of view, too. Linear distances between the stations are 87 km (MMA and SZO), 57 km (MMA and SOP) and 48 km (SOP and SZO).

Having performed factor analysis and also cluster analysis (i.e. (von Storch and Zwiers, 1999) in space mode (i.e. classification of the 10 stations with long-term PDSI values— see also in Fig. 1), both analyses yield separation of the three stations from the rest of the country. This means inter-annual variations of these stations are fairly similar to each other, but different from the more distant stations.

4. Methods

4.1. Regression to soil moisture

The Palmer indices are compared to the above defined soil moisture series of the upper one meter layer (Section 3.2), to validate usefulness of the indices from the viewpoint of agricultural hydrology. For this reason bi-variate correlation and regression coefficients between the PDSI and the synchronous soil moisture index are computed. The latter coefficients are also normalised by standard deviation of the soil moisture. This transformation leads to nearly similar coefficients in each month of the year.

4.2. Basic statistics

Basic statistical characteristics of the distributions for PDSI are demonstrated in the Results section. Since mean values of the PDSI differ from zero only for the reason, that the reference period of computations is not the full 99 years but the first 80 (1901–1980) only, standard deviations are presented in more details. The main question here is whether the standard deviations are identical in the different months and different stations, as it could follow from an ideal universal (climate-independent) soil moisture index.

Also for the higher order moments we expected normality of the distribution concerning month-by-month sub-sets of the series. Normality of the distribution is tested by the Kolmogorov–Smirnov test and χ^2-test.

Synchronous spatial correlation, which is the basis of the objective regionalisation, presented in Section 2, is also demonstrated for three stations to illustrate how different the water balance (drought indices) can be, even within this relatively small spatial domain. Time-correlation of 1 to 6 month lag is also computed to assess, whether the corresponding power function based on one-month lag do really fit the autocorrelation function depending on longer lags. This auto-correlation is largely influenced by the strongly recursive (auto-correlative) nature of PDSI.

4.3. Method of "slices"

Analysing the trends of drought indices an important question is the relation of their change to the global climatic trends. For this investigation we used the method of "slices". The method of "slices" was developed (Mika, 1988) to investigate connections between regional climatic elements and two hemispherical temperature characteristics, i.e. the average temperature ($\langle T \rangle$) and air-temperature difference between continents and oceans (DT) for the period 1901–1999. The original time series are sliced into sub-periods of the same length, and regression analysis is fulfilled using the time averages of the sub-periods: 5, 9, 13, 17 and 21 years long sets of time-slices, with no attention to trends, defined to randomise the possible data inhomogenities. Regional features of a global warming are estimated by linear regression that connects the regional variable, Y, to anomalies in the annual mean temperature, $\langle T\rangle$, of the Northern Hemisphere and temperature contrast, DT, above the continents vs. the oceans:

$$Y = Y_o + (\delta Y/\delta\langle T\rangle)\langle T\rangle + (\delta Y/\delta \mathrm{DT})\mathrm{DT}$$

The aim of "slicing" is to quantify the connections being non-significant on the year-by-year basis, not distorting the original coefficients. The temperature interval covered by the "slices" is 0.5 K. Regression coefficients are calculated by the method of least squares. Student's t-tests of the regression coefficients are performed. Hemispherical mean temperature and continent–ocean contrasts are derived from air temperatures above the oceans (Folland et al., 1984 and updated) and above the continents according to updated series of Jones (1994). The updates are taken from the Internet with reference to Jones et al. (2000); (*http://cdiac.esd.ornl.gov/trends/temp/jonescru/jones.html*). Air temperature series above the two domains are linearly weighted according to the areas of continents and oceans. Fig. 2 indicates the idea of the method and the transformed hemispherical series.

