ACTA UNIVERSITATIS SZEGEDIENSIS

PARS CLIMATOLOGICA SCIENTIARUM NATURALIUM

CURAT: G. PÉCZELY

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Conservation Probabilities of the Temperature Anomalies of Subsequent Months in the North-Atlantic-European Area

by

G. Péczely

Egymást követő hónapok hőmérsékleti anomáliáinak megmaradási valószínűsége az északatlanti-európai térségben. A tanulmány az északatlanti térség keleti felét és Európát magában foglaló, a 35 °W és 45 °E hosszúsági körökkel, s a 70 °N és 35 °N szélességi körökkel határolt szektor területéről kiválasztott 28 állomás 1901—1960 közötti havi középhőmérsékletei alapján vizsgálja az egymást követő hónapok hőmérsékleti anomáliáinak előjel szerinti megmaradási valószínűségét.

Megállapítja, hogy az anomáliák előjelének megmaradási valószínűsége az éven belül nem azonos, hanem erősebb—gyengébb évi menetet mutat. Ez az évi menet elsősorban Nyugat- és Közép-Európa térségében jellegzetes, ahol az anomáliák legnagyobb megmaradási hajlama nyáron, a legkisebb pedig ősz végén, tél elején mutatkozik. A reális statisztikai kapcsolatra utaló nagy megmaradási valószínűségekkel rendelkező hónappárok száma Közép-Európában jóval kevesebb, mint a vizsgált térség óceáni klimaterületein. Ezért ezen a területen a hőmérsékleti anomáliák hosszú távú előrejelzésénél felhasználható statisztikai módszerek alkalmazása korlátozott.

Die Erhaltungswahrscheinlichkeit der Temperaturanomalien der aufeinanderfolgenden Monaten in dem Nordatlantischen Raume. In der Arbeit wird auf Grund von 28 ausgewählten Stationen, welche sich in der östlichen Hälfte des Nordatlantischen Raumes und in Europa zwischen den Längenkreisen 35 °W und 45 °O und den Breitenkreisen 70 °N und 35 °N befinden, aus den Monatsmitteltemperaturen des Zeitraumes 1901—1960, die Erhaltungswahrscheinlichkeit des Vorzeichens der Temperaturanomalien von zwei aufeinanderfolgenden Monaten untersucht.

Es wird festgestellt, daß die Erhaltungswahrscheinlichkeit des Vorzeichens der Temperaturanomalien im Laufe des Jahres nicht die gleich ist, sondern einen schärfer oder schwächer ausgeprägten Jahresgang aufweist. Dieser Jahresgang ist in erster Reihe für West- und Mittel-Europa charakteristisch, wo die größte Erhaltungstendenz der Anomalien im Sommer und die geringste am Ende des Herbstes und am Anfang des Winters auftritt. Die Anzahl der Monatspaare, welche durch den hohen Wert der Erhaltungswahrscheinlichkeit auf die Existenz eines realen statistischen Zusammenhanges hinweisen, ist in Mittel-Europa bedeutend geringer als in den ozeanischen klimatischen Gebieten des untersuchten Raumes. Somit ist die Verwendungsmöglichkeit der statistischen Methoden für die langfristige Vorhersage der Temperaturanomalien in diesem Gebiete beschränkter.

In this study, we are investigating, on the basis of 28 selected stations situated in the eastern half of the North Atlantic and in Europe, between the longitudes 35 °W and 45 °E and the latitudes 70 °N and 35 °N, and the monthly temperature anomalies observed during the period 1901—1960, the conservation probabilities of the signs of temperature anomalies in subsequent months.

It is found, that the conservation probability values are not the same during a year, but they are exhibiting a more or less important annual variation. This annual variation is mainly in Western and Central Europe a characteristical one, where the highest tendency of the conservation of anomalies is appearing in summer, and the lowest is occurring at the end of autumn and at the beginning of winter. The number of high conservation probability values, indicating the existence of a realistic statistical relationship, is essentially lower in Central Europe than in the Oceanic area under investigation. Thus, the applicability of statistical methods in the field of long-range forecasting of temperature anomalies is rather a limited one in this area.

In the field of long-range forecasting of monthly anomalies of temperature, it is of a fundamental importance to elucidate the question, whether the time series of monthly anomalies can be considered as a chain of events that are independent from or, on the contrary, dependent on the antecedent values. Such investigations are also of further interest, in addition to their application in long-range forecasting, in studies on the general circulation of the atmosphere and from the point of view of a more fundamental knowledge of the statistical structure of meteorological time series. The latter application is yielding a possibility for a more precise formulation of the statistical models which are serving to the description of the time series of the anomalies.

In the present study, we are investigating, on the basis of monthly mean temperatures observed at 28 selected stations situated in Europe and in the eastern half of the Northern Atlantic area between the longitudes 35° W and 45 °E and between the latitudes 70 °N and 35 °N, during the period 1901 to 1960, the probabilities of the conservation of the signs of subsequent mensual temperature anomalies. The names and geographical co-ordinates of the stations used are listed in *Table I.*

Table I

Station network

Angmagssalik	65°37'N	37°34'W
Stykkisholm	65°05'N	22°44'W
Bergen	60°24'N	5°19'E
Bodo	67°16'N	14°22'E
Karasjok	69°28'N	25°31'E
Arkhangelsk	64°35'N	40°30'E
Valentia	51°54'N	10°15'W
Aberdeen	57°12'N	20°12'W
De Bilt	52°06'N	5°11'E
Berlin	52°27'N	13°18'E
Uppsala	59°52'N	17°38'E
Vilnius	54°42'N	25°18'E
Leningrad	59°58'N	30°18'E
Moscow	55°45'N	37°34'E
Okt. Gorodok	51°38'N	45°27'E
Paris	48°49'N	2°30'E
Marseille	43°27'N	5°13'E
Basel	47°33'N	7°35'E
Roma	41°48'N	12°36'E
Budapest	47°31'N	19°01'E
Sibin	45°48'N	24°09'E
Odessa	46°29'N	30°38'E
Ponta Delgada	37°45'N	25°40'W
Lisboa	38°43'N	9°09'W
Palma	39°35'N	2°41'E
Tunis	36°50'N	10°14'E
Athens	37°58'N	23°43'E
Nicosia	35°09'N	33°17'E

The basic material for these investigations was yielded by monthly mean temperatures of the stations contained in *Table I* observed during the period 1901—1960. In this way, the stochastic variables ξ_{ik} were available in the from of a matrix of the type $m \times n$, with

$$i = 1, 2, \dots 60$$

and

$$k = 1, 2, \dots 12$$

From the stochastic variables ξ_{ik} we derived the new stochastic variables ξ'_{ik} possessing only two values, namely $+1$ and -1. Designing by M_k the *median* of the distribution function of the stochastic variables situated in a given column of the matrix (that is, of the stochastic variables for a given month) we used in the course of the transformation the following values;

$$\begin{aligned} \xi'_{ik} &= 1 \qquad \text{when} \qquad \xi_{ik} \geqq M_k \\ \xi'_{ik} &= -1 \qquad \text{when} \qquad \xi_{ik} < M_k \end{aligned} \tag{1}$$

(corresponding to positive and negative anomalies, respectively).

It should be noted that this definition of the anomalies is at variance to that conventionally used in climatology, as the deviations are measured not from the arithmetic mean value. However, from the point of view of probability calculus, the use of the definition given under (1) is a more advantageous one, as in this way, the distorting effect of asymmetry is eliminated, and, as a consequence, both the positive and the negative anomalies are possessing one and the same probability of 0,5.

By forming the anomaly time series according to (1), we determined, for every consecutive pair of months k and $k+1$ the probabilities

$$P(+,+) \qquad \text{and} \qquad P(-,-)$$

that is, the probabilities for the occurrence of two subsequent positive and for the occurrence of two subsequent negative anomalies, respectively. Let us design the probability $P(+,+)$ by P_1 and the probability $P(-,-)$ by P_2. The probability of the conservation of the sign of anomalies is thus given by the sum P_1+P_2.

When assuming *independence* between the anomalies of subsequent months, and considering, that, according to (1),

$$P(+) = P(-) = 0{,}5$$

we have in this case

$$P_1 = P(+)\cdot P(+) = 0{,}25$$

$$P_2 = P(-)\cdot P(-) = 0{,}25$$

from which it is obtained, for the probability of sign conservation:

$$P_1+P_2 = 0{,}5$$

By selecting a confidence level of 95%, the assumption of independence should be, in the case of the data series under investigation, discarded when the empirically determined value of the sign conservation probability P_1+P_2 falls outside the following interval:

$$I = 0{,}5 \pm 1{,}96 \sqrt{\frac{0{,}25}{60}}$$

that is, when

$$P_1+P_2 > 0{,}627$$

or

$$P_1+P_2 > 0{,}373 \tag{2}$$

The conservation probabilities P_1+P_2 are contained in *Table II*. Let us consider the data collected in *Table* II. At first it may be stated that the conservation probability of the anomalies is not the same during the year, that is, a more or less

Table II

Conservation probabilities $P_1 + P_2$

	I—II	II—III	III—IV	IV—V	V—VI	VI—VII	VII—VIII	VIII IX	IX—X	X—XI	XI—XII	XII—I
Angmagssalik	0,63	0,60	0,53	0,60	0,57	0,55	0,67	0,65	0,55	0,55	0,43	0,58
Stykkisholm	0,48	0,63	0,50	0,58	0,63	0,67	0,67	0,55	0,65	0,43	0,55	0,56
Bergen	0,52	0,68	0,55	0,52	0,62	0,67	0,68	0,73	0,65	0,57	0,55	0,49
Bodo	0,48	0,68	0,62	0,53	0,55	0,65	0,63	0,57	0,63	0,58	0,62	0,58
Karasjok	0,50	0,55	0,62	0,55	0,58	0,53	0,57	0,53	0,60	0,50	0,60	0,68
Arkhangelsk	0,60	0,67	0,57	0,62	0,48	0,52	0,62	0,55	0,68	0,63	0,62	0,64
Valentia	0,63	0,63	0,78	0,67	0,52	0,63	0,58	0,63	0,58	0,53	0,63	0,62
Aberdeen	0,72	0,63	0,48	0,58	0,58	0,68	0,73	0,68	0,58	0,57	0,50	0,63
De Bilt	0,67	0,65	0,52	0,45	0,63	0,53	0,67	0,70	0,52	0,52	0,40	0,48
Berlin	0,58	0,63	0,60	0,60	0,60	0,48	0,62	0,60	0,72	0,47	0,50	0,54
Uppsala	0,67	0,67	0,57	0,65	0,65	0,60	0,70	0,62	0,62	0,58	0,53	0,54
Vilnius	0,62	0,60	0,65	0,60	0,63	0,55	0,58	0,65	0,68	0,57	0,47	0,63
Leningrad	0,60	0,65	0,63	0,62	0,57	0,50	0,72	0,57	0,67	0,62	0,53	0,63
Moscow	0,68	0,62	0,65	0,55	0,43	0,55	0,62	0,62	0,63	0,57	0,63	0,51
Okt. Gorodok	0,58	0,47	0,63	0,53	0,63	0,63	0,68	0,65	0,60	0,50	0,70	0,58
Paris	0,57	0,52	0,65	0,47	0,57	0,48	0,77	0,63	0,57	0,50	0,38	0,48
Marseille	0,60	0,62	0,52	0,58	0,63	0,60	0,65	0,70	0,57	0,43	0,48	0,58
Basel	0,63	0,53	0,63	0,47	0,58	0,55	0,77	0,55	0,60	0,53	0,40	0,58
Roma	0,63	0,65	0,52	0,55	0,65	0,67	0,70	0,58	0,53	0,65	0,48	0,54
Budapest	0,60	0,52	0,57	0’43	0,57	0,62	0,72	0,70	0,62	0,52	0,52	0,56
Sibin	0,67	0,52	0,53	0,50	0,53	0,57	0,63	0,68	0,52	0,60	0,53	0,61
Odessa	0,75	0,65	0,70	0,55	0,53	0,62	0,65	0,53	0,65	0,60	0,48	0,68
Ponta Delgada	0,58	0,50	0,47	0,65	0,65	0,62	0,70	0,68	0,65	0,48	0,48	0,66
Lisboa	0,57	0,55	0,67	0,65	0,53	0,53	0,55	0,58	0,58	0,63	0,50	0,54
Palma	0,68	0,60	0,63	0,57	0,45	0,63	0,55	0,63	0,53	0,77	0,65	0,53
Tunis	0,67	0,68	0,60	0,65	0,68	0,63	0,65	0,60	0,55	0,63	0,60	0,64
Athens	0,47	0,53	0,52	0,53	0,77	0,75	0,75	0,60	0,65	0,63	0,62	0,59
Nicosia	0,68	0,70	0,65	0,47	0,62	0,67	0,50	0,60	0,57	0,60	0,62	0,70

developed *annual variation* is experienced. The main characteristics of this annual variation are represented for the whole area under investigation, on *Fig. 1,* in which we have plotted the number of occurrence of the maximum and minimum values of the conservation probabilities on each of the 28 stations. The highest probability for the conservation of a temperature anomaly is occurring during midsummer from July to August, while the lowest one occurs in the transition period between autumn and winter, that is, from November to December.

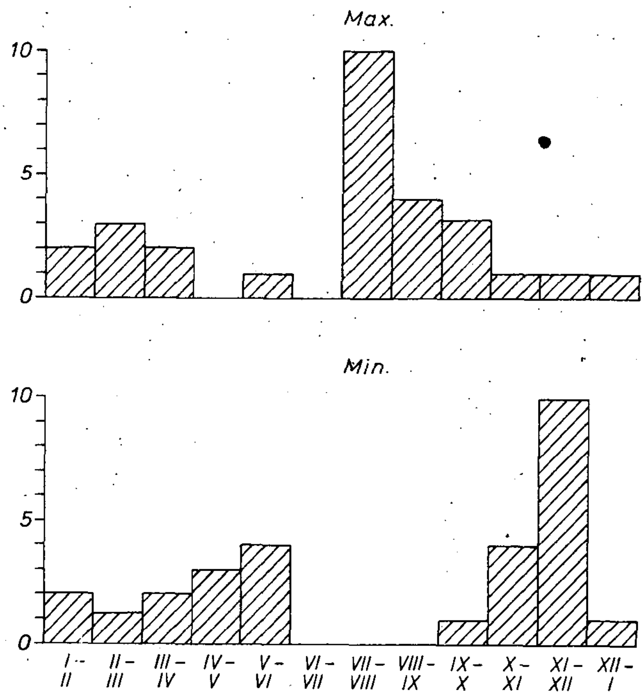

Fig. 1. Annual distribution of the maximum and minimum values of temperature anomalies
1. ábra. A hőmérsékleti anomáliák maximális és minimális értékű megmaradási valószínűségének éven belüli megoszlása

An essentially similar annual variation is exhibited by the frequency distribution presented on *Fig. 2,* which is yielding among the 28 stations the number of cases in which the value of the conservation probability was lying outside the limits of the independence interval according to (2), that is, it is yielding those months for which a statistical relation exists in respect to the signs of the anomalies. This annual variation is exhibiting, in agreement to the data presented on *Fig. 1,* a distinct accumulation of realistic relations in midsummer (July—August) and, in addition, on the end of the winter and the beginning of spring (February—March). Realistic relations are experienced rather seldom between the pairs of months November—December and April—May. Expressed in other words, this means that from the point of view of a prevision of the temperature anomaly of the subsequent month, July and February are yielding most information, while the least of information is obtained in April and November.

