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Drought indices in meteorology.
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Drought indices in meteorology

Proper quantification of drought events is required for either their comparative spatial and temporal evaluation or for relevant predictability studies. Drought indices appear to be the simplest tools in drought analysis. The objective of this paper is to collect more or less "popular" indices, and to compare them as to their theoretical and numerical effectiveness. Indices are classified into four groups: precipitation indices, supply/demand (water balance) indices, soil moisture indices, and "recursive" indices. For each group, a few typical expressions are given and analyzed for their performance and comparability. Some empirical relations are established among the different indices. Those indices proved to be of highest utility in the delineation of meteorologically determined droughts which possess "memory" that is which actual values depend also on preceding values of the related meteorological elements. Such indices are the soil moisture, and Bhalme-Mooley or Palmer indices. Results are illustrated for some recent drought events in Hungary.

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Aszályindexek a meteorológiában. Mind az aszályos időszakok területi és időbeli összehasonlításához, mind előrejelelhetőségüknek vizsgálatához szükséges e jelenségek megfelelő számszerűsítése. Az aszályok vizsgálatának a legegyszerűbb eszközei a különféle indexek. E tanulmány célja a többé-kevésbé ismert aszályindexek összehasonlítása mind elmélet, mind a számszerűsíthető hasznosíthatóságuk szempontjából. Az indexeket négy típusba soroltuk: csapadékindexek, vízmérleg-indexek, talajnedvesség indexek és „rekurzív" indexek. Minden egyes csoport néhány főbb indexének alkalmazhatóságát, illetve az indexek összefüggéseit elemeztük. A meteorológiailag meghatározott aszályok leírásában azon indexek bizonyultak a leghatékonyabbnak, amelyeknek bizonyos „memóriája" van, azaz aktuális értékük a megelőző időszak meteorológiai elemeinek az alakulásától is függ. Ilyen index a relatív talajnedvesség, a Bhalme-Mooley és a Palmer index. Az eredményeket az elmúlt évek konkrét adataival illusztráljuk Magyarország különböző területeire vonatkozóan.

1. Introduction

> "There is a strong desire to develop indices for all factors in nature and society. .. If it is properly formulated, if its limitations are recognized .. then an index can be very useful. Its misuse and misinterpretation , however, may lead to the development of policies that are inappropriate for combatting environmental and societal effects of droughts. In particular, there should not be an overreliance on any single index to monitor droughts. " (Katz and Glantz, 1986; p. 770)

Among the extreme meteorological events, droughts are possibly the most slowly developing, often have the longest duration, and probably are the least predictable of all atmospheric hazards. Due to these characteristics, particularly their temporal character, droughts cannot be compared with other weather and climate extremes such as floods, hurricanes, lightening hail, cold, winter stroms, frosts or windstroms, which also significantly contribute to a nation's annual loss due to weather (*Riebsame et al.*, 1986). Because of their peculiar character, droughts deserve great scientific investigation. The problem involves a large number and variety of definitions, indicators, indices and methods of evaluation. As a consequence, almost all agrometeorologists, climatologists and agronomists engaged in this field have their own time series, methods and conclusions about the characteristics of drought episodes, even for the same regions.

Drought is considered as one of the man's worst enemies (*WMO*, 1975), because of its potential to continue for a long time over a large area and to have long-lasting effects. It expresses some sort of imbalance, arising from either extraordinary climatic variations or human activities (overconsumption of water, overgrazing and soil erosion, migration of population to places with insufficient water resources etc.). This dual character of the drought concept is stressed by *Kraus* (1977, p. 1009) who states that "by definition droughts are anomalies-deviations from a rainfall regime to which people, plants and animals have adapted as the local norm."

In spite of its special character (severity, recurrence, spatial pattern), societies — even those that are regularly affected by this hazard — fail to effectively adapt themselves to this adverse phenomenon. One of the typical causes is the short-mindedness: "with a return to above-mean rainfall levels, interest in dealing with the chronic (Sahelian) problems (revealed by the drought) will dissipate" (*Glantz and Katz*, 1977; Lamb, 1982). Moreover, adjustment is rather costly and the cost/loss ratio can hardly be assessed. Similar problems repeated during the recent drought incidences in sub-Saharan Africa (*Glantz and Katz*, 1985). As regards our region, we note that a new research program has been initiated recently in Hungary, following a four-year period of drought (1983—86) in various regions of the country. Of course, the main purpose of investigations of droughts, be of regional or large-scale extension, is the development of forecasting methods. In this relation, only some success can be mentioned, e.g., for the subtropical area, which is based on the El Niño—Southern Oscillation (ENSO-) related processes (*Bhalme et. al.*, 1983; *Bell*, 1986).

From a scientific point of view, the concept of drought is ambiguous. Some of the reasons for this problem are: (1) the water/moisture deficit may occur in the atmosphere, in water basins, in the hydrosphere, in the soil in accordance with the multistate and multiphase character of the water cycle; (2) the complexity of causes leading to a particular form of drought; (3) the regionally relative nature of drought (i.e., its significance depending on the local water supply/demand relations); and (4) the differential effects of water/moisture shortages on various natural and socioeconomic processes.

It is obvious that any form of drought originates from precipitation deficits, and that all other aspects of droughts are related to these inadequate precipitation amounts. But, a hydrological drought can occur even if anomalous rainfall was observed in the distant catchment area of the rivers. This emphasizes the need for the complex analysis of the areal and temporal characteristics ordifferent forms of drought. Moreover, while there are special disciplinary fequirements in drought investigations, the various approaches to and the related definitions of drought may not be separated. As *Wilhite and Glantz* (1985, p. 6) note: "Although it is useful to compartmentalize the various views of drought, the boundary separating these views is often vague." Not even the concept of meteorological drought can be isolated and analyzed in a „pure" form; it is closely related to the incidence of hydrological, agricultural and other droughts. Varga-Haszonits (1985) also points out the differences and similarities among the approaches of various scientists to droughts. Thus, the meteorological analysis should encompass the extended idea of meteorologically determined droughts.

