IDŐJÁRÁS

QUARTERLY JOURNAL
OF THE HUNGARIAN METEOROLOGICAL SERVICE

CONTENTS

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Prof. Dr. Gábor Szász is 75

Professor Dr. Gábor Szász has recently celebrated his 75th birthday. On this occasion we wish to publish a special number of the journal Időjárás devoted to him and his lifetime activity. The articles of this number are written by their colleagues who shared the work with him under his long and extremely successful career, and by students who were educated by him to be such devoted persons to meteorology and especially agrometeorology as he was and remains all the time of his life. We can say without any reservation that he is one of the leading personalities of Hungarian meteorology and one of the rare ones who could build and can maintain efficient contacts with other sciences, which gives us – the community of Hungarian meteorologists – a living joint to the engineering practice of agronomy and industrial use of microclimatology. He was extremely successful in building a school and creating a unit at the Agricultural University of Debrecen dealing with these branches of our science. He has undertaken important leading positions at the mentioned university including the rector and prorector position for a long time, and this way he was an important personality of the country's scientific policy, and also a leading personality of science in the region.

In the name of the Editorial Board, the authors of this issue and all representatives of other disciplines, whose activities were encouraged and supported by Professor Szász, I wish to send our greetings to him and wish many good and successful years in his professional and personal life.

Prof. Dr. Gábor Szász's scientific career

He was born in 1927 in Békés. After finishing his studies in the course of biology and geography at the University of Debrecen, he obtained his teacher's degree in 1950. Already in the years 1948–1950, he worked for the Meteorological Institute of the University where he started his professional career. He was an aspirant between 1952 and 1955, then he obtained his CSc degree in 1956. In this period he also finished the course of mathematics-physics (1958). He became the doctor of the Hungarian Academy of Sciences in 1999.

His long scientific career he devoted to research in the field of agrometeorology, and inside it he dealt with macro- and micrometeorological problems of water supply of plant. For the aim of this research, in 1962 he developed, with significant investment, the climatic station of the – now already 100 years old – Debrecen-Pallag Agricultural Academy into a high level and even now state-of-the-art observatory, for which he is the head. His scientific work was always the part of some large country-wide scientific programs, presently he works on two big projects in the frame of the Széchenyi plan. In the years 1970–1980 he worked out the system of agroecological regions of Hungary, that is in constant use from that time in decision-making practice of the Ministry of Agriculture and Regional Policy. He maintained and continues to maintain intensive scientific contacts and co-operation with research institutions and universities from a wide range of countries, like Germany, the Netherlands, USA, etc., mainly in micrometeorology. Some specific areas of his research include

- *the characteristics and representativity of plant canopy climate,*
- *the role of micro-advection and turbulent diffusion in formation of plant canopy climate,*
- *the transport of energy and substances within plant canopy,*
- *the role of macro- and micrometeorological conditions in plant development and crop yield,*
- *the use of remote sensing in determining the water supply of plant canopies.*

During his entire professional life, he had affiliation with universities and never left Debrecen. From 1951 to 1956 he worked as lecturer in the Department of Meteorology, then became scientist at the same place. In 1961 he changed to the College of Agrarian Science, where he continued to work in the same capacity. He became associated professor in 1963, then professor in 1971. At both of his workplaces he was the lecturer of agrometeorology, but also made lectures in climatology, micrometeorology, hydrometeorology, and agroecology for decades. In the frame of the reform of the University he organized and became the leader of a multidisciplinary PhD educational program and postgraduate school in agroecology. He always actively participated in the public life of the Debrecen universities, held a number of leading positions. For ten years he was the dean of a faculty, then for nine years the rector and the prorector of the Agricultural Univerisity of Debrecen. From the end of the 1970s he was an active initiator of the integration of the universities in Debrecen, and in consequence he led for two years the Union of Debrecen Universities, or – simply – the Universitas. His long participation in scientific policy has given us such results, like the computer network connecting the universities, or the harmonization of education and research through bringing them nearer to practice. In 1997 he formally retired from education, and presently he is professor emeritus of the University. He is member of numerous scientific organizations, e.g., for 10 years he was the president of the Hungarian Meteorological Society. He is the author of a very long list of scientific publications whose number exceeds 140, with 3 big monographs among them. He was honored for his work with numerous orders and prizes, like the Order of Labor (two times), the Officers' Cross of the Order of Republic, Eötvös Loránd, Pázmány Péter, and Schenzl Guidó Prizes, Steiner L., Hatvani I., Bocskai I., Pro Universitatis Memorial Medals, etc.

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Effects of the weather and climate to the evapotranspiration of crop canopies

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Abstract—Water demand, water consumption and water efficiency of crop canopies are affected by numerous factors. Plant species, variety, developing phase, plant canopy density, leaf area, fertilizer- and water supply play important roles. On the other hand, changes in these active factors are controlled by the weather and climatic conditions.

This paper presents the relationships between the features characterizing crop canopies and the optimal evapotranspiration affected by the weather (the water demand of crop canopies), on the base of few decades agrometeorological experiments.

Key-words: evapotranspiration, water demand, water supply of crop canopy, water balance, irrigation water demand, surface water surplus.

1. Introduction

At a given place, crop canopy water balance is regulated by the plant properties and, significantly, by the meteorological and climatic factors. Under Hungarian climatic conditions, precipitation plays significant role in the shaping of water demand and water supply of plants. Precipitation, capriciously spreaded in time and space, often does not satisfy the water demand of plant canopies, on the other hand, from time to time water surplus is generated which can be resulted in inland waters. In dry, drought-stricken years different amounts of irrigation water are needed to enhance the yield safety. The crop canopy water balance is exposed primarily to the permanently changing weather. The role of the climate becomes conspicuous when we are interested in the changes in space or in the effects of a climate change or fluctuation.

Examination of the plant-water-atmosphere system is always a current topic in our country, but nowadays it gets to the center of interest all the more, since it is considered that the dry, droughty periods are more frequent and intense in the last one or two decades, affecting the crop canopy water balance. Also, global warming became a timely problem in the last 10–15 years, whose expectable consequences can cause more unfavorable agricultural water balance in our country. It is obvious, that these suppositions or facts arouse the interest of competent experts whether it can be expected a sensible (favorable or unfavorable) change in the plant-water-weather-climate system in the future in our country. The starting point of these kind of examinations is the survey of the present situation, to which the expectable future situation can be compared.

On the other hand, the research activity has significantly decreased on this field in Hungary. There are very few studies demanding financial and material resources, though, the spread and direct practical use of recent measuring instruments and computing techniques, as well as modeling techniques would make the more exact description and more correct biological and physical explanation of the soil-plant-atmosphere system possible.

This paper gives a short overview of the relationships between the crop canopy water balance and the weather, based on the results of agrometeorological researches carried out at the Hungarian Meteorological Service. We try to make a comprehensive study on the questions and answers raised in the course of the examination of evapotranspiration.

For an objective analysis of the effects of an expectable global warming on the evapotranspiration, and irrigation water demand it is essential to clear the possible answers given to the following questions.

2. How does the permanently changing weather influence the water demand (optimal evapotranspiration) of crop canopies?

There is an objective answer for this question only if the characteristic parameters of crop canopies (variety, plant density, fertilizer- and water supply) in the field experimental researches are the same year by year, whilst the weather conditions are changing. We had been carrying out studies for different species planted in compensatory evapotranspirometers for years (*Antal*, 1966). The water consumption of different plant species had been measured from sowing to harvest by lysimeters with good water supply. From the experimental series we chose maize and potato from the period of 1963–1968 (variety, fertilizer supply, plant canopy density, and water supply were the same during this period). There were extremely hot and dry (1963, 1968) as well as cold and wet (1965, 1966) years in the field experimental period.

Figs. 1 and *2* definitely show that the water demand changes within wide bounds depending on the weather conditions. (The same experimental conditions and varying weather conditions in the course of period 1963–1968). The water demand of both the potato and maize canopies is 30–40 percents higher in a hot and dry growing season than in a cold and wet vegetation period. The influence of the weather parameters is more obvious if we examine the course of the daily water demand in the growing season in years with different weather conditions (*Antal*, 1966, 1998) (*Figs. 3* and *4*).

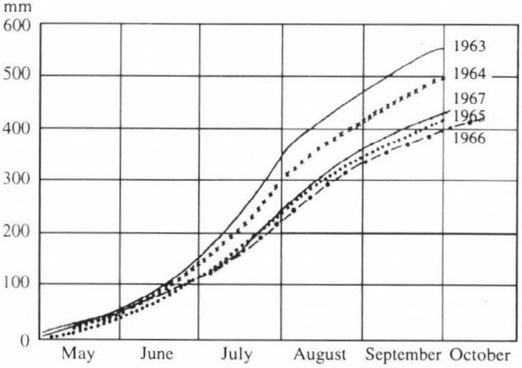


Fig. 1. Accumulated water demand (ET_{opt}) of maize canopy in the years 1963–1967 with different meteorological conditions (Szarvas).

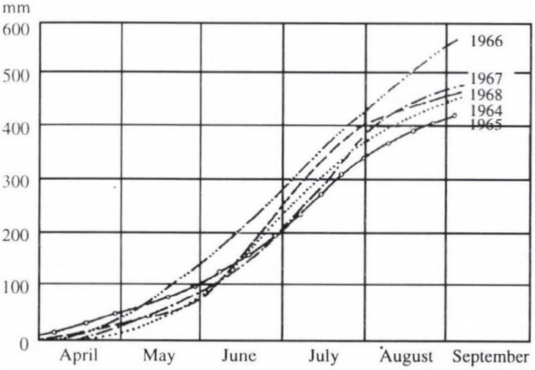


Fig. 2. Accumulated water demand (ET_{opt}) of potato canopy in years 1963–1967 with different meteorological conditions (Szarvas).

As *Fig. 3* shows, in the period between sowing and spring, the optimal evapotranspiration (ET_{opt}) of potato canopy is 1.5–2.0 mm/day, which is practically equal to the evaporation of moist soil surface. After this period, water consumption increases with the quick growing of the plants. In the critical development phase (blooming and intensive tuber growth), daily maximum of the water demand is 4–6 mm depending on the weather. On certain hot and dry days, the daily water loss was as high as 8 mm, while on cool and cloudy days it did not exceed the value of 2–3 mm, even in the period of apex development. After the ending of blooming, the daily water demand of potato canopy quickly decreases, and it less depends on the weather conditions.

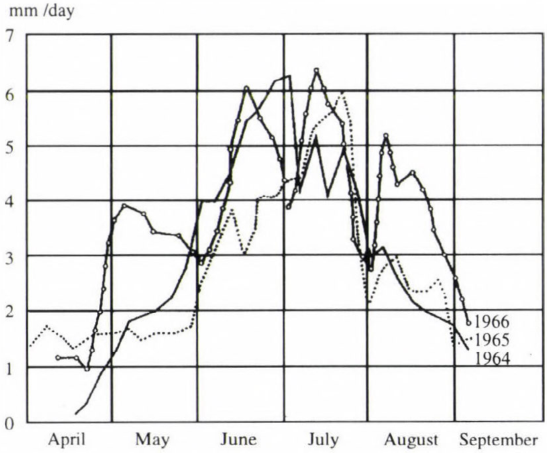


Fig. 3. The annual course of potato ET_{opt} in the period 1964–1966 with different weather conditions.

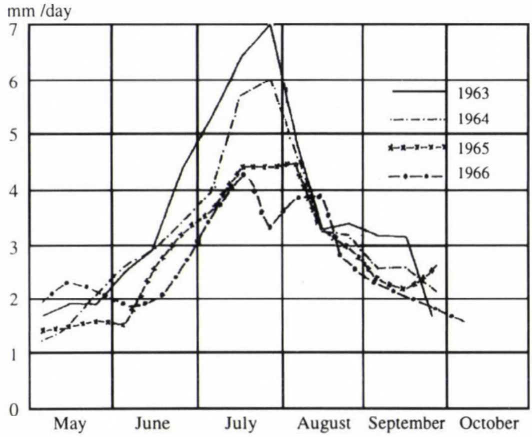


Fig. 4. The annual course of maize ET_{opt} in the period 1964–1966 with different weather conditions.

Similar conclusions can be drew on the water consumption of maize canopy (*Fig. 4*), with the difference that in the period following the critical development phase (head flowering, tasseling) of the individual years, the water consumption of maize canopy is less dependent on the weather conditions than the evapotranspiration (ET) of potato canopy.

Data presented in *Table 1* gives an other evidence on the considerable dependence of the water demand of crop canopies on the day by day changing quantity of the heat reversible to evaporation and, indirectly, on the temperature (*Antal*, 1972, 1981, 1998, 2003).

From the blooming and hand flowering phases those periods (3–4 days) were chosen, when there was a quick temperature increase or decrease after a frontal passage. Based on the data of the table it can be concluded, that the temperature change of 7–10°C resulted in an overnight 50–80% change in the water consumptive use of maize canopy, without any change in the soil moisture content of the root zone.

Table 1. Changes of the daily water consumptive use of potato- and maize canopy as a function of the air temperature (Szarvas, silty clay soil)

Period	**Water demand of potato mm/day**	**Temperature daily average °C**	**Period**	**Water demand of maize mm/day**	**Temperature daily average °C**
1964 June 28	7.1	24.2	1963 June 26	10.0	25.9
29	7.7	23.2	29	2.3	17.5
30	4.3	11.5	30	3.5	17.8
1965 June 24	6.2	23.4	1964 July 22	9.4	27.1
25	6.4	24.4	24	3.1	20.8
28	2.6	16.5	25	3.6	19.7
1966 June 3	2.8	14.4	1965 July 26	7.1	22.6
July 1	2.4	16.7	27	2.4	16.9
July 5	7.1	24.6	28	3.7	18.0
1967 June 8	7.0	22.0	1966 July 20	7.3	24.0
9	1.3	17.3	23	1.7	17.0
10	1.4	16.5	1967 Aug 7	2.2	18.8
11	0.6	11.4	10	8.2	25.3
1968 June 9	7.4	22.9	1968 July 15	11.0	24.7
10	2.5	13.1	18	2.5	17.1
11	2.0	12.4	19	2.7	16.1

The above presented temperature dependent water demand fluctuations could be detected also in vegetable-, fruit-tree-, and vineyard canopies (*Cselőtei*, 1959; *Gergely* and *Stollár*, 1980; *Stollár* and *Gergely*, 1978; *Füri* and *Kozma*, 1975, 1978).

On the base of the above described results, a general answer can be formulated for the first question. Namely, *the yearly and year-by-year changes of the weather are definitely followed by a change in the optimal evapotranspiration, i.e., water demand of crop canopies, even if the soil moisture content availability in the root zone is optimal in irrigated circumstances.*

An other question raised in connection with the determination of the irrigation water demand was the role of the species, variety, and developing phase in the shaping of water consumption of crop canopies. Therefore, the second basic question was the following:

3. How do the species, variety, and developing phase effect the shaping of the water demand (optimal evapotranspiration)?

There are many ways to study the effects for getting an answer to this question. A definite and quantitative answer can be given only on the base of the results of a water balance and ET measurements, which are carried out under the same meteorological conditions. On the base of the few decades long evapotranspirometer experiments carried out at the agrometeorological observatory of Szarvas, a definite answer can be given for the above expressed question. *Fig. 5*, which was drawn in our earlier studies (*Antal*, 1966, 1998), presents the accumulated curves of the optimal evapotranspiration (water demand) of five different plant species. The curves suggest that there is a significant difference in the water demand of individual crop canopies, even with the same soil texture, water supply, and meteorological conditions. If in thought we transform the sowing times into the same point of time, there are still decided differences among the accumulated curves. For example, the winter wheat finished its vegetation period with 400 mm accumulated water consumption, while the sugar beet used 25% more water in its vegetation period, and the accumulated water consumption of the alfalfa crop canopy was about 700 mm, in the same year. The most evident differences among the individual species appear in the accumulated water consumption. On the other hand, since all the field experimental conditions were the same, the possible cause of the significant differences experienced in the water demand can be searched in the different biological and plant physiological features, as well as in the individual characteristics of the crop canopy type (plant canopy density, leaf area, different radiation balance).

The effect of the plant species difference on the water demands are more clearly demonstrated by the curves of *Fig. 6*, which presents the daily amounts of the optimal evapotranspiration in case of different plant species (*Antal*, 1966, 1972, 1998, 2003; *Antal* and *Szesztay*, 1996).

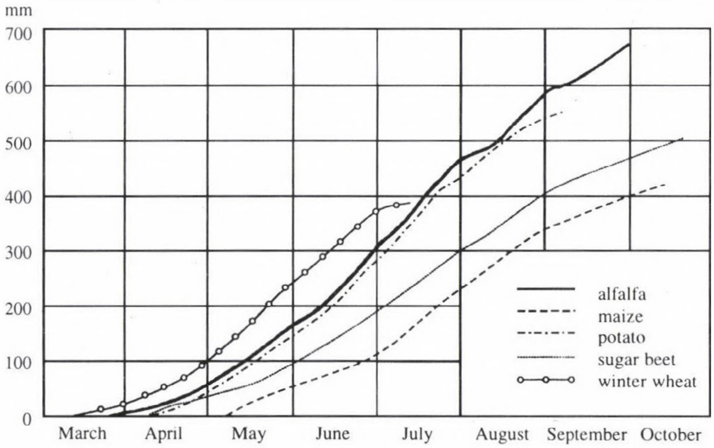


Fig. 5. Accumulated curves of the ET_{opt} in case of different plant canopies (Szarvas, 1966).

Three basic conclusions can be derived from the course of the curves of *Fig. 6*:

- there are definite differences between the daily water demands of the species (on certain days, a plant canopy with higher water demand uses 2–3 times more water than a plant with lower water demand, e.g., in the middle of May);
- the water demand of all crop canopies remarkably fluctuates during the growing season, which indicates the effect of changing meteorological conditions;
- it can be stated, too, that the maximum of daily water demand coincides with the critical development phase in the case of all species (it is not obvious in the case of pasture as it is mown a few times in the vegetation period).

Therefore, there is an unambiguous answer for the second question: *there are significant differences in the water demand and water consumption of the individual plant species, even in the same periods and meteorological conditions, and the period of maximal water demand definitely coincides with the time of the critical development (usually blooming) phase.* Taking into account these facts is of vital importance in the irrigation farming.

In both irrigation- and natural farming, the role of the plant canopy density in the formation of the yield harvest and water demand is a matter of many disputes. In irrigated conditions, the size of the optimal plant number per hectare is influenced by the light supply, whilst in circumstances without irrigation it is influenced by the soil moisture content in the root zone. This discussion emerged the third question of the evapotranspiration researches:

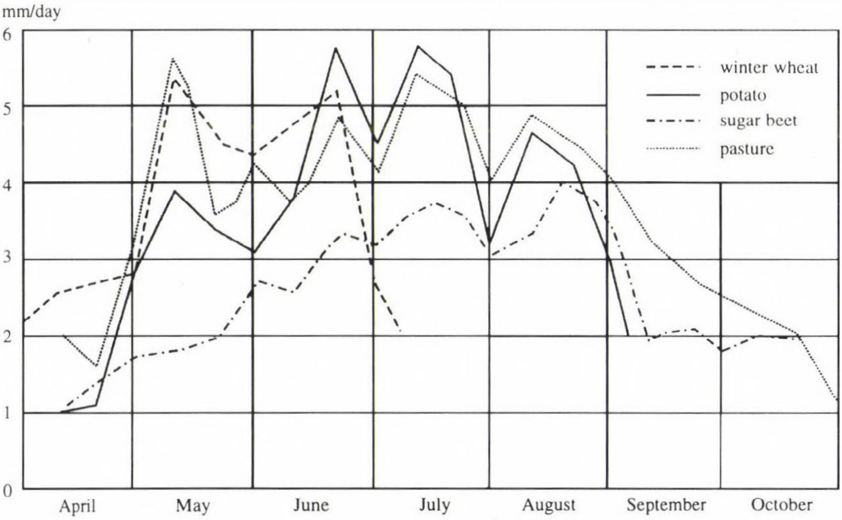


Fig. 6. Dependence of the daily ET_{opt} on the different plants species in course of the growing season (Szarvas, 1966).

4. What kind of relationship can be found between the plant canopy density and the development of water demand?

It is obvious, that in this case every effecting factors (water- and fertilizer supply, variety) have to be the same. Our study was based on data measured by evapotranspirometers in maize canopies with different plant density (*Antal et al.,* 1974, 1975).

The main features of the relationship between the evapotranspiration and plant stem density are summarized in *Table 2.*

The table shows that with increasing plant density (from 4 to 6 stems per m^2) the water demand raised with 19%. It means, that the corn canopy of 60,000 plant numbers per hectare used 1050 m^3 more water than that of 40,000.

It was not possible to determine an optimal plant density for water consumption and yield without any doubt. In cooler years with less sunshine, the less dense plant canopy assured more yield and water consumption, while in warmer vegetation periods, which were rich in sunshine (for example in 1968, see *Table 2*), the most dense corn canopy produced the biggest yield and water consumption.

Table 2. Optimal evapotranspiration measured by evapotranspirometers in corn canopy with different plant stem density per m^2, mm (Szarvas, 1968)

Month	**Plant number per m^2**			
	4	**5**	**6**	**7**
April	10	11	11	11
May	43	48	54	54
June	135	145	154	187
July	189	209	229	284
August	126	138	150	176
September	61	65	69	99
October	5	6	7	11
April–October	569	622	674	822

Note: April and October are fragmented months (few days only)

Therefore, there is no definite answer for the third question to the experts of irrigation. Results of the examinations can be summarized only generally: *with increasing plant number per hectare, the water demand definitely increases, however, the plant canopy can not be given with a particular value, since it depends on the probable weather of the vegetation period, and this dependency can not be compensate even with the most favorable water- and fertilizer supply.* It would help if a weather prediction with acceptable accuracy would be given in the time of sowing at least to the end of the critical development phase, and the sowing plant number per hectare would be determined as a function of that.

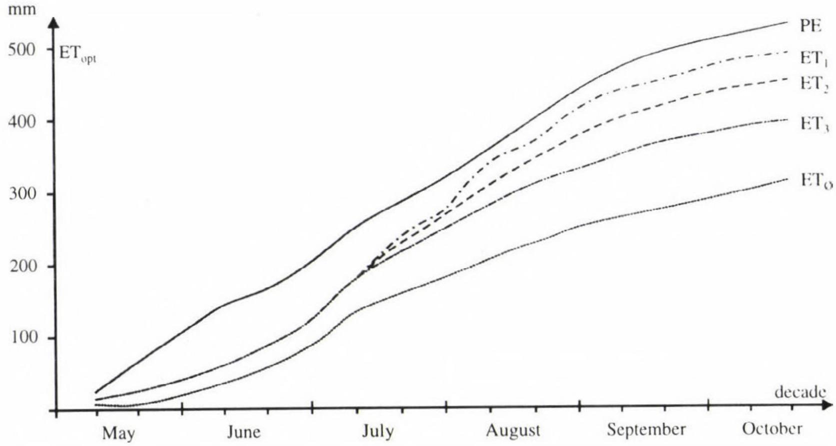
Development of the irrigation emerged the so far actual question, that farms working on soils with good productivity need to be preferred to those with weak productivity in the course of irrigation investments. An objective decision postulates studies on numerous economic, market, social, and farming technologic aspects. Experts of this field ask the fourth question concerning the water balance of the plants:

5. Does the fertilizer supply influence the water cycle of plant canopies, if yes, in what extent?

The question can not be narrow down to the analysis of interaction between the nutrition supply and water consumption, since the crop result appears as a standard of value. On this field, evapotranspirometer investigations have been carried out for years in Szarvas and Keszthely, examining few plant species with different amounts of combined nitrogen, phosphorus, potassium artificial fertilizers (NPP) (*Antal et al.*, 1975; *Endrődi*, 1978; *Tóth*, 1978).

From the decade long measurement series we pick out the results concerning maize canopy. In this case, we tried to give an answer on the basis of four measuring series with different fertilizer supply (0, 120, 240, 480 kg/hectare NPP efficient substance) and three series with different water treatment (*Antal et al.*, 1975; *Tóth*, 1978; *Posza* and *Tóth*, 1978).

Fig. 7 shows the accumulated curves of the optimal evapotranspiration (ET_1, ET_2, ET_3) of corn canopy in the case of different fertilizer supply, as well as the actual evapotranspiration of the plant canopy with natural water supply (precipitation) in the surroundings of the evapotranspirometers ($ET_ø$). The upper curve is the calculated potential evapotranspiration (PE).



Legend: PE potential evapotranspiration; ET_1 and ET_2 optimal evapotranspiration in case 480 and 240 kg NPP with irrigation; ET_3 optimal evapotranspiration of irrigated canopy without NPP; ET actual evapotranspiration of controlling plot (no irrigation, no NPP).

Fig. 7. Accumulated curves of the maize canopy evapotranspiration at the different NPP supply (Keszthely, 1973).

The results of investigations can be summarized as follows. In the case of medium fertilizer supply (240 kg/hectare NPP), the water demand of maize canopy increased with 13%, in the case of large amount (480 kg/hectare NPP), the water demand increased with 22% comparing to the plant canopy without fertilization. However, the excess fertilizer did not increased the water demand in the same degree, which is proved by *Table 3*.

Table 3. Increase of the water demand of maize canopy comparing to the controlling plot

Year	**Increase of water demand, 240 kg/hectare NPP, %**	**Increase of water demand, 480 kg/hectare NPP, %**
1971	14	22
1972	8	9
1973	7	13
Average	**9.7**	**14.7**

Effects of the nutrition supply to the water demand is well demonstrated by the water consumption coefficient as well (to be expressed in liters of water used to corn cob yield production), which is presented by *Table 4* on the base of the experiments (*Antal et al.,* 1975).

Table 4. Maize water consumption coefficient in the case of different fertilizer supply

Year	**Controlled plant liter/kg**	**240 kg/hectare NPP liter/kg**	**480 kg/hectare NPP liter/kg**
1971	592	431	412
1972	1090	467	430
1973	1404	539	445

The maize water consumption coefficient changes from year to year depending on the meteorological conditions. Similar results arised from examining the transpiration coefficient of vegetable canopies (*Cselőtei*, 1964).

It can be seen, that excess fertilizer makes the water efficiency better by means of bigger crop results (on the controlling parcel the water consumption coefficient declines from year to year, since the crop result decreases in the lack of fertilizer supply).