For obtaining a more detailed analysis of this interesting phenomenon, we are presenting the geographical distribution of the probability P_1+P_2 for the pairs of months July—August and February—March, respectively (*Fig. 3 and 4*). The higher conservation probability in midsummer is mainly characteristic for Western and Central Europe, where its value is exceeding 0,7 and on a smaller area, even 0,75. On the other hand, in the area of the Atlantic under investigation, and in Northern and Eastern Europe, lower probability values are encountered which are indi-

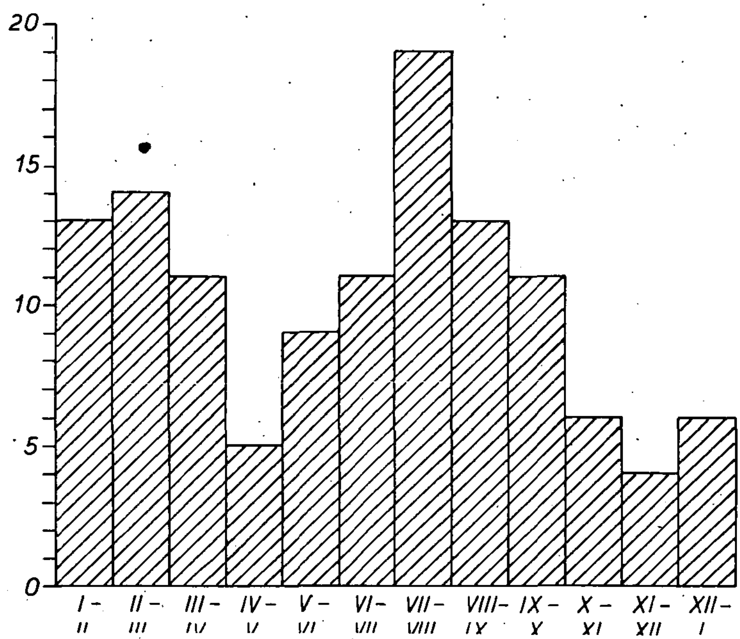

Fig. 2. Annual distribution of those values of conservation probabilities which are indicating the existence of a statistical relationship
2. ábra. Reális statisztikai kapcsolatot tükröző megmaradási valószínűségek éven belüli megoszlása

cating an independence of the subsequent anomalies. Thus it appears that the higher conservation probability of temperature anomalies in midsummer is primarily characteristic for the boundary regions between oceanic and continental climatic influences which is finally meaning that, in this area, a more prolonged persistence of both characteristics may occur. The exploration of the deeper causes of this phenomenon are affording further investigation. However the fact itself, as established here, should be taken intoconsideration during forecasting work. In the case of the second pair of months (February—March) an entirely different structure of the geographical distribution of conservation probabilities is exhibited. In this case, higher values of probability are appearing mainly in the North and, in our opinion, this is reflecting an effect of the disappearance of the winterly snow-cover. When, in the course of a winter, big snow masses are accumulated, then the melting of these snow masses is absorbing important quantities of energy, which is favourable for a prolonged subsistence of negative anomalies. At the same time, however, it is of interest, that even in the Mediterranean area, there is a higher probability for

VII. ⟶ VIII.

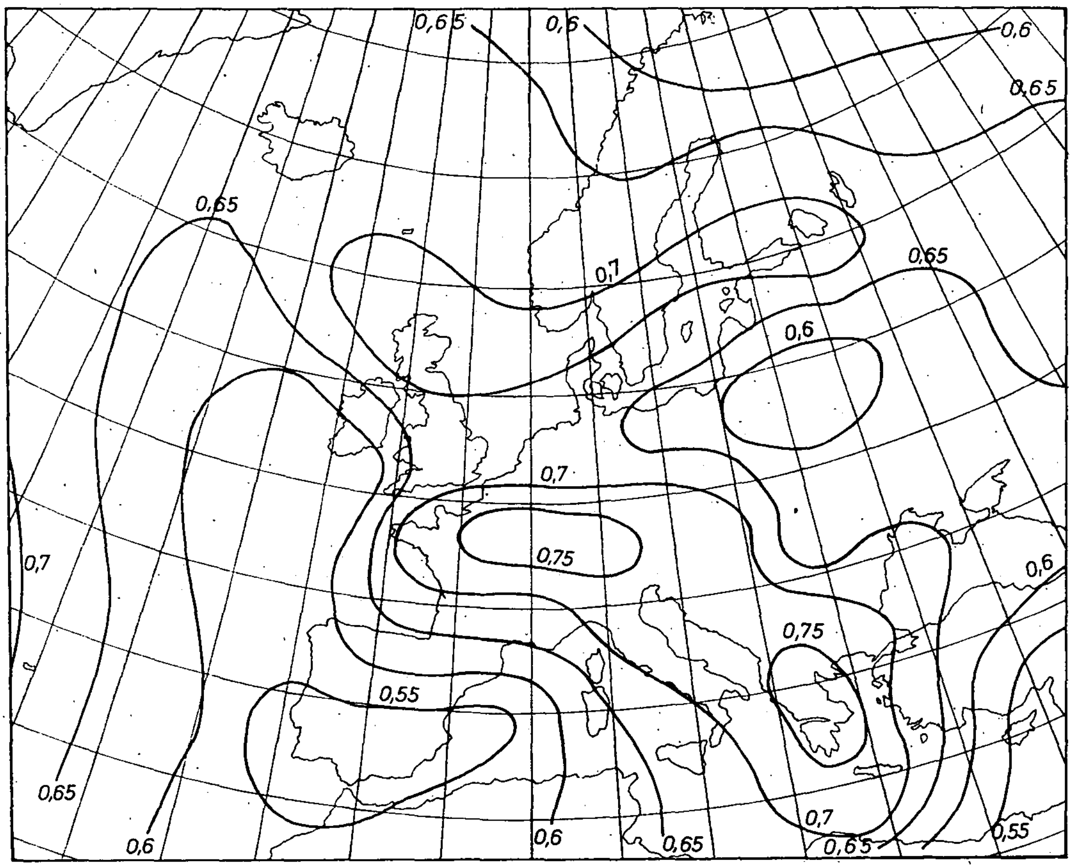

Fig. 3. Conservation probability of temperature anomalies for the pair of months July—August

3. ábra A hőmérsékleti anomáliák megmaradási valószínűsége júliusról augusztusra

the conservation of temperature anomalies, which is caused assumedly by the circumstance that the slower warming and cooling of water masses is equally favourable for the prolonged substistence of temperature anomalies having the same sing.

In the following we are dealing with the characteristics of the annual variation of conservation probabilities. As, already indicated, from *Table II* it appears that the conservation probability is possessing, on the majority of the stations under investigation, a characteristical annual variation. For solving the problem, whether this variation is in contradiction to the assumption of a uniform annual distribution of the probability (that is, whether the appearing annual variation is a realistic one), the following procedure may be used.

Let us determine the arithmetic mean value O_n of the conservation probabilities obtained for the 12 pairs of months. The maximum and minimum values of these probabilities will be designed by Q_{max} and Q_{min}. Let us consider the confidence interval of Q_m on the 95% level:

II. ⟶ III.

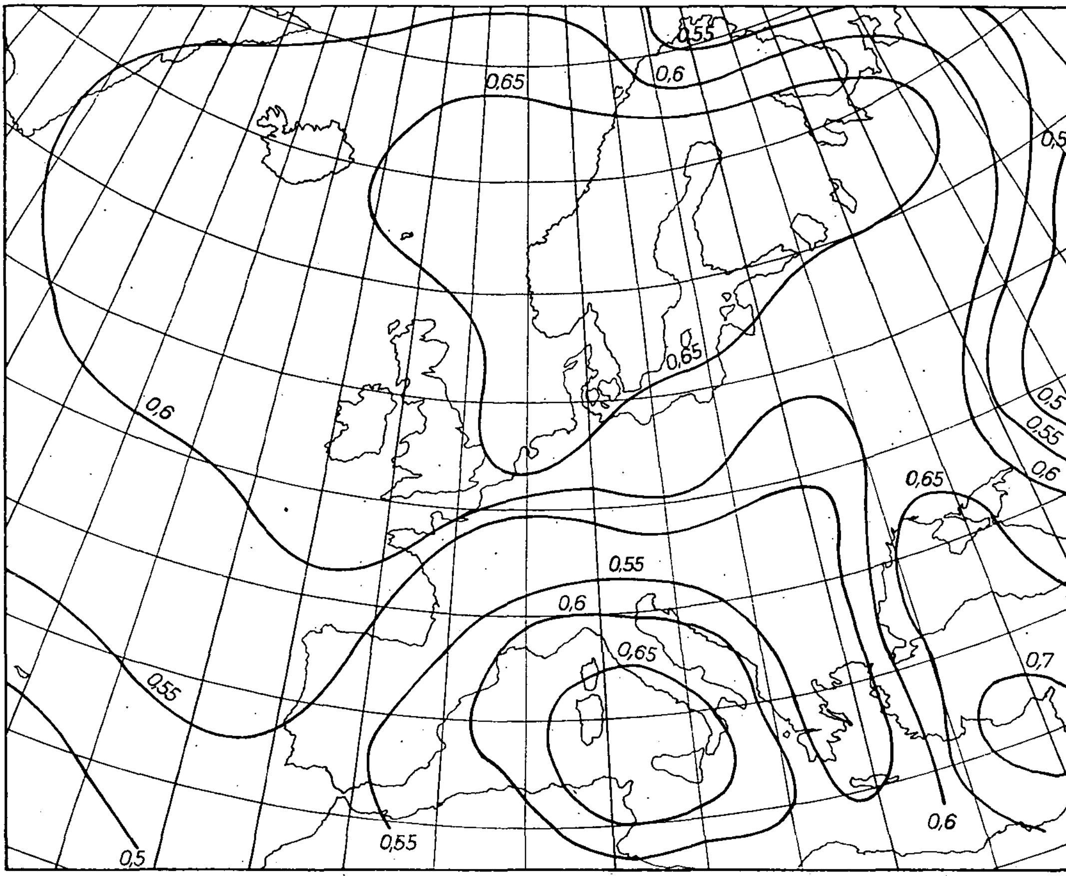

Fig. 4. Conservation probability of temperature anomalies for the pair of months February—March
4. ábra. A hőmérsékleti anomáliák megmaradási valószínűsége februárról márciusra

$$I = Q_m \pm 1{,}96 \sqrt{\frac{Q_m(1-Q_m)}{60}}$$

An annual variation is considered to be a realistic one, when the following conditions are fulfilled

$$\begin{aligned} Q_{\max} &> Q_m + 1{,}95 \sqrt{\frac{Q_m(1-Q_m)}{60}} \\ Q_{\min} &< Q_m - 1{,}95 \sqrt{\frac{Q_m(1-Q_m)}{60}} \end{aligned} \qquad (3)$$

A realistic annual variation of the conservation probabilities of temperature anomalies on the basis of the conditions (3) can be demonstrated for the following stations: De Bilt, Paris, Basel, Budapest, Odessa, Palma, Athens. We are separately

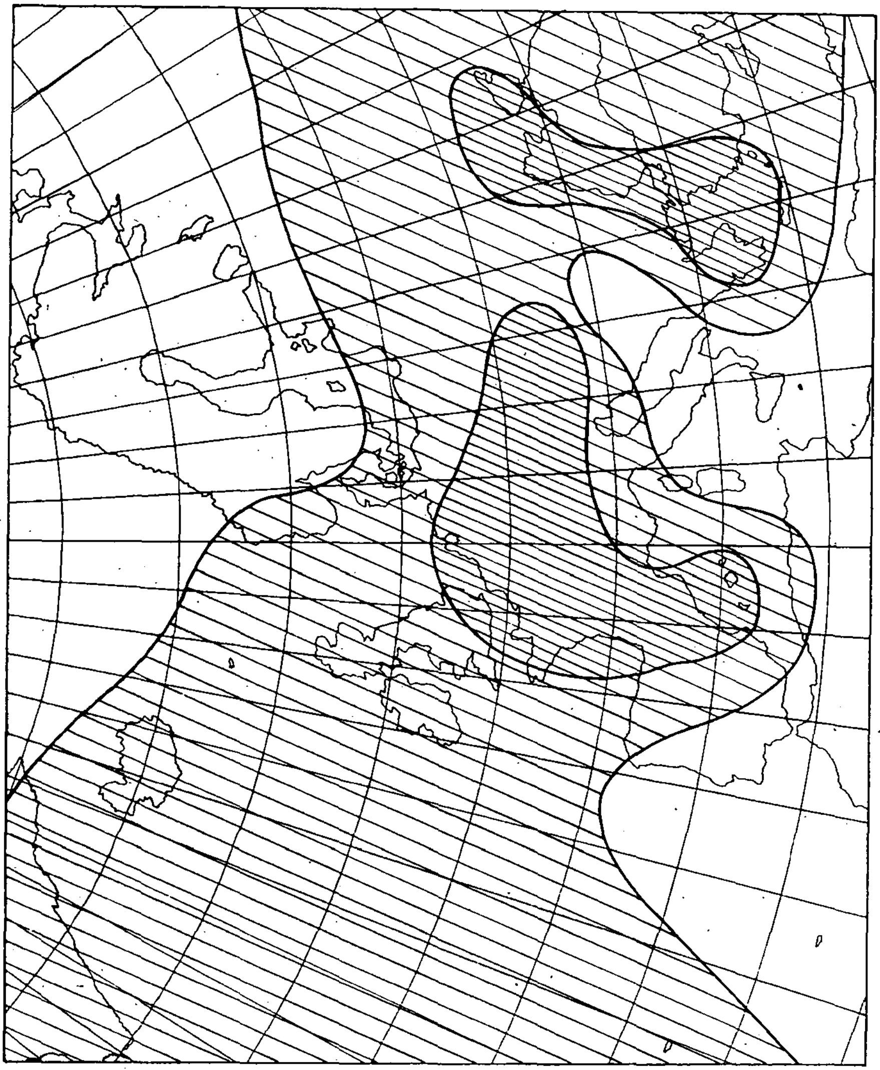

Fig. 5. Distribution of areas where there exists a realistic annual variation of the conservation probabilities of monthly temperature anomalies
5. ábra. Azon területek, eloszlása, ahol a havi hőmérsékleti anomáliák megmaradási valószínűségének reális évi menete fennáll

considering annual variations, for which only one of the conditions (3) is fulfilled. Such annual variations are found on the following stations: Angmagssalik, Stykkisholmur, Bergen, Valentia, Aberdeen, Berlin, Vilnius, Moscow, Okt. Gorodok, Marseille, Ponta Delgada, Nicosia. Separating the area represented by the above stations, we are obtaining a picture like that of *Fig. 5.* On this figure, we represented by a heavy shadowing the areas for which a realistic annual variation of the conservation probabilities is found on the basis of the conditions (3), while the areas are separated by a light shadowing where (3) is fulfilled only for Q_{max} or only for Q_{min}.