Our goal was to collect a set of "popular" indices for droughts for a few months or longer, and to compare their theoretical and practical advantages, limitations, interrelations, and numerical effectiveness. These investigations showed only limited agreement among the drought/moisture indices. Actually, that agreement is a direct consequence of both the relative nature of drought and the related specific characteristics of all droughts and indices. It has been revealed that the highest performance can be achieved by such recursive indices as for instance, the Palmer-index. In areas, where soil moisture measurements or its reasonably good estimations are available, the respective (soil drought) indices are more advantageous.

The analysis concerns the meteorological and the "meteorologically determined" droughts. More exact estimations could be made by means of more specific (e.g., plant-specific) indicators of moisture deficiency, taking into account the particular timing and regional extent of water demand during the consecutive developmental stages of plants of plants or for different forms of socioeconomic activities.

2. *Drought definitions and the introduction of indices*

There are many conceptual definitions of drought in the scientific literature. Quite a few of these definitions have been reviewed by *Wilhite and Glantz* (1985), by *Andreeva and Sazonov* (1987), *Girskaya and Sazonov* (1979) and by the WMO (1975), but according to the authors of the first work (see also *Sadowski*, 1984 and *Bagrov*, 1986) the lack of a precise (and objective) definition of drought can be an obstacle to understanding this phenomenon.

Meteorological drought is often identified with atmospheric drought and defined simply with below mean precipitation amounts, sometimes combined with parameters of air temperature, humidity and wind velocity. This type of apparently oversimplified definition is reported by *Sadowski* (1984): "Atmospheric drought, with rainfall deficit in comparison to normal values." Similarly, *Bell* (1986) notes: "drought... is the unfulfillment of expected rainfall." As a matter of fact, the general term drought is not identical to precipitation deficit, but rather it is usually the consequence of below-normal precipitation *(Oladipo*, 1985). These attributes have been quantified in a more general drought definition given in WMO (1986, p. 2) that relates all other forms of drought to the precipitation deficit: "Drought is by definition a sustained, regionally extensive, deficiency in precipitation... All other definitions of drought are related to the effect or impact of below normal precipitation on agriculture, water resources, social and economic activities." More precisely, atmospheric drought is a *state* of the atmosphere (not an instantaneous but an integrated state for a longer period of time) that results in less than average (for that peirod) precipitation amounts and/or below normal atmospheric humidity. Other authors extend the concept of meteorological drought to soil droughts. Indeed, the soil moisture balance is closely connected to the atmospheric processes in terms of recharge into and loss water from the surface layer of the soil. Thus, the low precipitation, high temperatures and consequently low air humidity may cause the soil to dry out. Nevertheless, soil drought is a specicif form of this phenomenon, which is closely but not exclusively controlled by meteorological processes.

Agricultural drought occurs when the available soil moisture is inadequate to meet the evaporative demand by crops. This deficiency may result either from an unusually low moisture supply or an unusually high moisture demand (WMO, 1975). For its numerical analysis, all the abovementioned parameters of the atmospheric and the soil dryness/wetness can be used, though relevant agrometeorological studies must be plant-specific. Some compromise can be achieved if the meteorological approach is refined to account for the plant-specific time period (vulnerable stages of plant development in regard to the water availability), or the depth of the actual root zone in the case of soil moisture evaluations.

Hydrological drought refers to a period of below-normal streamflow and depleted reservoir storage, and it can also be treated as a meteorologically determined phenomenon. A similar definition is given by *Das* (1983). From a meteorological standpoint, the snow accumulations in the catchment areas, the runoff term in the surface water balance, or the potential evaporation for a specific case are of significance.

For scientific analysis, the different types of droughts are usually quantified. For this purpuse various indices are introduced.

In the past, the concept of dryness/wetness was primarily developed for studies of the geographical distribution of climates, or the determination of climate zones. Numerical indicators of moisture availability introduced for the characterization of climatic zones are termed "aridity indices". Aridity expresses a relatively permanent state of climate. Drought is a "shorter-term" phenomenon. Moreover, drought means a moisture deficit relative to the local (climatically and ecologically/socially established) average moisture balance. Therefore, aridity indices must be used with proper care for drought studies. In other words, the intensities of drought incidence in various parts of the world can be compared only if the corresponding anomalies are normalized in relation to local expected conditions. Indices have been specially developed for drought analysis. Various indices are used, depending on the purpose of an investigation (whether persistence, intensity or recurrence is in question) or on the availability of meteorological data.

The above mentioned disciplinary aspects of drought studies also affect the choice and the utility of indices. From a meteorological standpoint, the different types of droughts can be delineated by means of meteorological parameters in a wider sense (parameters of the state of the atmosphere "in itself", and those which have significant effects on the agricultural or hydrological water balance through evaporation or evapotranspiration, and, in turn, through available soil moisture amounts or levels in water reservoirs, etc.).

3. *Drought indices*

3.1. Indices of cumulative precipitation anomalies. All forms of drought are related to some antecedent and relative precipitation amounts for a time period lasting from a month (or less) to years. But drought may occur well after this anomalous precipitation period. For example, agricultural drought may well be the consequence of a rainfall shortage during the previous autumn or winter. Therefore, the simplest measures of drought severity only account for precipitation characteristics.

P r e c i p i t a t i o n   i n d e x (AI): $\Delta P = P - \mu(P)$(mm), (1)

where P denotes the monthly (or longer-time) precipitation amount and $\mu(P)$ is its "normal" (or mean) value. This index is an appropriate variable for the analysis of the precipitation fluctuations at a single station. Kane and Trivedi (1986) investigated long series of precipitation anomalies for different sites to reveal significant periodicities.