The answer for the fourth question is the following: *Increasing of the NPP doses leads to the increase of the water demand and water consumption, nevertheless, the irrigation demand depends on the year-by-year changing meteorological conditions*. Moreover, irrigation can be built into the farming technology only if it is rewarding in average of many years. As our investigations pointed out (*Antal et al.*, 1975), beyond a certain extent, even the increasing of either the artificial fertilizer or irrigation water was not resulted in the efficiency of the yield.

A widely discussed question is the role of leaf area in water consumption of plant canopy. The majority of formulas, developed to evaluate ET, do not contain this feature of plants, nevertheless, many researchers tried to involve it in the computational methods more or less successfully. It is an understandable aspiration, as the leaf area projected to 1 m^2 soil surface changes within wide limits depending on species, canopy plant density, and developing phase. While the leaf area of a well developed maize canopy is 2.5–3.5 m^2/m^2 soil surface, leaf area of a closed, well developed alfalfa crop canopy is much bigger.

In this case the following question is raised:

6. What role does the leaf area play in the change of evapotranspiration amount of crop canopies?

Leaf area plays twofold role in influencing the evapotranspiration. On the one hand, it increases the transpiration area, on the other, by interception of precipitation and sprinkler irrigation water it decreases the water income of the root zone, because the interception water evaporates from the plant without utilization.

In this case the question is raised in the following form: what order of magnitude of the water, first intercepted from the precipitation and sprinkler irrigation by the leaf area, is evaporated without utilization, and whether this amount should be taken into account in drawing up of irrigation timetable. How should this water amount be taken into account in determination of irrigation water norm, single irrigation water dose, and irrigation time (day or night).

The evaporation deficit of the sprinkler irrigation is composed of the accumulated amounts of water evaporated from the irrigated water drops, plants, and soil surface. The quantity of the deficit depends on the leaf- and steam surface of plant canopy, as well as on the evaporative demand of atmosphere. On cloudy, cool days and in the night hours the evaporation deficit is minimal, under hot, dry, clear, windy weather conditions it can maximally reach the value of net radiation of the plant canopy expressed in water millimeter.

In Hungary, in the summer months, under average weather conditions the daily net radiation of plant canopies is 150–250 MJ/m^2 depending on the plant species (*Dávid et al.*, 1990). The evaporative equivalent of this value is 3.3–6.0 mm/day. Thus, in the case of continuous daytime irrigation, the maximum evaporation can fluctuate within this limits, supposing that the net radiation is fully consumed by the evaporation, i.e., the heat converted to warming up the soil and air is zero.

Comparing the 3.3–6.0 mm maximum amount of evaporative water to the 30 mm irrigation dose, the evaporation deficit of the irrigation water is given to be 11–20%, i.e., 24–27 mm water reaches the root zone from the 30 mm irrigated water. *Using smaller irrigation water doses, the relative evaporation loss (expressed in % units)is greater, with bigger water doses the loss is less. It is obvious, that in the case of plant canopies with bigger leaf area and under highly evaporative weather conditions, 10–20 % evaporation loss of the sprinkler irrigation needs to be taken into account in determining the amount of irrigation water.*

The above discussed questions can be numerically and objectively answered only by experimental measurements and analysis. These kind of investigations will be requisite in the future as well, as the results up till now have to be strengthened and corrected, and the continuously developing farming technologies emerge further questions. On the other hand, analyzing conditions of the water cycle of plant canopies improve, since the measuring instruments and data processing and analyzing methods are developed very quickly, too. It can also be supposed, that our membership in the European Union will be a great challenge for the Hungarian food industry as regards farming efficiency as well as qualitative expectations. Since under Hungarian climatic conditions the precipitation supply is a limiting factor for both efficiency and quality, prospectively the farmers, leaders of the management of water-supplies of the agriculture, and experts will claim for correct and reliable research results regarding water- and irrigation water demand as well as economical and effective water supply of plant canopies. Our researchers can fill these expectations if the concerning special fields take the initiative in closer co-operation. That is, as the above discussed tasks made it perceptible, analysis of the effects of the weather and climate to the soil-plant-water system is an interdisciplinary task.

7. The role of the climate in shaping of water demand and irrigation water need of plant canopies

As it was mentioned in the introduction, climatic features play important role in planning of the irrigation system of plant canopies. During the planning of the irrigation system (irrigation canal, pump capacity, size of the irrigated area, etc.), a claim is raised to necessary basic data, which are well-established in climatic point of view. Such data are, among others, the average of many years, expectations with different probabilities, fluctuation in time, areal distribution, etc., of the water demand and irrigation water need of plant canopies.

Based on the climatic database, being at our disposal, the parameters are suitable to be calculated for as long time as possible (possibly for 50, 100 years), partly for avoiding the presence of short time climate fluctuations in the averages and particular frequency threshold values. The other reason is that building up of the irrigation system for long time needs basic climatological data representing few decades.

For well-established planning of irrigation systems, the basic irrigation water need data have to be determined by regions and plant species. The most expedient method is to derive empirical distribution functions from the time series, as the function curves represent the irrigation water need expected with any possibilities.

As an example, *Fig. 8* presents the empirical distribution function of irrigation water need calculated from the almost 100 years long data base of our earlier studies (*Antal*, 1991, 1998), for three different climatic zones in Hungary and two plant species. Irrigation water need expected with any percentage frequencies and the average value belonging to the 50% can be read from the curves. The ending points of the curves indicates the maximum and minimum values occurred in the processing period.

If the irrigation water need data calculated for every vegetation period are put in order of magnitude and the year of occurrence is signed, we can reply to the question whether the droughty growing seasons were more frequent in the last one or two decades. Because of the size of the data table, detailed values are not presented, we only note that all of the three climatological stations reported more occurrences of extreme irrigation water need in the last two decades than in any other two decades of the 20th century.

Summarizing our investigations on water demand, water consumptive use, and irrigation water need of plant canopies, we can say that these features change by climatic zones and plant species depending on the weather conditions. Consequently, the recent irrigation systems have to be based on the following up the dynamically changing water demand and water deficit of the given plant canopy. The basis of planning of irrigation system developments can be the irrigation water norm determined by farming regions and plant species from long climatological time series. The average values and expectations with different probabilities of the irrigation water norm have to be known as well.

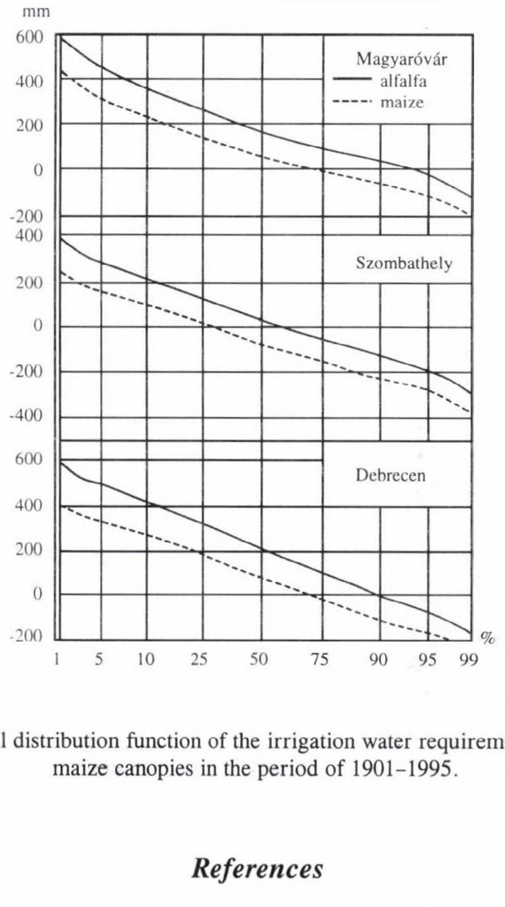


Fig. 8. Empirical distribution function of the irrigation water requirement of alfalfa- and maize canopies in the period of 1901–1995.

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Seasonal changes of soil moisture in Hungary

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Abstract—Soil moisture is a basic enviromental factor determining the water supply of plants. Amount of this meteorological element is influenced by precipitation and evaporation. When rainfall exceeds water loss, water content of soil increases, and this time spell is called as wet period. In a period dominated by evaporation, water uptake of plants becomes more difficult and it is designated as dry period. A critical period from agricultural point of view is when the water shortage in soil can coincide with a high temperature stress, since this phenomenon can be a serious risk factor for plant production.

We can see that seasonal changes in water balance of soil can be determined by the help of climatological data of precipitation and evaporation. This calculation is based on water balance equation, which can be described by models of different complexity. First, the different indices expressing changes in water supply were studied, then wet and dry periods of an average year were determined by the help of an index called relative evaporation.

Climatological data measured in observational network of Hungarian Meteorological Service during the time period of 1951 and 1990 were used for studying the question.

Key-words: soil moisture, relative evaporation, dry and wet period, drought, seasonal changes.

1. Introduction

The plants are fixed in soil by their roots, and their green parts are surrounded by air over the surface, therefore, the value of soil moisture and humidity of the air, as well as evaporative power of the air (potential evaporation) moving water from the soil to the assimilating organs (mainly to leaves) are equally important for them. Besides, the water plays role in moving nutrition solved in water to assimilating organs. Therefore, water supply of plants should be continuous. This is based on soil moisture, which is varying according to the rainfall amount and evaporation. In a wet period, when the rainfall amount exceeds the water loss, the water income is determinant, and the water content of soil is increasing. In a period without precipitation, the evaporation becomes prevailing, and the water content of soil begins to decrease.

The seasonal changes of precipitation and quantity of incoming energy determine the annual course of soil moisture, too. This is of basic importance for agricultural crops because of the knowledge of irrigation water need. Hence, an agroclimatological analysis of this phenomenon is essential in a country dealing with plant breeding.

2. Material and methods

Seasonal changes in water balance of soil can be determined by the help of climatological data of precipitation and evaporation. This calculation is built on water balance equation, which can be described by models of different complexity. A relatively simple model was elaborated earlier for Hungary using climatological data (*Dunay et al.,* 1968, 1969). This model makes up the data of observational network including potential evaporation data of pan „A", which are very useful for practical purposes (*Stanhill, 2002).*

A simple form of water balance equation is the following:

$$W = W_0 + P - E\,, \tag{1}$$

where W is the soil moisture measured at the end of investigated period, W_0 is the soil moisture measured in the beginnig of the examined period, P is the rainfall amount, and E is the actual evaporation of the same period. Using this model we have to assume, that there is no water flow in horizontal direction, there is no capillary rise, and the water surplus over field capacity will infiltrate to lower layer of soil, under the plant root zone. It is important to build up a water balance model, which can be run with readily available inputs (*Ritchie,* 1985; *Brisson et al.,* 1992).

It is practical to characterize the soil moisture conditions by the aim of relative value of soil water balance. This method is based on radiation aridity index (*ARI*) elaborated by *Budiko* (1956), and can be written in general form as

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

where E_0 is potential evaporation and P is the precipitation. The ratio of potential evaporation and precipitation was represented in graphical form and designated as climate diagram of *Walter* (1985) for characterizing of different types of climate. Earlier we have analyzed the values of this index for Hungary (*Varga-Haszonits,* 2002).

For a long time, it is usual to calculate the water balance of soil by using relative evaporation (*RE*) or evaporation index, which is ratio of actual evaporation to potential evaporation and can be written as

$$RE = \frac{E}{E_0}. \tag{3}$$

This value is in very close connection with the water content of soil. Therefore, it is a useful method for describing the seasonal changes of soil moisture.

Sometimes it is practical to use the crop water stress index (*CWSI*), which is a version of relative evaporation expressing the ratio of the difference between potential and actual evaporation to the potential evaporation (*Jackson,* 1982), that is

$$CWSI = \frac{E_0 - E}{E_0}. \tag{4}$$

The values of *CWSI* make the values of relative evaporation up to the value of 1. This is the reason why it is sufficient to use only one of two. In this paper, the relative evaporation is calculated.

Climatological data measured in observational network of Hungarian Meteorological Service during the time period of 1951 and 1990 were used for studying the water balance of soil.

3. Results

3.1 Effect of precipitation and evaporation on the change of soil moisture

The rainfall is the main source of water income of soil. Amounts of rainfall vary significantly during the year as *Fig. 1* indicates. The annual course of precipitation is very similar over the whole territory of Hungary. Differencies are found only in the amount of rain. South-western part of the country has the highest amount of rainfall, and southern part of central Hungary has the lowest values.

The evaporation is the basic loss of water content of soil. Potential evaporation shows what amount of water can be evaporated from the free water surface. The amount of evaporated water is mainly depending on solar energy reaching the surface, water vapor deficit in the air, and wind speed. As we can see in *Fig. 1*, potential evaporation indicates a seasonal variability within the year, which follows the annual course of solar radiation. The annual course of potential evaporation is similar in the whole territory of Hungary, but the amount of evaporation differs from area to area. Highest values of evaporation occur in the southern part of central Hungary, and the lowest are found in the south-western part of the country.

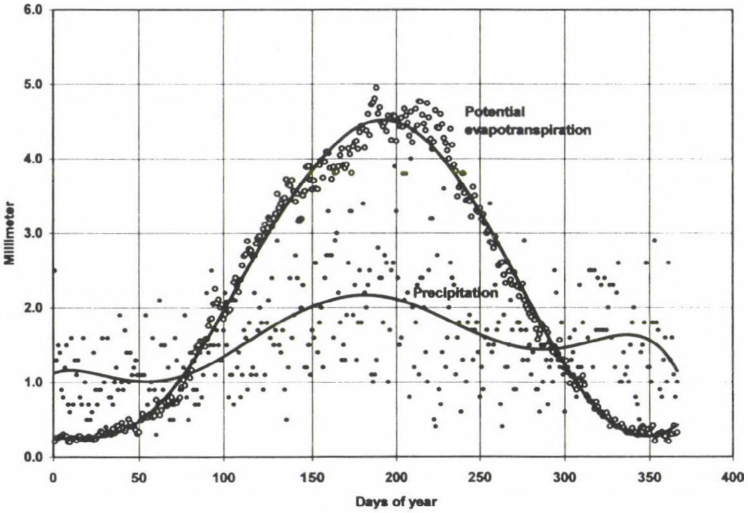


Fig. 1. Annual course of potential evapotranspiration and precipitation in Mosonmagyaróvár (1901–2000).

As we can see in *Fig. 1*, the rainfall amount during the cool period (from November to March) exceeds the amount of evaporation, therefore, this period is a wet period. In this period the rain water can be accumulated in soil. From March the evaporation exceeds the rainfall, thus, soil moisture continuously decreases. This period is called dry period and lasts to the November, when the rainfall amount becomes higher than the evaporation. During this time, the water is primarily the most important environmental factor for plants, and this is why the plant breeders have an increasing risk according to the drought.

Drought can be defined as a significant water shortage for a longer period (*Palmer*, 1965). Only shortage of rainfall itself does not cause dryness. Dry period, however, begins with a low level or total lack of rainfall. Especially in warm period of the year, the low level of rainfall is coupled with intensive radiation and low relative humidity of air, therefore, the evaporation is increasing. This is the reason why the soil water content is decreasing to such an extent, that plants suffer from water deficiency. Generally, the dryness results from the combination of low precipitation and high evaporation (*Larcher*, 2003).

Finally, rather close relationships can be determined between drought and crop yield. On the basis of these relationships we are able to characterize economically the drought as risk factor (*Varga-Haszonits* and *Harnos,* 1988).

3.2 Seasonal changes in the indices of relative evaporation

In many cases, the soil moisture conditions were characterized by relative evaporation. This was expressed — as we mentioned earlier — by the ratio of actual evaporation to the potential evaporation, where the actual evaporation was calculated as evaporation of bare soil, and the potential evaporation was determined as the evaporation of free water surface (measured by pan „A") under given meteorological conditions. This index value is in a very close connection with soil moisture.

We calculated the average daily values of relative evaporation from 40 years data series for Zalaegerszeg, which is in the wettest area of Hungary, and for Szeged, which is in the dryest territory in Hungary. We can see in *Fig. 2* that in the winter time, when precipitation amount exceeds the amount of evaporation, the values of relative evaporation vary most intensively. In this period, the values in both stations are very close to the maximum values. Consequently, there are no essential differences between the wettest and dryest area in the respect of the length of wet period. After the maximum in February, however, the values of *RE* decrease continuously to the minimum in August. The dryest part, Szeged, indicates a double minimum in August and September. In autumn the values of *RE* begins to rise again to the maximum at the end of winter, but variability of the values increases.

The curve of annual course shows a rise in values of *RE* between 150th and 180th days of the year. In this time a rainfall maximum is found in Hungary. In the area with various moisture content, the differences in soil moisture content are the highest in autumn. The reason of this fact is that in summer time precipitation falls mainly in the form of shower, that means a lot of quantity of rainfall in a short time, then a long dry and warm period follows with high evaporation values. Soil moisture decreases in this latter time. The values of evaporation also strongly decrease, the values of *RE* indicate low evaporation too.

It is worth noting, that annual course of *RE* can be represented by a polinom of six degree with a determination coefficient $r^2 = 0.97$ (as can be seen in *Fig. 3*). This shows a very close relationship. It makes possible to model this phenomenon climatologically.

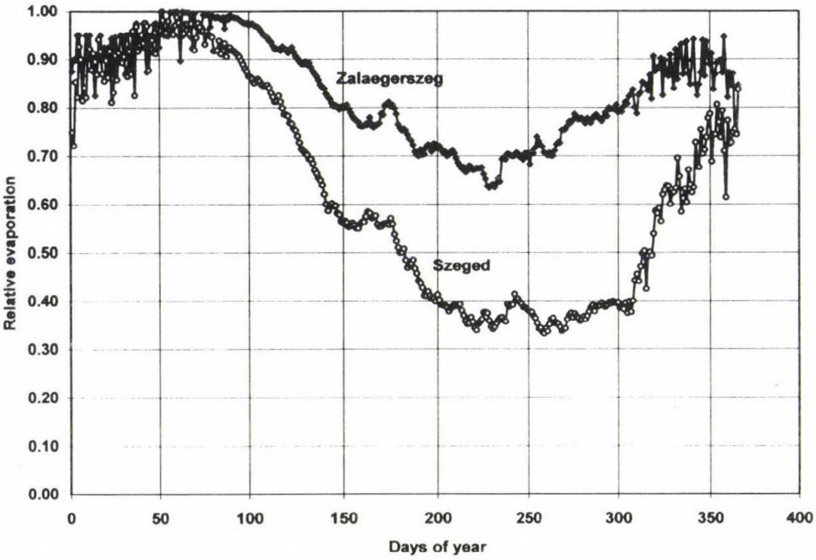
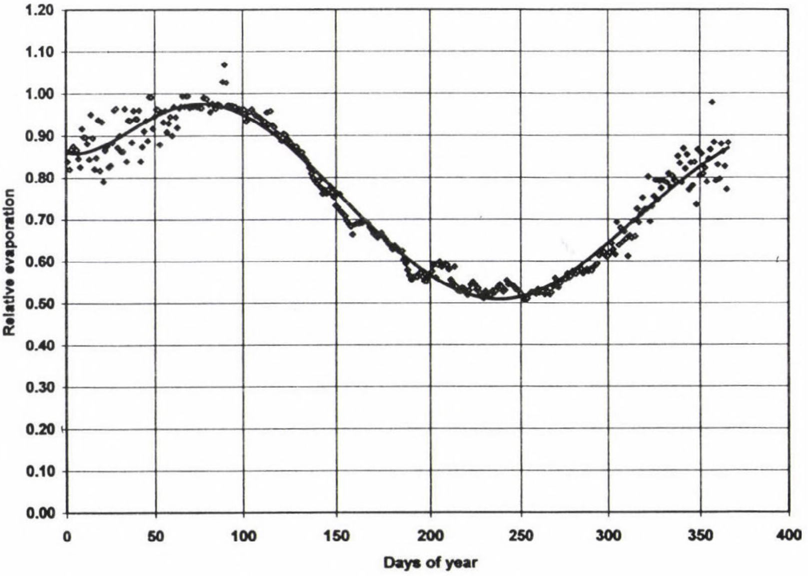


Fig. 2. Annual course of relative evaporation in a moist (Zalaegerszeg) and dry (Szeged) area (1951–1990).



3. ábra. Annual course of relative evaporation in Mosonmagyaróvár (1951–1990).

3.3 Seasonal changes of soil moisture

Finally, we determined the seasonal changes of soil moisture content by using a soil moisture model elaborated earlier *(Dunay et al.,* 1968; *Dunay et al.,* 1969). We calculated the daily average values of water content of soil from 40 years long (1951–1990) data series.

It is important to note, that the average values in winter months are very close to the maximum values of 40 years data series. From the beginning of spring, when the soil moisture begins to decrease, the mean values come more and more nearer to the minimum values. The minimum can be found nearly in the middle of August. The soil moisture values can, therefore, be characterized by the help of values of *RE*.

The southwestern part of Hungary is the wettest area of the country, where the maximum values of soil moisture are very close to the field capacity throughout the year (*Fig. 4*). A short dryer time can be found in the beginning of July and in September, but the dry period in September is more significant because of intensity of dryness and length of time spell. In the dryest area of the country, the maximum values could sometimes be quite low, near to 80% from the middle of summer to the end of autumn, even in some cases they fall short of 80% (*Fig. 5*).

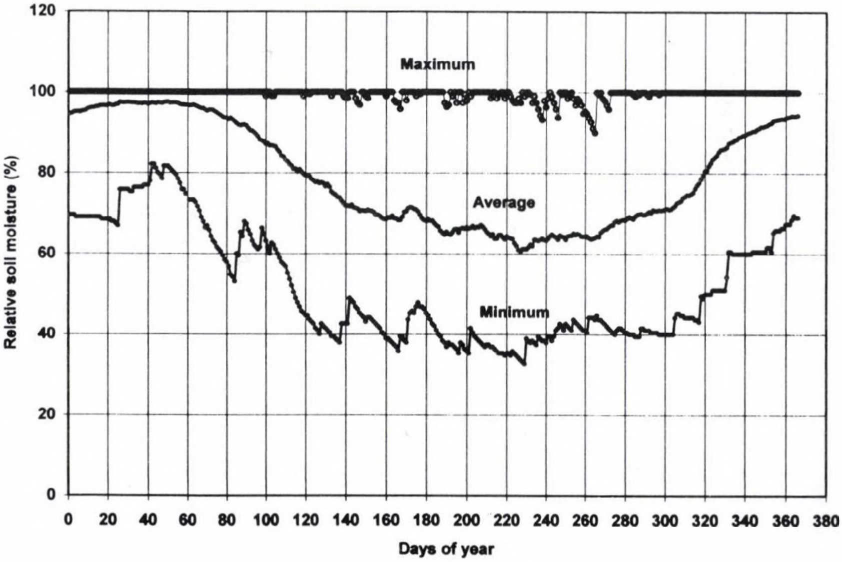


Fig. 4. Annual course of minimum, average, and maximum values of relative soil moisture in a moist area (Zalaegerszeg, 1951–1990).

Seasonal changes of minimum values are very similar to those of averages and maximum. Minimum values fall down under 40% only in the middle of summer in the wettest area, but in the dryest area the minimum values under 40% can occur from the end of spring to the end of autumn.

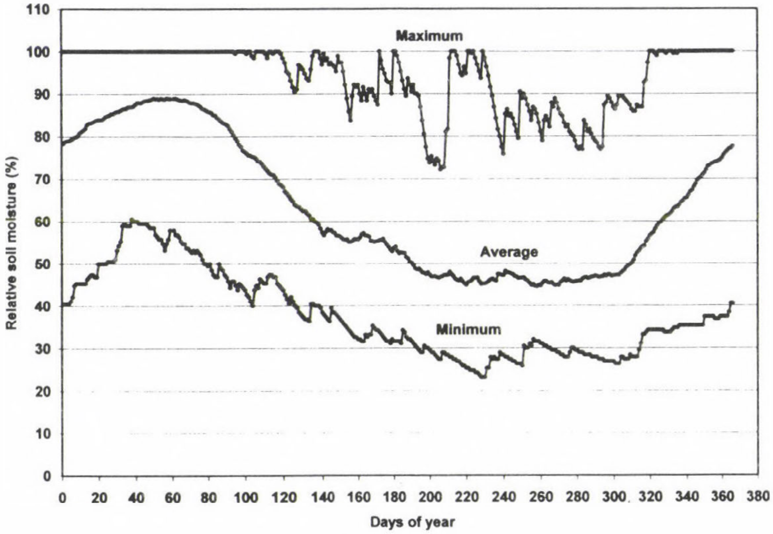


Fig. 5. Annual course of minimum, average, and maximum of relative soil moisture in a dry area (Szeged, 1951–1990).

4. Conclusions

An important feature in the climate of our country — as it can be seen in *Fig. 1* — is the length of period, when the amount of rainfall exceeds the amount of evaporation. This time spell begins with the secondary maximum of mediterranean type of rainfall. After the secondary maximum of rainfall in November, the rainfall is more than evaporation, therefore, if soil is not frozen, water continuously infiltrates into the soil. If soil is frozen and the precipitation falls in the form of snow, the water would infiltrate into the soil after melting. This is the reason why the maximum value of soil moisture occurs at the end of February and in the beginning of March (*Fig. 6*).

In spring the amount of rainfall rises intesively to the maximum of that element in June, but evaporation increases more rapidly because of the strongly rising quantity of energy, thus, soil moisture is gradually decreasing. Summer precipitation falls mostly in the form of shower, and long warm periods without rainfall can cause significant water shortage from the middle of summer to the end of autumn. It is a critical period from agricultural point of view, because the water shortage in soil coincides with a high temperature stress, and this phenomenon is able to rise the risk factor in plant breeding of our country.

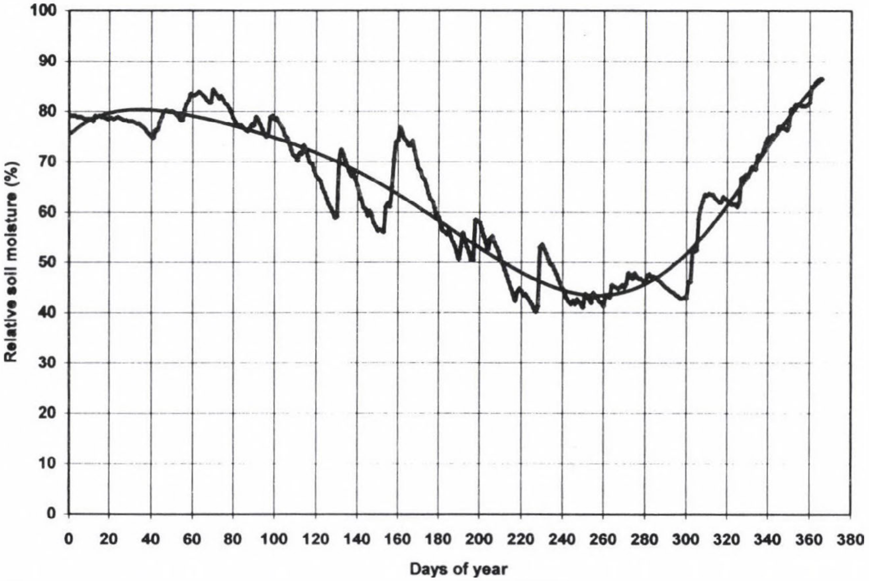


Fig. 6. Annual course of relative soil moisture in Mosonmagyaróvár (1950–1990).