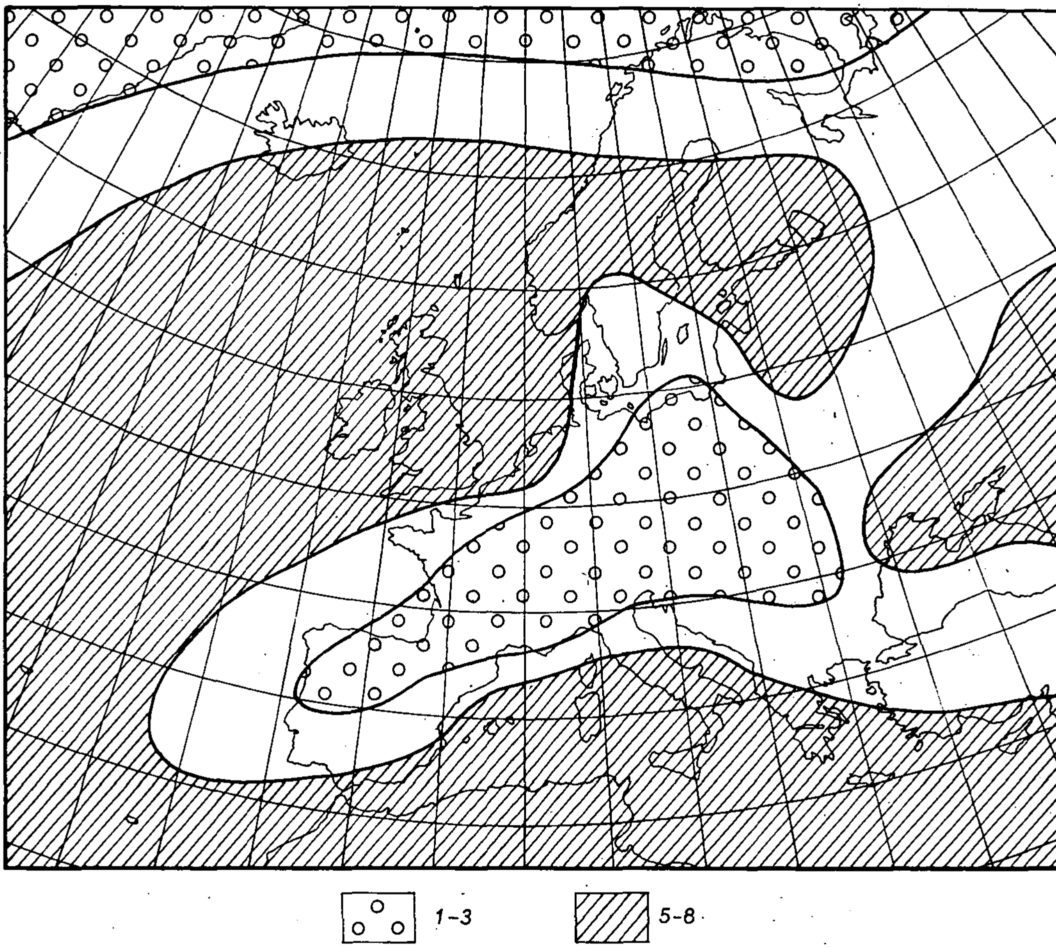

Fig. 6. Number of the pairs of months which are indicating the existence of a realistic statistical relationship between temperature anomalies
6. ábra. A hőmérsékleti anomáliák reális statisztikai kapcsolatára utaló hónappárok száma

The characteristic annual variation of conservation probabilities is primarily a feature of the middle-latitude regions of the European mainland, principally of Western and Central Europe and further it can be demonstrated in the south-eastern part of the Balkan peninsula and in the area of the Black Sea. These regions can be evaluated, from the point of view of the conservation of monthly temperature anomalies, as areas in which the anomaly of a previous month is yielding, in certain parts of the year, important information about the thermal character of the subsequent month (as seen before, this part of the year is, for Western and Central Europe, the midsummer).

In the following we are investigating the number of such pairs of months on the various stations, for which the conservation probability of the signs of temperature anomalies is indicating, on the basis of the criterion (2), the existence of a reâlistic statistical relation, that is, for which

$$P_1+P_2>0{,}627$$

The distribution of the numbers of such pairs of months is illustrated on *Fig. 6.* The number of the pairs of months indicating a statistical relation is lowest in Central Europe as well as in the area beyond the Arctic Circle (1 to 3), while in the central parts of the Atlantic, in the area of the British Isles, and in the Mediterranean basin, a much more higher frequency (5 to 8 pairs of months) is found. No doubt, this frequency distribution is indicating, that a higher persistence of temperature anomalies is mainly determined by the *distribution of oceanic and continental areas,* as a consequence of the well-known differences in the thermal balance of these two kinds of surfaces. However, it is very likely, that the characteristical band-like structure of the distribution of conservation probabilities which we have demonstrated is also connected to the characteristical types of the spatial distribution of montly temperature anomalies as discussed under [1], that is it may be also connected to circulatory causes.

Fig. 6. may be further evaluated from the point of view of long-range forecasting. As mentioned above, in the long-range forecasting of temperature anomalies, it is necessary to use, in addition to the dynamical-synoptical methods, also some statistical procedures. The figure is containing a warning, that the applicability of these methods may be, in the area of Central Europe, a much more limited one than in other areas of the world.

Reference

[1] *Péczely, G.:* A hőmérséklet havi anomáliáinak megmaradási hajlama az északatlanti-európai térségben (Tendency of persistence of monthly temperature anomalies in the area of the Northern Atlantic and of Europe). Időjárás, 80, 5. 267—273 pp.

Investigations on the Distribution of Days Required to Attain Cumulative Precipitation Amounts

by

I. Herendi

Kumulatív csapadékösszegek eléréséhez szükséges napok eloszlásvizsgálata. A gyakorlati munkában gyakran szükséges az adott időn belül várható csapadék előrejelzése. A dolgozat egy módszert mutat be a statisztikai becslésre. Négy csapadékmérő állomás adatai alapján, az ún. csapadéktelítődési függvény bevezetésével a szerző a kumulatív csapadékösszegek eléréséhez szükséges napok eloszlásfüggvényeit vizsgálta. Az adatsorokra vonatkozó statisztikai próbák (függetlenség-, homogenitásvizsgálat) elvégzése után a *Gamma*-eloszlás paramétereinek kiszámítása, értékelése következett. Ez elegendő volt egyetlen állomáson (és kis környezetében) az előrejelzéshez. A területi általánosításhoz át kellett térni a *Pearson-III* típusú eloszlás alkalmazására. Így a módszer alkalmas lett nagyobb földrajzi egységekre vonatkozó statisztikai becslésekre.

Verteilungsuntersuchang der zum Erreichen von kumulativen Niederschlagsmengen erforderlichen Tage. In der praktischen Arbeit taucht oft die Notwendingkeit der Vorhersage eines innerhalb einer gegebenen Zeitdauer herabfallenden Niederschlagsmenge auf. In der vorliegenden Arbeit wird eine Methode zur statistischen Schätzung beschrieben. Auf Grund der Angaben von vier Niederschlagsstationen und mit der Einführung der sog. Funktion der Niederschlagssättigung werden vom Verfasser die Niederschlagsverteilungsfunktionen der zum Erreichen von kumulativen Niederschlagsmengen erforderlichen Tage untersucht. Nach der Ausführung der statistischen Proben bezüglich der Datenreihen (Untersuchung der Unabhängigkeit und Homogeneität) wurde die Errechnung der Parameter der *Gamma*-Verteilung, sowie ihre Auswertung vorgenommen. Dies war genegend auf bloss einer Station (und ihrer Umgebung) die Vorhersage ausführen zu können. Zur territorialen Verallgemeinerung musste auf die Anwendung der Verteilung vom Typ *Pearson-III* übergegangen werden. In dieser Weise wurde die Methode auch zu statistischen Schätzungen grösserer geographischen Einheiten geeignet.

In the practical work the forecasting of the precipitation to be expected within a given time is often required. In the present paper a method of the statistical estimation is described. On the basis of the precipitation data measured by four stations and by introducing the so-called function of precipitation-saturation the author gives an analysis of the distribution-functions of the days required for attaining the cumulative precipitation amounts. After the statistical checking of the data-series (analysis of independence and of homogeneity) the computation, evaluation of the parameters of the *Gamma*-distribution was carried out. This was sufficient for making the forecast for a single station (and its close environment). For the areal generalization the distribution Type *Pearson-III* had to be applied. Thus the method became suitable for statistical estimations concerning larger geographical units.

1. Introduction

Both the meteorologist and the geographer, when analysing complex processes, give particular attention to the precipitation from among the different parameters.

In many cases it would be desirable to know in advance the precipitation amount to be expected for a given area at a given fixed time.

This practical requirement gave rise to the idea of elaborating a method enabling the specialist to determine, within certain limits of exactitude, the probable precipitation amount at a given geographical point within a certain time, or to state the probable precipitation income within a concrete limit of probability.

It was envisaged also to make the method to be elaborated applicable also for the general evaluation of well measurable geographical parameters (that can be expressed by data series). The aim of the method was also to give assistance in the analysis and, later, in the synthesis, when dealing with complex processes.

2. Objectives

Anwers are to be found to the following questions:

1. Which is the percental probability of a precipitation amount exceeding, within a fixed time, a value determined in advance?
2. Which precipitation amount can be expected with a given probability within a given period?
3. After how many days is a given precipitation amount to be expected with a given probability?
4. How can generalization in time be made for an optional day of the year but for the same place?
5. How is the areal generalization to be solved?

3. Selection of the observation places

Serious problems arose when selecting the stations disposing of suitable data series. For the present there are more than 700 precipitation observation points in Hungary but comparatively few of them dispose of time series of the completeness and length required for our purposes.

From among the stations Budapest, Szeged, Keszthely and Eger proved to be the most appropriate. By the more than 100 years old daily data series of the above station a good representation of the climatic peculiarities of our geographic regions can be obtained facilitating also the areal generalization (*s. Fig. 1*).

4. Mathematical bases

Our task is now to determine the probability of the excession of a certain precipitation amount at one and later at several points of the space, and also the connection

$$y = f(t, p(t))$$

of the above and the time [3].

The trend of precipitation values is the result of a stochastic process. Under a stochastic process the one parameter assembly of probability variables ξ_t is meant, where the parameter t runs through a — generally — T set and is denoted with

$$\xi_t,\ t \in T$$

If the variable ω (elementary occurrence) of the essentially two-variable-function $\xi_t = \xi_t(\omega)$ (t and ω) is fixed and t runs through set T, than a real function, a realiza

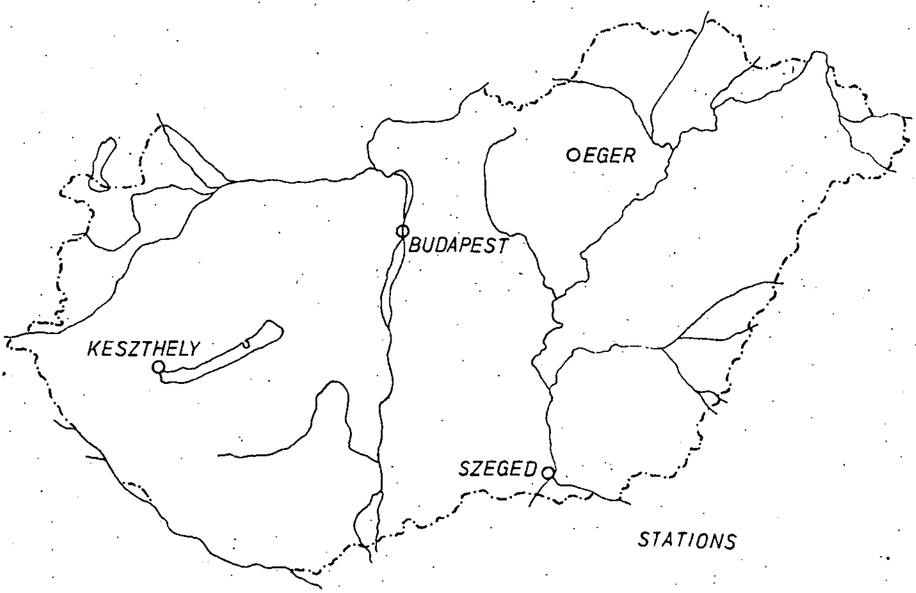

Fig. 1. The situation of the investigated stations
1. ábra A vizsgálatba bevont állomások elhelyezkedése

tion of the stochastic process will take place. One realization will characterize a concrete course of the process.

The measured values of the precipitation are concrete values of the probability-variable ξ_t.

The examined phenomenon, the precipitation, can be solved with discrete data series. From a different aspect our probability-variable is limited from below (it cannot assume negative values).

The distribution of the probability variable is unknown. As to its properties only inferences can be made from the finite data series called "model" permitting also computations, statistical evaluations concerning the attributes of the basic set.

In order to make inferences from our models as to the distribution of the basic set the models have to meet two basic conditions: they must be (with a good approximation) independent and homogeneous.

A most utilizable attribute of the precipitation data is that their function of distribution can be readily evaluated on the basis of models consisting of discrete data. This facilitates also the use and application of the method, to be presented by us.

Under the distribution-function of a probability variable the following is meant "per definitionem":

$$F(x) = P(\xi < x), \quad -\infty < x < +\infty$$

The basic features of the distribution functions are the following:
monotonously increasing:

$$F(x_1) < F(x_2), \quad \text{if} \quad x_1 < x_2$$

continuous from the left:

$$\lim_{x \to x_0} F(x) = F(x_0)$$

and also:

$$\lim_{x \to -\infty} F(x) = 0$$

and

$$\lim_{x \to +\infty} F(x) = 1$$

The distribution function $F(x)$ describing the distribution of the basic set is estimated with the empirical distribution function $F_n(x)$ computed from the model, if in the case of model $\xi_1, \xi_2, \ldots, \xi_n$

$$F_n(x) = \begin{cases} 0 & \text{if} \quad x \leqq \xi_1 \\ \dfrac{k}{n} & \text{if} \quad \xi_{k-1} \leqq x < \xi_n \\ 1 & \text{if} \quad \xi_n \leqq x \end{cases}$$

The distribution functions are to be ranged among the most frequently occurring classes of distribution functions. Thus the originally infinite set can be considerably narrowed down to a subset of the utilized types of distribution. The reason of the occurrence of the distribution types in nature is supported by several theorems of probability calculation — for instance the central limit distribution theorems.

On the basis of a finite model the determination of the type of $F(x)$ is facilitated when drawing the empirical distribution function of the model, i.e. $F_n(x)$ and comparing the curve with those of the more known and more manageable distributions. Selecting from among them the most appropriate one we have to determine, by the aid of the model elements, the form and parameters of the function. The empirical and theoretical distribution functions can be brought even more exactly together by the use of the methods of mathematical adaptation investigations.

5.1 The applied methods

The initial series, used for the computations, were the 100 years daily precipitation series of the selected stations. From these were produced the values determining the so-called "function of precipitation saturation". This is a step-function and such ones were attached to every fifth day of the year and to 5 precipitation amounts determined in advance. Thus the concrete values of these step-functions will yield, for the initial days being in a distance of five days one from the other, the number of the days within which 20. 30, 50, 70, 100 mm will fall down. (E.g.; the first initial day is the lst of January, the second one the 6th, the third one the 11 January etc.)

So at each station five figures were attached to every fifth day of the year. This means also that the 36 500 data collected from each station were transformed to further 36 500 data.

From the series produced by transformation we selected and ordered the number of excess days (or saturation values) of 100 different years but belonging to the same initial days and to the same cumulative amounts.

For these ordered series some simple statistical parameters too, have been computed which were helpful in our further work.