Deviation of the actual amount from normal is applied to drought/flood identification by many authors. *Glantz and Katz* (1985) used $(\sum_1^M \Delta P_m)/M$, the average of these anomalies for stations, as a regional index in the analysis of drought for Sub-Saharan Africa. *Adámyné and Csomor* (1986) retained only the signs of station anomalies to determine the regional extent of drought episodes, depending on which sign was most prevalent for different regions in Hungary.

It goes without saying that the anomalies for nonhomogeneous regions or larger areas with different climatic conditions are not comparable. To avoid this problem, either relative amounts, or standardized values should be used. The relative value is used in two forms:

R e l a t i v e   p r e c i p i t a t i o n   a m o u n t (rP): $P/\mu(P)$ ($\times 100\%$), (2)

R e l a t i v e   a n o m a l y   i n d e x (rAI): $[P - \mu(P)]/\mu(P)$ ($\times 100\%$). (3)

The rP-index is an indicator of drought if its value is less than some established threshold level, e.g., 75% on an annual basis, or 60% for a particular month. The use of other, slightly different levels was also suggested in *WMO* (1975). *Bunting* et al. (1976) have analyzed the long-term series of monthly and annual area-averaged rP-values to test the hypotheses of the existence of trends and periodicities for the Sahelian region. For a recent world-wide drought assessment (*WMO*, 1986), the following criterion was applied: annual $rP < 60\%$ for more than two consecutive years (combined with an additional condition that the area affected must be 50% or more of the region).

S t a n d a r d i z e d   a n o m a l y   i n d e x (SAI):

$$P' = [P - \mu(P)]/\sigma(P), \tag{4}$$

where $\sigma(P)$ is the standard devition of P. A very detailed analysis of SAI was accomplished by *Katz and Glantz* (1986). It proved aeffective for the delineation of regional precipitation anomalies because of the reasonable comparability of station SAI-values. For a single station, the usual probability levels can be applied to choose particular threshold values in accordance with the hypothetical distribution of SAI (Gamma or Gaussian distribution for longer periods of time provide a good statistical fit). The variance of a regional (average) index depends on the number of stations and their correlations.

Kraus (1977) applied area-averaged SAI on an annual basis for two monsoonal (African and Indian) regions to reveal persistence in the occurrence of rainfall anomalies. This hypothesis was later refuted by *Katz* (1978). *Nicholson* (1979) returned to the same problem and *Lamb* (1982, 1983) also intensively used this index for the Sub-Saharan area. *Bhalme and Mooley* (1980) introduced the rainfall anomaly index, weighted by the reciprocal of coefficient of variation, which is actually equivalent to the SAI multiplied by $\mu(P)$.

R o o y a n o m a l y i n d e x (RAI): $3P\,[h-\mu(P)]$, if $P>0$ (5)

$-3P/\,[l-\mu(P)]$, otherwise,

where h = mean of ten highest P-values and l = mean of ten lowest P-values, within the sample. For RAI, nine abnormality classes were given *(Oladipo*, 1985), ranging from extremely wet to extremely dry classes.

Similary for the standardization of station precipitation anomalies, an analogous procedure can be suggested for a sequence of monthly amounts:

A v e r a g e SAI (ASAI): $[\sum_1^K A(P_k)/\sigma(P_k)]/K$. (6)

3.2. Other indices of atmospheric drought. The water vapor saturation deficit is commonly used for the characterization of atmospheric drought, although the temporal scale for similar analyses is usually much shorter than a month *(Pasechniuk and Sennikov*, 1983; *Levitt*, 1958):

S a t u r a t i o n d e f i c i t: $d = E_s - E = E_s\,(1-\varrho)$ (hPa), (7)

where ϱ is the relative humidity, E is the vapour pressure, and E_s is the saturated vapour pressure. Threshold values for the identification of atmospheric dryness for a shorter period, the "sukhovei" are: 20 – 29 hPa = weak, 30 – 39 hPa = moderate, 40 – 49 hPa = intense, >50 hPa = very intense.

Dry conditions of longer time intervals can be described with the number of days of at least mild dryness. *Pasechniuk and Sennikov* (1983) have shown the strong correlation between this dryness measure and the precipitation amount during the growing season (correlation coefficient reached between 0.7 – 0.8). Air temperature and wind can also play important roles in the formation of severe atmospheric dry situations.

3.3. "Supply/demand" indices based on the water balance. Precipitation is the dominant form of water supply for the soil water balance. It is evident that for a more precise delineation of drought conditions, the "demand" side of this balance must also be considered. Water demand is usually expressed through the potential evaporation or, as a simplest approximation, the average temperature of the analyzed period. Maximum evaporation can be assessed from the radiation balance, too.

L a n g' s r a i n f a l l i n d e x (LI): P/T (mm/°C), (8)

where T is the mean air temperature for the given monthly (or longer) period.

d e M a r t o n n e a r i d i t y i n d e x (dMI);

$12P/(T+10)$, the monthly index (mm/°C) (9)

$P/(T+10)$, the annual index.

Originally, LI and dMI were developed as aridity indices, although they can be used for the detection of drought episodes (*WMO*, 1975). *Chowdhury* and *Hussain* (1983) apply dMI for aridity analysis (the select arid areas with annual dMI < 20) and choose a different criterion for drought studies.

S e l y a n i n o v' s h y d r o t h e r m a l c o e f f i c i e n t (SHC):

$P/(\sum_{t>10\,°C} t/10)$ (mm/°C), (10)

where t denotes the consecutive daily mean air temperatures. This index (known also as HTC) is often employed for agrometeorological investigations *(Sapozhnikova*, 1958; *Sinicina* et al., 1973; *Ajtay*, 1977.) The established threshold values for drought (and aridity) classification are: 0.4 – 0.7 = very dry, 0,7 – 1.0 = dry, 1.0 – 1.3 = insufficiently wet, >1.3 = wet. A simpler criterion is recommended in WMO (1975) for drought indentification, namely, SCH < 0.5. SCH can be directly formulated for monthly values (*Bagrov*, 1983) in the form P/(3T).

