IDŐJÁRÁS

QUARTERLY JOURNAL
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CONTENTS

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Dr. Emánuel Antal is 70

Dr. Emánuel Antal recently celebrated his 70th birthday. We wish to celebrate his anniversary with a volume containing articles written by his friends, colleagues. I feel this way of salutation is extremely appropriate in his case, because he was in all his life, and even now continues to be, a man who concentrates his work at building a bridge between meteorology and other sciences. He strongly believes that common, interdisciplinary efforts with well-defined practical aims will give meteorology greater recognition by the state and society. His outstanding scientific and official career is a living example of the successful realization of his ideas. He achieved great scientific and practical results not only in meteorology and climatology, but also in agricultural science, soil science, and hydrology working together with excellent representatives of these sciences. One cannot forget to mention, that his successes were highly recognized by the state through the State Prize and the Schenzl Guidó Prize, and he was a highest-ranking leader of the Hungarian Meteorological Service for a decade.

In the name of the Editorial Board, authors of this issue, and – feeling obliged to do so – the scientific communities of partner disciplines, I would like to send Dr. Emánuel Antal our greetings and wish him continuing successes in his scientific work and personal life.

Dr. Emánuel Antal's scientific career

Dr. Emánuel Antal was born in Jászárokszállás, in 1931. He finished his studies in meteorology at the Eötvös Loránd University of Budapest in 1955, and received his diploma with distinction. Immediately after that, Dr. Antal started to work at the Hungarian Meteorological Institute (which is now the Hungarian Meteorological Service). His scientific interest turned to the problems of micro-meteorology (radiation-, heat-, and water-fluxes in the near-to-the-surface layer of the atmosphere, heat and moisture balance of the surface and the vegetation, plants, etc.) and agrometeorology. He defended his university doctor's thesis very early, in 1960. Just after that, at the age of 31, he was promoted to serve as the Head of Department of Agrometeorological Research. At that time he started his scientific organizational work, which resulted in the re-thinking and re-establishment of Hungarian agrometeorology in the whole or, maybe, the establishment of the applied agrometeorology, as a new interdisciplinary field between agriculture and meteorology. It is impossible to give the whole scope of his extremely significant scientific career in the frame of such a short summary. As examples of the most important milestones of his career, Dr. Antal:

- *successfully defended his CSc thesis in 1968 in the field of agrometeorology, which contained outstanding scientific results, among them the so-called "Antal-formula" for estimation of potential and actual evapotranspiration. In this way, he became one of the rare scientists, who fixed their names in the science of meteorology;*
- *founded the Szarvas Agrometeorological Station in 1963 and the Observatory in 1974, with the aim to investigate the radiation, heat, and moisture exchange in the earth-plant-atmosphere system, and the demand and consumptive use of water by plants;*
- *initiated field research work from 1968 in the Lake Fertő area to explore the heat and water balance of this area rich in natural beauties but, also, extremely vulnerable region;*
- *organized and led, from 1977, the important national agricultural program with the aim to establish a complex decision-making system of growing vegetables the optimum way for the canning industry. For this work he was honoured by the National Prize;*
- *organized and led, from 1987, a common American-Hungarian project to investigate extreme climatic events, such as droughts, etc., which have great and adverse impact on national economy;*
- *served as the meteorological advisor of the Minister of Environment and Regional Policy in the years 1992–1994;*
- *prepared (with a co-author) the booklet "Role of Climate and Climate Change in the Life of Hungary" in 1996, which represented the Hungarian point of view at the Second World Conference on Climate;*
- *served as the elected member of the General Assembly of the Hungarian Academy of Sciences between 1995 and 2001.*

Meanwhile – in his official career – he was promoted to serve as Assistant Director of the Central Institute for Atmospheric Physics (1978), then Vice-President of the Hungarian Meteorological Service (1981), finally President in Charge (1990). He retired from this position in 1991. As the above list of his activities shows, this retirement was only formal, he acts in the same wide range and with the same energy to the present day. At this time he is the highly esteemed chairman or member of several scientific societies in meteorology, agriculture, and hydrology, intensively participates in university education at the Szent István University (Hungary), and follows his extremely fruitful activity in writing and editing scientific books and other publications. He has written more than 100 scientific publications.

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Water supply of growing seasons and maize production

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Abstract—Water is one of the most important factors of plant breeding. The rain water falling from the clouds is stored in the soil. This water is available for the plants. The most significant factor that can reduce the water content of the soil is the evaporation. The water income and the water loss determine the water balance of the soil which shows the quantity of water available for plants.

In agroclimatology, therefore, there is a need to find out a climatic characteristic for representing the water balance of the soil. Such characteristics are aridity indices (ARI) and humidity indices (HI), which can be calculated in a simple way, and they have a clear physical meaning.

These agroclimatic indices may be used for describing the wet or dry character of the years and the vegetation periods. Finally, rather close relationships can be found between the aridity indices and the yield of maize.

The investigations were carried out on the basis of meteorological and yield data collected parallel during the period of 1951–1995.

Key-words: agroclimatology, evaporation, water balance, aridity index, yield, trend function, trend ratio, regression analysis.

1. Introduction

Water is one of the major growing factors for plants. It is necessary for photosynthetic process because plants demand water besides carbon dioxide to produce organic matter. It is also an important element of transport processes since nutrients come to the assimilating organs dissolved in water.

Plants take up water from the soil where it fills the pores together with air. The ratio of the water and air in the pores exert a strong influence on the life of plants. If there is abundance of water, plants suffer the shortage of air since the oxygen is necessary for respiration. If there is a water shortage in the pores, plants cannot photosynthesize in a normal way because the transpiration process is not able to transport required quantity of water to the organs of assimilation. It is a general demand for plants, therefore, to have certain amount of water in the soil for photosynthesis and a sufficient quantity of air in pores for respiration.

To enable plants to take up water from soil, there is a need for sufficient water in soil pores, a temperature above threshold value that makes possible to take up water by the cells of roots, and continuous transpiration process which is able to transport the water from the soil to plant leaves where the assimilation takes place. Finally, the water not used up by plants goes away to the atmosphere through the stomata.

These phenomena and processes are strongly affected by meteorological factors.

2. Meteorological conditions and the water supply

The water necessary for plants is stored in the soil. The amount of water between field capacity and wilting point is available for plants, and this quantity mainly depends on the precipitation and evaporation. The former is the main source of water income and the latter is the most significant factor of water loss. These are the two meteorological factors by which the atmosphere can control the water amount available for plants.

Other important factors influencing the water content of the soil and the water supply of plants are illustrated in *Fig. 1*. The amount of water in soil is primarily effected by macrometeorological conditions. It is possible that humid air masses arrive at a given place where clouds are forming and significant rain water income occurs. In another case high atmospheric pressure with descending air movements hinders formation of clouds resulting in a clear weather and strong insolation, increasing the water quantity evaporating from the soil.

When the amount of rain water is abundant, the water accumulates in the upper layers of the soil. If the amount of rain water is continuously increasing, the soil fills up slowly with water and finally the water appears on the soil surface. In this case the plants suffer the air shortage.

The first step towards the formation of water shortage is the lack of rain. If this phenomenon is long-lasting accompanied by low air humidity and high temperature, the evaporation from the surface increases gradually.

The soil cultivation can also exert an influence on the water content of the soil.

Finally, the water in the soil has different significance for various plants since their water demand strongly differs (*Varga-Haszonits,* 1983).

In agroclimatology, therefore, there is a need to find out a climatic characteristic for representing the water balance of soil.

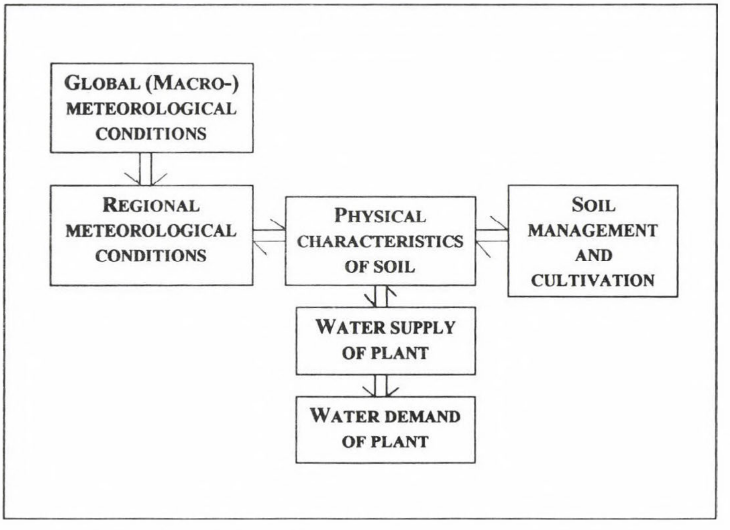

Fig. 1. Factors influencing water demand of plant.

2.1 Agroclimatic index of relative water balance

For characterizing the water balance of a time period at a place, it is practical to use rainfall amount (representing the water income), and the amount of evaporation (representing the water loss) for forming a ratio of two values, which expresses the relative water balance. Using these data we can calculate the relative water balance in the following way (*Varga-Haszonits et al.,* 1996):

$$HI = \frac{P}{E_0} \tag{1}$$

or

$$ARI = \frac{E_0}{P}, \tag{2}$$

where HI is the humidity index, ARI is the aridity index, both are dimensionless, P is the rainfall amount in mm, and E_0 is the evaporative power of air in mm.

The evaporative power of air (E_0) is the amount of water (in mm) evaporated from free water surface, measured by pan „A". This is a generally used equipment at the meteorological stations. At places, where no evaporation measurements have been made by pan „A", the evaporative power of air can be determined by using different local formulas developed for the given area (*Dunay et al.*, 1968). Essentially, the evaporative power of air is equal to the potential evaporation.

In Hungary water shortages occur more frequently than water surpluses, therefore, we use the aridity indices for analyzing the humidity conditions of growing seasons. The aridity indices show that how much more water could be evaporated by air than the actual amount of the rainfall in the same period.

2.2 Dry and wet periods of year

Examining the year from the point of view of water balance, we can divide it into two parts. One of these parts is the dry period when the water loss exceeds the water income. The other period is the wet one when the amount of rainfall during the period is higher than the water loss in the same interval. Consequently, the threshold value for separating dry and wet periods is 1, since this shows an equilibrium between water income and water loss. By using this climatic separator value (ARI=1), we can differentiate the periods of different humidity conditions (*Varga-Haszonits et al.,* 1997).

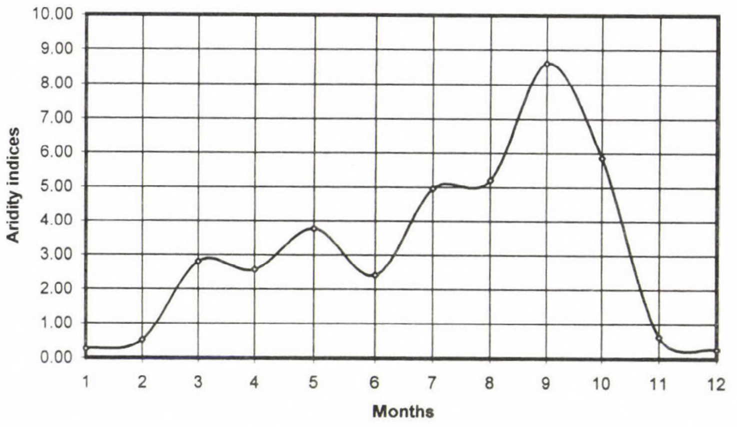

Fig. 2. Annual course of aridity index in Hungary (1951–1955).

The line of annual course of monthly aridity indices intersects the line of aridity index value 1 in February and November (*Fig. 2*). Between February and November the values of ARI are higher than 1, so this period is designated as a dry one. While the period from November to February, when ARI falls short of 1, is called wet period.

Looking at Fig. 2 we can see that an increasing dryness becomes dominant in spring finding a moderate decline in June, then the values of ARI begin to grow again to the peak of the curve in September. After the peak the values of aridity indices (ARI) decreases deeply as far as November in which time our country has a secondary maximum of precipitation. As we mentioned earlier, the wet period begins in November and lasts to the end of February.

2.3 Humidity conditions of growing season

Knowledge of meteorological conditions during the growing season is of basic importance for plant breeding (*Varga-Haszonits,* 1998). *Table 1* contains the statistical characteristics and frequency values of ARI during the vegetation periods of years under investigation. The wettest part of the country is the southwestern area (Vas, Zala, Somogy counties) of Transdanubia, where the mean values of ARI fall short of 2. The driest area can be found in central part of the country, in the counties of Pest, Bács-Kiskun, Jász-Nagykun-Szolnok and Csongrád. In this area the average values of ARI vary between 2.75 and 3.10.

Table 1. Distribution of aridity indices in Hungarian counties during the growing season of maize 1951–1995

County	Min.	Aver.	Max.	>1.00	1.01-1.50	1.51-2.00	2.01-2.50	2.51-3.00	3.01-3.50	3.51-4.00	4.01-4.50	>4.00
Győr-Moson-Sopron	1.35	2.43	4.16	3	12	11	9	7	2	1	0	0
Vas	0.74	1.68	3.62	20	18	3	3	0	1	0	0	0
Zala	0.80	1.53	3.34	27	12	3	1	2	0	0	0	0
Somogy	0.92	1.73	3.24	19	14	7	4	1	0	0	0	0
Veszprém	0.93	2.00	3.98	13	12	12	3	2	3	0	0	0
Komárom-Esztergom	0.99	2.21	3.94	6	16	8	8	4	3	0	0	0
Fejér	1.25	2.67	5.68	1	9	14	8	7	2	1	3	3
Tolna	0.99	2.09	4.02	9	14	12	5	2	2	1	0	0
Baranya	1.04	2.28	3.82	7	12	9	9	6	2	0	0	0
Bács-Kiskun	1.32	2.95	6.21	2	8	7	10	8	3	2	5	5
Pest	1.45	3.07	6.42	1	5	8	13	10	0	2	6	6
Jász-Nagykun-Szolnok	1.22	2.78	6.29	2	8	10	10	9	1	2	3	3
Csongrád	1.30	2.86	5.65	3	7	11	8	5	2	4	5	5
Békés	1.35	2.51	4.57	4	7	18	6	5	1	3	1	1
Hajdú-Bihar	0.85	2.44	5.19	8	5	14	8	6	0	1	3	3
Szabolcs-Szatmár-Bereg	1.05	2.45	4.81	5	10	13	5	6	3	1	2	2
Borsod-Abaúj-Zemplén	0.99	2.10	4.66	11	15	8	4	3	2	1	1	1
Heves	1.38	2.70	6.17	4	10	7	10	6	4	1	3	3
Nógrád	0.94	2.21	4.21	6	17	9	5	4	3	1	0	0

The ARI values having a minimum less than 1 are characteristic mainly for the western and northern counties. The growing seasons with ARI under the value 1 can be considered as the wettest ones in our country. This type of vegetation periods, however, does not occur in the middle of the country.

The maximum values of ARI are over 3 in the southwest area, while they vary between 5.5 and 6.5 in the middle of the country.

The frequency values also follow this spatial distribution. Values less than 1.5 are most frequent in the southwestern part of the country. Values between 2.01–2.50 are characteristic for the middle part of the country. Very dry growing seasons occur also in this area.

The humidity conditions of vegetation periods are especially important for the plants sensitive to water supply. It is well known that humidity is one of the most variable meteorological factor which has a significant influence on the yield formation and plays an important role in fluctuations of actual yield from year to year.

The fluctuations of the ARI in successive years are illustrated in *Fig. 3*. The values of ARI during the growing seasons of maize (the ratio of evaporative power of air to rainfall in the time spell from April to September) vary generally between 1.50 and 3.00, that is, the air would be able to evaporate 1.5–3 times more water amount than the actual precipitation in this period.

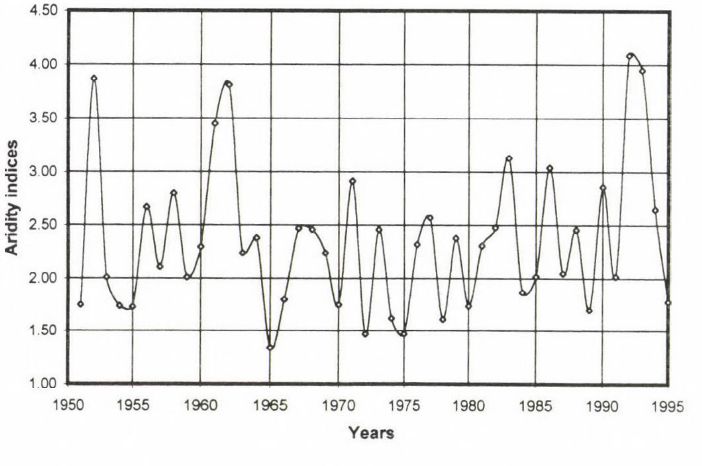

Fig. 3. Temporal fluctuations of aridity indices in Hungary (1951–1995).

There are some extremely dry growing season values some of which exceed the value of 3 and in a single case (in 1992) even 4. In that year the air would have been able to evaporate 4 times more water than the actual amount of rain water during the growing season. There were 7 years between 1951 and 1995 when the ARI values of vegetation periods were higher than 3. The ARI values went below 1.5 only in 3 years. The minimum value was 1.34 (in 1965).

3. *Agroclimatic analysis of humidity-yield connection*

We used meteorological data collected by the Hungarian Meteorological Service and yield data gathered by the Hungarian Statistical Office as average values for the counties of Hungary. All data belong to a 45 years long period from 1951 to 1995.

Since yield data are referred to a county, we have selected one representative meteorological station in each county for calculating the connections between meteorological and yield data. Knowing the close relationships among neighboring stations, the representativity is self-evident for temperature. The values of precipitation, however, are more variable but showers take place mainly in summer months, when the values of aridity indices are generally high, so the possible errors have no effect on the computation of the relationships. Considering the earlier mentioned facts, the following meteorological stations were selected: Győr, Szombathely, Zalaegerszeg, Kaposvár, Pápa, Tatabánya, Martonvásár, Iregszemcse, Pécs, Kecskemét, Budapest, Szolnok, Szeged, Békéscsaba, Debrecen, Nyíregyháza, Miskolc, Kompolt, and Balassagyarmat.

3.1 Analysis of yield

Fig. 4 shows that from early 1960s the yield gradually increased thanks to the introduction of new intensive varieties, as well as more up-to-date fertilizers, and methods of plant protection. It was so till the end of 1980s, when the technological level dropped due to the considerable lack of the rather expensive fertilizers, and the plant protection methods. These variations in technological level can be described relatively well by trend functions. The set of points in Fig. 4 can be determined by third-degree or fourth-degree polinoms. The various trend functions give an approximation with different accuracy.

As we can see in Fig. 4 the actual yield values represented by points are located around the trend curve. It is assumed that the yield is made up of two components, one of the result of technological factors and the other of meteorological factors, that is

$$Y(t) = Y_T(t) \cdot Y_M(t), \tag{3}$$

where $Y(t)$ is the actual yield, $Y_T(t)$ the part of yield caused by technological factors, $Y_M(t)$ is the part caused by meteorological factors, and the t is the time (in this case it is a given year).

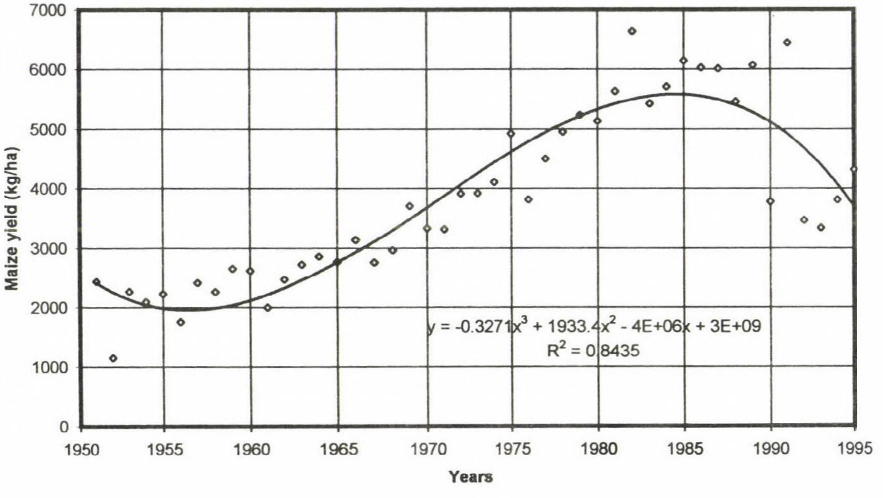

Fig. 4. Average maize yield in Hungary (1951–1995).

From Eq. (3) we can determine the meteorological effect as

$$\frac{Y(t)}{Y_T(t)} = Y_M(t)\,. \tag{4}$$

The actual yield is measured, we can calculate the trend values from actual values, so the left side of Eq. (4) is known and the $Y_M(t)$ can be determined. Using this value, the $Y_M(t)$ expresses the complex effect of all meteorological elements.

It is practical to give the annual fluctuations as deviations from the trend values. The less the annual fluctuations, the more stable the yields are. First, the deviations from trend should be investigated. As Fig. 4 indicates, the fluctuation increases in the case of higher yields, that is the yield stability becomes smaller. If the deviations will be greater with higher yield, the meteorological elements increase or decrease proportionally the part of yield determined by technological factors. That is the reason why Eq. (3) was described in multiplicative form, and the meteorological effect was calculated as a ratio of actual yield to trend value (Eq. 4).

3.2 *Complex meteorological effect on yield*

Under constant climatic conditions, similar meteorological effects are expected to produce similar yield fluctuations. If it is true, these are considered as additive effects, since the meteorological effects increase or reduce the technological effects to a certain degree, so the meteorological effects are defined by the deviations. As Fig. 4 shows, the yield fluctuations increased in spite of no change in climate. In this case, the meteorological effects can be described by the ratio of the actual yield to the trend value (trend ratio).

Thus we have a possibility to separate technological effects (the effects of hybrids, fertilizers, and plant protection), and meteorological effects influencing the yield.

Trend ratio describes the effects of all meteorological elements, so it is also called index of complex meteorological effects (ICME). In this form it also includes the random effects which only slightly modify the results. We have calculated the trend ratios for each county. The yearly fluctuations of the average trend ratio over the whole country are shown in *Fig. 5*.

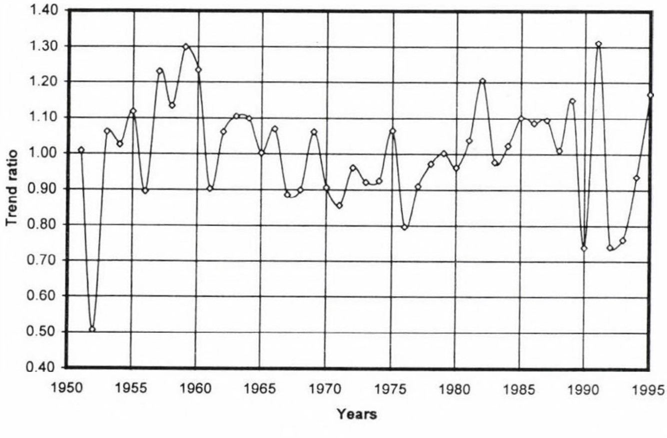

Fig. 5. Yearly fluctuations of the average trend ratio.

Values higher than 1 show that the effect of the weather was favorable for yield in that year. The values equal to 1 represent yields which are in accordance with the technological level, that is the weather neither increase nor reduce the yield. In the event of values over 1, the trend ratio numerically shows to what extent the weather increases the yield, in the case of values less than 1, the trend ratio indicates the extent of yield reduction.

Fig. 5 shows that the highest yield increases induced by the weather vary between 20% and 40%, which indicates a considerable fluctuation. In the period 1951–1995 there are 5 years (1957, 1959, 1960, 1982, 1991) when favorable weather conditions dominated in every county. There were years when positive yield deviations were observed in most counties, and only few counties showed no favorable weather effect.

The most unfavorable meteorological effects vary between 20% and 50%. So they seem to be more intensive. Five years were found (1952, 1976, 1990, 1992, 1993) when the weather conditions were unfavorable in every county, that is the unfavorable effects occurred more often. Most of the years were dominated by unfavorable effects in the greater part of counties, although parallel there could be favorable effects in one or two counties.

The frequency results show that trend ratios between 0.8 and 1.2 add up to 76–84% of all events, i.e., the deviations higher than ±20% account for 16–24% of all events. So approximately in every fifth year the weather may modify the yield by 20% either positively or negatively compared to the technological level.

3.3 Effect of different meteorological elements

The meteorological effects based on various elements can be determined in a particular year (*t*) in the following way:

$$\frac{Y(t)}{Y_T(t)} = Y_M(t) = f(m_1, m_2, \ldots, m_K). \tag{5}$$

Since the values of meteorological elements are measured continuously at numerous places of the country, the only thing we need to know is: which element for which period is to be taken into consideration.

It is well known that green plants as maize require solar radiation for photosynthesis. The water is also an essential element to photosynthesis besides the nutrients dissolved in it and moving from the soil to the organs of assimilation. The speed of biochemical reactions within the plants depends on the temperature. So there are three meteorological elements: photosynthetically active radiation, temperature, and water which have basic importance for plants (*Pethő,* 1993).

It is practical to choose the precipitation to determine the water amount which is necessary to form unit quantity of yield, and aridity indices for determining the connection between actual yield and humidity factor.

3.4 The time period

We have to make a decision about the time period in which the meteorological elements will be determined. The following obvious time periods can be selected:

- the whole vegetation period,
- phenophases,
- calendar time spell, and
- optional duration of time spell.

In this paper the whole vegetation period was chosen. This is a natural period lasting from planting to ripening, and, therefore, it may be a base of a primary investigation.

3.5 Temporal variations of water used by yield formation

Precipitation amount of growing seasons was used to determine the quantity of water utilized for yield formation. The water use (*WU*) can be calculated by the following equation (*Varga-Haszonits*, 1981):

$$WU = \frac{P}{Y_{BIO}}, \tag{6}$$

where P is the precipitation during the vegetation period (in mm), and Y_{BIO} is the biomass. Generally, we know only the economic yield, therefore, the biomass must be calculated. It is known that the economic yield (Y_{ECO}) is a part of total biomass which may be computed as follows:

$$Y_{ECO} = k_{ECO} \cdot Y_{BIO}, \tag{7}$$

where k_{ECO} is a coefficient of economic yield (harvest index). From Eq. (7) we can receive that

$$Y_{BIO} = \frac{1}{k_{ECO}} \cdot Y_{ECO}. \tag{8}$$

So we can calculate the value of biomass by using the value of economic yield. The accuracy of this calculation depends primarily on the value of k_{ECO}.

The results can be seen in *Fig. 6* where the values were averaged for the whole area of the country. Fig. 6 shows that the water use of maize yield formation gradually decreased by using new intensive varieties of plants, appropriate doses of fertilizers, and adequate methods of plant protection.

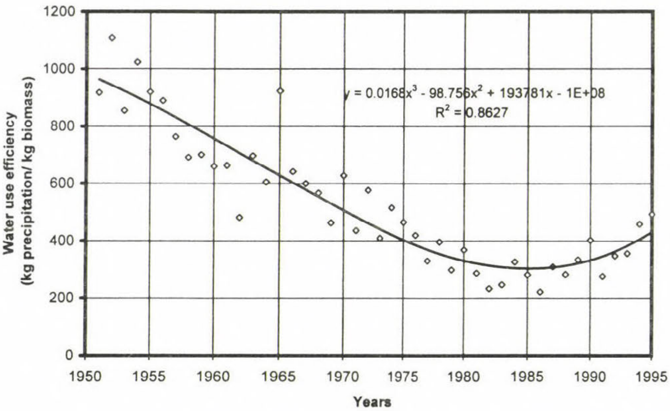

Fig. 6. Temporal changes of water use efficiency of maize in Hungary.

At the end of the 1980s, the application of fertilizers decreased as a consequence of financial problems, so water amount needed for the formation of unit mass of maize yield begun to increase again.

For the period 1951–1995 the water use of maize yield can be described by the aim of a third-degree equation.

3.6 Relationship between aridity index and maize yield

We investigated the connection between aridity index and yield in every county by regression analysis (*Varga-Haszonits* and *Harnos,* 1988). The connection might be described by second-degree equations. The correlation indices are indicated in *Table 2*. As we can see they vary between 0.40 and 0.65. All the values are significant on the 0.1% level of probability.

These values show not only close relationships but also an areal distribution. It is interesting to note that the correlation indices are the highest in the wettest area of the country (Vas, Zala, Somogy, Veszprém counties). It shows that maize yield has a higher sensitivity to the fluctuations of relative water balance in wet areas than in dry ones. In the eastern part of the country (in the eastern border of Hungarian Great Plain) which is a dry area, the correlation indices are the lowest showing less sensitivity of maize yield to the relative water balance.

Table 2. Correlation values (*r* values) of the relationship between aridity index of the growing season and maize yield

County	r values
Győr-Moson-Sopron	0.5591
Vas	0.6306
Zala	0.5771
Somogy	0.6158
Veszprém	0.6158
Komárom-Esztergom	0.6158
Fejér	0.6158
Tolna	0.6158
Baranya	0.6158
Bács-Kiskun	0.5442
Pest	0.5516
Jász-Nagykun-Szolnok	0.4958
Csongrád	0.5002
Békés	0.4626
Hajdú-Bihar	0.4204
Szabolcs-Szatmár-Bereg	0.4354
Borsod-Abaúj-Zemplén	0.5416
Heves	0.5335
Nógrád	0.4798

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Atmospheric wet deposition as a nutrient supply for the vegetation

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Abstract—The chemical composition of precipitation and wet deposition rates of different water soluble substances and metals are presented on the basis of precipitation chemistry programs carried out in Hungary. Data gained by bulk and wet-only collectors are compared and changes in precipitation composition during the last two decades are discussed. Decrease of acidity due to air pollution management and economical changes is emphasized. Hungarian atmospheric metal depositions are compared with other information obtained abroad. It is concluded that atmospheric wet deposition is an important nutrient source of plants in regions where no fertilizers are used.

Key-words: atmospheric precipitation chemistry, wet deposition, soluble matter, heavy metals, Hungary

1. Introduction: remarks on the water cycle

It is a well-known fact that water is an essential compound for life. Thus, the plants on the continents use water molecules, besides carbon dioxide and solar energy, to produce complex carbon species necessary for their metabolism. During this process called photosynthesis, a part of the oxygen is liberated which makes the respiration of animals and men possible. Consequently, the cycle of water in nature is a very important process for ecosystems in continental environment (*Antal* and *Szesztay*, 1996). The main characteristics of atmospheric water cycle is the fact that the budget in oceanic air is negative. In the atmosphere the balance is transported over the continents, where, due to this excess, the quantity of precipitation is higher than that of evaporation (actually mainly evapotranspiration). The total yearly continental precipitation amount is equal to about 100×10^3 km^3, while the evaporation is only 70% of this water volume (*Manning*, 1997). This implies that about 30% of the continental precipitation amount turns back in rivers to the oceans each year. Taking into account the oceanic precipitation amount ($\cong400\times10^3$ km^3/year), as well as the atmospheric water burden (13×10^3 km^3), a simple calculation shows that the turnover time of water in the atmosphere is equal to 0.026 year, that is about 9 days. This indicates that the quantity of atmospheric water is changed forty times during a year.

The atmospheric turnover of water has a considerable impact on the atmosphere itself. This is due to the fact that aerosol particles imbedded in cloud and precipitation elements and molecules of gases soluble in water are removed form the air during precipitation fall. This so-called *wet deposition* process constitutes an important atmospheric self-cleansing mechanism. Without wet (and dry) deposition the atmosphere would become very dirty in a short time interval owing to the accumulation of trace substances of biogenic and anthropogenic origin. However, due to wet deposition, water soluble trace gases as well as aerosol particles have residence times around 10 days or less in agreement with the turnover time of water in the air. Obviously, species removed from the atmosphere by precipitation are deposited onto the surface, that is the water cycle supplies not only water for continental vegetation, but also a lot of other materials which can be used as nutrients by aquatic and terrestrial ecosystems.

During the last million years of the Earth's history, atmospheric deposition removed exactly such quantity of different trace substances as the amount liberated by natural sources and the vegetation accommodated to the magnitude of this atmospheric input. However, during the last millennia, but mostly during the last century human activities have disturbed this steady state considerably. This is caused by the emission of different pollutants which are removed from the air similarly to biogenic compounds. Consequently, an important concentration increase of different species in precipitation water can be observed. This change alters significantly the atmospheric input into the biosphere which can be dangerous under special circumstances. Thus, the comparison of the nitrate concentration in precipitation water measured in Hungary at the beginning and in the second half of the 20th century shows an increase of seven times, which is obviously due to human activities like transport and energy production (*Horváth*, 1983).

The aim of this paper is to summarize the results of precipitation chemistry measurements carried out in Hungary during the last decades and to discuss the atmospheric input of different ions and metals into soils and vegetation.

2. Concentration and deposition of different ions

The first Hungarian measuring *network* for precipitation chemistry observations was operated between 1968 and 1970. In this early program precipitation water was collected at eight stations in the country by using *bulk collectors*. These collectors were open continuously, which means that the sedimentation of dust particles was not excluded during the sampling. The results of this program were published by *Kozák* and *Mészáros* (1971) who evaluated the data concerning their agricultural importance. This evaluation indicated that atmospheric deposition of nitrogen (actually ammonium and nitrate) is not an important source for the nitrogen uptake of agricultural plants caused by the magnitude of the nitrogen content of fertilizers. However, in areas where no fertilizers are used (e.g., in forests) the atmospheric input is very important. On the other hand, the wet deposition of sulfur contributes significantly to the sulfur uptake of plants even over agricultural regions. Finally, the potassium deposition is also much smaller than the doses given to the soils as fertilizers. One can speculate, however, that potassium ions reaching the leaves of the plant directly in precipitation water can possibly be taken up by the plants.

The results of the bulk network also demonstrated that the concentration of several ions increased from west to east. An exception was the amount of hydrogen ions which was lower in the eastern part of the country. This finding was interpreted as the effect of dustfall which is obviously more significant over the Great Plain as compared to more hilly western areas covered by denser vegetation. To exclude this problem, in 1977 a new network was organized which consisted of six stations. At these sampling sites precipitation was collected by *wet-only gauges* which opened at the beginning of precipitation fall and closed automatically at the end of precipitation events. The results obtained by this network were presented and discussed by *Horváth* and *Mészáros* (1984) who also compared them with those gained by the earlier bulk network. Among other things, it followed from this comparison that data gained by wet-only collectors did not show the spatial distribution emerged from concentrations obtained by the bulk network, which was previously mentioned.

Data given in *Table 1* are taken from the latter paper referenced. The table shows the average concentrations of different ions analyzed. Wet deposition rates calculated by multiplying mean concentrations by yearly precipitation amounts are also tabulated. In the rows for concentrations the values of specific conductivity, characterizing the total amount of ions, are also given. Further, in the case of hydrogen ions, pH data, defined as the negative logarithm of the concentration, can also be seen. By comparing bulk and wet-only concentrations it is evident that in the bulk network higher concentrations were measured than by the wet-only collectors. The difference are high in particular for sodium, potassium, and calcium which can be considered as elements typically of surface origin. The ratios of bulk concentration to wet-only concentration for sodium, potassium, and calcium are 2.8, 3.3, and 2.6, respectively. Accordingly, the specific conductivity is two times larger in the case of bulk measurements. It is interesting to note that this ratio is relatively high also for sulfate ions. This involves the possibility that coarse aerosol particles of soil origin contains a large amount of sulfate. Taking into account the high calcium concentration one can speculate that the sulfate is associated with calcium as calcium sulfate which is slightly soluble in water.

Table 1. Average concentrations (C in mg L^{-1}) and wet deposition rates (D in mg $m^{-2}a^{-1}$) according to bulk and wet-only networks. Note that the values of specific conductivity (κ) are expressed in $\mu S\ cm^{-1}$ (1 siemens is equal to 1/ohm), while the dimension of the yearly precipitation amount (P) is mm (=L m^{-2}).

Network		κ	pH	H	NH_4-N	Na	K	Mg	Ca	Cl	NO_3-N	SO_4-S	P
Bulk	C	62	4.7	0.02	1.4	1.5	1.1	-	4.4	1.4	0.83	3.7	638
	D	-	-	13	893	957	702	-	2807	893	530	2360	-
Wet-only	C	31	4.5	0.03	1.1	0.54	0.34	0.42	1.7	1.0	0.58	1.9	573
	D	-	-	18	630	309	194	241	974	573	332	1088	-

Considering the deposition rates, it is evident that the dominant elements are calcium and sulfur. The wet deposition rate of sulfur (wet-only network) is about 1 gm^{-2} yr^{-1} which results in a total Hungarian deposition of 93 Gg yr^{-1} (1 Gg=10^9g). It is self-evident that this important amount of sulfur deposition is due to sulfur dioxide emission in Hungary and abroad. According to *Horváth* and *Mészáros* (1984), the dry deposition of sulfur dioxide was similar to the wet deposition of sulfate (both expressed in sulfur) value at least in the years of the study. It is well documented that the main species responsible for acidic deposition is sulfuric acid. While the concentration of hydrogen ions depends directly on that of sulfate ions (it is speculated that sulfate deposition determines the soil pH), species of soil origin like calcium carbonate tends to decrease the acidity of atmospheric waters. However, we have to take into account that a part of calcium comes from anthropogenic sources (e.g., as fly ash from power plants). This statement is based on our aerosol measurements (*Molnár et al.*, 1993), according to which the enrichment factor of calcium is between 5–10 relative to the average soil composition. Anyway, we can conclude that precipitation in Hungary was rather acidic about twenty years ago: the hydrogen ion concentration was more than one magnitude higher than the equilibrium value controlled by the absorption of atmospheric carbon dioxide (pH≈5.7). A further peculiarity of the precipitation composition based on wet-only network is the fact that the sum of cations expressed in equivalents (238 μeq L^{-1} is higher than the sum of anions (189 μeq L^{-1}). This is not surprising, however, since hydrogen carbonate ions were not analyzed in the program. According to a parallel set of analyses, the concentration of these latter ions was found to be 67 μeq L^{-1} on an average. Taking into account this value, we can conclude that the sum of anions is equal to the sum of cations which is an absolute necessity of electric neutrality. If we calculate in a first approximation the hydrogen ion concentration as the sum of the amount of sulfate, nitrate, and hydrocarbon ions diminished by the sum of ammonium, calcium, and sodium concentrations (all expressed in equivalents), a pH value of 4.4 is obtained which is similar to the value directly measured (see Table 1).

It is well documented that during the years between 1980 and 1995 sulfur-dioxide emission decreased considerably in Europe (*Mylona*, 1996) and in Hungary (*Bozó*, 1998). Generally speaking, air pollution caused by energy production was mitigated significantly. For this reason it seems interesting to estimate changes in precipitation composition during the same time period. In *Table 2* the average chemical compositions of precipitation measured between 1977–1980 and 1997–1998 in Hungary are compared (the latter data are kindly provided by the Hungarian Meteorological Service).

Table 2. Chemical composition of precipitation in Hungary collected by wet-only samplers in two time periods. Data are expressed in mg/liter, except those for electric conductivity (κ: in μS cm^{-1}) and pH. The table also contains the average yearly precipitation amount (P: mm a^{-1}) observed at the sampling sites.

Period	κ	pH	NH_4-N	Na	Ca	NO_3-N	SO_4-S	P (mm/a)
1977–1980	31	4.5	1.1	0.54	1.7	0.58	1.9	573
1997–1998	20	6.0	0.61	0.63	0.76	0.38	1.0	602

First of all we have to note that the average precipitation amounts were similar in the two periods which makes the comparison reliable (the composition also depends on the amount of precipitation). One can see from data tabulated that an important change occurred between the two time intervals. Thus, the specific conductivity decreased by a factor of 1.5, while precipitation became slightly alkaline, practically neutral (if the effects of carbon dioxide is considered). This is obviously due to the decrease of nitrate, but mostly of sulfate concentration. At the same time calcium concentration also decreased (decrease of fly ash emission?), while sodium content of precipitation remained the same. This implies that sodium is of soil origin, consequently its concentration is independent of air pollution management. An interesting feature of ammonium is that the concentration of this ion, independent of energy production and industrial activity, also decreased. A similar situation was observed in Eastern Germany in the territory of former German Democratic Republic by *Möller et al.* (1996) who published an ammonium ion concentration decrease of 39% between data gained before 1990 and after the political changes. These authors attributed this change to the decreasing ammonia emission owing to changes in the structure of agriculture after the unification of Germany. In Hungary before and after 1990 the ammonia emission was reduced by a factor of two as discussed in more detail by *Horváth* and *Sutton* (1998). However, these latter authors stated that this reduction was not reflected in atmospheric ammonia/ammonium concentration. This means that further research is needed in this fields since ammonia (ammonium ions in precipitation) plays an important role in the control of nitrogen deposition and of acidity of atmospheric waters.

As it was previously mentioned, the significance of atmospheric deposition is important in particular for forest ecosystems where no fertilizers are used. For this reason the rate of dry and wet depositions was studied in detail at three forest sites in Hungary (Farkasfa: western Hungary in a hilly area; K-puszta: central part of the country at the Great Hungarian Plan; Nyírjes: north-east area in the Mátra-mountains) by *Horváth et al.* (1993). Data collected between 1988 and 1992 show that the concentration of sulfur dioxide and nitrogen dioxide as well as acidic deposition rate decreased during this time interval in agreement with our previous discussion. At the same time the importance of nitrogen (actually nitrogen dioxide and nitrate) deposition relative to sulfate-sulfur increased. This is obviously related to the fact that the decrease of the release of nitrogen oxide into the atmosphere was not as drastic as in the case of sulfur dioxide (see *Mészáros*, 2001).

3. Deposition of different metals

Beside essential components like water, nitrogen, and sulfur, the so-called microelements (e.g., metals) play an important role in the control of the life of terrestrial and aquatic ecosystems. The question of microelements has become an important environmental problem since human activities liberate a lot of metals into different domains of the environment including the atmosphere. Briefly, practically each element is needed for the plants. It should be emphasized, however, that the amount is a very important factor. If a given element is not available for a given plant, this causes deficiency disease. On the other hand, and this is the general case concerning pollution, too high quantities lead to poisoning the vegetation. This explains the great number of projects in the last decades aiming to estimate the atmospheric deposition of different metals onto the surface of continents and oceans.

The Hungarian program, the results of which is published in the literature (in Hungarian: *Mészáros et al.*, 1993; in English: *Molnár et al.*, 1995), was carried out in 1992. In this program precipitation water was collected by wet-only gauges in the country at three sampling sites including K-puszta mentioned above. Special care was made to avoid chemical contamination and transformations during sample storage before chemical analyses. Parallel to wet deposition observations, the concentration and size distribution of different elements in the atmospheric aerosol were also measured for estimating dry deposition rates. It was found, however, that the magnitude of dry deposition was much smaller than wet deposition rates, so they are not included into the following discussion. *Table 3* gives the concentration in precipitation of some environmentally important metals determined on the basis of K-puszta samples. In the tables the wet deposition rates can also be seen as calculated by multiplying the concentrations by the yearly amount of precipitation. For comparison some data from other parts of the world (Sweden, Ontario) are also included (see *Mészáros*, 1999).

Table 3. Concentrations (C in $\mu g\ L^{-1}$) and wet deposition rates (D in $mg\ m^{-2}a^{-1}$) of different metals at K-puszta station (Hungary) in 1992, where P gives the yearly precipitation amount expressed in mm. In the table appropriate data for Sweden and Ontario (Canada) are also given.

Parameter	Cd	Cu	Ni	Pb	V	P
C (Hungary)	0.94	3.6	1.8	13	2.4	477
C (Sweden)	0.14	1.4	-	8.8	-	-
C (Ontario)	<0.20	1.3	-	2.4	0.4	-
D	0.45	1.7	0.85	6.2	1.1	-

One has to consider that the elements listed in Table 3 are practically totally of anthropogenic origin in populated industrial areas like Europe and North America. Briefly, cadmium and copper emissions are due to non-ferrous metal production, although a part of cadmium is liberated into the atmosphere during fossil fuel use and waste incineration. The main nickel source is oil combustion, in a lesser way coal burning. Lead is a notorious product of the use of leaded gasoline, while atmospheric vanadium is created during oil combustion.

It follows from Tables 1–3 that metal concentrations in precipitation are one or two orders of magnitude smaller than ions concentrations. However, one should keep in mind that depositions calculated from concentration data give essentially the anthropogenic contribution to the atmospheric input into natural and agricultural ecosystems. The deposition rate of copper and lead is high in particular. We have to note, however, that lead emission has certainly decreased considerably during the last decade owing to the introduction of unleaded gasoline (see *Isakson et al.*, 1997). It is a fortunate thing that the deposition of cadmium, very dangerous for life, is relatively low. Finally, one can see from the table that metal concentrations in precipitation are rather high in Hungary as compared to information obtained in other parts of the world. Hungarian concentrations are roughly higher by a factor of 2 relative to those measured in Sweden and North America.

Results of atmospheric budget calculations for Hungary indicate (*Molnár et al.*, 1995) that the emission of cadmium, copper and lead is smaller than the respective depositions. This means that continental wide emission mitigation is needed to control their deposition values. On the other hand, caused by fossil fuel use, in Hungary more nickel and vanadium is emitted than deposited indicating that the country is a net atmospheric source of these elements.

It goes without saying that there is not intention here to discuss the importance of the above metal depositions in the uptake of microelements by plants. Information in Table 3 gives the necessary input data for such an evaluation which is badly needed for estimating the impact of atmospheric pollution on the microelement cycle of natural and agricultural vegetation.

4. Conclusions

On the basis of information presented and discussed in this paper we can conclude that atmospheric wet deposition carries an important amount of material continuously into terrestrial and aquatic ecosystems. Before the industrial era this material input certainly was an important source of different elements necessary for natural vegetation. Nowadays, however, the atmospheric flux is significant first of all in areas where no fertilizers are used (e.g., national parks, forests, and grazing grounds).

During about the last fifty years the wet deposition rate of sulfur and some other materials (e.g., ammonium and calcium) has decreased in a significant way due to economic changes and continental-wide air pollution management. It is a bit surprising that the nitrate deposition has also decreased in spite of the fact that no important change in nitrogen oxide emissions occurred. Further research is necessary to prove and/or elucidate this finding.

It is an important fact that the acidity of precipitation is lower now than about twenty years ago when acid rains caused important ecological damages in Europe, mostly in forest and lake ecosystems. It is certain that the improvement of the situation is a consequence of substantial decrease in sulfur dioxide emissions.

The comparison of metal depositions of anthropogenic origin in Hungary with those gained elsewhere indicates that Hungarian deposition rates are rather important. The estimation of the impact of these depositions on material budget of plants is an obvious need for further research.

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Climate change and soil processes

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Abstract—The changes in the gas composition of the atmosphere may lead to increasing temperature with heterogeneous spatial and time distribution. These changes are reflected in changes of vegetation and land use pattern with considerable impacts on soil formation properties and soil characteristics. These potential changes are discussed in the paper, analysing the impact on soil formation processes, the moisture regime, and soil degradation processes under the influence of four plausible climate scenarios.

Key-words: climatic scenarios, soil formation processes, soil moisture regime, soil degradation processes

1. Introduction

Human activities lead to changes in the global environment at virtually unprecedented rates, with potentially severe consequences to our future life. The study and solution of the problems of global environmental changes require urgent and efficient actions. This crucial task formulates a challenge for science: to describe and understand the interactive physical, chemical, and biological processes that regulate the Total Earth System, the unique environment for life (*Toward...*, 1988).

2. Potential changes in climate

In the last century considerable changes took place in the *gas composition of the atmosphere* due to natural processes and human activities, such as increasing energy consumption, industrialization, intensive agriculture, urban and rural development. This may lead to a *rise in global temperature* with a rate of 0.1–0.8°C per decade. The spatial and temporal patterns of temperature increases will be heterogeneous on Earth. The changing temperature regime pattern will be followed by considerable changes in *precipitation characteristics:* quantity of rain and snow, their spatial and temporal distribution pattern, rain intensity, etc. Their forecast is even more uncertain (*Scharpenseel et al.*, 1990).

3. Consequences of climatic change

Due to the increasing temperature, an increasing part of the mountain glaciers, the permafrost soil zone, and the Polar ice caps will melt. It leads to changes in the water flow dynamics, including flood waves and surface runoff will result in a *rise of the eustatic sea level.* The forecasted 0.20–1.40 m sea level rise will threaten low-lying, man-protected lands, settlements, agricultural areas, and extended seashores with low slope. Another consequence will be the further extension of salt affected territories under the direct effect of temporal sea water inundations or due to the rise of the sea level-connected water table of saline or brackish groundwaters.

The changing climate will result in considerable changes in the *natural vegetation* and in *land use practice*s (*Greenland* and *Szabolcs*, 1993; *Lal et al.*, 1994). The great vegetation zones will move into the direction of the Poles, with a predicted rate of 25–200 km/100 years. Vegetation — in many cases — cannot tolerate and follow this „velocity" and it leads to considerable changes in the species distribution, dynamics, diversity, and production capacity of various ecosystems. *Land use practices* will follow or modify the natural changes, depending on environmental and socio-economic conditions (*Lal et al.*, 1994).

Changes in the vegetation or land use pattern result in a feedback effect on climate, modifying albedo, surface roughness, micro-circulation processes, the heat and energy balance of the near-surface atmosphere, the characteristics of both temperature and precipitation. Vegetation changes will considerably influence the field water cycle and soil formation processes.