5.2 The investigation of independence

In the foregoings suppositions were made concerning our models. Their fulfilment is now to be checked.

The independence of the model elements means the following:

The probability variables ξ and η are called independent, if in the case of optional figures $a \leqq \xi \leqq b$ and $c \leqq \eta \leqq d$ the following equality is accomplished:

$$P(a \leqq \xi \leqq b,\ c \leqq \eta \leqq d) = P(a \leqq \xi \leqq b).\ P(c \leqq \eta \leqq d).$$

This is equivalent with the condition according to which:

$$H(x, y) = F(x) \cdot G(y),$$

where on the left side we have the joint distribution function of the probability variables ξ and η, and while $F(x)$ and $G(y)$, appearing on the right side, are the distribution functions of the random variables ξ and η.

The independence investigations were carried out with the method of *Wald-Wolfowitz.* [5]

According to the theorem functioning as the basis of the control:

In the case of an independent sample of a large number of elements, if the elements arise from an identical distribution, the sum,

$$R = \sum_{i=1}^{n-1} \xi_i \cdot \xi_{i+1} + \xi_1 \cdot \xi_n$$

formed from it will be, with a good approximation, of normal distribution.

Its expectable value is:

$$M(R) = \frac{S_1^2 - S_2}{n-1}.$$

Its variance

where n = the number of element of the sample

$\xi_i =$ the i-eth element of the sample arranged in the order of observation

$$S_i = \sum_{j=1}^{n} \xi_j^i.$$

For the control of the independence the distribution of $|R|$ is employed, and after that the required percentual probabilities are computed on the basis of the formula

$$p\% = 2 \cdot (100 - f(X))$$

where

$$|Y = \frac{R - M(R)}{D(R)}.$$

The "sample of large number of element" requires that $n \geqq 30$. Since in our case $n = 100$, this condition is fulfilled.

5.3 Investigation of homogeneity

In the foregoings reference was made to our hypothesis of the evenness of our samples. This too, is a most important criterion, since, although in our precipitation data series and at least on the examined places, no important change of climate could be proved, but e.g., a change of the surroundings of the observation place or of the type of the instrument could lead to inhomogeneities in our data. This investigation may appear as an important aspect also in evaluating other parameters.

The examination of homogeneity is based on the theorem of *Smirnov*. In this theorem it is stated that if two samples (with the element-nombers k and j) originate from a basic set of identical distribution, and they are independent from each another, so, when looking for the maximum absolute value of the differences ($d_{k,j}$) of the differences between the empirical distribution functions formed from them, the

$$Z = \sqrt{\frac{k+j}{k \cdot j}} \cdot d_{k,j}$$

product is, with a good approximation, the probability variable of *Kolmogorov*, in the case of $k, j > 30$.

The samples examined by us have been disjoined into such part-samples, and these investigations have been carried out.

The $L(Z)$ probability of the obtained values Z can be found in the standard tabulation of the *Kolmogorov*-distribution. The percentile probabilities, characteristic for the homogeneity, are to be computed from the formula

$$p\% = 100(1-L(Z))$$

5.4 The use of the distribution functions

On the basis of the initial concepts, in meteorology and water economics we choose the *Gamma*-distribution with the distribution function

$$F(x) = \frac{\lambda^k}{\Gamma(k)} \int_{x_0}^{x} t^{k-1} \cdot e^{-(t-x_0)} \, \mathrm{dt}.$$

Hypothetically it is supposed that the sample originates from the basic set of the theoretical distribution $F(x)$.

The curves of $F(x)$ and $F_n(x)$ will not correspond in all points but the deviations and their place within the domain of interpretation will be characteristical of the fitting.

Our initial hypothesis was that the two distributions are identical. This is the so-called zero-hypothesis. The question to be solved is whether or not this hypothesis is correct at the given significance level.

The hypothesis is to be rejected if in some part of the domain of interpretation the deviations are too large and unidirectional. If the hypothesis can be kapt, the deviations can be considered as random fluctuations of the sample.

The distribution-function of the *Gamma*-distribution has two parameters: the formal parameter "k" and the scale-parameter "λ". In its general form x_0 *too,* is a parameter, but in our case $x_0 = 0$, since that number of days is required to the precipitation of zero-mm.

For the estimation of the parameters "k" and "*lambda*" of the distribution function two methods can be used:

a) in the case of the so-called maximum likelihood:

$$k = \frac{1}{4 \cdot A}\left(1 + \sqrt{1 + \frac{4 \cdot A}{3}}\right)$$

$$\lambda = \frac{k}{\bar{x}}$$

where:

$$A = \ln \bar{x}$$

$$\bar{x} = \sum_{i=1} x_i/n$$

b) in the case of the method of moments:

$$k = \frac{m_1^2}{m_2^*}$$

$$\lambda = \frac{m_1}{m_2^*}$$

where m_1 is the value to be expected, while m_2^* is the second central moment or, as it is more often called, the variance. (It is customary to name the expected value also "first moment": hence the denomination of the method [4].

The question arises why our random variable (describing the precipitation data and being of continuous distribution) is treated discretely. This can be explained partly by the practical execution of the sampling, and partly by the circumstance that in the case of certain conditions the theoretical continuous distribution-function can be estimated very well with the discrete distribution.

6. Extension to any arbitrary day of the year and to arbitrary cumulative precipitation amounts

The original aim of the investigations was to find certain parameters for the statistical forecasting of the precipitation, and to extend these parameters in time and space to larger geographical units.

The processing has been made for the series of Budapest, Keszthely, Szeged and Eger. The transformation of the basic data, i.e. the production of the saturation values were carried out. After that, the produced samples were examined as to whether they satisfy the required, and from the aspect of the processing indispensable, conditions. The samples were found homogeneous, independent and well fitting (*Fig. 2; Tables 1—4*).

The analysis of the parameters (characteristic of the samples) has proved that the aspects fixed on the basis of preliminary considerations are good and suitable for the generalization (*Fig. 3*).

The results are already applicable to each fifth starting day and to five precipitation limits.

Now it is to be examined how the previous considerations could be generalized for an arbitrary day of the year and an arbitrary precipitation limit at a station or within its smaller surroundings.

Since the $[(n-1):5+1]$-eth day of the year has been chosen arbitrarily by us it seems obvious to condense the initial days (fixed when computing the saturation times) by changing the operandus "5". With an analogous procedure arbitrary precipitation limits can be established also instead of 20, 30, 50, 70 and 100 mm. But that would have meant a considerable increase of the time required to the processing.

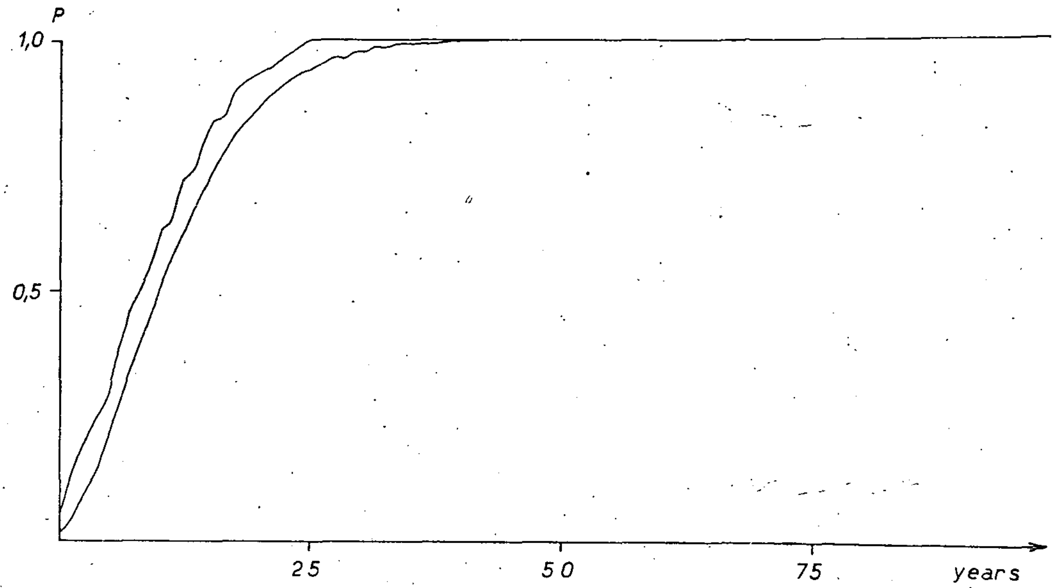

Fig. 2. Investigation of fitting. Budapest, lst starting day January 1. Fitting of distribution function belonging to the cumulative precipitation amount of 20 mm

2. ábra. Illeszkedésvizsgálat. Budapest, első kezdőnap január 1. 20 mm-es kumulatív csapadékösszeghez tartozó eloszlásfüggvény illeszkedése

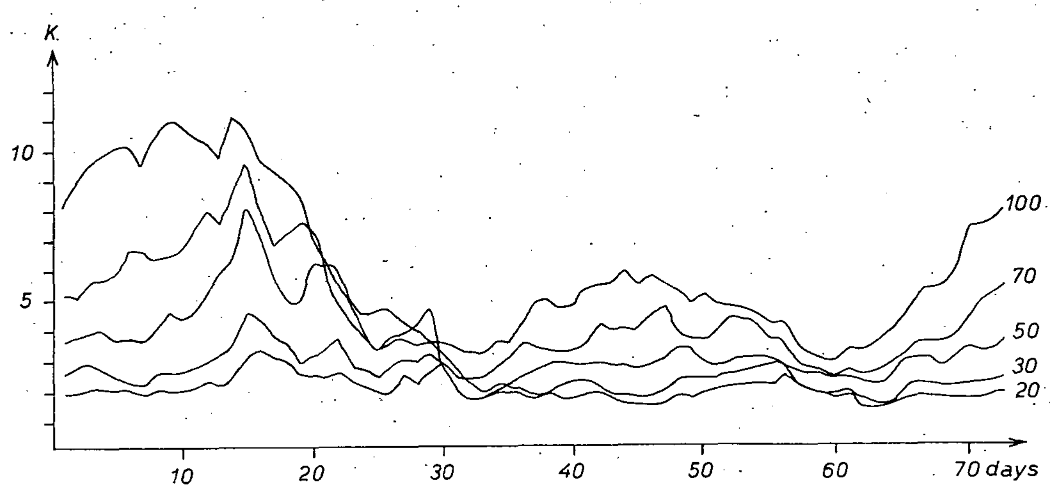

Fig. 3. Investigation of the parameters of the distributions. The „k" parameters of Budapest during the year, for 5 precipitation amounts

3. ábra. Az eloszlások paramétereinek vizsgálata. Budapest „k" paraméterei az év során, 5 csapadékösszeghez

Table 1

Investigation of fitting
Budapest, initial day 53, cumulative limit 20 mm empirical values

1.	0,03	0,05	0,10	0,11	0,12
2.	0,14	0,16	0,23	0,25	0,31
11.	0,32	0,34	0,43	0,48	0,53
16.	0,57	0,58	0,59	0,61	0,62
21.	0,65	0,67	0,69	0,70	0,73
26.	0,74	0,75	0,78	0,80	0,82
31.	0,84	0,85	0,86	0,88	0,89
36.	0,92	0,93	0,94	0,95	0,97
41.	0,98	0,99	1,00	1,00	1,00
46.	1,00	1,00	.	.	.
	.	.	.	.	.
91.	1,00	1,00	1,00	1,00	1,00
96.	1,00	1,00	1,00	1,00	1,00

Table 2

Investigation of fitting
Budapest, initial day 53. cumulative limit 20 mm computed values

1.	0,00	0,01	0,03	0,05	0,07
6.	0,09	0,13	0,16	0,19	0,22
11.	0,25	0,29	0,32	0,35	0,38
16.	0,41	0,44	0,47	0,50	0,53
21.	0,55	0,58	0,60	0,63	0,65
26.	0,67	0,69	0,71	0,73	0,74
31.	0,76	0,78	0,79	0,80	0,82
36.	0,83	0,84	0,85	0,86	0,87
41.	0,88	0,89	0,90	0,91	0,92
46.	0,92	0,93	0,93	0,94	0,94
51.	0,94	0,95	0,95	0,95	0,96
56.	0,96	0,96	0,96	0,97	0,97
61.	0,97	0,97	0,98	0,98	0,98
66.	0,98	0,98	0,98	0,98	0,99
71.	0,99	0,99	0,99	0,99	0,99
76.	0,99	0,99	0,99	0,99	0,99
81.	0,99	0,99	0,99	0,99	1,00
86.	1,00	1,00	1,00	1,00	1,00
91.	1,00	1,00	1,00	1,00	1,00
96.	1,00	1,00	1,00	1,00	1,00

Table 3

Investigation of homogeneity (Budapest)

Initial day	*Cumulative limit mm*	*Homogeneity %*
1	20	100,00
1	70	99,99
1	100	100,00
53	20	100,00
53	70	100,00
53	100	100,00

Table 4

Investigation of independence
(Budapest)

Initial day	Cumulative limit mm	Independence %
1.	20	82,58
1.	30	100,00
1.	50	96,80
1.	70	100,00
1.	100	95,22
53.	20	60,30
53.	30	79,48
53.	50	82,58
53.	70	93,62
53.	100	100,00

Table 5

Correlation coefficients (Szeged—Eger)
30. initial day
Correlation matrix, Szeged

	20 mm	30 mm	50 mm	70 mm	100 mm
20 mm	1,00	0,83	0,70	0,54	0,48
30 mm	0,85	1,00	0,81	0,63	0,57
50 mm	0,70	0,81	1,00	0,76	0,70
70 mm	0,54	0,63	0,76	1,00	0,86
100 mm	0,48	0,57	0,70	0,86	1,00

Correlation matrix, Eger

	20 mm	30 mm	50 mm	70 mm	100 mm
20 mm	1,00	0,85	0,65	0,60	0,53
30 mm	0,84	1,00	0,79	0,71	0,57
50 mm	0,66	0,79	1,00	0,92	0,78
70 mm	0,60	0,71	0,92	0,99	0,85
100 mm	0,53	0,57	0,78	0,85	1,00

Correlations between the corresponding limits (Szeged—Eger)

20 mm	0,39
30 mm	0,49
50 mm	0,55
70 mm	0,45
100 mm	0,35

In the case of a continuous parameter there is also a more serviceable method, but even then it must be considered whether it is worth while to apply a raster that is even more dense than a certain resolution.

On the basis of our experience it can be stated that two days, near to each other, e.g. two subsequent days, may be in close connection from the aspect of the precipitation. It is comparatively easy — but supportable also with computations — to suppose that there is a linear connection between the subsequent days and their respective saturation-, cumulation-values, or more exactly: between the distributions of the saturation values belonging to the days following each other rather densely.

So the linear connection has been supposed. To prove this, a new statistical method the correlation computation was applied (*Tables 3 and 5*).

In the examined cases comparatively high values with positive signs were obtained, and on the basis of the significance investigations it may be stated that according to the linear character and density of the connections the extension for time and quantities can be solved by means of the linear interpolation between the data. And this brings not only a gain in the time but facilitates also the practical application of the method. Thus if we are looking for a value for a given distribution in our tables as to an arbitrary initial day of the year or to arbitrary cumulative amounts, then we have to linearly interpolate between the corresponding data of two, already existing neighbouring distribution.