Thornthwaite Index (WMO, 1975): $1.65[P/(T+12.2)]^{10/9}$ (11)

Ped's Drought Index (PDI1) $\Delta T/\sigma(T) - \Delta P/\sigma(P)$. (12)

Bagrov (1983) transformed PDI1 into a wetness index to demonstrate the limitations of this index (and the SHC) for yield estimation. Recently, the statistical properties of the original PDI1 index were investigated (*Bagrov*, 1986). The simple criterion of drought, namely, the validity of inequality of PDI1 > 2, is appropriate only when the independence of T and P is assumed for a single month. If these requirements are not satisfied, the threshold value should be modified: (1) for a K-month period, the $2/\sqrt{K}$ threshold value is the proper choice; (2) the actual dispersion of the index for a single month equals $2(1-r)$, where r is the correlation between T and P.

It was mentioned above that the temperature in these indices simply substitutes for the more complex potential evaporation term. A few indices incorporate PE explicitly.

Potential water deficit: $PE-P$, (mm) (13)

Potential evapotranspiration ratio: PE/P. (14)

Moisture availability index (MAI): P/PE. (15)

Biswas and Khambete (1983) used MAI for climate potential assessment for crop production in the arid zone of India. It is believed that the same levels of MAI values can be used for meteorological drought evaluation: ≤ 0.3 = very dry, > 0.3 = classification depends on the relative length of time units with MAI > 0.3 within the vegetation period.

Dryness ratio: $(R/L)/P$, (16)

where R is the surface radiatoin balance. This ratio is primarily a measure of dryness or wetness that is commonly used for defining climatological regions (*Péczely*, 1979).

Finally, let us consider indices with ET instead of the precipitation amount (ET stands for the actual evapotranspiration). These indices more closely express the drought/flood conditions from the agricultural (or hydrological) perspective. The relation of these two indicators of water supply can also be evaluated by means of the Thornthwaite moisture index, $(P/ET-1)\times 100\%$.

Relative evapotranspiration: ET/PE (100%). (17)

This index was used by, among others, *Patel et al.* (1986) for agricultural drought classification in relation to crop yield estimation.

Bowen ratio: $(R-L\times ET)/(L\times ET) = (R/L)/ET-1$. (18)

This ratio expresses the relation of the heat flux to turbulent eddies in the boundary layer to that part of the radiation balance which is used for evaporation (*Hare*, 1983).

3.4. Soil moisture indices.

Relative soil moisture content: W/AWC ($\times$100%). (19)

with W and AWC denoting the actual soil moisture and the available (or disponsible) water capacity for a fixed soil depth (e.g., the upper 1-meter layer or the root zone for a given plant).

Besides this well-known ratio, an extended form of Ped's drought index incorporates the standardized value of the soil moisture amount *(Davidov and Ped*, 1983):

Ped's drought index, (PDI2):

$$\Delta T/\sigma(T) - \Delta P/\sigma(P) - \Delta W/\sigma(W), \tag{20}$$

3.5. „Recursive" indices. Indices describing the moisture conditions for a relatively long time period through the integrated values of the related meteorological elements provide only a rough picture of the potentially adverse conditions within this period. It is thought that above al the cumulative effect of prolonged moisture deficits (month by month) should be properly expressed. For this purpose "recursive indices" are formulated. The simplest way to do so is to consider the progression of the accumulation of precipitation anomalies, which is expressed by the "residual mass curve" introduced by *Foley (WMO*, 1975).

Foley's cumulative anomaly index (FAI):

$$FAI_1 = \Delta P_1,\ FAI_k = FAI_{k-1} + \Delta P_k \text{ for the } k\text{th month.} \tag{21}$$

For the comparability of FAI-values of various stations and periods, normalization can be made, $\sum_1^K[\Delta P_k/\tilde{P}]/K$, where $\tilde{P}$ is the annual precipitation amount for the given station.

Bhalme and Mooley (1980) defined a "recursive" drought/flood index based on monthly precipitation amounts. Conceptually, this index (the BMDI) can be considered as an essentially simplified version of the well-known PDSI, the Palmer Drought Index *(Palmer*, 1965).

Bhalme-Mooley Drought Index (BMDI):

$$i_0 = 0;\ i_k = c_1\, i_{k-1} + (SAI)_k/c_2 \text{ monthly index} \tag{22}$$

$$\text{BMDI} = (\textstyle\sum_1^K i_k)/K \text{ for a period of } K \text{ months.}$$

The coefficients were derived for Indian observations: $c_1 = 0.5$ and $c_2 = 0.4855$. The regional extent of drought is expressed as a Drought Area Index, being defined as the percentage of area with averaged BMDI value below a certain negative threshold *(Das*, 1983). Coefficients in the BMDI formula were recalculated using data from Nebraska for the growing season, April-September *(Oladipo*, 1985): $c_1 = 0.452$ and $c_2 = 0.4267$. Slight variations appeared for a larger data base observations in the interior plains of North America: $c_1 = 0.548$ and $c_2 = 0.4267$ *(Oladipo*, 1986).

Palmer Drought Severity Index, (PDSI). The PDSI is based on the thorough analysis of the elements of surface water balance and on the comparison of their actual values to their climatically or physically attainable (potential) values *($ET \sim PE$* for the evapotranspiration, $W \sim AWC$ to estimate soil moisture amount, etc.).

A particular locality can be characterized by the ratios of the averages of the actual and potential values for all constituents in the water balance: α = ratio of averages of ET and PE, β = ratio of averages of actual and potential recharge, γ = the same ratio for the runoff, δ = the same ratio for the moisture loss, where the recharge and the loss express the change of the moisture content in the root zone of soil with opposite signs (i.e., negative changes assigned to the loss term). For a particular month, the "expected" value of the precipitation, or the value "climatically appropriate for existing conditions," the "CAFEC value" (a term introduced by Palmer) is calculated from the water balance equation: $\hat{P} = \alpha\, PE + \beta\, PR + \gamma\, PRO - \delta\, PL$, where PE, PR, PRO and PL are the potential values of evaporation, recharge, runoff and loss for that month, respectively. The deviation of the actual P from its CAFEC value weighted by a climatic factor (valid for the month and the area) provides a moisture anomaly index, $Z = K\,(P - \hat{P})$.