4. Impacts of climate changes on soils and soil processes

Climate changes and their consequences will result in significant alterations in *soil conditions* (*Brinkman,* 1990; *Greenland* and *Szabolcs*, 1993; *Lal et al.*, 1994; *Rounsevell* and *Loveland,* 1994; *Scharpenseel et al.*, 1990; *Tinker* and *Ingram,* 1994). These impacts and their relationships are summarized in *Fig. 1*

(*Várallyay,* 1989, 1990a,b, 1994). The figure clearly indicates why the quantitative evaluation of the impact of any climate change on the soil conditions and soil processes is so difficult and far from being satisfactory. The uncertainities in the long-term global temperature and precipitation forecasts are combined here with the complex, integrated influences of changing vegetation and land use pattern and the changing hydrological cycle (*Arnold et al.,* 1990; *Toward* ..., 1988; *VITUKI*, 1989).

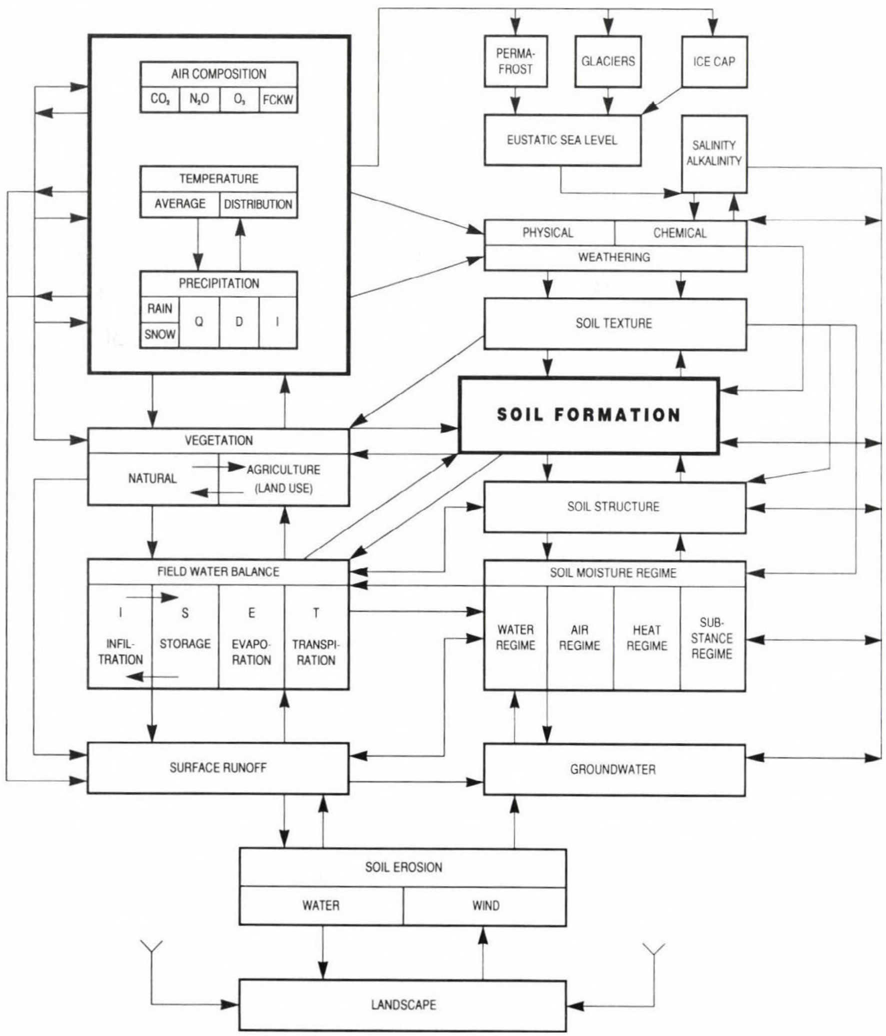


Fig. 1. The potential consequences of climate changes on soil processes.

4.1 Soil formation processes and soil properties

Climate, vegetation, and water regime determine (or at least strongly influence) *soil properties* (*Arnold et al.*, 1990; *Brinkman,* 1990; *Rounsevell* and *Loveland,* 1994; *Scharpenseel et al.*, 1990; *Várallyay,* 1990a).

Soil texture (particle-size distribution) is a rather constant soil parameter: characteristic response time is $>10^3$ years. *Physical and chemical weathering* are slow processes. Climate changes result in only slight and slow changes in these processes, and only moderately influence the "*biological weathering*". The influence is more expressed on the texture differentiation within a soil profile and on the organic matter turnover (*Arnold et al.*, 1990).

4.2 Moisture regime

The integral impact of climatic–hydrologic–vegetation–land use changes are reflected by the *field water balance* and *soil moisture regime* (*Antal et al.*, 2000; *Várallyay,* 1990a,c). Their components and the potential impact of 4 plausible climate change scenarios on these factors are summarized in *Fig. 2.*

An increase in *precipitation* will be followed by an increase of:

- surface runoff (R) in hilly lands with undulating surfaces and without permanent and dense vegetation, if the infiltration rate, permeability, and water storage capacity of the soil are limited;
- infiltration (I) and water storage (S) within the soil if they are not limited, in flat lands;
- groundwater recharge (G) if the soil has good vertical drainage, permeability is not limited, especially in low-lying areas;
- evaporation (E), if infiltration is limited;
- transpiration (T) in the case of well-developed plant canopies.

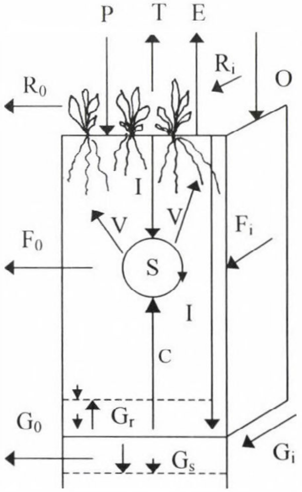
The decrease in precipitation results in adverse changes.

The rise in *temperature*

- increases the potential E and T, if the plant canopy is not suffering from limited water supply due to climatic or soil-induced drought, e.g., low precipitation or limited water storage capacity;
- decreases R, I, S, and G, especially if it is accompanied by low precipitation;
- decreases the intensity (depth) of permafrost; it will modify the geographical boundaries of permafrost, opening possibilities for increasing water storage and water movement, biological activity and soil formation processes within the unfrozen part of the soil.

The decrease in temperature will result in adverse changes.

These general influences are modified with the impact of vegetation characteristics (type, density, dynamics, species composition, biomass production, litter and root characteristics), and depend greatly on the type, intensity, spatial and temporal distribution of atmospheric precipitation. Man's influence is still more complex. Land use, cropping pattern, agrotechnics, amelioration (including water and wind erosion control, chemical reclamation, irrigation, and drainage), and other activities sometimes radically modify the field water balance and its components (*Arnold et al.,* 1990; *Lal et al.,* 1994; *Tinker* and *Ingram*, 1994; *Toward* ..., 1988).



P	= precipitation
O	= irrigation water
R_iR_o	= surface runoff (in and out)
F_iF_o	= filtration in the unsaturated zone (in and out)
G_iG_o	= groundwater flow (in and out)
G_r	= rise of water table
G_s	= lowering of the water table
I	= infiltration into the soils
i	= infiltration into the groundwater
S	= storage within the soil (recharge)
V	= filtration to the plant roots uptake by plants
T	= transpiration
E	= evaporation
C	= capillary transport from groundwater

Factors	**Cl**			
	Cold, wet	**Cold, dry**	**Hot, wet**	**Hot, dry**
P	I	D	I	D
R	I	d,D	I	D
G	i	d	i	D
I	I	d	I	D
I	i	D	(i)	D
S	I	d	(I)	D
E	D	E	E	I
T	D	E	i	I
F	-	-	-	-
G_r	i	-	(I)	-
G_s	-	I	-	I

Fig. 2. Components of the field water balance and soil moisture regime, and the influence of four potential climate scenarios on these factors: *i* and *I*: slight and great increase; *d* and *D*: slight and strong decrease; *E*: no change (equilibrium).

5. *Impacts of climate changes on soil degradation processes*

Soil degradation is usually a complex process in which several features of soil deterioration can be recognized. Soil degradation may lead to the loss of land or soil; limitations in normal soil functions; decrease in soil fertility and "productive capacity" (*Oldeman et al.,* 1991; *Várallyay,* 1989). Soil degradation may be the result of natural factors and/or human activities.

For the assessment of soil degradation a world-wide project was initiated by UNEP. In the framework of GLASOD (GLobal Assessment of SOil Degradation), a map was prepared in the scale of 1:10,000,000 on the present status and potential future development of the various human-induced soil degradation processes.

In *Fig. 3* an attempt was made to show the potential impacts of the four basic climatic scenarios on the main soil degradation processes. In the figure their main natural and human causative factors were summarized, as well.

	Sign	Climatic scenarios				Causative factors	
		Cold and dry	Cold and wet	Hot and dry	Hot and wet	Natural	Anthropogeneous
Soil erosion by water	E	4	1	4	1	1,2,3	9,10,11,12
Soil erosion by wind	D	3	4	[illegible]	4	3	9,10,11,12
Acidification	A	3	1	4	1	2,4	13,15
Salinization/Alkalization	S	[illegible]	4	1	4	5,6,8	14
Physical degradation	P	3	2	2	1	-	10,12
Extreme moisture regime (water logging)	M	4	1	4	2	5,6,7	11,12,14
Biological degradation	B	3	2	2	1	-	11,16
Unfavourable nutrient regime	N	3	2	2	1	(2,6)	13
Soil pollution (toxicity)	T	4	3	3	4	-	16

1 Strong

Medium

3 Slight

4 No or negligible

Causative factors:

<u>Natural</u>
1. Undulating surface
2. Parent rock
3. Lack of permanent and dense vegetation
4. Litter decomposition
5. Low-lying lands
6. Improper drainage
7. High water table (non saline)
8. High water table (saline)

<u>Anthropogeneous</u>
9. Deforestation
10. Overgrazing
11. Irrational land use
12. Improper tillage practices
13. Irrational fertilizer application
14. Improper irrigation
15. Acid deposition
16. Chemical soil pollution

Fig. 3. The influence of four main climatic scenarios on the main soil degradation processes, and their natural and anthropogenic causative factors

(a) Soil erosion

There are no linear relationships between mean annual precipitation, surface runoff, and the rate of denudation/erosion. The rate, type, and extension of *soil erosion* depends on the combined influences of climate (primarily the quantity and intensity of rainfall), relief, vegetation (type, continuity, density), and soil erodability characteristics.

The main impacts of potential climate change on soil erosion are as follows (*Lal et al.*, 1994; *Várallyay,* 1990a):

- higher precipitation may result in an increasing rate of erosion (→ higher runoff), if it is not balanced by the increasing soil conservation influences of better vegetation due to better water supply;
- lower precipitation generally reduces the rate of erosion, but it can be counterbalanced by the less intensive soil conservation influence of poor vegetation due to the non-adequate water supply of plants; this can be the consequence of increasing temperature, as well;
- lower precipitation (or higher temperature) may intensify wind erosion;
- increasing temperature may reduce the erosion hazard moderating the permafrost influence (limited infiltration rate and water storage capacity of the soil); but may considerably increase the erosion-risk reducing the snow: rain ratio in the cold regions and in high mountains.

(b) Acidification

Increasing precipitation may intensify downward filtration and leaching, consequently may help acidification. Climate determines the dominant vegetation types, their productivity, the chemical character and decomposition of their litter deposits, and influences the development of soil reaction in this way (*Brinkman,* 1990; *Scharpenseel et al.,* 1990).

(c) Salinization/alkalization

One of the well-pronounced consequences of the forecasted global "warming up" is the rise of eustatic sea level. Higher precipitation (→ increasing rate of downward filtration → leaching) will reduce, lower precipitation and higher temperature will intensify salinization/alkalization processes: higher rate of evapo(transpi)ration → increasing upward capillary transport of water and water-soluble salts from the groundwater to the root zone + no or negligible leaching (*Rounsevell* and *Loveland,* 1994; *Várallyay,* 1994).

Similar tendencies characterize the leaching and *accumulation of carbonates*, which may lead to the formation of compact and impervious hardpans, petrocalcic horizons.

(d) Physical degradation (structure destruction, compaction, sealing)

The influence of climate change on the changes in soil structure is a complex process with numerous direct and indirect impacts. The most important direct impact is the aggregate-destructing role of raindrops, surface runoff, and filtrating water. The indirect influences act through the vegetation pattern and land use practices.

(e) Biological degradation

Temperature, precipitation, and vegetation changes all considerably influence biological soil processes, but only a few data are available on these consequences (*Arnold et al.*, 1990; *Lal et al.*, 1994; *Tinker* and *Ingram,* 1994, *Toward ...*, 1988).

(f) Unfavourable changes in the biogeochemical cycles and the plant nutrient regime

One part of these changes is connected with changes in the soil moisture regime (the ratio between downward and upward water movement in the unsaturated zone; leaching–accumulation), another part is related to the abiotic and biotic transformation phenomena (fixation, immobilization–release, mobilization; changes in solubility and redox status; etc.) in the chemical-biological cycles of various elements (*Scharpenseel et al.,* 1990). High precipitation helps leaching, filtration losses (→ potential groundwater "pollution") and reductive processes; low precipitation → dry conditions may reduce the solubility, consequently mobility and availability of less soluble compounds.

6. Conclusion

Because of many uncertainities, more detailed, integrated multidisciplinary studies are required on the quantification of the existing facts and processes in the air–water–soil–geological deposits–plant continuum for a more real environmental forecast and for the rational control of soil processes under various potential and real climate change scenarios.

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Agrometeorology of maize — investigation carried out at the Hungarian Meteorological Service

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Abstract—Maize is one of the most important agricultural crops in Hungary. The high level of production is due to the introduction of new varieties as well as the proper agrotechnology which is the result of investigation of agro-ecological conditions of Hungary. In this review paper a draft overlook is presented about the works of the Hungarian agrometeorologists concerning the meteorological aspects of maize production. The results of field experiments, statistical investigations are presented as well as the results of the experiments carried out by dynamic simulation model of CERES-Maize. It was demonstrated by all of the methods that shortage in water supply is the critical factor in maize production under the climate of Hungary. The risk of drought varies between 20% and 40% among the different agroclimatological districts of the country.

Key-words: maize production, dynamic simulation model, statistical analysis, evapotranspiration, Hungarian investigations.

1. Introduction

Relationship between climatic elements and agricultural production had been revealed in different ways. Up to the sixties statistical methods were used to estimate yields from weather conditions, e.g., average air temperature and precipitation in various months were related to the final yield (*Berényi*, 1945). In the past decades, mechanistic models have been developed in which crop growth is simulated in relation to observed weather conditions. These models integrate knowledge of the most important effects of weather on individual crop growth processes like global radiation on photosynthesis, air temperature on development. Using these models it is possible to study the overall effect of weather on crop yield. In our study we use maize as test plant because it is one of the most important agricultural crops in Hungary. The sowing area is about 1 million hectars. Due to the introduction of new hybrids from the early seventies, yield level increased from 3 tons/ha up to 6 tons/ha as a country average. Hybrids can also be categorized according to the length of the vegetation period by FAO numbers. Climate of Hungary is suitable to grow hybrids with FAO number from 200 to 600.

Intensive agrometeorological experiments had been started at the Agrometeorological Research Stations in 1963 to assess the relationship among maize production, development, and climatic elements. The investigated subjects were description of phenological development, water consumption relating to irrigation, and estimation of yield production (*Antal,* 1966a; 1966b).

Investigation on maize production according to methodology belongs to one of the three direction: field experiments, statistical analysis and system analysis.

2. Field experiments

Majority of the field experiments carried out in the frame of the Hungarian Meteorological Service aimed to assess the optimal evapotranspiration and water requirement for irrigation (*Antal* and *Posza*, 1966; *Antal et al.*, 1971; *Posza*, 1973; *Posza* and *Tóth*, 1974; *Kádár* and *Szilágyi*, 1978). In agrometeorological observatories of the Hungarian Meteorological Service at Szarvas, Kecskemét, and Keszthely Thornthwaite type compensation lyzimeters (*Antal,* 1966b; 1968) were used to assess the optimal evapotranspiration of maize. The seasonal amounts of maize evapotranspiration varied between 350–680 mm depending on weather conditions and agrotechnics.

The importance of agrotechnics in water regime of plants also manifested in field experiments. The efficiency of fertilisation together with water demand of the plant were studied by *Antal et al.* (1975) and *Dávid* (1977). As a result of investigations of their work could be assessed that the optimal level of fertilization was about 150–200 kg nitrogen/ha, 50 kg potassium/ha, and 50 kg phosphorus/ha in average. When the amounts of nitrogen is higher than the optimum level, water demand increases without a reasonable yield increase.

Description of developmental rate in function of environmental variables was given by *Pletser* and *Szalai* (1979) for thirty maize hybrids.

3. *Statistical analysis*

Statistical analysis was used to compare the climatic requirements of different varieties (*Ábrányi*, 1976; *Ábrányi et al.*, 1977a, 1977b; *Ábrányi et al.* 1981). Using long time data series of county averages of yield, the probability of yield decrease due to aridity was provided by *Varga-Haszonits* (1979). The smallest number of years with yield decrease caused by drought was found in Zala county (20%), the largest number of years when yield decrease occured because of drought was found in Bács-Kiskun, Pest, and Győr-Sopron counties (40%). Stohastic relationships between climatic elements and yield level were found by *Kmetykó* (1984). The paper of *Kmetykó* presents the forecasting equations and results of the investigations concerning the relation between the relative yield and climatic conditions required by maize for county Tolna. 22 meteorological variables and indices were investigated as predictors. The statistical method applied by the author is linear regression analysis, the period of investigation is 10 years (1970–1979). The most important meteorological parameters were related to precipitation because this is the limiting factor in maize yield production in Hungary. Different methods of statistical analysis were applied by *Magyarits* (1981) who applied 19-year long data series of maize yield from the experimental field of Martonvásár. The frequency of "extremely good years" and "unfavorable years" was 20%, and 37%, respectively. According to monthly mean temperatures and precipitation amounts, high level of yield can be expected if the meteorological conditions follow the tendency presented in *Table 1*.

Table 1. Optimal conditions for high level of maize yield. ΔT – anomaly in monthly mean temperature, ΔP – anomaly in monthly precipitation sum, + above, – below the normal, O-indifferent.

	April	May	June	July	August
ΔT	–	+	–	+	+
ΔP	O	+	+	+	O

4. *System analysis*

The first efforts to describe biomass and yield production using a system analysis was done by *Ábrányi* (1976). Later *Dunkel* (1981), *Dunkel* and *Hunkár* (1985), *Hunkár* (1986), *Dunkel et al.* (1987), *Hunkár* (1987, 1990, 1994), *Hunkár* and *Dunkel* (1987), *Bacsi* and *Hunkár* (1994) used dynamic simulation models to study biomass and yield production of maize. Finally, the crop model CERES-Maize (*Jones* and *Kiniry*, 1986) was adapted and validated by data measured at the Agrometeorological Research Station of the Hungarian Meteorological Service at Keszthely (*Hunkár,* 1994). A sort of simulation experiments has been carried out for different purposes. The basic principles built into the model are as follows together with the results of the simulation experiments.

4.1 Climatic conditions of phenological development

The base temperature for maize germination is 8–10°C (*Ábrányi et al.*, 1977b). Growing degree days in the sowing-emergence phase are independent from the variety. The length of this phase is determined by soil temperature and moisture content with an average of 10–15 days (*Derieux* and *Bonhomme*, 1982). In Hungary the optimum sowing date is between April 20 and May 5.

The length of the vegetative period, from emergence until tasseling is determined by the variety mainly in the beginning of development. During this period leaf area increases, the rate of the increase and the final size of leaf area as assimilatory surface determines the radiation absorption as the source of energy for photosynthesis. Biomass accumulation can be derived directly from radiation absorption. Other environmental variables also influence the intensity of the photosynthesis. The optimum range of temperature for maize is around 30°C (*Vong* and *Murata*, 1977). According to the present climate of Hungary, the higher the temperature the faster the ontogenic development. Water demand of plants is the other important factor which has an effect on the development. Lack of water manifests itself as the vegetation period is shortened. The time of tasseling is usually between July 10 and 20. The phase from tasseling to maturity is the period of grain filling. The length of this period also depends on the variety. In general the "early type" varieties of short growing season are matured at the end of August, while the "late type" varieties of long growing season are matured in the middle of October. The length of the whole vegetation period from the emergence until maturity is important from the viewpoint of the potential biomass and yield accumulation. The longer the period the higher the potential yield. A warming climate may cause faster development, higher water demand and both of them may result in yield decrease.

4.2 Yield production of maize

The actual final yield of a crop is determined by many factors: weather, crop variety, fertiliser supply, soil conditions, occurrence of pests and diseases. After *Penning de Vries* and *van Laar* (1982) several production levels can be distinguished. In the potential production level the crop is optimally supplied with water and nutrients and free from pests, diseases, and weeds. Crop growth is only determined by crop factors, temperature, and radiation. In water limited condition when nutrients are in optimum level and the crop is free of pests, diseases, and weeds, yield is limited by the availability of water. The effects of climatic variables on yield can be studied on this second production level.

A crop simulation model must be capable of representing the actual performance of crops grown in any particular region before it can be applied to the prediction of agrotechnology or climatic variation. Crop analysis of maize was carried out at the Agrometeorological Research Station at Keszthely and Szarvas, Hungary in the years 1976–1991, and 1988–1997, respectively (*Hunkár*, 1994). Measured and simulated data of silking date, maturity date, leaf area maximum, final biomass, and grain yield ware compared in *Table 2*. The average differences between the predicted and observed plant characteristics are not more than 4% at Keszthely and 5.6% at Szarvas. The probability, that CERES-Maize model simulation results yield and biomass within 15% error in a given year, is 80%.

Table 2. Results of model validation. Averages and standard deviations of predicted and observed plant characteristics at Keszthely and Szarvas for the periods 1976–1991 and 1988–1997

	Predicted		Observed	
	Average	**Standard deviation**	**Average**	**Standard deviation**
Keszthely				
Silking date (day of the year)	199.7	8.3	199.5	8.0
Maturity date (day of the year)	260.4	14.6	257.0	14.2
Grain yield kg/ha	10,681	2190	10,294	2008
Szarvas				
Silking date (day of the year)	192.2	5.1	192.4	5.2
Maturity date (day of the year)	247.2	17.3	261.9**	13.1
Grain yield kg/ha	8476	4570	8020	4110

** harvesting day

Earlier results have been proven that under the climate of Hungary precipitation is the critical factor of the agricultural crop growing. Simulation model CERES-Maize provides an opportunity to estimate yield with optimal moisture supply. It means that the submodel of water balance is switched off. According to *Penning de Vries* and *van Laar* (1982) that is the production level No. 1. Simulation experiments had been carried out for Keszthely and Szarvas assuming no water stress during the vegetation season. It is called potential production. In *Fig. 1* the simulated potential production together with simulated actual production are presented for Keszthely, and Szarvas, respectively.

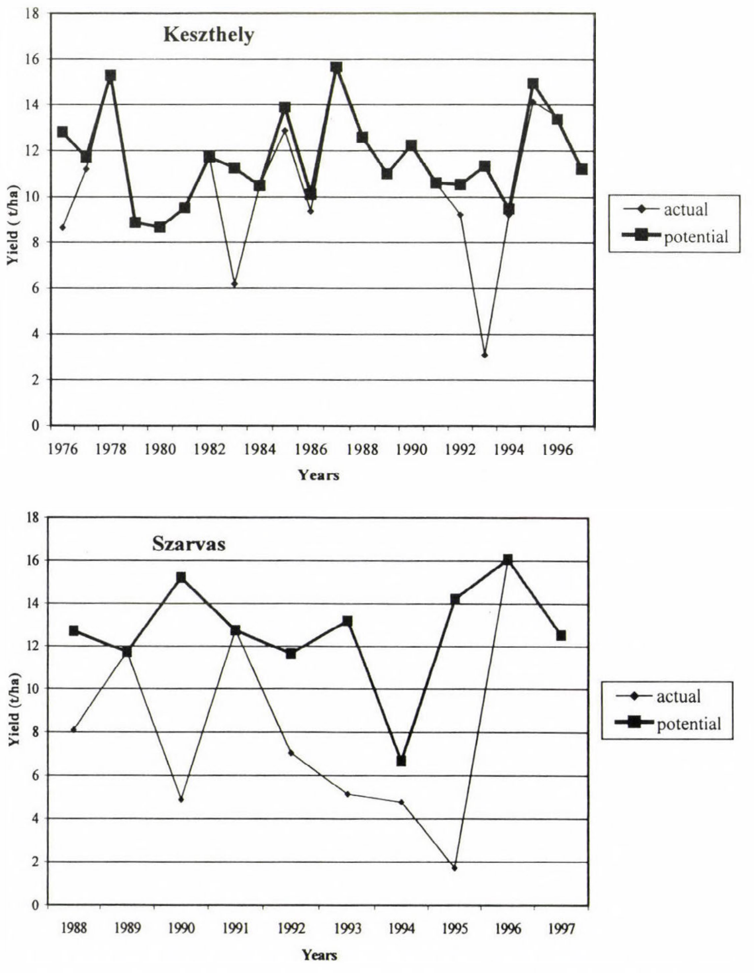

Fig. 1. Simulated yield production of maize with optimal and actual water supply for Keszthely and Szarvas.

The different climate of Keszthely and Szarvas is manifested in the potential production levels as well as in the frequencies of yield decrease due to water shortage. The average of the potential production is 11.7 tons/ha with CV=17%, and 12.7 tons/ha with CV=20% at Keszthely, and Szarvas, respectively. If the actual water supply, i.e., precipitation as driving force is taken into account in the model, there are years when significant yield decrease occurs. Under the climate of Keszthely, water shortage caused significant yield decrease in 14% of the years, while in Szarvas, which has a warmer and drier climate, in 50% of the years. The average yield on the production level No. 2. is 10.7 tons/ha (CV=29%) and 8.5 tons/ha (CV=54%) at Keszthely, and Szarvas, respectively.

It has to emphasize that these results came from simulation experiments, nevertheless they fit the empirical observations. To evaluate the model outputs in a proper way, some details about the plant-water relations have to be considered.

4.3 Behaviour of CERES crop model relating to evapotranspiration

Under the climate of Hungary, moisture supply is a critical and rather risky factor of agricultural crop production. It means that a crop growth model used in Hungarian circumstances must be capable to simulate water relations in satisfactory way. Details about the philosophy of CERES relating to the water balance is given in the *Appendix* after the works of *Ritchie* (1985).

To study the evapotranspiration submodel of CERES crop model a dry and a wet growing season were chosen. The growing season of maize takes from April 1 until September 30 in this relation. In the years 1975–1994 in Keszthely during the growing season, the average amount of precipitation (AVG) was 356 mm (the standard deviation (STD) was 71 mm). The season was assumed to be dry when the amount of precipitation was less than AVG–STD=285 mm, and wet when the amounts of precipitation were above AVG+STD=427 mm.

According to these conditions:

dry seasons	wet seasons
1977 (275 mm)	1975 (488 mm)
1981 (265 mm)	1984 (440 mm)
1982 (274 mm)	1987 (504 mm)
1988 (277 mm)	1989 (432 mm).

Simulation experiments were run for the year 1981 as a dry, and 1987 as a wet year.

Potential evapotranspiration calculated by CERES was compared to the values calculated by the local empirical formula developed by *Antal* (1968), which is widely used in Hungary.

In *Fig. 2* the accumulated potential evapotranspiration is shown. In the dry year CERES gives lower amounts, but in the wet year CERES gives higher amounts of potential evapotranspiration than the amounts calculated by the empirical formula of *Antal*.

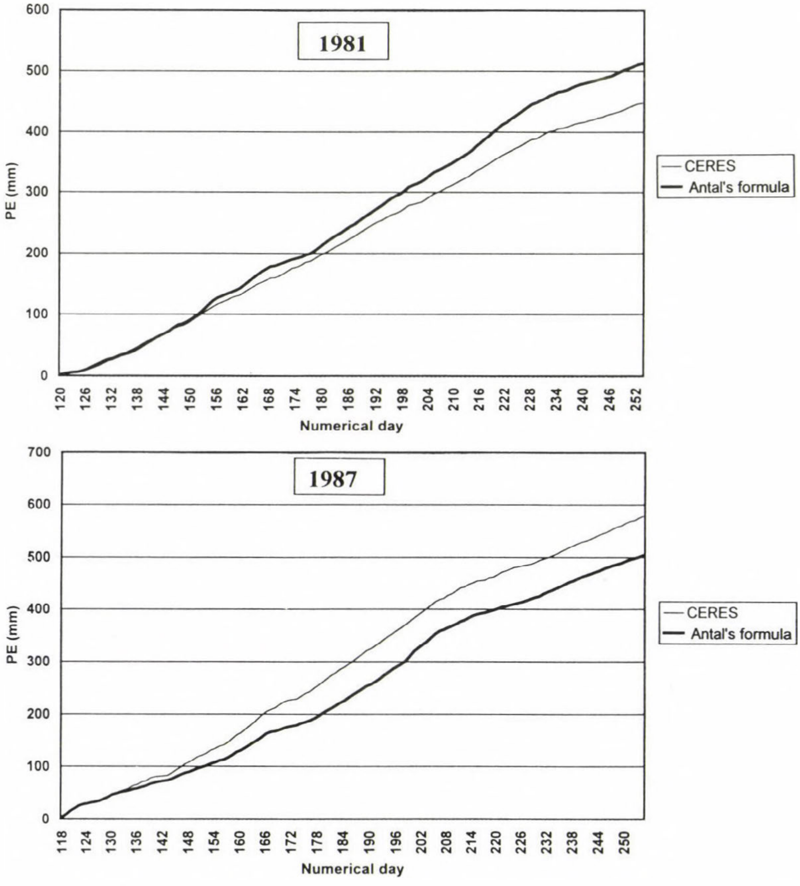

Fig. 2. Accumulated potential evapotranspiration calculated by CERES and *Antal's* formula in a dry (1981) and wet (1987) year at Keszthely.

The actual evapotranspiration simulated by CERES was compared to the measured evapotranspiration. In our case evapotranspiration measurement means that the soil moisture content was determined thermogravimetrically with about 10 days frequency, and the water balance was followed for those periods:

$$ET = W_1 - W_2 + P,$$

where W_1, W_2 are soil moisture content in the upper 1 m layer in mm, in the beginning and end of the period, P is the precipitation during the period.

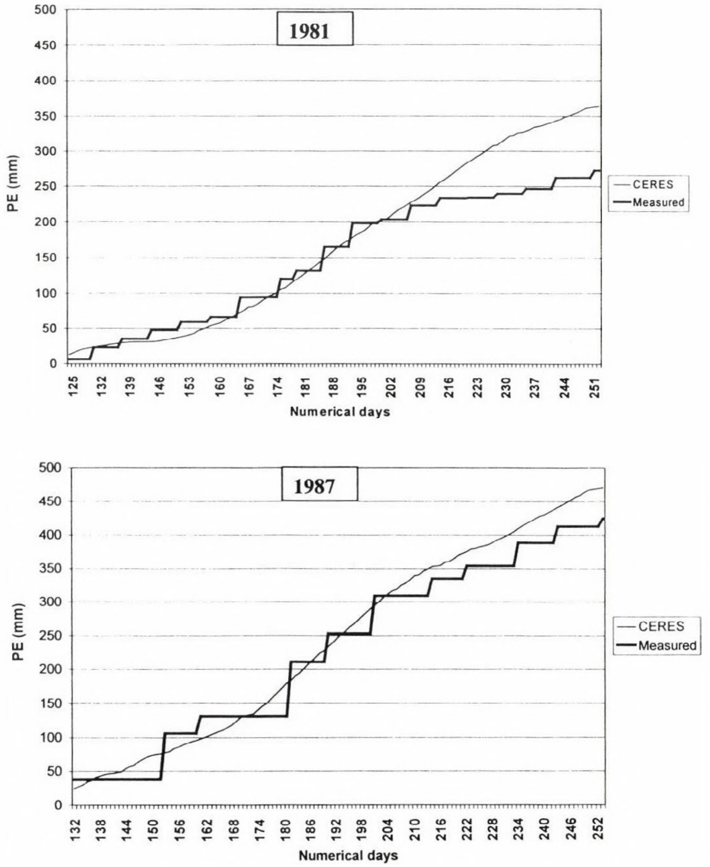

Fig. 3. Accumulation of calculated and measured evapotranspiration in a dry (1981) and wet (1987) year at Keszthely.

The final amounts of evapotranspiration simulated by CERES are higher than the measured ones both in dry and wet years, but the difference is significant only in the dry year. In the wet year the measured and simulated values show a good agreement (*Fig. 3*).

Considering that the actual evapotranspiration calculated by CERES is less than the potential one only when the amount of available water is below 25% of field capacity at least in one layer, even in a relatively dry soil the crop transpires water without any restriction. In reality the crop transpiration is more sensitive to the soil moisture conditions. Field experiments carried out by *Posza* (1973, 1975) demonstrated that when the amount of plant extractable water is below 50% of the maximum, the resistance against water uptake increases, therefore, the transpiration rate will be reduced. The plant physiological process of adaptation to arid conditions might play important role in water balance of that areas, and there is a need to include this process into crop models as well.

APPENDIX

Philosophy of CERES crop models relating to the water balance

In CERES crop models the submodel of water uptake, evaporation and transpiration is based mainly on the work of *Ritchie* (1985). Hereby a short summary of the concept is given.

Principles:

- The soil is divided into layers. Number of layers are optional, but the depth of the first upper layer is not more than 0.05 m. Number of layers is suggested between 9–15 depending on the maximum root depth which is required to know. Hydrological characterization of the individual layers is needed.
- Daily amounts of precipitation are input data.
- Initial soil moisture status has to be known.
- Equilibrium evapotranspiration rate is governed by the global radiation and daytime temperature:

$$E_q = G(2.04 \times 10^{-4} - 1.83 \times 10^{-4} \cdot \alpha)\,(T_d + 29),$$

where E_q equilibrium evapotranspiration,
G global radiation,
α albedo,
T_d daytime temperature.

The potential evapotranspiration (*PE*) is calculated from the equilibrium evapotranspiration as a function of daily maximum temperature

(T_{mx}), where the numeric constant 1.1 is used to account for unsaturated air, when T_{mx} is between 5°C and 24°C:

$$PE = 1.1 \times E_q.$$

When T_{mx} is greater than 24°C the constant is increased above 1.1 to account for advection, and

$$PE = E_q\ [0.05\ (T_{mx} - 24) + 1.1].$$

When T_{mx} is less than 5°C, the constant is reduced to account for cold temperature causing an additional decrease in *ET*:

$$PE = E_q\ 0.01 \exp\ (0.18\ T_{mx} + 20).$$

- Soil evaporation and plant transpiration are separated.

The potential rate of soil evaporation (PE_S) is then calculated from the leaf area index, *LAI*. When *LAI* is less than 1,

$$PE_S = PE\ (1 - 0.43\ LAI),$$

and when *LAI* is greater than 1,

$$PE_S = E_q \exp\ (-0.4\ LAI).$$

The actual rate of soil evaporation, E_S is limited by a local parameter and depends on the soil moisture of the upper layer.
The potential transpiration rate by the crop (PE_C), since germination is calculated based on the leaf area index (*LAI*), is

$$PE_C = PE\ LAI/3,$$

when *LAI* is less than or equal to 3. When *LAI* is greater than 3,

$$PE_C = PE.$$

If $PE_C + PE_S$ is greater than *PE*, then

$$PE_C = PE - PE_S.$$

- Root water absorption provides the water for transpiration. The root water absorption in CERES is calculated using a law of the limiting approach, whereby the soil resistance, the root resistance, or the atmospheric demand dominate the flow rate of water into the roots. All those processes are taking place in each layer of the soil. The number and depth of layers are given by the initialization of the model run.

  The actual plant transpiration (E_C) is limited by the potential water uptake values which are determined by the amount of plant extractable soil water in a given layer. The actual evapotranspiration is less than the potential value only in that case, when the model finds less than 25% of the amount of plant extractable water of the field capacity in a soil layer.

  Because the actual rate of transpiration is known only after the rate of water uptake by the roots is determined, the actual evapotranspiration rate (ET) is recalculated:

$$ET = E_S + E_C.$$

It has to be mentioned that the "water submodel" of CERES is under the way of a continuous development, and the later versions are more sophisticated in this context.

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Slices of plant–water relation in reflection to investigations carried out at Agrometeorological Research Station of Keszthely

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Abstract—Parallel with theoretical consideration, the investigation conducted at Keszthely Agrometeorological Research Station to study two important counterparts of the plant-water relation is discussed. The first part of the study contains the importance and role of the stomata in balancing plant-environment interaction. Methods of investigations of the stomata are also included. The modeling approach was also applied to justify a new consideration for average resistance estimation of maize. In the second half of the paper, method to assess the plant water supply by using the concept of the crop water stress index (*CWSI*) is introduced. Suitability of the index adaptation under Hungarian climatic and plant growing conditions for 13 consecutive years was evaluated. We present the influence of such scarcely experienced factor as nitrogen on the formation of determined *CWSI*.

Key-words: plant-water relation, stomatal resistance, crop water stress index (*CWSI*), irrigation timing.

1. Introduction

Methods of evaluating the plant watering level to assess irrigation cover a wide range from direct plant observations to theoretical consideration based on energy balance of canopies. Almost the whole plant growing area of Hungary is included in a special network for evapotranspiration determination using compensation lysimeters, which was established in the early 70-es. Operation of these lysimeter-stations for several decades revived a valuable data-set appropriate for producing studies, which contain a wide range of new results about the interactions in the soil-plant-water system. The observations carried out in the scope of the network include traditional evapotranspiration determinations (*Antal,* 1966a, 1966b, 1968a, 1968b, 1972; *Antal* and *Posza,* 1967, 1970; *Antal* and *Endrődi,* 1972), as well as other experiences on the crop-environment relationships (*Antal et al.*, 1975; *Tóth*, 1978; *Ruzsányi,* 1981, 1990) which highly influence the water balance of plants. Almost the whole range of locally grown arable and horticultural crops (*Kozma* and *Füri,* 1974; *Stollár* and *Gergely,* 1978; *Posza* 1980*)* were observed at the stations of the network. Results, as they comprise observations for the whole area of Hungary, are unique and irreplaceable on their own. In spite of the newly constructed sophisticated instruments, until now the lysimeters provide valuable information about temporal changes in water loss of plants when the number of studied cropping systems is limited (*Prueger et al.*, 1997; *Tolk et al.*, 1998). From the early 80-es, investigations have been built to the earlier experiments on water balance of plants conducted at Keszthely Agrometeorological Research Station.

In the past three decades, the techniques have greatly developed in sensor accuracy and availability in field use. These instruments combine the latest technology in plant observations and computer know-how into a small, portable, easy to handle unit. Observations by these instrument-units opened a new way in studying the plant–water–atmosphere continuum more precisely than we have ever done it in the past.

The purpose of the study was to present a selection of investigations on plant-water relation carried out at Keszthely Agrometeorological Research Station during the past two decades. Observations, aimed closely at related plant-water connection as irrigation timing and role of stomata, were focused. The study of the role of irrigation scheduling in plant growing has of primary importance, because the weather of the growing seasons became drier in the past decades (*Mika,* 1991). Until now, the stomatal resistance is one of the most popular parameters in investigating the plant-water relation. The paper contains only a few details on the experiments carried out at Keszthely Agrometeorological Station. Only those parts of the theory are partially conferred, which are necessary to understand the explained subject. Detailed description of the applied theory and methodology can be obtained from the cited publications.

2. Material and methods

Field study was conducted at Keszthely Agrometeorological Research Station, on a Ramann type brown forest soil of mean bulk density of 1.5 Mg m^{-3} in the top 1 m of the profile and an available water capacity of 210 mm m^{-1}.

Statistical design of the experiment was in a complete block design, because of the fixed installation of the lysimeters. Data were analysed using combined multiple analysis of variance.

Maize hybrid used was mainly the Norma SC (FAO 380), a dent variety, tolerant to water stress. In stomatal resistance experiment a second maize hybrid, MVNK 480, a dent variety was also grown similarly to Norma. While the hybrid Norma is water stress tolerant, the MVNK 480 is bred for irrigated conditions. More details on different behaviour of the two hybrids are in *Anda* (1998). Both hybrids are commercially grown in Hungary. Plant protection and other agronomic procedures — except of nitrogen fertilizer dosing — were established according to standard agronomic practices for optimum maize production in Hungary, mainly in the surroundings of Keszthely.

The size of the experimental area was 0.7 ha installed with 24 compensation evapotranspirometer pots in the southern edge of the field. The size of plots agreed with the surface area of the lysimeter growing chambers (4 m^2). In most of the years, the seed was sown at the end of April or in early May. Plant density was hand-thinned to 7.0 m^{-2} in each year. Fertilizer nitrogen applied was 100 kg N ha^{-1} at planting for the half of the treatments. The other half (controls) were grown without N fertilization. In the irrigation timing trials two experimental factors were investigated: the irrigation process itself, and the influence of nitrogen on *CWSI*. The following abbreviations of the treatments were applied in the figures:

- no N fertilization, C,
- 100 kg nitrogen ha^{-1}, N,
- irrigation, I,
- lysimeter growing chambers, ET,
- rainfed - or control treatments, P.

In the field experiment three watering levels were simulated:

- "Ad libitum" water supply in lysimeter growing chambers,
- rainfed - or control plants,
- irrigated plots by using the values of *CWSI*.

In stomatal resistance observations only crops with non limited watering and the rainfed ones were studied.

Lysimeter provides unlimited water supply from below the bottom of the pots. Surface area of the chambers of Thornthwaite type compensation evapotranspirometers were 4 m^2, and the depth of them 1 m. We filled the chambers with Ramann type brown forest soil, characteristic soil type of the surroundings. Daily sum of evapotranspiration was given by change in the volume of soil water in the chamber, additional water supply (irrigation) through the compensation pot, and precipitation amounts *(Antal,* 1968a).

Plots were irrigated when the *CWSI* exceeded the limitation value of 0.2–0.25, by using a drip irrigation system. (For easier data-handling we applied tenfold values of *CWSI* in our figures.) The amount of water used per irrigation was between 20 and 40 mm.

Stomatal resistance was measured with transient type porometer. We used a LI-COR 60 version between 1990–1992, and later on an improved model of Delta T Manufacturers, the AP4 type. We constructed a daily change in stomatal resistance by presenting hourly values randomly sampled on clear sky conditions. Two maize hybrids — see above — and three sugar beet varieties, Kawemaya, Gála, and Éva served as test plants. The variety Kawemaya is the most frequent grown sugar beet cultivar in Western Europe. Originally it was bred in Germany. The other two varieties, Gála and Éva are breeding in Hungary. We chose these three cultivars for the study, because they are close to each other in canopy architecture, and there is no literature about the difference in their water-relation. In general, the number of repetitions was 3 to 5. In maize, at the beginning of the seasons (May–June) both abaxial and adaxial leaf sides were sampled. Later on only the lower blade was applied for resistance determinations (amphystomatous plant). In sugar beet we measured the resistance of the lower epidermis only (hypostomatous plant). Daily mean resistance was calculated by averaging the hourly values.

After canopy closure, plant surface temperatures for *CWSI* determinations were resoluted by infrared thermometer RAYNGER II. with 2° field of view and an included 8 to 14 μm spectral band filter. The thermometer was hand held 1 m above the plants using an oblique angle of about 30° to the horizontal. The sampling time was 30 seconds with 5 repetitions in each treatment at around solar noon in all measuring days, including holidays. The presumed emissivity of plants was 0.96.

Plant parameters as leaf area and grain yield were also obtained. Weekly assimilation surface size was measured with LI-3000 type portable planimeter, or by using the Montgomery equation on 10 randomly selected sample plants. After harvesting, cobs were taken for dry matter estimation, and they were oven dried at 60°C to a constant weight (5 to 7 days).

3. Results and discussion

3.1 Theoretical introduction of stomata function

All of the water transpired by plants, as well as CO_2 absorbed in photosynthesis pass through the stomata, even though these pores occupy less than 5% of the whole leaf surface. There are two different types of pore on the epidermis with special capacity of moving: to open and to close. While in the higher plants the elliptical shape is the most common, Gramineae have guard-cells arranged in rows. The stomata are the most frequent on the lower surface of the leaf epidermis, but they may occur on other green tissues as stems, fruits, etc. Crops having stomata on both sides of the epidermis called amphistomatous, and those with stomata restricting to their lower leaf epidermis, hypostomatous plants.

The extreme environmental and physiological sensitivities of the stomata enables them to harmonize the balance between water loss and CO_2 uptake. The size and frequency of the stomata vary on both genetic and environmental conditions. Some examples for variation in frequency of stomata — the number of pores on unique leaf surface area — of different plants are presented in *Table 1.*

Table 1. Values of stomata frequency (number of pores per 1 mm^2 leaf area) on the two leaf surfaces after summarized data of *Sutcliffe* (1982) and *Jones* (1992)

	Upper (adaxial)	**Upper** (abaxial)
Plants	**Leaf epidermis**	
ARABLE AND OTHER CROPS		
	28	105–158
Tomato (high light)	0–2	83
Tomato (high light)	111	131
Tomato (high light)	81–74	242–385*
Soybean (mean of 43 cultivars)	149–158	357–418*
Soybean (watered-stressed)	0	135
Millet (mean of 6 cultivars)	54–98	60–89*
Barley (different cultivars)	52	54–68*
Maize	0	14
Spiderwort	33	14
Wheat	25	23
Oats	85	156
Sunflower	40	281
Bean	0	158
Ivy		
TREES		
	0	170
Carpinus betulus	0	350–600*
Malus pumila (different cultivars)	120	120
Pinus sylvestris	39	39
Picea pungens		

*Results published by more authors. For details see the original literature cited in the title of the table.

The stomatal aperture is linearly relates to guard-cell turgor pressure. There are two processes in regulation of the pore movements. The hydro-passive movement of the pores means change in all of the water potential outside the cells, while active variation in osmotic potential of the guard-cells produces the hydroactive term. We should take into account, that both stomata opening and closure are normally active metabolic processes requiring energy of plants.

There are different methods for studying the stomata:

(a) Microscopic observations

Epidermal imprints to characterize the size and frequency of the pores can frequently be applied. In most cases a solution (nail polish) is used to spread over the epidermis, allow it to dry, peel it off, and store for microscopic examination. This method is also suitable for in vivo investigations. The imprints sometimes do not produce reliable results, because different modifications may occur in pore apertures during the preparation of the observation.

Diffusion theory enables to derive estimates of the stomatal resistance by using the microscopically measured pore cross-section and diffusion coefficient. The modified version of the above approach can also be used for estimation of resistance of the intercellular spaces as well (*Jarvis* and *Mansfield,* 1981).

(b) Infiltration method

This approach is only used to study the qualitative differences in pore numbers within one species. To achieve this goal, graded solutions of differing viscosity (various mixtures of liquid paraffin and kerosene) are applied. Viscosity of the solution infiltrating into the pore provides measure of the aperture (*Larcher,* 1980).

(c) Porometry application

The instrument measuring the diffusive transfer of leaf gas exchange is called diffusion porometer. The diffusion porometer provides all of the leaf resistance to water vapor, including any cuticular part and boundary layer resistance in the porometer chamber (*Jarvis* and *Mansfield,* 1981; *Idso et al.,* 1987; *McDermitt,* 1990).

The principle of the transit-time instrument is to measure the humidity increase in a closed chamber resulted from the water loss of a given leaf section. The time taken for humidity increase over a fixed interval may be converted into a resistance value using a previously obtained calibration curve. For calibration the leaf is replaced by wet paper. The range of resistance is simulated by varying the number (size) of precision drilled holes with known resistance values obtained from the resistance theory.

In steady-state porometer air is flowing through a closed chamber — inside the chamber the leaf is also enclosed —, and the transpiration is determined from the flow rate and the water vapor difference in the inlet and outlet air-streams. There are two possible operation modes at using the steady-state porometer. When constant flow rate is applied, the outlet air is uniquely related to leaf resistance (*Day,* 1977). When the porometer operates in null balance mode, the flow rate is adjusted to give a particular relative humidity (*Beardsell et al.,* 1972). Both of them provide the most precise method in determination of the stomatal resistance in the field.

Among the others, errors in porometer application are discussed in details by *Idso et al.* (1988) and *Monteith* (1990). Application of diffusive resistance theory for maize under Hungarian climatic conditions is discussed partially in *Anda et al.* (1997) and *Anda* and *Lőke* (2002).

3.1.1 Diffusive resistance(s) on the water pathway in the plant-soil-atmosphere continuum

Stomatal resistance (r_s) is the most frequently mentioned counterpart of the diffusive resistances, but not the only component of the total resistance system. Transpiration rate can be expressed in the simplest way by using the Fick's law of diffusion. Complexity of diffusive resistances arising on water pathway from the soil to the open air makes the approach of dominator complicated. Plant water uptake depends on soil resistance, extent of root development, and sap movement of the xylem's vascular system. Roots are thought to offer the largest resistance of the whole soil-plant-atmosphere system, and in most cases they are responsible for the great intra- and interspecific variability (*Calveta,* 2000). The stem together with leaves nearly match the resistance of the roots (*Oke,* 1978). In leaves the water vapor diffuses from the wet mesophyll cell walls surrounding the intercellular air spaces to the stomata, allocated between two specialised guard-cells. Resistance of the cuticle is in a parallel circuit with stomatal resistance. When stomata are open, the cuticular resistance is much higher than the stomatal one, and for that most of the researchers ignore it. (Practically, to determine the true cuticular resistance for amphistomatous plants is very complicated, almost impossible). The true stomatal resistance consists of individual, but hardly approachable resistance of mesophyll cell walls and intercellular spaces connected in series. Using a theoretical consideration, the resistance of intercellular spaces is very small comparing to the other counterparts. Neglecting of this component involves about 2–5% error in the total stomatal resistance evaluation (*Pearcy et al.,* 1991).