7. Three examples for the application of the method

Thus on the basis of the applied methods, considerations and computations we are already in a position to give the answer to the questions relating only to *one* measuring point from among those enumerated in our program. Since the measuring points represent as a rule only restricted surroundings, the reliability of the answers will rapidly decrease in proportion with our moving away from the measuring point.

In the possession of the respective tables the following questions may be answered:

1. *What is the percentage of probability of a given precipitation amount within a given time?*

For that the number of the respective starting day and the respective limit must be looked for in the table.

Examining the ordered sample elements we will arrive to the value of the fixed time (i.e. days). The place where it will be found is the position index (POZ) and that shows the required probability on the basis of the following formula:

$$p\% = \frac{100 \cdot \mathrm{POZ}}{n}$$

where n means the number of years included in the elaboration.

2. *In how many days can be expected a given precipitation amount with a given probability?*

First of all the serial number of the initial day is to be determined.

The n-th initial day is the serial day

$$(n-1):5+1$$

of the calendar year.

In the table of the empirical distributions one has to find the required starting day with the required precipitation limit (cumulative precipitation amount). Such values are to be found e.g. in *Table 6*.

After that we will compute from the following proportion:

$$\frac{100}{\text{number of the sample elements}} = \frac{\text{given probability } p}{\text{serial number of the sample element}}$$

that is:

$$100:n = p:\mathrm{POZ}$$

From this:

$$\mathrm{POZ} = \frac{n \cdot p}{100}$$

and so the required value will be the number figuring at this place.

Our investigations were carried out for the data series representing $n=100$ years. Thus $n=100$. So in the tables at our disposal the measuring number of the given probability will be the serial number of the place where the number of the days required to the accumulation can be found.

Some difficulties arise merely from the correct setting of the initial day because in our tables only every fifth one has been fixed. However, as it has been shown, that can be helped by linear interpolation.

3. *A precipitation amount of how many millimeters can be expected within a given period and with a given probability?*

On the basis of the formula the actual starting day is to be looked for and after that we have to compute — if it is not given so — the number of days corresponding to the period and also the position corresponding to the given probability.

On the basis of that we select that sample from among the five ordered ones (precipitation limits of 20, 30, 50, 70 and 100 mm) in which at the so determined place the value, falling next to the obtained period, is found. That will be the required value.

8. On areal generalization

On the basis of the above considerations and computations the extension to an arbitrary day of the year can be considered as solved. A much more complicated problem is presented by the areal generalization.

By reason of the obtained results the application of the distribution functions seems to offer favourable possibilities.

The *Gamma*-distribution is determined by the parameters "k" and "*Lambda*". For producing them the expected value m_1 and the variance m_2 are needed, from which

$$\lambda = \frac{m_1}{m_2^*}$$

and

$$k = \frac{m_1^2}{m_2^*}$$

The computations have been carried out for four stations: Transdanubia is well represented by *Keszthely*, the Hungarian Lowland by *Szeged*, and the Northern Central Mountains (Északi Középhegység) by *Eger*.

The processing of the series of *Budapest* seemed to be serviceable because that station is situated at the meeting place of our mean geographical regions, and, on the other hand, almost all climatological-meteorological parameters measured there can be found here in well ordered and from many aspects in detail processed form. This is most of all from the aspect of the comparations of importance and use.

In addition to that we dispose of longer or shorter series concerning about 800 points in Hungary, or in other words: the monthly mean values of our presently investigated parameter (precipitation) of 800 geographical points are known to us.

As it can be seen, four stations are known to us in detail, while a certain smaller surrounding area of them — where the series of the measuring point can still be considered as representant — can be well characterized by the aid of our method.

Besides of the above we have at our disposal the nearly 800 measuring points but the fact is that either the series of these stations does not meet the statistical requirements of the sampling, or, apart from the monthly mean values we have possibly no other information whatever about them.

By this antagonism the following problems arise:

1. *How dense must be the network in order to allow a generalization of our method for our main regions and the whole country?*
2. *If at a given measuring station only the monthly means are known, how can the method be applied there and its smaller surrounding areas?*

8.1 How many stations are required at the minimum for the generalization for the whole country?

In order to give answer to the above question a detailed investigation of the basic data of the four selected stations, the series transformed from them, the different parameters, the distributions and the results obtained by the examination of fittings has been carried out (*Fig. 3, Table 6*).

Surprisingly many similarities were found during these investigations. We were mainly interested in the "attitude" of the parameters (describing these distributions) in the case of a common initial day or in that of a fixed precipitation limit.

The two parameters of the *Gamma*-distribution seemingly do not show any important variance: of course only in the case if out of the parameter-series of the four stations only the overlapping, values, i.e. those in identical position, are compared. The deviation is even in percentage not significant in the case of the compared data. The substantial conformity of the distribution parameters brought us to the idea of possibly construct parameter-series (valid for the entire territory of the country) from the values of "*k*" and "*Lambda*" previously computed for the investigated stations?

The checking has been made for five data series: for Budapest, Szeged, Eger, Keszthely, and for the arithmetical mean of the respective data of these four stations. In this way it was envisaged to represent Hungary by the fifth fictive station. The elements of this series was denoted by P_M.

In a formula the computing of the *i*-eth such parameter is the following:

$$P_M^{(i)} = \frac{P_{\mathrm{BP}}^{(i)} + P_{\mathrm{SZ}}^{(i)} + P_{\mathrm{K}}^{(i)} + P_{\mathrm{E}}^{(i)}}{4}.$$

By the checking computations, requiring a vast integration work, it has been proved that the indicated way means indeed the correct solution. The reproduced data series of the stations showed a good correspondence with the original ones and they cover each other very well. Thus, within a confidence- interval of some days, the individual stations may be substituted even with each other.

Table 6

Ordered saturation values, Szeged, 1. initial day

	20 mm	30 mm	50 mm	70 mm	100 mm
1.	4	4	12	19	28
2.	4	8	12	34	41
3.	4	8	12	33	41
4.	5	8	20	34	42
5.	5	8	23	35	45
6.	6	9	24	37	49
7.	7	10	25	37	50
8.	7	10	27	40	57
9.	7	11	27	40	57
10.	7	11	27	40	57
11.	7	12	29	40	57
12.	8	12	31	42	60
13.	8	14	31	42	61
14.	8	15	31	42	63
15.	8	15	32	43	63
16.	10	16	32	43	64
17.	10	16	33	43	64
18.	10	17	34	44	65
19.	10	17	34	44	65
20.	11	17	34	44	65
21.	11	17	35	45	66
22.	11	18	35	46	67
23.	11	18	36	47	67
24.	12	19	36	48	67
25.	12	19	36	49	69
26.	12	19	37	49	70
27.	13	19	37	50	70
28.	13	20	38	51	71
29.	13	21	38	51	73
30.	14	22	39	51	74
31.	14	22	39	51	76
32.	14	23	39	53	76
33.	14	24	39	53	79
34.	14	24	40	53	79
35.	15	24	41	53	81
36.	15	24	41	54	81
37.	15	24	41	59	83
38.	17	25	42	60	83
39.	17	25	42	60	85
40.	18	25	43	60	87
41.	18	25	43	60	87
42.	18	26	43	61	88
43.	18	26	44	62	89
44.	18	26	44	63	91
45.	18	27	44	63	92
46.	19	27	44	63	92
47.	19	27	45	66	93
48.	19	27	46	67	96
49.	20	30	47	69	97
50.	21	30	47	71	97
51.	21	30	48	71	98
52.	21	31	49	71	99
53.	22	32	50	72	99
54.	22	32	51	73	99
55.	23	33	52	75	99
56.	24	33	53	75	100
57.	24	33	53	77	101
58.	25	33	54	77	102

	20 mm	30 mm	50 mm	70 mm	100 mm
59.	25	33	55	78	102
60.	25	35	55	79	103
61.	26	35	55	79	104
62.	27	35	56	81	105
63.	27	35	57	82	105
64.	28	37	58	83	107
65.	28	37	58	85	107
66.	29	38	60	85	407
67.	29	40	61	86	108
68.	30	40	61	87	109
69.	30	40	62	87	110
70.	31	41	63	87	110
71.	31	41	68	88	112
72.	31	41	69	88	113
73.	32	44	71	89	113
74.	32	44	71	90	114
75.	32	45	73	93	115
76.	32	45	75	93	115
77.	33	47	76	93	116
78.	33	48	78	95	117
79.	34	48	79	96	118
80.	34	48	81	97	118
81.	34	50	83	97	118
82.	35	50	84	100	119
83.	35	52	85	100	119
84.	35	52	86	100	120
85.	39	55	86	101	124
86.	40	56	87	101	126
87.	42	65	89	102	128
88.	44	66	93	102	129
89.	65	67	95	106	131
90.	55	67	95	108	132
91.	55	69	97	111	133
92.	58	72	99	114	135
93.	59	74	99	118	135
94.	60	77	101	121	134
95.	64	78	102	123	135
96.	64	84	112	124	137
97.	72	90	115	125	144
98.	73	97	117	129	145
99.	76	107	122	132	150
100.	84	111	129	133	154

On account of the small areal variance the following considerations seem to be servicable both from the practical but even from the theoretical aspect:

1. The parameter-series of the detailedly investigated stations are to be considered as the series "k" and "λ" valid for the main geographical regions represented by the stations.
2. The values $P_M^{(i)}$, i.e. the mean values of "k" and "λ" are considered as characteristic distribution parameters valied for the whole country.

Thus we have determined the minimum station network density for our main regions, i.e. the distribution parameters required for the application of the method. However, it is also supposed that within smaller areas these investigated parameters may be possibly even more variable.

8.2 Territorial generalization with the knowledge of the monthly averages

To find an answer to this question is already considerably more difficult. It must be held in view that the mere knowledge of the monthly averages is not sufficient to the reproduction of the time series with the help of the parameters.

If the use of the *Gamma*-distribution is decided then a limit will be set up to the further computations by the requirement of the second central moment of the data in order to produce the parameters "k" and "λ". The variance is, however, not always accessible for us.

The following considerations were applied:

The parameters concerning the *Gamma*-distribution show, as it could be seen before, an extraordinarily small variance in the whole country. Even the data series of Transdanubia, having considerably more precipitation, and those of the Lowland (Keszthely — Eger) showing other characteristics, present only insignificant deviations. Thus on the basis of the preceding chapter stabile dataseries of "k" and "λ" were obtained satisfyingly characterizing the whole country.

Under point 5.4 and in connection with the distribution functions it was mentioned that several types of distribution functions occur in the practice and nature.

In the practice of hydrological work the distribution function of the type *Pearson-III* too, proved to be very good, in addition to the *Gamma*-distribution.

Accordingly, appropriately well elaborated tables are at our disposal in order to give assistance in the computations with it. Its greatest advantage is its generally more simple manageability in the practice, compared with the other types of distribution. In consequence of that, and in addition to its theoretical value, the practical specialists are first of all those who make a great use of it.

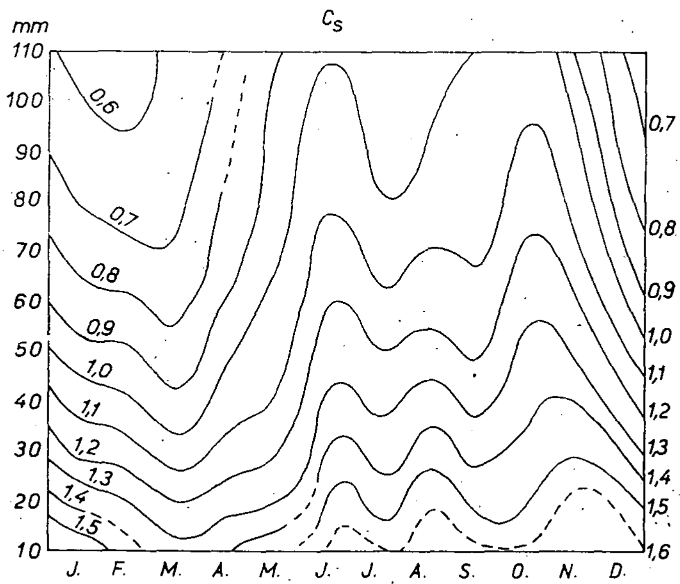

Fig. 4. Isoplethes of the values C_s of the distribution type Pearson-III
4. ábra. A Pearson-III eloszlás C_s értékeinek izoplétái

For our purposes the application of this type of distribution is essential because on the basis of the existing parameter-series one is enabled to easily switch over from the *Gamma*-distribution to the type *Pearson—III* which is rather similar to it.

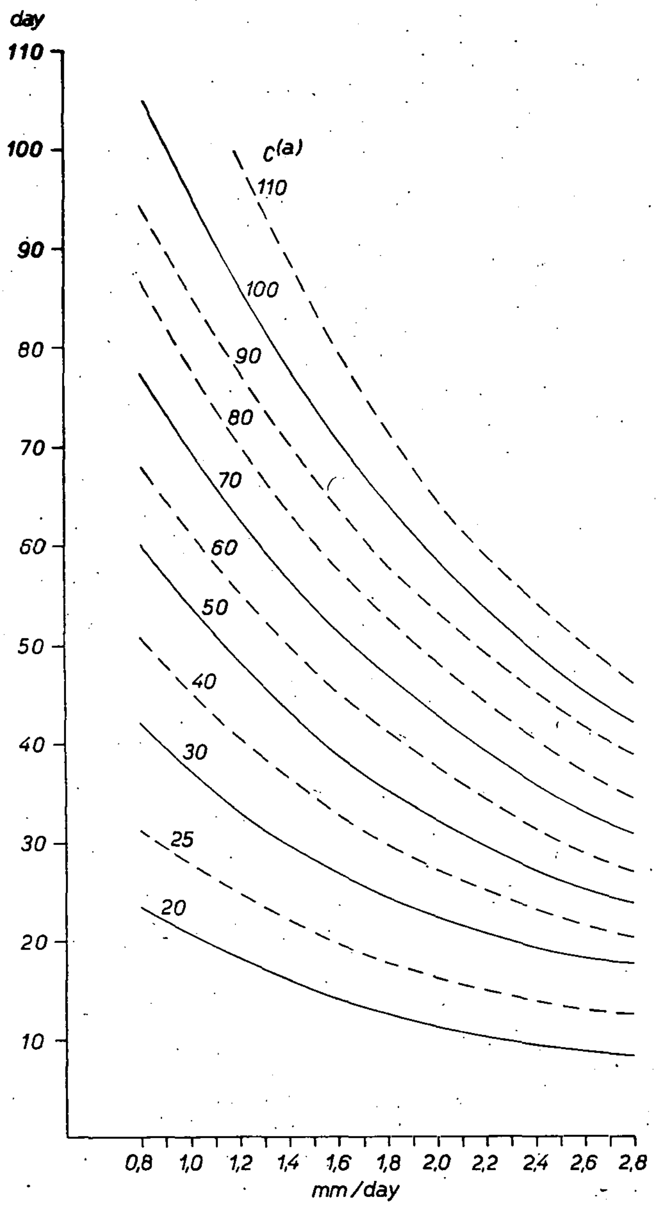

Fig. 5. The values $C^{(a)}$ of the distribution Pearson-III
5. ábra. A Pearson-III eloszlás $C^{(a)}$ értékei

The respective formulae are:

$$C_v = \frac{k}{m_1}$$

$$C_s = \frac{2 \cdot k}{m_1^3 \cdot C_v^3 \cdot \lambda^3}.$$

C_v is the variance factor, and C_s the asymmetry-factor of the *Pearson—III* distribution.