Correlation of the two hemispherical variables is negligible for all divisions of the investigated 1901–1999 period, which makes possible to avoid the problems of multi-collinearity of the two physically plausible variables. Regression coefficients found and validated by Student's t-tests at 95 and 80% levels we consider real ones. Those of 80% significance are listed only with their sign. If more then half of the cases (3 out of 5) yielded coefficients on at least 80% significance, the average coefficient was calculated as mean of the 5 slices' coefficient.

5. Results

5.1. The PDSI as a quantitative soil-moisture indicator

The linear correlation coefficients between PDSI and the two applied soil moisture series (SMC) are fairly high all along the year, with slightly higher values in October–November and lower values in February (Fig. 3). Considering the high (almost 100) degree of freedom, even the lowest values are significant at the 99% level. This means, strong linear relationship exists between PDSI and soil moisture, which can be used for physical interpretation of PDSI.

Fig. 4 represents the regression coefficients between water content of the upper one meter soil layer, as dependent variable, and the PDSI, for the independent variable of the regression. This coefficient indicates how much water is missing from the soil, comparing to the reference climate, if PDSI is −1, or the same in positive direction (Right axis of the figure). Hence, unit

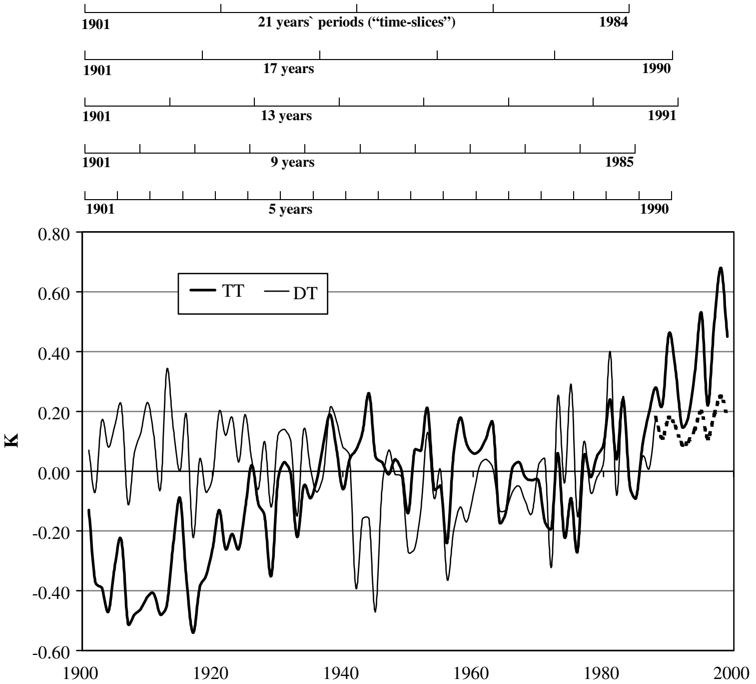

Fig. 2. The time-slices and the hemispherical trends (TT stand for the hemispherical mean temperature, $\langle T\rangle$, and DT stands for the continent–ocean temperature contrast, ΔT). The last 11 years of DT is calculated from correlation to TT.

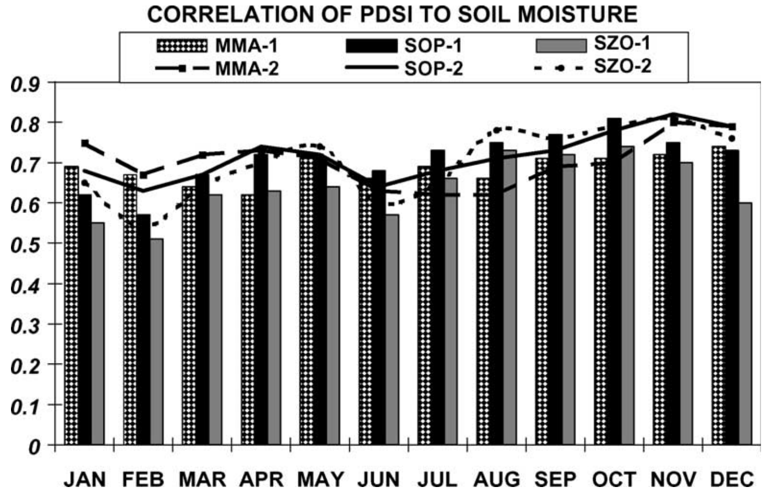

Fig. 3. Correlation of PDSI to soil-moisture series. (1) Dunkel, (2) Bussay series.