Determination of the mesophyll cell walls is too difficult, and many uncertainties still exist. Under normal outdoor conditions, the „general" resistance of the mesophyll cell walls was assumed to be about 20 s m^{-1} (another ~5% error in stomatal resistance estimation). Finally, it can be assumed, that nowadays there is no simple way to measure the components of the transfer system distinctly (*Anda* and *Burucs,* 1997). Constant attempts reveal in the literature to improve the estimation of the actual stomatal resistance. *Kang et al.* (2000) tried to improve the transpiration estimation in maize by applying the ratio of momentary and maximum resistances instead of using the simple stomatal resistance for different soil water status. Others constructed simple model to estimate stomatal conductance — reciprocal of the resistance — over a long term (*Yu et al.,* 1998).

The most plants have two evaporating surfaces with different transfer resistances on the abaxial and adaxial leaf surfaces. This difference must also be taken into account at the proper "average" resistance estimation. Theoretical consideration for maize is published by *Anda* and *Lőke* (2002) and *Anda* (2001).

Summering the above results, in most investigations we mean stomatal resistance as a measure of the leaf resistance (r_l), but we can measure sum of three — stomatal resistance, resistance of intercellular spaces, and resistance of mesophyll cell walls —, even though four different resistances, as the determined resistance often includes a part of the boundary layer resistance (r_b). Typical values of the most important parts of transfer resistance are totaled in *Table 2* originating and summarizing the data from different literatures (*Oke,* 1978).

Table 2. Typical values of the counterparts of transfer resistance after *Oke* (1978)

Type of diffusive resistances	**s m^{-1}**
Intercellular spaces (r_i) and mesophyll wall resistance (r_m) together	< 40
Cuticular resistance (r_{cu})	2000–10,000
Stomatal resistance (r_s)	
– minimum for succulents and xerophytes	200–1000
– minimum for mesophytes	80–250
– closed stomata (maximum)	> 5000
Boundary-layer resistance (r_b)	10–100

Although the most meaningful part of diffusive resistance is the stomatal resistance, techniques measuring it separately are not available. In most cases we measure a resistance mentioned stomatal resistance, but everybody knows, that this is not exactly equivalent to the corresponding resistance.

3.2 Scarcely studied factors influencing the stomata movements: the variety and the nitrogen application

Studying the stomatal behavior in its natural environment is very complicated because of the combination of influencing indoor and outdoor factors. The environment of living crops has great variability, and hardly known interaction exists in the plant-environment response. The stomatal response time can often be longer than the duration of the influencing external — environmental — factor. Large amount of literature was dealing with studying the stomata-environment relationship in the past. These results served as the basis for construction of the soil-vegetation-atmosphere transfer schemes, as for example the SVAT scheme (*Franks et al.*, 1997; *Calveta et al.*, 1998) or others (*Gottschalk et al.*, 2001). Little information is available on such factors like different cultivars or fertilization levels that may influence the stomata function as well. The selected beet varieties are widely used under Hungarian growing conditions. There was significant difference in stomatal resistance of three different sugar beet varieties (*Fig. 1*). Not only their water demands, but the length of their growing season were close to each other. The beet varieties were similar in their appearance (resembling surface size, leaf orientation, soil covering).

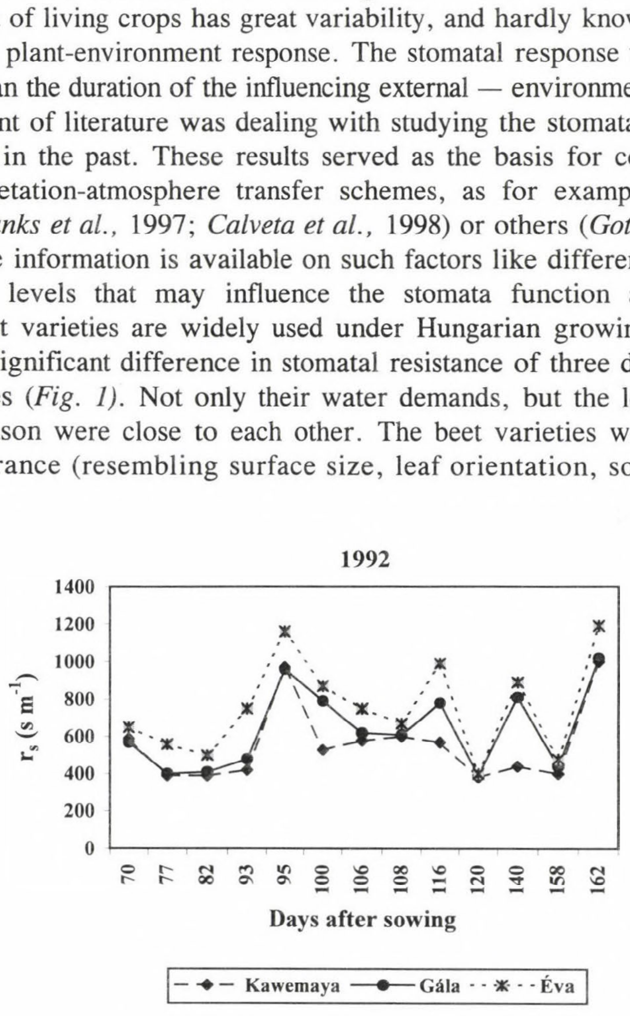

Fig. 1. Influence of the variety on daily average stomatal resistance measured in three different sugar beet cultivars (Kawemaya, Éva, and Gála) on some clear days during the 1992 growing season.

Differences in resistance of altered varieties were manifested during the dry and warm summer of 1992. Results of the stomatal resistance influenced by altered beet varieties are presented by the data measured during this season. The order of resistances measured in different beet cultivars was close to the stomatal resistance published by *Huzulak* and *Matejka* (1992) for sugar beet grown in Slovakia. The lowest resistance of variety Kawemaya produced the highest transpiration water loss and probably the most effective plant cooling at non limited watering in lysimeter growing chambers. At the end of the season, the sugar-yield of Kawemaya exceeded the sugar production of the other two studied sugar beet varieties (Éva, Gála) with increased stomatal resistance. In this experiment the variation in digestio — the sugar content of the roots in percents —, did not agree with the direction of change in the root production. There was an opposite connection between digestio and root yield of different beet treatments. The sugar beet growing is aimed to achieve the highest sugar yield, the third yield component, which is the multiplication of the digestio and the root yield. Although in 1992 the sugar production of Kawemaya was the highest, seasons of altered weather may revive other production results. As the earlier experiments show, the water level has to be in close connection with beet production, where the varieties may have differences in their sensitivity. There is the obligation of the plant growers to bring into harmony the need of the grown beet variety with the terms of the surrounding environment. The stomatal resistance can help the farmer to choose the best beet variety for his own conditions faster, than the other traditional factors used in the characterisation of the plant-water relation.

Experience of stomatal resistance observations for two maize hybrids was not so unambiguous as it was for sugar beet. In maize hybrids significant difference in stomatal resistance of altered hybrids could not manifested in each period of the investigation. In *Table 3* both the effects of nitrogen fertilization (0 and 100 kg N ha^{-1}) and use of two maize hybrids (Norma and MVNK 480) on daily mean stomatal resistance and evapotranspiration sum are totaled. The size of treatment influence considering the hybrids depends on the time of measurement (phenological development). In the very beginning of the growing season, the hybrid MVNK 480 had higher stomatal resistance than that of hybrid Norma. Later on the resistance determined in Norma increased above the resistance values measured in hybrid MVNK like a tendency. Effect of nitrogen on stomatal resistance was independent of the hybrid, and N always decreased significantly the stomatal resistance. Change in daily sums of evapotranspiration caused by treatment effects was in accordance with stomatal resistance alterations (see also in Table 3). Similarly to earlier results of *Howell et al.* (1998), our investigations confirmed that the nitrogen decreases the average stomatal resistance in accordance with the increase in plant water loss, independently of the hybrid.

Table 3. Daily averages of stomatal resistance (r_s, s m^{-1}) and daily sum of evapotranspiration (*ET*, mm day^{-1}) on the same sample days during 1997

Date in 1997	**June 28**		**July 14**		**July 31**		**August 7**		**August 14**		**September 12**	
	r_s	*ET*	r_s	*ET*	r_s	*ET*	r_s	*ET*	r_s	*ET*	r_s	*ET*
	Norma											
Nitrogen	461	5.1	384	5.3	268	5.71	221	5.3	192	6.3	582	5.1
Control	531	4.7	542	3.7	377	4.67	297	3.2	272	4.1	689	4.7
	MVNK 480											
Nitrogen	528	5.3	292	5.3	247	5.31	172	5.9	158	5.6	464	5.3
Control	756	5.0	435	4.4	350	4.15	255	4.0	263	4.2	429	5.0

There is also an endogenous rhythm of pores affecting the stomata movement independently of the current environment. Summarizing the different influencing factors, there are two main systems regulating the stomata, the fluxes of water vapor and CO_2. They need to be controlled and synchronized by pore movements.

3.2.1 Modeling approach of leaf resistance (r_l) evaluation

To test our field stomatal resistance observations, the modified version of Crop Microclimate Simulation Model of *Goudriaan* (1977) was applied. The original model was modified by *Chen* (1984). The basic of assumption of simulation the stomatal resistance is that mass transport processes — both water vapor and CO_2 — occur via stomata, so that the ratio between their resistances is equal to the ratio between their diffusivities. In maize a linear relationship exists between net CO_2 assimilation and inverse leaf resistance (leaf conductance) at constant CO_2 concentration in the sub-stomatal cavity. This connection provided the basis for simulation of leaf resistance, since the net CO_2 assimilation can be deducted precisely from the absorbed visible radiation (*Goudriaan,* 1977). Exceeding the saturation point of CO_2 assimilation (200 J m^{-2} s^{-1} for sunny maize leaves), the leaf resistance approaches the minimum value (*Stieger et al.,* 1977). Rate of net CO_2 assimilation (F, kg CO_2 m^{-2} s^{-1}) was considered empirically by *van Laar* and *Penning de Vries* in *Goudriaan* (1977) as follows:

$$F = (F_m - F_d)[1 - \exp(-R_v c / F_m)] + F_d, \qquad (1)$$

where F_m is the maximum rate of net assimilation (kg CO_2 m^{-2} s^{-1}),
F_d is the dark respiration (kg CO_2 m^{-2} s^{-1}),
R_v is the absorbed visible radiation (per LAI) (J m^{-2} s^{-1}) and
c is the slope of photosynthesis-light response curve at compensation point (kg CO_2 J^{-1}).

At F_m calculation the influences of leaf age and ambient CO_2 concentration were simplified, and their average values were applied. Leaf temperature was considered to depend on ambient air temperature. Dark respiration was at about -0.1 of F_m (*Goudriaan,* 1977). To calculate maize leaf resistance, Eq. (1) can be re-written as follows:

$$F = \frac{1.83\times10^{-6}(C_e - C_r)}{1.66 r_l + 1.32 r_{b,h}} \quad \Rightarrow r_l = \frac{1.83\times10^{-6}(C_e - C_r)}{1.66\,F} - 0.783\, r_{b,H}, \qquad (2)$$

where $r_{b,H}$ is the boundary layer resistance for heat (s m^{-1}),
1.66 is the ratio between diffusivities (for CO_2 and H_2O),
1.83×10^{-6} converts CO_2 concentration into kg CO_2 m^{-2} at 20°C,
C_e is the external CO_2 concentration (ppm),
C_r is assumed as „regulatory" CO_2 concentration (ppm),
1.32 originates from calculation of boundary layer resistance for CO_2.
The r_l was assumed to be equal to resistance measured by the porometer.

When stomatal resistance samples are taken in maize, the porometer chambers have to be placed on the three-thirds of the leaf blades (*Anda,* 2001), where the resistance tends to get close to the average of the whole leaf. In our observations this method markedly decreased the great variability of standard deviation of measured resistance values within one maize leaf.

The maize is an amphystomatous plant, having almost the same number of pores on both sides of the leaf blade. In the early phase of the vegetation period, in May and June, the maize stomatal resistance measured on the upper epidermis was about 30% higher than the resistance of the lower leaf side. It might be the special radiation environment attributed by the open canopy structure in spring that lets the radiation penetrative better into the stand decreasing the stomatal resistance. In fully developed canopy, in July and August, a linear relationship in stomatal resistance between adaxial and abaxial leaf surfaces existed (*Fig. 2*). In most studies the stomatal resistance of maize is only measured in one side of the leaf, assuming that the maize stomatal resistance of the two leaf surfaces are almost the same. In our study difference in resistance between the two leaf sides was registered, mainly at low solar angles, in early morning and late afternoon. There was hardly enough difference in maize stomatal resistance between the two leaf surfaces at high solar elevation.

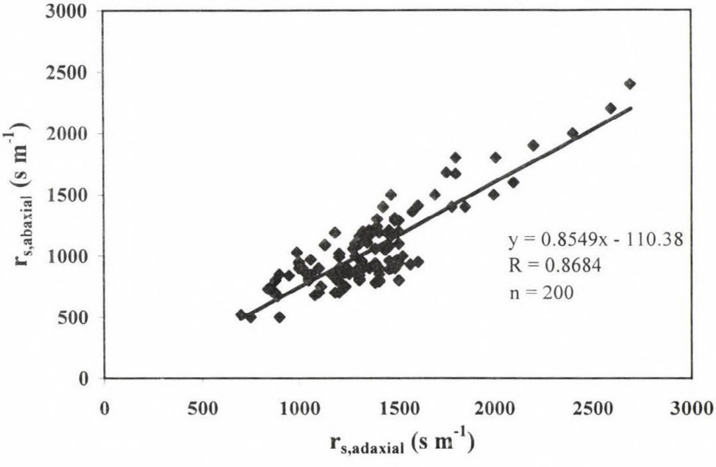

Fig. 2. Relationship in maize stomatal resistance between lower ($r_{s,abaxial}$) and upper ($r_{s,adaxial}$) epidermis. 200 samples (n) were taken at Keszthely, on clear days between 1992 and 1996. Hybrid Norma was applied in investigations.

Earlier results have shown that the influence of illumination exceeds the effects of leaf position or leaf age. We measured the stomatal resistance of totally sunny and shaded leaves separately, and the ratio of sunny (f) and shaded leaf sections were estimated by eyes. The average of the stomatal resistance (r_s) of the whole plant or leaf then will be as follows:

$$r_s = f\, r_{s,sunny} + (1-f)\, r_{s,shaded}, \tag{3}$$

where $r_{s,sunny}$ is the mean stomatal resistance of sunny leaves (s m^{-1}),
$r_{s,shaded}$ is the mean stomatal resistance of shaded leaves (s m^{-1}).

Finally the accuracy of estimation of maize average stomatal resistance was tested by using the model of *Goudriaan* (1977). Diurnal variation of stomatal resistance was simulated using ten soil water potentials from –0.1 to –14.0 bars for a bright summer day in July. At the same time, the lysimeter growing chambers and control field had -0.28 and –4.3 bars soil water potentials, respectively. The soil water potential in the field was measured by neutron probe. The weather of the sample day was excellent for measuring stomatal resistance. In the two watering levels the resistance values differed and had the highest deviation at around solar noon. Average stomatal resistance of plants grown in lysimeters was 21.1% less than the daily mean of the non-watering control treatments.

Independently of water levels, the measured resistance is below the 1:1 lines, which means that simulation produces higher resistance than the measured one, mainly at around solar noon (*Fig. 3*). Simulation of maize stomatal resistance at "Ad libitum" watering gave better result than that of the control. While the difference in daily mean resistance at non limited water supply was 8.1%, at control plants it was 12.0%, in favor of simulated values.

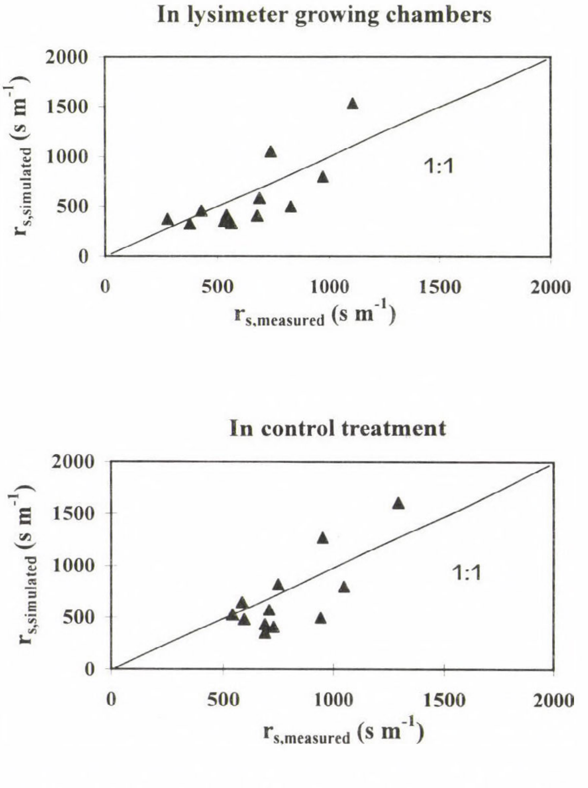

Fig. 3. Results of simulation of leaf resistance, r_l, on July 25, 1996. Samples were taken in hybrid Norma, at Keszthely, using non-limited watering level (lysimeter) and rainfed control stands.

For testing our stomatal resistance estimation, we chose an "ideal" sample day. When the sky was not completely clear, the difference in values of simulated and measured stomatal resistance increased, and the divergence sometimes exceeded even the 40–50%. These results need further clarifying.

3.3 Considerations of the Crop Water Stress Index (CWSI)

Radiation properties of substances — reflecting, absorbing, and transmitting radiation — vary considerably, thus presenting a method for extracting information about the substances. The infrared thermometer receives the reflected radiation from the surfaces in the direction within the field of view of the thermometer. The instrument yields an integrated temperature, which does not interfere with the plant surface.

The amount of reflected radiation (I, W m^{-2}) can be expressed by the Stefan-Boltzmann blackbody law:

$$I = \varepsilon \delta T^4 , \tag{4}$$

where T is the surface temperature (K),
δ is the Stefan-Boltzmann constant (5.674×10^{-8} W m^{-2} K^{-4}),
ε is the emissivity of the substance.

Most infrared thermometers have 8–14 μm wave-band filter, because the peak of blackbody emission at normal temperatures falls within this region, and the water vapor absorption is relatively low here.

Determination of surface temperature is complicated, because the emissivity of different plant surfaces has great variability in time and space. In most investigations emissivity for plant surfaces is adopted from the earliest literature of the subject.

Plant surface or canopy temperature can be a useful parameter in irrigation timing, because meteorological and soil factors indicate when plants may be stressed, as well as plant factors when they are stressed (*Jackson,* 1982). In quantification of plant water supply different indices are widely applied. These stress indices can be grouped into two categories. The combined energy balance-aerodynamic relation (*Penman,* 1948) may be re-written with surface temperature as a function of net radiation and vapor pressure deficit (*Jackson et al.*, 1981). The second category of stress indices comprises the empirical approach of *Idso et al.* (1981).

3.2.1 Theoretical consideration of the CWSI

The basis of assumption is the energy balance of the plants:

$$R_n = H + \lambda E + G , \tag{5}$$

where R_n is the net radiation (W m^{-2}),

H and λE are sensible and latent heat fluxes (W m^{-2}),

λ is the latent heat of vaporization of the water (J kg^{-1} water),

E is the transpiration intensity (J s^{-1} m^{-2}),

G is the heat flux into the ground (W m^{-2}).

Similarly to original literature, later on the soil heat flux is neglected.

Monteith (1973) gave detailed discussion and assumption for the members of Eq. (5). At non-limiting watering level, the plants will transpire at potential rate (potential evapotranspiration, *PET*). When there is not adequate water in the soil, the transpiration decreases below potential rate (actual evapotranspiration, *ET*). A measure of the ratio of actual to potential evapotranspiration should result in an index about the plant water status, the crop water stress index, *CWSI*. After expressing the members of Eq. (5), and solving for λE yields, the well known Penman-Monteith equation for *ET*, in terms of canopy (r_c, s m^{-1}) and aerodynamic (r_a, s m^{-1}) resistances is as follows:

$$\lambda E = \frac{\Delta R_n + \rho c_p \{e_s(T_c) - e\}/r_a}{\Delta + \gamma(1 + r_c/r_a)}, \tag{6}$$

where $e_s(T_c) - e$ is the difference in saturation and actual vapor concentrations of the air (hPa),

γ is the psychrometric constant (hPa K^{-1}),

c_p is the heat capacity of air (J kg^{-1} K^{-1}),

ρ is the air density (kg m^{-3}),

Δ is the slope of saturated vapor pressure-temperature relation (Pa °C^{-1}).

In most cases r_c may also be calculated with measurements of stomatal resistance and the leaf area index.

Producing the ratio of actual (λE for any r_c) and potential (λEp for $r_c = r_{cp}$ measured in lysimeter) evapotranspirations produces:

$$\frac{ET}{PET} = \frac{\Delta + \gamma}{\Delta + \gamma(1 + r_c/r_a)}. \tag{7}$$

Taking into account that plants are going from non-stressed to stress conditions, our index, the *CWSI* will present values from 0 to 1:

$$CWSI = 1 - \frac{ET}{PET} = \frac{\gamma(1 + r_c/r_a) - \gamma^*}{\Delta + \gamma(1 + r_c/r_a)}. \tag{8}$$

The ratio of the resistances can be obtained by re-arranging the energy-balance equation (Eq. 5):

$$\frac{r_c}{r_a}=\frac{\gamma r_a R_n/(\rho c_p)-(T_c-T_a)(\Delta+\gamma)-(e_s(T_c)-e)}{\gamma[(T_c-T_a)-r_a R_n/(\rho c_p)]}, \tag{9}$$

where T_c–T_a is the difference in the plant and air temperatures in °C.

The upper limit at potential transpiration, where $\gamma = \gamma^*$, and r_c is minimal with maximum transpiration:

$$\gamma^* = \gamma(1 + r_{cp}/r_a). \tag{10}$$

The lower limit, where $r_c \Rightarrow \propto$, and $r_{a,H} = r_{a,W} = r_a$ is

$$\Delta = \frac{\lambda E}{H} = \frac{\gamma(r_{a,W}+r_c)}{r_a}, \tag{11}$$

where $r_{a,W}$ is the aerodynamic resistance for water (s m^{-1}). More details are published in the original literature of *Jackson* (1982).

The empirical procedure in stress index determination suggested by *Idso et al.* (1981) seems to be less complicated than the theoretical one. Following the way of *Idso et al.* (1981), two empirical baselines (canopy and air temperature difference, T_c–T_a versus vapor pressure deficit, *VPD*) are proposed, where the lower baseline can be obtained when water level is unlimited. The upper baseline arises at wilting point, when there is no available soil moisture. The actual water supply of the plants is somewhere between these two bounds presenting a new method of producing stress level of plants (*Fig. 4*). Although this latter method seems to be less complicated, it does not account for changes in net radiation or wind speed. Neglecting of these two important environmental factors, the accuracy of stress evaluation of plants declines.

In the course of our 13-year experiment, both approaches were applied. When we used the *Jackson* (1982) method, the r_c/r_a, the Eq. (9) was determined and substituted into Eq. (8) to get the *CWSI*. To calculate the ratio of Eq. (9), net-radiation, canopy and air temperature difference, humidity (vapor concentrations of the air), wind speeds, and plant height (roughness parameter, displacement height) were measured or counted. In empirical consideration the elements of canopy and air temperature difference and the vapor pressure deficit are the only input parameters.

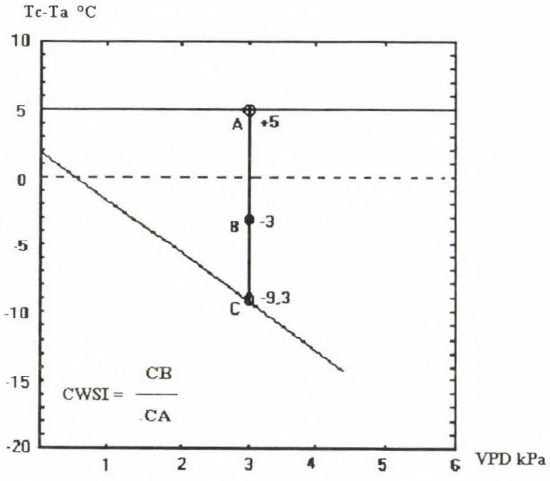

Fig. 4. Relations in plant and air temperature difference (T_c-T_a), vs. vapor pressure deficit (*VPD*). The lower and upper baselines are presented together with an example of how to calculate the *CWSI*. More details are discussed by *Idso et al.* (1981).

3.2.2 Some examples for practical use of CWSI

In our investigations both the empirical and theoretical considerations of *CWSI* determination were applied. When we calculated the index by using the method of *Idso et al.* (1981), the influence of convective heating on calculated stress index was also taken into account. Parallel investigations on wind and stress index variations were carried out, and a linear relationship between the temporary wind speed (u) and *CWSI* has been found. If the wind is blowing, the applied *CWSI* has to be raised with Δy :

$$CWSI' = CWSI + \Delta y, \tag{12}$$

$$\Delta y = 1.11 + 0.27u. \tag{13}$$

More details on the wind-*CWSI* relation are in *Anda* and *Ligetvári* (1993).

Seasonal variation of the *CWSI* depends on the environmental influences, weather, and mainly the amount and distribution of rainfall. The drier the weather of the season, the higher the value of the *CWSI* is. Our first example represents results in seasonal variation of *CWSI* during a humid growing season (*Fig. 5*). In 1997 the summer was cool, the seasonal mean temperature was 0.8°C lower than the climatic norm. Distribution of the seasonal rainfall was not very smooth, in August the precipitation was about 15 mm lower than in most of the years, but in June and July it was 28% higher than the 30-year average. As a result of the excess watering measured in June and July, no irrigation was necessary even in the drier August, during 1997. The tenfold values of *CWSI* never reached the limitation of 2.5 on the rainfed plots. During this humid year, an opposite tendency in *CWSI* appeared in lysimeters, where the excess watering of the chambers and the rainfall together caused increase in stress level of plants. The higher *CWSI* determined for lysimeters derived from the abundant watering of chambers, where the water ousted air from the soil.

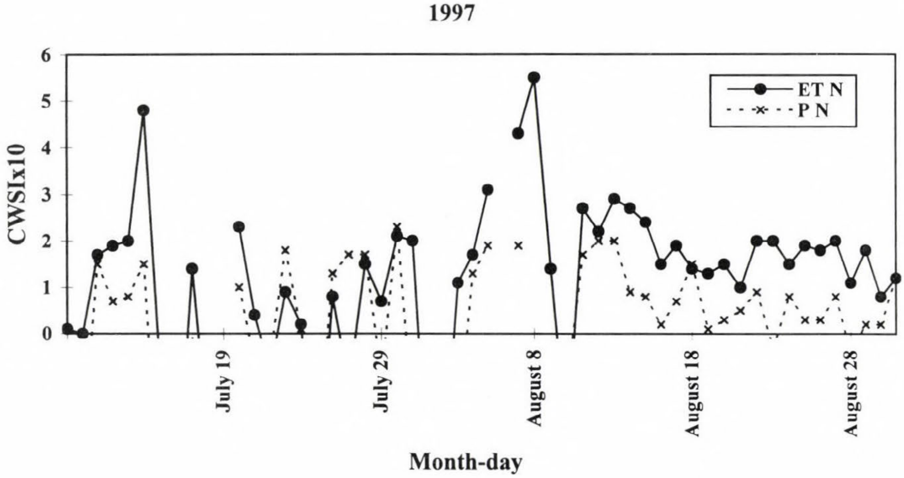

Fig. 5. Seasonal variation of *CWSI* tenfold values of for maize during humid summer of 1997. Continuous line represents indices measured in lysimeter (*ET*), dotted line are for rainfed control (*P*). The abbreviation *N* means nitrogen fertilization.

Traditional and other possibilities in use of *CWSI* are presented by the results determined for 1998. Although the seasonal mean temperature in 1998 was close to the climatic normal, in July and August precipitation was extremely low in the most important period of water stress in maize (*Fig. 6*). In dry summer of 1998 the seasonal mean of *CWSI* decreased by 131% in irrigated treatment comparing to its non-irrigated rainfed plots. According to the *CWSI* concept, the plants needed irrigation three times in the season (see arrows in Fig. 6). The irrigation produced 14.5% significant grain yield increase during the arid 1998.

Results of irrigation trials carried out at Keszthely by using the canopy temperature based *CWSI* are totaled in *Table 4*. Our test plant was the maize, duration of the study was 13 consecutive years between 1989 and 2001. We conclude for maize, that the *CWSI* is a useful tool in scheduling irrigation, but only in arid growing seasons. During humid seasons there is no need for irrigation and for computing the *CWSI*. In semi-humid summers the yield surplus produced by means of irrigation is not enough to cover the extra costs of watering under Hungarian growing circumstances. Other results can be obtained with different crops, mainly with those horticultural plants, which represent higher expense per unit area than the maize. In comparative observations on different types of irrigation timing, the *CWSI* application produced the most economic method with least amount of water and cheapest irrigation cost of all.

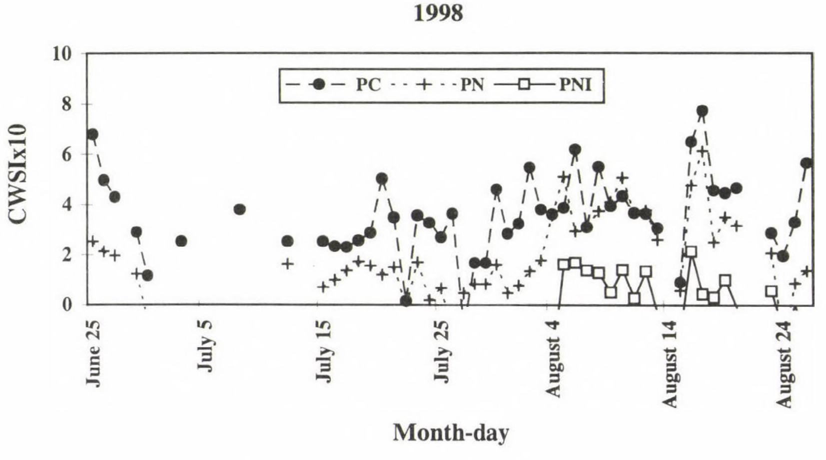

Fig. 6. Changes of tenfold values of *CWSI* in maize during arid 1998 resulted from altered watering (irrigated, *IP;* and non irrigated, *P*, plots) and fertilization levels. *C* and *N* sign the lack of nitrogen and fertilized plots, respectively. The arrows indicate the time of irrigation.

There are other than water problems derived stress factors that may influence the size of the *CWSI*. Effect of these stressors on calculated stress level of maize is demonstrated by artificially simulated N-shortage (Fig. 6). In 1998, the lack of N alone increased the seasonal mean of *CWSI* by 116% in rainfed plots. Surprisingly, the decreased amount of watering together with N limitation hardly increased in the average of *CWSI*. Probably the Liebig minimum-law or the close connection between nutrition and water level of plants might influence the final common stress index. To avoid overflowing the figure, and regarding the similarity of *CWSI* of the treatments without nitrogen, only indices determined for rainfed plots (PC) are presented.

Table 4. Results of the *CWSI* application for maize at Keszthely Agrometeorological Research Station, under different weather conditions between 1989 and 2001

Weather of the season	**Arid**	**Humid**	**Semi-humid**
Number of the seasons	**5**	**3**	**5**
		Deviation from the climatic normal	
Average air temperature (°C)	+	–	+ or –
Seasonal rainfall sum (mm*)	–	+	+ or –
VPD at solar noon (kPa)	2.5–3.0 <	< 1.5	< 2.0
Irrigation necessity	**Yes**	**No**	**Yes**
Irrigation water amount (mm)	100–140	No	< 100
Change in *LAI* (%)	+20 <	No	+8–15
Change in *CWSI* (%)	-25 <	No	(–10) – (–22)
Change in yield (%)	**+10** <	**No**	~ **5**

* Continuous season from April 1 to November 1

Our observations show, that attention should pay on *CWSI* determination, when non water shortage origin stress factors — plant diseases, nutrition or soil problems, etc. — arise. Proper irrigation timing may only be implemented after exclusion of these confusing elements. Neglecting of these influencing factors may lead to application of extra amount of irrigation water.

4. Summary

Field trial was connected to two maize hybrids and three sugar beet cultivars, at Keszthely Agrometeorological Research Station, in the growing seasons between 1989 and 2001. Investigations on two factors influencing the plant-water relation are discussed. The first theme of the study was possibilities in approaching the real stomatal resistance in the field. Theoretical consideration of stomatal resistance among the other counterparts of diffusive resistances was also presented briefly. Widely known, that to determine the value of stomatal resistance under field conditions is very complicated, because of high variability of the factor in both time and space. The modeling approach seemed to be a good assumption in average of leaf stomatal resistance determination. To achieve this goal, the concept of Crop Microclimate Simulation Model of *Goudriaan* (1977) was used. To test the model applicability under Hungarian ecological conditions, maize in lysimeters and rainfed plots was grown. The

accuracy of estimation of mean leaf resistance was 8.1% and 12.0% in rainfed and at non limited watering level, respectively.

The second subject of the paper was the irrigation timing by applying the crop water stress index, the *CWSI* concept. In dry summers the *CWSI* was a useful tool in scheduling irrigation of maize stand. In our trial the watering resulted in about 10% yield surplus during arid seasons. We found that the irrigation with the use of *CWSI* is an economic and environment-friend way even in maize. Attention should be paid for those non water-shortage origin factors, that might influence the calculated value of *CWSI* (plant diseases, fertilization, etc.).

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Energy budget between the atmosphere and the surface in the vegetation period during 1963–1994

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Abstract—The surface energy budget is characterized by the solar radiation, physical state of the soil, vegetation, sensible and latent heat fluxes, their proportion and periodicity. This paper summarizes the results of measurements carried out on black loam soil covered by natural short grass. The measuring point was located at 47°36'N, 21°36'E, 132 m above the sea level. The surface energy budget is determined mainly by the advective phenomena, however, near the surface plays a significant role. Continuous measurements have been taken at the Agrometeorological Observatory of the University of Debrecen during 1963–1994 for studying the components of the energy budget. Results are summarized by means of monthly mean values for the growing season.

Radiative conditions were examined through global radiation, reflectance of radiation, or albedo, and finally, the radiation balance. Relative distribution of daily values of the global radiation (Fig. 1) representatively indicates the radiation climate of the given area. Albedo is provided by the precipitation and annual course of the sun zenith angle (Table 2). Dynamics of the albedo depends on the soil state. Figs. 2 and 3 illustrate how different soils in wet or dry state modify the value of the albedo or the reflectance spectrum. Decrease of the soil moisture increases the reflectance, to different degrees at different wavelengths (Fig. 4). Ratio of R_n/G can be considered as climatic representative value. Table 3 shows the monthly values of this ratio.

The soil heat flux has a yearly cycle, however, it is only a small part of the radiation balance. Thermophysical layers of the soil differs from the mechanical layers because of the changing moisture content (Table 4). The energy circulation of the examined soil in the growing season is represented by the data of Table 5, showing that the soil heat flux gets the maximum in spring, and continuously decreases after July.

On the examined area, the intensity of the sensible heat flux is controlled mainly by the vertical temperature gradient (Table 6), since the wind speed is 1-4 m/s, and the diurnal fluctuation of the momentum flux is small during the growing season.

The biggest part of the incoming radiation is used for evapotranspiration. Potential evapotranspiration provides a good tool for evaluating the latent heat flux. Table 7 presents the values of the net radiation (R_n) and potential evapotranspiration (PE_0), and their ratio. According to the data, PE_0 gives the 70–90 % of the net radiation. Hourly values of the actual evapotranspiration for the 30 years long measuring experiment are summarized in Table 8 by means of monthly mean values, providing a useful basis for evaluating the time variation of the latent heat. Table 9 contains the values and ratios of the net radiation (R_n), potential (ET_0) and actual evapotranspiration (ET_a), which proves that the character of the budget and the fluctuation of its components are controlled by the water supply, as usual. Table 10 gives a tool for comparing the values of each components and studying their mean time course.

Key-words: energy budget compounds, heat balance in the soil, sensible and latent heat, potential and actual evapotranspiration, microadvection.

1. Introduction

Micrometeorological processes can be considered as interactions between the atmosphere and the surface, described by the components of the energy- and mass balances. Detailed examination of these processes was carried out at the Agrometeorological Observatory of the University of Debrecen in the years of 1964–1993. The measured data base was analyzed on different time scales. This paper presents the resulted characteristics, representing the micrometeorological processes near the surface in the growing season. On the basis of these characteristics, similarly to the climatic standards, a micrometeorological reference parameter system was determined (micrometeorological standards, MS). These standards represent the features of the surface layer (boundary layer) and the physical meanings of the widely used climatic constants. On the basis of the MS, the normality and extremity of the individual elements, as well as the interrelations of the microclimatic processes can be estimated. Climatic radiation and heat balance of Hungary were analyzed firstly by *Bacsó* (1959), on the other hand *Antal* (1974, 1982) presented the monthly values of the radiation-, heat-, and water balance components on the Lake Balaton (1931–1960), as well as Lake Fertő from 1970 to 1979.

During the 40 years history of the observatory all meteorological elements have been measured, from which micrometeorological parameters can be derived. As the technology improved, a digital measuring system with 6 s time scale was implemented giving a tool for more detailed analysis. For performing continuous measurements of the main meteorological elements, a 30 m high measuring tower was built with 8 stories. Turbulent exchange of the air was observed by sonic anemometer. The soil moisture was examined to 8 m depth.

This paper summarizes the resulted monthly values of the energy budget of natural surfaces. The experiments were carried out in Hajdúhát, at the eastern part of Hungary (47°36'N, 21°36'E, 132 m above the sea level), on lime-covered black loam (chernozem) with natural vegetation and agricultural crops.

2. Radiation budget of the natural surface

The only energy source of the surface energy budget is the solar radiation. The basic task is to determine the temporal change of the energy reaching the surface and the transform of the energy kept back by the surface. Another important question is the amount of energy stored in the few meters high layer above the surface, and the amount passed to upper layers. To solve this problem we have to analyze the energy budget of the surface.

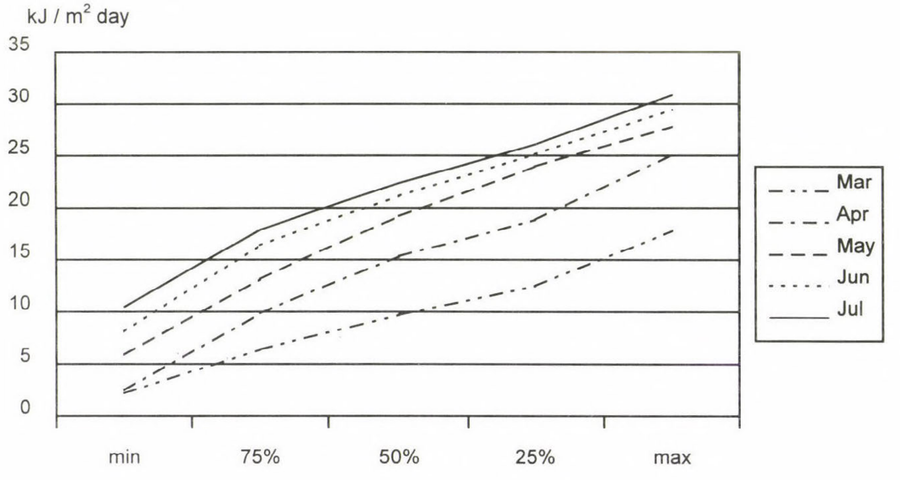
2.1 Global radiation

Our study was carried out on short-cut grassland as reference surface. Energy reaching the surface has a major role, in addition to advection, in the micrometeorological processes. Global radiation was analyzed on the basis of the 30 years long data base providing a useful tool to describe the micrometeorological processes. First the mean, daily, and monthly sums of the global radiation were determined using CM-2 and CM-5-6 type Kipp and Zonen radiation measuring instrument. *Fig. 1* shows the empirical distribution of the daily sums of global radiation in monthly figures. The data of this figure are very similar to the results of detailed measurements analyzed by *Takács* (1958) and *Major* (1976). It has to be emphasized that — especially in spring and autumn — the real distribution is a bit different, since the lenght of daylight modifies the value of the daily sum. Autocorrelation analysis of the daily sums indicates 8–10 days long significant parts, especially in the summer half-year, which is identical with the results of autocorrelation analysis of anticyclonic situations. The maximum of the empirical distribution of daily sums means the radiation sum of a sunny, cloudless day with minimum extinction.

Our study examined the relationship between the global radiation and sunshine duration. It was pointed out that the relation between the daily sum of the radiation and the number of sunny hours is not linear, so the Angström-type formulas overestimate the global radiation when the sunshine duration is low or high, and underestimate in the case of medium number of sunny hours (*Szász*, 1968). To correct this error, we worked out an empirical formula for middle geographical latitudes:

$$G = V\frac{1.163 C\omega G_{\max}}{2}, \tag{1}$$

where G_{max} is the intensity of the global radiation belonging to the solar zenith angle of the given day, C is a multiplier belonging to the relative sunshine duration ($C \leq 1.0$), ω is the duration of daytime in minutes, V is the attenuation coefficient (V=0.88–1.1). The applicability and accuracy of this formula were evaluated by many researchers (*Bohne* and *Klingenbert,* 1977). We founded that in the summer half-year, this method estimated the daily sums with 4.3–5.6% error. Comparative analysis of *Hinzpeter* (verbal note) confirmed this estimation. *Table 1* contains the quotient of the measured and calculated values of monthly sums. Correlation coefficient between the measured and calculated values is r=0.97–0.99.



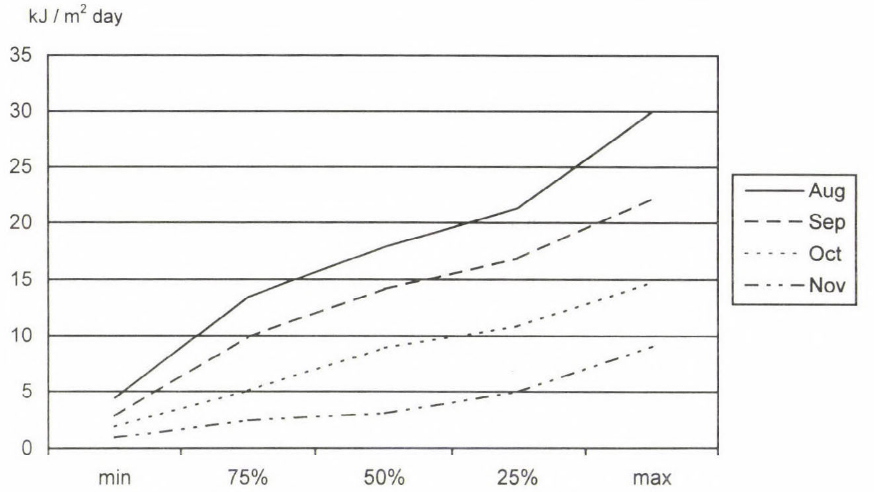


Fig. 1. Relative frequency of daily sums of global radiation in the months of the growing season.

Table 1. Mean values of the attenuation coefficient, *V*

Month	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
V	1.10	1.07	1.12	1.05	0.96	0.95	0.98	0.93	1.00	0.97	0.98	1.06

2.2 Reflection from the surface

Certain part of the radiation reaching the surface is reflected by the ground. The amount of the reflected radiation, relative to the incoming radiation, changes in wide range because of the heterogeneity of the surface. Values of the albedo in Hungary were examined by *Dobosi* (1961) and *Dávid et al.* (1990) providing a general overview, but it is not detailed enough for a particular situation. Our long time data series shows that on the experimental place the moisture content modified the albedo (*Table 2*). Amplitude in the annual course of the albedo is determined by the duration of snow-cover in winter and the precipitation amount in summer. In the vegetation period, the albedo is determined mainly by the vegetation-cover, which depends on the quality and water supply of the soil. Water depletion in the soil — independently from the soil type — increases the albedo. The amount of the reflected radiation is determined by the ratio of the leaf area and the area of the surface, the so called leaf area index. The reflectivity of the surface without vegetation widely varies, it is determined by the volume quotient of the clay in the soil and the soil moisture. The measured spectral distribution of reflection for three main soil structures is shown in *Fig. 2*. Results suggest the conclusion that the reflection increases with raising clay fraction. Another general phenomenon is that the reflectivity of the soil increases with growing wavelengths, independently from the clay- and moisture content. This finding is true for soil structures with few organic matters.

Table 2. Monthly mean values of the albedo of wet and dry soils (%)

Month	Albedo		
	Mean	Dry	Wet
April	16	17	14
May	18	20	16
June	20	23	17
July	22	26	16
August	23	28	15
September	21	24	13
October	18	20	12

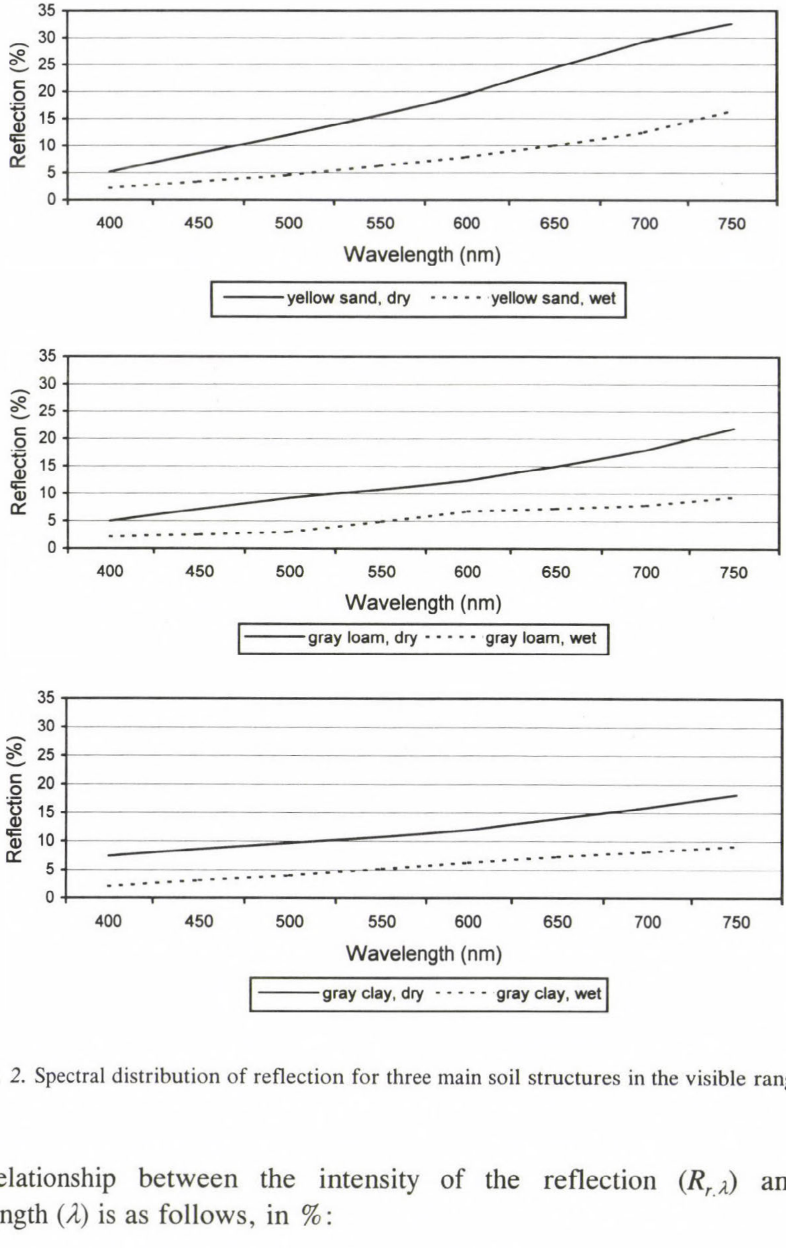


Fig. 2. Spectral distribution of reflection for three main soil structures in the visible range.

Relationship between the intensity of the reflection ($R_{r.\lambda}$) and the wavelength (λ) is as follows, in %:

$$R_{r.\lambda} = 5278 + 0.022\ \lambda, \tag{2}$$

where r is the correlation coefficient, r=0.986. The high value of the correlation coefficient indicates a nearly linear relation in the 400–1000 nm range. Bigger clay fraction results in closer linearity. The distribution function can be approached by a parabolic form as well, which makes the formula more accurate, r=0.985–0.999. Finally, the general formulas for the main soil structures are as follows, in %:

Sand: $R_{r.\lambda} = 5.4 \times 10^{-5} \lambda^2 + 0.124\,\lambda - 37.1$,
Loam: $R_{r.\lambda} = -3.5 \times 10^{-5} \lambda^2 + 0.091\,\lambda - 27.6$,
Clay: $R_{r.\lambda} = -0.6 \times 10^{-5} \lambda^2 + 0.036\,\lambda - 11.3$.