With the knowledge of the parameters the computation method of the saturation value $t_p^{(a)}$ belonging to the limit "a" and probability "p" is the following:

$$t_p^{(a)} = [(\Phi(C_s, p)) \cdot C_v + 1] \cdot M,$$

where (C_s, p) is the value of the Pearson—*III* distribution function in the case of an asymmetry factor C_s and a selected transgression probability p. The values (C_s, p) can be found in the *Foster-Ribkin* standard table [7].

When analyzing the time trend of the parameters it appeared that their formation shows a marked yearly tendency, so that their values ave been monthly averaged for the purpose to clearly see the characteristic form of the yearly trend by filtering out the random variations.

As an example the monthly averages of the parameters M, C_v and C_s belonging to $C^{(a)} = 50$ mm have been elaborated in detail for the station Szeged.

From among the parameters C_v and C_s do not show within a month considerable variations at the stations representing the different geographical regions of the cou-

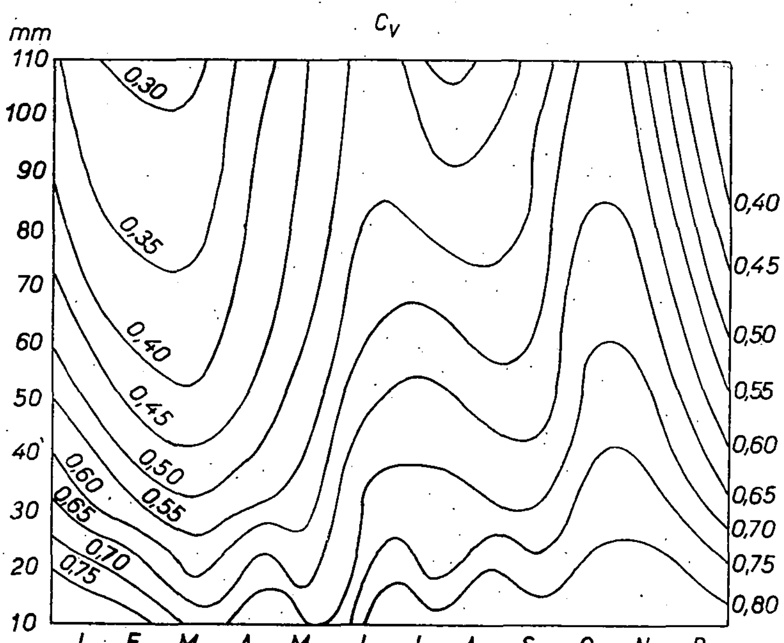

Fig. 6. Isoplethes of the values C_v of the Pearson-III distribution
6. ábra. A Pearson-III eloszlás C_v értékeinek izoplétái

try and so their arithmetical mean can be accepted as a characteristic standard value for the territory of the whole country. From the national averages of the parameters isoplethes were constructed (*Fig.-s 5 and 6*) from which not only the C_v and C_s values belonging to the selected precipitation amounts of 20, 30, 50, 70 and 100 mm but also those belonging to any intermediate precipitation amount $C^{(a)}$ can be found, because of their close connection with the amounts $C^{(a)}$.

The arithmetical mean M, on the other hand, shows considerable deviations according to the stations. However, the connection between M and K is obvious, where K is the mean precipitation amount of a given period after the time t_0, since the greater that average precipitation K the shorter time t will be required to reach the given cumulative precipitation amount $C^{(a)}$. For discovering the connection the most serviceable would be to know the average precipitation amount of periods of different durations after a certain given time datum. But such processings are not to our disposal, so only the known average precipitation amounts of the calendar months can be taken as a basis when analyzing the connection $M=f(K)$.

In our present investigations that the M values related to a given month N have been brought into connection with the average precipitation amounts of the months N, $N+(N+1)$, $N+(N+1)+(N+2)$. In order to eliminate the changing duration of the months, instead of the average precipitation amounts of K months only their part falling to 1 day have been taken into consideration, when carrying out the inves-

Table 7

Parameters of the distribution Pearson-III monthly
Eger, limit 30 mm

Month	Monthly average mm	Lambda	K	C_v	C_s
1.	30,70	0,07	2,22	0,69	1,34
2.	27,00	0,08	2,31	0,70	1,31
3.	22,40	0,13	2,99	0,59	1,15
4.	16,70	0,14	2,81	0,71	1,19
5.	16,20	0,17	2,45	0,56	1,27
6.	13,20	0,12	1,63	0,80	1,56
7.	16,20	0,12	2,00	0,72	1,41
8.	19,30	0,08	1,49	0,79	1,63
9.	22,50	0,08	1,83	0,75	1,47
10.	21,10	0,08	1,88	0,81	1,45
11.	21,20	0,07	1,43	0,80	1,67
12.	25,60	0,06	1,49	0,79	1,63

Keszthely, limit 20 mm

Month	Monthly average mm	Lambda	K	C_v	C_s
1.	25,90	0,08	2,26	0,72	1,33
2.	23,50	0,10	2,43	0,66	1,28
3.	20,80	0,12	2,52	0,63	1,25
4.	15,30	0,16	2,43	0,63	1,28
5.	13,50	0,15	2,07	0,71	1,39
6.	13,20	0,15	1,54	0,85	1,61
7.	14,00	0,15	2,08	0,68	1,38
8.	15,70	0,10	1,55	0,79	1,60
9.	16,40	0,10	1,72	0,74	1,52
10.	17,40	0,08	1,39	0,84	1,69
11.	22,30	0,08	1,88	0,76	1,45
12.	24,01	0,08	1,92	0,74	1,40

Budapest, limit 20 mm

Month	Monthly average mm	Lambda	K	C_v	C_s
1.	23,30	0,08	1,99	0,75	1,41
2.	23,20	0,09	2,11	0,69	1,37
3.	20,30	0,13	2,47	0,57	1,27
4.	15,70	0,15	2,37	0,65	1,29
5.	13,50	0,16	2,24	0,69	1,33
6.	15,00	0,12	1,79	0,74	1,49
7.	19,80	0,08	1,70	0,82	1,53
8.	21,10	0,07	1,51	0,83	1,62
9.	21,10	0,09	1,97	0,72	1,42
10.	18,50	0,10	1,96	0,75	1,42
11.	16,90	0,08	1,50	0,90	1,63
12.	19,01	0,08	1,65	0,84	1,55

Szeged, limit 20 mm

Month	Monthly average mm	Lambda	K	C_v	C_s
1.	26,30	0,08	2,27	0,71	1,32
2.	27,20	0,09	2,68	0,66	1,22
3.	22,70	0,14	3,28	0,56	1,10
4.	17,10	0,14	2,56	0,66	1,24
5.	14,50	0,18	2,62	0,62	1,23
6.	15,30	0,12	1,80	0,73	1,49
7.	19,70	0,10	2,02	0,72	1,40
8.	20,70	0,09	2,02	0,76	1,40
9.	22,30	0,07	1,74	0,84	1,51
10.	22,10	0,07	1,67	0;83	1,54
11.	19,40	0,09	1,93	0,79	1,48
12.	22,80	0,08	1,82	0,73	1'43

tigation of the functional connection. The mean monthly precipitation amounts of the examined stations — the mean values relate to the processed 100 years (1871—1970) — are contained in reference [2].

The analysis of the related values has shown that the best connection is yielded in the case of $C^{(a)}=20$ mm and $C^{(a)}=30$ mm with the precipitation amount of the month under consideration; in the case of $C^{(a)}=50$ mm and $C^{(a)}=70$ mm with that of the month under consideration and in the case of $C^{(a)}=100$ mm with the month under consideration and the two next following months, i.e. with the mean precipitation amounts of three months. The connection $M=f(K)$ for the given threshold value $C^{(a)}$ is shown by *Fig. 4.* On the horizontal axis the part falling to 1 day of the value K of the investigated period can be read while on the vertical axis the values of M.

So e.g. if we want to determine for an initial time t_0 in March and for the threshold value $C^{(a)}=70$ mm the arithmetical mean value M of a precipitation-saturation time $t^{(a)}$ concerning a station where the many years' average of the precipitation amount of May is 41 mm, and that of April 55 mm, then the value to be considered on the horizontal axis will be $(41+55):61=1{,}57$ and for that precipitation a mean saturation (accumulation) time of $M=52$ days is required.

For the control of the reliability of the approximative computations we selected from our processings at random the times t_0 for each of two stations, and to the empirical distribution functions of the precipitation-saturation time belonging to the two threshold values $C^{(a)}$ fitted the theoretical distribution functions computed with the parameters C_v, C_s, and M taken from the Figures 5 and 6.

The investigated empirical distribution functions are related to the cases

Szeged: t_0=January 1, $C^{(a)}=20$ mm;
Keszthely: t_0=July 15, $C^{(a)}=100$ mm .

After checking the fitting with the computation based on the *Kolmogorov* distribution function the values $p=0{,}11$ and $p=0{,}18$ have been obtained which are satisfying approximations since $p>0{,}05$ and thus the origin of the compared two distributions from an identical set should not be rejected.

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Relation between Nebulosity and Diurnal Temperature Amplitude

by

Á. Kiss

Az égbolt borultságának és a hőmérséklet napi ingásának kapcsolatáról. A tanulmány Magyarország északi, déli, nyugati és keleti szegélyéről egy-egy meteorológiai állomás 5 évi adatsorát használja fel. Minden hónapra meghatározza a lineáris regresszió egyenletét, a korrelációs együtthatót és egyéb paramétereket. Megállapítja, hogy a korrelációs együttható havonként és állomásonként —0,0 és —0,85 között ingadozik. A legerősebb korreláció szeptemberben és a nyári hónapokban mutatkozik. A regressziós együttható értéke 1° körül mozog, a reziduális szórás viszont 2°-nál is nagyobb.

Az azonos borultsággal számított napi ingás legnagyobb értékét a ténylegesen tapasztalt ingás havi átlagainak nyári és szeptemberi maximumával szemben májusban éri el.

A vizsgálatok eredményei arra vallanak, hogy a napi ingásnak a borultsággal való kapcsolata Magyarország különböző területein hasonló.

Über den Zusammanhang zwischen Bewölkung und täglicher Temperaturamplitude. In der Arbeit werden 5-jährige Beobachtungsreihen von vier meteorologischen Stationen verwendet, die je in den nördlichen, südlichen, westlichen und östlichen Grenzgebieten von Ungarn liegen. Für einen jeden Monat werden die linearen Regressionsgleichungen, die Korrelationskoeffizienten und manche andere statistische Parameter errechnet. Es wird festgestellt, dass der Korrelationskoeffizient nach Monaten und nach Beobachtungsstationen zwischen den Werten —0,50 und 0—85, schwankt. Die stärkste Korrelation zeigt sich im September und in den Sommermonaten. Der Wert der Regressionskoeffizienten schwankt um 1° herum, hingegen übertrifft die residuale Streuung selbst den Wert von 2°.

Der auf Grund der Annahme einer gleichbleibenden Bewölkung errechnete Wert der Tagesschwankung der Temperatur erreicht sein Maximum im Mai, im Gegensatz zu den beobachteten Werten, welche ihr Maximum in den Sommermonaten und im September erreichen.

Die Ergebnisse dieser Untersuchungen weisen darauf hin, dass der Zusammenhang zwischen der Tagesschwankung und der Bewölkung in den verschiedenen Gebieten Ungarns eine gleiche Struktur besitzen.

The paper is using 5-year data series from four meteorological stations situated respectively on the northern, southern, western and eastern boundaries of this country. For every month, the equation of linear regression, the correlation coefficient and some other parameters are determined. It is found that the correlation coefficient is fluctuating, according to the various months and various observing stations, in the range of —0,50 to —0,85. The strongest correlation is found in September and during the summer months. The value of the regressional coefficient is fluctuating around the value of 1°, while the residual scatter is higher than 2°.

The highest values of the diurnal amplitude computed under the assumption of identical nebulosities are occurring in May, in contrast to the observed amplitudes, which are reaching their highest values in the summer and in September.

The results of this investigation are indicating that the relation existing between diurnal temperature amplitude and nebulosity is a similar one in the different areas of this country.

The diurnal variation of temperature may be either a periodic or an aperiodic one. The periodical diurnal amplitude is identical to the measure of the diurnal rise of temperature and thus it is primarily depending on the diurnal amplitude of solar radiation, i.e. it is finally depending on solar elevation at noon. However, solar elevation is determining only an upper limit to the radiation intake: the radiation amount actually received by the terrestrial surface, as well as the emitted radiation of the soil surface and of the lower strata of the atmosphere, that is, the whole atmospheric radiation balance, is the result of the mutual effects of several atmospheric factors. Among these factors, primarily nebulosity is to be taken into account. In the present paper, we are analysing the relation existing among solar elevation at noon (i.e. solar declination), the amount of nebulosity, and the diurnal variation of temperature.

In this investigation, we selected meteorological stations located respectively in the northern, southern, western, and eastern parts of this country, namely the stations Budapest-Lőrinc, Szeged, Szombathely and Debrecen, and we used the data series collected during the five years 1970 to 1974 (Fig. 1). The data were processed without any alteration, disregarding the periodical or aperiodical nature of the amplitudes.

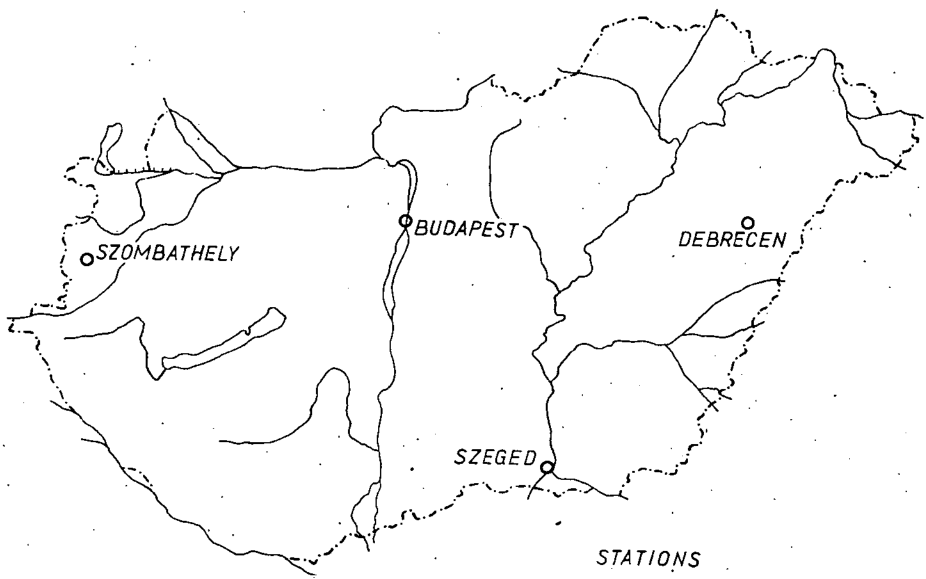

Fig. 1. The situation of the investigated stations
1. ábra. A vizsgálatba bevont állomások

As a first step, we determined the equation of the regressional straight line of the stochastic relation existing between the monthly average values of solar elevation at noon on the one hand, and the monthly average values of the amplitude of temperature (as, according to the results of a preliminary graphical investigation, this relation is a linear one); further, we determined the correlation coefficient, the standard deviation of average monthly amplitudes, the residual scatter and the quotient of the residual scatter by the regressional coefficient. The results are shown in *Table I.*

It is a well-known fact, that the value of diurnal amplitude is grosso-modo following the variations of solar declination. However, the high value of the correlation coefficient obtained here is still somewhat surprising. The average value of the four correlation coefficients corresponding to the four observing stations is

$$0{,}8663$$

Table 1

Annual scatter of the monthly average values of the diurnal amplitude of temperature [S_y], correlation coefficient between the monthly average values of solar elevation at noon and the monthly average values of diurnal amplitude of temperature [r_{xy}], equation of the straight line of regression [y=value computed for the diurnal amplitude of temperatur, x=value of the solar elevation at noon in degrees], residual scatter [S_{yx}] and the quotient of residual scatter by the regression coefficient [S_{yx}/a], for the stations Budapest-Lőrinc, Szeged, Szombathely and Debrecen, 1970—74

	S_y	r_{xy}		S_{yx}	S_{yx}/a
Budapest-Lőrinc	2,27°	0,9003	y=3,21+0,126x	0,99°	7,86
Szeged	2,31°	0,8403	y=5,04+0,119x	1,25°	10,50
Szombathely	2,33°	0,8777	y=4,34+0,126x	1,12°	8,89
Debrecen	2,32°	0,8469	y=4,61+0,121x	1,23°	10,17

The differences among the various correlation coefficients corresponding to the various stations are not exceeding the value of 0,0600. Similarly to the correlation coefficients, the regressional coefficients are varying also very slightly, and the regressional coefficients obtained from the data series Budapest-Lőrinc and Szombathely are identical to the fourth decimal digit. The quotients of the residual scatter by the regressional coefficient are indicating that an amplitude variation corresponding to a variation of 7,9 to 10,5 degrees in solar declination or, respectively, of solar elevation angle, is corresponding to the magnitude of the residual scatter.