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On the pointing error of pyrheliometers

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Abstract—In this paper the results of calculations are shown on the error caused by the mispointing of pyrheliometers. The penumbra functions of three pyrheliometers (absolute cavity, KIPP, and NIP) are used for two atmospheric situations (high solar elevation and low turbidity as well as low solar elevation and high turbidity) to calculate the output values of pyrheliometers as functions of the pointing error.

Key-words: cavity pyrheliometer, NIP pyrheliometer, KIPP pyrheliometer, penumbra function, pointing error

1. Introduction

The direct radiation is the shortwave (solar) radiation coming from the solid angle determined by the solar disk. The pyrheliometers are designed to measure the direct radiation. Their view limiting angles (slope, opening, and limit angle) are larger than the visible radius of the solar disk. This is partly due for the easier tracking of the Sun: if the limiting angles are larger than the solar disk, it is not necessary to follow the Sun quite precisely.

How large pointing errors or inaccuracies occur in the everyday practice? Let us take a hand-operated pyrheliometer. If its adjustments are made once in a minute, its largest mispointing in azimuth angle would be one quarter of a degree. The deviation from the right position in elevation angle is in the same order. Regarding the pointing devices of the pyrheliometers, 1 mm deviation of the illuminated spot from its proper position could be regarded as large mispointing. Depending on the length of the pointing path, this deviation means about half a degree of pointing error.

The purpose of this document is to present calculated values of the errors in the output of pyrheliometers caused by pointing uncertainty up to 2 degrees.

Two atmospheric conditions are taken into account:

- mountain aerosol, optical depth: 0.07, solar elevation: 60 degrees, direct radiation: 1000 W/m^2;
- continental background aerosol, optical depth: 0.23, solar elevation: 20 degrees, direct radiation: 461 W/m^2.

The calculations have been made for 3 pyrheliometers: the Packrad size cavity instrument (ABS), the KIPP and NIP pyrheliometers. Their slope angles are: 0.75, 1.0, and 1.78 degrees, respectively.

2. The method of calculation

The calculation is based on the Pastiels' theory (see, for example, in *Major,* 1994). The irradiance given by a circular pyrheliometer can be written as:

$$I = \frac{V}{KS} = \pi \int_0^{z_l} F(z)\, L(z) \sin(2z)\, dz\,, \tag{1}$$

where V is the output of the pyrheliometer,
K is the average sensitivity of the receiver,
S is the area of the receiver,
z_l is the limit angle of the pyrheliometer,
$F(z)$ is the penumbra function of the pyrheliometer,
$L(z)$ is the radiance (=sky function) and
z is the angle between the direction of radiance and the optical axis of the pyrheliometer.

Circular pyrheliometer means that all the view limiting diaphragmes and the receiver are circular in shape, that is the whole pyrheliometer has a rotational symmetry around its optical axis. In the equation, the same rotational symmetry is supposed for the solar disk and the circumsolar sky.

If the optical axis of the pyrheliometer is not directed to the solar center, then the angle measured from the solar center ($z1$) differs from the angle measured from the optical axis (z). The transformation is

$$\cos(z1) = \cos(d)\cos(z) + \sin(d)\sin(z)\cos(\varphi), \tag{2}$$

where d is the deviation between the solar center and the optical axis, that is the pointing error,
φ is an azimuth angle measured in the plane of the receiver, it is zero if the radiance comes from the solar center.

2.1 Radiance along the solar disk

Photospheric models of the Sun produce one-dimensional radiance distribution across the solar disk, that is the so called limb darkening function. According to theoretical calculations (*Allen*, 1985; *Zirin*, 1988), the radiance depends near linearly on the cosine of the zenith angle at the solar "surface". Taking into account some observations too (*Zirin*, 1988) and using $z1$ as variable instead of the aforementioned zenith angle, the following radiance distribution along the solar disk has been used:

$$L(z1) = L_0[0.3 + 0.7\,SQR\,[1 - (z1/0.26)^2]]\,, \tag{3}$$

where L_0 is the radiance at the solar center,

0.26 is the radius of the solar disk in degrees.

This way the atmosphere affects the absolute value of the radiance coming from the solar disk, but not the relative distribution along it. If the direct radiation is 1000 W/m^2, then L_0 = $2.01565 \cdot 10^7$ W/(m^2 sr), while at 461 W/m^2 it is $9.29216 \cdot 10^6$. Since the gradient at the solar edge is very large, the step of integration in Eq. (1) has to be 0.0001 degree to obtain 0.1 W/m^2 accuracy.

2.2 Radiance along the circumsolar sky

For several atmospheric aerosol contents and solar elevation angles the radiances coming from the circumsolar sky have been calculated by *Putsay* (1995). To make our calculation more practical, second order polynoms have been fitted to the logarithm of the two selected circumsolar sky functions. The fit is not quite perfect, but it is not significant since we want to obtain the effect of the shift caused by the uncertain pointing.

In *Fig. 1* the whole (solar and circumsolar) sky functions are shown for the two selected atmospheric models.

2.3 The penumbra functions

To make the computations faster, the penumbra functions have been approximated by third order polynoms in the interval between the slope and limit angles. Again, the fit is not perfect, but this has small effect on the deviations of the values calculated for different pointing uncertainty.

3. Results

In the calculations the effect of the solar disk and that of the circumsolar sky could be separated. In *Figs. 2* and *3* the actually direct irradiance of the pyrheliometric sensor can be seen. If the pointing error is smaller than the slope

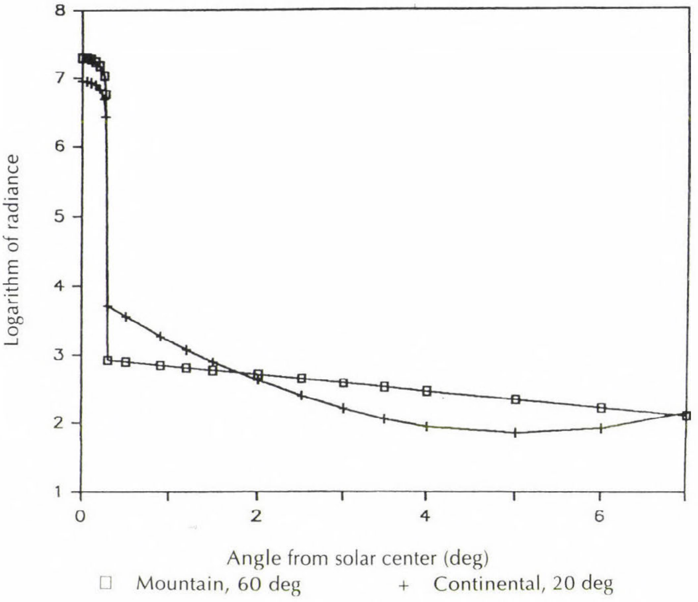

Fig. 1. The sky functions used in this calculation.

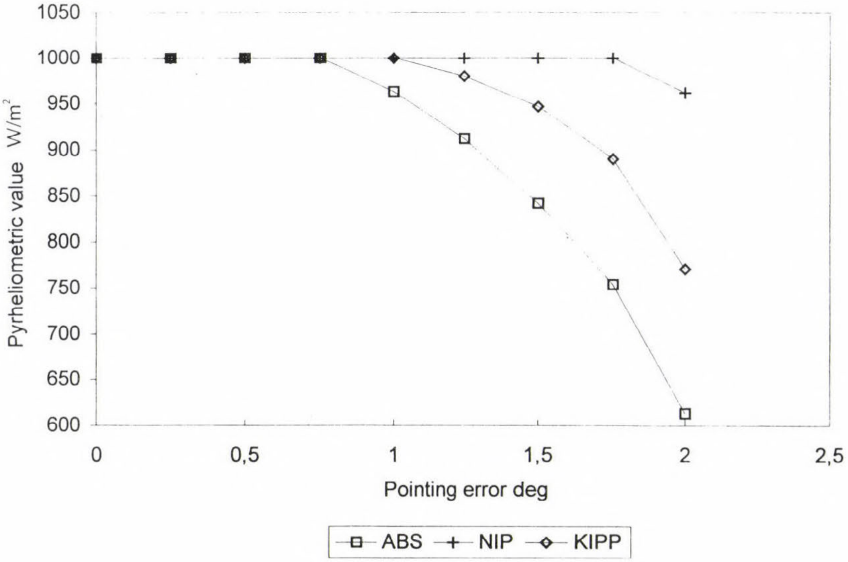

Fig. 2. The contribution of the solar disk to the irradiance of pyrheliometric sensors depending on the pointing error. Case of mountain aerosol and 60 degrees solar elevation.

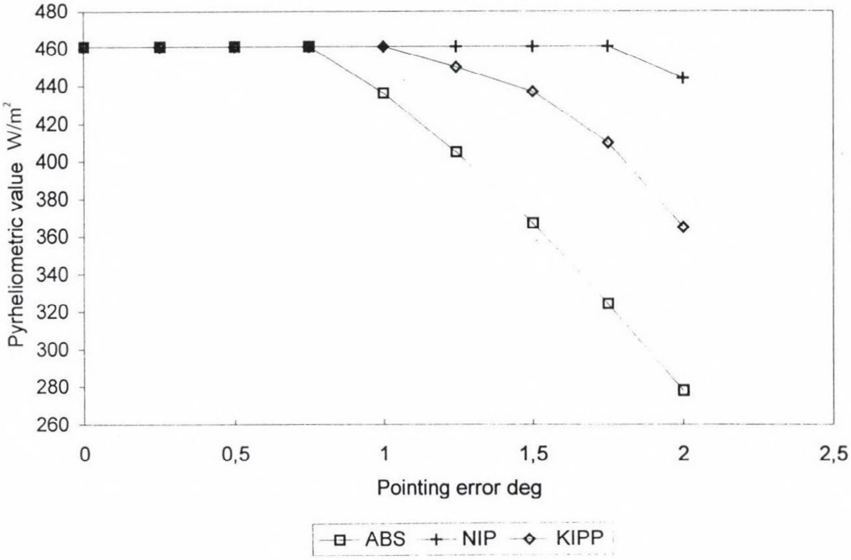

Fig. 3. Same as Fig. 2 except that the case is continental background aerosol and 20 degrees solar elevation.

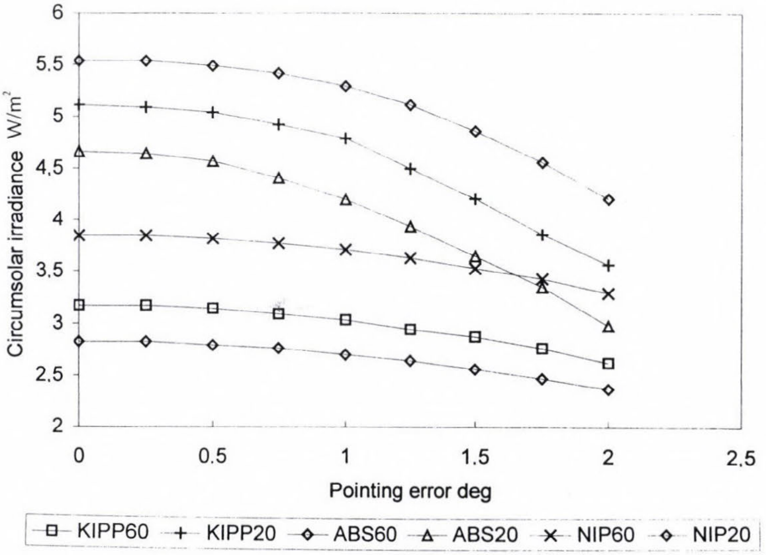

Fig. 4. The contribution of the circumsolar sky to the irradiance of pyrheliometric sensors. The upper 3 curves belong to the case of continental background aerosol and 20 degrees solar elevation, the lower 3 curves belong to the case of mountain aerosol and 60 degrees solar elevation. In both group of curves the instruments are from the top to down: NIP, KIPP, and ABS.

angle, the irradiance is not affected. If the solar disk is in the penumbra region of the pyrheliometer, the irradiance decreases rapidly with the increasing pointing error.

In *Fig. 4* the irradiance coming from the circumsolar sky is seen for all pyrheliometers and both atmospheric conditions. The decrease is continuous, but the effect is not significant compared to that of the solar disk.

4. Conclusions

- If the pointing error of a pyrheliometer is smaller than its slope angle, the effect is negligible.

- If the pointing error of a pyrheliometer is larger than its slope angle, the irradiance of the pyrheliometric sensor decreases rapidly with increasing mispointing. The value can be estimated using *Fig. 2* and *3*.

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Data on the short wave radiation balance and temperature of the Síkfőkút forest

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Abstract—Radiation and temperature measurements were carried out between 1978 and 1994 in oak forest at Síkfőkút, Hungary. The experimental station is located on the southwestern part of the Bükk Mountains at a height of 270-290 meters above sea level. Since about 1980, the forest decay has affected 50-60% of the sessile oak trees (Quercus petrea), which is the most important species (85% of the trees belong to that species); while Turkey Oak trees (Quercus cerris) have not been damaged. The impacts of the deforestation are detectable on the radiation balance and temperature regime. The healthy forest (July 1978) let through 8% of the incoming short wave radiation and absorbed 92% of that. The decayed forest (July 1994) let already 22% of short wave radiation and absorbed 78%. The annual and summer mean temperatures have increased more rapidly in the forest than above the neighboring treeless grassland during the 17 years period examined here. The ten years summer mean temperatures have increased by 2.2°C inside the forest and by 1.8°C above the treeless area. This fact can be explained partly by that the deforestation have decreased the tree density of the forest and partly by the higher frequency of warm years in the examined period. In the leafless forest, high temperatures can be found at the level of the crown of the trees and at the soil surface as well, while in the leafy forest high temperatures can only be found in the foliage. Because of the forest decay, sometimes when skies are clear and weather is calm in the afternoon, a second warm air layer can form near the surface.

Key-words: forest, forest decay, short wave radiation balance, temperature.

1. Introduction

The Department of Meteorology of the University of Debrecen (previously Kossuth Lajos University) joined in the complex ecological study of the Síkfőkút forest (*Jakucs*, 1973) in 1978, when a 25 meters tall meteorological tower had been completed. The project was started by the Department of Ecology.

The forest is situated on a slope of 2–3 degrees, which has a southern exposure at a height of 270–280 meters above see level. It is placed in the Bükk Mountains in Hungary. The average height of the trees was 19.5 meters at the beginning of the ecological studies. The experimental area is a homogenous Turkey-Sessile oak forest (Quercus-petrea-cerris). 84.5% of the woodland consists of Sessile oak (Quercus petrea). About 63% of the mass of the foliage of the forest can be found at the height between 14 and 21 meters. This layer absorbs a considerable amount of precipitation and incoming solar (global) radiation. In the forest there are two shrub levels that are not very thick, at heights of 0.5–1.0 and 1.5–2.0 meters.

The 25 meters high meteorological tower was mounted on the study area, which has an area of 1 hectare, while on the treeless area (grassland) a weather station was mounted (*Fig. 1*).

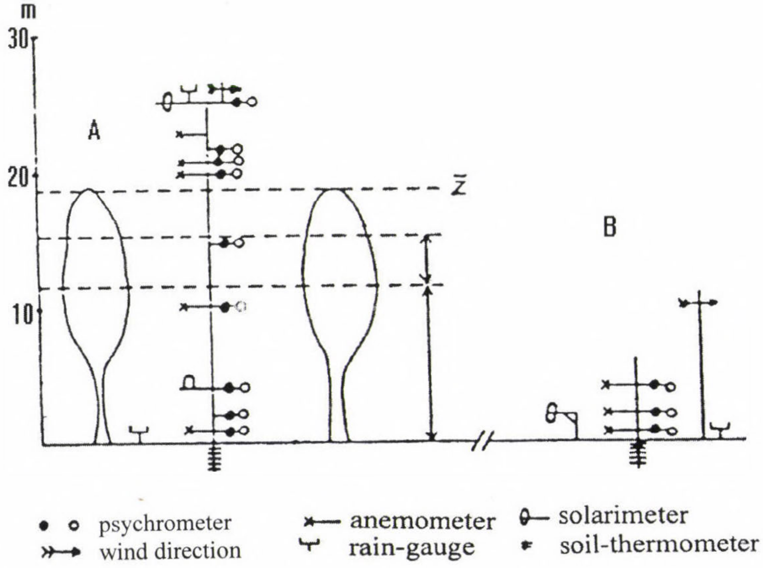

Fig. 1. The mounting of the meteorological instruments, and the levels of measurements inside the forest (A) and above the treeless grassland (B) at Síkfőkút.

Inside the forest, a digital data logger with 80 channels was installed in a wooden cabinet. It carried out meteorological measurements (radiation, temperature humidity, wind velocity, etc.) along a vertical profile inside the forest and above the treeless area. Data had been recorded on telex tape. Later digital data logging had been used (*Justyák*, 1995).

Forest decay appeared in about 1980 in Hungary. In the Síkfőkút forest 50–60% of the Sessile oak decayed until 1995, and a part of the trees had fallen causing detectable changes in the climate of the forest. In the followings we deal with the impacts of the forest on the radiation and temperature elements.

2. Solar radiation

Data on the short wave radiation balance of the Síkfőkút forest can be found in *Table 1*. It can clearly be seen that in 1978, when the forest was healthy, in leafless state (April), the lower layer (2 meters) of the foliage let the 48% of the global radiation through the forest, while the branches and trunks of the trees absorbed 52%. Many years later, in April 1994, 54% of global radiation were let through and 46% were absorbed due to the decay and fall of the trees. Therefore, the healthy forest lets through less and absorbs more radiation. In the decaying forest the situation is reversed. In both cases enough energy reaches the surface under the forest to form its *own independent energy balance* (the air of the forest warms up mostly from beneath). Since the near-surface level of the forest is abundant in light and heat, the growth of the plants in that level is accelerated.

Table 1. Short wave radiation components in the Síkfőkút forest (1978–1994), MJ m^{-2}

Years	**Forest leafless (April)**					**Forest leafy (July)**				
	G	**r**	**G–r**	**A**	**T**	**G**	**r**	**G–r**	**A**	**T**
1978	369	48	321	154	167	548	86	462	425	37
1980	367	47	320	160	160	536	83	453	399	54
1982	386	49	337	172	165	555	85	490	421	69
1984	397	50	347	180	167	570	89	4487	404	83
1986	399	49	350	186	164	581	81	500	4056	95
1988	395	49	346	183	163	609	86	523	429	94
1990	400	50	351	190	161	556	78	478	382	96
1992	411	50	361	195	166	625	86	540	416	124
1994	378	47	331	179	153	628	85	453	424	119
Average	**389**	**49**	**340**	**177**	**163**	**579**	**84**	**487**	**412**	**82**

Legend: *G*= global radiation, which reaches the surface of the forest, *r*=reflex radiation from the surface of the forest, *G–r*=short wave radiation balance of the forest, *A*=the amount of radiation absorbed by the forest, *T*=the amount of radiation let through by the forest, measured at a height of 2 meters in the forest.

In leafy state (July), the forest let through 8% and absorbed 92% of the global radiation in 1978. In the decayed forest in 1994, 22% were let through and 78% were absorbed. These figures mean that energy transport processes in the lower level of the healthy forest are restricted and *an independent energy balance can not form there* (*Justyák,* 1987). On the other hand, in the near-surface level of the decayed forest an *independent energy balance can develop* due to the additional incoming radiation. For this reason, the decayed forest in cloudless dry and hot days gets heat not just from the sides and above but from the warming surface beneath.

3. Temperature

In the trunk space (at a height of 2 meters) of the Síkfőkút forest, usually lower temperatures occur than above the neighboring treeless area (*Table 2*). The annual mean temperature of the forest is 10.0°C, while that of the treeless area is 10.3°C. In the coolest year (1980), the annual mean temperature in the forest canopy was 8.4°C, while above the treeless area it was 8.8°C. In the warmest year (1994), the annual mean temperature in the forest was 11.3°C, while above the treeless area it was 11.5°C. The difference between the coolest and warmest year in the forest was 2.9°C, while above the grassland area it was 2.9°C. The warming up was only a little bit stronger in the forest than its environment.

Table 2. Annual and seasonal mean air temperatures (°C) inside the forest and above the treeless grassland at a height of 2 meters (1978–1994)

Temperature	**Forest**				
	Year	**Winter**	**Spring**	**Summer**	**Autumn**
Maximum	11.3	1.8	12.7	22.9	11.8
Minimum	8.4	-4.1	8.4	17.3	8.0
Average	**10.0**	**-0.5**	**10.6**	**19.6**	**10.0**
			Treeless area		
Maximum	11.5	1.8	12.5	23.6	12.4
Minimum	8.4	-4.1	8.2	18.2	8.4
Average	10.3	**-0.4**	**10.7**	**20.3**	**10.4**

In *Fig. 2* the changes of the annual mean temperatures, their trend lines, the equations of the trend lines, and the R^2 values are presented during the studied period inside the forest canopy and above the treeless area. On the base of the figure, it is visible that during the 17 years of the studied period, the annual mean temperature increased faster in the forest than above the treeless area: the ratio of steepness of the lines is 1:13. The ratio of the temperature increase in a 10 years period is 1.0°C. The increase of the mean temperatures in the forest stand can be explained by two reasons. One of them is that the forest stand became rare due to the decay of the trees. The other reason is that there had been some years warmer than the average during the studied period. As it is visible, at the beginning of the period the difference between the two trends is 0.4°C, while at the end it is only 0.2°C.

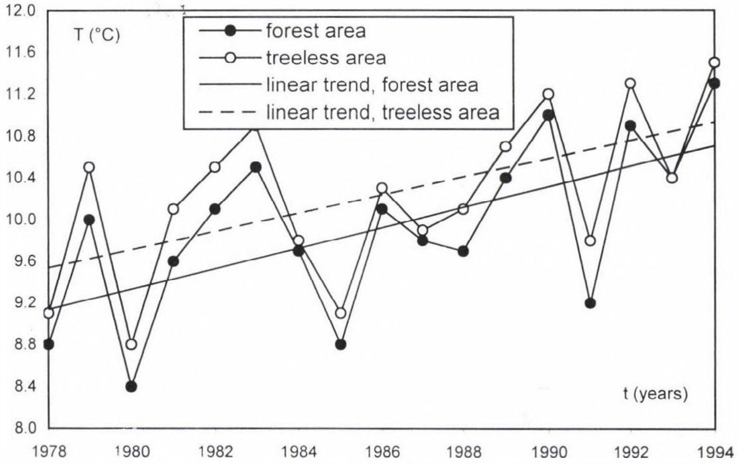

Fig. 2. Fluctuations and trends in the annual mean temperature in the oak forest.

In the forest the mean temperature in the summer period (Jun–Aug) is 19.6°C, above the treeless area it is 20.3°C. In the coolest summer (1978) it was 17.3°C in the forest and 18.2°C above the treeless area. In the warmest summer (1994) the mean temperature was 22.9°C and 23.6°C above the treeless area. The difference between the warmest and coolest summer was 5.6°C, while in the case of the treeless area it was 5.4°C. In this case there was a stronger warming up in the forest again.

As it can be seen in *Fig. 3*, the increase of the mean temperatures is the most rapid and significant in summer. The 10 years average increase of the mean temperatures in the treeless area is 1.8°C, while in the forest it is 2.2°C and R^2 reaches its highest values in this case. The ratio of steepness (1.19) is maximum as well. The ratio of the increase is more than double of the increase the annual mean temperatures: the ratios of the corresponding lines are 2.18 in the forest and 2.08 above the treeless area (*Tar*, 1995; *Antal et al.,* 1997).

In the studied period (1978–1994) it is characteristic for the decayed Síkfőkút forest, that its heat balance system had been altered, and its microclimate had become similar to the microclimatic system of the treeless area (*Justyák*, 1995; *Justyák* and *Víg*, 1997).

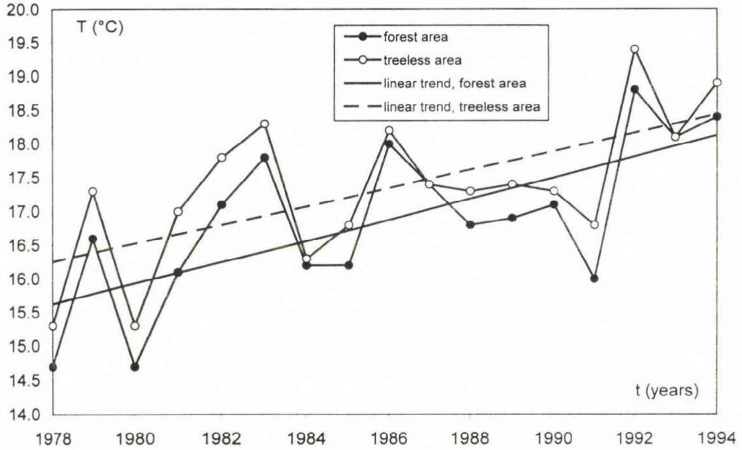

Fig. 3. The change and trend in the summer mean temperatures in the oak forest.

In both the healthy and decayed forest in leafless state, the warmest (active) surfaces can be found in the level of the leafless crowns and on the soil surface of the forest. In the leafless forest more solar energy reaches the soil surface warming it up, and the surface emits heat into the surface layer of the atmosphere (see part A in *Fig. 4*).

In both the healthy and decayed forest, the air temperature is lower under the foliage of the forest than over it, because the foliage absorbs and retains considerable amounts of solar radiation and heat. The active surface, which transfers the heat, is the foliage. In leafy state, only one active surface forms. The 19.5 meters high level of the foliage takes the place of the soil surface in warming the lower layer of the atmosphere (see part B in *Fig. 4*).