The effect of the moisture content becomes evident in the slope of the linear function fitted into the spectrum. Reflection of sand and clay in wet and dry conditions is illustrated in *Fig. 3.* The interval between sand and clay is valid for loam soils.

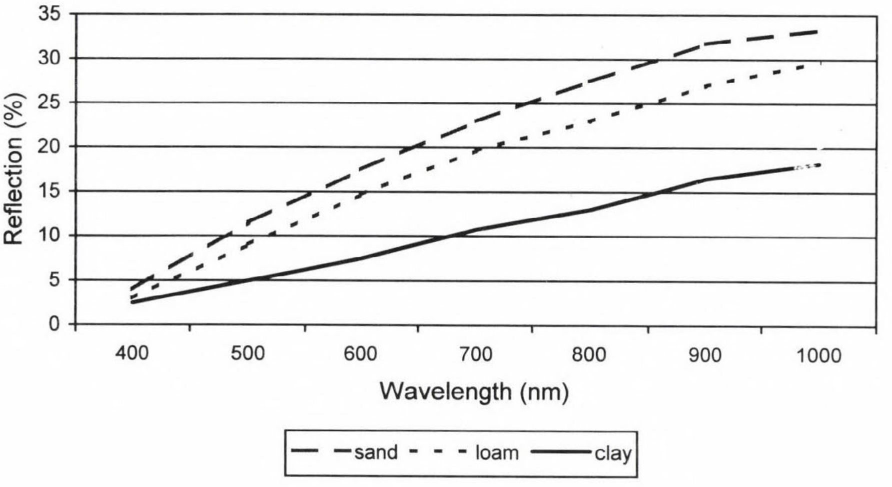


Fig. 3. Reflection of the three main soil structures in the 400–1000 nm range.

Whilst the spectral reflection of the soil linearly increases in the visible range, the reflection of the vegetation is quite different: it is very small in the visible range, then raises to 40–60% in the 750–2500 nm range.

2.3 Effective long wave radiation of the surface

The effective long wave radiation emitted by the surface has the following features:

(1) The long wave radiation is lower at night than in the daytime during the whole year, because of the local climatic temperature, air humidity, and cloudiness conditions.
(2) The minimum of the long wave radiation occurs at dawn, and often there is a secondary minimum late at night.
(3) The main minimum occurs at dawn in the winter half-year and in the evening in summer and autumn. Thermal stratification of the near-surface air layer plays an important role in the temporal variation of the minimum.
(4) The long wave radiation has its maximum early in the afternoon except the summer months, when the maximum occurs late in the morning because of convective cloud formation.

The long wave radiation strongly depends on the temperature and humidity profiles at night. On the basis of profile measurements we pointed out that there are strong effects modifying the vertical profiles of both temperature and humidity. Measurements, up to 10 m height with lifting technique, indicate horizontal exchange processes, especially late at night. As the inversion is getting stronger at dawn, the exchange processes cease (*Szász*, 1964b). These effects significantly influence the degree and diurnal variation of the emitted radiation. The lowest mean value of the long wave radiation is not below –25 W/m^2. The lowest values occur in December at dawn, because at this time the temperature of the surface is the lowest and the stratus clouds become stable. The mean of the highest effective long wave radiation is above 70 W/m^2, and it can be developed only at the biggest zenith angles in summer.

2.4 The net radiation

The relation between the net- and global radiation is an important question in the study of the micrometeorological processes. The quotient of the net- and global radiation (R_n/R_g) is essential to know, since often we can not calculate the net radiation for lack of necessary data. In that case the net radiation can be calculated from global radiation data, in MJ/m^2 month:

$$R_n = a\, R_g, \tag{3}$$

where the value of a varies between 0.3–0.6 according to Hungarian authors. *Table 3* presents the published monthly mean values of the a factor. This data well represent the effects determining the net radiation, first of all the effects of cloudiness. Both the mean values and different frequency values can be considered as important parameters of radiation climatology, noting that they are not valid under other climatic conditions. The values presented by this paper are valid only for the experimental area.

Table 3. Ratio of the monthly mean values of the net radiation and global radiation (a)

Author	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct
Bacsó (1959)	0.37	0.42	0.51	0.54	0.59	0.54	0.35	0.22
Major and *Tárkányi* (1969)	0.33	0.44	0.49	0.50	0.49	0.47	0.39	0.26
Dávid et al. (1993)	0.28	0.46	0.51	0.52	0.50	0.49	0.47	0.40
Szász (1999)	0.35	0.46	0.52	0.53	0.54	0.50	0.42	032

The intervals, where the values of the net radiation fluctuate, is not known. For this reason, we calculated the relative distribution of the global radiation and the mean values of the global radiation and net radiation for the days of the summer half-year. It can be concluded that the net radiation on the individual days in the summer months is very variable (*Fig. 4*). Nevertheless, the variability of daily net radiation is smaller than that of daily global radiation.

The curve of the empirical distribution of global radiation has a strong right hand side asymmetry, therefore, days with big radiation energy occur with 15–25% frequency in the months of the summer half-year. Effect of the changing amount of middle layer clouds causes the wide interval of the left hand, downsloping side of the frequency distribution curve. At the same time, in the case of almost total cloudiness, the frequency values steeply decrease. Relative distribution of the daily net radiation — as it seems from data of Fig. 4 — is considerably shifted to the lower energy ranges. It is worth to emphasize, that frequency distribution of the net radiation is closely symmetrical, and the distribution curve is forced into a much tighter energy range. This phenomenon is caused by the local climatic conditions, therefore, it is representative only for our data base, and can not be generalized. Because of the essential differences of the two distribution curves, the calculation of the net radiation from the global radiation data, using the a factors given in Table 3, is only a rough estimation. On the other hand, the differences between the calculated and measured daily global radiation data can be considerably big, caused by the very different weather situations. For these reasons, the net radiation was determined by means of measurements.

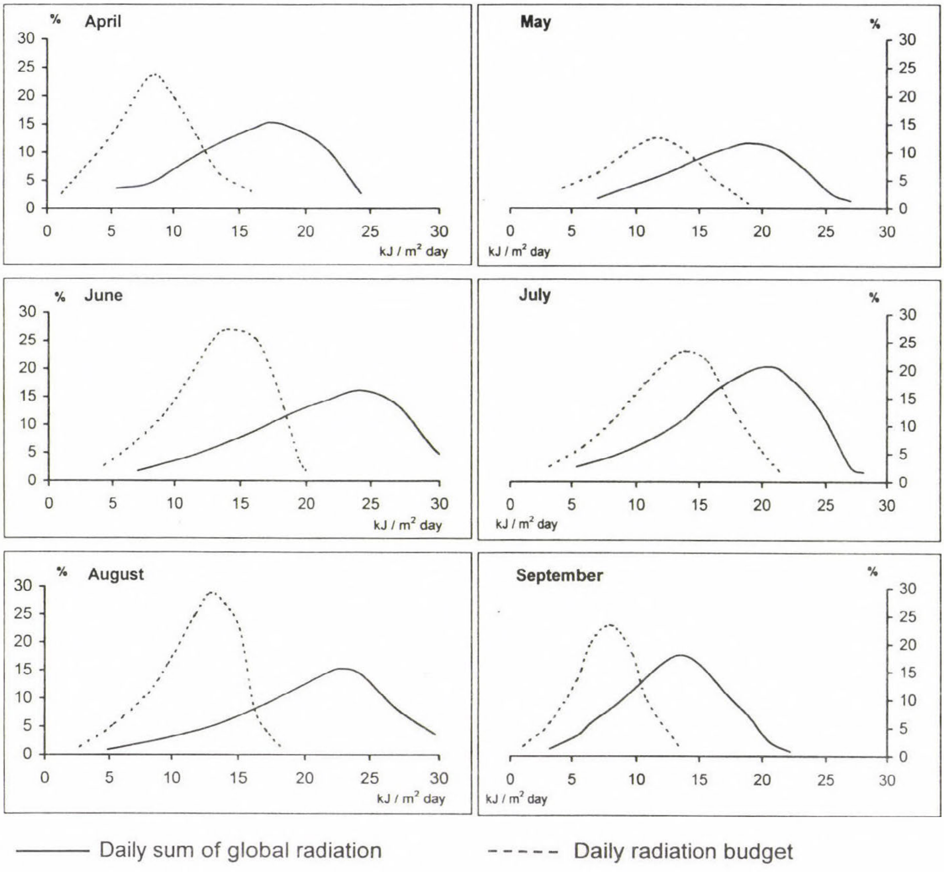


Fig. 4. Frequency distribution of the daily values of global radiation and net radiation.

3. Heat balance in the soil

Heat exchange in the root zone and soil temperature are very important agroecological factors. The soil stores and transports both energy and water. Therefore, there is a close relationship between the water- and heat exchange, where the water content of the individual soil layers deterministically influences the heat exchange.

Analysis of the soil heat flux was based on the Fourier's equation of heat conduction. We determined the most important factors evoking the soil temperature changes of energy gain and loss.

Determination of the soil heat transport is a difficult task, as the heat fluxes not only by molecular conduction, but infiltration of water in the soil.

The amplitude of the diurnal variation of the soil temperature can be derived by Fourier analysis, but only in case of enough effecting factors. Two or three factors are used in practice. Analyzing the weighed spectral distribution of the frequency, we found that the generally used calculation method is only an approaching estimation, because much more factors are needed. The molecular heat conduction can be determined by instrumentally (*Szász*, 1964b) providing more accurate results. The width of the frequency distribution of the molecular heat conductivity coefficient is 0.3–1.6 W/m^2 K. The large standard deviation of the coefficient is caused only by the moisture fluctuation. Determination of the heat capacity and conductivity was a much easier task. A representative data base is available about these features, calculated by widely used methods (*Szász*, 1970).

One of the most important thermophysical parameters describing the soil heat flux is the decline in amplitude (D) of the diurnal variation of the soil temperature with depth. The value of D can be calculated by the formula $D=(2k/\omega)^{0.5}$ (*Rose*, 1966), where k is the temperature conductivity and ω is the length of the period. There are numerous data measured at our experimental station about the temporal variation of thermal conductivity, molecular heat conductivity (λ), and heat capacity of the soil (*Szász*, 1970, 1988; *Molnár*, 1977).

The heat balance in loess soil can be described by soil temperature and moisture data measured by a so-called 'soil heat-flow transducer' instrument during ten years. On the basis of these data, we determined the phase shift of diurnal variation of heat conductivity and temperature, as well as the values of D. Data of each layers were examined and compared to each other, concluding that the soil is thermally anisotropic (*Table 4*). Searching for the reasons of the differences, above all one can think on the differences in the soil moisture profiles. The thermal anisotropy can be very high, mainly in the soil layer near the surface, nevertheless, there is a significant ratio-gradient in the deeper layers of the examined soil, because at the measuring place the soil water table situates in depth of 10–12 m, from where the water income, raised by capillarity as well as upstreamed by water potential, is very small. It means that the moisture content of the upper soil layers at our experimental station is controlled by the weather only.

Annual variation of thermophisycal parameters of the soil is peculiar. Minimal values of the annual variation of the *heat capacity* (2.1 J/cm^3 K) occur in the second part of summer caused by summer soil moisture depletion. Maximal values are consequences of snow melting. Annual variation of *heat conductivity* also has only one peak, as it is the function mainly of soil moisture. The *amplitude of temperature* is controlled by the moisture content, too. The diurnal variation of the temperature has a minimum at the end of

June, beginning of July (0.9 m), since the soil is dried up in the following period and the heat conductivity decreases. In spite of the previous variations, the annual variation of *temperature conductivity* can be described by a curve with two peaks. As both the big and small soil moisture contents decrease value of temperature conductivity, its minimum evolves at the time of minimal and maximal soil moisture, while its maximum develops at the beginning of soil moisture depletion and in the period of filling up with precipitation water.

Table 4. Ratios of temperature conductivities in the successive soil layers

Soil layer z/z_{+1}	**Temperature conductivity** kz/kz_{+1}
2–5/5–10 cm	0.31
5–10/10–20 cm	1.08
10–20/20–50 cm	0.57
50–100/100–150 cm	1.00
100–150/150–200 cm	0.71

Table 5. Values determining the energy budget of the loam soil in different months

(G_M is the daily balance of the heat flux, G_n is the total amount of energy, G^+ is the downward energy, G^- is the daily amount of energy conducted to the surface, G^+/G^- is the ratio of the upward and downward conducted energy amounts, and E_a is the duration of energy uptake)

Month	G_M	G_n	G^+	G^-	G^+/G^-	**Unit**	E_a
April	561.0	4264.4	2412.7	164.5	14.7	KJ/m^2 day	10^{h}15'
May	641.1	5365.3	3003.2	2362.1	1.3	KJ/m^2 day	10^{h}30'
June	573.6	6386.1	3458.8	2885.1	1.2	KJ/m^2 day	12^{h}00'
July	253.1	7027.2	3640.1	3387.1	1.1	KJ/m^2 day	11^{h}10'
August	-341.7	6592.7	3125.5	3467.2	0.9	KJ/m^2 day	9^{h}50'
September	-164.5	4171.6	2003.6	2168.1	0.9	KJ/m^2 day	8^{h}10'
October	-801.4	4378.3	1796.9	2598.3	0.7	KJ/m^2 day	7^{h}50'
November	-615.8	1577.5	480.9	1096.7	0.4	KJ/m^2 day	6^{h}10'

Table 5 shows an overview of monthly, layer-by-layer soil temperature values in the 200 cm layer. On the basis of the daily balance of soil heat flux, temporal variation of soil temperature can be explained. Table 5 presents the daily average values of the heat balance for different months. The presented data suggests that the soil heat flux plays more important role in the energy

budget mainly in spring and autumn, however, its part is only few percents of the net radiation. In July and August, when the sign of the balance changes, daily balance has a very small share in the energy budget. These values are 4–5% by monthly means, the amount of energy reaching the surface from deep layers compared to the net radiation is significant only in October. The soil heat flux is 12–18 percent of the net radiation, and the main part of it is transferred to the air in the afternoon.

4. Sensible and latent heat balances

As it was mentioned, the soil heat flux shares the sum of the net radiation only in a few percents, the main part of the net radiation is used for maintaining the latent and sensible heat fluxes. Ratio of the sensible and latent heat fluxes is climatically characteristic value, therefore, we analyze these two processes separately.

4.1 The sensible heat flux

Vertical and temporal variations of the temperature, caused by the transform of sensible heat energy are important processes in agroecological point of view. The sensible heat depends on two factors: horizontal advection and the amount of the heat energy transmitted by the surface. It is certain, that only a small part of the sensible heat, transmitted to the air from the surface, remains in the near surface layer, but even this energy causes significant air temperature changes mainly in vertical direction. Approaching the surface, the temperature extremes significantly increase, which can be expressed by the difference between the minimum and maximum temperature values. To prove this fact, *Table 6* presents some measured data. Thermal stratification of the surface air periodically changes during the day in the summer half-year almost without exceptions, day-time lability is changed by night-time stability. The maximal lability is developed at noon, while the maximal stability is observed late in the evening. Cyclic variation of the equilibrium determines the direction of the sensible heat flux in the surface air. It is both theoretical and practical problem, that at the time of the transformation of equilibrium, the inflexion point of the vertical temperature profile falls to the measured height domain, therefore, the profile has an extreme value in this layer. In these cases, the estimation of the heat flux is very difficult, very often impossible. Unfortunately, one of the biggest the disadvantages of the calculations based on gradient method is that the error in the heat flux intensity value is the biggest in the morning and evening when the inversion develops and disappears. Mainly this is the explanation for the facts, that the vertical temperature difference is small between the two above mentioned layers and the dispersion among the individual days varies in wide range, explained by the different weather features. In the morning, the air temperature gradients change without disturbations, except for the disturbing effects of the clouds caused by the morning lability in the summer half-year. The units and amount of the gradients are expressed by the means of the adiabatic gradient. Thus, the value of the effectively measured gradient is $\Delta T = 0.0098 \times 10^2/\Delta z$.

Table 6. Daily temperature range on clear days in different heights above surface (°C)

Height (cm)	April	July	September
800	12.4	13.2	12.1
400	12.6	13.7	12.5
200	13.0	14.1	13.0
50	13.6	14.9	13.9

Note: The sample is based on differences of minimum and maximum values of 60 days

The mentioned facts provide only an informative summary, but it can be generalized since the measured data base is extensive. On the basis of the variation of vertical temperature differences, we can conclude that the vertical heat flux varies in very wide range. The change of the sensible heat flux depends mainly on the soil moisture content. When the soil moisture content is big, the heat energy is used for evaporation. In case of smaller soil moisture content, the vertical heat flux is bigger. It can be proved by the next data, calculated for different moisture contents (different values of minimal water capacity in percents) of the upper soil layer:

$$0.8\ W K_{\min}:\ H/(R_n - G):\ 15 - 25,$$
$$0.5\ W K_{\min}:\ H/(R_n - G):\ 25 - 40,$$
$$0.3\ W K_{\min}:\ H/(R_n - G):\ 40 - 90,$$

where H is the vertical heat flux, R_n is the net radiation, and G is the heat flux in the soil. These data strengthen the relation between the soil moisture and the vertical heat flux, which is strongest in the summer months. This relation declines with the decrease of global radiation and daily net radiation.

There is a close relation between the vertical sensible heat flux and the turbulent exchange, since the transmitted heat amount is determined by the vertical air temperature gradient as well as the intensity of the exchange processes.

As the wind speed can has very variable values, the sensible heat flux has big fluctuation during the day. Therefore, in addition to soil moisture, the wind speed is an important factor which influences the heat exchange in the air.

Vertical sensible heat fluxes have medium order of magnitude between the magnitude of soil heat exchange and latent heat flux in the energy balance of the surface. Although both have great variability, the distribution curves of the values partly cover each other. The reason of great variability is that the heat exchange with the air depends on many factors. We have point out the heat capacity of the upper soil layers, since the warming and cooling of light soils cause intense day-time lability, while late in the evening, when inversion is developed, a strong temporal stability is evolved, which weakens for the night hours. According to our statistical analyses, there is a strong correlation between the daily fluctuation of the soil temperature in the upper 10 cm layer and the vertical heat flux in summer (r = 0.89; n = 95).

Regarding monthly sums of sensible heat fluxes, the average heat amount passed to the air is generally small in spring because of the moderate amount of radiation energy and the big soil moisture content. In summer, the sensible heat increases with the soil moisture depletion and the increase of the energy reaching the surface. In autumn a smaller part of the radiation balance is shared to the sensible heat amount again, because the exchange processes slow down.

4.2 The latent heat flux

In agroecological point of view, the latent heat flux has great importance, because this is the energy equivalent of the evapotranspiration and dew. Diurnal changes provide useful information in the analysis of daily dynamics in 1–2 hours range. First of all, our purpose was to determine the evaporation loss. For the calculation we used the concept of *Penman* (1956) and *Monteith* (1973). The value of the evaporation has an upper limit, since, supposing that there is no advection, this value should not be greater than the water equivalent of the net radiation. Agroecological researches were carried out on both small and big resolution of temporal variations. Terminology connected to evapotranspiration is as follows:

- *potential* evaporation and -evapotranspiration (ET_0, in W m^{-2}, MJ/m^2 units), or its water equivalent;
- *effective* or *actual* evaporation and -evapotranspiration (ET_a, in W/m^2, MJ m^{-2} units), or its water equivalent;
- *generally*: the energy used for evaporation of water (evaporation+ transpiration – LE, in W/m^2 time, kJ/m^2 time units), or its water equivalent.

The *Penman-Monteith* method, modified by *Monteith* (1973, 1975), was used in our study:

$$LE = \frac{\Delta(R_n - G) - \rho c_p (e_0 - e)/r_a}{\Delta + \gamma(1 + r_s/r_a)}, \tag{4}$$

where Δ is the value of saturation vapor pressure for 1°C (de_0/dT), R_n is the net radiation, G is the soil heat flux, ρ is the air density, c_p is the specific heat of the air at constant pressure, e_0 is the saturation vapor pressure, e is the real vapor pressure, γ is the psychrometric constant, r_a is the aerodynamic resistance, and r_s is the plant diffusion resistance.

If r_s refers to limited soil moisture exchange, the formula results in the actual evapotranspiration, otherwise, if the water supply in the root zone is near the field capacity the evapotranspiration is potential.

4.2.1 Potential evapotranspiration (ET_0)

In the first period of our investigation we worked out an empirical formula for estimating the potential evapotranspiration (*Szász*, 1973). It can be used even if daily average air temperature and daily average relative humidity data are available only. The value of the correlation coefficient between the measured potential evapotranspiration and the mentioned air temperature and humidity values is $R = 0.94$ on days without precipitation. On the basis of this close correlation, we could develop a formula which was exact enough for our investigations. The basic relation is: $ET_0 = Y = \mathrm{f}(t, R, u)$, where Y is a function of t (temperature in °C), R (relative saturation), and u (wind speed in m/s). Limiting condition of this relation is: $ET_0 = Y = 0$, if $R = 1$. In this case, the function which satisfies the condition and fits the relation is:

$$ET_0 = A(t)\,(1 - R)^{\mathrm{b}} \;\rightarrow\; \log ET_0 = \log A(t) + b\log(1 - R). \tag{5}$$

This relation can be considered linear with good approximation, the slope of the fitting line is independent of the temperature. The value of b is monotonously increasing with the increasing temperature. Accepting the values below 0°C, the reduced determination coefficient of the function fitted to the common b is $r^2 = 0.989$, the common b value is $b = 0.66$. Temperature has a major role in these relations, its computational formula is

$$A = a\,(t - t_0)^2. \tag{6}$$

Extracting square root of both sides and completing the operation we get: $a = 0.005356$ and $t_0 = 20.89°\mathrm{C} \cong 21°\mathrm{C}$. As final result we get the approaching equation for the potential evapotranspiration, in the function of temperature and relative humidity, as follows:

$$ET_0 = 0.005356\,(t + 21)^2\,(1 - R)^{2/3}\,. \tag{7}$$

There is no linear relation among the daily average values of potential evapotranspiration, temperature, and relative humidity. In this way, this relation differs from the generally used formulas.

Comparing the calculated potential evapotranspiration with the values measured by evaporation pans, it can be concluded that the evaporation of pans is subject to significant microadvective effect. We examined the energy exchange of the evaporation pans in details calculating the values of microadvection. There was no significant difference between the measured potential evapotranspiration, corrected with the microadvection effects, and the calculated values. This empirical relation was controlled by the Penman-Monteith equation, which proved a good agreement in the case of considering microadvective effects. A so-called oasis-effect, a microadvective phenomenon develops in radiation conditions and the summer half-year, which can be both negative and positive, because in the case of a rapid cooling the soil looses its storaged heat sooner than the water. On the basis of numerous comparative measurements and computations we determined the numerical value of the microadvective effect, which was integrated into our evapotranspiration estimating model in the following form, in mm/day:

$$ET_0 = c\,[0.005356\,(t + 21)^2\,(1 - R)^{2/3}]. \tag{8}$$

The effect of microadvection increasing the evapotranspiration, $c \geq 1$, which can reach the 1.5 mm/day value in summer. This is 35% of the physically real value.

It is worth to compare the water equivalent of net radiation with the daily average values of evapotranspiration (*Table 7*). The ratios presented by the table indicate that the energy remaining for the heat exchange between the soil and air is 26% at the end of spring, about 20% in summer, and 4–10% at the end of the growing season. Furthermore, in the average of the eight months, 85% of the net radiation is shared by the potential evapotranspiration, on our experimental area from March to November. In the agrometeorology and the energy- and water management practice the potential evapotranspiration can be considered as a frame number providing an information about the water shortage and surplus, and their degree. This is the reason for considering potential evapotranspiration as an essential factor in the investigations of water balance. We note, that in irrigation practice the value of potential evapotranspiration is considered as the upper limit of the water consumptive use of plant stands.

Table 7. Water equivalent and ratio of net radiation and potential evapotranspiration (R_n is net radiation, PE_0 is potential evapotranspiration in mm/day)

	March	April	May	June	July	August	September	October
R_n	1.43	3.34	4.26	4.49	4.89	3.82	2.23	1.27
PE_0	1.34	2.47	3.19	3.67	3.94	3.61	2.01	1.22
PE_0/R_n	0.93	0.74	0.74	0.82	0.81	0.94	0.90	0.96

4.2.2 *Actual evapotranspiration* (ET_a)

The energy used for the actual evapotranspiration and water consumption of the vegetation is the component taking the biggest part of the heat balance over Hungarian climatic conditions. This part is estimated as 65% in yearly course by the researchers, nevertheless, its daily value can vary between 35–95% depending on weather and vegetation conditions.

For a short period (hours) the actual evapotranspiration is calculated with the aerodynamic method, Bowen-ratio, or Penman-Monteith equation. In the case of calculating daily averages, the following simplified formula can give a reasonable result:

$$ET_a = ET_0 \frac{\Delta + \gamma}{\Delta + \gamma(1 + r_s \; r_a)}, \tag{9}$$

where ET_0 is the value computed by Eq. (8), Δ is the slope of the saturation vapor pressure curves in hPa °C unit, γ is the psychrometric constant, and r_s/r_a is the ratio of diffusive resistances of the vegetation and air. When estimating the diffusive resistance of the vegetation, r_s, values concerning xerophilous plants are suitable to use ($r_s > 6$ m/s). r_a is the reciprocal of the exchange coefficient.

Aerodinamical resistance is the function of wind speed and the height of vegetation. In the case of 5–10 cm high grass, its approaching values at different wind speeds are as follows:

u	1	2	3	4	5	m/s
r_a	125	75	60	45	40	s/m

Stomatal resistance can be exactly estimated by *Jarvis'* method (1976). Knowing the daily amounts of evapotranspiration and the mass of the vegetation, we can separate the values of the evaporation from bare soil and the transpiration. In the case of short-cut grass as reference surface, the transpiration is 40–60%, the remaining 60–40% is for the evaporation from bare soil. In the case of agricultural crops, the evaporation from bare soil is 10–30%.

Table 8 shows the hourly average values of evapotranspiration as the average of 21 years of the 1964–84 period in W/m^2. The hourly average values provide good overview of the dynamics of diurnal variation of evapotranspiration and the amount of used heat energy. According to our calculations, the biggest value of the evapotranspiration is about 0.3–0.4 mm/hour measured in July, but in the most cases the amount of the loss of water is much smaller. The table contains the values of the night-time negative evapotranspiration, i.e., the amount of energy released during dew formation. It is remarkable, that maximum amount of dew is formed at dawn, just before the temperature minimum occurs. Evapotranspiration data given in the table provide a useful overview of every hours of the vegetation period.

On the basis of the data of *Table 9*, diurnal variation of the water equivalents of the potential and actual evapotranspiration and net radiation can be determined. Values in the first column of the table ($(R_n - G)/L$) expresses the amounts of energy, which can be used for evaporation, i.e., the possible maximum values of the daily potential evapotranspiration. Second column contains the measured average values of evapotranspiration, third column shows the ET_a values. An important climatic information is that how large part is the ET_a of the energy usable for evaporation. There is a characteristical dynamics in the changes: in the month following the winter the evaporation becomes large, but for April and May, the water supply significantly decreases, so the energy used for evaporation is only the 40% of the potential amount, then at the time of precipitation maximum, this part is more than 50%, which is continuously increasing in autumn in consequence of atmospheric evapority.

5. Energy exchange between natural surface and surface air

To describe the interaction between the surface and the atmosphere is possible with analyzing the amount of energy reaching the natural surface and the transport processes of the surface air. The simplest and most certain method of this analysis is to determine the energy balance components. Monthly values can be determined on the basis of the few decades long study. *Table 10*

Table 8. Hourly average values of the actual evapotranspiration (in W/m^2), measured at Debrecen, during 1964–1984

Hour	March	April	May	June	July	August	September	October	November
0–1	-18.0	-20.2	-26.9	-37.7	-51.8	-10.8	-6.4	-3.4	-7.1
1–2	-15.7	-19.2	-26.9	-42.4	-42.1	-10.8	-7.2	-28.7	-8.2
2–3	-16.9	-24.5	-40.8	-54.9	-43.3	-8.8	-4.6	-21.2	-7.5
3–4	-12.9	-32.7	-45.6	-35.7	-13.5	-6.0	-17.7	-22.2	-3.9
4–5	-18.8	-11.9	-13.7	-54.2	-6.7	-35.0	-23.8	-30.9	-3.8
5–6	-25.5	-25.7	-0.2	4.2	-5.1	-74.4	-44.0	-43.4	-3.6
6–7	-20.7	16.9	50.1	70.5	85.5	7.6	-20.8	-47.9	-5.7
7–8	10.6	62.8	105.4	112.2	122.7	37.4	20.4	-0.6	-13.9
8–9	65.8	117.8	145.5	144.6	145.1	70.5	52.0	54.0	6.5
9–10	90.1	152.5	191.8	197.5	200.0	106.5	98.6	78.8	30.3
10–11	109.2	182.3	220.4	223.2	232.6	136.2	131.0	99.4	46.4
11–12	129.5	193.9	231.6	229.4	264.0	151.3	149.7	107.1	62.7
12–13	131.8	180.9	203.4	213.4	256.1	163.8	147.3	82.2	52.3
13–14	126.4	175.6	199.2	211.6	245.9	160.6	122.1	80.8	48.1
14–15	103.4	144.4	171.1	174.0	222.4	144.4	118.6	55.9	19.6
15–16	54.5	92.4	139.1	191.0	182.4	111.7	86.1	-20.1	-11.4
16–17	3.9	46.5	127.2	149.6	125.2	87.6	23.5	-24.1	-7.5
17–18	-10.3	8.2	56.4	116.4	36.3	-1.0	-27.0	-16.0	-4.3
18–19	-13.4	-7.1	30.7	8.4	-54.9	-22.7	-4.0	-7.3	-3.9
19–20	-16.4	-10.8	-11.1	-14.0	-17.9	-9.1	-71.0	-7.7	-17.0
20–21	-16.6	-23.5	-8.8	-21.2	-14.9	-8.3	-4.2	-10.2	-17.6
21–22	-16.3	-17.2	-9.4	-21.1	-17.4	-8.3	-8.0	-9.1	-15.3
22–23	-14.4	-18.0	-20.8	-27.4	-17.3	-11.0	-5.4	-4.0	-14.2
23–24	-16.1	-16.6	-25.7	-31.0	-20.1	-11.5	-7.8	-8.2	-14.9

Table 9. Average water equivalents of the potential and actual evapotranspiration and net radiation for the different months

Month	$(R_n\text{-}G)/L$ mm/day	ET_0 mm/day	ET_a mm/day	ET_a/ET_0	$(ET_a/R_n\text{-}G)/L$
March	1.43	1.34	1.19	0.89	0.83
April	3.34	2.47	1.42	0.57	0.43
May	4.26	3.91	1.78	0.46	0.42
June	4.49	3.67	2.32	0.63	0.52
July	4.89	3.94	2.47	0.62	0.51
August	3.82	3.61	2.32	0.64	0.61
September	2.23	2.01	1.38	0.67	0.62
October	1.27	1.22	0.81	0.66	0.64

Table 10. Average values of the components of energy exchange between the surface and the boundary layer, measured at the Agrometeorological Observatory of Debrecen, during 1961–1993

Components		Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov
R_n	(month)	105.6	202.5	298.4	322.9	347.2	291.0	190.8	102.4	-2.1
	(day)	3.41	6.75	9.63	10.76	11.2	9.39	6.36	3.30	-0.1
G	(month)	3.6	-9.7	-15.0	-20.1	-18.8	-11.1	2.81	11.1	14.5
	(day)	0.12	-0.31	-0.50	-0.65	-0.61	-0.36	0.09	0.39	0.48
H	(month)	-17.1	-86.0	-145	-128	-136	-99.5	-89.2	-51.5	22.0
	(day)	-0.55	-3.24	-4.66	-4.27	-4.39	-3.21	-2.97	-1.66	0.73
LE	(month)	-92.1	-107	-139	-175	-192	-180	-104	-63.6	-34.4
	(day)	-2.97	-3.56	-4.48	-5.82	-6.20	-5.82	-3.48	-2.05	-1.11
H/R_n		0.16	0.42	0.48	0.40	0.39	0.34	0.47	0.50	-0.10
LE/R_n		0.87	0.52	0.47	0.54	0.55	0.62	0.55	0.62	-
Temperature °C/day		4.8	10.3	17.5	18.4	20.0	19.2	15.2	9.7	3.8
Vapor pressure mbar		6.5	8.5	12.2	15.2	16.2	16.4	13.3	9.7	6.9
Equivalent temperature °C/day		16.3	21.6	30.5	38.0	40.5	41.0	33.3	24.3	14.2
Change of enthalpy kJ/kg day		2.57	.37	5.99	4.52	2.89	-1.34	-5.53	-4.65	-4.31

summarizes the components of the radiation balance and energy exchange in the boundary layer. Theoretically we assume that the components of energy balance are free from advection. This is an acceptable consideration, as the examined area is a plain surface with natural vegetation, and the agricultural crops are farther away. For this reason, the assumption, that surface air represents the friction (planetary boundary) layer suitable for the physical features of the surface, and that the components of the energy balance express the natural conditions of the soil-plants-atmosphere system, seems real.

The first row of Table 10 contains the 30 years averages of the monthly and daily values of the energy balance components. Considering that the examinations were carried out for agroecological reasons, the late autumn and winter months are not represented. Changes in the monthly values of the energy balance components represent the amount of energy transferred by the surface. Ratios of the different components well represent the local climate characteristics of this area.

Temporal change of the components is a useful information, but analyzing their ratios is more interesting. Thus, the share of the sensible and latent heat in the net radiation can be considered as the most important climatic factor. Ratios H/R_n as well as LE/R_n showed in Table 10 are values determined by daily measurements, and they are representative climatic parameters of the observational area. Latent energy shares a big part in the net radiation in the beginning of spring, and then at the end of spring it halved. It alludes to the relatively small amount of soil moisture storaged from the winter precipitation, and that it significantly decreases until June in the upper soil layers. At our experimental area the precipitation maximum occurs in June, therefore, the ratio of latent energy increases and becomes stable depending on the precipitation amount, moreover, it can exceed 60% at the end of summer. This later ratio does not mean the increase of the evaporation, but that the decrease of the evaporation is slower than that of the net radiation.

In the average of the three summer months, H/R_n is 51% and LE/R_n is 57%, thus a near equilibrium state is formed between the two ratios. This is a reason for Hungary to be classed among the climatic categories with medium dry summer. It has to be mentioned that there are significant deviations from the average values in the individual years, caused by different prevailing weather conditions of different years.

In the years with Mediterranean weather, when there is a precipitation maximum in winter and a minimum in summer, the mentioned ratio is about 30–40%, and what is more, in extremely dry years it can be considerably decreased below 30%. If cool, oceanic effects influence the year, the value of the LE/R_n ratio can reach 75–80%, thus the fluctuation around the average value is 20%. This ratio and its large fluctuation around the average value are good evidences of the inclination of the Hungarian climate to have extreme situations. This feature can not be proved by comparison of climatic elements only. Therefore, the study of the ratios provides a useful tool for not only to describe the features of the surface air layer, but it gives exact, considerable data for physical description of the climate.

Summarizing the results, we can state that the systematic, 30 years long research of the energy budget have made numerous important conclusions possible on the fields of climate, micrometeorology, and agroecology.

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Hydroecological risks of crop production in Hungary

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Abstract—It is through hydroecology and the soil moisture regime that water accomplishes her major planetary role: the support of plant and animal life. Within the evolution of terrestrial life forms, nature has elaborated a sophisticated and efficient mechanism for the harmonization of the regulatory processes of the soil-water-vegetation system within the broad framework of the climate-lithosphere-biosphere interaction. Man is now intervening into this regulatory mechanism at a rapidly increasing scale and without a clear and reliable knowledge of the long term consequences of his interventions. The expectable significant and rapid change of climate due to the increasing greenhouse effect is a good, and perhaps the last, opportunity for learning how to restrain our technological and economic capabilities according to the limits set by the carrying capacity of the biosphere. As assessments reveal, the conventional step-by-step analytical research must frequently be supplemented by and combined with management-oriented empirical solutions.

Key-words: soil moisture, evapotranspiration, water balance, irrigation water demand, deforestation, salinization.

1. Hydroecology and crop production

Crop production — lying at the heart of agriculture, food supply, and population growth — can be looked upon in two ways: as the greatest success story of mankind removig the ecological barrier of population size and holding the promise of Paradise for all, or as the largest human intervention into the nature's functioning lying at the hearth of the present environmental crisis and hiding the danger of a catastrophic falldown of our civilization. Both visions are valid and might become our common future pending on motivations, attitudes, and decisions of individuals and governments during the years and decades to come.

History offers ample examples for both, the successful harmonization of crop production with landscape ecology, and an almost total and irreversible deterioration of soil resources and natural ecosystems caused mostly by *complete deforestation* followed by intensive cultivation. Total deforestation can increase the rate of *soil erosion* by a factor of 100 or even 1000 (from 1–20 t km^{-2} in the forests to 1000–2000 t in cultivated areas) as documented, e.g., by depositions of the Neolithic agriculture in Europe, as well as the late Bronze age and Roman agriculture in Southern Europe and Britain.

A slower and more subtle form of deforestation-caused soil degradation has been experienced in the flatlands of Hungary. Due to basin-type hydrogeological structure, large parts of this region have an *upwards directed* pressure gradient and *groundwater supply*. The original deep rooted forests "pumped" this slow flow into the atmosphere without its high salt content reaching the soils and the surface. After deforestations (which were accelerated during the wars of the 16th and 17th centuries), the salt content of the upwards moving groundwater reaches the surface and degrades the formerly good quality soils by salinization (*Fig. 1*).

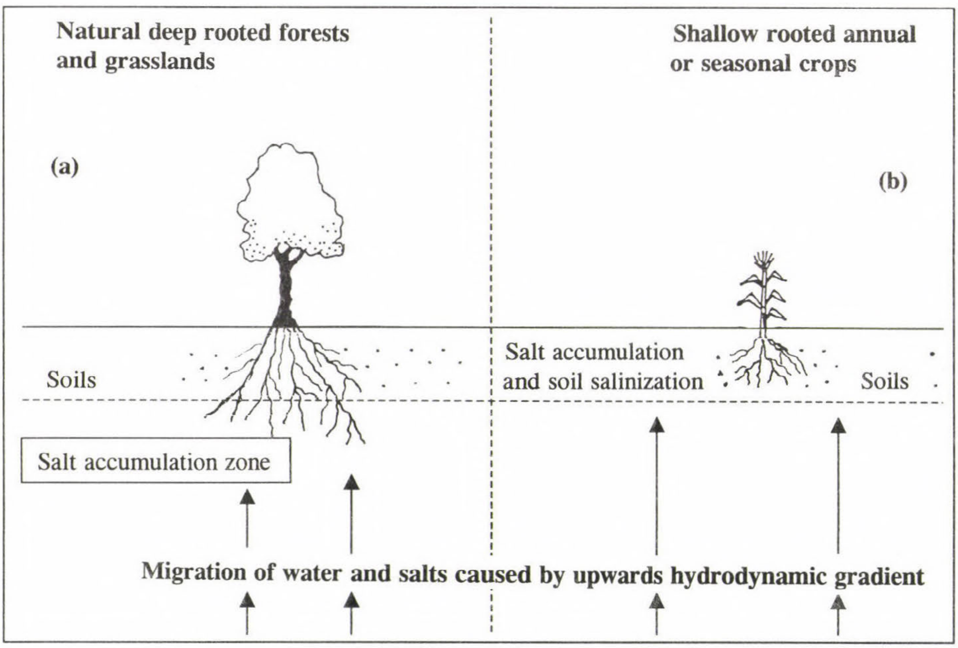


Fig. 1. Salt migration driven by upwards hydrodynamic gradient does not reach the soils in case of deep rooted natural vegetation (panel a), but it reaches the soil and causes salinization in case of shallow rooted crops (panel b).

Within the terrestrial water cycle, hydroecology is structured according to sets of smaller and larger fluvial basins. This integrating role of water manifests itself as the "short", "long", and "deep" branches of the subsurface water cycle. The near-surface short and long branches are dominated by climate and can be classified according to the major climatic and vegetational zones. The deep branch of the scheme is governed by the forces and processes of groundwater hydrodynamics, and it gains ecological and economic significance particularly in case of large fluvial basins, such as the Great Plain of Hungary within the Carpathian Basin. Because of the great variability of geohydrological structure within the Great Plain, the deep branches of the water cycle reach large proportions and are of decisive significance for plant ecology and crop production (*Szesztay,* 1991).

2. *Analytical planning for integrated water management*

In a historical perspective, it is the second time that water is becoming a critical factor of human evolution, the first being of age of fluvial civilizations within which water was a limiting constraint in the form of irrigated agriculture and inland transportation. These ancient constraints were eliminated during the subsequent ages by technological innovations revolutionizing the realms of economic and social development. Today water is becoming a critical factor of human evolution through its integrating roles within and among the technosphere, the biosphere, and the geosphere. The constraints and hazards stemming from these roles can now be overcome through integrated planning of analytical nature supported by improved societal and political water awareness.

The integrated roles of water have a twofold societal implication: within the individual managerial activities, the high level of interrelatedness requires readiness for co-operation and compromises in order to reach mutual advantages, as well as to avoid unnecessary confrontations; in the domain of national and regional policies, the holistic nature of the hydrological cycles urges the legal and economic recognition of water as a fundamental survival factor, as well as the establishment of an informational infrastructure as the basis for expressing and validating the broad public interest within water-related planning and decision-making.

The last decades are frequently labeled as the beginning of the age of information and knowledge. In terms of data handling and modeling capabilities, all technological and computational barriers seem to be overcome in constructing electronic information systems of any scale and complexity. The real bottlenecks of the implementation of societal water awareness seem to derive from the fragmented nature of our water-related concepts and information. The specific form and content of a holistic knowledge on water is, of course, site and time specific.

The following questions might give, however, a tentative and general indication of a holistic approach (*Orlóci,* 1978; *Orlóci et al.*, 1985):

- What are the socially significant roles (the valued components) of the country's hydrological processes? How can these components be analytically described and quantitatively assessed?
- What are the natural factors and human activities within and outside the country area that have a significant impact on those valued hydrological components?
- What is the role of water in the utilization of land and other natural resources?
- What is the extent of human interventions in the country's water balance and water quality processes, and what are the critical levels of eventual future interventions?
- What are the major social and economic demands for water and water-related services, and what are the major alternatives for satisfying, or influencing these demands?
- What are the major possibilities and modalities for the protection and development of the country's water resources?

The answers to these and other similar questions require (*Fig. 2*):

(i) a set of basic studies on water-related implications of major policy factors as a point of departure;
(ii) analytical planning and impact assessment exploring future scenarios with alternative options of human responses and their environmental consequences, and
(iii) series of carefully designed publications disseminating findings and messages of these studies in easily accessible form and language for the various groups of addressees.

Within the basic studies, water-related implication of present and expectable future technologies deserve particular attention. These preparatory studies should also include a systematic performance analysis of existing water systems and services (*OVK*, 1984).

It should be emphasized that the above advocated ecological orientation of hydrology in no way undermines its relevance and efficiency in supporting water management and hydraulic engineering activities. On the contrary, it shall enhance such applications in several important ways. First of all, improved understanding and modeling of water balance and water quality dynamics offers more reliable and more widely applicable methods for the design and

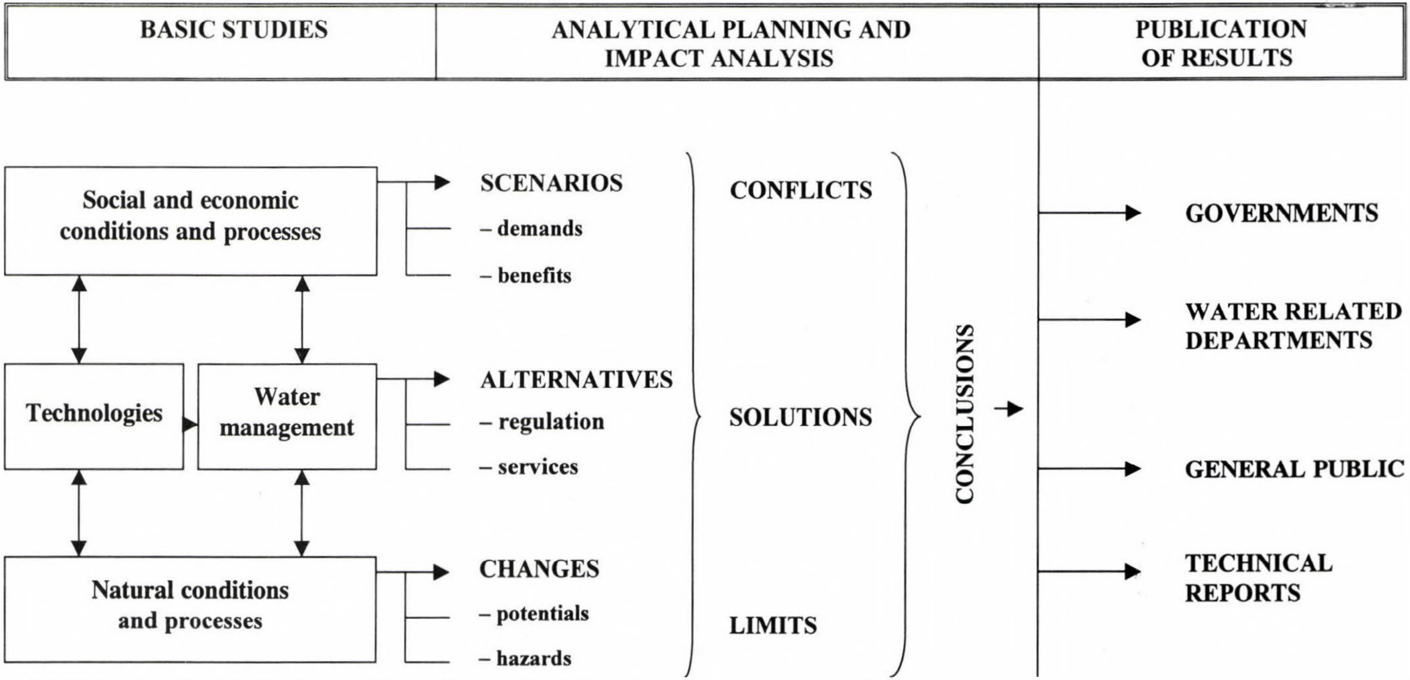


Fig. 2. Formation of programmes and policies for integrated water management (*Orlóci,* 1978; *Orlóci et al.,* 1985).

operation of engineering projects. In addition, its integration with landscape ecology, crop production sciences, and other aspects of land management enables hydrology to significantly broaden the scope and improve the efficiency of water management through a shift from "runoff and riverflow management" towards "precipitation management". Such a shift in concept and practice is particularly important for Hungary and other alluvial flatlands of the temperate and semi-arid zone. Under such conditions precipitation is rather unequally partitioned between infiltration and runoff, and groundwater flow is a locally diversified component of the water cycle.

Integration with landcape ecology also opens new perspectives in the assessment of human impacts on water balance and runoff formation. As indicated by data of *Fig. 3a*, relatively slight but opposite changes in potential and actual evapotranspiration (by about 8 to 10%) caused in Hungary by a decrease in forest covered lands from about 40% in the 10th century to about 12% in 1926, gave rise to a substantial change in local climate (as measured by climatic water deficiency) and led to a drastic increase in seasonal runoff (from about 10 mm to 25–30 mm). Water-balance dynamics of the land surface and the unsaturated zone is particularly sensitive to deforestation (and other land use changes) within the Great Plain region. This is reflected by historical changes in the percentages of wetland regions, as well as by the capacity of drainage systems needed for maintaining intensive crop production (*Fig. 3b*).

3. Risk analysis of the soil moisture regime

A water-centric investigation of theoretical and practical questions arising in the broader context of crop production and hydroecology usually leads sooner or later to a comprehensive analysis of the water balance processes taking place in the root zone (*Fig. 4*). Understanding and quantifications resulting from this analysis should be firm and detailed enough to allow the formulation of a *water balance simulation model and procedure* as a basic methological tool for answering the great number and variety of questions asked by agriculturalists or land use planners.

From point of view of time horizon and practical orientation, these questions tend to fall into two groups. One directed towards longer perspectives of preserving and enhancing the region's soil and other ecological resources and linked usually to the realms of land allocation, regional planning, and nature preservation. The other, generally more privileged under conditions of market economies, considers the root zone primarily through its potentials for crop production and is guided by the principles of achieving an optimal compromise between maximizing incomes and minimizing risks.