From the very close correlation found between the monthly average values of temperature amplitude on the one hand and solar elevation on the other hand, it appears probable that in a good correlation to the solar elevation at noon, there should be a variation in another meteorological element which is of influence on the magnitude of the diurnal amplitude of temperature. Such a meteorological factor is the nebulosity. Although the diurnal temperature variation is not independent from the types, altitudes and thicknesses of the clouds which are present, we are omitting, in this analysis, these factors and we are taking into account exclusively the degree of nebulosity. We are using the monthly average values of nebulosity expressed in the unit "octa". We are presenting the equation of the regression line of the relation existing between this quantity and the monthly average values of the temperature amplitude, as well as the correlation coefficient, the standard deviation of the average values of the diurnal amplitude, the residual scatter and the quotient of this latter quantity by the regressional coefficient in *Table II*, while the same characteristics for the relation existing between solar elevation at noon and the degree of nebulosity are shown in *Table III.*

Table 2

Correlation coefficient between the monthly average values of nebulosity and the monthly average values of diurnal amplitude of temperature [r_{xy}], equation of the straight line of regression [y=value computed for the diurnal amplitude of temperature, x=value of nebulosity in octas], residual scatter [S_{yx}] and the quotient of the residual scatter by the regression coefficient [S_{yx}/a] for the stations Budapest-Lőrinc, Szeged, Szombathely and Debrecen 1970—74

	r_{xy}		S_{yx}	S_{yx}/a
Budapest-Lőrinc	—0,8739	y=20,86—2,55x	1,11°	0,435
Szeged	—0,8997	y=21,77—2,43x	1,01°	0,416
Szombathely	—0,9114	y=23,36—2,67x	0,96°	0,360
Debrecen	—0,8270	y=23,20—2,53x	1,30°	0,513

Table 3

Correlation coefficient between the monthly average values of solar elevation at noon and the monthly average values of nebulosity [r_{xy}], equation of the straight line of regression [y=value computed for nebulosity, in octas, x=value of the solar elevation at noon in dgrees], the residual scatter of the monthly average values of nebulosity [S_{yx}] and the quotient of the residual scatter by the regression coefficient [S_{yx}/a] for the stations Budapest-Lőrinc, Szeged, Szombathely and Debrecen, 1970—74

	r_{xy}		S_{yx}	S_{yx}/a
Budapest-Lőrinc	—0,6717	y=6,18—0,032x	0,58	18,13
Szeged	—0,6946	y=6,34—0,037x	0,62	16,76
Szombathely	—0,6668	y=6,47—0,032x	0,58	18,13
Debrecen	—0,6164	y=6,54—0,029x	0,60	20,69

Between nebulosity and the monthly average values of diurnal amplitude there is similarly a very close correlation. The average of the four correlation coefficients corresponding to the four observing stations is

$$-0{,}8780$$

and the highest difference among them is lower than 0,0850. There are two stations for which the correlation between the diurnal amplitude and the solar elevation at noon is slightly closer, while for the two other stations, this is the case for the correlation between the diurnal amplitude and the nebulosity. Both types of correlations are significant ones at a probability level which is lower than 0,1 per cent. Among the regressional coefficients, the highest difference is 0,25°, which is a value that is even lower than 10 per cent of the smallest coefficient. On the basis of the quotient of residual scatter by the regressional coefficient, the residual scatter is not exceeding 36 to 51 per cents of the variation of the amplitude for 1 octa of nebulosity variation.

Between the monthly average values of solar elevation at noon on the one hand, and of nebulosity on the other hand, there is also a rather good correlation, however it is significantly looser than the correlation between the same variable and the diurnal amplitude. The average of the four correlation coefficients corresponding to the four observing stations is

$$-0{,}6624$$

with a significance level lower than 5 per cents. The difference among the correlation coefficients are lower than 0,0800. From the quotient of residual scatter by the regressional coefficient it can be stated, that the magnitude of residual scatter is only reached by a change in nebulosity which corresponds to a change of 16,8° to 20,8° in solar declination.

As the correlations of the diurnal amplitude with both solar elevation at noon and nebulosity are much more close ones than the correlation between the latter two variables under themselves, it appears to be advisable to carry out also correlation and regressional calculations for three variables. The three dimensional regressional equations, which are expressing the mutual relation existing among the quantities (average values of diurnal amplitude, average solar elevation at noon, and average nebulosity) as well as the mutual correlation coefficients are shown in *Table IV*.

Table 4

Mutual correlation coefficient among the following quantities: monthly average values of the solar elevation at noon, monthly average values of the nebulosity and monthly average values of diurnal amplitude of the temperature [$r_{x_1 x_2 y}$] and the regression eqution with three variables [y = value computed for the diurnal amplitude of temperature, x_1 = value of the solar elevation at noon in degrees, x_2 = value of the nebulosity in octas] for the stations Budapest-Lőrinc, Szeged, Szombathely and Debrecen. 1970—74

Budapest-Lőrinc	0,9708	$y = 12{,}07 + 0{,}080x_1 - 1{,}43x_3$
Szeged	0,9482	$y = 15{,}48 + 0{,}059x_1 - 1{,}65x_2$
Szombathely	0,9808	$y = 15{,}73 + 0{,}070x_1 - 1{,}76x_2$
Debrecen	0,9348	$y = 14{,}46 + 0{,}078x_1 - 1{,}51x_2$

The average of the common correlation coefficients corresponding to the four observing stations is

0,9586

a value which is by 0,0923 higher than the average of the four correlation coefficients for the solar elevation at noon and by 0,0806 higher than the average of the four correlation coefficients for nebulosity. It is seen that taking simultaneously into account the solar elevation at noon and the nebulosity, only a slight increase in the correlation coefficient is experienced. This slight increase, however, is decreasing by some tenths of degrees the value of residual scatter, and, in the case of the residual scatter obtained from the observations taken at Szombathely, even a decrease of more than 0,5 degrees is occurring, which is more than the half of the residual scatter calculated for the two variables.

On the basis of the regressional coefficients, it can be stated that from the point of view of the effect on the magnitude of diurnal amplitude, at Budapest-Lőrinc, a variation of about 18°, at Debrecen of 19°, at Szombathely of 25° and at Szeged of 28° in solar elevation at noon (or, respectively, in solar declination) is equivalent to a variation of nebulosity by 1 octa. These figures are, at the same time, also expressing, that the variation of about 47° in solar declination, which is experienced between two solstices, is equivalent only to a variation in nebulosity by 1,7 to 2,6 octas. At a first glance, this may appear to be an irrealistic statement. However, it should be taken into account, that we are dealing with monthly average values and the annual variation of the monthly averages of nebulosity is comporting only some 2 to 3 octas, to which is corresponding the annual variation of the diurnal amplitude as well as it is corresponding to a variation in solar declination of 47 degrees. (However, it should be noted, that the annual variation of the monthly average values of solar declination is equal not to 47, but only to 46 degrees.) A different result should be obtained when using not monthly average values, but daily values in the calculation. In addition, the reliabilities of the values of the solar elevation at noon and of the nebulosity, used in the calculation, are not of the same order. The values of solar declination are known, with suitable exactitude, without making measurements; while the determination of nebulosity is the kind of meteorological observations which is exhibiting the highest numbers of errors, without mentioning the circumstance that a nebulosity consisting of clouds of various altitudes and thicknesses is characterized by one and the same value.

In the further discussions, the relation between daily values of diurnal amplitude and nebulosity is analysed, by grouping the data according to months. The relation existing between the solar elevation at noon and the diurnal amplitude will not be investigated on the basis of daily values.

A similar analysis has been carried out earlier by *E. Antal* [*1*] on the basis of diurnal amplitudes observed at Budapest. *Antal* proceeded by sifting very carefully his data and taking into account only the days which were synoptically free from advection. This author is calling the temperature amplitude "the measure of temperature rise". Hi is approximating the relation existing between nebulosity and this quantity by a curve of the third order, stating, that the variation of diurnal amplitude with nebulosity is a different one for low, medium and high values of nebulosity. Thus, the values calculated by *Antal* are not differing by more than 0,5 degree from the observed values of the temperature amplitude. In the course of our own investigations, a preliminary graphical test showed that the unsifted diurnal variation data are not exhibiting such a degree of discrepancy from the linear relation with nebulosity, that a search for linear equations should not be justified.

In the course of our investigations, we determined, for the four observing stations, the monthly values of the nebulosity and of the diurnal amplitude; the standard deviations of the monthly distributions of the nebulosity and the diurnal amplitudes; the variation coefficients after *Pearson* of the monthly distributions of the diurnal amplitude; the regressional equations which are describing the relation existing between the nebulosity and the diurnal amplitude; the correlation coefficients, the quotients of residual scatter by the regressional coefficient; the amplitude values corresponding to the nebulosities of 0, 4 and 8 octas; and finally the values of the residual scatter expressed in per cents of the value of the diurnal amplitude computed for 4 octas of nebulosity. These data are contained in the *Tables 5—8*.

Table 5

Correlation coefficient between the nebulosity and the diurnal amplitude of temperature [r_{xy}], equation of the straight line of regression [y=value computed for the diurnal amplitude of temperature, x=value of the nebulosity in octas], residual scatter of the diurnal amplitude [S_{yx}], quotient of the residual scatter by the regression coefficient [S_{yx}/a] and the values of the residual scatter expressed in percents of the computed values of the amplitude of temperature under the assumption of 4 octas of nebulosity [$100 \cdot S_{yx}/y_{4octas}$] for various months of the year at the station Budapest-Lőrinc, 1970—74

	r_{xy}		S_{yx}	S_{yx}/a	$\frac{100 \cdot S_{yx}}{y_{4octas}}$
Januar	—0,6675	y= 9,07—0,74x	1,76°	2,38	28,8
February	—0,6198	y= 9,75—0,71x	1,94°	2,73	28,1
March	—0,6760	y=13,89—1,01x	2,67°	2,64	27,1
April	—0,6603	y=14,91—0,99x	2,28°	2,30	20,8
May	—0,6735	y=15,52—1,07x	2,21°	2,07	19,5
June	—0,6791	y=14,81—0,98x	1,99°	2,03	18,3
July	—0,6438	y=14,73—0,97x	2,15°	2,22	19,8
August	—0,8025	y=14,69—1,08x	1,76°	1,63	17,0
September	—0,7144	y=14,19—1,02x	2,25°	2,21	22,3
October	—0,7602	y=14,18—1,15x	2,38°	2,07	24,8
November	—0,5804	y=11,05—0,83x	2,56°	3,08	33,1
December	—0,4756	y= 8,26—0,55x	2,22°	4,04	36,6

It is conspicuous, that, while between the nebulosity and the monthly averages of the diurnal amplitude, a very close correlation exists: the correlation becomes very much looser when using daily values of the same quantity, even within a period of a month.

The values of the correlation coefficient are fluctuating within the following limits:

at Budapest-Lőrinc	−0,8025	and	−0,4756
at Szeged	−0,8028	and	−0,5685
at Szombathely	−0,8392	and	−0,6016
at Debrecen	−0,8522	and	−0,5040

The highest values are occurring at Szeged, at Szombathely and at Debrecen in September and at Budapest-Lőrinc in August; however, the September value is also on the latter station a very high one. The lowest value of the correlation coefficient occurs at Szeged and at Szombathely in February, at Budapest-Lőrinc in December,

Table 6

Correlation coefficient between the nebulosity and the diurnal amplitude of temperature [r_{xy}], equation of the straight line of regression [y = value computed for the diurnal amplitude of temperature, x = value of the nebulosity in octas], residual scatter of the diurnal amplitude [S_{yx}], quotient of the residual scatter by the regression coefficient [S_{yx}/a] and the values of the residual scatter expressed in percents of the computed values of the amplitude of temperature under the assumption of 4 octas of nebulosity [$100 \cdot S_{yx}/y_{4octas}$] for various month of the year at the station Szeged, 1970—74

	r_{xy}		S_{yx}	S_{yx}/a	$\frac{100 \cdot S_{yx}}{y_{4octas}}$
January	—0,6061	y = 12,36—1,05x	2,68°	2,58	32,7
February	—0,5685	y = 12,46—0,85x	2,70°	3,18	29,8
March	—0,6638	y = 16,96—1,26x	3,59°	2,85	30,1
April	—0,6342	y = 17,69—1,24x	3,27°	2,64	25,7
May	—0,6953	y = 18,14—1,30x	2,67°	2,05	20,6
June	—0,6707	y = 16,39—1,11x	2,44°	2,20	20,4
July	—0,7686	y = 16,54—1,16x	2,13°	1,84	17,9
August	—0,7790	y = 15,54—1,21x	2,18°	1,80	18,6
September	—0,8028	y = 17,20—1,24x	2,25°	1,81	18,4
October	—0,7795	y = 16,78—1,35x	2,97°	2,20	26,1
November	—0,6128	y = 14,71—1,08x	2,37°	3,12	32,4
December	—0,5905	y = 11,06—0,84x	2,57°	3,06	33,4