Heat is emitted into both layers: the air over the forest and inside the forest from that level of the foliage. For this reason, daytime and in the summer the air is cooler under the active surface of the foliage than inside the upper level of the foliage.

Nevertheless, in the decayed forest, where 50% of the trees had perished, especially in early afternoon in the drought-stricken '90's, when the weather was clear and calm, sometimes another active warm layer formed in the lower air layer of the forest. In such cases two active surface developed: *one* in the upper level of the foliage and *another* in the level of the dry fallen leaves and soil surface. In such weather conditions, a heat surplus of 1.0–2.0°C formed at a height of 2 meters compared to the treeless grassland.

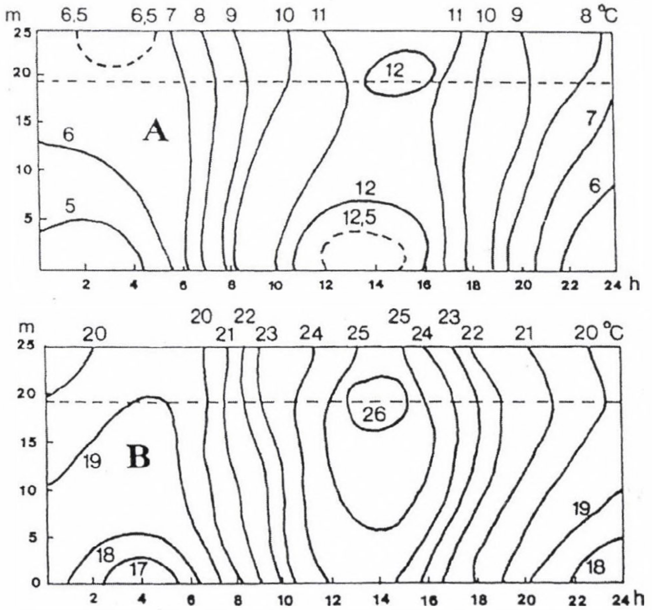

Fig. 4. The diurnal changes of air temperature profile in the leafless (A) and leafy (B) oak forest.

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IDŐJÁRÁS

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Relationship between the stability of wind directions and the mean wind velocity under various weather conditions

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Abstract—In this study we summarize the initial results of our examinations on the wind climate of the growing season. We have selected those weather conditions, when wind directions are stabile and wind velocity is low. These are preconditions for effective irrigation or spraying. These conditions – theoretically – cannot be satisfied at the same time, for this reason we have selected those ones, where the stochastic relationship is weak – it does not differ significantly from 0 – between the stability of the wind directions and mean wind velocity. The database consists of the 5-year (1991–95) hourly wind direction and velocity records of three meteorological stations: Debrecen, Békéscsaba, and Szeged. Analysis had been carried out for each weather station for the whole data set, for the growing season (from April 1 to September 30), and for certain subsets of them. These subsets had been established on the base of macrosynoptic type groups and the so-called central types described by Péczely. We can state that the most favorable for spraying and irrigation is macro synoptic type A and the MS type group. The time of irrigation and spraying can be planned a few days in advance, on the base of the results of the studies on the lifetime of the types, time pattern of their fluctuations, and fluctuations of their transitions into each other.

Key-words: stability of wind direction, index of stability, growing season, irrigation, spraying, macrosinoptic type by Péczely.

1. Introduction

"When dealing with the microclimates of the plant stands, it is not advisable to separate the characteristics of the inner microclimatic spaces of the plant stands from the features of the microclimatic spaces over the plant stands, especially from the wind velocities there. Wind velocity and radiation are the governing factors of the fluxes, which develop in microclimatic spaces. It means that the before mentioned elements control the formation of fluxes." (*Szász,* 1995). Therefore, the wind conditions inside the plant stand, which have an effect on the thermal conditions of the plant stand for instance, are not independent from the characteristics of the wind field over the plant canopies. On the base of this idea, we started the detailed analysis of the wind conditions during the growing season in Hungary. In a previous paper (*Tar* and *Szegedi,* 2002*)* we dealt with the initial results on meteorological conditions of the sprinkling irrigation and sprinkling, which are in direct connection with the developing of the plants. In the present paper we focus on those questions that were not answered there, for instance, the clustering of the data (the establishment of subsets).

2. The aim of our research

For effective sprinkling irrigation or sprinkling, those weather conditions are suitable, when wind directions are stabile (less variable) and wind velocity is low. These two preconditions are contradictory, since presumably, those wind directions are stabile that have high velocities. How can we select those macrosynoptic situations, when conditions are suitable for the irrigation or sprinkling? The aim of our research is to select those weather situations, where the stochastic relationship between the measure of the stability of the wind directions and mean wind velocity does not differ significantly from zero or very weak.

3. Database

The database consists of the 5-year (1991–95) hourly wind direction and velocity records of three meteorological stations: Debrecen, Békéscsaba, and Szeged, which was made available for us by the Hungarian Meteorological Service. The dataset – according to the climatologic standard – consists of 16 wind directions. When preparing the data for the analysis, it was taken into consideration, that not each value was measured at the standard 10 meters height. In those cases data were transformed to the 10 meters height using the equations proposed for meteorological wind measurements (Mezősi and Simon, 1981).

Analysis had been carried out for each weather station for the whole data set, during the growing season (from April 1 to September 30), and for certain subsets of whole data set and growing season. These subsets had been established on the base of the macro synoptic type groups described by Péczely. In opposition to *Péczely* (1983), the central types are handled separately on the base of its obviously different air flow characteristics. According to this, the following categories are used: meridional northern (MN), meridional southern (MS), zonal western (SW), zonal eastern (ZE) type groups, and anticyclone center (A) and cyclone center (C) types. Their characteristics are summarized in *Table 1*.

Table 1. Codes, letter codes, and short descriptions of the Péczely's macrosynoptic types

No.	Code	Description
		Types connected with northerly current (type group MN)
1	mCc	Hungary lies in the rear of an Eastern European cyclone
2	AB	anticyclone over the British Isles
3	CMc	Hungary lies in the rear of a Mediterranean cyclone
		Types connected with southerly current (type group MS)
4	mCw	Hungary lies in the fore part of a Western European cyclone
5	Ae	anticyclone in the east from Hungary
6	CMw	Hungary lies in the fore part of a Mediterranean cyclone
		Types connected with westerly current (type group ZW)
7	zC	zonal cyclonic
8	Aw	anticyclone extending from the west
9	As	anticyclone in the south from Hungary
		Types connected with easterly current (type group ZE)
10	An	anticyclone in the north from Hungary
11	AF	anticyclone over the Fennoscandinavian region
		Types of pressure centers
12	A	anticyclone over the Carpathian Basin
13	C	cyclone over the Carpathian Basin

As it can be seen, the base of the classification is the place of the cyclone or anticyclone centers as compared to Hungary. The types, when a cyclone governs the weather of Hungary, can be grouped in the cyclonal type group (CG). The types of the CG are: 1, 3, 4, 6, 7, 13. Anticyclone type group (AG) can be defined similarly; its elements are: type 2, 5, 8, 9, 10, 11, and 12. For the classification of the individual days, we used the macrosynoptic codes of *Károssy* (1998). The size (the number of days) of the macrosynoptic types/type groups for the whole five years long period and for five growing season together is shown in *Table 2*. It can be seen that in the frequency of the occurrence of the macrosynoptic types/type groups, the greatest differences (4–5%) are in the meridional group: the number of the days in the MN group has increased, while in the MS group it has decreased relative, to the whole period. It is remarkably that in the relative frequency of the cyclonal and anticyclonal type groups there are practically not differences between the whole period and growing season.

Table 2. The number of days of the various type groups (days), and their length relatively to the whole length of the period (%) in the whole 5-year period (1991–95), and in the growing season

Type groups	**1991–1995**		**Five growing seasons**	
	days	**%**	**days**	**%**
MN	401	22.0	244	26.7
MS	434	23.8	161	17.6
ZW	390	21.4	182	19.9
ZE	235	12.9	142	15.5
A	269	14.7	120	13.1
C	97	5.2	66	7.2
SUM	**1826**	**100.0**	**915**	**100.0**
AG	1259	68.9	627	68.5
CG	567	31.1	288	31.5
SUM	**1826**	**100.0**	**915**	**100.0**

4. Methods

On the diurnal changes of the wind directions in Hungary, *Hegyfoky* (1904a-d) had published data first. On the base of the records of ten weather stations, he established that near the surface layer of the atmosphere there is a positive change in the wind directions in the morning, and a negative change in the afternoon, if we determine positive direction as a clockwise turn facing the wind. He assumed that the reason for this phenomenon is the mechanical effect of the air pressure system drifting to the east on one hand, and the diurnal changes of the temperature conditions on the other hand. Based on this, in a previous paper we studied the change of the wind direction as a random variable (*Tar*, 1980) using the five years long dataset (1968–72) of five weather stations situated on the Great Hungarian Plane (Kisvárda, Debrecen, Kecskemét, Békéscsaba, and Szeged), which are affected by orography to a lower degree. The method of the examinations, which we follow in the present paper, too, is that we create elemental events from the hourly wind directions. The wind direction of the t hour of a day of a given period (subset) is compared to that of the previous (t–1) hour. If these are uniform (that is the difference between them expressed in degrees is lower than 360/16=22.5), then the t hour is considered to be a *stable (permanent) point of time* from the aspect of the change of the wind directions, and it is marked by S_t. Otherwise, the t hour is an unstable (fluctuating) point of time, and the event is I_t. The wind direction of the first hour of the day is compared to the 24th hour of the previous day, even though it belongs to another subset. If in each hour of a given time interval the wind direction is stable or unstable, it is considered to be a stable or instable period.

Therefore, in our case the maximal daily number of stable or instable hours can be 24. This should be considered as a lower limit from any aspect. The number of stable or instable hours increases if the number of observations and/or wind direction categories increases. Both are true, for instance, for the case of automatic weather stations. Our results would obviously change if we used such datasets, that would have not only climatological consequences but effects on the utilization of wind energy as well (*Tar,* 2003).

5. The measures of the stability of the wind directions

First we define the measure of the stability of the wind directions, i.e., the *stability index*. It is the ratio of hours with stabile wind direction in a given period to the whole length (in hours) of that period. The values of the stability index for the whole studied period growing season, as well as different type groups are presented in *Table 3*. Generally it can be stated, that on the Great Hungarian Plane wind directions show a stronger trend to be stable than to fluctuate, since, with the exception of three cases (the ZE type group, types A and C in Békéscsaba during the growing season), they take 0.5 or higher values. It can also be seen in the table, that in Békéscsaba the stability indexes are lower than those in the same subsets at the two other stations for the years and growing season, with the exception of the ZW type group. That is because the bent for fluctuation is the strongest there. The reason for this is presumably orografic, because that station is the nearest to the Carpathian Mountains, therefore, the development of thermal circulations is the most possible there. This fact and that the sequence (from the lowest to the highest) of the stability indexes for almost all cases is: Békéscsaba-Debrecen-Szeged, are quite in line with the observations of Péczely (*Péczely,* 1963) on mountain and valley winds.

From the table it is also visible, that stability indexes for all subsets and stations are lower during the growing season than in the whole year. The strong fluctuation of the wind directions during the growing season presumably has thermal reasons, because the mean temperatures during the growing season are much higher than in the whole year. The five-day mean temperatures are between 10 and 20°C during the growing season and under 10°C in the other part of the year (*Szász,* 1988).

The minimum of the stability index in the whole year occurs – understandably – at all the three stations in the C type. During the growing season the picture is not so simple: in Békéscsaba and Debrecen this value in type A is near the minimum. The logic of the maxima generally speaking is the following: in a north-south direction through the zonal (ZW, ZE) type groups it arrives to the meridional southern type group (MS). This logic is disturbed by the values of the whole year in the A type in Szeged and the values of the growing season in the MN type in Debrecen. The absolute maximum occurs in both periods in the MS type group in Szeged, but these values are very close to those of type A.

Table 3. Stability index (st.i), mean wind velocity (v_{mean}), and mean daily number of stabile hours (st.h) in various periods

Stations	**Charac-teristics**	**Whole**	**MN**	**MS**	**ZW**	**ZE**	**A**	**C**	**AG**	**CG**
					Year					
	st.i.	0.56	0.55	0.57	0.56	**0.59**	0.56	*0.53*	0.57	0.54
Debrecen	v_{mean} (m/s)	2.7	3.1	2.4	2.9	2.8	*2.0*	**3.3**	2.5	3.2
	st.h.	13.5	13.1	13.6	13.5	**14.1**	13.4	*12.6*	13.7	13.1
	st.i.	0.54	0.54	0.55	**0.56**	0.53	0.52	*0.51*	0.54	0.55
Békéscsaba	v_{mean} (m/s)	2.8	3.2	2.6	3.1	2.5	*2.0*	**3.5**	2.7	3.4
	st.h.	13.0	13.0	13.2	**13.4**	12.7	12.5	*12.3*	12.9	13.2
	st.i.	0.58	0.55	**0.61**	0.58	0.57	**0.61**	*0.54*	0.59	0.56
Szeged	v_{mean} (m/s)	3.1	3.5	3.3	3.2	2.8	*2.2*	**3.9**	2.9	3.7
	st.h.	14.0	13.3	**14.6**	13.9	13.8	14.5	*12.9*	14.2	13.4
					Growing season					
	st.i.	0.52	**0.54**	0.52	0.51	**0.54**	*0.50*	0.51	0.53	0.51
Debrecen	v_{mean} (m/s)	2.6	3.0	2.4	2.6	2.6	*2.0*	**3.2**	2.5	3.0
	st.h.	12.5	**12.9**	12.5	12.3	**12.9**	*12.0*	12.2	12.6	12.3
	st.i.	0.50	0.50	0.51	**0.52**	*0.48*	*0.48*	*0.48*	0.50	0.50
Békéscsaba	v_{mean} (m/s)	2.9	3.1	2.9	2.9	2.7	*2.5*	**3.3**	2.7	3.2
	st.h.	12.0	12.1	12.1	**12.5**	*11.5*	*11.6*	*11.5*	12.0	12.0
	st.i.	0.54	0.52	**0.56**	0.55	0.54	0.55	*0.51*	0.55	0.52
Szeged	v_{mean} (m/s)	3.0	3.3	3.2	3.0	2.7	*2.2*	**3.6**	2.8	3.4
	st.h.	12.9	12.5	**13.5**	13.2	13.0	13.1	*12.2*	13.1	12.5

bold: the maximum value, *italic:* the minimum value in the macrosynoptic type groups and types

On the base of the data in the table, the strength and trend of the stochastic relationships between the stability indexes and average wind velocities for the individual subsets (MN, MS, ZW, ZE, C, A) can be determined. It can be stated, that the linear correlation coefficient does not differ from 0 significantly in any cases, since its absolute value does not reach the critical value ($r_{0.05}=0.81$) of the probability level of 0.05. The closest to the critical value occurs in Szeged in the case of the year (-0.706). On the base of the before mentioned facts, we can establish that type C is quite unsuitable from the aspect of the sprinkling irrigation of the plant, because in that type low stability indexes are combined with high wind velocities. On the other hand, there are no types or type groups where the opposite is true, that is, where maximal stability indexes and minimal wind velocities occur. To find the solution, we have to analyze the relationships between the *daily* parameter of stability and *daily* mean wind velocity.

6. The relationship between the stability and daily mean wind velocity

As daily characteristic parameter of the stability, we choose the daily number of hours with stabile wind directions. Its average values are presented in *Table 3*. The relationships between daily stability and mean wind velocities are examined by two methods. In the first case we correlate the *actual* daily number of stabile hours of the periods with the mean wind velocity of the same day. Two examples for this are presented in *Fig. 1* and *Fig. 2*, where the regression line is presented as well (it will be called *case of day-to-day* in the followings). In the second method we gather the days that have the same number of stabile hours, and then we correlate their mean wind velocity with the daily number of stabile hours (*case of classification*). The frequency of those days, that have the same number of stabile hours, their daily mean wind velocities and the regression line applied for them are presented in *Fig. 3*. In *Table 4,* linear correlation coefficients determined by the two methods, and their critical values at the significance level of 0.05 ($r_{0.05}$) are shown in the different cases.

Critical values – the minimal values of the correlation coefficient that significantly differs from 0 –, up to about 100 sample elements, can be most simply determined using the Fisher-Yates table, in some cases by interpolation (*Dobosi* and *Felméry*, 1971). Dealing with more than 100 elements – that occurs using the first method – the significance analysis must be carried out using the *t*-test that is $r_{0.05}$ was calculated from the value of $t_{0.05}$. Therefore, the test function is

$$t = \frac{r}{\sqrt{1-r^2}}\sqrt{n-2}\ , \qquad (1)$$

where n is the number of the elements (*Yule-Kendall*, 1964). The critical t values ($t_{0.05}$) of the 0.05 probability level over 400 elements are 1.96–1.97 – they practically do not change –, but because of the different number of the elements, the $r_{0.05}$ changes. The critical values of the correlation coefficient between the daily number of the stabile hours and mean wind velocity of the same day (case of day-to-day) are the same in the three stations, because the number of the elements is the same. It is not true for the clustering of the days that show the same stability (case of classification), since the number of stabile hours – that theoretically can take values between 0 and 24 – can be different in each case (subset) and each station. The occurrence of low (0, 1, 2) and high (22, 23, 24) values can even be 0.

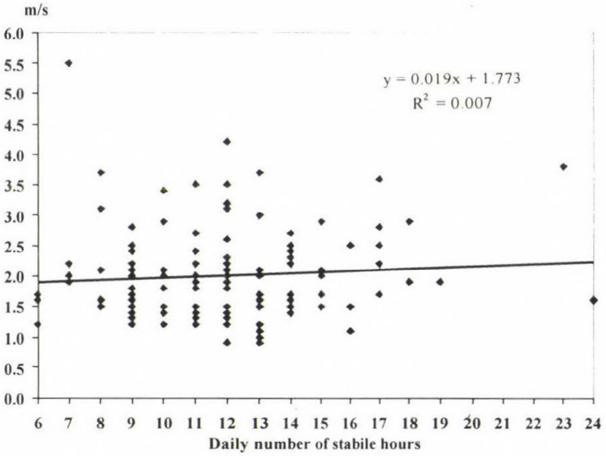

Fig. 1. Linear regression and correlation between the daily mean wind velocity (m/s) and daily number of stabile hours in the case of day-to-day (growing season, the case of minimum correlation: Debrecen, type A).

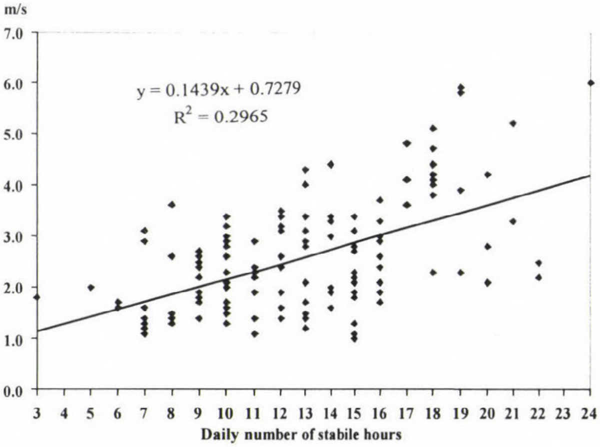

Fig. 2. Linear regression and correlation between the daily mean wind velocity (m/s) and daily number of stabile hours in the case of day-to-day (growing season, the case of maximum correlation: Debrecen, type group ZE).

On the base of *Table 4* it can be stated, that in the *case of day-to-day,* values of the correlation coefficient between the daily number of the stabile hours and daily mean wind velocities in the six subsets (MN, MS, ZW, ZS, C, and A) of both the whole year and growing season are the lowest in type A at all the three stations, and they are not significant in Debrecen (see *Fig. 1*) and Szeged. The "second lowest" but significant values at all the three stations occur in type group MS. In the *case of classification* there are slight differences between the whole year and growing season: for the whole year there are not significant correlation coefficients in type A, in Debrecen and Szeged (see *Fig. 3*), in Békéscsaba the lowest value can be found in type group ZE. The next correlation coefficient by the value is in type group ZW in Debrecen, it falls into type group MS in Békéscsaba and Szeged, but it is not significant in Szeged. The picture is more clear during the growing season: it is similar to the daily event with the exception of that in Szeged in the case of type group MS and type C the values are not significant.

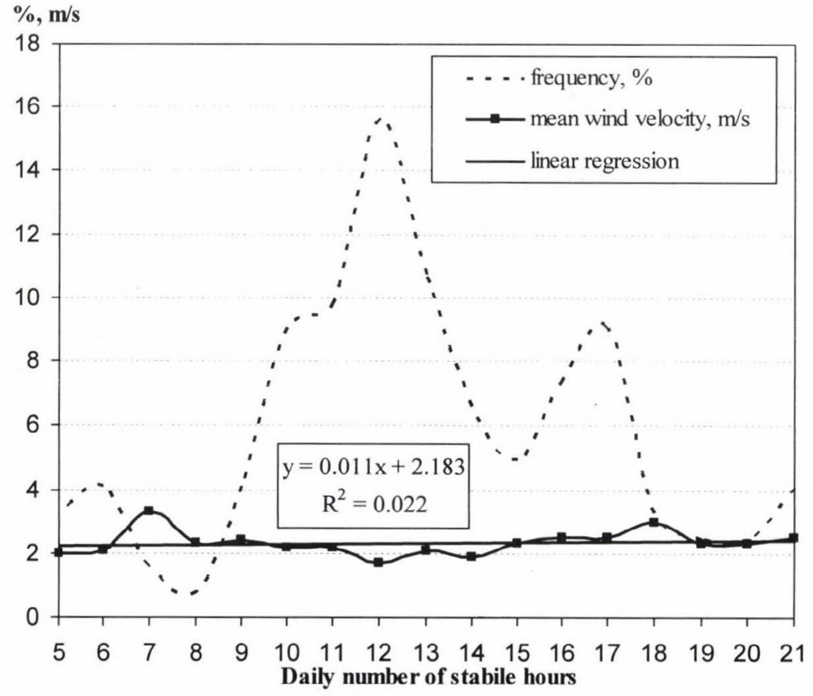

Fig. 3. Frequency of days having the same number of stabile hours (%), as well as the linear regression and correlation between the mean wind velocity (m/s) and daily number of stabile hours in the case of classification of the days (Szeged, type A, growing season).

Table 4. Linear correlation coefficients and their critical values at the 0.05 significance level in the case of day-to-day and classification of the days, in various periods

Method of the data groups	**Whole**	**MN**	**MS**	**ZW**	**ZE**	**A**	**C**	**AG**	**CG**
Year									
Debrecen									
Day-to-day	0.29	0.42	0.27	0.27	0.46	**0.11**	0.41	0.30	0.36
Classification of the days	0.91	0.94	0.88	0.80	0.87	**0.43**	0.82	0.87	0.95
Békéscsaba									
Day-to-day	0.44	0.56	0.39	0.48	0.44	0.34	0.53	0.32	0.33
Classification of the days	0.88	0.91	0.88	0.96	0.80	0.91	0.90	0.92	0.77
Szeged									
Day-to-day	0.21	0.39	0.20	0.21	0.32	**0.03**	0.38	0.19	0.38
Classification of the days	0.84	0.79	**0.27**	0.62	0.60	**0.20**	0.60	0.78	0.90
Critical values – day-to-day	0.05	0.10	0.09	0.10	0.13	0.12	0.20	0.06	0.08
Critical values – classification of the days									
Debrecen	0.40	0.42	0.42	0.44	0.42	0.45	0.50	0.42	0.42
Békéscsaba	0.41	0.41	0.44	0.44	0.42	0.42	0.47	0.41	0.41
Szeged	0.43	0.43	0.43	0.44	0.43	0.45	0.43	0.43	0.41
Growing season									
Debrecen									
Day-to-day	0.37	0.440	0.26	0.37	0.54	**0.08**	0.40	0.41	0.35
Classification of the days	0.89	0.92	0.67	0.86	0.75	**0.07**	0.79	0.78	0.92
Békéscsaba									
Day-to-day	0.34	0.41	0.26	0.35	0.33	0.26	0.44	0.35	0.36
Classification of the days	0.80	0.85	0.48	0.78	0.90	0.77	0.58	0.80	0.62
Szeged									
Day-to-day	0.25	0.32	0.20	0.29	0.36	**0.17**	0.33	0.27	0.29
Classification of the days	0.80	0.820	**0.32**	0.73	0.67	**0.15**	**0.19**	0.59	0.62
Critical values – day-to-day	0.06	0.130	0.16	0.14	0.16	0.18	0.24	0.08	0.11
Critical values – classification of the days									
Debrecen	0.42	0.47	0.43	0.52	0.47	0.53	0.50	0.42	0.43
Békéscsaba	0.43	0.47	0.44	0.50	0.47	0.50	0.50	0.44	0.44
Szeged	0.41	0.48	0.48	0.47	0.48	0.48	0.50	0.43	0.41

bold: no significant coefficients

For both periods examined here it can be stated, that the stronger correlations can be found for the case of day-to-day in Békéscsaba in the whole year, and at the two other stations during the growing season. Higher correlations can be found at all the three stations during the growing season only in the antyicyclonal type group (AG). In the AG and CG groups and A and C types, considering the whole period in the case of classification, the situation is the opposite: the correlations for the whole year are stronger. Consequently, our assumptions discussed in Chapter 2 are not true for type C, because in that type two fronts appear together.

7. Conclusions

It can be stated, that on the Great Hungarian Plane the conditions for the sprinkling and sprinkling irrigation are the most favorable in the A type and MS type group (its elements can be found in *Table 2*). In these cases there is no correlation, or only very weak linear correlation can be found between the parameter characteristic for the stability of the wind directions, daily number of stabile hours, and daily mean wind velocity. On the base of the studies on the life span of the macrosynoptic types (*Péczely,* 1983), time patterns of their fluctuations (*Tar* and *Kircsi*, 2002), and frequencies of their transformations into each other (*Mika* and *Domonkos*, 1994), the time of the sprinkling or sprinkling irrigation can be planned in a few days in advance.

We have proved that the increase of the size of the subsets (groups) decreases the amount of information useful for solving the problem (see, e.g., type groups AG and CG). Presumably, we could gain better results in this aspect carrying out the above detailed analysis for the individual macrosynoptic types. This analysis, however, requires more abundant database.

The relationship between the wind conditions and water supply of plant canopies can be examined from a completely different point of view. Wind machines, which use wind energy for water pumping, are suitable for dripping irrigation. It is proven that the specific wind power is in a – not obviously linear – stochastic relationship with the stability of the wind directions.