The so-called final PDSI is derived by a recursive equation:

$$PDSI_k = PDSI_{k-1} + Z_k/3 - 0.103\ PDSI_{k-1}. \tag{23}$$

The initialization depends on criteria for determining and ending times of wet and dry periods. Palmer scaled his index in accordance with a prescribed set of categories, which was recently adopted for the derivation of the BMDI: < -4.0 = extreme drought, $(-4, -3)$ = severe drought, $(-3, -2)$ = moderate drought, $(-2, -1)$ = mild drought, $(-1, +1)$ = near normal, and similar categories of slightly, moderately, very wet and extremely wet periods. The growing season can be characterized with the average PDSI values for the corresponding months.

PDSI is rather popular because of its quasi-invariance under different changes in climatic conditions for different regions (*Wilhite*, 1983; *Hecht*, 1983; *Karl*, 1983; *Oladipo*, 1985; *Rao et al.*, 1986).

4. Some properties and limitations

To evaluate the abnormality of P, the precipitation amount for a particular time period (month, season, etc.), this value must be interpreted in terms of its empirical probability distribution. For this purpose, either certain levels (thresholds) are used, such as the record amount (P_{min}, P_{max}); the lowest decile (P_{ld}), quartile (P_{lq}), median; or some "integrated" measures of dispersion of the distribution, e.g., the standard deviation, the mean absolute deviation, the average of the ten lowest values in a record; or the mean (expected) value of the distribution may also be mentioned among these characteristics.

Keeping in mind that the regional extent of a drought episode is commonly determined by mapping the individual index values or calculating areal averages, the values should be standardized first. This means transforming the original values into $P' = [P - \mu(P)]/d(P)$, where $d(P)$ denotes some general measure of dispersion. Typical examples of such a standardization use one of the above thresholds values, as in the case of indices defined by (3), (4) and (5): $\mu(P)$ for rAI, $\sigma(P)$ for SAI or $[l(P) - \mu(P)]/3$ for RAI. Such a procedure is performed to eliminate regional differences in local climates, or, in other words, to archieve regional comparability of various meteorological stations. Implicity, one would also expect that standardized indices for different areas (stations) ought to quantify drought events with similar socioeconomic (agricultural, hydrological, etc.) impacts in nearly identical ways. For the evaluation of the consequences, it can be assumed that the natural ecosystems and societies have adjusted to the "surrounding" local or regional climate and other environmental conditions (*Palmer*, 1965; *WMO*, 1986).

Among the precipitation-based indices, the rAI has a formally different character because its normalizing term $[\mu(P)]$ is not a measure of dispersion. This problem is partly solved using the finding reported by *Katz* and *Glantz* (1986), who have shown that for the Sahelian area the mean and standard deviation of precipitation amounts are closely related: the rainfall variability (for the April–October period) tended to increase. A similar statistical relationship takes place in Hungary (*Figure 1*). This approximate linear proportionality between the mean and standard deviation has a more general basis. Supposing that the monthly precipitation amounts obey the gamma distribution with the density function, $f(P) = P^{\gamma-1} \exp(-P/\beta)/[\beta^\gamma \Gamma(\gamma)]$, where β, γ are the scale and shape parameters, respectively, the expected value equals $\beta\gamma$, while the standard deviation is $\beta\sqrt{\gamma}$. If for the case of two stations only the means differ, $\mu(P_1) = c\,\mu(P_2)$ (i.e., their density functions are of the same form with identical shape parameters), then the relations between the scale parameters and the two standard deviations are given by $\beta_2 = \mu(P_2)/\gamma = c\mu(P_1)/\gamma = c\beta_1$, $\sigma_2 = \beta_2\sqrt{\gamma} = c\beta_1\sqrt{\gamma} = c\sigma_1$. Instead of this linear relationship, *Drozdov* (1956) came to a slightly different formula, $\sigma^2(P) = \alpha\mu^2(P) + k\mu(P)$, which again leads approximately to the previous result if the second term is small. For very dry regions, the ratio $\sigma(P)/\mu(P)$ differs considerably from a constant and significantly increases with decreasing $\mu(P)$. Based on this relationship, a nearly equal performance can be assumed for all the "standardized". rainfall indices. In particular, for an individual station, after the actual precipitation (for the given period) is classified (compared with a threshold value C), say $P < C$, then all other simple moisture anomaly indices determine a corresponding class on a different scale (different units according to the transformation definition of the index); that is, $SAI < C/\sigma(P)$, $rP < C/\mu(P) + 1$, $rAI < C/\mu(P)$, etc.

In spite of the advantageous properties described above, these indices do not meet the more rigorous requirements concerning the independence of the distribution of regional indices from any changes in the set of individual stations. Katz and Glantz (1986) noted that the average of local SAI values for M stations has a variance $1/M$ for uncorrelated and equally distributed data. For rAI, a similar expression takes place: var $[\Sigma_1^M(rAI)]/M] = \sigma^2(P)/\mu^2(P)/M$.