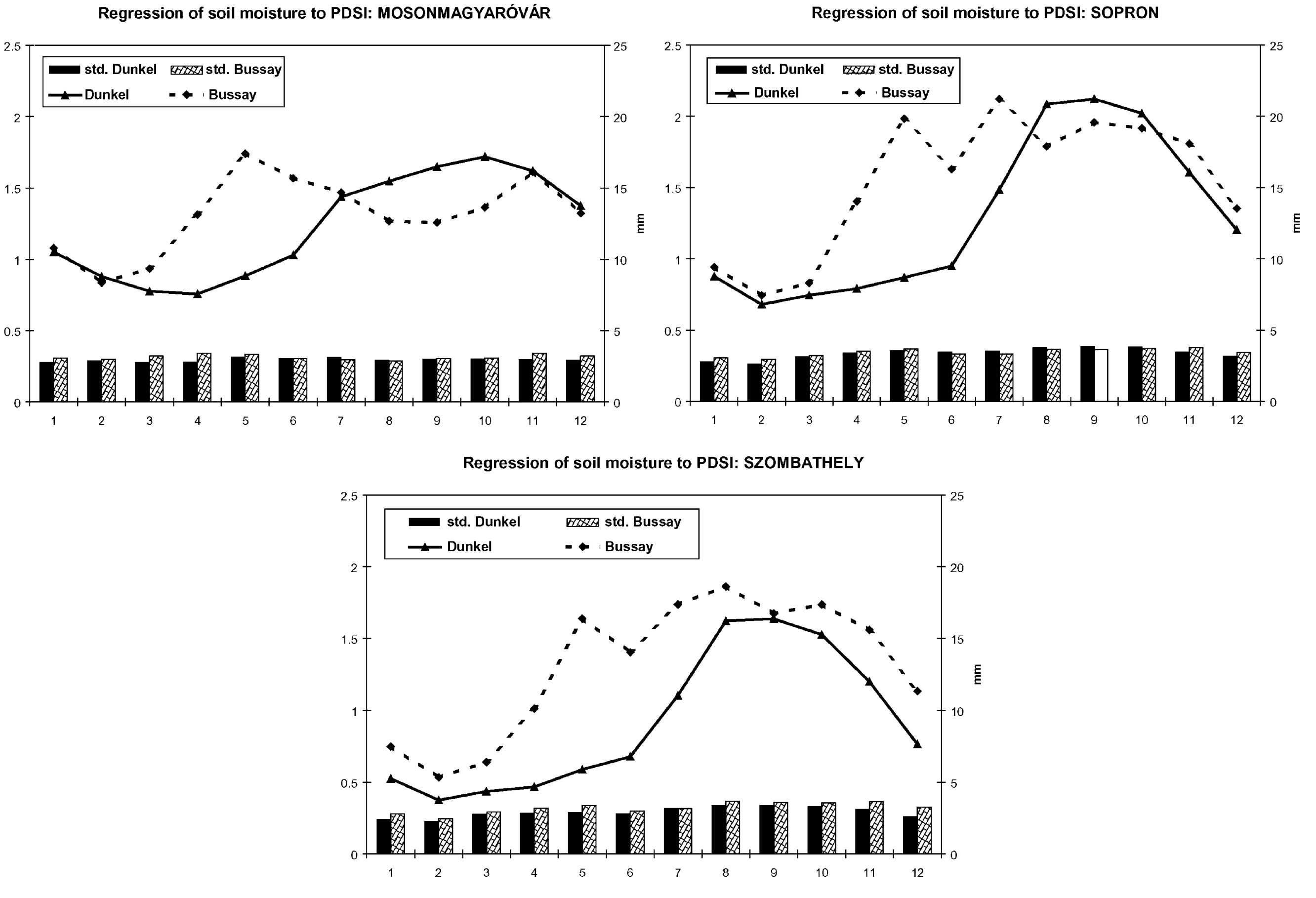

Fig. 4. Regression coefficients between soil moisture and PDSI, and also their standard values, divided by standard deviation of the soil moisture.