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An evapotranspiration calculation method based on remotely sensed surface temperature for agricultural regions in Hungary

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Abstract—The surface temperature measured by satellite can be the basis of evapotranspiration (ET) computation. The possibility of the daily sum of the regional ET using surface temperature was examined under Hungarian weather conditions. Simplified relationship, namely $R_{nd} - ET_d = A + B(T_C - T_A)$, which relates to the daily ET and daily net radiation with one measurements of surface and air temperature was used for the calculation. No information was obtained about the surface inequality using NOAA AVHRR satellite data. To collect any information describing the distribution of surface temperature, infrared thermometer was used to scan the surface from the board of a hang-glider, ultra-light-aeroplane, and light aeroplane. The limited field observations were made during the vegetation period of 1992, 1993, and 1994. In eastern part of the country, a homogeneous field (1 km × 1 km) and a larger, and relatively homogenous area was scanned, before noon and afternoon. In the western part of the country, a much larger area (45 km × 45 km) was investigated. The distribution of the surface temperature in special cases is shown. The paper presents a calculation system which can produce the daily sum of evapotranspiration using the daily maximum surface and standard air temperature. The results of a simplified validation method proves the usefulness of the shown method. The limitation of the use and possible development of the method concludes the paper.

Key-words: evapotranspiration, latent heat flux, near-surface measurements, satellite born information, surface temperature.

1. Introduction

One of the basic task of agricultural meteorology is the determination of water supply of plants using simply measurable meteorological elements. The ratio of transpired water and potential evaporation shows the plant water supply, the measure of plant water stress. Determining the water stress status of the plant, agrometeorology can give useful information to the users. Nowadays, three great changes have happened in meteorology. The first two are connected with the instrument and measurement technique, the third one is related to the data transfer and processing. The operative application of remote sensing gives new, previously impossible methodology to the agrometeorology. The satellite born information promotes the regional investigation as much as it was not possible earlier. The computer technique and data transfer allow so quick data process which was unthinkable 10–15 years ago.

The application of remote sensing in agrometeorology started when the first instruments were issued to measure the surface temperature. The equation using the surface temperature (*Jackson et al.,* 1977, 1981, 1988), the developed methods are valid for more or less small field experiment places. The question we have to answer is how to use the results of field experiments for larger scale processes (*Caselles et al.,* 1992, *Hurtado et al.*, 1994). Two demands triggered the present work. The first one was the investigation of possible use of directly measured surface temperature in agricultural meteorology (*Dunkel et al.*, 1989). The second one (*Kerényi* 1993, 1994) was the demand of ground truth for calibration of satellite born surface temperature measurements. Hungarian experiments (*Dunkel* and *Vadász*, 1993) were carried out to determine the water stress status of plants using the surface temperature.

The investigations were continued not in this direction. Previously the application of satellite born surface temperature for calculation of evapotranspiration was investigated. More or less the same approach is how to exploit the recent meteorological development in agricultural meteorology. The present study gives the methodology of the calculation of areal evapotranspiration. Neither the important calibration problems of the measurement of surface temperature close to the surface nor the evaluation of evaporation calculation methods are shown. To choose surface control method we had to evaluate few classic calculation methods, but only the result of the comparison with *Antal* (1968) method is presented. Combining the different systems, a method was developed which is suitable for calculation of daily areal evapotranspiration.

The surface temperature measured by remote sensing can be the basis of territorial investigation of water stress status of any kind of canopy. ET can be calculated following *Seguin* and *Itier* (1983) or *Lagouarde* (1991) by solving the surface energy equation in a simplified way, or *Jackson et al.* (1981, 1988) calculating directly the water stress status of the canopy. In the present paper water stress map is not shown. An example for calculation of territorial ET for Hungary is presented. There is no information about the changes within a pixel. There is no information about the emissivity of the investigated surface.

Knowledge about the water vapor content and its influence on the measured surface temperature is poor. Surface reference values to calibrate the ET calculation equation are inadequate. To answer at least partly the problem, a surface controlled measurement was made in Hungary.

2. Description of the field experiments

In order to get a better spatial resolution for estimating the temperature distribution, a near-surface trial was executed during the vegetation period of 1992, 1993, and 1994. KT-24 type infrared thermometer was used every year. The instrument was put on board of a hang-glider in 1992, a light (CESSNA 105), and ultra-light (TUCANO) aeroplane in 1993. Later, in 1994 and 1995, the measurement program was extended to measuring, besides the surface temperature short wave reflected, and the long wave irradiated radiation, by Kipp-Zonen CM-5 type radiometer, and Eppley pyrgeometer, respectively. In the first two years, the data were collected using one-channel data-logger. In 1994, the instruments were built into the wing of an AN-2 type aeroplane, and the data were loaded directly to a PC. The cruising height was 100 m in case of the hang-glider, and 500 m in case of the aeroplane. To understand the effect of water vapor on surface temperature, the runway was scanned from 600, 400, 200, and 20 m heights.

As reference to ET for surface control measurements in 1992 and 1993, only the class A-pan evaporation observation was used. The ET reference value, using empirical formula derived for Hungarian climate was calculated. In 1994, it was possible to measure the latent heat flux directly using Bowen ratio system, however, for only one place. In 1992, the basic goal was to investigate the temperature change within one field. Two fields were chosen as control area. Their size was about 1 km × 1 km. In the eastern part of Hungary sometimes such a large field with approximately homogeneous vegetation was found. In the present investigation the surface temperature scanning over irrigated and non-irrigated maize canopy was executed. The scanning routes had north-south direction in 1993 and 1994. In 1992, a relatively small area was scanned in the eastern part of Hungary close to the town Szarvas, where an agrometeorological observatory was maintained by the Hungarian Meteorological Service (HMS). The scanned surfaces were maize and wheat canopies. We tried to measure a more or less homogeneous surface as large as it could be identified in the satellite picture. In 1993, the philosophy was changed. We tried scan as large area as it was possible. The length of one scanning route was 45 km. The experiment area was in the western part of the country. Eleven scanning routes completed the examination area. Few scanning route crossed the lake Balaton. In 1994, the place of the field experiments was the Hortobágy puszta in the eastern part of the country close to Debrecen, the second biggest city of Hungary. The surface of the "puszta" is more less natural grass, it looks like a steppe. The assumption of homogeneous surface worked more or less in this case as well. The length of scanning route was shorter in 1994, 25 km.

To calibrate the infrared thermometer, a "black body" model was used. This "black body" was a sooty copper plate. The instrument and the "black body" were placed together into a climate chamber. The near-surface temperature measurements were not all simultaneous. Because of the achieved cruising speed, the time-leg between the first and last measured value could be few hours. If a comparison of the result with any kind of satellite picture is desired, the measured values need to be homogenized. Following *Sellers et al.* (1986), a simplified model was used for this purpose (*Dunkel et al.*, 1991). The main problem in the use of the suggested model is the estimation of heat capacity of the canopy. The calculated and measured surface temperatures were compared for sunflower canopy under different experimental conditions. Using the developed diurnal temperature change model, the measured surface temperature values were homogenized for the same time. This homogenization is necessary, because airborne measurements are not synchronous.

Comparing the morning and near afternoon temperature distribution along one cruising route during the 1993 field experiments, characteristically different distribution was observed. Few hours after sunrise almost no change in canopy temperature was observed, nevertheless, the swamp can be well identified, and of course the surface of the lake Balaton shows characteristic values.

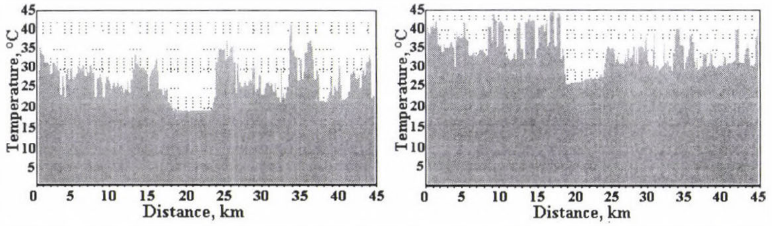

Fig. 1. The distribution of the surface temperature along a Balaton crossing route.
July 9 and August 5, 1993.

The hypothesis before the flight was that no change would be found in the water surface temperature. As *Fig 1.* shows, the surface temperature of the shallow lake is not homogeneous, but comparing with the other surfaces it is much more characteristic. Repeating the flight over that area near noon, opposite situation was observed. The swamp was much cooler than the surrounding area, and the variation in temperature was more than 10 degrees. The temperature distribution shows the same change along the same route during July–August across the lake Balaton. The lake surface temperature shows almost no change. The temperature difference outside the lake could be more than 10 degrees, but the results of the scanning showed that the homogeneous surface conception is not valid even in case of a shallow lake such as Balaton.

3. The model

The calculation method is based on the simplified surface energy balance equation. The energy balance equation of any given surface could be written as

$$R_n = D + G + J + \mu A + H + LE\ , \tag{1}$$

where R_n is the net radiation, D is the horizontal energy transfer, G is the soil heat flux, J is the energy storage in the "surface", μA is the energy used for biochemical processes, H is the sensible heat flux, LE is the latent heat flux. In our terminology the evapotranspiration expressed in W m^{-2} unit, later converted into traditional unit mm. A very simple and useful solution of the equation was suggested by *Jackson et. al* (1977). Later *Seguin* and *Itier* (1983) used only the larger elements of energy budget and neglected the smaller terms, namely μA (*Hunkár,* 1985), the rate of energy stored biochemically, and G, the ground heat flux is more or less zero on a daily basis. After the simplification the equation shows the following form:

$$R_n = H + LE\ . \tag{2}$$

The term H, the sensible heat flux could be expressed in the near-surface layer as,

$$H = -\rho\, c_p \frac{T_C - T_A}{r_A}, \tag{3}$$

where T_C is the surface temperature, later to be determined using satellite information, and T_A is the near-surface temperature practically to be taken equal with the air temperature measured in the standard wind screen above the surface in 2 m height at the synoptic station. The ρ is the density, c_p is the heat capacity of the air at constant pressure, r_A is the air resistance in the lower layer. In the present approximation every value is instantaneous. After the simplification, we have found a linear expression between the temperature difference latent heat flux, and net radiation difference, as

$$LE - R_n = -b^* (T_C - T_A), \tag{4}$$

where all physical constants are united into one parameter, b^*. We would like to calculate the daily evapotranspiration, but only instantaneous values are measured. The field experiments, in case of cloudless day, proved that the presented linear expression is good approximation (*Seguin* and *Itier*, 1983; *Lagourde*, 1991) for the calculation of daily sum of evapotranspiration using only one temperature difference as it is shown in *Fig. 2*. The final expression will be

$$ET_d - R_{nd} = A + B\,(T_C - T_A) \qquad \text{(mm/day)}, \tag{5}$$

changing the *LE* abbreviation to *ET*, as usually accepted in agricultural meteorology for the searched quantity, and signing the daily values with *d*. In the expression *A* and *B*, parameters can be estimated by linear regression, T_C is the surface (canopy, skin) temperature measured by satellite or any kind of remote sensing technique, and T_A is the near-surface (daily maximum) temperature. The determination of T_C, the surface (canopy, skin) temperature measured by satellite, is one of the basic tasks of satellite meteorology (*Becker* and *Li*, 1990; *Kerényi*, 1994).

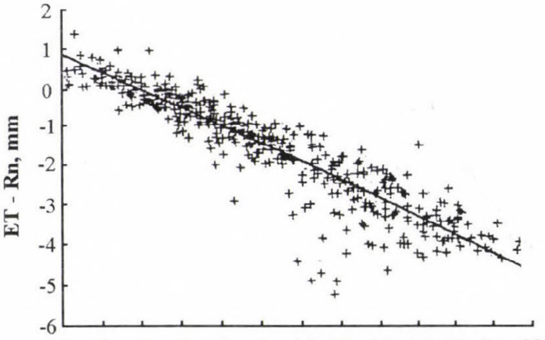

Fig. 2. The relationship between the two differences, ten-day average (*Lagourde*, 1991).

The parameters can be estimated experimentally, simulated, or calculated by a combination of the two methods. The regression coefficients depend on the surface roughness as it is shown in *Table 1*.

Table 1. The value of "A" intersection (mm day^{-1}) and "B" slope (mm day^{-1} $grad^{-1}$) by *Lagourde* (1991), z_0 is the roughness coefficient

z_0	A	B
0.001	0.41	0.145
0.002	0.43	0.166
0.005	0.47	0.204
0.01	0.5	0.244
0.02	0.53	0.296
0.05	0.55	0.394
0.1	0.59	0.498

The simple linear regression between the daily sum of ET and the midday difference of surface temperature minus air temperature is proven by many investigators (*Caselles et al.*, 1992; *Jackson et al.*, 1977, 1981, 1989; *Hurtado et al.*, 1994; *Kustas et al.*, 1982; *Lagourde* 1991; *Sandholt* and *Andersen*, 1993).

4. Calculation method

Satellite Research Laboratory of HMS has been receiving NOAA-AVHRR data digitally, since 1992. The data are calibrated and geographically identified. For the ET calculation, NDVI is derived from Channel 1 and 2, which is the calculation procedure widely used all over the world. The surface temperature was calculated by split-window method from the data of Channel 4 and 5 (*Kerényi* and *Putsay*, 2000). The calculated and measured surface temperature were compared in case of few situations during summer time. For estimation of ET, following *Seguin* and *Itier* (1983) suggestion, besides the midday surface temperature, the maximum air temperature measured by standard meteorological network is also needed. During the 1995/96 period, the traditional meteorological observations were replaced by automatic synoptic stations. If the observation time was very different from the date of maximum air temperature, a shifting correction was applied following the *Sellers et al.* (1986) model.

The basic goal of the research was to establish a method which could be used in operative practice to determine the daily sum of evapotranspiration. The base of the method is a simplified solution of surface energy balance equation as it was introduced earlier in Eq. (5). This equation is valid not only for instantaneous but longer period, namely for daily values, too, as showed by many field experiments. If we apply this equation for the whole day, the daily evaporation minus net radiation is the linear function of maximal air and surface temperature difference. To apply the method in the practice, we need several further informations.

The elements of the calculation system are shown in *Fig 3*. The near-surface air temperature and surface temperature are measured by HMS operatively. The necessary A and B constants are originated partly from field experiments and mainly from the cited literature, because we had possibility to carry out the necessary field experiment only in few cases of the possible land surface covering situations. The air-born surface temperature measurements were carried out above different parts of Hungary, as it was mentioned in the previous chapter. The measurements were done in summer during the time of well developed vegetation. We tried to carry out the measurements above homogeneous surface and use as large experimental area as it could be identified on satellite picture.

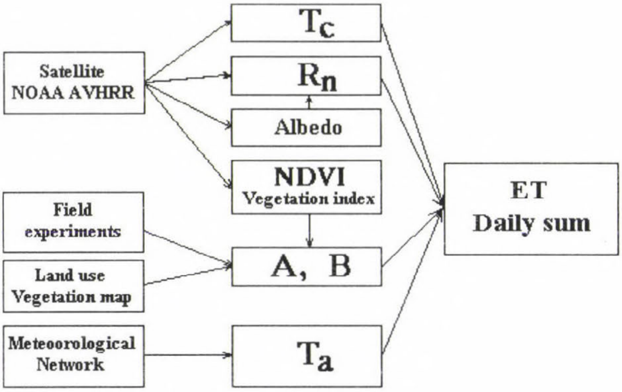

Fig. 3. The flow-chart of daily evapotranspiration calculation.

The satellite born surface temperature could be useful for our purpose if they were done near noon. Sometimes we need the temporal change of surface temperature. Using the surface energy equation, a simplified solution was developed to calculate the surface temperature. The calculation method was controlled using lysimeter data derived from the experimental sites. The evaporation calculation system could be compared with other calculation methods. Few potential evaporation calculation formula were evaluated to decide which one could be used for our evaluation. Taking into consideration the difficulties of many ET calculation methods, the comparison with the *Antal* (1968) potential ET calculation method is shown as verification of the method. For the calculation we used only the meteorological stations equipped with standard A pan. In *Fig. 4*, the stations are shown where Pan A open water evaporation is measured operatively by HMS.

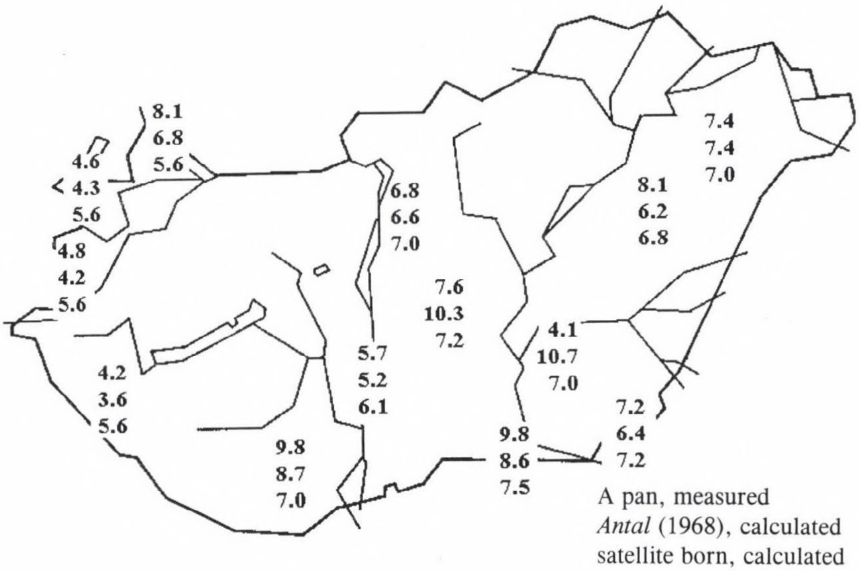

Fig. 4. Comparison of values derived from satellite data with surface measurements and calculation, July 23, 1998.

The calculation were carried out in dry and wet years as well. In present paper the results of the 1998 year are shown. The main problem concerning the application of the method is whether it is possible to determine realistically the daily sum of evapotranspiration using only one daily maximum value. The other problematic case is when the radiation balance will not follow sinus function when any kind of cloudiness appears. In this situation, to calculate the daily sum of ET using only one instantaneous value of temperature does not seem to be correct approach.

Nevertheless, when areal conditions are optimal for application of the method, the accuracy limit is approximately ±1 mm day^{-1}. It does not seem to be very good approach, if we take into consideration that the maximum value of daily evapotranspiration exceeds 10 mm day^{-1} very few times. But the accuracy of any classic method is more or less the same. We can accept this limitation, because in agrometeorological practice the 10 days time step is the standard, and on this time scale the calculation error seems to be not so bad.

Anyway, our basic approach is to exploit the standard meteorological observation system as simple as possible and the introduced method is applicable. *Fig. 3* summarizes the structure of the whole methodology of the introduced method.

5. Validation of the method

The introduced method produces areal values. The resolution of the prepared ET map is equal to the resolution of the satellite images. Practically, we can calculate with an approximately 3 km resolution. It means that we have got cca. 9,000 data every day for the whole territory of the country. The only possibility of the validation is, if the satellite born data are compared with the measured values in individual points. We have got only few places where A-pan measurements are available. For these stations the ET values using *Antal* (1968) method were calculated and compared with the satellite born values. As an example, the distribution of the values are presented on July 23, 1998 in *Fig. 4*. As it was mentioned, the method could be applied only on cloudless days. A really cloudless day occurs only few times during the vegetation period. *Table 2* shows a comparison for the days which met the requirements in 1998. In few cases the table shows good coincidence, but in few cases not. The basic reasons of the not too good fitting is the use of universal constant. But taking into consideration the difficulties of the determination of coefficients, the permanent lack of the necessary ground truth, we could be satisfied with the results as a first guess.

Table 2. Comparison of measured (Pan A), calculated by the method of *Antal* (1968), and satellite born values for cloudless days. The given values are the simple arithmetic averages of the values of the available stations (as shown in *Fig. 4*)

Day of calculation	**Pan A mm/day**	**Method Antal mm/day**	**Satellite born mm/day**	**A versus Antal %**	**A versus Satellite %**	**Satellite/ Antal %**
June 03, 1998	5.7	5.7	5.7	100	140	139
June 10, 1998	6.1	6.1	7.9	100	128	130
July 01, 1998	4.1	3.8	4.9	92	120	129
July 23, 1998	8.3	8.2	7.9	98	96	97
July 25, 1998	13.2	12.9	12.6	98	96	97
July 29, 1998	1.2	1.2	1.8	102	141	137
July 31, 1998	6.4	6.5	8.1	101	124	123
Aug 01, 1998	6.3	6.5	7.8	102	123	120
Aug 02, 1998	7.6	7.7	8.6	99	111	111
Aug 04, 1998	15.0	15.0	16.3	100	108	108
Aug 07, 1998	5.9	5.9	8.6	101	147	145
Aug 09, 1998	8.1	7.1	8.5	88	106	119
Aug 10, 1998	9.2	8.1	9.0	88	98	112
Aug 12, 1998	6.8	6.9	8.5	101	124	122
Aug 13, 1998	8.1	7.8	8.2	96	101	105

6. Conclusion

Based on NOAA-AVHRR visible, near-infrared, and thermal infrared data in combination with meteorological data, it is possible to calculate daily sum of (actual) ET for days when the satellite born surface data are available. The calculated values of areal ET were compared with surface evaporation measurements values. There are no direct methods to "calibrate" either the satellite born surface temperature or the daily sum of ET satisfactorily. The results of our method were compared with the standard surface A-pan network, and in few cases with the measurements and calculated ET at agrometeorological research sites. The method could be used in case of cloudless summer days to estimate territorial distribution of ET values for large area, but its accuracy is about ±1 mm day^{-1}. In our case, the main goal is not the precise determination of actual ET but to inform extension service about the possible drought situation or about area water stress status. To prepare the application of the method, air-plane born surface temperature measurements were carried out in some years. The goals of these measurements were double-folded. The first goal was to help the establishment of satellite calibration. The second one was to discover the surface temperature inequality. Cloudy weather could limit the application of the method. The longer cloudy period can mitigate the water stress status or completely stop it, when the soil moisture content is satisfactory. In this case, the areal evapotranspiration from the practical point of view is less important. The main parts of the investigation were the near-surface skin temperature measurements and the comparison of collected data with satellite born surface temperature. The determination of areal evapotranspiration for a large region was the main goal.

The results could be summarized that a system was introduced to determine the daily sum of evapotranspiration. The system is suitable for everyday operative application. Using the introduced method one day after the measurements, we can get the daily sum of evapotranspiration. The method is useful to calculate the areal evapotranspiration for a large region with the given error and limitation factors. Continuous application of the method can act as drought monitoring system as well.

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The impacts of the increasing drought frequency on the agricultural water management

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Abstract—The high risk of drought in Hungary can be illustrated by the high crop failure in the year 2000 when compared to the average crop of the past years. The crop failure of grain and industrial plant species varied between 15–40%. About 1.3 million hectares of arable land were affected by drought.

Increase in the frequency, length, and intensity of drought is more likely during the warming periods. The series of droughts in the past decades may be explained as a sign of global warming. The studies have shown that a small degree (+0.5 and +1°C) of global warming can cause significant climate change in Hungary. There will be a marked increase in summer temperature and a decrease in annual precipitation of about 10% (60 mm). Increase in sunshine hours can also be expected, and the length of periods with low relative humidity will probably grow, too. The drier air is accompanied with increasing potential evapotranspiration, that results in increasing transpiration intensity. Therefore, it is to be expected that the water demand of crops will increase.

As a result of global warming, the more frequent occurrence of drought will be an important challenge of the Hungarian society and economy during the following decades. That is why it is urgent to make efforts towards the development of a national drought strategy, that will coordinate and maximize the effectiveness of the work of all Ministries involved.

In this paper the authors present a survey of the drought situation in Hungary, and outline the most urgent tasks for reducing its harmful effects for the country.

Key-words: drought, drought frequency, drought strategy, climate change, irrigation water requirement.

1. Introduction

In Hungary, mainly in the Great Plain, the climate and weather have significant roles in the agricultural sphere, first of all through the variability of precipitation. Although the average climate conditions are favorable for the agricultural production in the Great Plain, there are some years or periods or months of years when the crop is decimated by drought, flood, or surplus water (*Szlávik* and *Fejér*, 1999). Nowadays, when there are permanent extreme climate occurrences or they last for years, the whole national economy can be shaken by the caused harms, while in the previous past they were accompanied by starvation, epidemics, and other social and economic tragedies (*Vágás*, 1982). However, nowadays it is concluded that unfavorable alterations of weather and climatic conditions and the potential coincidences of events observed in the past could lead to extraordinary consequences in the Tisza watershed (and in this manner on the Great Plain), as it was unfortunately evidence by large floods in 1998–2000 (*Szlávik*, 2002).

Because of the great climate sensibility of our society and economy, when there are some extreme or permanent weather oscillations, people instantly suspect a climate change, and they try to proclaim it as a result of human activities. Through the past centuries, deforestation, flood protection, drainage and riverbed controls were thought to be driven to serial, long lasting, and intensive extremes of weather. In the last century, building of dams and water basins, irrigation and other water management activities, nuclear explosions, air pollution, and changes in agricultural cultivation have become the supposed or real causes of the proceedings of extreme weather conditions or supposed climate change.

In Hungary, the question of climate change and the components of water balance was brought up in connection with flood release in the Great Plain after a permanent droughty period in the 19th century. During the last two or three decades, when some extreme weather events occurred, the scientists and the society concentrated on a possible global warming. While earlier the question was that whether the drainage of marshlands, moors, and flood inundated area of the Great Plain and the important riverbed controls had changed the climate of the country and caused a start of turning into desert. From the last two decades, people and scientists wonder whether the beginning of a warming up period is statistically significant? If yes, what kind of climate and water balance changes can be expected in this area? Will there be a change in the climate? Will there be a warmer and drier climate type? What kind of climate scenarios will be expectable in the following decades? What about temperature increase and the amount of precipitation? How will our economy and society change as a result of the climate change? What changes can be expected in the nature? Will there be an increasing drought frequency and intensity on the Great Plain?

In the last 150 years, some experts took up the question of climate change in the Great Plain, that was supposed to be the consequence of flood protection in particular after a series of droughty periods in the first half of the 1860s. In the 1930s, our acknowledged experts on climatology (*Róna*, 1936; *Réthly*, 1936) had to prove repeatedly, that the series of droughts in the first part of the '30s were not the consequence of flood protection, but concomitant characteristics of our climate. They could analyze a century-old time series of climate monitoring, because the systematic instrumental monitoring had been started in the first part of the 19th century. They both found out, that the control of the River Tisza and its creeks did not cause any turning into desert in the macroclimate of the Great Plain, however, the local microclimates became drier.