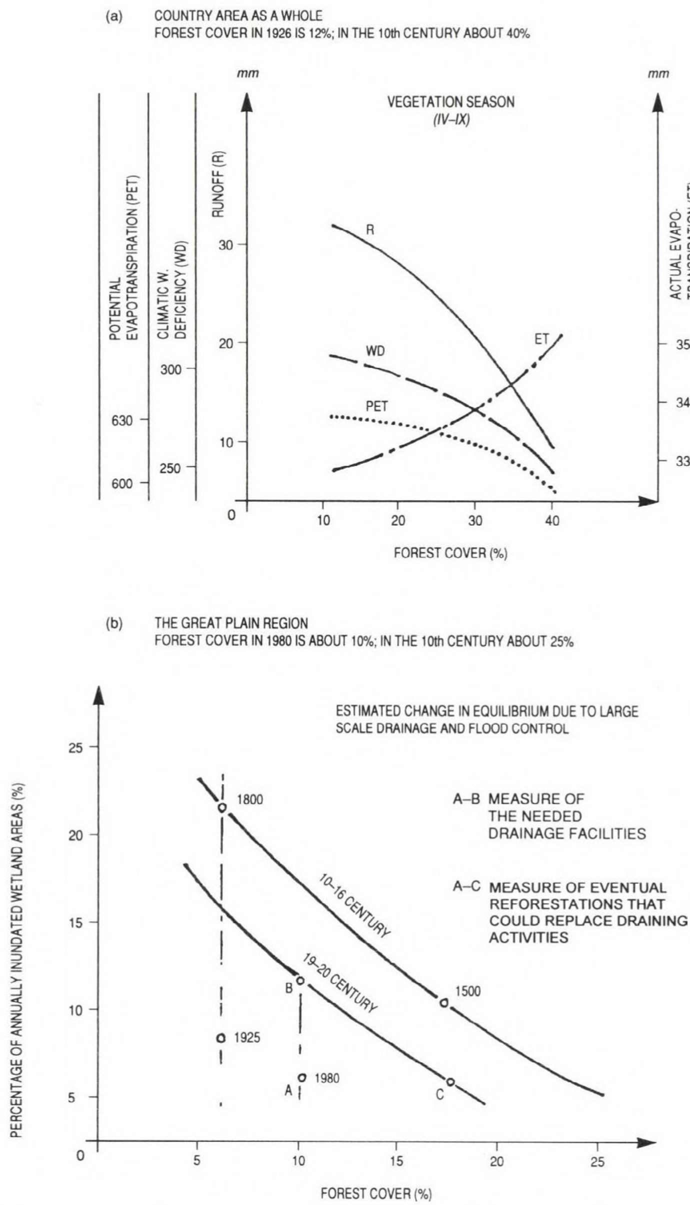


Fig. 3. Tentative relationships describing the effect of deforestations on water balance regime of Hungary (*Orlóci et al.*, 1985). Based on consultant report of *Antal* (1983) and *Pachner* (1981) for *OVK* (1984).

Regarding the results and the outputs of the investigation, the former groups of initiatives usually lead to an ordered set of recommended land use options. The latter groups of inquiries, linked closely to the motivations and decision problems of the farmer and the agricultural communities, require an ordered set of probability functions describing the expectations for crop-endangering shortages and excesses of water within the root zone, first in terms of hydrological events, then translating these into losses in crop yields — such as elaborated by *Orlóci* and *Pintér* (1981) as part of the 1984 "National Water Plan" (a broadly formulated Water Policy Analysis) for Hungary (*Fig. 5*).

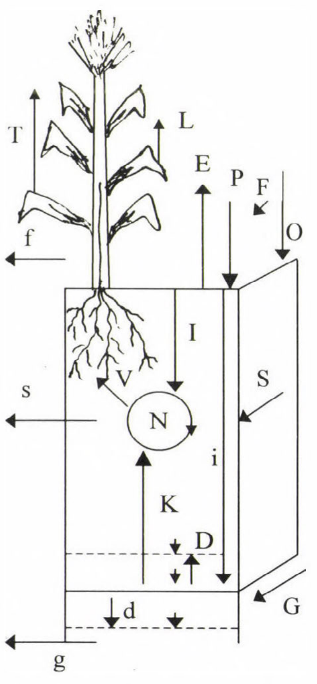
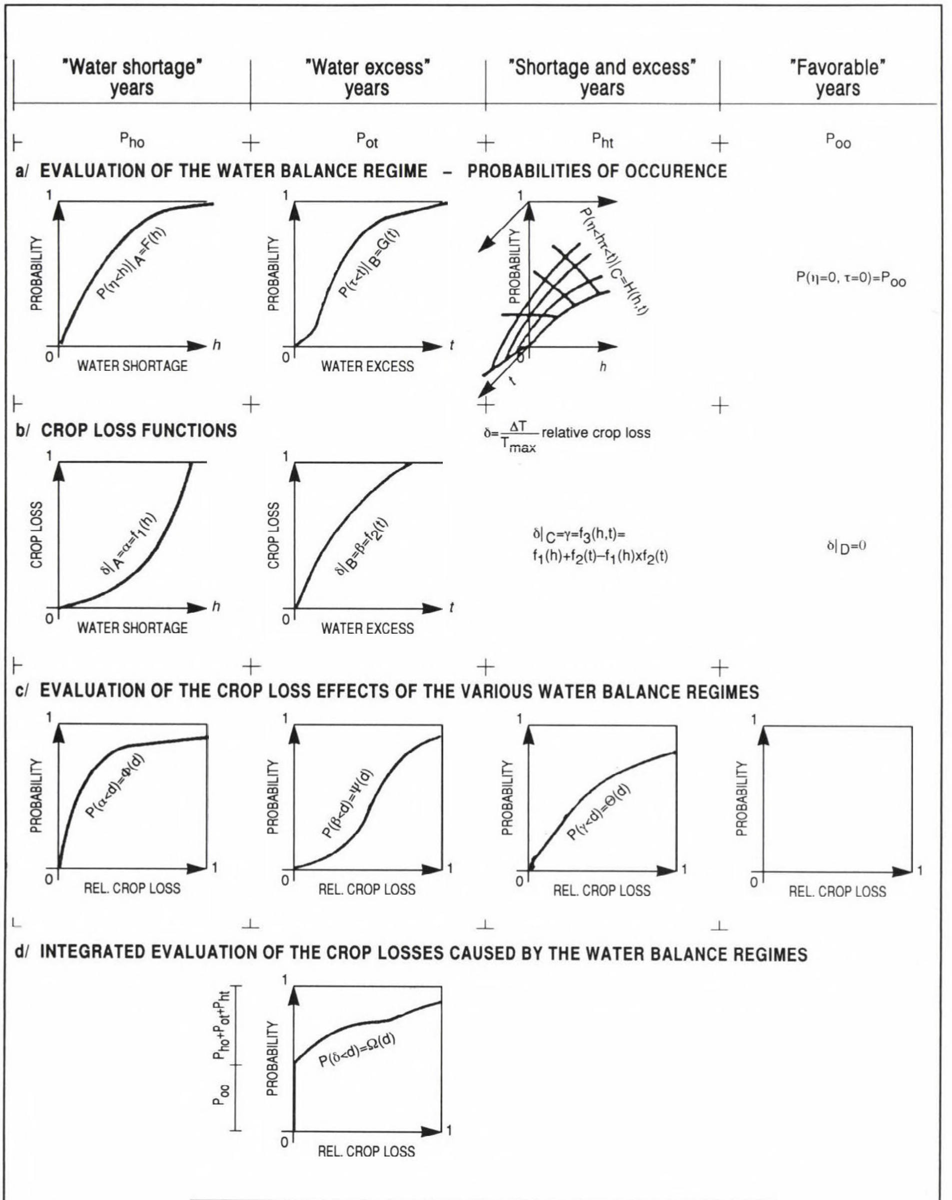


Fig. 4. Water balance components of the root zone (*Várallyai,* 1987).
(P=precipitation and irrigation; F and f=inflow and outflow through surface runoff; i=infiltration reaching the groundwater; I=infiltration; N=soil moisture storage; V=soil moisture available for plants; T=transpiration; E=soil evaporation; L=wet canopy evaporation; S and s=inflow and outflow through soil moisture flow; G and g= inflow and outflow through groundwater flow; D=rising groundwater level; K=capillary rise of groundwater; d=falling groundwater level)

In fact, these two directions of questioning and problem solving are closely interlinked not only through the water balance simulation as a major tool of methology, but also in their practical outcomes as no meaningful land use options can be formulated without prior knowledge of the risks and consequences of various potential crop production technologies.



$$p(\delta<d)=\Omega(d)=P_{ho}\Phi(d)+P_{ot}\Psi(d)+P_{ht}\Theta(d)+P_{oo}$$

Fig. 5. Evaluation of the crop loss impact of the water balance conditions (*Orlóci* and *Pintér,* 1981).

The needs for and the possibilities of water balance simulation stems from the experimentally and theoretically well documented notion that for given climate and soil conditions plant communities tend to evolve according to well defined margins of deficiencies and excesses of heat and water, whereby the extent and frequency of these limiting extremes vary from species to species, and also according to the phenological phases of plant development.

In order to identify these margins and apply them in predictions or planning, one has to quantify the annual and seasonal variability of heat and water availabilities in probabilistic terms. Such quantifications can be made relatively easily for radiation and heat factors on the basis of regular meteorological observations of sunshine and air temperature. For quantifying critical deficiencies and excesses of soil moisture, the required continuing and detailed observational data are, however, not available on a regular basis (*Antal,* 1988). Modeling and simulation of the soil moisture balance seems to be, therefore, the only feasible solution to this effect (*Orlóci et al.,* 1993).

With regard to possibilities and ways of quantifying evapotranspiration as the key element of the soil moisture balance equation, there seem to be important differences between the water-stress controlled and the heat controlled strategy of plant growth. When soil moisture is unlimited and the system maximizes biomass production under given radiation and heat constraints, the use of water in the form of evapotranspiration can be well approximated for a given plant species and given phase of phenological evolution through atmospheric data alone. The possibility of such approximation is probably a reflection of the coincidence of factors and mechanism guiding stomatal regulation of evapotranspiration with those promoting maximal biomass productivity.

The usual simulation procedure is based on a step by step compilation of the soil moisture balance equation of the root zone for continuous ten days periods. Wherever the soil moisture content drops below the level of unrestricted water availability — which varies widely according to crop type, soil conditions, and the phases of plant growth —, the root zone is hypothetically filled up to field capacity. The amount of irrigation water needed to this effect is registered in the simulation procedure as a measure of natural (climatic) water deficiency event, whereby the accumulated amounts of the hypothetical irrigations provide a quantitative measure of the dry weather events corresponding to the actual local conditions.

For the purposes of the National Water Policy Study for Hungary, some 2000 computerized soil moisture simulations have been made for the country's farmlands according to 13 crop varieties, 7 soil types, 4 categories of groundwater depth, and 3 levels of farming technologies.

The simulations are based on the 1928–1977 data of 23 meteorological stations characterizing climatic differences. The major results were summarized in the form of maps quantifying deficiencies of the soil moisture balance according to major crop types for selected characteristic periods. Water shortages belonging to risk values other than those indicated on the maps, as well as the impact of cultivation intensity (crop yield), soil properties, and groundwater depth are specified by tables or graphs obtainable directly from the simulation results.

The above outlined simulation is not a full description of the natural soil moisture regime. It does not clarify how stomatal regulation minimizes moisture stress within the root zone and it does not quantify the water-controlled evapotranspiration regime. Yet it can provide indirect solution for practical questions as it offers a comprehensive and quantitative description of soil moisture management measures which can keep the soil moisture within the desirable range. In this way the simulation procedure identifies long term equilibrium criteria for the given conditions and it can be envisaged for wide scale use in the allocation of available farmland resources among the required major crops, and in the selection of optimal combination of soil moisture regulation measures and cultivation technologies for given land conditions and crop type.

The proposed methodology could also be applied in climate change assessments by repeating the simulations for various assumed or predicted climates. A more simple solution of assessing hydroecological impacts of climate change can also be implemented by assuming equivalence of the climate differences of various regions and those of various time periods. Based on this assumption, climate dependent changes of irrigation water demands have been recently assessed for the country's agricultural areas by analyzing regional differences in water demands in dependence of the corresponding changes in average temperature and precipitation during the growing season (*Szesztay,* 1995).

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Relationships of water- and nutrient supply, yield and evapotranspiration of maize

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Abstract—The cycle and utilization of water were examined in lysimeters and on plots in a long-term experiment with four different water- and nutrient supply levels. The given nutrient doses were 100, 200, 300, and 400 kg/ha NPK substance in ratio 2:1:1, and the water-supply levels were rainfall (control), optimum water supply, and two intermediate levels. The given irrigation water levels in average of 18 years were the following: 0, 57, 126, and 173 mm.

It was established that the yield of maize is in close linear correlation both with the amount of precipitation in the growing season (between 150–470 mm), and annual precipitation (between 300–700 mm). The more favorable the nutrient supply, the closer the relationship is between the yield and precipitation.The closest correlation occurred between the evapotranspiration and yield. The irrigation and nutrient supply increased in a larger extent than the evapotranspiration, and the productivity of precipitation and irrigation was improved significantly by the two factors. The amount of water used per 1 kg of grain yield (Q) decreased with the increasing yield by a power function.

The utilization of precipitation is the worst at dry weather, when the evapotranspiration of maize (FAO 400) remains under 350 mm. In such case the yield is raised more than once by 100 to 200 mm supplementary water supply, and the specific water-use (transpiration coefficient) is decreased in an expressive way.

Key-words: evapotranspiration, utilization of irrigation water, utilization of precipitation, yield, maize.

1. Introduction

It is well-known that the yield is influenced by numerous ecological, biological, and agrotechnical factors. Among them the water supply is the most changeable factor. In Hungary the annual precipitation varies between 300–900 mm, while in the growing season it changes between 120–700 mm. 30 days long periods without precipitation are not rare, and more than 100 mm precipitation is also possible in one day. It often happens that wet and dry periods follow each other during the same growing season. Sometimes there are consecutive droughty or wet periods for several years.

The effect of fluctuation of precipitation is well-known, but the exact quantity relations between the amount and utilization of precipitation and irrigation water are yet missing or they are contradictory.

The increase of precipitation's productivity and rational control of plant water balance have to be examined concerning:

- the water requirement of plants,
- the frequency and yield-limiting effects of water shortage,
- the possibilities of improving precipitation's productivity and moderating the fluctuation of yield,
- the conditions and methods of efficient and environment-friendly water management control,
- the possibilities of harmonization of ecological, biological, and agrotechnical factors.

The report deals with the relations of precipitation, irrigation water, and nutrient supply to the actual evapotranspiration, yield, and productivity of water in maize.

2. Literature overview

The study of ecological needs of plants started with the searching of their water requirement. As early as the end of the 19th century, exact cylinder experiments were carried out to examine the water cycle and transpiration of plants. Measurements of consumptive use of water, namely the evapotranspiration began early in the 20th century (*Blaney,* 1954; *Thornthwaite* and *Mather,* 1961; *Armin,* 1969).

In Hungary the evapotranspiration measurements with compensation lysimeters were started by *László Erdős* (1966) in 1959 and *Emánuel Antal* (1966a, 1966b, 1966c) in 1963. However, in its popularization and wide range application *Emánuel Antal* has the biggest role. His outstanding research activities and personal contribution resulted in building of compensation lysimeters at least in ten different parts of the country within a few years (*Antal,* 1968a, 1968b; *Petrasovits,* 1966; *Balogh* and *Petrasovits,* 1970; *Possgay,* 1983; *Füri,* 1968; *Posza* and *Tóth,* 1970; *Gergely,* 1976), although, only a few pieces of lysimeters (6–12) were built in one place, which were used for establishment of water requirement of plants at quasi-optimal conditions. *Ruzsányi* (1973) studied the water requirement modifying effect of different nutrient supplies at field crops.

Parallel to investigations carried out by the Hungarian Meteorological Service, measurements of actual evapotranspiration have been started by *Szalóki* (1971) in 188 groundwater lysimeters at the Irrigation Research Institute, Szarvas, Hungary in plants supplied with different levels of water as well as by different ways of water supply. For the purpose of studies of evapotranspiration moderating effect of different agrotechnical factors (variety, plant density, water- and nutrient supply, etc.), a lysimeter station of 320 lysimeters was built in 1971, and this paper is based on the results of experiments carried out at this lysimeter station.

The actual evapotranspiration and water requirement of plants depend on a number of ecological, biological, and agrotechnical factors (*Antal,* 1968c, 1972; *Antal* and *Posza,* 1967; *Szász,* 1995; *Varga-Haszonits* and *Harnos,* 1988). The change in water requirement of plants can be determined — within the potential possibility — mostly by the surface area of plant-stand, and to a lesser extent on its age. There is a connection between the leaf area and the increasing *ET*, which can be characterised by a saturation curve *(Szalóki,* 1970, 1989), where the evapotranspiration on a unit area of leaves shows a decreasing tendency, even if it is caused by a more favorable water- and nutrient supply. The water requirement is modified by agrotechnical factors, primarily through the effect on surface of plant stand. Besides this, the direct effect of soil cultivation is also indisputable as far as the vegetation does not cover the soil.

Most researchers agree that mostly the nitrogen influences the size of leaf area and yield, and thereby the amount of evapotranspiration and its utilization. Many researchers stated — first of all those making experiments in cylinders — that the yield increase generated by fertilizing required more water only in absolute amount, at the same time the transpiration coefficient decreased (*Frank* and *Hank,* 1952; *Debreczeni*, 1965; *Szlovák,* 1968; *Cselőtei*, 1977).

In Hungary the relationships of fertilizer, water consumption, and yield of maize were examined by *Szalóki* (1971, 1989), *Ruzsányi* (1973), *Antal et al.*, (1975), *Szalóki* and *Németh* (1985), and *Szalókiné Zima* (1992) with regression analyses. It was found that the leaf area — in the case of no other limiting factors — was increased by the improvement of water- and nutrient supplies, and as a result, the evapotranspiration is enhanced by an exponential or saturation curve.

The accumulation of dry matter grows proportionally to the extension of leaf area. At the same time the ratio of dry matter content of grains increases within the total dry matter content, i.e., the harvest-index increases. Coming from the previous statement, the water use per unit of grain yield decreases by a hyperbolic function, i.e., the productivity of water improves (*Szalókiné Zima,* 1992; *Szalókiné Zima* and *Szalóki,* 1998).

It was proven unanimously by the long term fertilizer experiments, that the precipitation (if it is enough) and the irrigation water are utilized better by plants — in case of adequate nutrient supply —, and there is a positive interaction between them (*Debreczeni,* 1994; *Berzsenyi,* 1993; *Ruzsányi,* 1992; *Nagy,* 1994; *Lásztity* and *Csathó,* 1994; *Sárvári,* 1999; *Szalóki,* 1983). Considering the connection between water and nutrient supply, there are contradictory opinions among the researchers.

From the tendentious increase of yield and improving utilization of water some researchers concluded that the enhancing yield does not imply the raising of water consumption (*Klapp,* 1962; *Prenk,* 1967). On the other hand, *de Wit* (1958) has proved that the better water supply increased the water consumption by the same proportion as the quantity of yield, while the utilization of water remained practically the same. Regarding the productivity of irrigation water, it generally agrees with the getting of the diminishing returns, where the utilization of raising irrigation-water is more and more decreased (*Cavazza,* 1963). According to several authors' opinion, the productivity of irrigation water can be higher than that of a small amount of precipitation (*Ruzsányi,* 1975; *Possgay,* 1983; *Szalóki,* 1983).

The relationship between the water shortage and harvest deficiency, and the possibility of its reduction were examined by several authors (*Petrasovits,* 1989; *Ruzsányi,* 1989, 1992; *Szőke Molnár,* 1989; *Szalóki,* 1989; *Varga-Haszonits,* 1989; *Berzsenyi,* 1993). *Szalóki* (1989) found a very close relation between the actual water shortage and the harvest deficiency, which can be characterized by a quadratic equation. *Szalóki and Szőke Molnár* (1981) and *Szőke Molnár* (1989) also completed a simulation model of water shortage and harvest deficiency. Using this model, the probability dispersion curves of potential water shortage and harvest deficiency were determined with lot of plant species and for many different sites.

3. Materials and methods

The experiments have been carried out at the two hectare area of the Lysimeter Station of the former Irrigation Research Institute (now Research Institute for Fisheries, Aquaculture, and Irrigation) Szarvas, Hungary, since 1971, in 320 lysimeters. The experiments have been arranged in five blocks with 64 lysimeters in each. Four different water- and nutrient supplies were applied in 16 treatment-combinations. The number of repetitions was four.

The size of lysimeters was 1 m × 1 m × 1 m (1 m^3), and they were built into the middle of each plot with 32 m^2 area. The plots and lysimeters got the same treatments. The groundwater level is not kept in lysimeters, which is characteristic to the Thornthwaite-type lysimeters. The infiltration water is leaching from lysimeters to the dishes placed in cellars by gravitation, the amount of which is measured and its chemical content is analyzed in laboratory conditions.

In the three blocks two-factor long-term experiments have been carried out for 30 years, four repetitions, with the following 16 combinations of treatments.

The water supply in the three main blocks was the following:

I_1 - non-irrigated **control** (natural rainfall),

I_2 - irrigated with **one third** of optimum water supply,

I_3 - irrigated with **two thirds** of optimum water supply,

I_4 - **optimum** water supply (irrigated according to the demand of plants).

The amount of on-demand irrigation means the amount of supplementary water supply which is required for keeping the disposable water content in the active root zone to be higher than 50%.

The soil moisture difference was determined at the beginning and end of the growing season by gravimetric method. The continuous control of soil moisture content during the growing season was carried out by a BWK-Lanze Soil Moisture Meter. Water supply was carried out by drip irrigation through pipes laid on the soil surface, or by a portable plot irrigating device.

The calculation of actual evapotranspiration (ET_a) was carried out according to the following equation:

$$ET_a = P + I^{+} - Sm_{diff} - Inf, \tag{1}$$

where P is the precipitation, I is the irrigation water, Sm_{diff} is the soil moisture difference between spring and autumn, and Inf is the infiltration.

Daily evapotranspiration was measured by floating lysimeters supplied with optimum water- and nutrient level. Within the main treatment (water-supply), the nutrient levels were 100, 200, 300, and 400 kg/ha NPK substances in ratio 2:1:1.Type of the soil is chernozem meadow, which is well supplied with phosphorus and potassium, and has a medium nitrogen content. Its natural water capacity is 40 volume percent, half of which is disposable water. Such experiments are carried out in three blocks, in 192 lysimeters, and plots with three plant species. In the other parts of the experimental area different type experiments are run.

Hereafter the results of 18 years long experiments in water balance and water use are evaluated in maize (Volga, Florencia, Columba, Nónius, Reseda).

4. Results and discussion

It is visible from the literature, that there are differences in respect of quantity relations between water and yield, coming from great number of influencing factors. However, it is obvious, that using small amount of water is insufficient, and the precipitation or irrigation water above the water requirement of plants is not utilized, moreover, it may cause depression in yield. As a consequence, it is not advisable to study the effect and productivity of precipitation and irrigation water without considering the requirement of plants and its harmonization to the other factors.

The objectives of the study were realised by examination of results collected in a long-term experiment in maize. In the three blocks of the experiment, the effects of precipitation and three different levels of supplementary water supply on evapotranspiration, yield, and water utilization were investigated in four (same every year) different nutrient supply treatments. Results are demonstrated in *Tables 1* and 2, and *Figs. 1* to *4*.

Each element of the water balance was measured, so the water consumption could be estimated with reliable accuracy. The moisture content of the soil and its changes were measured in every 10 cm of soil layer as far as 2 m depth at the time of sowing and harvest, but the infiltration into more than 2 m deepness and amount of surface run-off water were not calculated. Based on our observations and calculations, the infiltration under the 2 m soil layer could be significant only at the I_4 treatment.

During the 18 years long experiments, the average precipitation was 297 mm in the growing season, and its extreme values changed between 150–450 mm. The amounts of annual precipitation (October 1–September 30) fluctuated between 300 and 700 mm around an average of 491 mm.

The given irrigation water has also varied not only by treatments but also by years, and in the I_4 treatment it changed between 60–300 mm. In average of 18 years the irrigation water amounts were 57, 126, and 173 mm in each treatment (*Table 1/a*). In the growing season the amount of irrigation water together with the precipitation was 354–470 mm in average, and with the annual precipitation it was between 548–664 mm (*Tables 1/b* and *1/c*). The winter evaporation (*E*) was averaged around 100 mm (*Table 1/e*).

The infiltration could have been measured only in lysimeters. The amount of infiltration fluctuated between 40–100 mm in a many years average, which occurred mainly early in spring at melting of snow and at the beginning of the growing season (April and May). The amount of infiltration was increased by the precipitation and irrigation and was decreased by the dose of fertilizers (*Table 1/f*).

Table 1. Water balance components in lysimeter- and in plot-experiment in average of 18 years

Treatment	Dimension	Fertilization treatment: NPK 2:1:1				Average	
		100	200	300	400	abs., mm	relative, %
1/a Irrigation water (I)							
I_1	mm	0	0	0	0	**0**	**0**
I_2	mm	57	57	57	57	**57**	**33**
I_3	mm	126	126	126	126	**126**	**73**
I_4	mm	173	173	173	173	**173**	**100**
1/b Precipitation in growing season (P_1) + irrigation water (I)							
P_1	mm	297	297	297	297	**297**	**100**
P_1+I_2	mm	354	354	354	354	**354**	**119**
P_1+I_3	mm	423	423	423	423	**423**	**142**
P_1+I_4	mm	470	470	470	470	**470**	**158**
1/c Annual precipitation (P_2) + irrigation water (I)							
P_2	mm	491	491	491	491	**491**	**100**
$P_2 + I_2$	mm	548	548	548	548	**548**	**112**
$P_2 + I_3$	mm	617	617	617	617	**617**	**126**
$P_2 + I_4$	mm	664	664	664	664	**664**	**135**
1/d Evapotranspiration in growing season in lysimeters ($ET_a ly$)							
$ET_a ly_1$	mm	344	356	358	355	**353**	**100**
$ET_{al} y_2$	mm	385	409	408	410	**403**	**114**
$ET_a ly_3$	mm	436	452	465	464	**454**	**129**
$ET_a ly_4$	mm	453	476	499	509	**484**	**137**
Average	**mm**	**405**	**423**	**433**	**435**	**424**	
1/e Evaporation (E) in lysimeters in winter season							
E_1	mm	102	94	95	98	**97**	**100**
E_2	mm	105	94	96	97	**98**	**101**
E_3	mm	111	108	104	109	**108**	**111**
E_4	mm	110	102	104	103	**105**	**108**
Average	**mm**	**107**	**100**	**100**	**102**	**102**	
1/f Infiltration (Inf.) in lysimeters							
Inf_1	mm	45	41	38	38	**41**	**100**
Inf_2	mm	58	45	44	41	**47**	**115**
Inf_3	mm	70	57	48	44	**55**	**134**
Inf_4	mm	101	86	61	52	**75**	**183**
Average	**mm**	**69**	**57**	**48**	**44**	**54**	
1/g Evapotranspiration in growing season on plots ($ET_a p$)							
$ET_a p_1$	mm	389	397	396	393	**394**	**100**
$ET_a p_2$	mm	443	454	452	451	**450**	**114**
$ET_a p_3$	mm	506	509	513	508	**509**	**129**
$ET_a p_4$	mm	554	562	560	561	**559**	**142**
Average	**mm**	**473**	**480**	**480**	**478**	**478**	

The annual actual evapotranspiration (*ETa*) of maize varied between 270–600 mm during the analyzed period. During the 18 years of experiments, the average *ETa* has been changed between 344–509 mm depending on the applied treatments (*Table 1/d*). The irrigation increased the evapotranspiration, but in the well-watered treatments (I_3 and I_4) it also raised when higher doses of fertilizers were applied (Table 1/d). The *ETa* of maize was estimated to be between 390–560 mm on plots (*Table 1/g*), on the basis of examination of water balance in the 2 m soil layer. It is worth to mention that the upper 2 m soil layer was saturated several times to water capacity in spring at the I_4 treatment. Presumably, certain infiltration appeared during these years under the 2 m soil layer as well, which has not been taken into consideration in the determination of *ET* on plots. However, in non-irrigated treatments and at less supplementary water supply, the calculated values of *ET* are close to the actual values. It is supported not only by the values of average yields, but also the tendency of water productivity.

The average yields of many years are shown in *Tables 2/a* and *2/b*. Comparing the yields in lysimeters with the ones collected on plots shows that the yields are lower in lysimeters than on plots, especially at low water- and nutrient supply levels. The reason is that the lysimeters are only 1 m deep, and significant amount of water and nutrients is released from them.

It is indicated by the about 5 t/ha average yield of non-irrigated maize, that the formation of yield is often strongly limited by water shortage. There were droughty years when there was no grain yield at all in the lysimeters or it was very low. However, in wet years and in treatments with adequate water supply, the yield exceeded 10 t/ha. The increase of fertilizer dose compared to the basic (100 kg/ha NPK) treatment had a depressive effect in the non-irrigated treatments in dry years. In wet years and in treatments with adequate water supply (I_3; I_4) the fertilizer also had a good utilization.

The average yield was increased by 3–4 t/ha in lysimeters and 1–2 t/ha on plots in the irrigated treatments, if higher NPK doses were applied and the water supply remained the same. The yield was increased to higher extent by irrigation than by raising of nutrient levels. Nevertheless, the effect of fertilizers was continuously higher in the recent years because of the differences appearing in the nutrient reserve of the soil, and both factors proceeded in a positive interaction with each other.

The correlation between raising yield and water supply is evaluated on the one hand by a regression analysis, and on the other hand by a coefficient expressing the productivity of water.

The relations between the precipitation and yield at different nutrient treatments can be seen in *Figs. 1* and *2* based on the annual data of non-irrigated plots. It also emerges from the data, that the yield of maize has

Table 2. Productivity of precipitation and irrigation water

Treatment	Dimension	Fertilization treatment: NPK 2:1:1				Average	
		100	200	300	400	absolute	relative, %
2/a Yield in lysimeters (Yly)							
Yly_1	t/ha	4.4	5.0	5.2	4.8	**4.9**	**100**
Yly_2	t/ha	6.0	7.2	7.9	7.9	**7.3**	**149**
Yly_3	t/ha	6.9	9.0	10.6	10.9	**9.3**	**192**
Yly_4	t/ha	8.0	9.7	11.6	12.4	**10.5**	**215**
Average		**6.4**	**7.7**	**8.8**	**9.0**	**8.0**	**164**
2/b Yield on plots (Yp)							
Yp_1	t/ha	6.9	7.0	7.3	7.0	**7.1**	**100**
Yp_2	t/ha	8.9	10.1	10.3	10.2	**9.9**	**139**
Yp_3	t/ha	11.0	12.0	12.7	13.0	**12.2**	**171**
Yp_4	t/ha	11.1	12.3	13.4	13.8	**12.7**	**178**
Average		**9.4**	**10.3**	**10.9**	**11.0**	**10.4**	
2/c Productivity of the evapotranspiration in lysimeters (Y/ET_a)							
Yly/ET_{a1}	kg/ha/mm	12.8	14.1	14.4	13.7	**13.8**	**100**
Yly/ET_{a2}	kg/ha/mm	15.7	17.5	19.4	19.3	**18.0**	**130**
Yly/ET_{a3}	kg/ha/mm	15.8	19.8	22.9	23.4	**20.5**	**149**
Yly/ET_{a4}	kg/ha/mm	17.8	20.4	23.3	24.3	**21.6**	**156**
Average		**15.7**	**18.2**	**20.4**	**20.7**	**18.8**	
2/d Productivity of the evapotranspiration on plots (Y/ET_ap)							
Yp/ET_ap_1	kg/ha/mm	17.7	17.6	18.5	17.8	**17.9**	**100**
Yp/ET_ap_2	kg/ha/mm	20.0	22.2	22.8	22.5	**21.9**	**122**
Yp/ET_ap_3	kg/ha/mm	21.7	23.5	24.8	25.5	**23.9**	**134**
Yp/ET_ap_4	kg/ha/mm	20.0	22.0	23.9	24.6	**22.6**	**127**
Average		**19.8**	**21.3**	**22.5**	**22.6**	**21.6**	
2/e Productivity of the annual precipitation + irrigation on plots							
$Yp/(P_2+I_1)$	kg/ha/mm	14.0	14.3	15.0	14.2	**14.4**	**100**
$Yp/(P_2+I_2)$	kg/ha/mm	16.2	18.4	18.8	18.6	**18.0**	**125**
$Yp/(P_2+I_3)$	kg/ha/mm	17.8	19.4	20.6	21.1	**19.7**	**137**
$Yp/(P_2+I_4)$	kg/ha/mm	16.7	18.6	20.2	20.8	**19.1**	**132**
Average		**16.2**	**17.7**	**18.6**	**18.7**	**17.8**	
2/f Effect of the irrigation (more yield, t/ha)							
I_2ly	t/ha	1.6	2.2	3.1	2.7	**2.4**	**49**
I_3ly	t/ha	2.5	4.0	5.4	5.7	**4.4**	**92**
I_4ly	t/ha	3.6	4.7	6.4	7.2	**5.5**	**115**
I_2p	t/ha	2.0	3.1	3.0	3.2	**2.8**	**39**
I_3p	t/ha	4.1	5.0	5.4	6.0	**5.1**	**71**
I_4p	t/ha	4.2	5.3	6.0	6.8	**5.6**	**78**
2/g Productivity of the irrigation (Y/I)							
I_2ly	kg/ha/mm	28.5	37.9	55.0	47.7	**42.3**	**294**
I_3ly	kg/ha/mm	19.8	31.4	43.2	44.9	**34.8**	**242**
I_4ly	kg/ha/mm	20.9	27.3	37.2	41.5	**31.7**	**220**
I_2p	kg/ha/mm	34.6	53.9	51.8	56.0	**49.1**	**341**
I_3p	kg/ha/mm	32.6	39.4	42.8	47.7	**40.6**	**282**
I_4p	kg/ha/mm	24.3	30.9	34.9	39.6	**32.4**	**225**
Average	kg/ha/mm	**26.8**	**36.8**	**44.1**	**46.2**	**38.5**	**267**

changed between 2–12 t/ha at the same NPK supply depending on weather conditions, and it is in close correlation both with the rainfall in the growing season and annual precipitation. Furthermore, it can be observed that in treatments with low supplementary nutrient supply, the fluctuation of yield was less, as in these treatments nutrient shortage was the limiting factor in rainy years. The degree and closeness of the relation, i.e., the values of regression and correlation factors increase with nutrient levels, which refer to higher utilization of precipitation caused by the fertilizers.

There is an even closer connection between the yield and evapotranspiration than between the yield and precipitation. Regression analysis for the two variables was made using 72 pairs of data of the 18 years and four nutrient- and water supply treatments. Results are shown in *Fig. 3*. It is visible that the correlation between *ET* and grain yield is loose at low NPK level, because in this case the average yield is limited not only by the water shortage but also the shortage of nutrients. The nutrient supply in the soil was highly modified also by the previous crop. At a better nutrient supply, the variability of *ET* and the average yield is more and more determined by the precipitation and irrigation, which is indicated by the increasing regression and correlation factors. The increase of the regression factors means that a yield increase per 1 mm evapotranspiration is got with rising nutrient levels (Fig. 3), i.e., the grain yield of maize enhanced to higher extent with the improving water- and nutrient supply than the *ET*, which means that the productivity of water increased and the amount of *ET* per unit of yield decreased. It is shown by functions between the average yield and transpiration coefficient (Q = evapotranspirated water per unit of grain yield) in *Fig. 4*.

The results prove definitely that the value of Q depends primarily on the average yield, and it is in a very close negative correlation with the growing yield, which can be characterized by a logarithmic- or hyperbolic curve. Furthermore, it has been established that every factor which increases the yield has a favorable effect on the productivity of water.

Without irrigation the average grain yield is 13.8 kg/ha in lysimeters and 17.9 kg/ha on plots per 1 mm evapotranspiration. At the same time these values increased as high as 24–25 kg/ha in the irrigated treatments.

From the previous observations it is concluded that in case of heavy rainfalls, the productivity of precipitation does not decrease at adequate nutrient supply but increases to some extent. It is supported by the regression factors of Fig. 2 (i.e., the excess yields per 1 mm increase in rainfall) which are continuously on the rise and exceed the 18 years average of I_1 treatment (Table 2). Consequently, not only the average yield but also the productivity of water were increased by the precipitation, as it served the satisfaction of water requirement of plants and improved the other life conditions of plants as well.

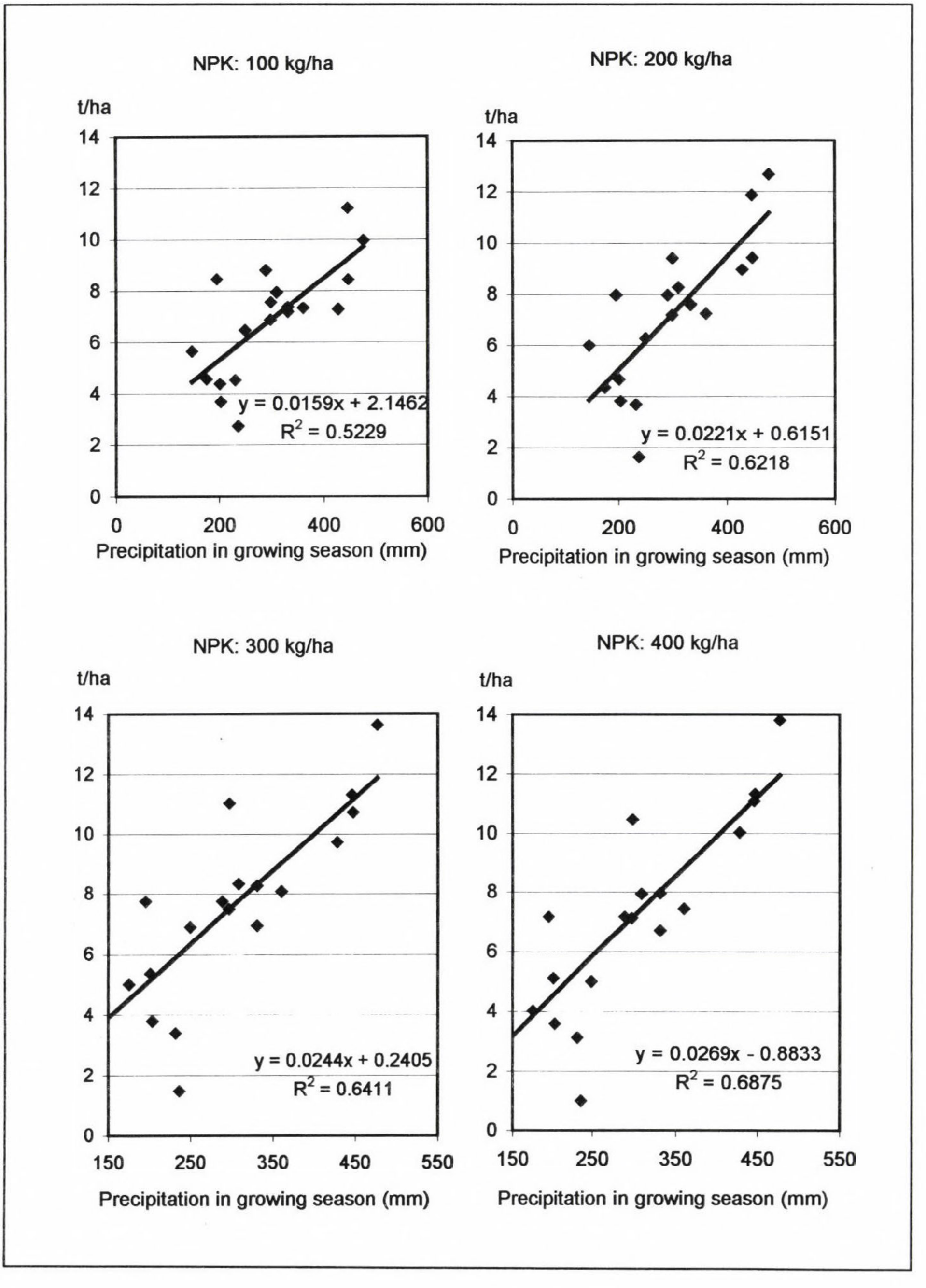

Fig. 1. **Relationship between the precipitation and yield during the growing season at different NPK levels in non-irrigated maize.**

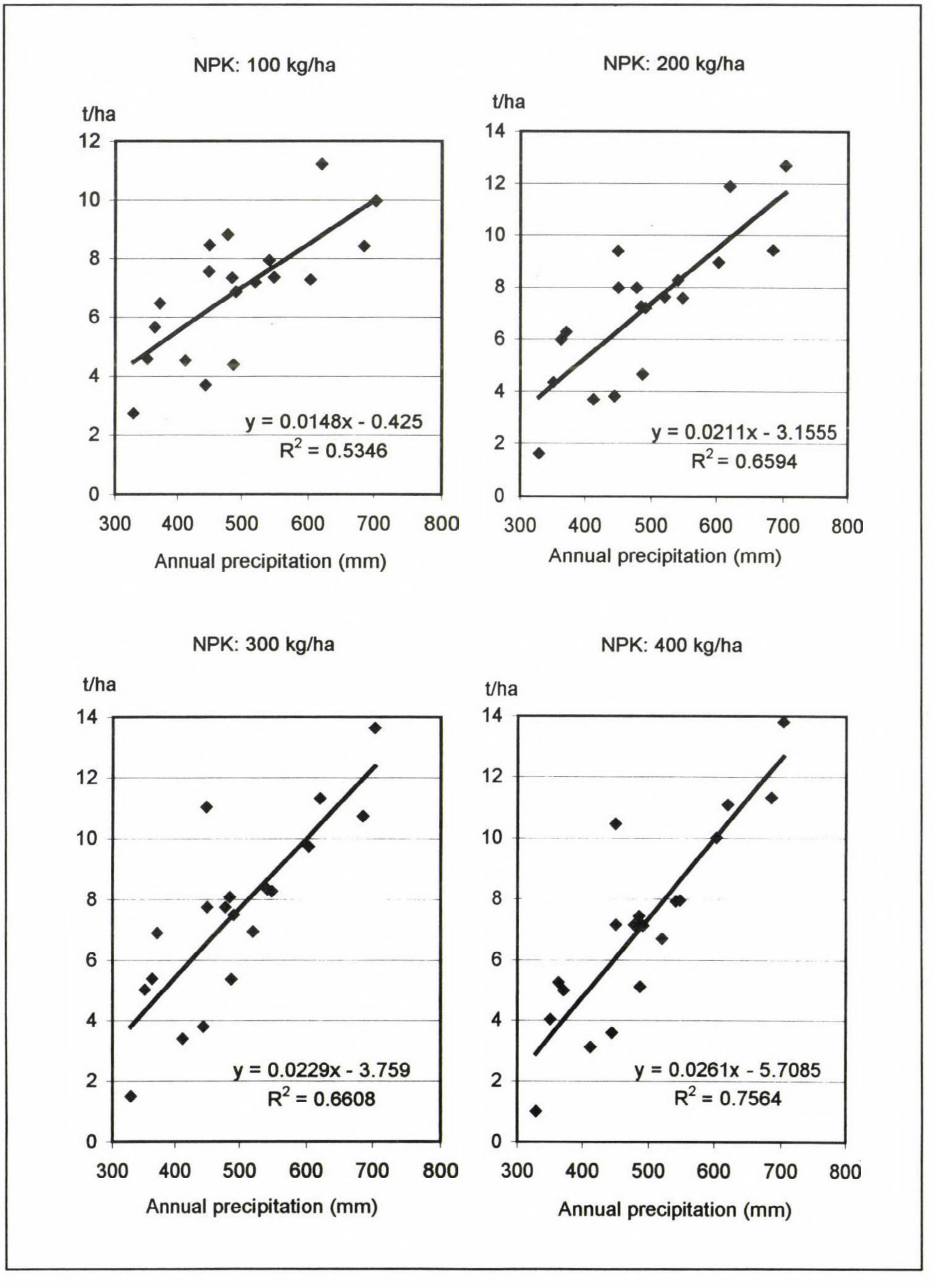

Fig. 2. **Relationship between the annual precipitation and yield at different NPK levels in non-irrigated maize.**

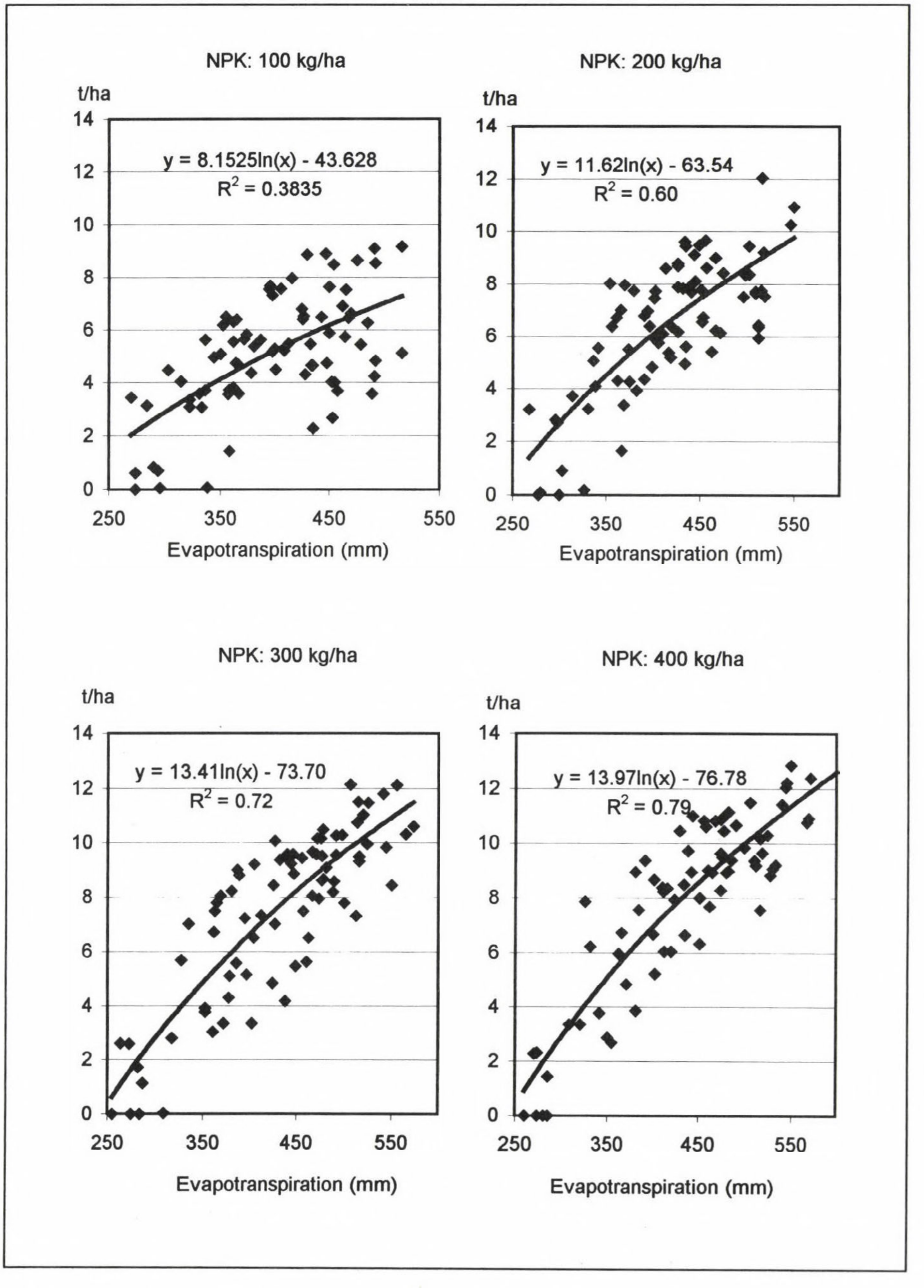

Fig. 3. Correlation between the evapotranspiration of maize and its dry grain yield at different NPK levels.

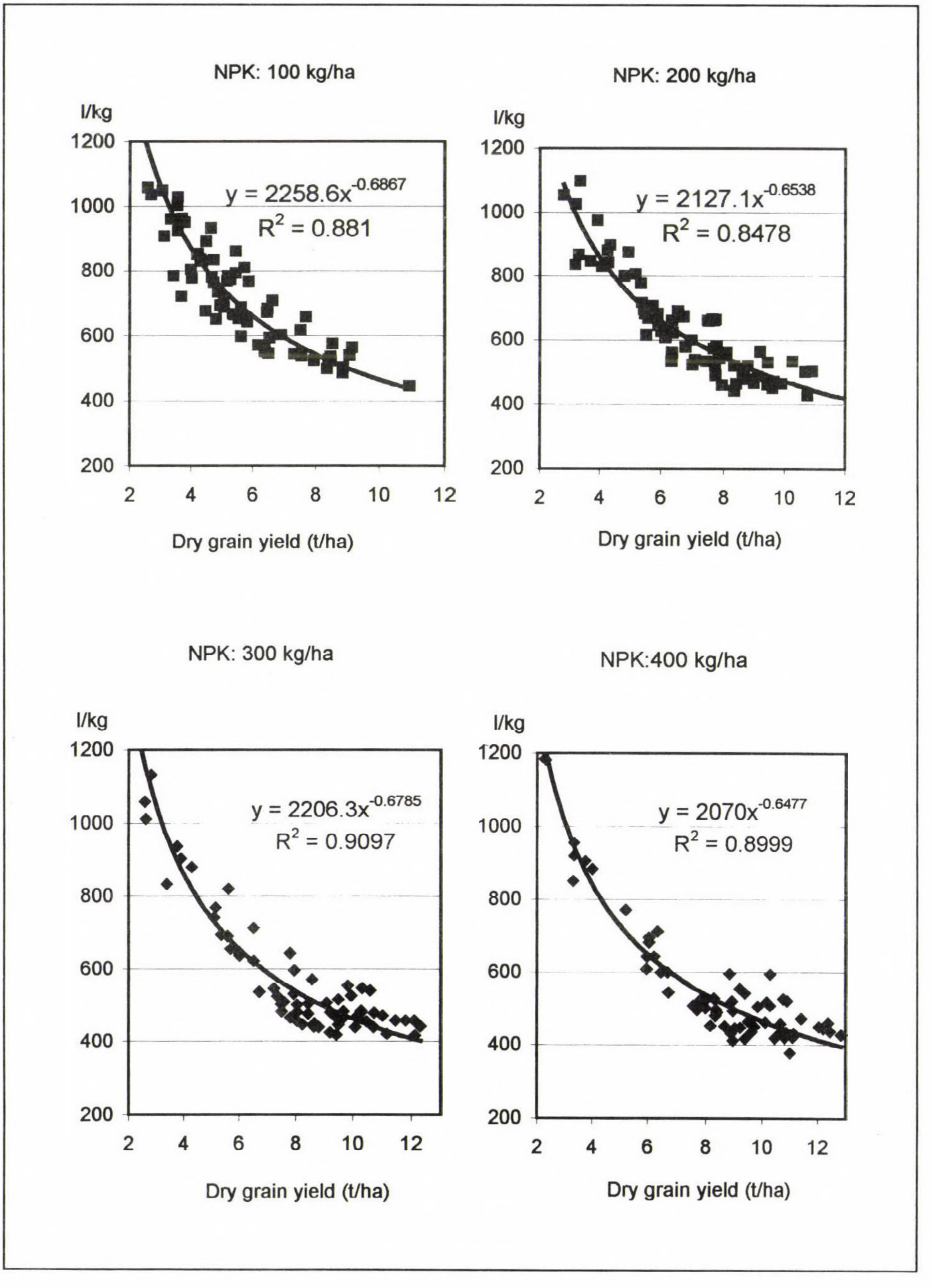

Fig. 4. Correlation between the dry grain yield and the evapotranspiration coefficient of maize at different NPK levels.