increase of the PDSI, corresponds to 4–21 mm surplus according to the Dunkel's, and 5–21 mm surplus concerning the Bussay's soil moisture content, respectively, depending also on season and place.

If comparing the two soil-moisture series, one can establish that sensitivity of the latter series, by Bussay, is generally higher, especially in the late spring period, which coincides with the peak of precipitation (May–June). In the drying-out phase of the year the sensitivities are fairly similar.

A further feature of the regression coefficients, referring to the deep meaning of the PDSI, is the following. If we divide these regression coefficients by the corresponding standard deviation of soil moisture content, then the obtained dimensionless coefficients are in the interval 0,33 ± 0,08 in the data sets of the Dunkel's, and, in the 0,35 ± 0,08 in those of the Bussay's soil moisture content, respectively. Consequently, unit change of the PDSI in each month and station equals to almost the same change in the unit of the soil moisture content standardised by the standard deviation. This feature also helps to accept that PDSI universally represents soil-moisture conditions of the given region.

5.2. Basic statistical characteristics of PDSI

5.2.1. Seasonal variation and normality of distribution

PDSI is claimed to be a non-seasonal characteristic of water availability. However, since the empirical constants of the computation are determined for other regions (continental USA), it is worth checking if statistical parameters of PDSI are really independent from month.

Fig. 5 displays annual cycle of standard deviation at the three stations. One may observe a slight annual cycle with lower values in the warmer half of the year. The differences between highest and lowest standard deviations are 10–15% compared to the lowest values.

No identical standard deviations can be established in spatial respect, either, since Sopron exhibits the lowest, whereas Mosonmagyarovar the highest standard deviation, with Szombathely always showing intermediate standard deviations along the year.

Another question is the shape of statistical distribution of PDSI. If not having the values of different months separated, PDSI may behave quite irregular. In some regions bi-modality of the distribution was reported (Alley, 1984). For Central and South-Eastern Europe, Mika et al. (1994) found no good fit with the common distributions for 110 years‘ long series of PDSI.

Considering these negative results and also the above described slight seasonal variation of the standard deviation, normality of the distribution is investigated for the monthly sub-samples. Non/separated annual sample is used just for comparison.

Table 2 indicates the results of distribution tests performed by Kolmogorov–Smirnov and also by χ^2-tests. The first column reports no normality in the whole-year sample according to the more strict χ^2-test, that indicates strongly significant deviation from normality. In the monthly sub-samples, however, the latter test yields highly significant non/normality in the last three months of the year in one of the stations (Szombathely). Slightly non-significant cases also fall into the drying-out phase

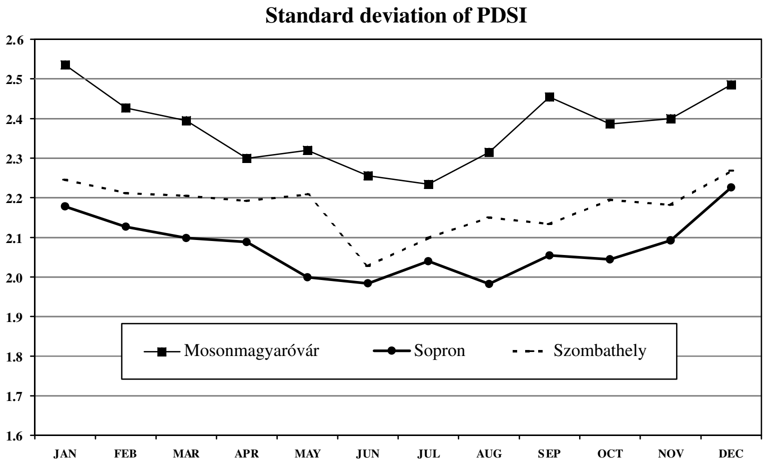

Fig. 5. Annual cycle of the PDSI standard deviation at the three stations.