The weather of the last years can be characterized by the water balance of our climate with its long-lasting floods (*Szlávik* and *Fejér*, 1999), surplus waters besides unusual droughts and long-lasting heat waves. On the basis of climate records we can state that these climate extremities can be caused clearly neither by the flood protection or control of rivers in the 19th century, nor human activities in the 20th century (*Szlávik*, 2002).

It is founded on facts, that the frequent droughts and extreme weather events (floods, surplus waters, long-lasting periods of heat) from the last one or two decades are characteristics of our climate, but the increasing frequency and intensity of them can be the consequence of the global warming in the Carpathian Basin. However, we should consider the possibility of a warming up rising phase of a period of natural climate fluctuation, too.

2. Expectable climate changes increase the drought suspectibility in Hungary

Apart from the consequences of global warming, experts state that among the circumstances that cause drought in Hungary, we should emphasize the weather, hydrological, soil, and agronomical conditions equally. These factors can be particularly affected by the geographical position of a farm or the relief conditions. As mentioned above, the drought suspectibility is a typical characteristic of the climate in Hungary. The drought frequency shows significant regional variability, as it can be seen in *Fig. 1*. Drought affects the western part of the country which is richer in precipitation, as well as the much drier counties of the Great Plain. In summer, which is most important for cultivation, there is high drought frequency in the region of the Great Plain, and it causes significant damage to the fields with poor water management.

The high risk of drought in Hungary can be illustrated by the high crop failure in the year of 2000 compared to the average crop of the past years. The crop failure of grain and industrial plant species varied between 15–40%. About 1.3 million hectares of arable land were affected by drought that caused more or less crop failure.

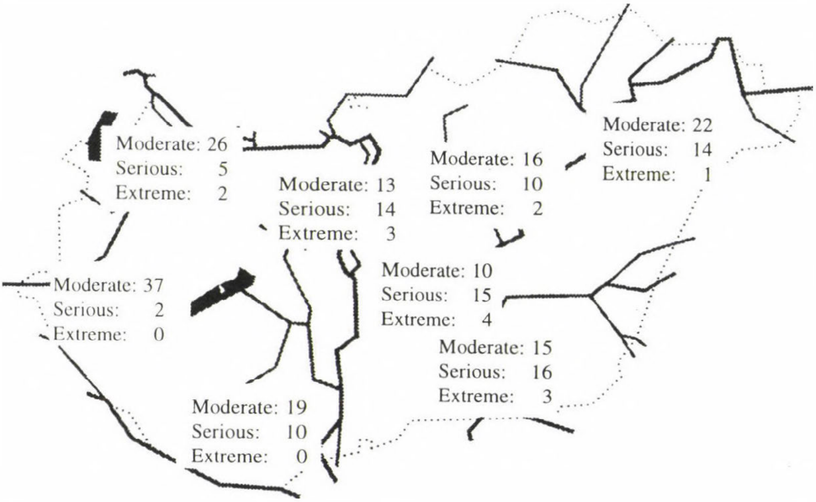

Fig. 1. Moderate, serious, and extreme drought occurrences in Hungary between 1951 and 1992 (after *Dunay* and *Tölgyesi*, 1993).

Table 1. Expectable changes in temperature and precipitation in Hungary till 2050 (after *Mika*)

Global change, °C	**Change in temperature in Hungary, °C**		**Change of annual precipitation, mm/ °C**
	summer	**winter**	
0.5	1.5	0	-60 (±20)
1.0	0.8	1.0–2.5	-60 (±40)
2.0	0.8	1.5	about 0
4.0	0.6–0.9	1.5	+10 (±100)

According to the researches on climate in Hungary (*Antal*, 1988, 1991, 1992, 2002; *Mika*, 1988, 1991, 1993), an increase in frequency, length, and intensity of drought can be supposed during the warming period to come. On one hand, the series of droughts in the past few years and decades may probably be explained as a sign of global warming. The studies have shown that a small degree (+0.5 and +1°C) of global warming can cause significant climate change in Hungary (see *Table 1*).

A marked increase in summer temperature and a significant decrease in annual precipitation of about 10% (60 mm) can be expected. Increase in sunshine hours can also be expected, while the length of periods with low relative humidity will probably grow. The drier air is accompanied with increasing potential evaporation, that results an increase in the intensity of transpiration. Therefore, it is to be expected, that the water demand of crops will increase, while the amount of precipitation and probably the number of rainy days will decrease during the warmer and drier growing season (*Antal*, 1988, 1998).

Table 1 shows that in summer, in case of little global warming (0.5°C), about one and a half times higher (that is 0.75°C) rising in temperature is expectable, while the annual precipitation can be reduced with 60 mm by each degree rise. In case of 1 or more degrees rise in temperature in the northern hemisphere, more moderate warming up is expectable in Hungary in summer, while in winter higher warming up can be expected (1.0–1.5°C). It is remarkable, that in case of higher global warming, the decrease of precipitation is not so significant, but its deviation in time increases (*Mika*, 1991, last line of *Table 1*).

3. Indirect and direct impacts of climate change on the agricultural water management

The indirect and direct impacts of climate change in Hungary is considerable by the following aspects (*Antal* and *Szesztay*, 1994):

- From the point of view of the agricultural sphere, the supposed climate change (first of all an increase in the frequency of drought) is one of those uncertain factors that have direct practical importance to the long-distance planning of agricultural water management.
- In the land ecological parts of a given area, the *water supply* for plant production that gathers precipitation, appears for farmers as the provider of complicated interactions between the climate, soil, and plants that in the end becomes the question of cultivation, pasture management, forestry and regional planning through water balance processes of the root zone.
- Finally, the examination of the possibility of feedback is essential. That is to find out the relationship between the role of climate regulation of atmospheric water cycle and the expectable climate change (climatic scenario).

When we try to analyze the effects of the increasing drought frequency on the agricultural water management, and to give a rough estimate of them, we have to emphasize that the changes in surface and subsurface water resources, caused by climate change or agricultural (or other anthropogen) activity in catchment areas, can be found out exactly only by the cognition and numerically definition of interactions between the water balance units and land ecological units divided into three parts of agricultural region.

The climate change can have a direct effect on the water supply of cultivated fields through the changes of hydrological cycle. In a qualified sense, besides the potential water supply, the whole water demand of the agricultural sphere may alter as a result of warming up, and because of it, an unfavorable change in the elements (first of all precipitation and evaporation) of climate may appear that will have an influence on the agricultural water management technologies (*Orlóci* and *Szesztay*, 2002).

In the future, the conditions of the agricultural water management can be modified by the change in climate, first of all the change in the elements of agricultural water balance. According to the latest examinations, the trends of the possible changes in the climate and water balance of our area are as follows:

- Global warming have a modifying influence on *water cycle and water supply* (surface and subsurface) that is commensurable with the effects of human activities on them, and sometimes it is more significant. As our climate is turning into drier, the damages caused by drought could be increased, because 1% decrease of precipitation (annual 6–7 mm) causes 1 m decrease of karst water level. A long-lasting period of decrease of ground water level could increase the sensitivity of plants in such areas, where it supplies more water to the stocks of plants nowadays.

- While *the soil moisture content* of the root zone follows the weather conditions (mainly precipitation) fast, the influence of climate change appears later as a change in the average humidity of soil supply of many years (turning into one direction).

- The change of *ground water content* during the time is influenced by precipitation and potential evaporation, and as these meteorological elements have a marked annual course, the ground water level also follows an annual rhythm. However, the long lasting warming up periods and the decrease of precipitation cause slow, but considerable decrease in the average of subsurface water resources of several years. The change can be dramatic in the areas having high ground water level, while in fields with lower ground water level it can be a bit more favorable.

4. Climatic water shortage and irrigation water requirement

As drought appears more often and more intensively, first of all, it has a marked influence on the development of need for irrigation water of plant stands in those fields, where there is a possibility for irrigation. However, the estimation of the expectable changes in the plant water requirements (numerical data) is one of the most complicated water management modeling tasks. The irrigation water requirement of plants changes in time and space, because it is controlled by the changing water demand of plants through the changeability of weather, water storage capacity of the root zone, depth of water table, physical properties of soil, and biological characteristics of plants (species, type, phase of development, population density, nutrient supply, etc.).

The available water supply for plants in cultivable lands can be well characterized by *climatic water demand* and its change in space and time. According to surveys in Hungary, as a consequence of worse precipitation supply, the climatic water demand is growing in the whole Carpathian Basin in the future. So the agricultural water management will ask for more and more water for irrigating purposes, while the neighbouring countries use more and more water from streams flowing through Hungary. As a consequence, our surface water supply decreases because of the decrease in precipitation and the decrease of water output of streams (*Antal*, 1988; *Nováky*, 1988; *Szesztay*, 1995). It is to be remarked that 10% decrease of precipitation causes more than 10% decrease of surface runoff. It is true for the climatic conditions in Hungary, that the modification effects of soil surface and cropcover (that causes a progressive change in crop rotation structure because of the increasing drought frequency) can cause a change of surface runoff in space and time, which can be equal with the change of precipitation, or it can be higher than that (*Nováky*, 1988; *Várallyay*, 1990).

According to the above, it is clear, that the supposed (predicted) climate change in Hungary has an unfavorable influence on almost all contaminants of plant water balance - supposing warming up in the Carpathian Basin and decreasing of precipitation as a consequence of it. Besides them longer, more intensive and more frequent droughts, as well as wider range of extremities in climate elements are expectable - so the availability of water supply will be less for the agricultural water management, while the quality of water will grow worse. This means that the costs of irrigation water and other water utilization will grow by the cost of purification of water. According to the water directives of EU, during the compulsory adaptation and implementation, additional cost growth is expectable.

5. The expectable impacts of the increasing drought frequency on biomass production

The effect of climate change on cultivation is significant from the point of view of the society, economy, and ecology, because the crop, the effectiveness and success of farming, and the annual fluctuation of it are caused by weather, and they have an effect on food industry, export market, the price structure of the domestic food market, and the employment. You should know that the agricultural sphere is almost the only stage utilizing the primer natural resources and converting its power (*Antal*, 1984; *Láng et al.,* 1983; *Szász*, 1985).

From the point of view of potential production, the temperature increasing is the smaller risk, although we can not neglect it. According to surveys in Hungary, in case of 2°C of temperature rise, the number of those days when the temperature reaches 30°C (that is they can be characterized by atmospheric drought) can be doubled (at present it is 15–30 days yearly). It will have a harmful influence on dry matter production (high breathing rate, turgor loss in the midday hours, intensive transpiration, disadvantageous water transport in plants etc.), mainly in the periods of droughts.

The expectable decrease of precipitation amount will cause more troubles for the agricultural water management with its worse water balance system. Supposing an annual 1 mm decrease of precipitation for 2020, the annual precipitation amount can fall under 470 mm in the middle of the Great Plain. Thus, producing of intensive plant species with more drought resistance is recommended, or we should calculate that the crop-time chart will be saturated, that is the crop will not grow in case of more intensive supplying of nutritive. It seems that in the next few decades of the 21st century, the agricultural water management (decrease of precipitation, irrigation) will be the most critical scientific problem of primer biomass production (*Varga-Haszonits* and *Harnos*, 1988).

On the basis of the above, it is evident that working out a well-adaptable ecological strategy and practical application of new technologies will be essential for our agriculture (*Szász*, 1993), because in case of high frequency of drought, the degree of supplying of precipitation will become one of the most important factors of ecological crisis. The indications of it have been shown since the beginning of the '80s and every Hungarian climate scenario has shown us the possible extrapolation of droughty tendency of the past one and half decades. They equally prognosticate rising in temperature and decreasing in precipitation. In *Fig. 2*, three categories of Palmer Drought Severity Index (PDSI; *Palmer*, 1965) are shown by way of illustration of increasing drought frequency since the beginning of the '80s (*Bussay et al.,* 1999). To the classification of droughts according to PDSI, they use three categories - *moderate* (PDSI<-2.0 and the drought extends to 50% of Hungary), *serious* (PDSI<-3.0 and the drought extends to 33% of the country) and *extreme* (PDSI<-4.0 and the drought extends to 20% of our country). These categories have been determined by *Bussay et al.* (1999). It can be seen that in the past two decades moderate and serious droughts have been ensued almost continuously in every year. On country-wide scale, drought occurred more frequently and more clustered at the end of the investigated period (1881-1995). The climatic stations exhibited a drying tendency verified by regression analysis. At most of the examined meteorological stations, the majority of the months show decreasing PDSI values (*Fig. 3*).

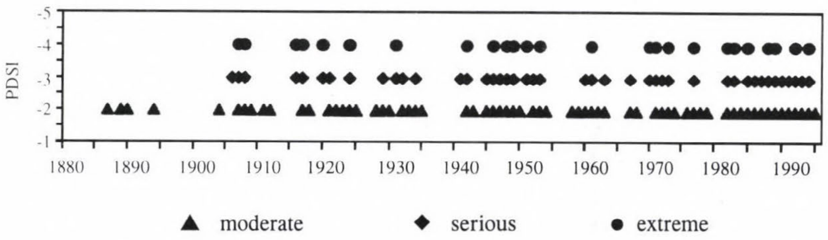

Fig. 2. Classifications of droughts according to the Palmer Drought Severity Index (PDSI) (after *Bussay et al.*, 1999)

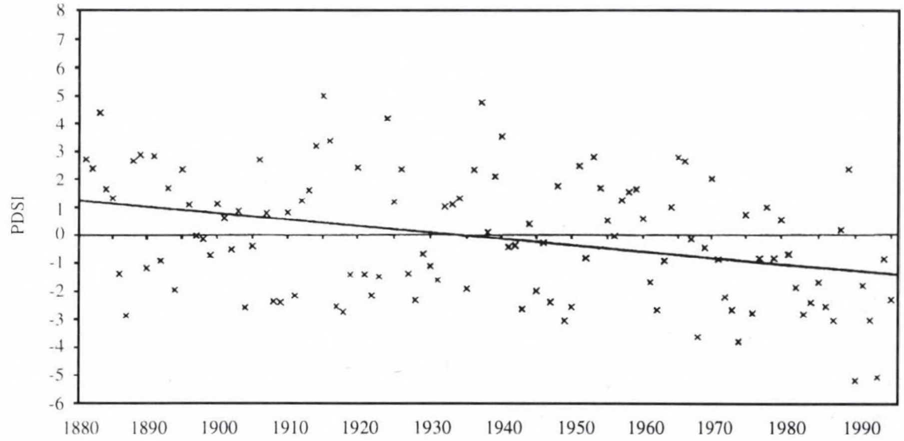

Fig. 3. The values and trend of PDSI in July (after *Bussay et al.*, 1999)

6. Increasing frequency of drought – national drought strategy

On the basis of experience of drought frequency in the last 1–2 decades, we can state as a summary that if we count on the decrease only of precipitation amount in the following decades, and a significant temperature rise will not occur in 20–30 years, we can not avoid working out a new ecological strategy to keep and even increase biomass production because of the expectable changes in the use of land.

The basis of the new ecological strategy is an up-to-date strategy of defending against drought (mainly passively), that is due to summarizing as a scientifically established national drought strategy (as a part strategy of the agricultural water management) and dealing with it as a government program (*Vermes*, 1998; *Vermes et al.*, 2000). On the other hand, it is necessary to increase the active agricultural water management interventions (irrigation, keeping precipitation in water basin, reducing evaporation and transpiration, reducing the surface runoff, managing the humidity of the soil more effectively, etc.).

Among the active processes of defence, we must review the technological and economical conditions of irrigation, and the practical and technical state of it. The present situation can not be preserved for a long time (preference, effectiveness, quality of crop, supply of liquid nutrients, soil-physical and chemical problems, etc.).

As in Hungary the agricultural sector remains a significant part of the economy for a long time (because of natural conditions of Hungary), it is essential to draw the political and economical decision makers' attention to all risk factors that can have influences on the agriculture of Hungary as the consequences of global environmental changes, which can endanger the use of land and primer plant production.

Among the natural factors we must emphasize the consequences of global warming in Hungary, but while working out the agricultural strategies (including national drought strategy), nowadays, and even in the future in connection with joining the EU, it is compulsory to considerate other anthropogenic effects, e.g., agricultural consequences of acid rains, acidification of soil as part of soil degradation, low ground water table and other environmental effects caused by global changes. The aim of a national drought strategy is to serve the expert and social cooperation, summarizing all the necessary concepts, methods, steps, and financial sources that are important in the fight against drought damages in a longer term (*Antal et al.*, 2002; *Vermes*, 1998). There will be a governmental body, which could maintain all those coordination tasks which are indispensably necessary in due time for the complex measures against drought damages and for an effective drought mitigation.

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IDŐJÁRÁS

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Estimate of the dry deposition of atmospheric nitrogen and sulfur species to spruce forest

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Abstract—A field campaign to determine the dry fluxes of nitrogen and sulfur compounds between the atmosphere and a Norway spruce forest was conducted during the spring of 1998. The field site was located in the Mátra mountains in north-eastern Hungary. Fluxes of particulate phase ammonium, nitrate, and sulfate were determined in addition to fluxes of gas phase ammonia, nitric acid, and sulfur dioxide. Dry deposition velocities determined by the gradient method are 1.9 and 1.3 cm s^{-1} for nitric acid, 3.7 and 1.1 cm s^{-1} for ammonia and 0.6 and 0.3 cm s^{-1} for sulfur dioxide in unstable and stable cases, respectively. For fine ammonium and sulfate particles there were no detectable concentration differences between the different measuring heights. For nitrate particles, existing mostly in the coarse fraction the gradient method is not applicable to determine the dry flux.

Key-words: dry deposition velocity, gradient method, sulfur dioxide/sulfate, nitric acid/nitrate, ammonia/ammonium.

1. Introduction

Knowledge of the dry deposition flux of different nitrogen and sulfur containing species is necessary for estimating the nitrogen and sulfur balance in forest ecosystems. However, there are substantial differences among the few

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existing data sets caused in part by differences in climatic conditions during the collection of the data.

Field studies carried out mostly during the last decade over forests have derived dry deposition velocities for gaseous ammonia ranging typically between 0.8 and 4.5 cm sec^{-1} (*Andersen et al.*, 1993; *Erisman et al.*, 1995; *Duyzer et al.*, 1994; *Wyers et al.*, 1995). In some cases emission was reported (*Andersen et al.*, 1999). Dry deposition velocities for nitric acid determined in field studies are much higher (*Janson* and *Granat*, 1997) than had been estimated on the basis of theoretical and field studies (*Baldocchi et al.*, 1992) due to the low canopy resistance of these species.

In the case of sulfur dioxide the dry deposition velocity measured over forests ranges between 0.05–3 cm s^{-1} (*Mennen*, 1995). For aerosol particles, limited field studies (*Erisman*, 1995; *Wyers*, 1995) have obtained systematically higher deposition velocities than there were determined by theoretical calculations and wind tunnel experiments (*Ruijgork et al.*, 1993 and *Borrell et al.*, 1997).

The aim of this paper is to provide data set for dry deposition velocities of these compounds, to increase the available data based on experimental investigations.

2. Measurements and calculations

Gradient flux measurements were carried out in the Mátra mountains, Hungary (Nyírjes station) in a Norway spruce forest, during a field campaign in May 1998. Characteristics of the measuring site are:

latitude = 47° 54' N,
longitude = 19° 57' E,
altitude = 560 m a.s.l.,
type of vegetation: Norway spruce (8 ha),
surroundings: Pine, Beech,
average height: 16 m,
age 33–35 years,
leaf area index (LAI): 3.3.

Concentrations of gaseous and aerosol species were measured with parallel three stage filter packs, located at heights of 13 and 23 meters on an instrumented tower for 8 hour sampling periods, during day and night (corresponding to unstable and stable conditions). Since the average calculated displacement height was 12 m, the lower height (13 m) is also above the canopy layer, therefore the turbulent motions are not disturbed by the canopy between the two measuring heights. A total of 18 measurements were carried out partly during daytime, partly in the night. A Teflon filter collected particles on the first stage of the filter packs, and on the second and third stages Whatman filters with basic and acid coatings collected acidic (nitric acid, sulfur dioxide) and alkaline (ammonia) gases, respectively. Concentrations of ammonium ions were determined by spectrophotometry using the indophenol-blue method, while nitrate and sulfate ions were determined by ion chromatography. The bulk precision of sampling and analysis was determined by parallel sampling at the same (23 m) level. The results of 10 measurements show that the mean relative error of sampling and measurements for all components is around $\pm$ 5 %.

The eddy-diffusivity of heat was calculated from the measured wind and temperature profiles on the basis of the *Monin-Obukhov* similarity theory (*Horváth et al.*, 1998; *Weidinger et al.*, 2000). The Monin-Obukhov length was calculated using the Richardson number derived from profile measurements and the universal functions for momentum and heat transfer (for details see *Horváth et al.*, 1998). It was assumed that the average bulk eddy-diffusivity for the trace materials is same as for the eddy-diffusivity of momentum and heat transfer. The values of the heat (K_T) were calculated on the basis of profile-flux relationship using the measured wind and temperature profiles. During the evaluation of the profile measurements the similarity theory provides the more appropriate estimations for the calculation of turbulent momentum and heat flux and hence for the calculation of the eddy-diffusivity (*Weidinger et al.*, 2000).

The eddy-diffusivity depends on the surface roughness as well as on meteorological conditions in the surface layer, especially the on stratification. During stable conditions (mostly during nighttime) the rate of exchange is generally lower, when the eddy-diffusivity is generally lower (0.3 $m^2 s^{-1}$) by a factor of three as compared to unstable conditions (0.9 $m^2 s^{-1}$ as an average, daytime). The eddy-diffusivity of the heat was considered as the measure of the turbulent exchange for all species. Stable and unstable cases were separated by the sign of the Richardson number.

3. Results and discussion

Dry deposition velocities of gases were calculated according to the gradient method. Concentration gradients of gases were multiplied by eddy-diffusivities. Eddy-diffusivities (K_T) and concentration gradients were determined separately for unstable (mainly daytime) and stable (nighttime) conditions. Concentration differences between the two sampling heights ranged between 4 and 82 per cent as shown in *Table 1*. The precision of concentration measurements was determined to be about ± 5 per cent in field and laboratory tests. The mean eddy-diffusivities for unstable and stable stratifications were 0.89 and 0.17 $m^2 s^{-1}$, respectively.

Table 1. Deposition velocities, mean concentrations and mean concentration differences of gases and aerosols

Compound	**Period**	**Mean deposition velocity ($cm\ s^{-1}$)**	**Mean concentration ($nmol\ m^{-3}$)**		**Mean concentration difference (%)**
			23 m	**13 m**	
HNO_3	day	1.9±0.4	11	8.1	27
	night	1.3±0.1	2.7	0.5	82
NH_3	day	3.7±0.4	28	18	35
	night	1.1±0.1	16	8.2	47
SO_2	day	0.6±0.4	134	124	8
	night	0.3±0.1	59	49	18
NO_3^-	day		6.9	4.8	30
	night		12	7.6	37
NH_4^+	day	<0.4±0.4	68	72	6
	night		82	78	4
SO_4^{2-}	day	<0.4±0.4	54	51	6
	night		51	49	5

According to the *Table 1*, in the case of nitric acid the deposition velocities for both unstable and stable conditions are similar, indicating substantial cuticular uptake. These figures are in the range of the deposition velocities collected by *Baldocchi* (1992), 0.5–5.0 cm s^{-1} for deciduous forests on the basis of model calculations compared to experimental results. The deposition velocity provided by *Janson* and *Granat* (1997), 7 cm s^{-1} for coniferous forest, is substantially higher, suggesting that canopy resistance is higher than it was expected by these authors.

For ammonia, where the stomatal uptake is the dominant deposition process, the deposition velocity figure (*Table 1*) is about three times higher when stomata are open (daytime) than during nighttime. These figures are in a good agreement with other field measurements above forests: *Andersen et al.*, 1993: 2.6 cm s^{-1} or 4.5 cm s^{-1} in the intensive vegetation period; *Duyzer et al.*, 1994: 2–3 cm s^{-1}; *Erisman et al.*, 1995: 2.5 cm s^{-1}; *Wyers et al.*, 1995: 1.5–4.3 cm s^{-1} at night, 0.8–4.0 cm s^{-1} in daytime, however the last authors did not found any difference between the deposition velocities during day and night hours. *Andersen et al.* (1999) pointed out the importance of ammonia emission from the forest, demonstrating that deposition dominates in the net flux when the ammonia concentration exceeds the compensation point.

The deposition velocity of sulfur dioxide is lower than that of nitric acid and ammonia during both stable and unstable conditions (*Table 1*). The sulfur dioxide deposition rate determined for forests ranges two orders of magnitude (*Mennen et al.*, 1995: 0.05–3 cm s^{-1}; *Erisman et al.*, 1995: 1.5 cm s^{-1}; *Horváth et al.*, 1997: 0.6–1.6 and 0.2–0.34 cm s^{-1}, during unstable and stable conditions, respectively).

As to the nitrate particles, annular denuder system (ADS) (cyclone, denuder tubes, filter pack) measurements made in parallel with the filter pack measurements show that nitrate particles are found mainly in the coarse fraction (*Table 2*). Therefore, the gradient method for determining the dry deposition of nitrate flux needs correction for the effects of gravitational settling.

Table 2. Share of nitrogen compounds in different phases in nmol m^{-3}

Phase	Nitrate/nitric acid	Ammonium/ammonia
Coarse particles d>2.5 m	18	<2
Fine particles d<2.5 m	<2	188
Gas	12	26

In contrast, ammonium particles were found solely in the fine fraction (d<2.5 m). The majority of the total nitrogen, consisting of ammonia gas, nitric acid, and particulate nitrate and ammonium, was contained in the form of ammonium fine particles. The molar ratio of ammonium to sulfate in the samples was between 1 and 2, i.e., sulfuric acid has been neutralized by ammonia to ammonium bisulfate or ammonium sulfate. Because ammonium (bi)sulfate was present mainly in the fine fraction, the gradient method is applicable provided gravitational settling can be neglected.