Both precipitation and nutrients have weak utilization if the water- and nutrient supplies are insufficient. If adequate and harmonious nutrient and water supply is ensured, the positive interaction of factors predominates, and the yields grow to large extent. Together with the increase of yields the productivity of water also improves. It is demonstrated by the data of the 18 years long experiments as well. The effect of improving water- and nutrient supply of these experiments is presented in Table 2. These data also support the conclusion that with the aid of supplementary water supply the yield was increased on a larger scale than the evapotranspiration, i.e., the utilization of water improved.

The yield increase per 1 mm evapotranspiration, 1 mm precipitation, as well as annual precipitation + irrigation water is also shown in *Tables 2/c, 2/d,* and *2/e.* According to the exact measurement data, the utilization was weak at low levels of precipitation which was far behind the requirement of plants (12.8–14.4 kg/mm), especially in lysimeters having shallow productive layer, where even the increase of fertilizer dose was inefficient.

Compared to the above mentioned non-irrigated treatment, the 57 mm supplementary water supply increased the productivity of evapotranspiration to 18 kg/ha, the 126 mm water supply to 20 kg/ha, and finally the 173 mm irrigation water increased the productivity of *ET* to 21.6 kg/ha. The productivity of water also increased on the plots as far as the third level of irrigation.

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Impact of the climate change on the hydrological regime. A case study on the Danube River

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Abstract—Many Hungarian national studies related to the impact of the climate changes on the stream flow have been published in the last decade. This paper made an attempt to summarize the results on the investigations of potential change of the runoff.

The model system used for operational hydrology data was applied for the climate change study.

The Danube River, its tributaries, and the Hernád River have been investigated. The model system was used for three scenarios of the upper Danube Basin. Some fragments of the studies have been selected as examples for the applicability of the method. The effects of the change in precipitation and temperature on the change of streamflow (runoff) for four scenarios have been applied. The effect of the change in runoff can be predicted particularly. The method is introduced by examples on real observed data (Appendix). The method can be applied for comprehensive studies on the effects of the climate change.

Key-words: case study, climate change, discharge, flood wave, flood, impact modelings, runoff, snow melt, water intakes.

1. Introduction

Hungarian national studies (*Antal* and *Starosolszky*, 1990; *Starosolszky,* 1989) on the effects of the climate change go back to the eighties. A comprehensive report covering several aspects was published in 1994 (*Starosolszky,* 1994a,b). In a study for the International Institute for Applied Systems Analysis (IIASA, Laxenburg, Austria), Water Resources Research Centre (VITUKI) submitted a report on the climate change impact covering the upper Danube Basin (*Starosolszky,* 1994b). Another report within the PECO Project of the *European Union* (1997) includes the following items:

- Impact of a potential climate change on long-term discharge data series;
- The expectable impact of the climate change on the characteristic discharges;
- Effects of the temperature change on the catchment over the Nagymaros section and on the flow conditions of the Danube.

The basic aim was the analysis of the regime of the Danube River at Nagymaros. Special emphasis was devoted to the change of temperature in the drainage basin with special regard to snowmelt induced runoff (*Starosolszky,* 1989).

For the studies three periods, each of them covers one year from September 1 to August 31, were selected. The first study for the daily temperatures was fulfilled with (1) observed values, (2) temperatures 1°C higher, and (3) 3°C higher than the observed values all over the periods. The following questions have been tackled:

- seasonal and monthly distribution of runoff;
- changes in the winter low-flow period;
- changes in the peak discharges of floods in winter;
- changes in the peak discharges of the spring snowmelt induced floods.

The calculations with constant temperature rises are oversimplified and can not express the complicated cross-relations among temperature, precipitation and runoff. These aspects have been neglected in the previous studies (*Starosolszky* and *Gauzer,* 1996).

2. Scenarios

The voluminous data sets are decomposed for selected periods for modeling climatological and hydrological phenomena. The basic changes on the Danube upstream from Nagymaros were selected for temperature, precipitation, and runoff. The data set can cover simulated values originated from the three scenarios as follows:

- UK Meteorological Office equilibrium experiment (UKHI),
- UK Meteorological Office transient experiment (UKTR),
- Canadian Climate Centre high resolution equilibrium experiment (XCCC).

The fourth scenario was based on measured temperature, and precipitation (*Gauzer,* 1993a,b).

Simulated values can be considered in the data sets. The results gained include tables for daily, monthly, and seasonal streamflow data and their graphical presentation.

3. General trends

Climatologists and hydrologists produced some basic data and these studies were described in a special comprehensive publication (*Antal* and *Starosolszky,* 1990). The higher snowmelt induced winter flood wave occurs, the larger increase in the ratio of the volume of winter runoff to the annual volume can be expected. If no winter flood wave passes on the river, the changes remain insignificant (i.e., under 1–2%).

Changes of a winter flood wave depend on the role of snowmelt. If a winter flood wave induced by liquid precipitation passes on the river, increase of peak discharge remains insignificant (December flood wave in the wet period). In case of mostly snowmelt induced winter flood waves, peak discharges rise significantly (15–30%, December flood wave in the average period).

In case of several spring and early summer snowmelt induced flood waves, the rise of temperature results in an increasing peak discharge for the first flood wave, while consecutive flood waves in such a period pass on the river with decreasing peak discharges (see flood waves in the dry and average period). Peak discharge of a single major snowmelt induced flood wave shows increasing tendency (spring flood in the wet period).

These oversimplified conditions do not follow the realistic cases when temperature changes are combined with precipitation, snowmelt, and the consequence in the change of the runoff. The snowmelt, the rainfall-runoff, the coupled structural stochastic and unsteady flow models have been applied, and observations on rainfall-runoff were used (*Gauzer,* 1993a,b). The so-called EU (*European Union,* 1997) scenarios were selected for the three-year periods and applied in the PECO Project of the EU.

The simulation of the time series is excepted for three periods and the output is summarized for the Nagymaros gauging station on the Danube.

Fig. 1 shows the discharge time series for a selected year. *Fig. 2* demonstrates the monthly average runoff for a selected period using the results of the different scenarios. *Fig. 3* shows the seasonal averages of the observed and simulated values with special attention to the EU simulations.

Monthly average runoff values are tabulated in *Tables 1–3*. The values are variable according to the simulation method. The data originate from the comprehensive study (*VITUKI*, 1997).

Table 2 shows the seasonal average values of runoff. The variations of seasonal values do not show regular character. Similar investigations were executed for a medium size river (gauging station Gesztely, on the Hernád River). Modeling for a smaller river has more difficulties in the simulation grid (Appendix).

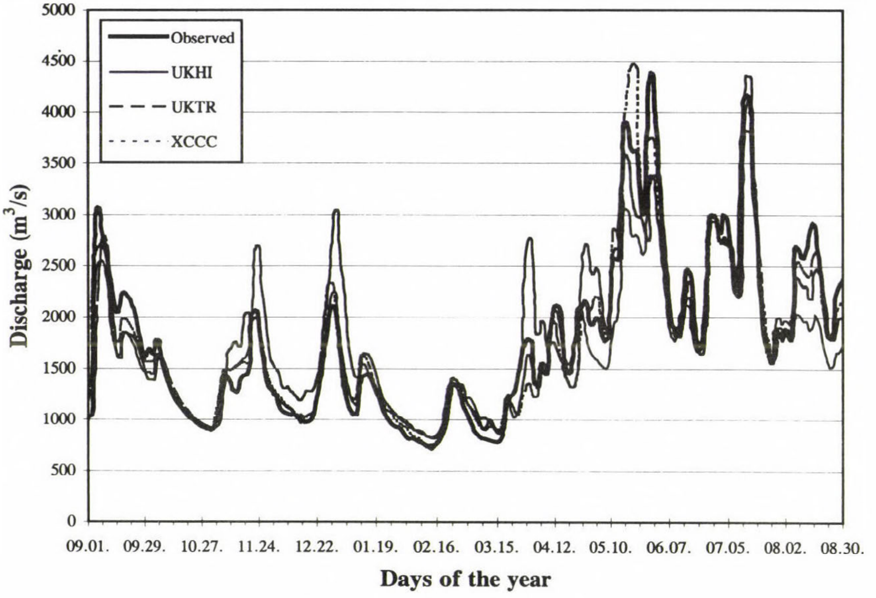

Fig. 1. Runoff simulation for the period 1995–96, at station Nagymaros, Danube River.

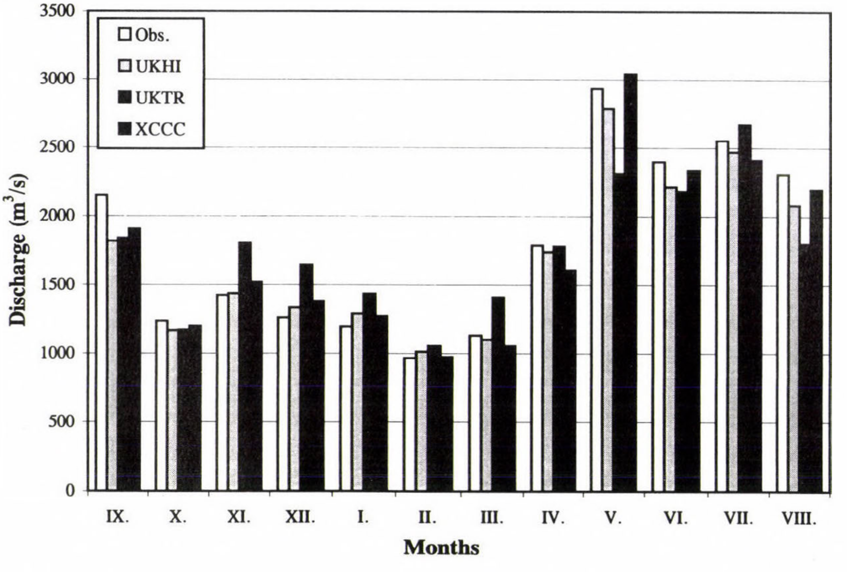

Fig. 2. Monthly average runoff for the period 1995–96, at station Nagymaros, Danube River.

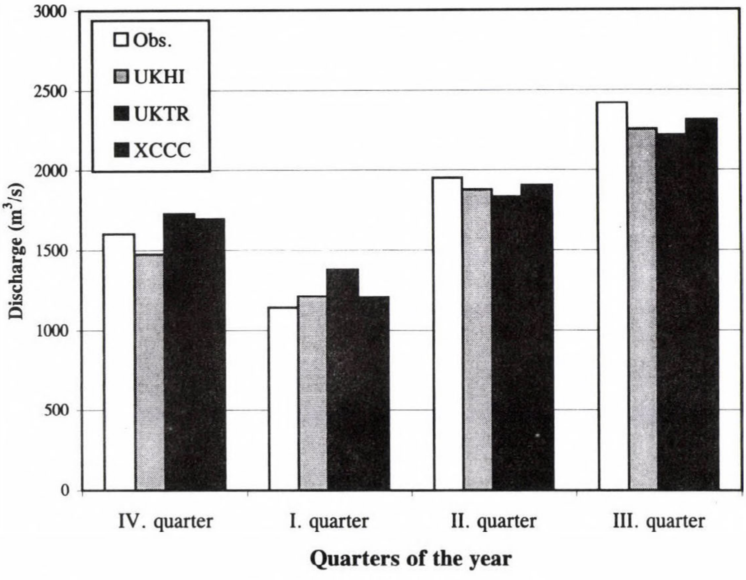

Fig. 3. Seasonal average runoff for the period 1995–96, at station Nagymaros, Danube River.

Table 1. Monthly average values of runoff for cross section of Danube at Nagymaros (m^3/s), for the period 1995–96

Scen.	IX.	X.	XI.	XII.	I.	II.	III.	IV.	V.	VI.	VII.	VIII.
Obs.	2151	1234	1424	1263	1195	967	1132	1787	2935	2395	2552	2307
UKHI	1818	1167	1436	1335	1291	1013	1101	1741	2787	2214	2469	2082
UKTR	1838	1170	1801	1644	1435	1054	1410	1779	2311	2182	2669	1800
XCCC	1906	1198	1518	1378	1273	974	1054	1608	3044	2335	2412	2197

Table 2. Seasonal average values of runoff for cross section of Danube at Nagymaros (m^3/s), for the period 1995–96

Scenario	IX–XI.	XII–II.	III–V.	VI–VIII.
Observed	1603	1142	1952	2419
UKHI	1474	1213	1876	2255
UKTR	1727	1378	1834	2218
XCCC	1697	1208	1902	2315

Table 3. Statistical characteristics of runoff for cross section of Danube at Nagymaros (m^3/s), for the period 1995–96

Scenario	Q_{min}	Q_{max}	Q	σ
Observed	741	4426	1681 (100%)	786
UKHI	798	4032	1707 (102%)	680
UKTR	833	4369	1760 (105%)	651
XCCC	746	4516	1744 (104%)	758

4. *Experiences*

For the period 1995–96 the following experiences have been gained, as examples: Since air temperatures were not very low during November and December in the observed series representing present day climate, a significant (approximately 5°C) rise of air temperature had a considerable effect on two insignificant floods passed during this period. In spite of the global warming air temperature increase, due to low air temperatures between January and March, changes of runoff remained insignificant. It can be concluded that during this period, due to the low air temperatures, the investigated scenarios have only a slight impact on the safety of power plant operation (Thermal Power Station at Százhalombatta, Nuclear Power Station at Paks).

The number of days gained in the scenarios with discharges higher than those according to the present day climate is 160, 193, and 188, respectively. These are less than values gained by previous periods investigated; this fact is the consequence of slight changes during winter.

The significant decrease of peak discharges of the bi-modal flood for UKHI and UKTR scenarios were the consequence of the earlier snowmelt. Considerable increase of the first peak for the XCCC scenario was produced by smaller air temperature changes.

Since the July flood originated from liquid precipitation, peak discharges were strongly related to the changes of precipitation values.

The decrease of effective precipitation during August was partly compensated by the increase of subsurface inflow as a consequence of higher effective precipitation during winter and spring. As a consequence, 30 percent decrease of precipitation resulted in 20 percent decrease of runoff on the upper catchment.

The rise of the annual average runoff was most significant for the UKTR scenario, due to the most expressed increase of precipitation. The change of standard deviation was influenced by the peak discharge values.

Concluding the present analysis, it is necessary to mention that the simple technique used to transform monthly changes into daily values and also the short data series used may limit the possibility to derive general conclusions from the results received.

Generalization of the results was made for upstream stations for the three selected years. The simulations of the Danube flow regime for three selected periods of one year demonstrated the change in the time series as a consequence of the variation of daily temperature and precipitation. The three scenarios based on global circulation models redistribute the flow time series for the selected periods.

Major changes were found in the freeze and snowmelt periods, thus the flow originated from snowmelt generated runoff was shifted and the spring flows changed characteristically. The minimum and maximum flows were distorted, e.g., minimum flow in the autumn period was further reduced, and maximum flow in springtime was increased.

The simulation of the flow time series offers some general trends in the flow redistribution during the year (e.g., from September 1 to August 31). The arbitrary nature of the selection of the years for the Danube study may hinder the exact generalization of the changes, but some characteristic change could be detected.

The extension of the river basin from the two upstream stations (Pfelling and Wasserburg) downstream to Nagymaros also shows the different nature of the upstream and downstream sub-basins. This is reflected by the smaller runoff coefficients of the catchments located in lower altitudes (like rivers Rába, Morava, Vah). The outputs of the study may be extracted for each gauging station but for the latest (PECO) study at Nagymaros gauging station is considered as reference downstream station (*European Union,* 1997).

5. Climate change and water intakes

Considerations of water intakes on the Danube can be useful for the water management. The crucial question of the Hungarian Danube is the change of the minimum flows at the large cooling water intakes for a thermal (Százhalombatta, rkm 1622) and a nuclear (Paks, rkm 1526) power station. Their pumping stations have been constructed in diversion (intake) canals, considering a design water level for the pumps. Due to water level depletion in the main Danube, the water levels in the derivations can be influenced.

For the sake of simplicity, flow relations have been derived between Nagymaros, and Százhalombatta, and Paks, respectively. A simple linear correlation was derived for the stations and the effects of the changes of the minimum flows were investigated (*Gauzer* and *Starosolszky,* 1997).

The simple flow interrelations are, as follows:

$$Q_{sb} = 104.5 + 0.996\, Q_{nm}\,,$$

$$Q_{pa} = 20.7 + 0.965\, Q_{nm}\,,$$

where

Q_{sb} - discharges at Százhalombatta,
Q_{pa} - discharges at Paks,
Q_{nm} - discharges at Nagymaros.

It was concluded that the lowering of the low flows will be transferred to the locations of the intakes, but the reduction of the flow in low water periods will do basically no harm to the water intakes. The effect of the riverbed degradation can be much stronger than the effect of the reduction of low flows.

The changes of the flow due to the climate change expressed by the three temperature and precipitation scenarios can not danger the basic water intakes in Hungary. However a combined effect of low flow reduction and bed degradation can be sensitive for further developments, particularly for new cooling water intakes. It is therefore suggested, that bed degradation due to river morphology and climate change may be simultaneously investigated from the point of cooling water intakes.

For demonstrating the application of the method, an *Appendix* is attached. The river size may limit the applicability. The Hernád River demonstrates a medium size catchment. The Appendix shows the results of modeling climate data according to the three scenarios.

APPENDIX

Application of the model for the simulation over the Hernád River Basin

The Danube study has a regional character. It was obvious to make a study on a tributary representing a medium catchment.

The methodology applied for the Danube catchment over Nagymaros was used for a local study of a subcatchment of the Tisza River, the catchment of the Hernád River (in eastern Slovakia and north eastern Hungary). The total area exceeds 5359 km^2, 4300 km^2 is in Slovakia, and the downstream part of 1013 km^2 is in Hungary. This can be the minimum size of a catchment where the method is applicable.

The average temperature varies between 8 to 10°C in consequence of the prevailing northeast dry and cool climatic fronts. Interesting to mention that the NE-SW oriented valley sometimes produces a very strange climatic situation, as a dry and hot front from SW direction initiates a sudden temperature rise and low humidity rates in the southern part of the valley. Also these circumstances contribute to the relative low precipitation rate, which amounts to 550 mm/year only in long-term average.

In the Slovakian part, the river drains agricultural areas and some major cities, with Spiska Nova Ves, Presov and Kosice being the three largest ones. The Hungarian part of Hernád is dominated by flood plains. The upper layer of **the** soil consist of clay, peat land, and in some areas sand.

The average temperature for the coldest month in the Hungarian part of Hernád is approximately -3.5°C, while the warmest month is July with 18–21°C. The rainfall is 530–580 mm/year, and the specific runoff 21 l/s km^2. The maximum rainfall is during the summer months.

Forests cover 200 km^2 (20%) of the total catchment area, and grassland extends to 260 km^2 (26%). The crop area is predominantly tilled, and consists of 410 km^2 (41%). Built up areas (roads, towns, and villages), water areas, riverbanks, etc. cover 137 km^2 (13%).

Agriculture is one of the major activities, and the main crops are winter-wheat, maize, sunflower, lucern, rape, potatoes, and barley.

The total population in the river basin is approximately 64,000. There are 72 municipalities within the Hungarian catchment, where the largest villages are Bőcs (2643 inhabitants), Encs (6663 inhabitants), Gesztely (2753 inhabitants), Gönc (2378 inhabitants), and Forró (2239 inhabitants). These five villages constitute 26% of the total population, showing that the total numbers of small villages are high. The largest villages are situated along the Hernád River.

The aim of the Hernád study was to demonstrate the applicability of the simulation method for a relatively small catchment of the Danube basin, where the number of the stream gauging stations is rather limited.

The reference periods selected for the simulation are from September 1, 1992 to August 31, 1993 and from September 1, 1995 to August 31, 1996.

For examining the climate change impact, the three selected scenarios UKHI, UKTR, and XCCC were also used. One may note that relatively small figure of the grids involved may influence the accuracy of the simulation. Thus, a refinement of the grid-system would be desirable. The variety of the temperature and precipitation over the catchment may influence the credibility of the simulation.

The results of the simulation are given in tabulated form (*Tables A1–A8*). The effect of the climate change scenarios is rather characteristic for the daily, monthly, and seasonal flows at the gauging station of Gesztely (Hungary).

The conclusion is that the change in flow does not influence heavily the hydropower generation neither at present, nor in the future. Thus, the annual hydropower generation will not be strongly effected.

Changes of air temperature and precipitation according to the scenarios

Table A1. Changes of the monthly air temperature (°C)

	IX.	X.	XI.	XII.	I.	II.	III.	IV.	V.	VI.	VII.	VIII.
UKHI	3.04	3.73	3.59	3.04	1.66	1.52	1.10	1.79	1.93	2.07	1.93	2.62
UKTR	4.20	4.50	7.40	3.70	2.30	1.50	2.30	2.50	3.60	2.50	4.60	5.50
XCCC	1.52	1.66	2.21	1.38	0.97	0.97	1.10	1.10	1.24	1.24	1.38	1.10

Table A2. Changes of the monthly precipitation (%)

	IX.	X.	XI.	XII.	I.	II.	III.	IV.	V.	VI.	VII.	VIII.
UKHI	14.08	23.74	9.38	4.97	1.38	2.35	−0.55	−5.38	−5.80	4.14	4.42	15.04
UKTR	20.30	28.20	45.90	26.00	−5.90	9.00	12.80	−12.8	−3.00	16.50	27.80	31.10
XCCC	9.25	15.18	11.18	9.80	4.00	−0.97	−6.21	0.14	−6.35	0.00	16.84	15.46

Table A3. Calculated monthly maximum flows (m^3/s)

	Observed	UKHI	UKTR	XCCC
1992–93	27.213 (July)	27.202 (July)	31.453 (July)	25.523 (July)
1995–96	32.110 (Aug)	31.234 (May)	31.643 (June)	32.282 (May)

Table A4. Calculated monthly minimum flows (m^3/s)

	Observed	UKHI	UKTR	XCCC
1992–93	3.311 (Feb)	4.240 (Feb)	4.405 (Feb)	3.881 (Feb)
1995–96	4.121 (Oct)	4.028 (Oct)	4.242 (Oct)	4.026 (Oct)

Monthly mean effective precipitations and discharges according to the selected scenarios

Table A5. Monthly sums of the effective precipitation (mm) (1992/93)

	IX	X	XI	XII	I	II	III	IV	V	VI	VII	VIII
Obs.	26.20	59.40	25.90	13.70	5.10	4.50	15.50	6.60	27.90	36.40	69.40	50.80
UKHI	24.00	62.10	27.30	17.00	6.00	7.50	16.50	6.70	28.60	37.70	68.80	47.50
UKTR	25.10	70.80	34.90	20.00	6.90	7.60	23.50	8.80	26.00	41.30	81.40	42.20
XCCC	24.00	59.60	31.50	16.40	5.50	6.80	18.10	7.30	29.30	35.40	64.40	50.60

Table A6. Monthly sums of the effective precipitation (mm) (1995/96)

	IX.	X.	XI.	XII.	I.	II.	III.	IV.	V.	VI.	VII.	VIII.
Obs.	42.90	2.60	20.40	15.70	19.70	17.20	13.80	41.90	76.50	66.80	44.40	111.5
UKHI	39.20	2.70	21.40	18.60	21.70	22.20	15.00	44.30	77.50	68.30	44.40	104.4
UKTR	41.20	3.30	25.80	22.50	23.20	23.00	21.40	54.30	70.40	74.70	53.20	94.90
XCCC	39.20	2.60	24.80	18.60	21.60	19.70	15.30	47.30	80.00	66.40	40.70	110.7

Table A7. Monthly mean discharges (m^3/s) (1992/93)

	IX.	X.	XI.	XII.	I.	II.	III.	IV.	V.	VI.	VII.	VIII.
Obs.	11.17	22.58	17.46	8.158	4.562	3.311	5.505	7.530	7.461	18.17	27.21	21.81
UKHI	10.56	23.25	18.15	9.490	4.939	4.240	6.320	8.003	7.641	18.70	27.20	20.84
UKTR	11.14	26.05	21.82	10.83	5.501	4.405	7.907	10.03	7.228	19.33	31.45	20.28
XCCC	10.60	22.43	19.51	9.402	4.756	3.881	6.402	8.257	7.799	18.20	25.52	21.44

Table A8. Monthly mean discharges (m^3/s) (1995/96)

	IX.	X.	XI.	XII.	I.	II.	III.	IV.	V.	VI.	VII.	VIII.
Obs.	18.21	4.121	10.07	6.724	10.49	9.712	6.221	13.47	30.54	29.33	17.95	32.11
UKHI	16.85	4.028	10.43	7.698	11.27	11.33	7.192	14.25	31.23	30.01	18.01	30.57
UKTR	17.66	4.242	12.24	8.984	12.18	11.81	9.059	17.90	30.17	31.64	20.75	29.14
XCCC	16.90	4.026	11.64	7.749	11.40	10.59	6.966	15.17	32.28	29.62	16.99	31.69

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Mapping of mean annual actual evaporation on the example of Zagyva catchment area

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Abstract—A method for the mapping of the mean annual actual evaporation is demonstrated on the example of the Zagyva catchment. For the calculation of the mean annual actual evaporation, the original and modified Turc's formula and two modified types of Budyko's formula were compared. Calculated by the listed formulas, mean annual actual evaporation was compared with its values given by Eq. (1), as a difference between annual precipitation and runoff for drainage areas using, observations within the Zagyva rivershed. The Turc's formula underestimates the evaporation, the best fitting is provided by the modified Budyko's formula (Eq. (10)). Using this formula, the model parameter α was validated by observed runoff data. In the mapping process any drainage area is covered by orthogonal grids, and using the digitized maps of precipitation and temperature, the actual evaporation is calculated for all the grids individually, then step by step a continuous and consistent map of actual evaporation is developed by interpolation between calculated grid values (Fig. 2).

Key-words: mean annual evaporation, Turc's formula, Budyko's formula, water balance, mean annual runoff, grid mapping.

1. Introduction

Some portion of the precipitation over a given area evaporates and returns to the air after a temporal accumulation on land surface or in the soil. The rate of the actual evaporation depends on the capacity of the air to take the vapor, that is, on the potential evaporation and the actual moisture content of soil. The soil moisture content changes mainly with weather conditions, but depends on the soil and vegetation type, too. In many agrometeorological tasks like the calculation of total water demand or irrigation water demand of plants, the soil and plants play very important role in the actual evaporation, so the calculation of the later needs equations, which take into the consideration those factors. Many methods of such type exist in the world. In Hungary one of the most frequently used formula was proposed by *Antal* (1968). In the calculation of the actual evaporation for larger areas and longer time period (for average year), the climate elements play the main role, probably exceeding the role of the soil and vegetation. This follows from that the soil properties and particularly the vegetation usually are in a good relation with the climate. As the purpose of this paper, to demonstrate the mapping of long term average annual actual evaporation using the formulas based on the climate elements is highly suitable.

The map of the spatial actual evaporation of Hungary was firstly constructed as a difference of maps of mean annual precipitation and mean annual runoff (*Szesztay,* 1957). Later this map was refined using a similar method (*Kardos,* 1975). The mapping is based on the mean annual water balance equation

$$P = ET + R, \tag{1}$$

where P, ET, and R are the mean annual precipitation, actual evaporation, and runoff, respectively.

2. Methods and data

Literary sources suggest many different formulas for the calculation of the actual evaporation. In our investigations the Turc's and Budyko's formulas were selected.

Using the *Turc*'s method (1954), the calculation of the actual evaporation consists of two consecutive steps. Firstly it is necessary to calculate the *EPI*-index expressing the spatial variation of the potential evaporation (*EP*) by means of the formula

$$EPI = 300 + 25T + 0.051\,T^3, \tag{2}$$

which needs to have only temperature data. The second step is the calculation of the actual evaporation by the following equation:

$$ET = P[c + (P/EPI)^n]^{-1/n}, \tag{3}$$

where P is a the precipitation, c and n are the model parameters. Using the observed data of precipitation, temperature, and runoff at the 254 different catchment areas under various climate conditions, the model parameters were suggested by *Turc* as follows, $c = 0.9$ and $n = 2$, so the specified equation has the form

$$ET = P[0.9 + (P/EPI)^2]^{-0.5}. \tag{4}$$

Application of the Turc's formula in climate conditions of Slovakia, being a neighbouring country to Hungary, showed that the actual evaporation was gradually underestimated using this formula in its proposed form (*Parajka* and *Szolgay,* 1998). Trying to adapt Eqs. (2) and (4) to Slovakian climatic conditions the model parameters were refined. For the calculation of the *EPI*-index and actual evaporation

$$EPI = 237.38 + 40.67\,T + 0.0864\,T^3, \tag{5}$$

and

$$ET = P[0.9 + (P/EPI)^{2.67}]^{-1/2.67} \tag{6}$$

were proposed. The fact that Turc's formula underestimates the evaporation was shown in Hungarian drainage basins using precipitation and runoff data (*Nováky,* 1991).

The *Budyko*'s (1948) formula was selected for further investigation. The formula is

$$ET = P[1 - \exp(-r/LP)], \tag{7}$$

where r is the net radiation, L is the latent heat, P is the precipitation, and r/LP is the aridity index. The Budyko's formula is the improved version of the relationship between precipitation and runoff discovered by *Schreiber* (1904) nearly 100 years ago. In our investigations the formula was modified, when the net radiation was substituted by the temperature or climate index given by any combination of the temperature and precipitation.

In the first version of the modification, using the equation

$$r/L = P(\ln P - \ln R) \tag{8}$$

given by the rearrangement of the Budyko's formula, the values r/L were calculated for observed drainage areas, and a relationship between calculated values r/L and the temperature was developed. (The value r/L can be accepted as one of the variant of the *EPI*-index.) This relationship was formulated as follows:

$$r/L = m(T - T_o)^n, \tag{9}$$

where T_o is a threshold temperature, below it the (potential) evaporation can be accepted as it equals to 0, m and n are the model parameters to be validated. In an earlier study for the Rába catchment, these parameters were validated: m = 15.95, n = 1.66 (*Bergman et al.*, 2001).

The other adapted version of the Budyko's formula was elaborated in the process of mean annual runoff mapping (*Nováky*, 1985) at the time of the compilation of National Master Plan of Water Management in Hungary (*OVH*, 1984). The value r/L (practically the net radiation) was substituted in the proposed method by the value of the pan evaporation using a correction factor (α), taking into consideration that both net radiation (r/L) and pan evaporation (E_{pan}) are in close correlation with potential evaporation, consequently also with each other. The relationship between net radiation and pan evaporation is described with equation

$$r/L = \alpha E_{pan}. \tag{10}$$

In the next step, using the observed values of 13 pan evaporation stations under different climatic conditions in Hungary, the relationship between pan evaporation (E_{pan}) and a simple climate index, formulated as the rate of the mean annual temperature to the mean annual precipitation, was developed, as follows

$$E_{pan} = 36{,}400TP^{-1} + 104 \tag{11}$$

(*Nováky*, 1985). Substituting the Eqs. (10) and (11) to Eq. (7), a formula

$$ET = P\{1 - \exp[-\alpha(36{,}400TP^{-1} + 104)P^{-1}]\} \tag{12}$$

is given for the calculation of the actual evaporation. The formula needs the validation of model parameter α. Validating this parameter the calculation of the actual evaporation requires the knowledge only of mean annual precipitation and temperature.

The validation of the model parameter α is possible by the optimisation of values α calculated individually for the observed drainage areas, where the values of mean annual precipitation, mean annual temperature, and mean annual runoff exist for the same periods of observations used in Eqs. (1) and (12). The optimised value of model parameter α for the whole country without any regionalisation is equal to 2 (*Nováky,* 1988). While parameter α is in the exponent of Eq. (12), naturally the calculation of the actual evaporation is highly sensitive to its value.

For mapping the actual evaporation, a formula is selected, which allows to calculate its value being best fitted to actual evaporation values calculated as the differences between precipitation and runoff in observed drainage areas.

Besides the model parameter, all formulas for calculation of the actual evaporation need essentially to have two meteorological elements, the precipitation and temperature, so the mapping of the actual evaporation also needs the maps of mean annual precipitation and mean annual temperature. During the process of mapping, all of the catchments are discretized using the net of orthogonal grids. The mean annual precipitation and temperature are determined for all grids separately, that means that continuous maps are digitized according to the space interval of grids. Using the digitized maps of precipitation and temperature, the actual evaporation is calculated for all the grids individually and a continuous and consistent map of actual evaporation is developed by interpolation between these calculated grid values.

Following the calculation of the actual evaporation, the grid runoffs are also calculated as the difference between grid precipitation and evaporation, further by summing up the grid runoffs for the grids belonging to any river section with runoff observations, the catchment runoff is also given. In this process according to the rows and columns of orthogonal grids, a matrix notation can be adopted to identify the grids (*Fig. 1*).

The grid in row i and column j can be denoted by e_{ij}, the calculated runoff by R_{ij}. Let N_{ij} is the set of grids upstream to e_{ij}, so grids belonging to N_{ij} compose the catchment upstream to grid e_{ij}. If a grid e_{kl} belongs to set of grids N_{ij}, the grid runoff R_{kl} reaches the grid e_{ij}, otherwise it does not. The catchment runoff in grid e_{ij} can be calculated as

$$R^* = \Sigma\Sigma\ \delta_{kl} R_{kl}, \tag{13}$$

where δ_{kl} is a Kronecker function, its value is 0, if $e_{kl} \notin N_{ij}$, and 1, if $e_{kl} \in N_{ij}$. Eq. (13) provides a connection between the grid runoff R_{ij} and catchment runoff R^*_{ij}. In the section of observed catchments, the calculated catchment runoff R^* can be compared by its observed value, and on the basis of differences the suitability of the model can be evaluated.

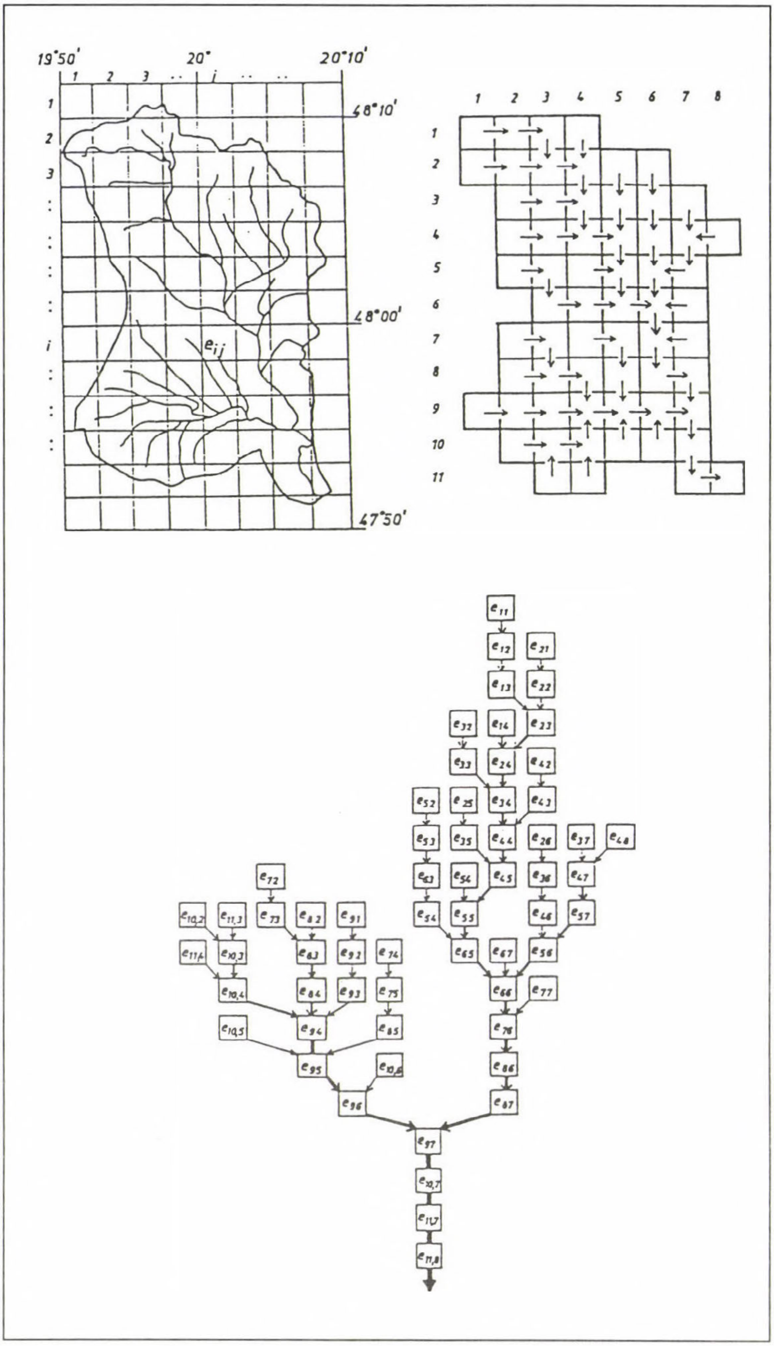

Fig. 1. Spatial discretization of the catchment.

3. Data sources

The map of the actual evaporation was constructed for the Zagyva catchment, where there are some hydrological stations with runoff observation. Using runoff data, the formula mostly available for calculation of actual evaporation can be chosen and its model parameter can be calibrated (*Table 1*). Runoff data were taken from the study describing hydrological conditions of the Zagyva catchment (*Nováky,* 1985) compiled in the process of the elaboration the National Master Plan of Water Management in Hungary (*OVH,* 1984).

Table 1. Mean annual water balance of subcatchments in the Zagyva River basin

Watercourses	**Hydrological stations**	**Catchment area km²**	**Precipitation P (mm)**	**Temperature T (°C)**	**Runoff R (mm)**
Zagyva	Nemti	149	631	8.7	68
	Pásztó	488	636	8.6	94
	Szentlőrinckáta	1955	600	9.1	54
	Jásztelek	4207	605	9.2	56
Tarna	Verpelét	574	645	8.4	79
	Jászdózsa	1810	621	9.1	64
Galga	Galgamácsa	242	585	8.9	54
	Hévízgyörk	416	577	9.0	64
Tápió	Tápiószele	771	562	9.9	44
Bene-patak	Nagyfüged	133	671	8.9	105
Gyöngyös-patak	Jászárokszállás	290	637	9.1	93

In the calculation of mean annual runoff, the period of observation were very different, and its length ranged from 9 (Zagyva, Nemti) to 76 (Zagyva, Jásztelek) years. Therefore, the mean annual precipitation and temperature were calculated for the same period from which the mean annual runoff was originated. Precipitation data before 1970 were taken from the collection of precipitation data compiled by *Hajósy et al.* (1975), and after 1970 from the Hydrographical Bulletins of VITUKI. The mean annual temperature for given periods was taken with appropriate correction from the Climatic Atlas compiled by the Hungarian Meteorological Service (*Kakas,* 1960). The correction factors were calculated using the temperature data of meteorological stations having long term observations.

After selecting the formula mostly available for calculation of actual evaporation, the mapping of the latter was made using digitized maps of precipitation and temperature of the mentioned Climatic Atlas. The grid of the digitization has a size of 6 km × 6 km.

4. Results

To choose the best fitting formula for calculation of the actual evaporation, the original and modified Turc's formulas, also the modified Budyko's formulas were used and compared. Before using the modified Budyko's formula, it was necessary to calibrate the parameters m and n in Eq. (9). Based on the observed data of 11 catchments, the model parameters were calibrated, $m = 10.4$ and $n = 1.85$, so the formula for the calculation of the actual evaporation specified for Zagyva catchment was given as

$$ET = P\{1 - \exp[-10.4(T + 5)^{1.85}P^{-1}]\}. \tag{14}$$

For the calculation of the actual evaporation by Eq. (12), the model parameter α was selected so that the differences between calculated annual evaporation and annual evaporation given by Eq. (1) for the observed catchments were minimal. The calibrated model parameter $\alpha = 2.14$, and the actual evaporation can be calculated by specified formula

$$ET = \{1 - P\exp[-2.14\alpha\,(36{,}400TP^{-1} + 104)P^{-1}]\}. \tag{15}$$

The values of the EPI index calculated by Turc's, modified Turc's, two types of modified Budyko's formulas range in order of these formulas in intervals of 540–596 mm, 630–723 mm, 1265–1540, and 1237–1595 mm. In all cases the minimum value is given for the catchment of Tarna at Verpelét and the maximum for Tápió at Tápiószele. The potential evaporation indices given by Budyko's method are significantly higher than values calculated by Turc's formulas. The significant difference can be explained by fact that the EPI index calculated by Budyko's formulas really expresses the energetically possible maximal evaporation, which is formed in case when the whole net radiation energy is directed to evaporation. In fact, in climatic conditions similar to Hungary, only about 70% of net radiation is used for evaporation (*Péczely,* 1981). Taking into consideration this rate, the values of the EPI index range about between 830–1050 mm, which are higher than values given by Turc's formulas, but are closer to the potential evaporation typical for Hungarian climatic conditions.

The actual evaporation was calculated by four different equations for 11 observed drainage basins (*Table 2*). The results were compared to the values of actual evaporation given by Eq. (1) as a difference between observed precipitation and runoff. The average error is 21.8% using the original Turc's formula, 8.5% for the modified Turc's formula, and less than 2% for modified Budyko's formulas. A similar order is received when formulas are compared to the errors of calculated runoff, but the errors of runoff calculations are significantly higher (*Table 3*). The average error is 178% using the original Turc's formula, 69% using its modified version based on Slovakian runoff data, and about 12% using whichever modified Budyko's formula. The Turc's formula significantly underestimates the actual evaporation for Hungarian climatic conditions.

Table 2. Comparison of the formulas for the calculation of actual evaporation

Water-courses	Hydrological stations	Observed actual evaporation mm	Calculated actual evaporation by							
			Turc's formula Eq. (4)		Modified Turc's formula Eq. (6)		Modified Budyko's formula Eq. (14)		Modified Budyko's formula Eq. (15)	
		E	E_c	%	E_c	%	E_c	%	E_c	%
Zagyva	Nemti	563	427	24.1	501	11.0	553	1.8	550	2.3
	Pásztó	542	427	21.2	500	7.7	554	2.2	550	1.5
	Szentlőrinckáta	546	423	22.5	494	9.5	541	0.9	542	0.7
	Jásztelek	549	426	22.4	499	9.1	546	0.5	546	0.5
Tarna	Verpelét	566	427	24.6	500	11.7	554	2.1	550	2.8
	Jászdózsa	557	430	22.8	504	9.5	555	0.3	552	0.9
Galga	Galgamácsa	531	415	21.8	483	9.0	527	0.7	532	0.2
	Hévízgyörk	513	413	19.5	481	6.2	523	1.9	529	3.1
	Tápiószele	518	419	19.1	488	5.8	526	1.5	529	2.1
Bene-patak	Nagyfüged	566	442	21.9	521	8.0	582	2.8	568	0.4
Gyöngyös-patak	Jászárokszállás	544	435	20.3	512	5.9	566	4.0	559	2.8
Mean				**21.8**		**8.5**		**1.9**		**1.6**

Table 3. Errors of calculation of mean annual runoff using different formulas for the calculation of mean annual actual evaporation

Water-courses	Hydrological stations	Observed mean annual runoff mm	Mean annual runoff calculated by							
			Turc's formula Eq. (3)		Modified Turc's formula Eq. (4)		Modified Budyko's formula Eq. (7)		Modified Budyko's formula Eq. (10)	
		R	R_c	%	R_c	%	R_c	%	R_c	%
Zagyva	Nemti	68	204	200.0	130	91.2	78	14.7	81	19.1
	Pásztó	94	209	122.3	136	44.6	82	12.8	86	8.5
	Szentlőrinckáta	54	177	227.8	106	96.3	59	9.3	58	7.4
	Jásztelek	56	179	219.6	106	89.3	59	5.4	59	5.4
Tarna	Verpelét	79	218	175.9	145	83.5	91	15.2	95	20.3
	Jászdózsa	64	191	198.4	117	82.8	66	3.1	69	7.8
Galga	Galgamácsa	54	170	214.8	102	88.9	58	7.4	53	1.8
	Hévízgyörk	64	164	156.2	96	50.0	54	15.6	48	25.0
	Tápiószele	44	143	225.0	74	68.2	36	18.2	33	25.0
Bene-patak	Nagyfüged	105	229	118.1	150	42.8	89	15.2	103	1.9
Gyöngyös-patak	Jászárokszállás	93	202	117.2	125	34.4	71	23.7	78	16.1
Mean				**179.6**		**70.2**		**12.8**		**12.6**

The modified Budyko's formula (Eq. (14)) was accepted for mapping the actual evaporation in the Zagyva catchment (*Fig. 2*). According to the map, the mean annual actual evaporation ranges from 490 to 560 mm. The lowest values are typical for flatland parts of Zagyva close to the mouth of the river, and the highest values are characteristic for mountainous parts of the catchment. The actual evaporation in the Zagyva catchment is determined and limited mainly by the rate of annual precipitation, so the actual evaporation in all catchments is lower than the potential evaporation.

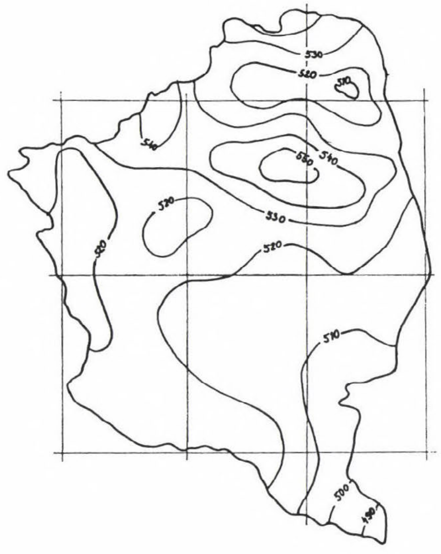

Fig. 2. The map of mean annual actual evaporation for the Zagyva catchment.

5. Discussion

The spatial variation of the actual evaporation reflects the spatial variation of two main meteorological elements, the precipitation and temperature, but does not reflect the mosaic-like spatial variation of land surface properties. For a given locality the map can provide only a first approximation of the actual evaporation, and for a given farm-sized cultivated land, to calculate the actual evaporation, the method proposed by *Antal* (1968) remains as one of the most applicable methods. Nevertheless, there is a possibility for the refinement of the proposed map with taking into consideration at least the spatial differentiation of mean land use categories within the catchment. Further investigations to that direction may lead to the desired results.

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International efforts for drought mitigation

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Abstract—Drought has harmful effects not only on plants and animals, but on all living organisms and the whole human society as well. In the light of this recognition, many organizations and international associations started to deal with the methods of drought mitigation not only in agriculture, but also in wider environment, including socio-economic effects. The paper gives an outlook on some of the most important international and national events concerning drought mitigation in the last few years. Detailed information is given on the activity of ICID European Regional Work Team on Drought (ERWTD) and its guide for drought mitigation strategy. We point out the importance of the construction of the European drought sensitivity map as well as the new evaluation and synthesis of the results of the research work related to different drought effects. For solving all these multi-disciplinary tasks, a well coordinated international co-operation is needed with active participation of the experts of the most interested countries, especially in the frame of the UNCCD.

Key-words: drought mitigation, international co-operation.

1. Introduction

In the last period of time it could be clearly recognized that the rhythm of the changes in the natural and social environment became more and more rapid, and the effects of the changes became increasingly bright, complex, and permanent. All these processes influence the everyday and future human life more and more directly compared to previous time. Dryness and drought are results of the special interaction between natural and social environment. Man and society play active as well as passive roles in this process, which influences the global development of a region. In the last few decades it became also obvious that effects of drought can cause damages not only in agriculture and plant production, but, at the same time, in all living organisms including domesticated and wild species of plants, animals, microorganisms, and also human beings. Consequently, there is a global need to find out the means and measures against the harmful effects of drought, and to create some variables in space and time for influencing the preparedness of the whole society, including politics, economy, ecology, justice, and ethics, as well as the private and public behavior for the sustainable development of the society (*Vermes,* 1997).