Table 2
Significance of the deviation (% of randomness) from normal distribution. (n.s. = non-significant. Significance thresholds: 10 and **5**%)

1901–1999	test	year	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
Moson-magyaróvár	*K–S*	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.
	χ^2	**0.0036**	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	8	n.s.	n.s.	6
Sopron	*K–S*	5	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.
	χ^2	**0.0001**	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	10	9	n.s.	6	8
Szombathely	*K–S*	**1**	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.
	χ^2	**0.0000**	n.s.	n.s.	n.s.	n.s.	n.s.	n.s.	9	n.s.	n.s.	**1**	**0.04**	**0.08**

of the year, while no month and station differs from normality in the first six months of the year. The overall proportion of significant deviations and also its regularity in time points out the fact, that normality is a rather rude approximation in the second half of the year concerning PDSI in the given region.

5.2.2. Space and time correlation

Spatial correlation values among the stations are indicated in Table 3, with different months above and below the diagonals. High 0.59 – 0.75 spatial correlation values are not surprising, since the stations are fairly close to each other (see Section 3). On the contrary one could expect even stronger similarity among the parallel indices. The correlation values also exhibit slight annual course with a minimum in June–July, i.e. after the precipitation maximum.

Table 3
Spatial correlation of PDSI in each month of the year and the one-month autocorrelation between the indicated six pairs of months

	January		
	M.m.óvár	Sopron	Szomb.hely
February			
M.móvár	***0.90***	0.70	0.74
Sopron	0.73	***0.90***	0.69
Szomb.hely	0.73	0.72	***0.91***
	March		
	M.m.óvár	Sopron	Szomb.hely
April			
M.móvár	***0.91***	0.73	0.74
Sopron	0.72	***.87***	0.73
Szomb.hely	0.70	0.72	***.91***
	May		
	M.m.óvár	Sopron	Szomb.hely
June			
M.móvár	***0.88***	0.62	0.62
Sopron	0.62	***.86***	0.70
Szomb.hely	0.59	0.66	***.80***
	July		
	M.m.óvár	Sopron	Szomb.hely
August			
M.móvár	**0.89**	0.65	0.59
Sopron	0.75	***.82***	0.65
Szomb.hely	0.63	0.65	***.85***
	September		
	M.m.óvár	Sopron	Szomb.hely
October			
M.móvár	***0.88***	0.75	0.69
Sopron	0.72	***.86***	0.63
Szomb.hely	0.68	0.67	***.92***
	November		
	M.m.óvár	Sopron	Szomb.hely
December			
M.móvár	***0.98***	0.72	0.69
Sopron	0.73	***.95***	0.74
Szomb.hely	0.72	0.74	***.95***

This table also shows time autocorrelation values of one-month lag in every second month. They are all very high, close to 0.9, which is a direct consequence of the strongly recursive (auto-correlative) definition of PDSI.

In case of a stationary one-step Markovian process, autocorrelation values of the longer lags would be equal to the given power function of the initial, constant one-month auto-correlation. The model could be extended for the case of season-dependent autocorrelation in the way, that instead of the power function (chain multiplication of the one-step values), we sequentially construct the production from the corresponding one-step autocorrelation values. Fig. 6 gives an example of such modelling and comparison to the empirical multi-step autocorrelations. PDSI values of August correlated to the previous July (one month lag), June (2 months), … , February (6 months). Empirical autocorrelation values are fairly similar to those, computed according to the above multiplication rule. This means, that besides the strong one-step memory, no independent multi-step signal appears in autocorrelation function of PDSI.