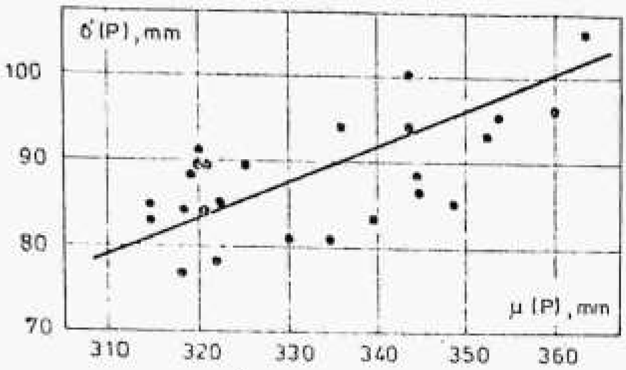

Fig. 1. Relationship between μ(P) and σ(P) based on 80 years averages for 25 meteorological stations in the Hungarian Great Plane (precipitation amounts for the growing season: Apr–Sept)

All indices treated above characterize the precipitation anomaly of a period of K months as a whole, neglecting the time dependence of the monthly anomalies during that period.

A refined (and more effective) approach is performed in the simplest way with the use of the sequence of accumulated anomalies (or the elements of the "residual mass curve"): ΔP_1, $\Delta P_1 + \Delta P_2$, $\ldots$, $\sum_1^K \Delta P_k$ (WMO, 1975). However, the precipitation anomalies for different months produce different impacts; thus the amounts of the natural water supply should be compared with the "established" water demand for the particular time period. This means that it is more reasonable to use the standardized precipitation values so that we can introduce the series of cumulative monthly standardized anomalies, $\sum_1^k \Delta P_j/\sigma)P_j)$, or, for a period of K months, the average of monthly values: $\mathrm{ASAI} = \sum_1^K[\Delta P_k/\sigma(P_k)]/K$.

Among the indices based on the precipitation term, the BMDI defined by *Bhalme and Mooley* (1986) appears to be a unique example reflecting the development of a dry episode through a "recursive" definition: $i_k = c_1 i_{k-1} + [\Delta P_k/\sigma(P_k)]/c_2$, $\mathrm{BMDI} = (\sum_1^K i_k)/K$. The BMDI for a given period can be expressed in the form ($c_2' = 1/c^2$, $P' = \Delta P/\sigma(\sigma P)$:

$$i_k = c_1 i_{k-1} + c_2' P_k' = c_2'(c_1^{k-1}P_1' + c_1^{k-2}P_2' + \cdots + P_k') = c_2' \sum\nolimits_1^k c_1^{k-j} P_j',$$

$$\mathrm{BMDI} = \sum\nolimits_1^K i_k/K = c_2'[(1 + c_1 + c_1^2 + \cdots + c_1^{K-1})P_1' +$$

$$+ (1 + c_1 + c_1^2 + \cdots + c_1^{K-2})P_2' + \cdots + P_K'].$$

Obviously, the "weights" for previous months are gradually reduced with increasing values of K (the length of the period): $(1 + c_1 + c_1^2 + \ldots + c_1^{K-1})/K \to 0$.

However, the "weight" of the first month increases with respect to that of the last month, so that their ratio tends from an initial value of 1 to 2 as K increases, which is possibly an unreasonable property.

One question remains: is the "recursive index" more straightforward than the "directly" determined indices (like SAI or rP) for the identification of drought episodes. Of the latter indices, the ASAI corrected by a factor of c'_2 is "closest" to the BMDI because both are based on consecutive SAI values of months within a given period. (The correction by c'_2 results in the equality of ASAI and BMDI for the first month). As a matter of course, an ASAI-type moisture index contains more information on the precipitation "history" of the period than any other index of the total precipitation anomaly for the whole time interval. Let us consider, for example, the SAI-value corrected by the $\sqrt{K}$ factor ($P = \sum_1^K P_k$):

$$\frac{1}{\sqrt{K}} SAI(K) = \frac{1}{\sqrt{K}} \Delta P/\delta(P) = \frac{1}{\sqrt{K}} \sum_1^K P'_k \frac{\sqrt{K}\sigma(P_k)}{\sigma(P)}.$$

Thus, for uncorrelated consecutive P_k-values:

$$SAI(K)/\sqrt{K} = \sum_1^K P'_k/K = ASAI(K).$$

But these assumptions are usually not valid; as illustrated in *Table 1* for three Hungarian stations (from an 80-year observation period; K = 6).

TABLE 1

Ratio of standard deviations of precipitation amounts for sequence of months and for the whole period

$\sqrt{K}\sigma(P_k)/\sigma(P)$	Apr	May	Jun	Jul	Aug	Sept
Baja	0.78	1.15	1.07	0.95	1.04	0.98
Kalocsa	0.77	0.97	1.15	1.03	0.97	0.89
Nyiregyháza	0.66	0.82	1.05	0.97	0.97	0.84

Let us introduce two periods lasting for K months (e.g., two growing seasons) so that all months are "normal" with zero SAI values, except for the first and last month:

$1: SAI_1 = -2, \Delta P_2 = 0, \ldots, \Delta P_{K-1}, \Delta P_K = 0$

$2: \Delta P_1 = 0, \Delta P_2, \ldots, \Delta P_{K-1} = 0, SAI_K = -2.$

The index of total anomaly is independent of the time when the single anomalous month occurred. Therefore, for both periods we obtain the same value: $ASAI' = c'_2\, ASAI = c'_2(-2/K)$, while the BMDI values are significantly different: $BMDI_1 = c'_2(-2/K)[(1-c_1^k)/(1-c_1)]$, $BMDI_2 = c'_2(-2/K)$. Obviously, the two indices coincide only when the precipitation deficit takes place during the last month of the period, $ASAI' = BMDI_2$. In this case, $BMDI_2 = -0.7$ (provided $K = 6$), which characterizes a near-normal period. With respect to the former period, the consecutive BMDI values are as follows: -4.0 (extreme drought), -3.0 (severe drought), ..., at last for the sixth month, -1.3 (mild drought). Thus, the "recursive" index reflects the fact that the precipitation anomalies during the given period have their agricultural or hydrological impacts mainly on the water balance of the soil. That is why these effects are delayed and smoothed.