2. The role and activity of ICID

Recognizing the above mentioned facts, some of the international organizations, especially ICID, the International Commission on Irrigation and Drainage, and its working groups started to deal with drought problems from the year of 1992. ICID — focusing on irrigation, drainage, and flood control questions — is practically active in all problems related to agricultural water management. It has been established as a scientific, technical, professional and voluntary, international non-profit and non-governmental organization, and dedicated inter alias to enhance the worldwide support of food and fiber for all people. ICID strives to achieve this by improving water and land management, and the productivity of irrigated and drained lands through the appropriate management of water, environment, and the application of irrigation, drainage, and flood control techniques. ICID is interested in matters relating to the planning, financing, and economics of the mentioned fields.

In the frame of ICID several working groups are operating, among which two are closely connected with drought problems: (1) the Working group on Irrigated Agriculture under Drought and Water Scarcity (WG-IADWS), and (2) the European Regional Work Team on Drought (ERWTD) in the frame of the European Regional Working Group (ERWG). Hungary is represented in both, but we are more active in the last one, which was established on a Hungarian proposal. This is a task force of the ERWG to help and coordinate member countries interested in the fight against drought damages.

3. An ICID guide on drought mitigation

One of the main tasks of the ERWTD was to compile a guide — entitled How to Work out a Drought Mitigation Strategy — for summarizing all the necessary information which is important to drought prevention, which should be taken into consideration during the elaboration of a drought mitigation strategy, and which can be used in the case of the occurrence of drought for reduction and/or tolerance of the caused damages. The guide was completed on the basis of several earlier initiatives taken and proposals made at former ICID meetings, and by individual experts, members of the ERWTD and other organizations (*Vermes,* 1998).

In the introductory part of the Strategy, it is important to determine the clear definitions of the relevant concepts concerning drought. There should be distinguished the differences between, e.g., dryness and drought, and here should be formulated the aim of the strategy as well. After this, some drought occurrences have to be quoted from different parts and countries of the world. The ICID guide gives examples from recent drought events from Hungary, Yugoslavia, Romania, and Croatia.

The following part should deal with the drought analysis, examining first the climatic conditions and hydrological factors causing drought, secondly quoting the human made effects, especially the agronomic circumstances, and thirdly analyzing the role of soil types and soil conditions in drought occurrence.

The next part of the Strategy gives the inventory of the harmful impacts and damages of drought, specifically in the given country or region. Among these, in first case the directly effected water users should be introduced and analyzed, namely the different branches of agricultural production, the different branches of industry, especially food industry, and the different services, which mostly suffer from drought. Also the environmental impacts on natural resources, habitats, and ecosystems, natural protected areas, national parks should be determined, and combined environmental effects studied in this part of the Strategy.

Among indirect effects, the trade conditions, especially the import-export relations, financial affairs, and social impacts on public health, on employment/unemployment, on politics and foreign affairs, and on tourism should be evaluated. Some beneficial effects of drought should be examined, too, e.g., mosquito reduction, reduced cost for clearance of snow during winter drought, improvement efficiency in water use and water quality control, and the control of overproduction in agriculture. It has to be strongly pointed out that intensive research work is urgently needed in the field of further beneficial effects of drought.

Also the measures taken so far against harmful impacts of drought damages should be taken into account. In most cases unfounded, not quite well consolidated, highly improvised measures have been taken during — or even after — recent drought events. The steps were mainly succeeding and not preventing the damages, the actions were mainly stop-gap type actions with partial effects. Also the attitude and reaction of the society on drought should be analyzed, which is in most cases strange and not understandable. In this respect the role of the media has to be pointed out as well.

In the most important part of the Strategy the means and methods of the complex fight against drought damages have to be discussed. First the assessment and forecast of drought events should be studied, determining the calculation methods and indices used in the forecasting process. Establishment of a monitoring system in meteorology and water management, as well as in other branches of the economy is an important step. To draw up management plans in agriculture and forestry, it can help to mitigate drought damages in these sectors. Extension services and media programs should be used for increasing awareness of farmers and other professionals on potential drought events. Key question is the determination of prevention methods, damage reduction instruments, and toleration of drought damages. Among prevention methods, supply-oriented and demand-oriented measures, and impact- and losses minimization methods can be distinguished.

The instruments of damage reduction are manifold. There are agricultural methods, like optimum land use, better crop pattern, changes in the elements of agro-technology, complex land reclamation methods, improving plant breeding for better drought tolerant crop varieties, elaborating of authority system for control, regulation and support of drought damaged farmers, determining the sources and means of compensation or disaster aid for those who suffer from great income losses caused by drought, developing special insurance system for drought damages, establishing relief funds and/or guarantee funds for those who lost their yields or properties because of drought, etc.

Toleration means consciously counting on some risk, limitation or losses of yield, or income. Therefore, it is important to determine the tolerable level of losses by risk assessment, to set up priority lists for toleration of deficiencies caused by drought, and to draw up case studies on farm and/or company level for determining toleration, prevention, and reduction measures for damage minimization.

Key questions are the organization and coordination questions, too, among which, first of all, the followings should be answered:

- How to organize the formulation and interpretation of the national drought strategy?
- How to organize the determination of the tasks of different participants?
- How to organize the compilation of the action program on the systems of measures?
- How to establish the National Drought Commission for the realization of the Strategy?

Especially the latest has of great importance, the detailed explanation of which should be included in the Strategy.

As far as the *international cooperation* is concerned, the role of the neighboring countries have to be cleared, where the potential collaboration of the countries concerned should be organized. The contact with international organizations, like UNCCD, ICID, or others, should be determined either.

Also, the needs for research and development, as well as education and training should be analyzed in the Strategy. In most cases new types of research work are needed for summarizing and synthesizing the results gained so far, for systematization of the results and experiences according to a new classification of the themes: ecological, technological, economical, and sociological questions of drought should be differentiated. It is important to find out new sources and funds for financing drought research.

Educational and training programs should be started for better understanding of drought problems among the public, to increase public awareness and preparedness for the coming drought events. Special media awareness program is necessary to convince journalists about drought problems and mitigation possibilities, and long-term educational programs should be established for all groups and economic sectors.

A glossary of terms and indices, and a collection of recommendations for potential risk reduction actions are advised to attach in appendices to the Strategy.

One of the major *future tasks* of the cooperation under the guidance of the ERWTD of ICID could be the preparation of a Network for Central and Eastern European Countries, and to draw up the European drought sensitivity map which can be a good basis of any further actions against harmful drought effects.

4. Fight against drought in the USA

Drought was one of the most underestimated and least understood natural disasters in the United States, too. Based on the recognition, that many of the worst effects of drought can be reduced or even eliminated when proper mitigation measures are introduced in advance of drought events, American climatologists and other scientists in the Institute of Agriculture and Natural Resources at the University of Nebrasca-Lincoln decided to establish the *National Drought Mitigation Center* in 1995, and they founded later the *International Drought Mitigation Center* as well (*Wilhite,* 1998). Both centers are going to help people and institutions develop and implement measures to reduce societal vulnerability to drought. They stress prevention and risk management rather than crisis management. This approach promotes self-reliance to achieve greater resilience to drought, and this way their activities are of great interest for us and give an example to be followed.

In 1998 the Congress of the USA adopted the *National Drought Policy Act* to establish an advisory commission to provide advice and recommendations on the creation of an integrated, coordinated Federal policy designed to prepare for and respond to serious drought emergencies. Main task of the *National Drought Policy Commission* is to conduct a thorough study and submit a report to the Congress on national drought policy in accordance with the duties and tasks determined by the Act (*National...*, 1998).

In the same year the *Western Drought Coordination Council* was organized as an intergovernmental forum that focuses on drought preparedness in the Western United States, especially in New Mexico, Texas, and Colorado. Drought experiences in the West and Southwest in the former years highlighted the need for a long-term, region-wide drought planning and streamlined access to government services. The Western Governors' Association responded to those needs when spearheading it for the Council through a Memorandum of Understanding with representatives of federal, tribal, and local governments. The United States Department of Agriculture (USDA) serves as the lead federal agency, and a Steering Group implements the Council's work plan. The Steering Group oversees four working groups:

- Monitoring, Assessment, and Prediction,
- Preparedness and Mitigation,
- Response, and
- Communications (*Western...*, 1998).

The working groups are staffed by volunteer professionals from federal, state, local, and tribal entities. The Council's purpose is to reduce the effects of future droughts in the Western United States, specifically its goals are

- to foster better intergovernmental coordination and communication on drought planning,
- to help state, local, and tribal governments develop drought preparedness and mitigation plans,
- to contribute an efficient drought monitoring and information delivery system so that decision makers learn about low water supplies and drought prospects as soon as possible, and
- to heighten awareness of drought management issues and promote efficient use of water.

Due to the well organized actions, nowadays most of the States have a drought plan elaborated on certain level, and some of them are even under revision and redesigning.

5. Drought mitigation actions in the United Kingdom

In the frame of the International Decade for Natural Disaster Reduction (IDNDR), the National Committee founded a special working group, the *Drought Mitigation Working Group* (DMWG) in the UK for the organization and coordination of different activities against drought damages and water shortages, and for a more effective and rational use of water resources. The water industry's response to the severe drought events has been to reduce water demands by a mixture of appeals for voluntary restraint, hosepipe bans, and drought order, and to increase supply by bringing forward works which increase the flexibility of supply and allow transfer of water from areas of surplus. The Department of Environment maintained a coordinating role holding monthly meetings with the National Rivers Authority and water companies. Drought orders are issued by the Secretary of State, providing the powers to ban non-essential uses of water, and make changes in abstraction licence conditions. Some conservation bodies, like the Royal Society for Nature Conservation (RSNC), have suggested that legislation should be changed to allow drought orders to be made on environmental grounds (*Dealing with ...*, 1993).

Under a drought order, water companies may first seek to have their conditions of abstractions changed, e.g., a drought order may permit the reduction of the "compensation water" released from reservoirs, allow increased abstraction from some sites, or reduce the minimum river flow above which abstractions are allowed. The National River Authority is unlikely to support such an application unless a hosepipe ban is already in operation.

Other powers which may be sought under the drought order are restrictions on the non-essential use of water, specified by the Secretary of State. These restrictions can include the washing of buildings, watering public gardens, sports grounds, and golf courses, automatic car washes, the filling of swimming pools, ornamental ponds, and the operation of fountains. Drought orders may also be used to change the conditions in discharge consents, but this is not commonly done. The National River Authority also has powers to impose restrictions on abstractions made for spray irrigation — except for greenhouses and container-grown plants. In order to assist farmers, a system of alerts is in operation in the Anglian region, which provides warning of impending partial and total bans of abstraction. Water restrictions are monitored by the Office of Water Services (OFWAT), and form one of the levels of service indicators for the water companies.

6. UN Convention to Combat Desertification and Drought

Because drought and natural processes do not consider political boundaries, the influences from the neighboring territories should be taken into consideration. A good drought strategy deals with international relations and counts on international cooperation in the fight against drought damages. This is promoted even by the United Nations trying to help the countries involved, and call the attention of the states and governments on a better international cooperation in this field. This is declared in the *United Nations Convention to Combat Desertification (UNCCD) in those countries experiencing serious drought and/or desertification, particularly in Africa (United...,* 1999). The Convention was adopted and open for signing in 1994, Paris, and up till now many countries — altogether 115, including Hungary — joined as full Parties. The main objective of this Convention is to fight against desertification and drought through effective actions at all levels, supported by international cooperation and partnership arrangements, in the framework of an integrated approach which is considered with Agenda 21, with a view to contributing to the achievement of sustainable development in affected areas. Achieving this objective will involve long-term strategies that focus simultaneously, in affected areas, on improved productivity of land, and the rehabilitation, conservation and sustainable management of land and water resources, leading to improved living conditions, in particular at the community level.

Beside of determining the general provisions, the Convention deals in a separate part with *action programmes, scientific and technical cooperation, and supporting measures,* as one of the major fields of implementation of the goals and objectives. All statements and recommendations of the Convention should be taken into consideration and are desired to be built into the national drought strategies and action plans. In the Carpathian Basin the *Fifth Annex of the Convention* is of great interest, adopted in the Conference of the Parties in 2000, which is devoted to the regional implementation of the Convention in the Central and Eastern European countries. Fulfilling the aims of the Convention, Hungary should work actively in the preparation and implementation of national action programmes on drought mitigation.

7. Conclusions

The paper dealt with only some of the very important international and national efforts and events concerning drought mitigation. There was no place to mention all of such efforts, however, in many other countries more and more efforts are made against drought damages, e.g., in Australia, countries in the far and middle East, as well as the countries in Central and Eastern Europe. In Hungary, the National Drought Strategy is under preparation and the organization of the National Drought Committee was started. All these activities show that a well coordinated, complex work is necessary in all effected countries for an effective fight against drought, and international cooperation is needed either in research work or in practical actions and organizational fields for the minimization of drought damages.

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Characteristics of the economy of water supplies in an oak forest

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Abstract—The components of the water balance in an oak forest at Síkfőkút were examined in the frame of the *Síkfőkút Project* between 1978 and 1995. The sample site, which has an area of 1 hectare, is situated on a slope of 2–3 degrees in the Bükk Mountains of Hungary at a height of 270–280 meters above sea level. It has a southern exposure. At the beginning of the project, the average height and age of the trees were 21.5 meters and 71 years. The soil is brown forest soil with a vast humus layer. The humus content is 2–5%, field water capacity is about 30 percents. Soil samples were taken by soil strata and decades repeatedly three times. Soil moisture content was determined using gravimetry. In the growing season the soil moisture content decreases in the upper 60 cm layer of the forest soil from April to August. Between 1978 and 1995, soil moisture and precipitation showed a decreasing trend in the sample area. The most humid is the upper layer of the soil (0–5 cm, 4–5% humus). Less water can be found in the next layer between 30–50 cm, since roots take up most of the water in that depth. Moisture content of the upper 60 cm of the soil follows the changes of the amount of precipitation and soil moisture in the growing season. The same is true for the course of mean annual water content of the different horizons of the forest soil. Mean annual interception of the forest is 17%. It is 9% when the trees are leafless and 18% when they are leafy. In the growing season the monthly maxima of actual evapotranspiration are 50–100 mm, while minima are between 30 and 60 mm. The total amount of it is between 297 and 422 mm in the growing season. The Síkfőkút forest consists mostly (84.5%) of sessile oak (Qercus petrea), which had decayed dramatically between 1978 and 1995. Possible reasons for this are the decrease of the pH, bacteria, and fungi content of the soil, the decrease of precipitation amount and soil moisture content, and the increase of annual mean temperature. These changes and especially the complex impacts of the chain of these effects have lead to the decay of the trees.

Key-words: effective precipitation, soil moisture, water-balance, surface runoff, actual evapotranspiration.

1. Introduction

Productivity of a forest, condition of the trees, and their resistance to harmful effects from outside depend considerably on the economy of water supplies of the forest. The examinations carried out on the water balance of the Síkfőkút forest are important, because 50% of the sessile oak trees in that forest had decayed between 1978 and 1993. The decay of the forest has decreased and finally stopped only in the previous years due to more precipitation. The simplified water balance equation of the forest,

$$\pm\Delta W = P - ET \pm R - F - I, \tag{1}$$

describes that the change of the moisture content in the root zone ($\pm\Delta W$) is a function of the amount of precipitation reaching the foliage of the forest (P), the evapotranspiration of the forest (ET), the surface runoff ($\pm R$), the infiltration into the deeper layer (F), and the amount of precipitation retained by the whole forest vegetation (trees, shrubs, and undergrowth), the interception (I).

Since during the growing season the forest can gain water from the soil moisture retained in the root zone, it is important to know what kind of changes occur in time in the moisture content of the deep layer of the forest soil within this period.

It is well known that due to global changes that affect the biosphere (acid precipitation, green house effect, etc.), from the beginning of the 80's, interest have been focused on long term biomonitoring researches worldwide. These studies concentrate on the detection and exploration of slow, sometimes hardly traceable processes of ecosystems. In Hungary, this kind of long term biosphere research project, the so-called "Síkfőkút Project", was started by the Department of Ecology of the Kossuth Lajos University of Debrecen in 1972, in the frame of IBP, later the MAB programs (*Jakucs,* 1973, 1985). It was a complex ecological and meteorological study of a climate-zonal Austrian oak-oak joining near the city of Eger. The Department of Meteorology of the Kossuth Lajos University joined the project establishment intensely in 1977. By that time the department had installed a 25 meters tall meteorological tower and a digital data logger with 80 channels (*Justyák,* 1987).

The environmental-biological model site has an area of 1 hectare and it is situated on a slope of 2-3 degrees, with southern exposure. It is placed near the town of Eger in the Bükk Mountains, at a height of 270-280 meters above see level. It is a brush-wood origin, homogeneous Austrian oak-oak joining. The average height of the trees was 19 meters in 1972, at the beginning of the project. It reached 21.5 meters in 1978. In 1972 the forest was 65 years old. There are a high (3–4 meters) and a low (1–2 meters) shrub level in the forest.

About 60% of the mass of the forest foliage can be found at a height between 15 and 20 meters.

2. Soil moisture

The type of the soil in the model area is lessivated brown forest soil. Its characteristics play an important role in the water balance of the root zone. Its main characteristics are:

- The soil is covered with 2–3 cm thick layer of litter in the forest.
- The 0–10 cm layer (subhorizon A_1) is dark brown loam with 2–3% humus. It is densely laced with the roots of the shrubs.
- The 10–25 cm layer (subhorizon A_2) is brown compact polyederic loam.
- The 25–40 cm layer (subhorizon A-B) is brown compact clay-loam.
- The 40–70 cm layer (subhorizon B_1) is slightly reddish brown compact polyederic loam. It is densely laced with the roots of the trees.
- The 70–90 cm layer (subhorizon B_2) is clay-loam, while horizon C is compact loamy clay without structure.

The pH of the soil is between 4 and 6, hygroscopicity (hy) 3–5%, while humus content is 2–5%. The most important soil characteristics are summarized in *Table 1*.

Table 1. The most important soil characteristics on the area of Síkfőkút Project

Depth (cm)	**pH**		**Acidity**		**hy (%)**	**Humus (%)**
	H_2O	**KCl**	**Y_1**	**Y_2**		
0–10	6.3	4.6	28.1	1.1	3.1	5.4
10–25	4.7	3.8	38.3	7.2	2.9	2.7
25–40	5.2	4.0	22.0	4.4	3.5	2.2
40–70	6.2	6.0	15.3	-	4.7	1.8

The pH of the different soil layer profiles is characteristic for the lessivated brown forest soil. The values show considerable *acidification*, especially in subhorizon A_2 (10–25 cm), where pH decreases to 4.5. From subhorizon B_1 (40–70 cm), pH increases again.

Bulk density (BD) of the soil increases with the depth from 1.0 to 1.5. *Field capacity* of the soil (FC) is about 30%, while *wilting point* (WP) is between 12 and 18%, therefore the ratio of *available soil water* (AW) is high in the soil. The soil moisture characteristics in the different layers are summarized in *Table 2*.

Bulk density (BD), fild capacity (FC), minimal water capacity (WCmin), and wilting point water content (WP) increase with the depth, while saturation water content (SWC) decreases with increasing dept. Bulk density values were taken into consideration when calculating on mm values from soil moisture content.

Table 2. Bulk density (BD), field capacity (FC), and available water (AW) in the soil layers

Depth cm	BD g/cm^3	FC v/v%	FCmax v/v%	FCmin v/v%	AW v/v%
0–10	1.02	38.5	60.1	29.2	12.4
10–25	1.37	42.1	51.0	30.3	11.6
25–40	1.44	40.8	47.6	29.2	14.0
40–70	1.47	42.7	48.2	30.8	18.8

Particle size distribution (%) of samples in the soil profiles are shown in *Table 3.*

Table 3. Particle size distribution in the soil layers

Depth cm	1–0.25 mm	0.25–0.05 mm	0.05–0.01 mm	0.01–0.005 mm	0.005–0.001 mm	<0.001 mm
0–10	4	3	33	13	26	21
10–25	2	4	30	13	24	27
25–40	2	3	26	11	22	36
40–70	3	2	25	8	19	43

The ratio of the clay fraction is 43% in the depth of 40–70 cm. There is the half of it (21–27%) in the upper subhorizons (0–10 and 10–25 cm), while humus content is higher (*Stefanovits,* 1985).

Summarizing the above-mentioned facts, it can be stated that the physical characteristics and structure of the soil are important, because the soil of the forest is rich in colloids (clay). The productive layer of the soil is thick and free of stones and pebbles, therefore, if there is adequate amount of precipitation, the soil is able to receive and store a significant amount of water.

In order to determine the soil moisture content, soil samples were taken during the growing seasons between 1978 and 1995 by drilling repeatedly three times from each soil layer from the depths of 0–5, 5–10, 10–20, 20–30, 30–40, 40–50, 50–60 centimeters. Soil moisture content was determined by gravimetric method. Soil moisture content in percentage of the weight of the dry soil was determined for each growing season (from April to October), on the base of three samples taken by decades (monthly averages are calculated from the values of the three decades).

Soil is able to retain and store a certain amount of water against the force of gravity. This amount of water is called *soil moisture content*. The difference between field capacity and wilting point water content is only a part of this amount, which is available for the trees and shrubs of the forest. This amount of water is usually called *available* or plant extractable *water*. The soil water content will be discussed here as *soil moisture content*.

Mean moisture content of the soil (%) in the Síkfőkút forest for the months and growing seasons are shown in *Table 4*. ($P_{eff.aver.}$ is the amount of effective precipitation, that reaches the soil falling through the foliage.)

Table 4. Mean soil moisture content (%) and effective precipitation (mm) in the Síkfőkút forest for the months and growing seasons

Years	Apr	May	Jun	Jul	Aug	Sep	Oct	Apr–Oct	$P_{eff.}$ Apr–Oct
1978	24.2	29.1	25.6	26.9	21.4	17.3	17.0	23.1	376
1979	29.4	25.5	23.3	21.7	19.1	21.6	17.6	22.6	265
1980	30.7	28.5	28.4	27.2	21.9	20.4	21.6	22.5	361
1981	26.8	26.1	22.5	21.1	17.6	19.3	20.9	22.0	288
1982	27.7	26.6	23.4	20.1	15.5	15.0	17.2	20.8	236
1983	27.5	26.7	22.6	15.0	14.7	17.5	17.0	20.1	276
1984	21.3	27.6	26.3	16.0	15.5	15.7	20.2	20.4	319
1985	27.2	27.2	28.5	19.6	18.0	14.5	13.9	21.3	314
1986	25.7	20.8	18.3	15.8	15.6	14.1	14.1	17.8	209
1987	25.9	27.2	22.2	14.7	15.4	14.1	15.9	17.9	270
1988	26.3	24.8	25.6	20.7	15.9	17.2	16.7	21.0	300
1989	26.6	22.6	22.0	15.8	16.4	16.3	16.5	19.5	334
1990	19.2	16.7	16.9	15.9	18.4	18.3	23.4	18.5	290
1991	30.4	29.9	24.0	18.3	19.1	13.4	18.7	22.0	322
1992	27.2	20.1	16.5	14.6	13.7	16.2	18.9	18.2	154
1993	21.9	18.0	14.9	14.9	15.5	20.3	24.3	18.4	233
1994	28.4	21.9	19.8	19.5	14.9	15.6	17.8	19.7	291
1995	24.6	26.6	18.4	13.6	18.3	21.9	12.8	19.5	329
Average	**26.2**	**24.8**	**21.6**	**18.2**	**17.1**	**17.2**	**18.1**	**20.5**	**285**
$P_{eff.\ aver.}$	39	55	52	36	43	30	30	–	285

Since the area of the Síkfőkút forest is free of the effects of subsoil water, the soil moisture originates directly from the atmospheric precipitation. It can be seen in Table 4, that average monthly moisture content in the upper 60 cm of the forest soil decreases from April to August by 9.1%. In dry growing seasons, like in 1992 and 1994, the decrease was more than 13%. The soil moisture content, therefore, reaches its minimum in August or September. Monthly average soil moisture contents begin to increase again in October, when weather is cool and the precipitation is more than the evaporation.

The soil moisture content decreases from April to August, because the temperature increases and relative humidity decreases. The growth of the saturation deficit in the air increases the transpiration force on the plants' leafs, while soil moisture content decreases. The difference between the potential and actual evapotranspiration increases in such situation. The energetic background of the process is that higher ratio of the radiation balance is devoted to the warming of the air of the forest with the decrease of soil moisture content, since there is a significant decrease in the amount of energy devoted to evapotranspiration in line with the decrease of soil moisture content. The effective precipitation, which reaches the soil through the foliage in the forest compared to the soil moisture content, does not reflect the increase of the amount of monthly mean precipitation from April to May–July, which is characteristic for Hungary. In other words, the precipitation maximum at the beginning of summer does not appear in the course of soil moisture content, since the forest uses it for increasing biomass production and evapotranspiration.

Moisture content values of the different soil layers are shown in *Table 5*. It can be seen in the table, that the maximum values of soil moisture can be found in the upper soil layer, which contains the most humus. The highest moisture content is found in the topsoil (0–5 cm), since the soil surface is covered with a thick layer of forest litter. For this reason, the soil surface is not dried by the solar radiation and wind. On the other hand, the thick cover of fallen leafs decreases the evaporation from the soil surface. There is less moisture content in the 30–50 cm layer, since it is densely interlaced by the roots of the trees, and the roots take up the most water in this layer. The soil moisture content, its oscillation and standard deviation decreases with the increasing depth. For more details on the soil moisture content in the Síkfőkút forest, see *Justyák* and *Nagy*, 1988, 1990; *Antal et al.,* 1995; *Justyák* and *Vig*, 1997; *Kiss*, 1995.

Trends of the changes of soil moisture content in the root zone of the forest between 1978 and 1993 were analyzed by the following method. Measured soil moisture data were transformed in order to screen seasonal anomalies in a way, that each value was divided by the 15 years average of its own decade. The set of the transformed data was divided into two heaps for each year. The first contained the data below 1.00 (the decade average), the other contained the values above 1.00. On the basis of the ratio of these heaps it can be decided for each year, whether the soil of the Síkfőkút forest were dry, wet, or it had average water supply compared to the average of the given period. For annual trend analyses, ratios of the soil moisture content values higher than the decade averages within the whole dataset were taken into consideration *(Fig. 1)*. The decreasing tendency of soil moisture content is clearly visible in the figure.

Table 5. Mean soil moisture content (%) in the growing seasons in the Síkfőkút forest in different depths

Years	0–5 cm	5–10 cm	10–20 cm	20–30 cm	30–40 cm	40–50 cm	50–60 cm	0–60 cm
1978	32.3	26.3	23.1	20.2	19.3	19.9	20.5	23.1
1979	30.0	23.7	21.8	21.9	20.3	18.3	22.2	22.6
1980	38.0.	29.8	23.8	21.9	21.3	21.6	21.8	25.5
1981	32.1	24.3	21.5	22.2	18.7	18.9	20.2	22.0
1982	29.1	21.9	20.4	18.6	18.2	18.4	18.8	20.8
1983	28.5	21.5	19.2	17.8	17.8	18.2	18.1	20.1
1984	27.0	23.0	20.7	18.5	17.8	17.8	17.8	20.4
1985	29.5	23.6	20.7	19.7	18.3	18.7	18.6	21.3
1986	20.6	18.9	17.4	16.5	17.2	17.3	17.3	17.8
1987	26.8	20.1	18.2	17.0	17.3	18.0	18.0	17.9
1988	28.2	23.4	21.3	19.8	17.8	18.4	18.3	21.0
1989	25.3	21.2	19.2	17.7	18.0	17.7	16.9	19.5
1990	23.4	20.1	18.6	17.5	16.9	16.2	16.1	18.5
1991	30.2	24.2	21.1	20.4	19.6	19.5	18.8	22.0
1992	22.2	19.9	17.7	16.7	17.5	16.8	16.4	18.2
1993	25.1	22.5	19.0	16.8	15.9	15.3	15.2	18.4
1994	31.1	23.4	17.5	15.9	15.2	15.6	15.5	19.7
1995	27.5	22.2	18.7	18.1	17.1	16.5	15.9	19.5
Average	**28.2**	**22.8**	**20.0**	**18.5**	**18.0**	**18.0**	**18.1**	**20.5**
Maximum	**38.0**	**29.8**	**23.8**	**22.2**	**21.3**	**21.6**	**22.2**	**25.5**
Minimum	**20.6**	**18.9**	**17.4**	**15.9**	**15.2**	**15.3**	**15.2**	**18.2**
Oscilation	**17.4**	**10.9**	**6.4**	**6.3**	**6.1**	**6.3**	**7.0**	**7.3**
Standard deviation	**4.0**	**2.5**	**1.8**	**1.8**	**1.4**	**1.5**	**2.0**	**2.1**

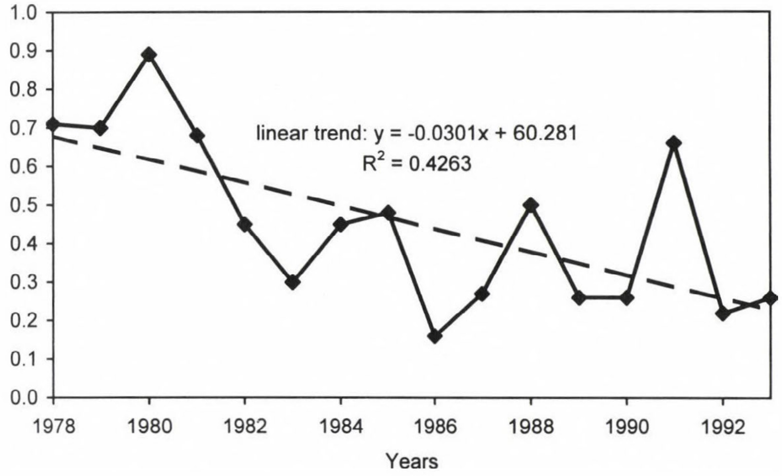

Fig: 1. Ratio of soil moisture content values higher than the average of the decade and total soil moisture content values in the 0–60 cm layer.

3. Water supply

The amount of water could be stored in the soil is determined by the above mentioned physical characteristics of the soil. The supply of water in the forest soil was determined on the basis of soil moisture (%), bulk density (g/cm^3), and thickness of layer (cm), the following way (*Szász,* 1988):

$$W\,(\mathrm{mm}) = Sm\,(\%)\cdot BD\cdot Th/10, \tag{2}$$

where *Sm* is the soil moisture content (in percentage of dry soil weight), *BD* is the bulk density of the soil (g/cm^3), and *Th* is the thickness of the soil (cm).

Additional information can be obtained by analyzing the monthly water supply (mm) of the soil layers of the forest (*Table 6*).

Table 6. Annual mean water supply (mm) of the soil layers in the Síkfőkút forest (1978–1995)

Depth (cm)	Apr	May	Jun	Jul	Aug	Sep	Oct	Average
0–10	40	39	34	28	25	25	26	31
10–20	32	30	29	21	20	21	22	26
20–30	29	27	23	20	18	18	19	22
30–40	31	32	28	20	20	21	23	25
40–50	38	30	27	23	24	23	24	27
50–60	23	35	33	30	27	27	27	29
0–60	203	193	174	142	134	135	141	160

It can be seen in the table, that the upper rich humus layer (0–10 cm) and the 50–60 cm deep one, which is not plentiful of living roots, contain the same amount of water. In the other layer, where the root mass of the trees and shrubs can be found, there are only 22–27 mm of water due to the strong water uptake. In line with the soil moisture content, the soil water supply decreases from April to August, which is a consequence of energy transport processes mentioned before, in connection with the soil moisture content.

4. Interceptional water

The difference between the amount of atmospheric precipitation (P) falling onto the foliage of the forest and the amount which reaches the surface falling through the foliage ($P_{eff.}$) is called *retained water* (Pb). The process of the retention of the precipitation is *interception* (I). Monthly averages of these characteristics are presented in the *Table 7.*

Table 7. Monthly averages of the precipitation characteristics (1978–1995)

Charac-teristics	Apr	May	Jun	Jul	Aug	Sep	Oct	Apr–Oct	Dim.
P	43	68	63	44	54	36	37	345	mm
$P_{eff.}$	39	55	52	36	43	30	30	285	mm
Pb	4	13	11	8	11	6	7	60	mm
I	9	19	17	18	20	17	19	17	%

(P – atmospheric precipitation, $P_{eff.}$ – precipitation reaching the surface, P_b – retained water, I – Interception)

It can be seen from the characteristics presented in Table 7, that the forest retained 60 mm of atmospheric precipitation on the average. This interceptional water is not useful for the forest, since it is evaporated into the atmosphere from the trunks, branches, and leafs of the trees. The indirect use of the interceptional water is that it decreases the intensity of transpiration till it evaporates. The annual average *interception* (1978–1995) of the forest during the growing season is 17%. In leafless state it is 9 % (April), while in leafy state (May–Oct) it is meanly 18%. These values are characteristic for the forest canopy decayed by 50%. Before forest decay, the average interception of the healthy, closed foliage forest was 21%. In leafless state (April) it was 13% (*Szabó,* 1985).

The water of abundant rains is conducted to the trunks by the branches, then it flows down on the trunks onto the soil surface. The amount of water flowing down on the trunks in leafy state was 3.4% of the annual average amount of precipitation. In leafless state it was 3.9%, which is not ignorable from the aspect of water balance of the root zone.

Table 8 shows the average interception retained by the Síkfőkút forest in leafy state, in case of different rains, over and inside the forest. It is visible, that the forest can retain higher ratio of small rains and smaller ratio of large rains.

Table 8. Average interception in case of different rains over and inside the forest

Precipitation over the forest (mm)	Precipitation inside the forest (mm)	Water retained by the forest (mm)	Interception (%)
50.0	41.8	8.2	16.4
45.0	35.8	9.2	20.4
38.8	29.3	9.5	24.5
17.5	18.8	4.7	26.9
14.8	10.0	4.8	32.4
7.5	4.7	2.9	37.3
2.5	1.2	1.3	52.0

One hectare of oak forest at Síkfőkút retained 700–2400 m^3 water annually depending upon the amount of precipitation over the canopy. Most of it evaporates into the atmosphere. Due to the many influencing factors, this value oscillates within a broad interval. Interception capacity of the Síkfőkút forest was 2300 m^3/ha in 1978, and 720 m^3/ha in 1982.

In the forest, a part of the precipitation is retained by the litter, while the larger part of it gets into the soil through the litter. When the litter is saturated, the water flows away on the surface, this is the surface runoff. The ratio of the water amounts flowing away and got through the forest is called *surface runoff factor*, which is characteristic for the use of precipitation water. In the Síkfőkút oak forest (on a slope of 3°, intermediately permeable soil), it varied between 0.01 and 0.04 depending on the amount of precipitation. The amount of surface runoff from 1 hectare, from May to November, varied from 40 to 80 m^3 (*Justyák* and *Bihary,* 1993; *Justyák* and *Vig*, 1997).

5. Actual evapotranspiration

Another important component of the water balance in the forest is the *actual evapotranspiration* (*AET*). The actual evapotranspiration of the forest is determined by the sum of the *transpiration* from leaves and *evaporation* from the soil surface. The actual evapotranspiration of the forest can be determined using the following simple relationship:

$$AET = P_{eff.} + (W_s - W_f) + I - R, \tag{3}$$

where W_s is for the starting and W_f is for the finishing water supply of the soil (mm), $P_{eff.}$ means the effective precipitation, which reaches the soil and litter falling through the foliage of the forest (mm), I is the interceptional water (mm), and R means the surface runoff (mm).

Mean actual evapotranspiration values from the forest canopy are presented in *Table 9*, where the amount of evaporated interceptional water is taken into consideration as well, while the amount of runoff is ignored.

Table 9. Monthly mean actual evapotranspiration (mm) from the Síkfőkút forest canopy during the growing seasons of the 1978–1995 period

Years	Apr	May	Jun	Jul	Aug	Sep	Apr–Sep	$P_{eff.}$ Apr–Sep
1978	60	77	83	88	74	40	422	367
1979	60	59	74	68	80	35	376	239
1980	54	70	97	77	54	42	376	294
1981	44	62	66	82	77	50	381	256
1982	46	63	92	79	58	37	375	236
1983	47	81	69	69	65	38	369	205
1984	47	85	67	62	55	52	368	281
1985	48	87	94	67	70	40	406	309
1986	43	69	70	79	62	39	362	201
1987	57	73	66	63	58	43	360	246
1988	45	80	83	67	79	41	395	291
1989	46	84	97	79	65	36	407	326
1990	55	65	63	69	62	41	355	213
1991	46	78	79	81	66	33	383	267
1992	40	57	50	62	53	35	297	101
1993	50	59	52	60	64	40	325	185
1994	59	71	63	59	75	43	370	250
1995	68	72	85	72	78	46	421	329
Average	**51**	**72**	**75**	**70**	**66**	**47**	**374**	**255**
Maximum	**68**	**87**	**97**	**88**	**78**	**52**	**421**	**367**
Minimum	**43**	**57**	**50**	**59**	**53**	**33**	**297**	**101**
Oscillation	**25**	**30**	**47**	**29**	**25**	**19**	**124**	**166**
$P_{eff.\ aver.}$	**39**	**55**	**52**	**36**	**43**	**30**	**255**	

During the growing season, the monthly mean actual evapotranspiration increases from April to June and then decreases from July to October in line with the amount of precipitation falling onto the soil surface of the forest. Monthly mean maxima are between 50 and 100 mm. Monthly mean minima are between 30–60 mm. Annual sums of monthly mean evapotranspiration during the growing season are between 297 and 422 mm. Actual evapotranspiration showed a decreasing trend in the examined 18 years period between 1978 and 1995. There were growing seasons (1978, 1995) when a considerable amount of water evaporated (422 and 421 mm). There were few cases as well, when only a small amount of water evaporated (297 mm in 1992).

It is obvious from the data, that the monthly mean actual evapotranspiration is always higher than the effective precipitation, which reaches the soil surface in the forest. The surplus derives from the winter precipitation amounts. The difference originates from the decrement of the soil's water supply and interceptional water. The monthly mean actual evapotranspiration reaches its maximum in May and June (72–75 mm), which is in line with the summer maximum of precipitation. In this period the water uptake of the forest from the root zone increases due to the intense assimilation processes caused by rapid growth of healthy trees and shrubs. The evapotranspiration of the dissoluted forest of Síkfőkút is affected by two factors. The *first one* is that the faded trees do not transpirate, so the actual evapotranspiration of the forest is lower than that of the healthy forest. The *other factor* is that soil is covered with a 2–3 cm thick layer of litter in the forest, so meteorological elements can affect it only indirectly. The litter layer restricts many water molecules, which are able to get into the air of the forest from the soil. This way the evapotranspiration is decreased by the fading of the trees on one hand and process of evaporation, affected mainly by the characteristics and state of the litter, on the other hand (*Justyák*, 1987, 1995; *Antal et al.*, 1997).

6. The degree of tree dissolution

At the beginning of the complex environmental-biological studies (1972) in the Síkfőkút forest, there were 816 trees forming the foliage on one hectare area. 84.5% (689 trees) of them were sessile oak (Quercus petrea), while 15.5% of them (127 trees) were Austrian oak (Quercus cerris). The ratio of tree dissolution in the Síkfőkút forest is presented in *Fig. 2*. On the left vertical axis of the figure the number of living trees, while on the right vertical axis the number of dead trees are shown (tree fading affected mainly the sessile oak trees). According to Fig. 2, tree dissolution occurred in three main waves in the periods of 1979–80, 1983–86, and 1988–94 (*Kiss* and *O'Heix*, 1998).

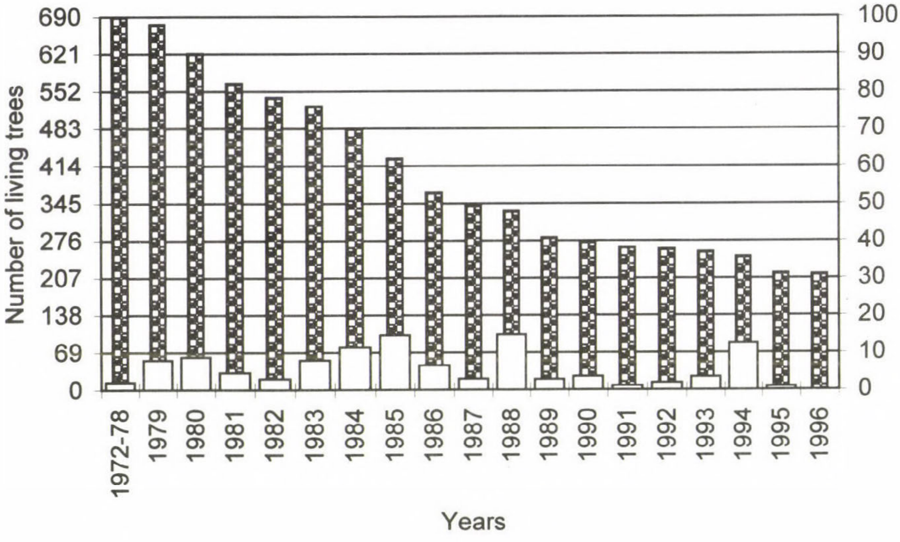

Fig. 2. The degree of fading of sessile oak trees on the 1 hectare sample area of Síkfőkút.

Causes of tree dissolution have been studied by many researchers. Some researchers think that the main reason is an epidemic caused by *viruses*. They specify *fungi*, which attack the root trunks of the trees as pathogenic organisms (*Redfern*, 1973; *Rütze* and *Liese,* 1980). Others attribute tree dissolution to soil acidification caused by acidic rain (*Ulrich et al.*, 1979; *Ulrich* and *Pankrath*, 1983; *Jakucs*, 1990). Others again explain the process by the *lack of precipitation* (for instance in the northern mountains in Hungary, where 50% of sessile oak trees have faded due to the lack of precipitation, while in Transdanubia, which gets more precipitation, only 10–20% of sessile oak trees have faded). There are some researchers who believe that the lack of precipitation causes harm to the *mikorrhiza-fungi*, which live in positive symbiosis with thin roots of sessile oak and cause problems in the nitrogen supply of the trees (*Berki*, 1995).

In connection with tree dissolution, the analysis of *Figs. 3* and *4* is interesting. Fig. 3 shows the annual variation and trend of temperature, while Fig. 4 presents the annual amounts and trend of precipitation, the trend equation and correlation coefficient, during the studied period (1978–1994), measured in a treeless area and inside the forest at a height of 2 meters. During the examined 17 years long period, mean temperatures in the forest show stronger increase than those in the treeless area, ratio of steepnesses of the two straights is 1:1.3. Increase of the mean temperature in the forest during a 10 years period is 1.0°C. It can be explained by several reasons. One reason is that due to the tree fading, density of the forest stand became lower. Other reason could be the consequence of warm weather series of the last 10–20 years. Fading of several trees decreased the shading effect of the foliage, and for this reason the difference between the forest and treeless area has decreased as well (*Antal et al.* 1997).

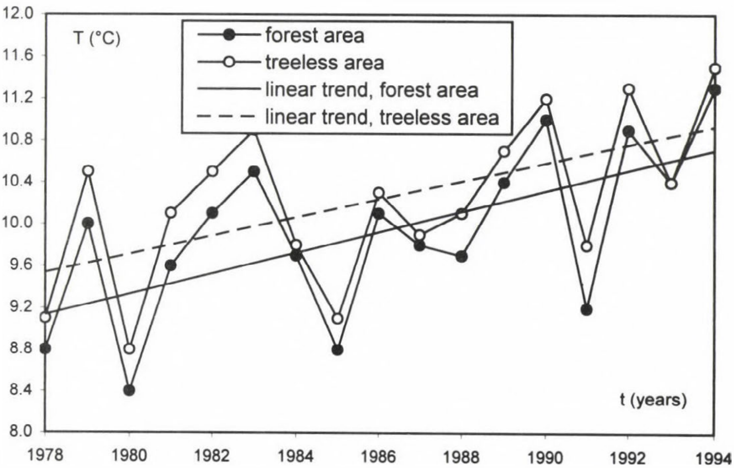

Fig. 3. Annual fluctuation and trend of temperature in the forest and treeless area.

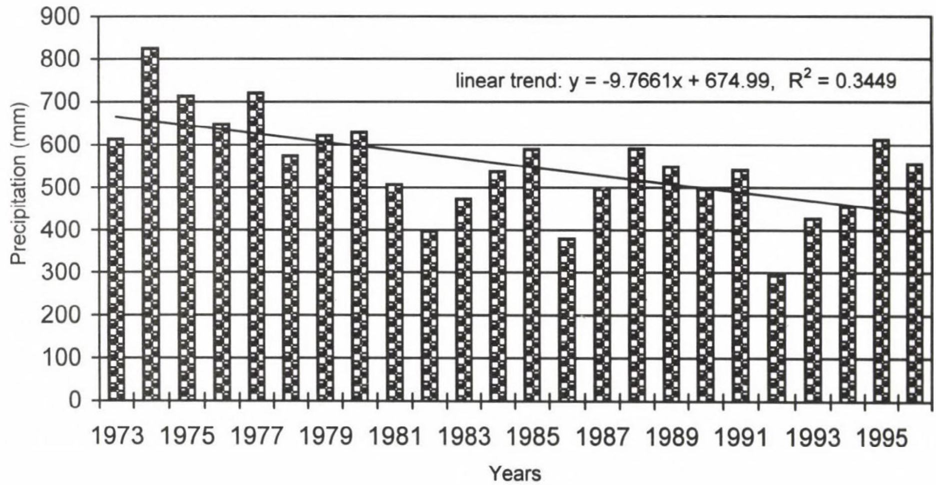

Fig. 4. Annual amounts of precipitation geting into the forest.

In Fig. 4 amounts of annual mean precipitation between 1973 and 1996 are presented. While annual mean temperature has increased, the amount of precipitation shows decreasing trend during the examined period. The annual amount of precipitation decreased yearly by 11.2 mm (with 1% error). Decrease of annual amount of precipitation is in line with decrease of the number of the trees (Fig. 2).

Warmer and drier climate presumably could launch processes, which caused the depreciation of the forest, and in combination with other factors they caused the death of the trees.

It seems however, that since 1995, due to the wetter climate and other factors, tree fading has stopped. These establishments are valid for the Síkfőkút forest only. The deterioration of the forest is in connection with environmental changes, acidification, warming up, decrease of precipitation, damage of root fungi. These changes and the complex effect of these damage-chains could lead to the fading of the trees.

Under natural conditions, the regeneration of the Turkey oak-oak joining requires few decades. Till then, researchers can study the interesting biological and ecological processes of the transformation.

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Probability of drought occurrence in Hungary

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Abstract—The paper gives a brief review on the drought parameters used in the Carpathians' Region, and on the previous investigations of their frequency. It deals with the calculation method of the aridity index (*PAI*) constructed by the author. It contains maps showing the results of a distribution-investigation (the average returning time period of the different severe droughts) of a 70 years long *PAI* data-base determined on 73 meteorological observation stations. Drought is a serious risk factor especially on the Great Hungarian Plain.

Key-words: drought, water scarcity, aridity index, returning time period

1. Introduction

Hungary, situated in Central Europe, belongs to the continental climatic zone, but its weather conditions are influenced sometimes by Atlantic and Mediterranean effects, too. The natural-geographical conditions are favorable for agricultural production from several points of view (about 70 per cent of the total area is cultivated), but drought is a serious risk factor. For making this risk countable, it is necessary to know — among other important factors — the probability of drought occurrence on the different regions of the country. Such an investigation is essential for the elaboration of the national drought strategy (*Harnos,* 1993, *Vermes,* 1998, *Vermes et al.,* 2000).

However this topic is of high interest, it was relatively neglected in the past home literature. The main reason of this is probably the fact that in Hungary till now there is no consensus in the best indices or parameters usable under home circumstances for the quantification of drought (*Faragó et al.,* 1988, 1990; *Bussay et al.*, 1999; *Domonkos et al.*, 2001). On the other hand, in most cases it is difficult to have long-term data-base for the distribution-investigations at great number of observation stations.

In the years of the 1960-s the probability of the occurrence drought or water scarce situations has been investigated for determining irrigation water demand.

Szesztay (1966) — based on the data observed between April 1 and October 31 of the years 1871–1960 in Debrecen — calculated the water deficit in the summer period as the difference of the potential evaporation and the precipitation, and constructed the probability curve of the 90 years data-base. From this curve the probability of a given water deficit can be read (in average after how many years this water deficit occurs). The value of the potential evaporation has been determined by the sum of number of days with temperature above 10°C (*Szesztay*, 1964).

Antal (1969, 1988) calculated the climatic water deficit for the whole year and also for the cultivation period (April 1–September 30) of the years 1901–1965 at three observation stations (Debrecen, Mosonmagyaróvár, Szombathely), and constructed the probability curves. The potential evaporation was calculated by his own formula, in which the main factors are as follows: the mean daily air temperature, the saturation vapor pressure, and the actual vapor pressure.

The National Master Plan of Water Management (*Varga*, 1984) shows the 80 per cent probability values of the water deficit in the cultivation season on a map-series of different crops. The simulation model-investigation as a basis of the mapping was carried out on 13 groups of plants, for two or three different yield-level of each. The soils have been ranged into four classes according to the water storage capacity of them, and groundwater level has been taken also into consideration. Meteorological effects have been characterized by the 50 years data of the decade heat-sum and precipitation observed in 23 stations. In case of two observation stations (Zalaegerszeg and Kecskemét) the probability distribution-functions of the water deficit in the cultivation season of alfalfa was given as well.