5.3. Correlation to the global temperature trends

Regression of the local PDSI to the hemispherical temperature is quantified by the method of slices to find significant regression coefficients between selected monthly PDSI values, as dependent variables, vs. hemispherical mean temperature, $\langle T\rangle$, and continent–ocean air temperature contrast, DT, as parallel independent variables (see Section 4.3). Table 4 presents the partial regression coefficients of PDSI to the unit (1 K) increase of the Northern Hemisphere mean temperature.

The basic period of regression estimation is 1901–1999. The regression coefficients are calculated in the cases when more then half of the cases (at least 3 out of 5 slices) yielded significant regression according to the *t*-test at the minimum 80% level. If the level is less than 95%, the table displays the sign of the coefficient, only. In the months and stations, where this criterion is fulfilled, the average coefficient ("**Σ/5**") contains all the five coefficients, irrespectively to their individual significance.

One can establish that all partial regressions of monthly PDSI to the Northern Hemisphere mean temperature, $\langle T\rangle$, are negative, i.e. *the warmer the Hemisphere, the drier the soil is in West Hungary*. Comparing levels of significance or numerical values of the coefficients one can not establish any annual cycle in the reliability or intensity of the connections.

The coefficients are rather high, especially if we consider that the extreme drought threshold of PDSI is only −4. This means that almost all cofficients are stronger than that which would turn the present average situation

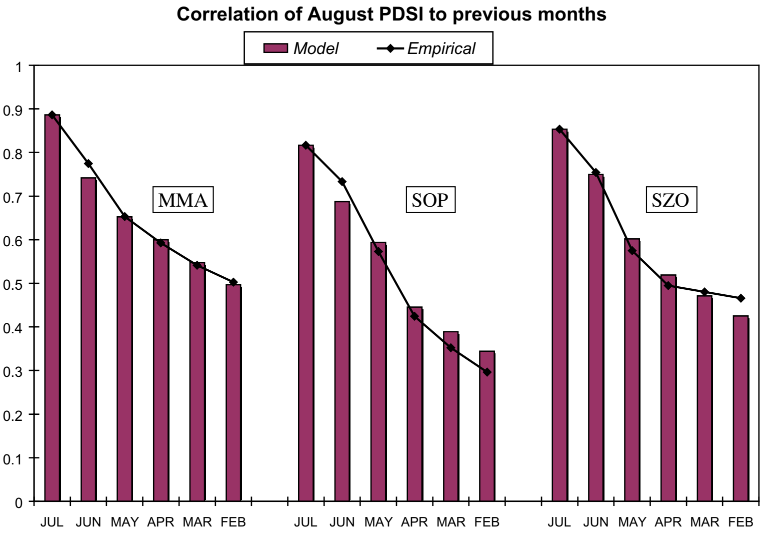

Fig. 6. Autocorrelation of August month PDSI to the previous months (empirical) compared to the product of the corresponding one-step autocorrelation ("Model").

to continous dry extremities (comparing to the present climate, of course). On the other hand, however, one should note, that during the 20th century mankind observed only ca. 0.5 K warming. This means, that even if the relationship between the local and the hemispherical variables remained the same in the future (i.e. for the next half a centigrade), extrapolation of the results for 1 K may be still misleading.

Nevertheless, the regression coefficients indicate strong decrease of soil moisture in the region parallel to the global warming of the 20th century. This consequence is in coincidence with similar studies, performed for Hungary with PDSI data (Szinell et al., 1998; Domonkos et al., 2001; Horváth, 2002; Makra et al., 2002), or simply from precipitation and temperature tendencies (Molnár and Mika, 1997; Kertész and Mika, 1999). This effect of climate change is highly concerned in the neighbouring regions, as well (Szolgay et al., 1997, 2002).

6. Discussion

There are two potential limitations on the PDSI computations presented above. The first one is the identical value of the available water capacity, AWC, in Step 1. Spatial variability of AWC may also add to the regional picture drawn by the factor analysis.