However, the average concentration difference between the upper and lower levels was between 4 and 6 per cent for ammonium and sulfate during both unstable and stable conditions. Considering the precision in sampling and measuring the concentrations with the filter pack method (about 5 per cent), and the precision (bulk relative error) of (estimated as 5–10 per cent) deposition velocities for ammonium and sulfate could not be determined. In *Table 1* estimated limits for deposition velocities of ammonium and sulfate can be found. According to the measurements concerning forests the dry deposition velocity of these compounds are: 2.4 cm s^{-1} for sulfate (*Sánchez et al.*, 1993); 1.2–1.5 cm s^{-1} for ammonium particles over 0.8 m size (*Wyers et al.*, 1995); 1–2 cm s^{-1} for fine and 5 cm s^{-1} for coarse sulfate and ammonium particles (*Erisman et al.*, 1995). Different research groups agree with the high uncertainty of these figures (*Lopez,* 1994). *Borrell et al.* (1997) suggests over 1 cm s^{-1} deposition velocity of particles according to the re-evaluation of theoretical, wind tunnel and field estimations (*Ruijgork et al.*, 1993). Our results contrast with most of these studies where relatively high ammonium deposition velocities were found suggesting that generalization of particle dry deposition field measurements is limited.

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On the relationship between the spatial variability of soil properties and transpiration

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Abstract—The sensitivity of transpiration, E, characteristics to areal variations of soil properties is analyzed. The study is performed by a *deterministic* model and a *statistical-deterministic* model. The core of the models is based on the *Penman-Monteith* concept. The transpiration characteristics are investigated in terms of analyzing the change of E versus soil moisture content, θ, relative frequency distribution of $E(\theta)$, and the algorithms for relating E obtained for a homogeneous and an inhomogeneous areal distribution of θ. The heterogeneity of soil characteristics is considered in terms of soil texture, areal variation of θ, and areal variation of soil hydraulic parameters (field capacity and wilting point soil moisture contents). In the study sand and clay are used as soil textures. The soil hydraulic parameters are evaluated for both the North-American and Hungarian soils. In the simulations strong and weak atmospheric forcing conditions are used.

According to the results, transpiration characteristics seem to be most sensitive to the areal variation of θ. The E characteristics are sensitive to the changes of soil texture and areal variation of soil hydraulic parameters at about the same rate, though the characteristics of the sensitivity are different. The analysis is valid when there are no advective effects and mesoscale circulation patterns. The results obtained can be useful for estimating area-averaged transpiration.

Key-words: transpiration, sand, clay, soil hydraulic parameters, heterogeneity, parameterization.

1. Introduction

A number of factors contribute to generate surface heterogeneity: (1) variability associated with surface type (e.g., vegetated surface, bare soil, inland water, snow, ice, urban area), (2) variability associated with terrain morphology (e.g., elevation, slope, and orientation), (3) spatial variability of climatic forcings (e.g., precipitation pattern on macro, -meso, and microscale, wind field), and (4) spatial variability of soil characteristics (e.g., soil color, soil texture, hydrophysical and thermal properties of the soil). Among the factors mentioned, we focus on the spatial variability of soil characteristics.

For evapotranspiration, among soil characteristics soil texture and soil hydrophysical properties are the most important (*Ek* and *Cuenca*, 1994). The spatial heterogeneity of these factors is pronounced. The spatial heterogeneity of soil texture is mostly represented by soil texture maps. The maps have different scales (for instance 1:100 000 or 1:500 000 scale) (*Várallyay et al.*, 1980), and they are based on different data sources (*Mika et al.*, 2002). These data refer to soil surface layer. Subsurface soil texture data are rarely available. This missing information can be important in some cases, because soil evaporation is sensitive to the changes of soil texture not only in the soil surface layer but also in the soil subsurface layer (*Ács* and *Lőke*, 2001a). Among soil hydrophysical properties, soil moisture content, θ, and the hydrophysical parameters (field capacity and wilting point soil moisture content, θ_f and θ_w, respectively) are the most important. Soil moisture content shows spatial variability on both the microscale and macroscale (*Robock et al.*, 1998). The macroscale variability of θ is determined by the macroscale precipitation pattern. The microscale variability of θ is observed on a few m^2 (*Rajkai*, 1991; *Seyfried*, 1998) and also on a few km^2 (*Nielsen et al.*, 1973; *Bell et al.*, 1980; *Hawley et al.*, 1983) of the soil surface. This inhomogeneous θ-field can be characterized by normal distribution (*Erdős* and *Morvay*, 1961; *Bell et al.*, 1980). In the point and close vicinity of it, the variability of θ can be neglected, therefore, the θ-field is homogeneous. The hydrophysical parameters depend not only on soil texture but also on soil type. Therefore, they also depend on geomorphological characteristics of the location (*Tomasella et al.*, 2000; *Ács*, 2002b; *Ács* and *Drucza*, 2003).

The impact of areal variations of these soil characteristics on evapotranspiration or transpiration is extensively investigated (*Ek* and *Cuenca*, 1994; *Boulet et el.*, 1999; *Braud et al.*, 1995; *Kim* and *Entekhabi*, 1998; *Giorgi*, 1997a, 1997b; *Mika et al.*, 2002; *Ács*, 2002a; *Ács* and *Lőke*, 2001b; *Ács* and *Szász*, 2002). In these studies, the impact of only one soil characteristic is analyzed. There is no study dealing with a comparative analysis between transpiration and all the three above mentioned soil properties: areal variation of θ, soil texture, and areal variation of soil hydraulic parameters. This comparative aspect is the novelty of this study. Such a comparative analysis would give an insight into the relative importance of the soil heterogeneity effects mentioned above.

In the study, we applied a diagnostic approach. The transpiration models for both the homogeneous and inhomogeneous areal distribution of θ are based on the *Penman-Monteith* concept (*Monteith*, 1981; *Ács* and *Hantel*, 1999; *Ács et al.*, 2000a; *Ács* and *Szász*, 2002, *Ács*, 2002a). In the study, we assumed that there are no advective effects accompanied by occasionally observed internal boundary layers (*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*, 1988).

In this study I use terms *homogeneous* and *inhomogeneous* θ rather than θ at the *point* and *local scale* as in the former studies (*Ács et al.*, 2000b; *Ács*, 2002a; *Ács* and *Szász*, 2002). I think the expressions "homogeneous θ" and/or "inhomogeneous θ" are more accurate than the expressions θ at the "*point*" and/or "*local*" scale. Namely, there is no any specific length scale to separate the so called point and local scales.

2. Models

Two model types are used for diagnozing transpiration: *deterministic* (D) and *statistical-detrministic* (SD) models. In the D-model there is no areal variation of θ, that is θ is homogeneous over the surface area under consideration. In the SD-model, in contrast to the D-model, there is areal variation of θ, that is θ is inhomogeneous over the surface area under consideration.

2.1 Deterministic model

In the model, we suppose that the vegetation canopy is completely closed, and that there is no interception. So the water vapor flux above vegetation is formed only by transpiration. The core of the model is based on the *Penman-Monteith* concept (*Monteith*, 1981) by the following equations:

$$H = A_e - \lambda E, \tag{1}$$

$$\lambda E = \frac{\Delta \cdot Ae + \rho \cdot c_p \delta e / r_a}{\Delta + \gamma\,(1 + r_v / r_a)}, \tag{2}$$

$$r_a = \mathrm{f}_1(u_*, L, \text{constants}), \tag{3}$$

$$u_* = \mathrm{f}_2(u_r, L, \text{constants}), \text{ and} \tag{4}$$

$$L = \mathrm{f}_3(u_*, H, E, \text{ constants}) = \frac{-\rho \cdot T_r \cdot u_*^3}{g \cdot k \cdot \left(\dfrac{H}{c_p} + 0.61 \cdot T_r \cdot E\right)} \tag{5}$$

The latent heat flux is determined by Penman-Monteith's equation, and the sensible heat flux is obtained as residual from the energy balance equation. The surface and aerodynamic transfers are parameterized using a resistance representation. The aerodynamic transport is parameterized using Monin-Obukhov's similarity theory taking into account the atmospheric stability (*Ács* and *Kovács*, 2001). The surface resistance of vegetation canopy is parameterized by *Jarvis* (1976) formula. The moisture availability function, Fma, is expressed as simply as possible via soil moisture content θ (e.g., Eq. (6) in *Ács et al.* (2001) or Eq. (35) in *Ács* and *Hantel* (1998)), that is I used the *Theta*-parameterization. The model is described in details in work of *Ács* and *Szász* (2002). A somewhat shorter description of the model is given in work of *Czúcz* and *Ács* (1999), *Ács* and *Hantel* (1999), and *Ács* (2002a).

2.2 Statistical-deterministic model

The model 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 $E(\theta)$. As input it uses atmospheric boundary conditions, soil/vegetation parameters (roughness length, zero plane displacement height, leaf area index, moisture content at field capacity, and wilting point), and the averaged value and standard deviation of the soil moisture content. It is described in details in works of *Ács et al.* (2000b), *Ács* (2002a), and *Ács* and *Szász* (2002), therefore, it will not be presented.

3. Numerical experiments

The numerical experiments are performed by both the *deterministic* and *statistical-deterministic* models for grass covered surface. The simulations are made for different atmospheric forcing conditions, soil textures, and soil hydraulic parameters. Concerning atmopheric forcing conditions, we distinguished strong and weak atmospheric forcings (*Table 1*).

Concerning soil textures, the simulations are performed for sand and clay. It has to be mentioned, that there are differences in the textural composition between the North-American and Hungarian soils. This is presented in *Table 2*. For sand, there are practically no differences. In spite of this, the differences between the North-American and Hungarian clay are somewhat greater. The silt and the clay fraction of the North-American clay is 20 and 58 percent. For Hungarian clay, the silt and clay fractions are 30 and 50 percent (*Ács* and *Drucza,* 2003). These differences are presumably caused not only by the differences in the geomorphological characteristics but also by the differences in the soil texture classification systems. A more detailed analysis of this subject can be found in the paper of *Ács* and *Drucza* (2003).

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
Wind velocity at reference level (m s^{-1})	6.0	2.0

Table 2. Average values of textural fractions in percentage for sand, loam, and clay of North-American and Hungarian soils

Soil texture	**Country**	**Sand fraction**	**Silt fraction**	**Clay fraction**
Sand	Hungary	92	5	3
	USA	92	5	3
Loam	Hungary	52	26	22
	USA	43	39	18
Clay	Hungary	22	30	50
	USA	22	20	58

The soil hydraulic parameters (field capacity and wilting point soil moisture contents) are chosen for parameterizations of *Clapp* and *Hornberger* (1978) (hereafter CH-parameterization) on one hand, and for parameterization of *Várallyay et al.* (1979) and *Rajkai* (1984, 1988) (hereafter VR-parameterization) on the other hand. The former parameterizations refer to North-American, while the latter parameterizations to Hungarian soils. The soil hydraulic parameter values for both the North-American and Hungarian soils are presented in *Table 3*. The values are presented for sand, loam, and clay assuming both the homogeneous and inhomogeneous areal distribution of θ. Note that the parameters for loam are also given in *Table 2* and *3*, although they are not used in the analysis. The $\theta_{hom, f}$-values, irrespectively of the parameterization used, are defined for $K(\theta) = 0.1$ mm/day. Similarly, the $\theta_{hom, w}$-values are defined for $\log \Psi(\theta) = 4.2$, when $\Psi(\theta)$ is given in cm water column height.

Table 3. Boundary values of soil moisture content obtained for North-American (parameterization of *Clapp* and *Hornberger* (1978); briefly CH) and Hungarian (parameterization of *Várallyay* (1979) and *Rajkai* (1984, 1988); briefly VR) soils for sand, loam, and clay as soil textures

Boundary values of soil moisture content ($m^3\ m^{-3}$)	**Sand**		**Loam**		**Clay**	
	CH	**VR**	**CH**	**VR**	**CH**	**VR**
$\theta_{inhom,w}$	0.03	0.02	0.08	0.08	0.15	0.14
$\theta_{hom,w}$	0.07	0.02	0.15	0.15	0.29	0.25
θ_c	0.09	0.06	0.18	0.20	0.31	0.30
$\theta_{hom,f}$	0.14	0.15	0.24	0.31	0.37	0.40
$\theta_{inhom,f}$	0.27	0.27	0.36	0.39	0.47	0.50
θ_S	0.40	0.42	0.45	0.48	0.48	0.51

Abbreviations: $\theta_{inhom,w}$ = the wilting point soil moisture content for inhomogeneous θ, $\theta_{hom,w}$ = the wilting point soil moisture content for homogeneous θ, θ_c = a soil moisture content value in the transition region for which $E_{hom}(\theta_c)=E_{inhom}(\theta_c)$, $\theta_{hom,f}$ = the field capacity soil moisture content for homogeneous θ, $\theta_{inhom,f}$ = the field capacity soil moisture content for inhomogeneous θ, and θ_S = saturated soil moisture content.

During numerical experiments, 2 × 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 because of the generation of statistical variables.

4. Simulation results

Verification of $E(\theta)$ model based on *Theta* parameterization assuming homogeneous areal distribution of θ is performed on the well known Cabauw data set from 1987 (*Beljaars* and *Bosveld*, 1997) . These results are presented in *Ács* and *Szász* (2002) and *Ács et al.*, (2000a).

The transpiration characteristics are considered in terms of analyzing the change of E versus θ, relative frequency distribution of $E(\theta)$, and the algorithms for relating E obtained for homogeneous and inhomogeneous θ. These features are analyzed for both North-American and Hungarian soils using sand and clay as soil texture.

4.1 Analysis of transpiration curves

The $E_{hom}(\theta)$ and $E_{inhom}(\theta)$ curves for strong and weak atmospheric forcing conditions are presented in *Fig. 1* and 2, respectively. From the point of view of the inhomogeneity of θ, the main characteristics are as follows:

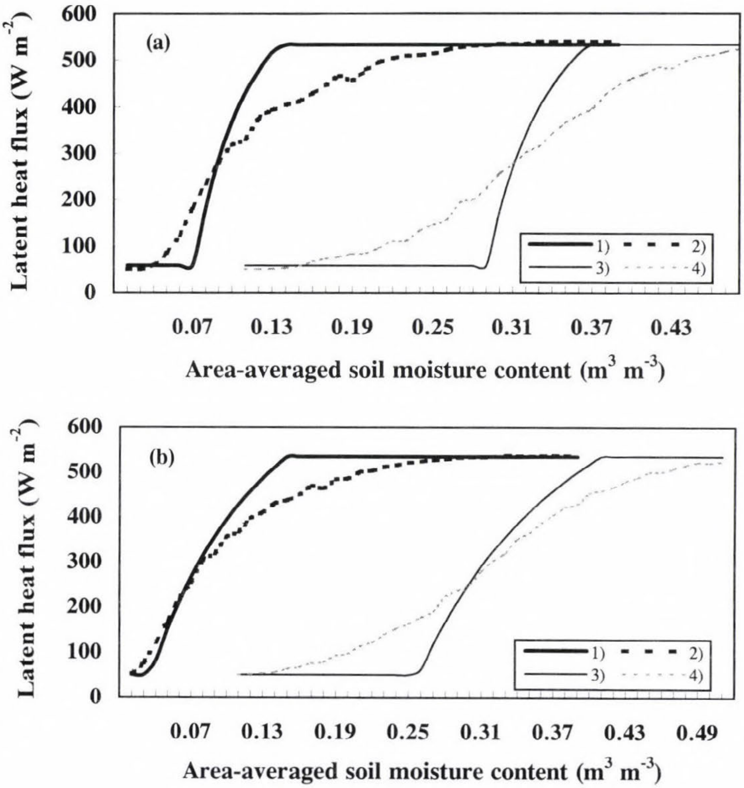

Fig. 1. Transpiration curve as obtained (*a*) by parameterization of *Clapp* and *Hornberger* (briefly CH-parameterization) (1) for sand and homogeneous θ (black continuous line), (2) for sand and inhomogeneous θ (black dashed line), (3) for clay and homogeneous θ (grey continuous line), and (4) for clay and inhomogeneous θ (grey dashed line); (*b*) by parameterization of *Várallyay* and *Rajkai* (briefly VR-parameterization) (1) for sand and homogeneous θ (black continuous line), (2) for sand and inhomogeneous θ (black dashed line), (3) for clay and homogeneous θ (grey continuous line), and (4) for clay and inhomogeneous θ (grey dashed line). The curves refer to strong atmospheric forcing conditions.

(1) the slope $S_{hom}=\partial E_{hom}(\theta)/\partial\theta$ is greater than the slope $S_{inhom}=\partial E_{inhom}(\theta)/\partial\theta$ in the transition region between the soil-controlled and atmospheric-controlled transpiration. This is caused because the transition region of $E_{inhom}(\theta)$ is much greater than the transition region of $E_{hom}(\theta)$. (2) The greatest $|E_{hom}(\theta) - E_{inhom}(\theta)|$ differences appear at θ_w and θ_f. At θ_w, $E_{inhom}(\theta) > E_{hom}(\theta)$, while at θ_f, $E_{inhom}(\theta) < E_{hom}(\theta)$. The course of $E(\theta)$ curves for North-American loam is also analyzed in details in the paper of *Ács* and *Szász* (2002).

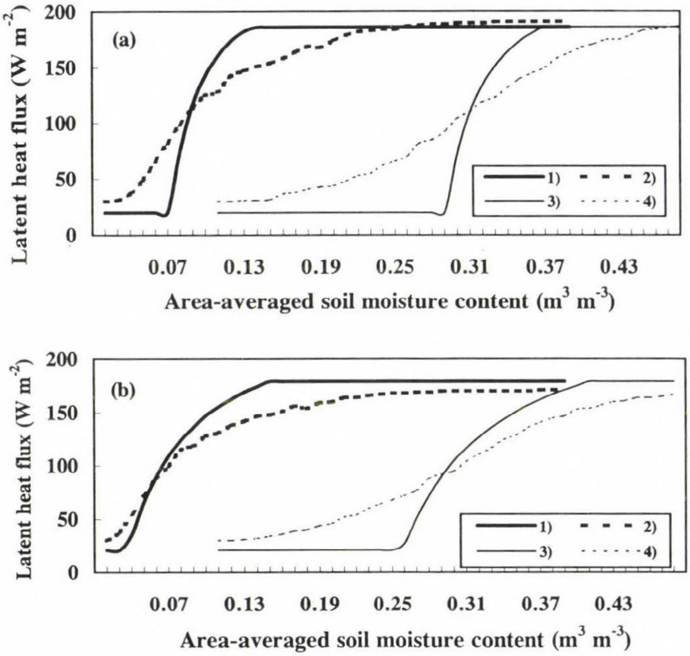

Fig. 2. As in Fig. 1 but for weak atmospheric forcing conditions.

From the point of view of the soil texture, the general characteristics of the course of $E(\theta)$ curves are as follows: (1) Potential (high θ-values ($\theta > \theta_f$) and atmospheric control) and water-stressed (low θ-values ($\theta < \theta_w$) and soil control) transpiration does not depend on the soil texture. (2) The slope $S_{hom} = \partial E_{hom}(\theta) / \partial \theta$ do not depend on soil texture, that is, $S_{hom}^{s} = S_{hom}^{c}$. In spite of this, $S_{inhom} = \partial E_{inhom}(\theta) / \partial \theta$ depends on soil texture, that is, $S_{inhom}^{s} \neq S_{inhom}^{c}$. The dependence of $E(\theta)$ curves on soil texture for North-American soils is also analyzed in details in the paper of *Ács* (2002a).

The course of $E(\theta)$ curves obtained for both the CH-parameterization (North-American soils) and VR-parameterization (Hungarian soils), separately for sand and clay is presented in *Figs. 3* and *4*, respectively. Inspecting the curves, the main characteristics are as follows: (1) The slope $S_{hom,CH} = \partial E_{hom,CH}(\theta)/\partial\theta$ is greater than the slope $S_{hom,VR} = \partial E_{hom,VR}(\theta)/\partial\theta$. (2) The potential and water-stressed transpiration are independent not only from the soil texture but also from the parameterization used. (3) The width of the transition region between the soil-controlled and atmospheric-controlled transpiration is different for North-American and Hungarian soils (see *Table 3*).

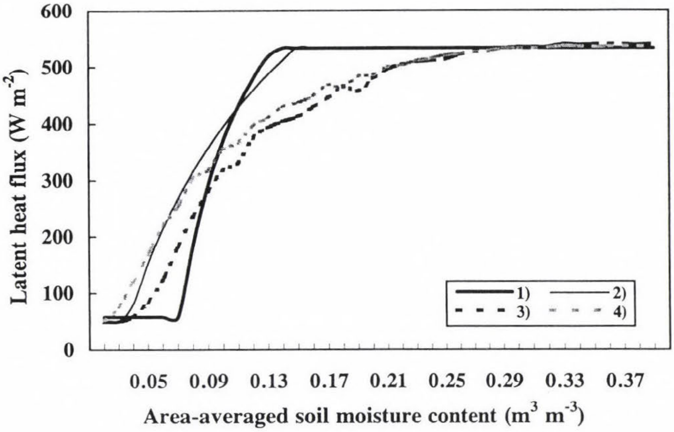

Fig. 3. Transpiration curve as obtained (1) for homogeneous θ using sand as soil texture and CH-parameterization (black continuous line), (2) for homogeneous θ using sand as soil texture and VR-parameterization (grey continuous line), (3) for inhomogeneous θ using sand as soil texture and CH-parameterization (black dashed line) and (4) for inhomogeneous θ using sand as soil texture and VR-parameterization (grey dashed line). The curves refer to strong atmospheric forcing conditions.

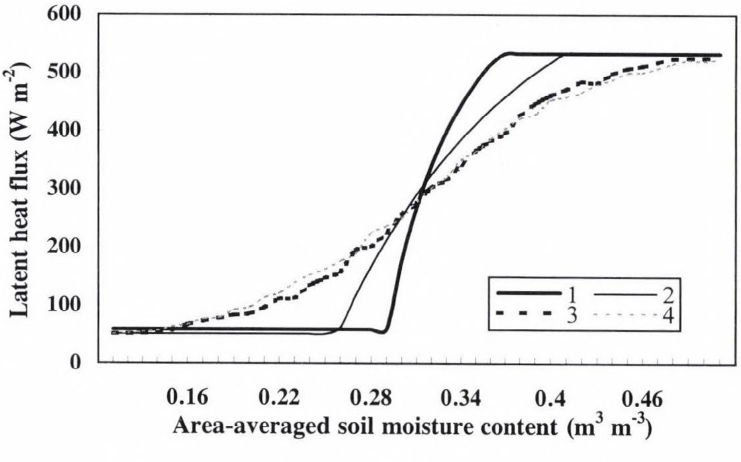

Fig. 4. As in Fig. 3 but for clay as soil texture.

The average width for all three textures is about 0.08 m^3 m^{-3} for CH-parameterization, while the same parameter for VR-parameterization is about 0.15 m^3 m^{-3}. The greatest differences between CH-parameterization and VR-parameterization appear for sand at $\theta_{hom,w}$. In this case, the $\theta_{hom,w,CH} - \theta_{hom,w,VR} \approx$

$0.05\ m^3\ m^{-3}$. (4) The course of $E_{inhom,CH}(\theta)$ and $E_{inhom,VR}(\theta)$ curves are very similar; the differences between them are very small. The greatest differences (about 80–90 W m^{-2}) appear for sand and strong atmospheric forcing in dry conditions ($\theta_{hom,w}$–$\theta_{inhom,w}$ region).

4.2 Areal variations of transpiration

Areal variation of $E(\theta)$ is analyzed inspecting its relative frequency distribution RF. The estimates are performed for strong atmospheric forcing conditions and two different θ_m-values. The characteristic θ_m-values are $\theta_{hom,w}$ (0.07 and 0.02 $m^3\ m^{-3}$ for sand and 0.29 and 0.25 $m^3\ m^{-3}$ for clay) and $\theta_{hom,f}$ (0.14 and 0.15 $m^3\ m^{-3}$ for sand and 0.37 and 0.40 $m^3\ m^{-3}$ for clay). The histograms of RF_{CH}^s and RF_{CH}^c for $\theta_{hom,f}$ and $\theta_{hom,w}$-values are presented in *Fig. 5a* and *Fig. 5b*, respectively. The histograms for VR-parameterization are shown in *Figs. 6a* and *6b*.

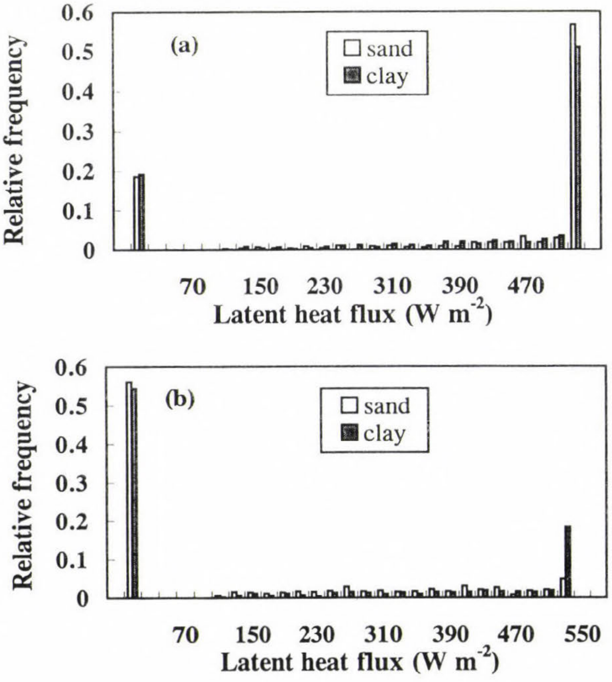

Fig. 5. Relative frequency distribution of latent heat flux from vegetation as obtained by CH-parameterization using sand (light columns) and clay (dark columns) as soil texture for (*a*) $\theta_{hom,f}$ and (*b*) $\theta_{hom,w}$. The results are obtained for strong atmospheric forcing conditions.

The RF^{s} and RF^{c} are very similar. In dry regime ($\theta_m = \theta_{hom,w}$), the RF maximum is on the left hand side of the spectrum. In wet regime ($\theta_m = \theta_{hom,f}$), the RF maximum is on the right hand side of the spectrum. For both parameterizations, the greatest deviation between the peaks referrig to sand and clay appear at $\theta_{hom,w}$ (see *Figs. 5b* and *6b*). For CH-parameterization, RF^{s} shows unimodal, while RF^{c} shows bimodal distribution. For VR-parameterization, both RF^{s} and RF^{c} show unimodal distribution.