Similar "recursive" indices can be deduced in another, and possibly more natural, way. The effects of serial precipitation anomalies accumulate as a soil moisture deficit (surplus) or a particular state of a plant. The simplest soil moisture balance can be given by $W_k = W_{k-1} - ET_k + P_k$, where W denotes the available (disponsable) soil moisture. A soil moisture deficit index can be formulated, for instance, with the following equation: $\tilde{i}_k = (W_k - AWC)\eta$, $= 4/AWC$, which gives -4 for the driest situation, $W = 0$, like PDSI or *BMDI*. Substituting a simple paramererization scheme for the actual evaporation (*Faragó*, 1985), a BMDI type recursive index is defined, which depends only on the monthly precipitation amounts: $\tilde{i}_k = \tilde{i}_{k-1} - \varepsilon W_{k-1} + P_k = a\,\tilde{i}_{k-1} + bP'_k + c$ where $a = 1 - \varepsilon$, $b = \sigma(P_k)$, $c = \eta\mu(P_k) - 4\varepsilon$.

Indices based exclusively on precipitation provide a simplified insight into the drought phenomenon. Actually, it is a supply/demand problem. For an agriculturist, the water demand refers to the amount of water for evapotranspiration (ET). Because data for ET are not routinely collected, temperature-based expressions are used for the estimation of the potential evaporation (PE); for instance (*Antal*, 1968): $PE = 0.74\ (E_s - E)^{0.7}(1 + aT)^{4.8}$. (mm/day; $a = 1/273$). If the purpose of quantification is the determination of climatological or agroecological regions in a general sense, then the radiation balance is used for the (potential) demand estimation.

The water supply/demand balance is expressed either in the form of a ratio or as a difference, e.g.:

	Lange's LI	Selyaninov's SHC	Ped's PDI1	Pot. water deficit	Dryness ratio
Supply:	P	P	$\Delta P/\sigma(P)$	P	P
Demand:	T	$\Sigma t/10$	$\Delta T/\sigma(T)$	PE	R/L

There is only a slight relationship between the precipitation-based and the balance-type indices. *Bagrov* (1986) has analyzed the statistical properties of Ped's index. Besides the estimation of the expected recurrence time of drought, the statistical distribution of this index was also investigated.

This distribution is close to normal (Gaussian) with zero mean and dispersion $\sigma(PDI1) = 2(1 - r)$, where r denotes the correlation coefficient between T and P (for a single month). During the summer season, this coefficient is usually slightly below zero. Bagrov's conclusions probably hold for other locations at middle latitudes and definitely have a physical background (through the cloud-insolation coupling and the cooling effects of precipitation). For a Hungarian station, Baja, we obtained the empirical value $r = r(P, T) = -0.2$. This gives some basis for connecting the "P-based" and "supply/demand" indices (at least for the summer or growing season) in drought idenfification, however, the later indices evidently include more information for the analysis of meteorologically determined hydrological or agricultural droughts.

Most straightforward indices incorporate the elements of the soil moisture balance. The most sophisticated of those indices is the rather popular Palmer Drought Severity Index (PDSI). Despite its regular computation in some countries, no attempt has been yet made to adopt it in Hungary. Despite of its advantageous properties, even the PDSI cannot be considered universa Raol.

and Subramaniam (1986) refer to a severe drought event in the Indian subcontinent during 1918 for which Bhalme and Mooley showed the unreliability of the PDSI; to correct this problem, they recalculated Palmer's coefficients. *Oladipo* (1985) demonstrated the nearly comparable performance of simple precipitation based indices, such as Roy's RAI and, to an even greater extent, the BMDI and PDSI for the purposes of single-station analyses.

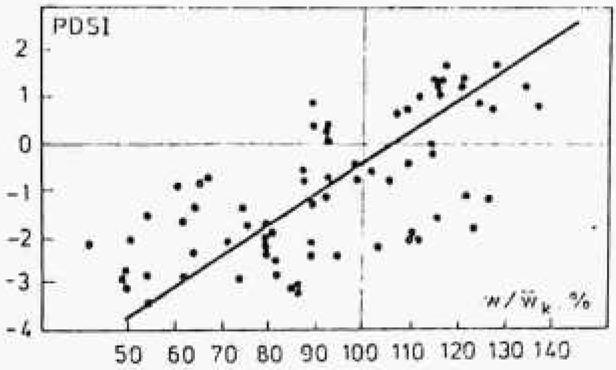

Fig. 2. PDSI versus relative soil moisture content divided by its normal value (%) for the corresponding month. The sample consists of all months of years 1981 – 86 for station Baja. The coefficient of correlation: r = 0.66

The PDSI is a "recursive" index, and is related by definition to the soil moisture balance in the way shown above. If the "established" water supply/demand regime for a given region is identified with the long-term average of the relative (available) soil moisture w then we may expect a good relationship between PDSI and the deviations of w from its normal, or their ratio. The latter is depicted in *Figure 2* for the station of Baja in Hungary. Fitting a linear regression between these variables provides a coefficient of correlation 0.66 for monthly values from the period 1981 – 86.

5. Illustration of the performances of some indices

We have analyzed the general aspects of the performance and comparability of different drought (moisture) indices in the previous section. We now summarize some results of the numerical studies. All calculations were based on Hungarian data series of precipitation, air temperature and soil moisture measurements.