Concerning these patterns, one might think that the obtained regions are obvious consequences of the methodology, if there is considerable spatial correlation among the stations. But this is not the case, as it was experienced by analyzing monthly temperature series at 16 stations in the period 1951–1990 in Hungary (Mika, 2000). For this element we got one single region in each month of the year from an area of the whole country (93 000 km^2). This means, that in that example the strong long-term variations occurred in fairly similar way in all stations. Contrary to this behaviour, fluctuations of PDSI exhibited strong spatial variations, as well. Hence objective separation of the investigated region from the rest of the country was not an artefact of the methodology.

Concerning PDSI as a universal index, there have been several reflections addressed in the literature. Most of the criticisms are collected in one or both of two publications by Alley (1984) and by Heddinghaus and Sabol (1991) on the PDSI. Some of the specific weaknesses addressed by the two publications, among others, are:

- the arbitrary designation of drought severity classes;
- the sensitivity of PDSI to exact determination of AWC;
- the bimodal distribution of the PDSI;
- the original potential evapotranspiration is estimated using the Thornthwaite method, not considering specific vegetation.

Our results may add to this discussion in the following way: The selected four threshold for extremity is quite realistic in the given region, since PDSI does not often exceed this threshold in any direction. Exact values of the AWC might really be important, but even in case of its identical values the neighbouring stations may exhibit partly independent fluctuations. Bimodality of the distribution might be, at least partly, the consequence of some remained seasonality in the series. Use of monthly sub-samples could largely reduce this feature of the distribution.

Finally let us terminate the Discussion, and the paper itself, with a comparison of the PDSI series based on the

Table 4
Relative sensitivity (partial regression, K^{-1}) of PDSI in relation of Northern Hemispheric mean temperature, $\langle T\rangle$ and the continent–ocean contrast, ΔT in selected stations of East-Hungary (1901–1988). The coefficients are shown by their value (>95%) or sign (>80%) according to their significance levels