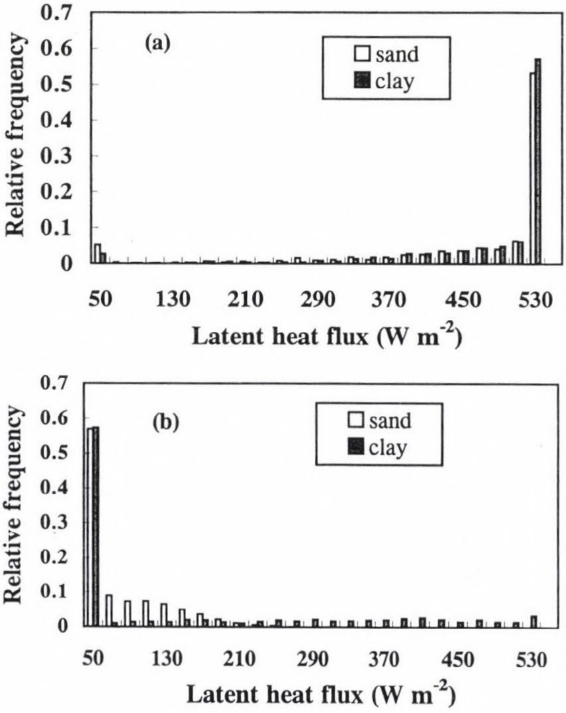

Fig. 6. As in Fig. 5 but for VR-parameterization.

Inspecting RF_{CH}^{s} versus RF_{VR}^{s} and RF_{CH}^{c} versus RF_{VR}^{c}, the main characteristic is as follows: RF_{CH} (excepting sand at $\theta_{hom,w}$) possesses a bimodal, while RF_{VR} rather an unimodal distribution. Note that there is a difference between RF_{CH}^{s} and RF_{VR}^{s}, though both relative frequency distributions are unimodal. It is hard to explain these RF-differences; they can presumably be related to the differences in soil moisture potentials obtained by CH- and VR-parameterizations (see *Ács* and *Drucza,* 2003).

4.3 Aggregated soil moisture content

The aggregated soil moisture content θ_{ag} is defined by

$$E(\theta_{ag}) = \langle E(\theta_m, \sigma_\theta) \rangle , \tag{6}$$

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. The relationship between θ_{ag} and θ_m can be obtained comparing $E_{hom}(\theta)$ and $E_{inhom}(\theta)$ curves (*Figs. 1* and *2*). The relationships obtained by such comparison refer to the transition zone of $E(\theta)$-curves. The change of $\theta_{ag,CH}^{s}$, $\theta_{ag,CH}^{c}$, $\theta_{ag,VR}^{s}$, and $\theta_{ag,VR}^{c}$ versus θ_m for strong and weak atmospheric forcing conditions is presented in *Fig. 7*.

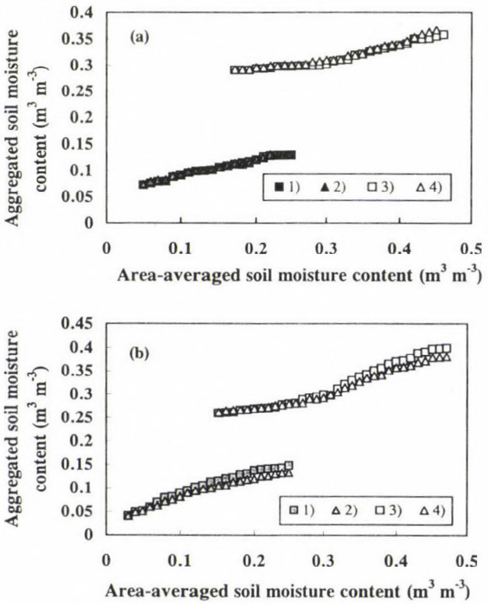

Fig. 7. Aggregated versus area-averaged soil moisture content as obtained (*a*) by CH-parameterization for (1) sand, using strong atmospheric forcing conditions (dark squares), (2) sand, using weak atmospheric forcing conditions (dark triangles), (3) clay, using strong atmospheric forcing conditions (light squares), and (4) clay, using weak atmospheric forcing conditions (light triangles); (*b*) by VR-parameterization for (1) sand, using strong atmospheric forcing conditions (grey squares), (2) sand, using weak atmospheric forcing conditions (grey triangles), (3) clay, using strong atmospheric forcing conditions (light squares), and (4) clay, using weak atmospheric forcing conditions (light triangles).

The relationship between θ_{ag}^{s} and θ_m is linear, while the relationship between θ_{ag}^{c} and θ_m is slightly non-linear. In all cases there is a dependence on the atmospheric forcing conditions. This dependence is weak in dry regime ($\theta_m < \theta_c$), but it is more stronger in the wet regime ($\theta_m > \theta_c$). This is in accordance with simulation results of *Shao et al.* (2001) made by a mesoscale atmospheric model.

The impact of atmospheric forcing conditions can be analyzed comparing $\theta_{ag}(\theta_m)$ courses for strong and weak atmospheric forcing conditions. The differences between $\theta_{ag,\,VR}(\theta_m)$ courses for strong and weak atmospheric forcings are greater than the differences between $\theta_{ag,\,CH}(\theta_m)$ courses for strong and weak atmospheric forcings. Therefore, the dependence on the atmospheric forcing conditions is stronger for VR-parameterization than for CH-parameterization. In spite of this fact, it has to be mentioned, that the course of $\theta_{ag,\,CH}(\theta_m)$ is quite similar to the course of $\theta_{ag,\,VR}(\theta_m)$.

5. Conclusions

The relationship between transpiration characteristics and areal variation of soil properties is analyzed. The transpiration characteristics are analyzed in terms of analyzing the change of *E* versus θ, relative frequency distribution of $E(\theta)$ and the algorithms for relating *E* obtained for homogeneous and inhomogeneous θ. The heterogeneity of soil characteristics is considered in terms of areal variation of soil moisture content, soil texture, and areal variation of soil hydraulic parameters (field capacity and wilting point soil moisture contents). In the analysis, sand and clay are used as soil textures. The soil hydraulic parameters are evaluated for both the North-American and Hungarian soils. The North-American soils are represented by parameterization of *Clapp* and *Hornberger* (1978), while the Hungarian soils by parameterization of *Várallyay et al.* (1979) and *Rajkai* (1984, 1988).

The transpiration for homogeneous θ is evaluated by a *deterministic* model. The transpiration for inhomogeneous θ is estimated by a *statistical-deterministic* model. The simulations are performed for strong and weak atmospheric forcing conditions. The results obtained can be summarized as follows:

- The slope of $E_{hom}(\theta)$ curves in the transition zone is much greater than the slope of $E_{inhom}(\theta)$ curves, irrespectively of the parameterization and soil texture used. In spite of this, the course of $E_{inhom,\,CH}(\theta)$ curves and $E_{inhom,\,VR}(\theta)$ curves are very similar. Further, the slope of $E_{hom}(\theta)$ curves in the transition zone, potential and water-stressed transpiration is independent from the soil texture used.

- The histograms of RF^s and RF^c for $\theta_{hom,f}$ and $\theta_{hom,w}$ are very similar. The greatest deviations between them appear for sand in dry conditions. In most cases the RF_{CH} possesses a bimodal distribution. In spite of this, the RF_{VR} shows an unimodal distribution.
- The $\theta_{ag}^{s}(\theta_m)$ relationship is linear, while the θ_{ag}^{c} (θ_m) relationship is slightly non-linear. In all cases there is a dependence on the atmospheric forcing conditions. This dependence is stronger for VR-parameterization than for CH-parameterization.

The results obtained are valid when there are no advective effects and mesoscale circulation patterns. In the latter cases the transpiration characteristics are more complex.

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List of symbols

H - sensible heat flux,
λE - latent heat flux,
λ - latent heat of vaporization of water,
A_e - available energy of vegetation surface,
Δ - slope of saturated vapor pressure curve,
ρ - air density,
δ_e - vapor pressure deficit,
r_a - aerodynamic resistance,
γ - psychrometric constant,
r_v - surface resistance of vegetation canopy,
u_* - friction velocity,
L - *Monin-Obukhov* length,
u_r - wind speed at reference height,
T_r - air temperature at reference height,
c_p - specific heat of air at constant pressure,
g - gravitational acceleration,
k - von Karman constant,
f_1, f_2 - functions which depend on the stratification and choice of the universal functions,
θ - soil moisture content,
θ_m - area-averaged soil moisture content,
σ_θ - standard deviation of soil moisture content,
θ_{ag} - aggregated soil moisture content,
$\theta_{ag,CH/VR}$ - aggregated soil moisture content obtained by CH- or VR-parameterization,
$\theta_{ag,CH}^{s/c}$ - aggregated soil moisture content obtained by CH-parameterization for sand or clay,
$\theta_{ag,VR}^{s/c}$ - aggregated soil moisture content obtained by VR-parameterization for sand and clay,
$\theta_{hom,w}$ - homogeneous θ at wilting point,
$\theta_{hom,f}$ - homogeneous θ at field capacity,
$\theta_{inhom,w}$ - inhomogeneous θ at wilting point,
$\theta_{inom,f}$ - inhomogeneous θ at field capacity,
$\theta_{hom,w,CH/VR}$ - homogeneous θ at wilting point obtained by CH- or VR-parameterization,
$\theta_{inhom,w,CH/VR}$ - inhomogeneous θ at wilting point obtained by CH- or VR-parameterization,

$\theta_{inom,f,CH/VR}$ – inhomogeneous θ at field capacity obtained by CH- or VR-parameterization,

$E_{hom}(\theta)$ – area-averaged transpiration flux in W m^{-2} for homogeneously distributed θ,

$E_{inhom}(\theta)$ – area-averaged transpiration flux in W m^{-2} for inhomogeneously distributed θ,

$E_{hom,CH/VR}$ – area-averaged transpiration obtained by CH- or VR-parameterization for homogeneously distributed θ,

$E_{inhom,,CH/VR}$ – area-averaged transpiration obtained by CH- or VR-parameterization for inhomogeneously distributed θ,

$S_{hom}(\theta)$ – slope of $E_{hom}(\theta)$ transpiration curve in the transition zone between the soil-controlled and atmospheric-controlled regime,

$S_{inhom}(\theta)$ – slope of $E_{inhom}(\theta)$ transpiration curve in the transition zone between the soil-controlled and atmospheric-controlled regime,

$S_{hom,CH/VR}$ – slope of $E_{hom}(\theta)$ transpiration curve obtained by CH- or VR-parameterization,

$S_{inhom,CH/VR}$ – slope of $E_{inhom}(\theta)$ transpiration curve obtained by CH- or VR-parameterization,

RF – relative frequency distribution of $E(\theta)$,

$RF_{CH/VR}$ – relative frequency distribution of $E(\theta)$ obtained by CH- or VR-parameterization,

$RF^{s/c}$ – relative frequency distribution of $E(\theta)$ obtained for sand or clay

$RF_{CH}^{s/c}$ – relative frequency distribution of $E(\theta)$ obtained by CH-parameterization for sand or clay.

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Reliability of estimated global radiation for crop model input

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Abstract—There are several estimation methods to calculate daily global radiation using different input data. The Szász method uses the daily sum of sunshine hours, the Ritchie-Fodor method uses the daily thermal oscillation. It was investigated whether global radiation estimated by the Szász method or Ritchie-Fodor method can substitute the measured global radiation for biomass and yield predictions in the 4M crop simulation model. The reliability of estimated global radiation is much greater if the data is used for yield prediction than for biomass prediction. The global radiation estimated by the Szász method can substitute measured data in 4M and other CERES based crop models with 89–96% of reliability, when yield prediction is the goal. The Ritchie-Fodor method made less reliable radiation estimations. This method needs further development.

Key-words: radiation estimation, 4M crop model, yield predictions.

1. Introduction

The primary purpose of crop models is to describe the processes of the very complex atmosphere-soil-plant system using mathematical tools and to simulate them with the help of computers. The ultimate aim of using crop models, however, is to answer questions that otherwise could only be answered by carrying out expensive and time-consuming experiments.

In the year 2000 a workshop, called 4M, was set up within the new System Modelling Section of the Hungarian Soil Science Society, with the specific purpose of creating an easy-to-use package for modeling cropping systems. The 4M package was developed at RISSAC using predominantly the results of Hungarian scientists from various institutes in the country. The CERES model was chosen to be a starting point for this project, since it has an open source code, and several studies have proved its competence in describing the soil-plant-atmosphere system (*Kovács et al.*, 1995; *Németh*, 1996; *Jamieson et al.*, 1998).

The accuracy of a crop model is judged mostly by how precise it is in estimating the production. The preciseness of a crop model is determined, on one hand, by the authenticity of the algorithms describing the processes of the real world, and on the other hand, by the quality of its input data. Even the "perfect" model would not be able to simulate the real processes precisely if inaccurate input data were fed into it.

Air temperature, global radiation, and precipitation are the key inputs for most crop models. In 4M, a number of calculations are based on the global radiation, such as soil evaporation, plant transpiration, and daily amount of photosynthesis. Despite of this, a considerable source of error is made while making direct measurements of solar radiation, all over the world. Accurate measured data is rarely available even in the United States. What adds to the difficulties of inaccuracy in Hungary is the limited number of the places where observations are made. This hinders the modelers in extending the spatial validity of crop simulation, which explains why modelers are interested in using estimated data instead of measured ones, that being the case when a good model is at hand for estimation, and easily accessible data are available. There are estimation methods to calculate daily global radiation using different input data: *Ångström*, 1924; *Szász*, 1968; *Bristow* and *Campbell*, 1984; *Fodor et al.*, 2000; *Donatelli* and *Bellocchi*, 2001. The first two methods use the daily sum of sunshine hours, the other three use the daily thermal oscillation. In our present study it was investigated whether the global radiation estimated by the Szász method or Ritchie-Fodor method can substitute the measured global radiation in 4M (*Fodor et al.,* 2003) crop simulation model.

2. Materials and methods

The 4M crop model was used for the test. Since the basis of 4M is the well-known CERES model, that was used as a starting point for several other crop models in the world, the result of this test gives relevant information for a wide range of scientists.

A daily global radiation estimation method was elaborated by *Szász* (1968). It calculates the daily global radiation from the daily sum of sunshine hours, depending on which month the observed day belongs to.

$$R = 0.024347 \cdot V1 \cdot C \cdot \Omega \cdot G_{max}\,, \tag{1}$$

where R is the daily global radiation (MJ m^{-2}),
$V1$ is the correction of Albrecht's loss factor,
C is the ratio of sunshine and sunlit hours,
Ω is the length of sunlit period of a day (h),
G_{max} potential value of daily global radiation (MJ m^{-2}).

For a given day of the year, knowing the sum of sunshine hours, the values of $V1$, C, Ω, and G_{max} can be read from a table given by Prof. Szász.

The Ritchie-Fodor method (*Fodor et al.*, 2000) uses the difference between the daily maximum and minimum temperatures for predicting the global radiation.

$$R = ETR \cdot FCD \cdot CDT^{OM}, \tag{2}$$

where R is the daily global radiation (MJ m^{-2}),
ETR is the extraterrestrial radiation (MJ m^{-2}),
FCD is the fraction of clear day (varies between 0 and 1),
CDT is the clear day transmissivity (varies between 0 and 1),
OM is the optical airmass (varies between 1.58 and 3.89 in Hungary).

Since ETR and OM are the functions of the latitude and day of the year solely, and CDT can be expressed as a function of the daily maximum temperature, and FCD can be expressed as a function of the daily thermal oscillation, with the help of Eq. (2) the daily global radiation can be estimated for a given location and day of the year – knowing the daily maximum and minimum temperature.

The algorithms of the two procedures were incorporated into the 4M cropping model package, therefore, there is an option to select the measured or estimated data directly for model runs. The Szász method was tested using a twenty-year long (1968–1987) independent dataset coming from the meteorological station of Pestlőrinc, Hungary (*Fig. 1*).

Since the Ritchie-Fodor method was developed using the dataset from Pestlőrinc, it was tested on a different, independent, ten-year long (1970–1979) dataset from Debrecen (*Fig. 2*), that was offered by Prof. Szász.

We were not only interested in the direct comparison of the observed and estimated values, since it had already been done from a meteorological point of view. Rather, the focus of this study was to figure out the effect of errors of radiation estimations on the crop model outputs. The obvious aim of this is to see the expected reliability of the generated radiation data for those locations, where there was no observation of global radiation, but there was observation of sunshine hours or temperature, respectively.

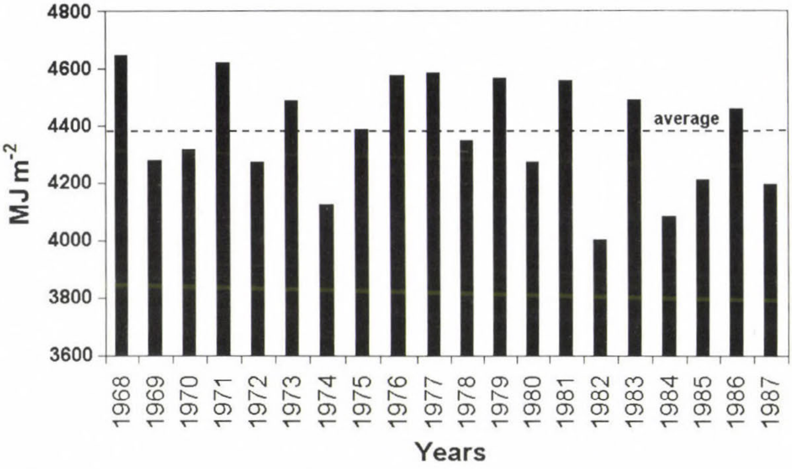

Fig. 1. The yearly sum of observed global radiation at Pestlőrinc, Hungary, during the studied period.

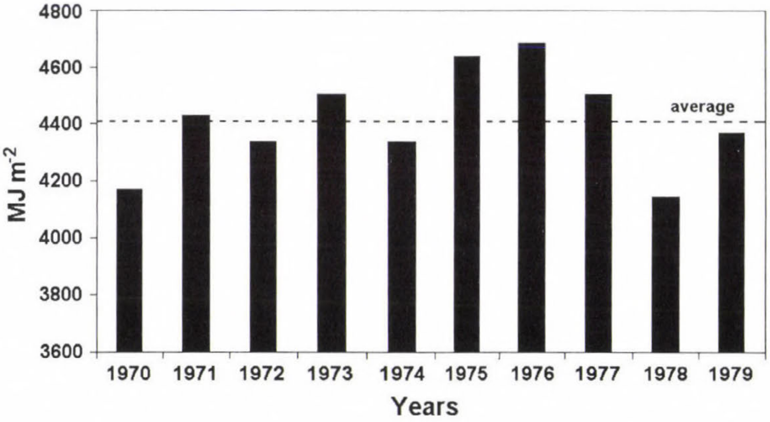

Fig. 2. The yearly sum of observed global radiation at Debrecen, Hungary, during the studied period.

A recent study on the sensitivity of crop models to the inaccuracy of meteorological observations (*Fodor* and *Kovács*, 2003) showed that the uncertainty caused by the errors of the measured global radiation can be up to 6% and 11% for the calculated biomass and yield, respectively. These thresholds (acceptance limits) were used for deciding whether the radiation estimation is acceptable for the crop model or not. If the difference between the model results obtained by using estimated and measured radiation is less than the above mentioned limits, the radiation estimation is said to be acceptable.

Since global radiation indirectly affects the water balance of the soil, two soil profiles, having different water regimes, were selected for the model runs: a calcaric chernozem soil on loam (*Table 1*) and a meadow soil on clay (*Table 2*). Soil data were provided by the Research Institute of Soil Sciences and Agricultural Chemistry (*Rajkai et al.*, 1981; *Várallyay*, 1987).

Table 1. Some characteristics of the loamy chernozem soil profile. Ks stands for hydraulic conductivity

Horizon	**Bulk density** **($g\ cm^{-3}$)**	**Humus** **(%)**	**Ks** **($cm\ d^{-1}$)**	**Sand** **(%)**	**Silt** **(%)**	**Clay** **(%)**
A_p	1.49	4.23	8.82	17.7	60.1	22.2
A	1.45	3.69	4.63	19.4	58.1	22.5
B	1.32	2.49	6.33	18.1	57.5	24.4
BC	1.32	0.00	6.35	16.9	60.1	23.0
C	1.43	0.00	3.50	21.8	56.4	21.8

Table 2. Some characteristics of the clayey meadow soil profile. Ks stands for hydraulic conductivity

Horizon	**Bulk density** **($g\ cm^{-3}$)**	**Humus** **(%)**	**Ks** **($cm\ d^{-1}$)**	**Sand** **(%)**	**Silt** **(%)**	**Clay** **(%)**
A_p	1.22	4.48	5.96	6.9	32.4	60.7
B1	1.28	2.01	2.38	8.9	31.6	59.5
C	1.48	0.00	0.28	7.3	47.0	45.7

The genetic parameters of the Pi3978 cultivar (maize) were used as crop specific inputs. Each run started on the 1st of April, the initial water content of the soil profiles was set to 90% of the field capacity.

To test the Szász method, 80 model runs (one year – one run) were made using the measured and estimated global radiation input of the selected 20 years (Pestlőrinc: 1968–1987), on two soil types. Similarly, to test the Ritchie-Fodor method, 40 additional model runs were made (Debrecen: 1970–1979). For the simulated biomass and yield, the difference between the runs with measured and estimated radiation was calculated.

3. Results

First, the measured and estimated global radiation values were compared for the two methods (*Figs. 3–4*). As one can see, the correlation coefficient is very high for the Szász method, even though it slightly underestimates the global radiation (*Fig. 3*). The correlation coefficient is much worse for the Ritchie-Fodor method, the dots on the graph are much more scattered. This method slightly overestimates the global radiation (*Fig. 4*). This does not come as a surprise since the connection between the global radiation and daily sunshine hours is much stronger than between the radiation and daily thermal oscillation.

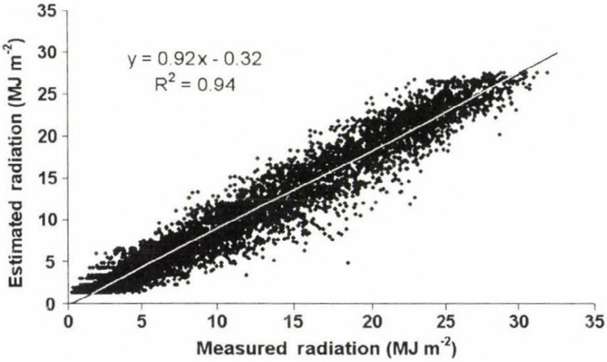

Fig. 3. Measured and estimated (with the Szász method) global radiation values at Pestlőrinc, Hungary, 1968–1987.

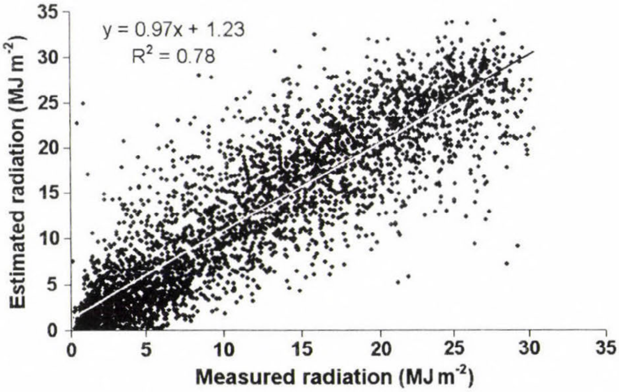

Fig. 4. Measured and estimated (with the Ritchie-Fodor method) global radiation values at Debrecen, Hungary, 1970–1979.

The 4M model was run for biomass and yield predictions using measured and estimated radiation. For the two soil types the same results were obtained (*Table 3*).

Table 3. Average difference between the run results with measured and estimated radiation as a percentage of the results obtained by using measured radiation. The ± values stand for the confidence interval having $\alpha = 0.05$

Method	**Soil type**	**Biomass (%)**	**Yield (%)**
Szász	Loamy	5.7 ± 2.1	8.9 ± 3.2
Ritchie-Fodor	Loamy	10.3 ± 3.6	15.7 ± 5.3
Szász	Clayey	5.4 ± 2.1	8.4 ± 2.9
Ritchie-Fodor	Clayey	8.7 ± 3.9	11.8 ± 5.2

Taking the average difference for the two soil types, the Ritchie-Fodor method made unacceptable radiation estimations causing 9.5% and 13.8% "errors" in average in the calculated biomass and yield, respectively. That means that even the average differences (for the biomass and yield) are greater than the acceptance limits. There were only two years (20% of the cases), where the radiation estimations were acceptable for the loamy soil, and five years (50% of the cases), where the radiation estimations were acceptable for the clayey soil.

If we take the average difference for the two soil types, we can conclude that the Szász method made acceptably good radiation estimations causing 5.6% and 8.7% "errors" in average in the simulated biomass and yield, respectively. Since these values are within the acceptance limits, this result is somewhat misleading. We calculated the level of confidences at which the whole confidence intervals are within the acceptance limits (*Table 4*).

Table 4. Level of confidences for the Szász method at which the whole confidence intervals are within the acceptance limits, and the corresponding percentages of the acceptable cases (years)

Soil type	**Biomass**	**Yield**
Loamy	$\alpha = 0.76 \rightarrow 62\%$	$\alpha = 0.22 \rightarrow 89\%$
Clayey	$\alpha = 0.60 \rightarrow 70\%$	$\alpha = 0.08 \rightarrow 96\%$

Only 62–70% of the radiation estimations were acceptable for biomass predictions and 89–96% for yield predictions depending on the soil type. The reason for this is that the Szász method slightly underestimated the global

radiation at Pestlőrinc. Perhaps the method is not valid for this site, or the "radiating circumstances" have been changed since 1968.

We modified Eq. (1) by simply multiplying it with 1.08, so that the slope of the graph in *Fig. 1* would be 1. The modified Szász method gave 100% acceptable radiation estimations for crop model input for Pestlőrinc.

5. Conclusions

Daily global radiation is one of the key inputs for most crop models. Since radiation is measured only in a few places, and the measurements are often loaded with errors, there is a great need for radiation estimators for crop models.

The reliability of estimated global radiation is much greater if the data is used for yield prediction than for biomass prediction.

Based on this study, the global radiation estimated by the Szász method can substitute measured data in 4M and other CERES based crop models with 89-96% of reliability, when yield prediction is the goal. The Ritchie-Fodor method is not reliable for radiation estimations for crop models yet, it needs further development. Since temperature is measured at many more locations than sunshine hours, our group continues to work on the method.

The Szász method should be revisited, so that it would be a 100% reliable method both for yield and biomass predictions for the whole country. It should be recalibrated by using the latest databases of more meteorological stations from the country.

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