TABLE

Drought indices in the growing season, Baja (1975 – 1986)

Year	rP%	RAI	SAI	SHC (mm/°C)	PDII	PE/P	ET/P	ET/PE (%)	W/AWC (%)	BMDI	PDSI
1975	143	2.7	1.3	1.5	−1.6	1.3	0.9	70	90	2.5	3.0
1976	80	−1.7	−0.8	0.9	−0.1	2.5	1.4	56	59	−1.4	−0.2
1977	78	−1.9	−0.9	0.9	−0.1	2.6	1.5	57	60	−1.5	−0.8
1978	95	−0.4	−0.2	1.1	−1.7	1.9	1.2	65	72	−0.1	0.7
1979	81	−1.6	−0.8	0.9	0.6	2.6	1.5	58	53	−1.1	−0.7
1980	98	−0.1	−0.1	1.2	−2.0	1.8	1.2	64	69	0.3	0.5
1981	93	−0.6	−0.3	1.0	0.2	2.2	1.4	64	67	−0.8	−0.0
1982	85	−1.3	−0.6	0.9	0.9	2.4	1.5	63	62	−0.9	−0.1
1983	93	−0.6	−0.3	0.9	1.3	2.5	1.2	48	38	−0.9	−2.7
1984	88	−1.1	−0.5	1.0	−1.0	2.1	1.2	62	65	−0.8	−2.3
1985	78	−1.9	−0.9	0.9	−0.1	2.4	1.5	62	72	−1.2	−1.5
1986	97	−0.3	−0.1	1.0	0.0	2.1	1.4	68	75	−0.1	0.2
Drought	(50; 70]	(−3.0; −2.0]	(−1.0; −0.5]	[0.7 1.0)	(1.5; 2.0]	[2.5; 3.5)	[1.1; 1.9)	45; 55)	[40; 50)	(−4.0; −2.0]	(−4.0; −2.0]
Extreme drought	≤ 50	≤ −3.0	≤ −1.0	< 0.7	> 2.0	≥ 3.5	≥ 1.9	< 45	< 40	≤ −4.0	≤ −4.

First, the relatively long-term series (1901 – 1986) of various indices were compared for a single meteorological station Baja located in the SW-part of the Hungarian Great Plain. As an illustration, *Table 2* summarizes the values of several indices calculated for the period 1975 – 1986. Apparently, the rainfall amount during the last eleven growing seasons were more or less anomalous according to most indices. Even this short period exhibits the occasionally considerable difference in the performance of drought indices. The growing season of 1983 was exceptional in accordance with the values of indices possessing with "memory". Though the rainfall amount was near normal (rP = 93%) during this period, the values of W/AWC and PDSI (38% and −2.7, respectively) are the smallest for this years, which is a consequence of the precipition deficit of the previous months. Such properties of these indices are very important, especially when we compare their values for growing seasons. During these months, most areas in Hungary receive only about 60 percent of the annual totals (for instance, at Baja it is 58 percent); therefore, the preseason precipitation during the winter months has a very importent role in the water balance of soils. We note *(Table 2)* approximetely parallel runs of rAI and SAI, that for the recursive indices BMDI and PDSI, and to a less extent, for the relative soil moisture (W/AWC) and PDSI. (The latter is most evident for the extremely dry events, i.e. for the extremely small values of these indices.)

More detailed investigations of the drought episods have been made for the years 1983 and 1986, including the analysis of spatial distribution of the index values. In 1983, the drought period started on the Great Plain in the previous winter and lasted through the spring and summer seasons. In Transdanubia the previous winter was wet, and this accumulated wetness in the soil was enough to sustain good moisture storage during the following spring. The summer was very dry in this part of Hungary, as it is shown by the soil moisture index W/AWC *(Fig. 3)* according to which an extreme drought incidence is indicated in the Great Plain area. The potential evapotranspiration ratio (PE/P) gives a much milder classification even for the longer period of the whole vegetation season. Most values of this index do not exceed the threshold 3.5, i.e. a value that separates the classes of "drought" and "extreme drought" events. Such a contradictionary behaviour of these indices is a trivial consequence of their properties. The soil moisture balance and the related index express in a certain way even the below normal-normal moisture conditions of the preceding (winter) period that was characteristic for the most part of the Great Plain (just in contrary to the situation in the Transdanubian area).

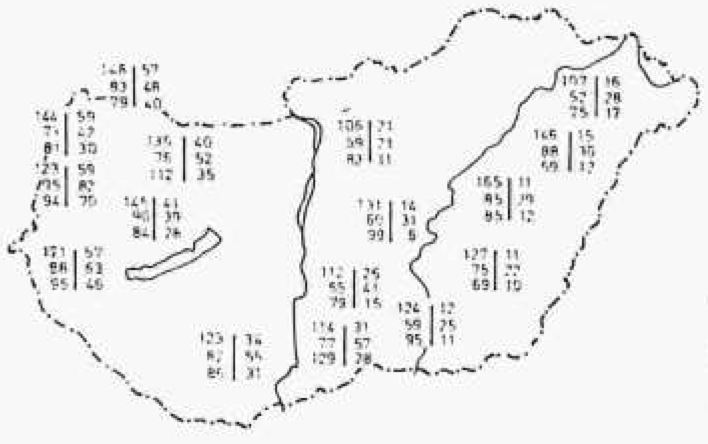

Fig. 3. Spatial distribution of some indices characterizing the 1983 growing season:

rp for previous winter	rp in Apr – Sept
rp for previous spring	PE/P in Apr–Sept
rp for previous summer	W/AWC in June – Aug

Fig. 4. Values of some indices characterizing the 1986 growing season:

rp for previous winter	rp for autumn
rp for previous spring	W/AWC in Jul – Sept
rp for previous summer	W/AWC in October

The character of the drought during the 1986 growing season was different to some extent because the previous winter season and the summer as well were considerably wetter in the whole country (*Fig. 4*). However the autumn brought about an extremely dry period throughout the Eastern part of the country, as well as in the E–SE part of Transdanubia.

6. Conclusions

Summarizing the main results of the above comparison, we have found some similarities among the different types of drought indices. Apparently, the precipitation indices possess similar features when applied to individual stations, yet some standardization is needed before determining regional indices. There is a much weaker connection between the precipitation indices and the balance-type expressions, and even less of a connection with the recursive indices. Therefore, it seems impossible to reduce the existing multitude of drought (moisture) indices to a single one. At the same time, for the delineation of meteorologically determined droughts, those indices that possess "memory" (the soil moisture indices and the recursive indices) are of highest utility.

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