Station		Mosonmagyaróvár						Sopron						Szombathely					
Month	Regression coefficient K^{-1}	5 years	9 years	13 years	17 years	21 years	Σ/5	5 years	9 years	13 years	17 years	21 years	Σ/5	5 years	9 years	13 years	17 years	21 years	Σ/5
January	∂**PDSI**/∂⟨**T**⟩	–	–	–	**8.7**	–	**−5.1**	**−4.1**	**−6.2**	**−5.5**	**−11.4**	–	**−7.1**	**−4.6**	**−5.7**	**−5.2**	–	**−8.0**	**−7.2**
	∂PDSI/∂Δ*T*				*−14.1*				*−8.9*	–	–		*−9.5*				–	*−6.1*	
February	∂**PDSI**/∂⟨**T**⟩		**−4.0**	–	**−8.5**		**−4.7**	**−3.9**	**−5.4**	**−4.4**	**−9.2**		**−6.1**	**−4.9**	**−6.2**	**−5.3**	–	**−7.7**	**−7.2**
	∂PDSI/∂Δ*T*			–	–				*−6.7*		–						–	–	
March	∂**PDSI**/∂⟨**T**⟩		–	–	–		**−5.2**	**−3.9**	**−6.0**	**−4.4**	**−10.6**		**−6.4**	**−4.0**	**−5.9**	**−4.8**	–	–	**−6.8**
	∂PDSI/∂Δ*T*			–	–			–	*−10.0*		*−17.4*		*−8.8*				–		
April	∂**PDSI**/∂⟨**T**⟩	–	**−5.1**	**−4.7**	–		**−5.2**	**−4.4**	**−6.2**	–	–		**−6.1**	**−4.8**	**−6.8**	**−4.9**	–	–	**−6.6**
	∂PDSI/∂Δ*T*			–	–				*−8.0*		–				–				
May	∂**PDSI**/∂⟨**T**⟩		–	–	–		**−4.8**	–	**−5.5**	–	**−10.6**		**−5.6**	–	**−5.6**	**−4.1**	–		**−5.3**
	∂PDSI/∂Δ*T*			–	–				*−8.1*		*−18.7*								
June	∂**PDSI**/∂⟨**T**⟩	–	**−5.1**	–	**−7.7**		**−5.3**	–	**−4.7**		**−8.4**		**−4.6**	**−2.7**	**−4.1**	–	–	–	**−3.7**
	∂PDSI/∂Δ*T*				*−11.2*				*−8.1*		–				–				
July	∂**PDSI**/∂⟨**T**⟩	**−4.0**	**−5.9**	–	**−9.4**	–	**−6.3**	**−4.3**	**−6.3**	–	**−10.8**	–	**−6.5**	**−3.6**	**−5.0**	–	–	–	**−5.0**
	∂PDSI/∂Δ*T*		–	–	–	–	*−7.5*		*−9.4*		–				–		–		
August	∂**PDSI**/∂⟨**T**⟩	–	**−5.4**	**−4.8**	**−8.1**		**−5.3**	–	**−5.5**	–	**−10.0**		**−5.6**	**−3.3**	**−5.2**	**−3.5**	**−10.2**		**−5.2**
	∂PDSI/∂Δ*T*		*−8.7*	–	*−15.4*		*−6.8*		*−9.7*		–						*−15.8*		
September	∂**PDSI**/∂⟨**T**⟩	–	**−6.3**	**−5.7**	**−9.4**		**−6.1**	**−4.2**	**−5.8**	**−4.6**	**−10.5**		**−6.3**	**−3.8**	**−5.7**	–	**−10.5**		**−5.7**
	∂PDSI/∂Δ*T*		–		*−14.1*				–		*−15.4*						*−13.6*		
October	∂**PDSI**/∂⟨**T**⟩		**−5.8**	–	**−9.2**		**−5.3**	**−3.6**	**−5.7**	–	**−10.3**		**−5.8**	**−3.6**	**−5.6**	–	**−10.7**		**−5.6**
	∂PDSI/∂Δ*T*		–		*−17.8*				*−7.2*		*−17.4*						*−15.1*		
November	∂**PDSI**/∂⟨**T**⟩		**−5.2**	–	–		**−4.8**	**−4.2**	**−5.8**	–	**−10.2**		**−6.3**	–	**−5.1**	–	**−10.4**	–	**−5.5**
	∂PDSI/∂Δ*T*		–	–	–		*−6.6*		*−7.0*		*−16.0*						–		
December	∂**PDSI**/∂⟨**T**⟩		**−5.5**	–	–		**−5.2**	**−4.5**	**−6.2**	–	**−10.7**		**−4.5**	**−3.8**	**−5.5**	–	**−11.3**	–	**−6.0**
	∂PDSI/∂Δ*T*		–		–				*−8.4*		*−17.4*						–		

Σ/5: Mean of all (significant and non-significant) $\partial\text{PDSI}/\partial\langle T\rangle$ and $\partial\text{PDSI}/\partial\Delta T$ coefficients.

Table 5
Proportion of standard deviation (S) computed from Thornthwaite (Th) and Blaney–Criddle (BC, parameterised for maize) version of PDSI: $qS = S(\mathrm{Th})/S(\mathrm{BC})$; and correlation coefficient (R) between the two PDSI version, for 1901–1999

	Szombath	Sopron	Mmóvár
qS	1.08	1.04	1.09
R	0.896	0.868	0.878

The standard deviation is somewhat larger in the non-vegetation (Th) case. Correlation is high, but not deterministic.

Thornthwaite method of potential evapotranspiration to those computed for maize plant applying the Blaney–Criddle formula. As indicated in Table 5, standard deviation of the previous PDSI series is slightly higher than of those implying plant coverage. In other words, plants can decrease the time fluctuation of the index. The proportion, however is not far from 1 and the correlation between the two series is also very high. This means, majority of our conclusions could also be valid for the alternative PDSI series, as well.

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