The probability of drought occurrence in the years of the 1980-s has been studied using our own aridity index (*PAI*, see 2.1) which was not quite the same as the present form of the index (*Pálfai*, 1987). Using the data calculated from the values of 1901–1986 years at 67 meteorological stations, first we determined the spatial distribution of the 10 per cent occurrence probability values of the index (values occurring in average every ten years), second the distribution-curves of the national spatial averages were determined according to the Gauss-function. As a result, country-wide drought occurs in every 3.3 years.

Varga-Haszonits (1989) calculated an aridity index as a quotient of the potential evaporation and precipitation data of 1881–1980 years from 25 observation stations, and he determined the frequency values for the vegetation period at each stations. The potential evaporation has been determined by the formulas containing the empirical constants based on the monthly mean air temperature values of the years 1951–1980, differentiated by the counties.

The study of *Dunay* and *Tölgyesi* (1993) gives the drought frequency data in three classes to nine regions (nine typical observation stations) of Hungary, based on the soil moisture data of the years 1951–1992. As an example, in the region between the Danube and Tisza rivers from the 42 years investigated, 10 years were favorable from the view-point of water supply (it means the soil moisture content did not decreased below 40 per cent of the utilizable water capacity of the soil in the upper 1 m layer), 15 years were classified as droughty (soil moisture content decreased below 20 per cent), and 4 years were severely droughty (soil moisture content decreased below 10 per cent). The frequency of the same categories was found in the western part of Hungary 37, 2, and 0 years, respectively.

Szász (1994) elaborated different formulas for expressing dryness, among which his aridity index is the quotient of the precipitation and potential evaporation, but for certain extent it depends on the value of the winter-spring precipitation and the soil type (water retention capacity of the soil), too. From the average values of the index the Hungarian distribution map has been constructed. Based on the long-term investigations made in Debrecen, it was found that occurrence of dry periods was considerably more frequent in the last decades of the 20th century than at the end of the 19th and beginning of the 20th century. The author intensively evaluated the temperature and precipitation data of the years between 1901–1960 at 15 observation stations, and the connections among them. He determined—among others—the frequency per cents of the occurrence of extremely hot and dry months in the summer period as an indicator of the droughty character of the weather.

An example for the statistical evaluation of the extremely dry periods in the neighboring Vojvodina Province in Yugoslavia can be seen in the paper of *Beric* and *Neskovic-Zdravic* (1995). A series of maps are showing the length of the water-scarce periods returning in every 2, 5, 10, 20, 50, and 100 years. As a result, the length of these periods is changing between 21 and 72 days.

Another study (*Bussay et al.,* 1999) determined the drought frequency from the relative sum of precipitation, based on the data of the 110 years between 1881 and 1990 of 25 observation stations. According to the recommendations of the WMO, drought has been determined if the monthly relative precipitation values were found below 60 per cent, and the yearly values below 75 per cent. In this case the spatial distribution of the drought frequency showed surprisingly small variability. The different months were droughty in 30 per cent of the total, while in the whole year investigations the drought frequency remained below 10 per cent.

In the past few years — apart from the Palmer Drought Severity Index (PDSI) —, another index: the Standardized Precipitation Index (SPI) has been used more and more worldwide (*Bussay et al.*, 1999). Both of these indices have been calculated to some Hungarian stations as well (*Domonkos et al.*, 2001).

Vrânceanu et al. (2000) published a map of areas affected by drought, which was constructed by Stanescu and coworkers for the territory of Rumania using our aridity index (*PAI*). The authors gave the average returning time of the drought according to three value-categories of the index, separated on the map.

In the next chapter the calculation method of the *PAI* is given, and the results of the investigations carried out with the index are presented.

2. Method

2.1 Calculation of the Pálfai Aridity Index (PAI)

There are indices for the characterization of the severity of an arid situation (dryness) by single digit derived from only few meteorological and/or hydrological parameters. The great advantage of such indices is that long-term data series could be produced by them.

The formula to calculate the base-values of the aridity index has been introduced by *Pálfai* (1984) as follows:

$$PAI_0 = \frac{t_{IV-VIII}}{P_{X-VIII}} \cdot 100, \tag{1}$$

where PAI_0 - base-value of the aridity index (°C/100 mm),

$t_{IV-VIII}$ - mean value of air temperature of the period of April–August (°C),

P_{X-VIII} - precipitation depth summed up by the weighed monthly values of precipitation of the period of October–August (mm).

Monthly weights for the precipitation values were based on the conditions of moisture-storage and on the changing general water demand of the crops. Estimates of the weighing factors are the following (with due regard on the overall natural conditions of the Carpathian Basin): 0.1 in October, 0.4 in November, 0.5 from December to April, 0.8 in May, 1.2 in June, 1.6 in July, 0.9 in August. It is evident that month July is the most critical period from the point of view of water supply.

When the values of PAI_0 were compared to the well-known Palmers „drought-index", strict correlation was founded between them *(Pálfai,* 1990).

For the more accurate expression of aridity the base-value of PAI_0 should be corrected by the following factors (*Pálfai et al.*, 1995).

Temperature (hot days) correction factor:

$$k_t = \sqrt[6]{\frac{n+1}{\bar{n}+1}}, \tag{2}$$

where k_t – temperature correction factor, n – number of the hot days ($t_{max} \geq 30°C$) in period of June–August (d), $\bar{n}$ – long-term country-wide average of the n value (d); in Hungary this value is 16 days.

Precipitation correction factor:

$$k_p = \sqrt[4]{\frac{\tau_{max}}{\bar{\tau}_{max}}}, \tag{3}$$

where k_p – precipitation correction factor, τ_{max} – the longest precipitation poor period (if the sum of precipitation in the successive days does not exceed max. 5–6 mm) between the middle of June and middle of August (d), $\overline{\tau_{max}}$ – long-temp country-wide average of τ_{max} (d); in Hungary this value is 20 days.

Groundwater correction factor:

$$k_{gw} = \sqrt{\frac{H}{\bar{H}}}, \tag{4}$$

where k_{gw} – groundwater correction factor, H – mean depth of the groundwater table below ground level in the period of November–August (m), $\bar{H}$ = long-term value of H on the given area (m).

The use of this correction factor is important on plain area. Practically it is the best to use the data of the nearest 2 or 3 groundwater wells in the surrounding of the meteorological station or observation point.

The final value of the aridity index — defined as *PAI* — is obtained from the base-value (PAI_0) by corrections:

$$PAI = k_t \, k_p \, k_{gw} \, PAI_0 \,, \tag{5}$$

where the correction factors are those described above.

According to the Hungarian experiences, the threshold value of the Pálfai Aridity Index shall be at $PAI=6.0$. Smaller values are for wet years at a particular site, larger values would indicate the different severities of dryness. These may be categorized as follows: $PAI=6$–8: moderate dryness, 8–10: medium dryness, 10–12: heavy dryness, >12: extremely heavy dryness.

2.2 Calculation and statistical evaluation of PAI data-series

The values of the aridity index explained above of the years between 1931–1998 for 68 meteorological stations were published in one of the periodicals of the Hungarian Meteorological Service *(Pálfai et al.,* 1999*)*. Some typing mistakes found in the publications were corrected, plus 5 mountain-stations were included, and the data-base was completed up to 2000. As a typical example, a diagram is given here with the data of Szentes (*Fig. 1*), where — during the 70 years examined — the most severe drought happened in 1952 ($PAI=15.9$°C/100 mm).

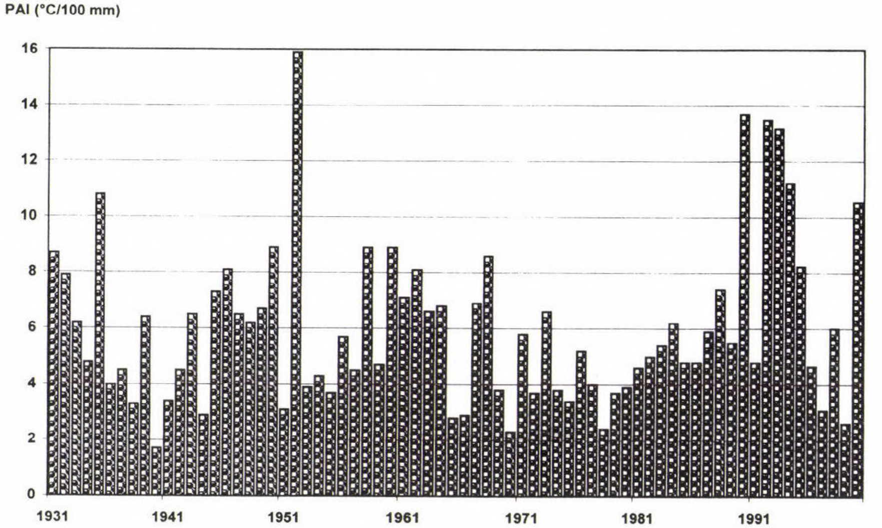

Fig. 1. Values of the Pálfai Aridity Index (*PAI*) between 1931 and 2000 at Szentes station.

Homogeneity of the 70 years data has been investigated by the *Smirnov-Kolmogorov* test. Taking into account the 30 and 70 per cent significancy boundaries, the data-base was taken homogeneous in the majority of the stations. Inhomogeneity (slightly increasing trend) was observed only in the case of 8 stations, mainly because of the differences in the lower value-categories of the index. Therefore, the data-base was accepted in all stations from the view-point of the determination of drought occurrence probability. Other investigations (*Szalai* and *Szentimrey,* 2001) showed much more inhomogeneity.

For the determination of the latest (as the average returning time) the so-called logarithmic Pearson-distribution function has been chosen, because — according to our examinations — this fitts best the empirical distribution.

3. Results

The results of the distribution-investigations carried out with the data of the 73 observation stations are given in three figures (Figs. 2, 3 and 4). On these the lines of the returning times of 3, 5, 10, 20, 50, and 100 years are shown for the cases of different severe droughts.

On the spatial distribution of the average returning time of the moderate drought (*PAI* $\geq$ *6*) can be stated (*Fig. 2*), that this type of droughts occurs in every 3 years on the Great Hungarian Plain, which covers the half of the country, while only in 5–20 years on the western and northern part of Hungary (near to the mountain area of the Alps).

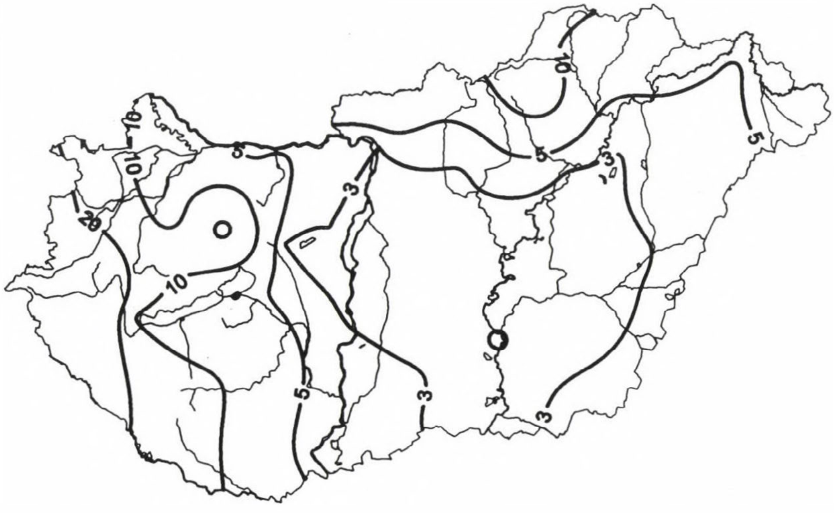

Fig. 2. Spatial distribution of the average returning time of moderate drought (*PAI* $\geq$ 6).

According to *Fig. 3* showing the average returning time of medium drought ($PAI \geq 8$), one can count on this type of droughts in less than 10 years on the Great Hungarian Plain, and about 20–100 years on the western part of Hungary.

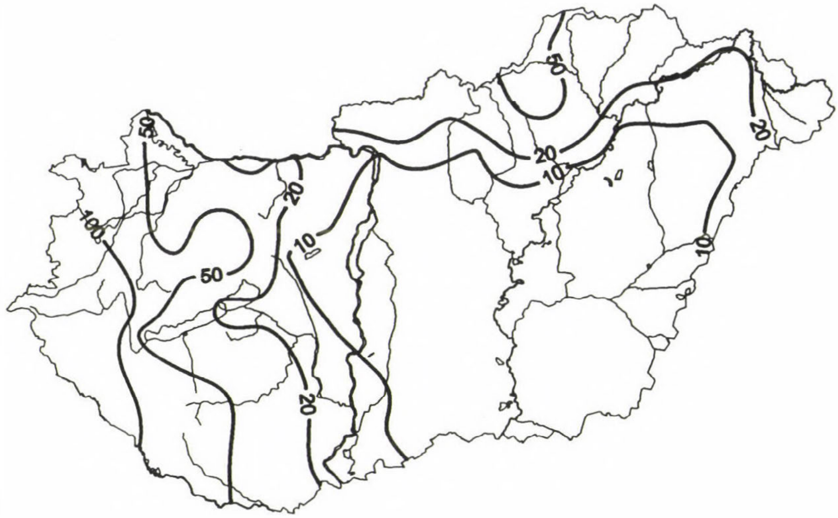

Fig. 3. Spatial distribution of the average returning time of medium drought ($PAI \geq 8$).

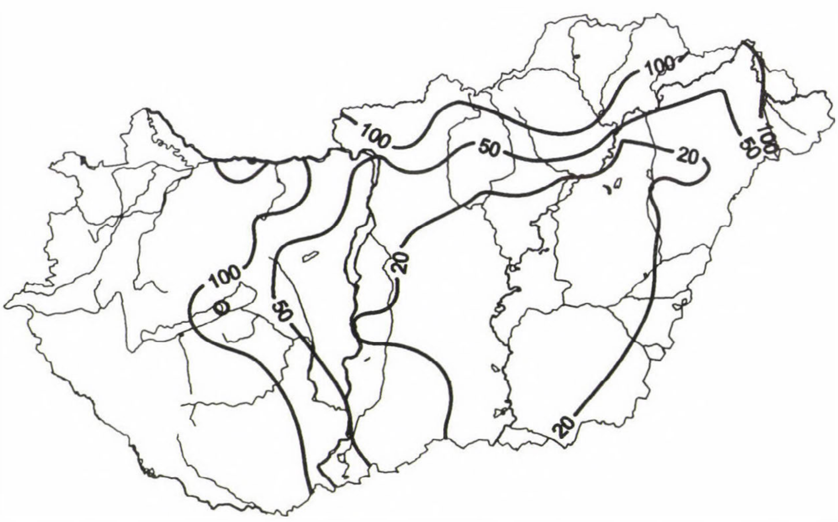

Fig. 4. Spatial distribution of the average returning time of severe drought ($PAI \geq 10$).

As it is seen in *Fig. 4*, severe drought (*PAI* $\geq$ *10*) occurs in 10–50 years on the Great Hungarian Plain (except its far eastern part), and 50–100 years on the rest areas of the country.

Extremely severe drought (*PAI* $\geq$ *12*) occurred during the 70 investigated years only on the Great Hungarian Plain, therefore the construction of a country-wide map would be illusory. The average returning time of such a drought is about 20–50 years on the middle area, and 50–100 years on the edge of the Plain. In *Table 1* the places and years are listed where and when the *PAI*-values were higher than 12. In the majority of the stations extremely years 1952 and 1992 occur most frequently. At the stations examined, but not included into the table, the *PAI*-values did not reached the number 12.

Table 1. Extremely severe droughts in Hungary (*PAI* $\geq$ *12*) between 1931–2000

Meteorological stations	Years
Ásotthalom	1952, 1992, 1993, 2000
Békéscsaba	1935, 1950
Budapest	1952, 1992
Cegléd	1952, 1983, 1990, 1992, 1993
Debrecen	1935, 1992
Fegyvernek	1935
Gyöngyös	1992
Hajdúdorog	1952, 1992
Hortobágy	1952, 1990, 1992
Izsák	1952
Jászberény	1992
Kalocsa	1993
Karcag	1952, 1962, 1992
Kecskemét	1952, 1983, 1990, 1992, 1993, 1994
Kiskunfélegyháza	1935, 1952, 1990, 1992
Kistelek	1952, 1993, 2000
Kunszentmiklós	1952
Mezőhegyes	1952, 1992
Örkény	1952, 1983, 1992
Paks	1952
Polgár	1992
Poroszló	1990, 1992
Siófok	1992
Szarvas	1952, 1992, 1993
Szeged	1952, 2000
Szeghalom	1935, 1952, 1992, 1993
Szentes	1952, 1990, 1992, 1993
Szolnok	1992, 1993
Tiszafüred	1992, 1993
Tiszakécske	1952, 1992, 1993
Túrkeve	1935, 1950, 1983

The probability of drought occurrence will be presumably influenced by the future climatic changes. According to several investigations, slight warming up is expected in Hungarian territories, and the probability of precipitation will decrease (*Antal,* 1997; *Bartholy* and *Mattyasovszky,* 1998; *Mika,* 2002). If this will be the case, the frequency of drought will increase, while the components of the water balance will decrease (*Antal,* 1992; *Domonkos,* 1997; *Nováky,* 2000; *Mika,* 2002).

4. Conclusions

The results of our investigations interpreted — with the consideration of other factors — give more reliable basis for the analysis of the risks of agricultural production, more precisely of plant production, but even for the examination of some ecological situations and planning of irrigation. Drought is a serious risk factor especially on the Great Hungarian Plain.

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On the relationship between transpiration and soil texture

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Abstract—The relationship between microscale transpiration E characteristics and soil texture is analyzed. Microscale means point (few hundred meters × few hundred meters) and local (few kilometers × few kilometers) scales. In the analysis, sand and clay is used as soil texture. The point scale transpiration E_p is evaluated by a *deterministic* model, while the local scale transpiration E_l is estimated by a *statistical-deterministic* model. The core of the models is a diagnostic energy balance submodel based on the *Penman-Monteith* concept. The characteristics are considered in terms of analyzing the change of E versus soil moisture content θ, relative frequency distribution of $E(\theta)$, and the aggregation algorithms for estimation of (θ). The analyses are performed for different atmospheric forcing conditions.

According to the results, the E characteristics for sand and clay show great similarity. They differ from each other only quantitatively. These quantitative differencies are extensively discussed. The results obtained can be useful for estimating E for different soil textures.

Key-words: transpiration characteristics, point and local scale, upscaling, sand, clay.

1. Introduction

The soil texture determines indirectly the hydraulic and thermal properties of soil, and also the movement of soil water and the process of evapotranspiration (e.g., *van de Griend et al.*, 1985; *Kondo* and *Xu*, 1997; *Kim* and *Entekhabi*, 1998; *Braud et al.*, 1995; *O'Kane*, 1991). These interrelations are investigated experimentally and theoretically. The experimental studies are rare. One of the the most interesting experimental study was performed by *Idso et al.* (1974). The theoretical studies are more widespread. They investigate the chain of the interrelations between the soil hydraulic properties, water movement in the soil, and the evapotranspiration by numerical experiments. The experiments have been made either in prognostic or diagnostic mode. The former ones are widespread. In most cases they investigated the effect of soil texture on surface fluxes. In general, the experiments are performed in an off-line mode using fictitious data. The sensitivity proved to be strong, especially on the annual scale (*Irannejad* and *Shao*, 1998; *Wilson et al.*, 1987; *Pitman*, 1994). The sensitivity on a short (several days) time scale has been shown by *Mihailovic et al.* (1992), *Ács* and *Lőke* (2001a, b). The experiments performed in a coupled mode by a meteorological model are rare (e.g., *Kim* and *Entekhabi*, 1998). They commonly use fictitious soil physical data but there are also studies dealing with real databases (e.g., *Mika et al.*, 2002). In these studies, the sensitivity of meteorological variables is investigated.

In this study, we applied a diagnostic approach to study the relationship between transpiration and soil texture. The objective of this study is to analyse the dependence of microscale transpiration characteristics on soil texture. The characteristics are considered in terms of analyzing the change of E versus soil moisture content θ, relative frequency distribution of $E(\theta)$, and the aggregation algorithms for estimating transpiration on the local scale $E_l(\theta)$. The point (few hundred meters × few hundred meters) and the local (few kilometers × few kilometers) scale models are based on the *Penman-Monteith* concept (*Ács* and *Hantel*, 1999; *Ács et al.*, 2000). To our best knowledge, there is no study investigating the relationship between the soil texture and all the important aspects of microscale transpiration characteristics. In the study, we assumed that there are no advective effects accompanied by occasionally observed internal boundary layers (e.g., *Garratt*, 1992; *Hupfer* and *Raabe*, 1994), and there are also no mesoscale circulation patterns induced by surface discontinuities. Then the atmosphere can be assumed to be horizontally homogeneous with constant meteorological boundary conditions above a certain level (*Shuttleworth*, 1998).

2. Models

The two model types used are the *deterministic* model (DM) for diagnosing transpiration on the point scale and the *statistical-deterministic* model (SDM) for estimating transpiration on the local scale.

2.1 Deterministic model

The model does not treat interception. Further, we suppose that the vegetation canopy is completely closed. So the water vapor flux above vegetation is formed only by transpiration. The model is based on the equation system as follows:

$$H = A_e - \lambda E \tag{1}$$

$$\lambda E = \frac{\Delta \cdot Ae + \rho \cdot c_p \delta e / r_a}{\Delta + \gamma(1 + r_v / r_a)} \tag{2}$$

$$r_a = \mathrm{f}_1(u_*, L, \text{constants}) \tag{3}$$

$$u_* = \mathrm{f}_2(u_r, L, \text{constants}) \tag{4}$$

$$L = \mathrm{f}_3(u_*, H, E, \text{ constants}) = \frac{-\rho \cdot T_r \cdot u_*^3}{g \cdot k \cdot \left(\dfrac{H}{c_p} + 0.61 \cdot T_r \cdot E\right)}, \tag{5}$$

where H and λE are the sensible and latent heat flux, A_e is the available energy of vegetation surface, Δ is the slope of saturated vapor pressure curve, ρ is the air density, δe is the vapor pressure deficit, r_a is the aerodynamic resistance, γ is the psychrometric constant, r_v is the vegetation canopy resistance. r_a depends on the friction velocity u_* and the stratification expressed by the *Monin-Obukhov* length L. u_r and T_r are the wind speed and air temperature at reference height, c_p is the specific heat of air at constant pressure, g is the gravitational acceleration, and k is the von Kármán constant. The functions f_1 and f_2 depend on the stratification and choice of the universal functions. For *Businger et al.* (1971) functions, their exact form is given, e.g., in *Ács et al.* (2000) or *Ács* and *Kovács* (2001). But some other functions can also be applied (see e.g., *Antal,* 1962). The available energy flux of vegetation surface can be easily expressed using energy balance equation. r_v is parameterized by *Jarvis* (1976) formula. The moisture availability function F_{ma} is expressed as simply as possible via soil moisture content θ (e.g., see Eq. (35) in *Noilhan* and *Planton* (1989) or Eq. (35) in *Ács* and *Hantel* (1998)). Hereinafter this will be referred to as *Theta*-parameterization.

The equation system of five equations contains five unknowns: H, λE, r_a, u_*, and L. The equation system is solved applying an iterative procedure. In most cases, the solution can be easily obtained after 4–5 iterations, but in some physically unreal situations (for instance for strong radiation and weak wind) the procedure does not converge (*Czúcz* and *Ács*, 1999).

2.2 *Statistical-deterministic model*

The local scale transpiration is estimated by the *statistical-deterministic* model. It consists of a *deterministic* submodel for estimating transpiration (see section 2.1), a *statistical* submodel for generating θ as a random variable (*Wetzel* and *Chang*, 1988), and a submodel for calculating the area-averaged $E(\theta)$. In the following the two latter submodels will be briefly considered.

2.2.1 *Modeling soil moisture variability*

According to observations of *Bell et al.* (1980) and *Hawley et al.* (1983), area variations of θ on the local scale can be characterized by a normal distribution. According to *Wetzel* and *Chang* (1988), the corresponding standard deviation is

$$\sigma_\theta = \min\left(0.08,\ \theta_\mathrm{m}/2\right), \tag{6}$$

where θ_m is the area-averaged value of soil moisture content. The area variations of θ on the local scale are generated by Monte-Carlo runs applying a standard random number generation algorithm (see *Dévényi* and *Gulyás,* 1988) using θ_m and σ_θ as inputs.

2.2.2 *Calculation of area-averaged transpiration*

Since θ is a statistical variable, so are the turbulent heat fluxes H and λE. The statistical distribution of λE or E is analyzed by its relative frequency distribution. The area-averaged value of E is estimated by numerical integration of its relative frequency distribution function $RF(E_j)$ as follows:

$$E_l = \langle E \rangle = \sum_{j=1}^{n} RF(E_j)\, E_j\,, \tag{7}$$

where $\langle E \rangle$ is the area-averaged E, j is the interval number, and E_j is the corresponding E-value for the *jth* interval. The length of E-interval is chosen as 25 W m^{-2}. The submodel is applied in each step for $0.02 \leq \theta_m < \theta_S$ cycle to obtain the $\langle E(\theta_m, \sigma_\theta)\rangle$ curve. θ_S is the saturated soil moisture content.

3. Numerical experiments

The numerical experiments are performed by both the *deterministic* and *statistical-deterministic* models for grass covered surface using sand and clay as soil texture. The simulations are made for different atmospheric forcing conditions. In this study, we distinguished strong and weak atmospheric forcing conditions. They are presented in *Table 1*. During numerical experiments 2 × 2 runs are performed using both the *deterministic* and *statistical-deterministic* models. Of course, the computation time of the latter model is much longer with respect to the former one because of the generation of statistical variables.

Table 1. Atmospheric forcing conditions

Variables	Strong atmospheric forcing	Weak atmospheric forcing
Net radiation flux (W m^{-2})	700	300
Air temperature at reference level (°C)	25.8	25.8
Vapor pressure at reference level (hPa)	18.0	32.0
Vapor pressure at reference level (hPa)	6.0	2.0

4. *Simulation results*

Verification of $E(\theta)$ model based on *Theta* parameterization on the point scale is performed on the well known 1987 Cabauw data set (*Beljaars* and *Bosveld*, 1997). These results are presented in *Ács* and *Hantel* (1998) and *Ács et al.* (2000).

The microscale transpiration characteristics are analyzed in terms of the $E(\theta)$ curve (relationship between transpiration and soil moisture content), the relative frequency of $E(\theta)$, and the aggregation algorithm for $E(\theta)$. These features are analyzed for both sand and clay.

4.1 Analysis of transpiration curves

On the point scale, $E(\theta)$ does not show area variations, therefore, it is obtained by running the *deterministic* model. So $E(\theta)= E_p^s(\theta)$ for sand and $E(\theta)= E_p^c(\theta)$ for clay. On the local scale, the area variations of θ are represented by the normal distribution via θ_m and σ_θ. $E(\theta)$ is obtained by running the *statistical-deterministic* model. So, analogously to the point scale transpiration, $E(\theta) = <E^s(\theta_m, \sigma_\theta)> = E_l^s(\theta)$ for sand and $E(\theta) = <E^c(\theta_m, \sigma_\theta)> = E_l^c(\theta)$ for clay.

$E_p^s(\theta)$, $E_p^c(\theta)$, $E_l^s(\theta)$, and $E_l^c(\theta)$ (including the factor λ) for strong and weak atmospheric forcing are presented in *Figs. 1* and *2*. Inspecting the curves, it is obvious that the curves obtained for sand ($E_p^s(\theta)$, $E_l^s(\theta)$) and clay ($E_p^c(\theta)$, $E_l^c(\theta)$) are very similar. The similarity between $E_p^s(\theta)$, and $E_p^c(\theta)$ is more pronounced than between $E_l^s(\theta)$ and $E_l^c(\theta)$. The similarity can be analyzed introducing some characteristic soil moisture content values. They are presented in *Table 2* for sand, loam and clay. Note that the corresponding loam-referred curves are not presented here.

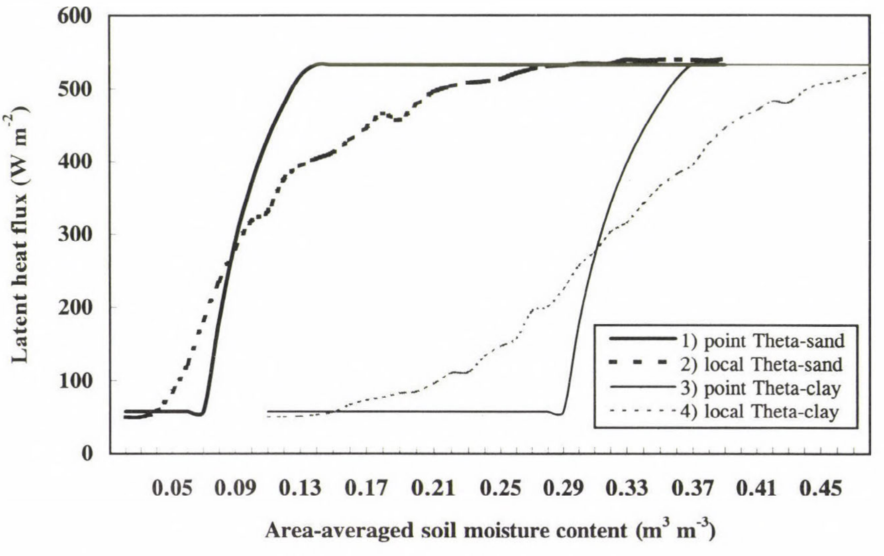

Fig. 1. Transpiration curve as obtained by Theta parameterization (1) on point scale using sand as soil texture (black continuous line), (2) on local scale using sand as soil texture (black dashed line), (3) on point scale using clay as soil texture (grey continuous line), and (4) on local scale using clay as soil texture (grey dashed line). The curves refer to strong atmospheric forcing conditions.

In total there are five boundary values of soil moisture content: the wilting point soil moisture content on the local scale $\theta_{l,w}$, the wilting point soil moisture content on the point scale $\theta_{p,w}$, the θ_c soil moisture content value in the transition region (region between the soil-controlled and the atmospheric-controlled transpiration), namely $E_p(\theta_c)=E_l(\theta_c)$, the field capacity soil moisture content on the point scale $\theta_{p,f}$ and the field capacity soil moisture content on the local scale $\theta_{l,f}$. The shape of $E_p(\theta)$ curves is strongly determined by $\theta_{p,w}$ and $\theta_{p,f}$. In the 0.02–$\theta_{p,w}$ region (the so-called soil-controlled region by low θ-

values on the point scale), $E_p^s(\theta)=E_p^l(\theta)=E_p^c(\theta)$=const. This constant does not depend on the atmospheric forcing conditions. Its value is determined by the parameterization of moisture availability function $F_{ma}(\theta)$. In the $\theta_{p,w}$-$\theta_{p,f}$ region (the transition region between the soil-controlled and the atmopheric-controlled regions on the point scale), the slope $\partial E_p^s(\theta)/\partial\theta = \partial E_p^l(\theta)/\partial\theta =$ $\partial E_p^c(\theta)/\partial\theta$. Note that the width of the region is about the same (in average about 0.08 m^3 m^{-3}) for all three soil textures. Therefore the slope of $E_p(\theta)$ curves is somewhat greater for stronger atmospheric forcing conditions. In the $\theta_{p,f}-\theta_S$ region (the so-called atmospheric-controlled region by high θ-values on the point scale), $E_p^s(\theta) = E_p^l(\theta) = E_p^c(\theta)$ =const. This constant depends only on the atmospheric forcing conditions. In this θ-region, the transpiration is called as potential transpiration. Note that the value of potential transpiration does not depend on the soil texture.

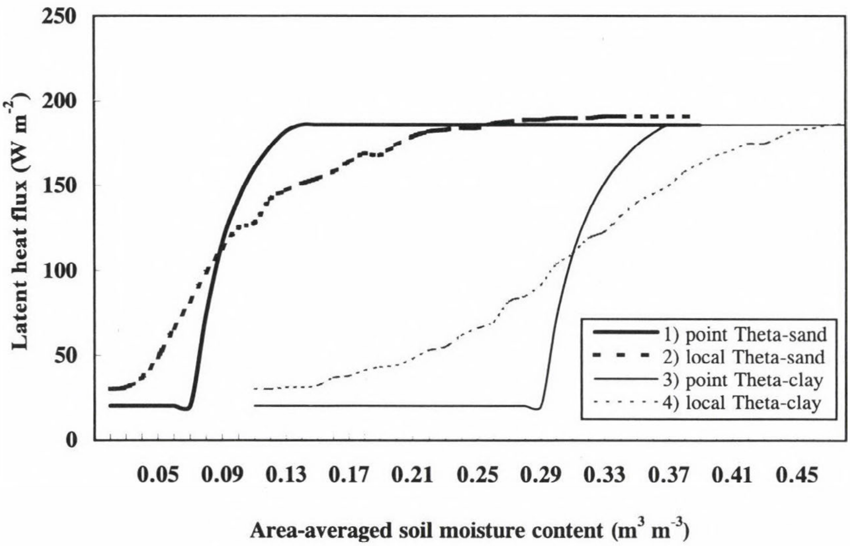

Fig. 2. As in Fig. 1 but for weak atmospheric forcing conditions.

Similar analysis can also be done for $E_l(\theta)$ curves. In the 0.02–$\theta_{l,w}$ region, (the so-called soil-controlled region by low θ-values on the local scale), $E_l^s(\theta)=E_l^l(\theta)=E_l^c(\theta)$=const. This constant — as in the former case — does not depend on the atmospheric forcing conditions. Since $\theta_{l,w} < \theta_{p,w}$, it obvious that the parameterization of $F_{ma}(\theta)$ is different on the point and the local scale.

The $\theta_{l,w}$–$\theta_{l,f}$ region (the transition region between the soil-controlled and the atmopheric-controlled regions on the local scale) has to be separated into two subregions. In the $\theta_{l,w}$–θ_c subregion, $\partial E_l^s(\theta)/\partial\theta > \partial E_l^c(\theta)/\partial\theta$ (the slopes $\partial E_l^c(\theta)/\partial\theta$ and $\partial E_l^l(\theta)/\partial\theta$ are about the same, this is not presented here) while in the θ_c–$\theta_{l,f}$ subregion, $\partial E_l^s(\theta)/\partial\theta \approx \partial E_l^c(\theta)/\partial\theta$. Since $\partial E_l^s(\theta)/\partial\theta >$ $\partial E_l^c(\theta)/\partial\theta$ the $E_l^s(\theta)$ and the $E_l^c(\theta)$ curves are not pronouncedly similar. In the $\theta_{l,f}$–θ_S region (the so-called atmospheric-controlled region by high θ-values on the local scale), $E_l^s(\theta) = E_l^l(\theta) = E_l^c(\theta)$=const. This constant — as in the former case — depends only on the atmospheric forcing conditions. Note that the potential transpiration on the point and the local scale is equal. As in the former case, the value of potential transpiration does not depend on the soil texture.

Table 2. Boundary values of soil moisture content obtained for sand, loam, and clay as soil textures. Symbols: $\theta_{l,w}$=wilting point soil moisture content on local scale, $\theta_{p,w}$=wilting point soil moisture content on point scale, θ_c=soil moisture content value in the transition region for which $E_p(\theta_c) = E_l(\theta_c)$, $\theta_{p,f}$=field capacity soil moisture content on point scale, and $\theta_{l,f}$=field capacity soil moisture content on local scale

Boundary values of soil moisture content ($m^3\ m^{-3}$)	**Sand**	**Loam**	**Clay**
$\theta_{l,w}$	0.03	0.08	0.15
$\theta_{p,w}$	0.07	0.15	0.29
θ_c	0.09	0.18	0.31
$\theta_{p,f}$	0.14	0.24	0.37
$\theta_{l,f}$	0.27	0.36	0.47

4.2 Area variations of transpiration

Area variation of $E(\theta)$ is examined analyzing its relative frequency distribution. *RF* obtained for sand and clay is RF^s and RF^c, respectively. The estimates are performed for strong atmospheric forcing conditions and three different θ_m-values. The characteristic θ_m-values are $\theta_{p,w}$ (0.07 $m^3\ m^{-3}$ for sand and 0.29 $m^3\ m^{-3}$ for clay), θ_c (0.09 $m^3\ m^{-3}$ for sand and 0.31 $m^3\ m^{-3}$ for clay) and $\theta_{p,f}$ (0.14 $m^3\ m^{-3}$ for sand and 0.37 $m^3\ m^{-3}$ for clay). The histograms of RF^s and RF^c for $\theta_{p,w}$-, θ_c- and $\theta_{p,f}$-values are presented on *Fig. 3a–c,* respectively. Obviously the RF^s and RF^c are very similar. They show bimodal distribution.

In dry regime ($\theta_m = \theta_{p,w}$), the *RF* maximum is on the left-hand side of the spectrum. In wet regime ($\theta_m = \theta_{p,f}$), the *RF* maximum is on the right-hand side of the spectrum. For $\theta_m = \theta_c$, the peaks on the left-hand side and on the right-hand side of the spectrum are quite large. Since the peaks on left-hand side of the spectrum are larger than the peaks on the right-hand side of the spectrum, the wetness regime represented by θ_c is more dry than wet.

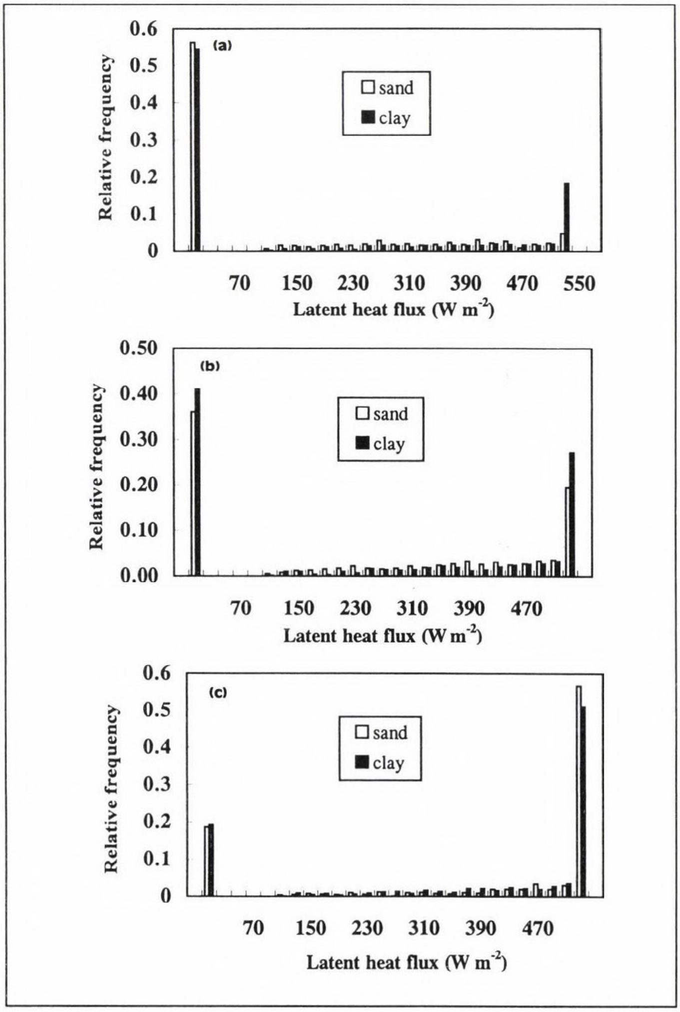

Fig. 3. Relative frequency distribution of latent heat flux from vegetation using sand (light columns) and clay (dark columns) as soil texture for (a) $\theta_{p,w}$, (b) θ_c, and (c) $\theta_{p,f}$. The results are obtained for strong atmospheric forcing conditions.

4.3 Aggregated soil moisture content

The aggregated soil moisture content θ_{ag} is defined by

$$E(\theta_{ag}) = \langle E(\theta_m, \sigma_\theta) \rangle , \tag{8}$$

where $E(\theta_{ag})$ is the area-averaged transpiration calculated by the *deterministic* model using θ_{ag} and $<E(\theta_m, \sigma_\theta)>$ is the area-averaged transpiration calculated by the statistical-deterministic model using θ_m and σ_θ. θ_{ag} obtained for sand and clay is $\theta_{ag}^{\ s}$ and $\theta_{ag}^{\ c}$, respectively. The relationship between θ_{ag} and θ_m can be obtained comparing $E_p(\theta)$ and $E_l(\theta)$ curves (see Figs. 1 and 2). The change of $\theta_{ag}^{\ s}$ and $\theta_{ag}^{\ c}$ versus θ_m for strong and weak atmospheric forcing conditions is presented in *Fig. 4.* The plots obtained for sand and clay are similar. The relationship between $\theta_{ag}^{\ s}$ and θ_m can be treated as linear, while the relationship between $\theta_{ag}^{\ c}$ and θ_m can be characterised as slightly non-linear. But it has to be noted that in the dry regime ($\theta_m < \theta_c$), the relationship can also be characterized as linear. Further, the relationship between $\theta_{ag}^{\ s}$ and θ_m does not depend on the atmospheric forcing conditions. In spite of this, the relationship between $\theta_{ag}^{\ c}$ and θ_m weakly depends on the atmospheric forcing conditions. In dry regime ($\theta_m < \theta_c$) this dependence is very small. In the wet regime ($\theta_m > \theta_c$), the more wet the surface, the more atmospheric-dependent is the $\theta_{ag}^{\ c}/\theta_m$ relationship. This is in accordance with simulation results of *Shao et al.* (2001) made by a mesoscale atmospheric model. Summarizing, the θ_{ag}/θ_m relationship for sand and clay can be characterized by regressions as follows:

$$\theta_{ag}^{\ s} = 0.30 \cdot \theta_m + 0.06, \ r^2 = 0.98 \tag{9}$$

and

$$\theta_{ag}^{\ c} = 0.93 \cdot \theta_m^{\ 2} - 0.32 \cdot \theta_m + 0.32, \ r^2 = 0.99, \tag{10}$$

where r^2 is the correlation coefficient. Eq. (10) refers to weak atmospheric forcing conditions.

4.4 Upscaling strategies

The results obtained for $E_p(\theta)$, $E_l(\theta)$ and θ_{ag} determine the possible upscaling procedures of transpiration from point to local scale. Since there is a pronounced similarity between $E_p^s(\theta)$ and $E_p^c(\theta)$, $E_l^s(\theta)$ and $E_l^c(\theta)$ and $\theta_{ag}^{\ s}$ and $\theta_{ag}^{\ c}$, the upscaling strategies for sand and clay will be also very similar. The upscaling procedures to be used are determined by the wetness regime.

In extremely dry (between 0.02 and $\theta_{l,w}$) and wet (between $\theta_{l,f}$ and θ_S) conditions (see Table 2), $E_p(\theta)$ is equal to $E_l(\theta)$, so there is no need for upscaling $E(\theta)$. In the transition region (between $\theta_{l,w}$ and $\theta_{l,f}$) between the extremely dry and wet conditions, the upscaling procedure depends on the characteristics of the θ_{ag}/θ_m relationship. The upscaling of $E^s(\theta)$ is to be performed by Eqs. (8) and (9). In this case, the procedure does not depend on the atmospheric forcing conditions. Analogously, the upscaling of $E^c(\theta)$ is to be performed by Eqs. (8) and (10). In contrast to the former case, the procedure slightly depends on the atmospheric forcing conditions.

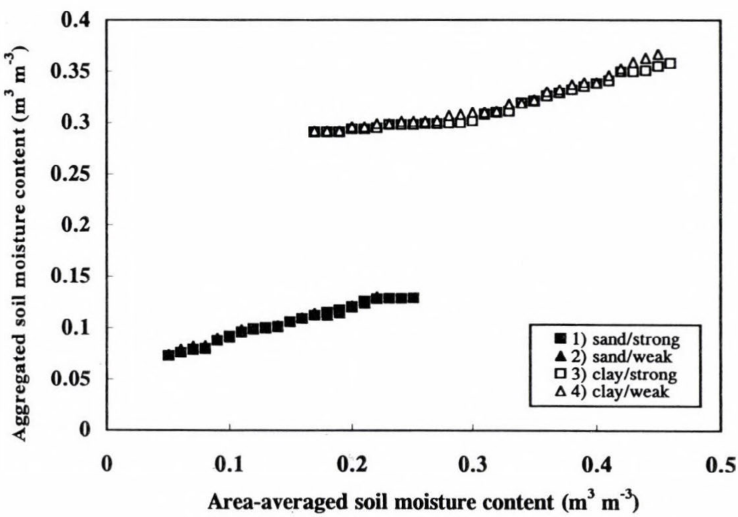

Fig. 4. Aggregated versus area-averaged soil moisture content for (1) sand using strong atmospheric forcing conditions (dark squares), (2) sand using weak atmospheric forcing conditions (dark triangles), (3) clay using strong atmospheric forcing conditions (light squares), and (4) clay using weak atmospheric forcing conditions (light triangles).

5. Conclusions

The relationship between microscale transpiration characteristics and soil texture is analyzed. Microscale means point and local scale transpiration. In the analysis sand and clay are used as soil textures. The point scale transpiration is estimated by a *deterministic* model, while the local scale transpiration is simulated by a *statistical-deterministic* model. The deterministic E_p-model is based on the *Penman-Monteith* concept (*Ács* and *Hantel*, 1999), that is, transpiration is determined by the *Penman-Monteith* equation, and the sensible heat flux is obtained as residual component from the energy balance equation. The *statistical-deterministic* E_l-model consists of an E_p-submodel, a statistical submodel for generating θ as random variable, and a submodel for calculating the area-averaged $E_l(\theta)$. The transpiration characteristics are considered in terms of analyzing transpiration E versus soil moisture content θ, relative frequency distribution characteristics of $E(\theta)$, and the aggregation algorithms for its estimation. The analyses are performed for different atmospheric forcing conditions. The results can be briefly summarized as follows:

- $E_p^s(\theta)$ and $E_p^c(\theta)$ curves are pronouncedly similar to each other. They can be treated as parallel curves translated to each other. The soil texture does not determine the value of E_p in the soil-controlled and atmospheric-controlled regimes, and the slope of E_p $(\partial E_p(\theta)/\partial\theta)$ in the transition region. $E_l^s(\theta)$ and $E_l^c(\theta)$ curves are also similar to each other. They are translated but not parallel to each other. The greatest deviation between them is in the $\theta_{l,w}$–θ_c region. Similarly to the former case, the soil texture does not determine the value of E_l in the soil-controlled and atmospheric-controlled regimes.

- Normally ditributed soil moisture variations produced bimodal RF^s and RF^c. Relative frequencies depend strongly on the wetness regime. According to the expectations, RF^s- histograms obtained for $\theta_{p,w}{}^s$, θ_c^s, and $\theta_{p,f}{}^s$ are very similar to RF^c-histograms obtained for $\theta_{p,w}{}^c$, θ_c^c, and $\theta_{p,f}{}^c$.

- The relationship between $\theta_{ag}{}^s$ and θ_m is linear. Inspite of this, the relationship between $\theta_{ag}{}^c$ and θ_m is non-linear. It has to be noted that in the dry regime ($\theta_m < \theta_c$), this latter relationship can also be characterized as linear. Furthermore, the relationship between $\theta_{ag}{}^s$ and θ_m does not depend on the atmospheric forcing conditions. However, the relationship between $\theta_{ag}{}^c$ and θ_m weakly depends on the atmospheric forcing conditions. In dry regime ($\theta_m < \theta_c$), this dependence is very small. In wet regime ($\theta_m > \theta_c$), the wetter the surface, the more atmospheric-dependent the $\theta_{ag}{}^c/\theta_m$ relationship is. This is in accordance with simulation results of *Shao et al.* (2001) using a mesoscale atmospheric model.

- Since there is a pronounced similarity between $E_p^s(\theta)$ and $E_p^c(\theta)$, $E_l^s(\theta)$ and $E_l^c(\theta)$, and $\theta_{ag}{}^s$ and $\theta_{ag}{}^c$, the upscaling strategies for sand and clay will be also very similar. The upscaling procedures to be used are determined by the wetness regime. In extremely dry (between 0.02 and $\theta_{l,w}$) and wet (between $\theta_{l,f}$ and θ_S) conditions (see Table 2), $E_p(\theta)$ is equal to $E_l(\theta)$, so there is no need to apply a procedure for upscaling $E(\theta)$. In the transition region (between $\theta_{l,w}$ and $\theta_{l,f}$) between the extremely dry and wet conditions, the upscaling procedure depends on the characteristics of θ_{ag} / θ_m relationship. A crucial aspect is wether the relationship between θ_{ag} and θ_m depends on the atmospheric forcing conditions. If this relationship is simple (there is no atmospheric dependence or this dependence is weak), the upscaling procedure to be applied is also simple (Eqs. (8) and (9) for sand and Eqs. (8) and (10) for clay). If the relationship is complex, for instance, it depends on the atmospheric forcing conditions, the construction of a simple $E_l(\theta)$ formula seems to be unlikely.

It has to be mentioned that soil moisture characteristics on the point scale ($\theta_{p,w}$, $\theta_{p,f}$) presented in Table 2 refer to soils of North-America (*Cosby et al.*, 1984). Obviously, the soil moisture characteristics on the local scale and the microscale transpiration characteristics refer also to the soils of North-America. The extension of the method to Hungarian soils is in process. Of course, these results are valid only — as mentioned in the introduction — when there are no advective effects and mesoscale circulation patterns. In the latter cases the transpiration characteristics and its upscaling strategy are more complex.

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