Table 7

Correlation coefficient between the nebulosity and the diurnal amplitude of temperature [r_{xy}], equation of the straight line of regression [y = value computed for the diurnal amplitude of temperature, x = value of the nebulosity in octas], residual scatter of the diurnal amplitude [S_{yx}], quotient of the residual scatter by the regression coefficient [S_{yx}/a] and the values of the residual scatter expressed in percents of the computed values of the amplitude of temperature under the assumption of 4 octas of nebulosity [$100 \cdot S_{yx}/y_{4octas}$] for various months of the year at the station Szombathely, 1970—74

	r_{xy}		S_{yx}	S_{yx}/a	$\frac{100 \cdot S_{yx}}{y_{4octas}}$
January	—0,7506	y = 14,44—1,40x	2,39°	1,71	27,0
February	—0,6016	y = 12,34—0,87x	2,33°	2,68	26,3
March	—0,6849	y = 17,37—1,37x	3,39°	2,46	28,5
April	—0,6705	y = 17,91—1,38x	3,31°	2,39	26,7
May	—0,6722	y = 18,20—1,29x	2,92°	2,26	22,4
June	—0,6511	y = 17,08—1,21x	2,66°	2,19	21,7
July	—0,7631	y = 17,88—1,31x	2,55°	1,95	20,2
August	—0,6909	y = 17,18—1,29x	2,90°	2,25	24,1
September	—0,8392	y = 17,89—1,54x	2,31°	1,50	19,7
October	—0,6712	y = 15,25—1,13x	3,09°	2,73	28,8
November	—0,6721	y = 14,79—1,16x	3,21°	2,77	31,6
December	—0,5598	y = 12,01—1,01x	2,43°	2,41	30,5

Table 8

Correlation coefficient between the nebulosity and the diurnal amplitude of temperature [r_{xy}], equation of the straight line of regression [y=value computed for the diurnal amplitude of temperature, x=value of the nebulosity in octas], residual scatter of the diurnal amplitude [S_{yx}], quotient of the residual scatter by the regression coefficient [S_{yx}/a] and the value of the residual scatter expressed in percents of the computed values of the amplitude of temperature under assumption of 4 octas of nebulosity [$100 \cdot S_{yx}/y_{4octas}$] for various months of the year at the station Debrecen, 1970—74

	r_{xy}		S_{yx}	S_{yx}/a	$\frac{100 \cdot S_{yx}}{y_{4octas}}$
January	—0,5851	y=12,05—0,94x	2,72°	2,89	32,8
February	—0,5559	y=13,09—0,93x	2,54°	2,73	27,1
March	—0,6075	y=17,33—1,12x	3,54°	3,16	27,6
April	—0,5040	y=16,37—0,91x	3,14°	3,45	24,7
May	—0,6298	y=17,02—1,04x	2,78°	2,67	21,6
June	—0,6960	y=16,43—1,04x	2,25°	2,16	18,3
July	—0,6201	y=16,00—0,95x	2,35°	2,47	19,3
August	—0,6546	y=16,04—0,98x	2,62°	2,67	21,6
September	—0,8522	y=16,66—1,10x	1,93°	1,75	15,7
October	—0,7471	y=16,14—1,29x	3,03°	2,35	27,6
November	—0,4446	y=13,40—0,99x	2,76°	2,79	29,2
December	—0,5888	y=12,10—1,06x	2,30°	2,17	29,3

end at Debrecen in April. There appears to be a relation between the monthly average value of nebulosity and the monthly correlation coefficient. In the rather cloudless months, the correlation is a closer one, while in cloudier months, it is looser. All of the correlation coefficients are significant at a probability level lower than 0,1 per cent.

The value of the regressional coefficient is fluctuating, for all of the four observation stations, around 1°. The variation in the amplitude corresponding to 1 octa of variation in nebulosity is possessing its highest value at Szombathely (annual average: 1,25°). For Szeged, this average value is 1,14°, for Budapest-Lőrinc, 0,93° and for Debrecen, 0,92°. The regressional coefficient is reaching its highest value at Budapest-Lőrinc, at Szeged and at Debrecen in October, and at Szombathely in September. Lowest values of the regressional coefficient are characterizing the relation between nebulosity and diurnal amplitude in the winter months, mainly in February.

The values of the residual scatter are relatively high ones, their annual average is on all of the four stations exceeding the value of 2°. They are possessing a definite annual variation with maxima in the spring and in the autumn, and minima in the winter and the summer; however, the values obtained for various months are only slightly differing from each other. This annual variation is similar to that of the (whole) standard deviation, which is again characterized by maxima in spring and in autumn, a fact stated already by *Antal* [*1*] and also by our earlier work [*2*]. The residual scatter, however, possesses a main maximum in the spring, in March. Actually, the correlation is the closest one in the autumn, and thus in the autumn, namely in September and in October, the residual scatter is a lower one than in the spring. On two stations, the September value is even lower than the similar values of the summer months.

The quotient of the residual scatter by the regressional coefficient is for all of the four stations exceeding the value of 2. In annual average, a variation of the diurnal amplitude corresponding to a variation in nebulosity of 2,28 to 2,61 octas is reaching the magnitude of the residual scatter. The value of the quotient is lowest in September and in the summer months, while it is highest in November, in February and in March.

By using the regressional equation, we computed the values of diurnal amplitudes to be expected in the various months and on the various stations under the assumption that the nebulosity will be everywhere the same, namely, 4 octas, which is representing the middle point of the nebulosity range extending from 0 octas to 8 octas. Comparing

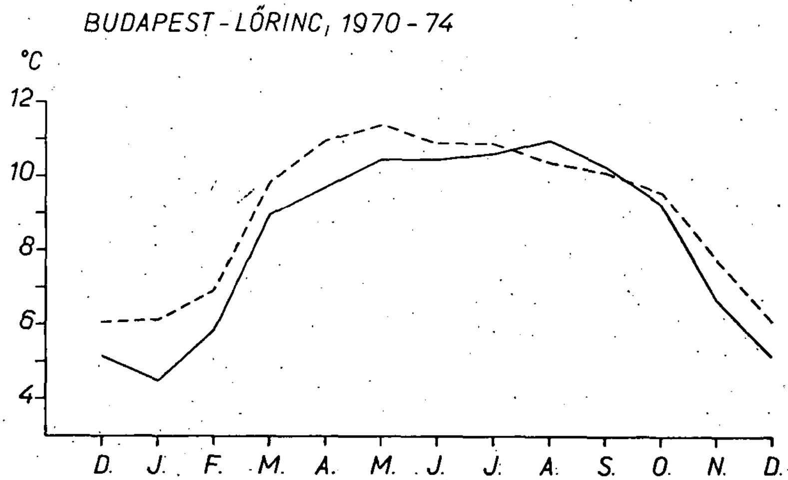

Fig. 2. Annual variations of the diurnal amplitudes as computed for a nebulosity of 4 octas (— — — —) and of the observed diurnal amplitudes (———) on the station Budapest-Lőrinc, 1970—74
2. ábra. A 4 okta felhőzetre számított napi ingás (— — — —) és a tapasztalt napi ingás (———) havi átlagainak évi menete Budapest-Lőrinc állomásról, 1970—74

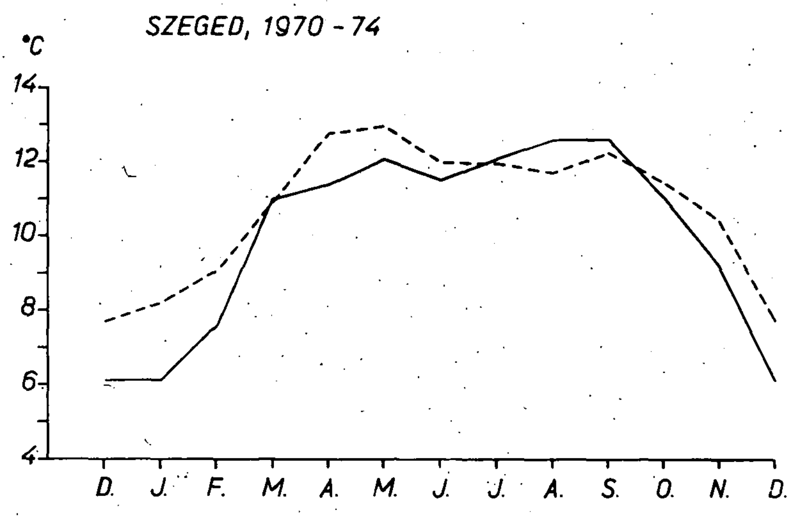

Fig. 3. Annual variations of the diurnal amplitudes as computed for a nebulosity of 4 octas (— — — —) and of the observed diurnal amplitudes (———) on the station Szeged, 1970—74
3. ábra. A 4 okta felhőzetre számított napi ingás (— — — — —) és a tapasztalt napi ingás (———) havi átlagainak évi menete Szeged állomásról, 1970—74

the values obtained in this way to the monthly average of observed amplitudes (*Fig. 2—5*) it is found that the annual variation of amplitudes computed on the assumption of a nebulosity of 4 octas is smaller than that of the monthly averages of the really

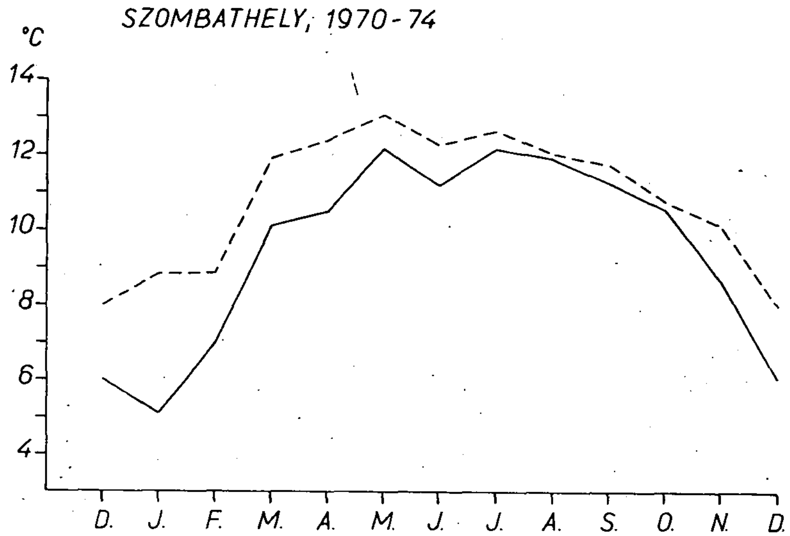

Fig. 4. Annual variations of the diurnal amplitudes as computed for a nebulosity of 4 octas (— — — — —) and of the observed diurnal amplitudes (———) on the station Szombathely, 1970—74
4. ábra. A 4 okta felhőzetre számított napi ingás (— — — — —) és a tapasztalt napi ingás (———) havi átlagainak évi menete Szombathely állomásról, 1970—74

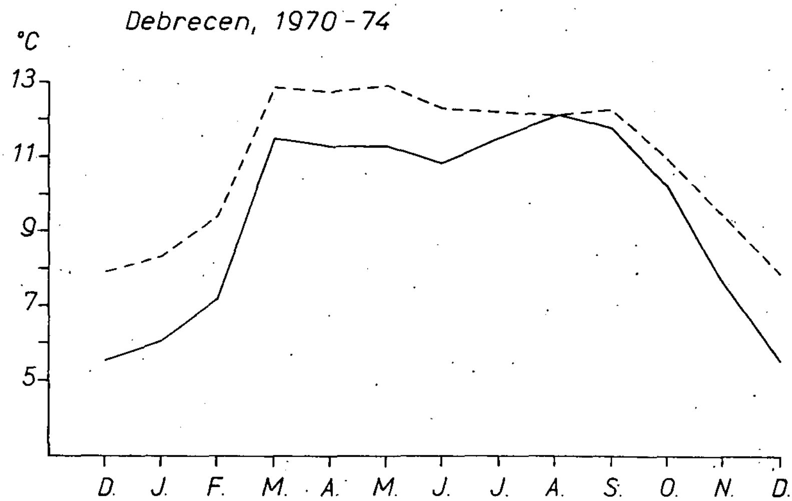

Fig. 5. Annual variations of the diurnal amplitudes as computed for a nebulosity of 4 octas (— — — — —) and of the observed diurnal amplitudes (———) on the station Debrecen, 1970—74
5. ábra. A 4 okta felhőzetre számított napi ingás (— — — — —) és a tapasztalt napi ingás (———) havi átlagainak évi menete Debrecen állomásról, 1970—74

observed amplitudes. This is indicating again that the discrepancies found between the summer and winter values of the diurnal amplitudes are partly to be attributed to a variation in nebulosity. Let us compare the observed annual average values of diurnal amplitudes to those computed on the basis of a nebulosity of 4 octas.

	Real average	Computed under the assumption of a nebulosity of 4 octas
Budapest-Lőrinc	8,57°	9,23°
Szeged	10,26°	11,01°
Debrecen	9,74°	11,10°
Szombathely	9,72°	11,04°

From the table above it appears that when assuming a uniform nebulosity of 4 octas, then the annual average, as compared to the real one, is at Szombathely and at Debrecen increased, at Szeged it is decreased, and among the average values for the three stations, there is not even a discrepancy of 0,1 degree. In the case of Budapest-Lőrinc, not only the real, but also the computed average values are different ones, a fact which should be attributed to local environmental inffuences.

Further it can be stated, that, while the monthly average values of the observed amplitudes are highest in the summer months and in September, the diurnal amplitudes computed on the assumption of a nebulosity of 4 octas is reaching on all of the four stations its highest value in May.

By using the regressional equation, we computed for every month those amplitude values, which are to be expected on entirely clear days (nebulosity 0 octas). From the monthly values obtained in this way, an annual average is derived with the following results: Budapest-Lőrinc, 12,94°; Szeged, 15,57°; Szombathely, 16,03° and Debrecen, 15,52°. The average values for Szeged and Debrecen are almost the same ones, however, that for Szombathely is by a half degree higher. (Among the four stations, Szombathely is characterized by the highest value of the regressional coefficient between nebulosity and diurnal amplitude.)

The annual maximum of the amplitude computed under the assumption of a nebulosity of 0 octas is in May, similarly as in the case of amplitudes computed under the assumption of a nebulosity of 4 octas. In an earlier paper [3] we investigated the highest amplitudes that occurred in the middle months of the four seasons, that is, in January, April, July and October, by using 60-year data series from the meteorological stations Szeged and Kecskemét. The average of these "monthly maximal amplitudes" is, among the four months, the highest at Szeged in April, while at Kecskemét in October, and on neither of these stations in July. The results mentioned above are rendering it desirable to carry out a study of the maximum amplitudes which should extend to every month. However, from the fact, that the diurnal variation computed under the assumption of a nebulosity of 0 octas has a higher value in May than in April, it is still not following that the maximum monthly amplitude should be also necessarily higher in May than in April. The residual scatter is higher in April than in May, and the April value of the diurnal amplitude computed under the assumption of a nebulosity of 0 octas, when increased by the residual scatter, is, according to the data of Szeged and Szombathely already higher than the corresponding value in May.

In addition to the values on entirely clear days and on days possessing a nebulosity of 4 octas, we determined, using again the regressional equation, also the daily values of the diurnal amplitude to be expected on overcast days (nebulosity 8 octas).

The annual average of these values is, for Budapest-Lőrinc, 5,42°; for Szeged, 6,45°; for Szombathely, 6,06° and for Debrecen, 6,06°. The ratio of the annual averages of this quantity on the two stations has indeed an inverse character to those computed under the assumption of a nebulosity of 0 octas, a circumstance which is connected to the fact that, among the three stations, Szombathely is possessing the highest value of the regressional coefficient, while Dbrecen is possessing the lowest one.

In contrary to the amplitudes computed for the two former values of nebulosity (0 and 4 octas), in the case of amplitudes computed under the assumption of overcast days, the maximum values are occurring on three stations not in May, but in April.

For a further characterization of the relation existing between nebulosity and diurnal amplitude, we are introducting another parameter which is derived in an analogous way to Pearson's variation coefficient. This parameter is expressing the

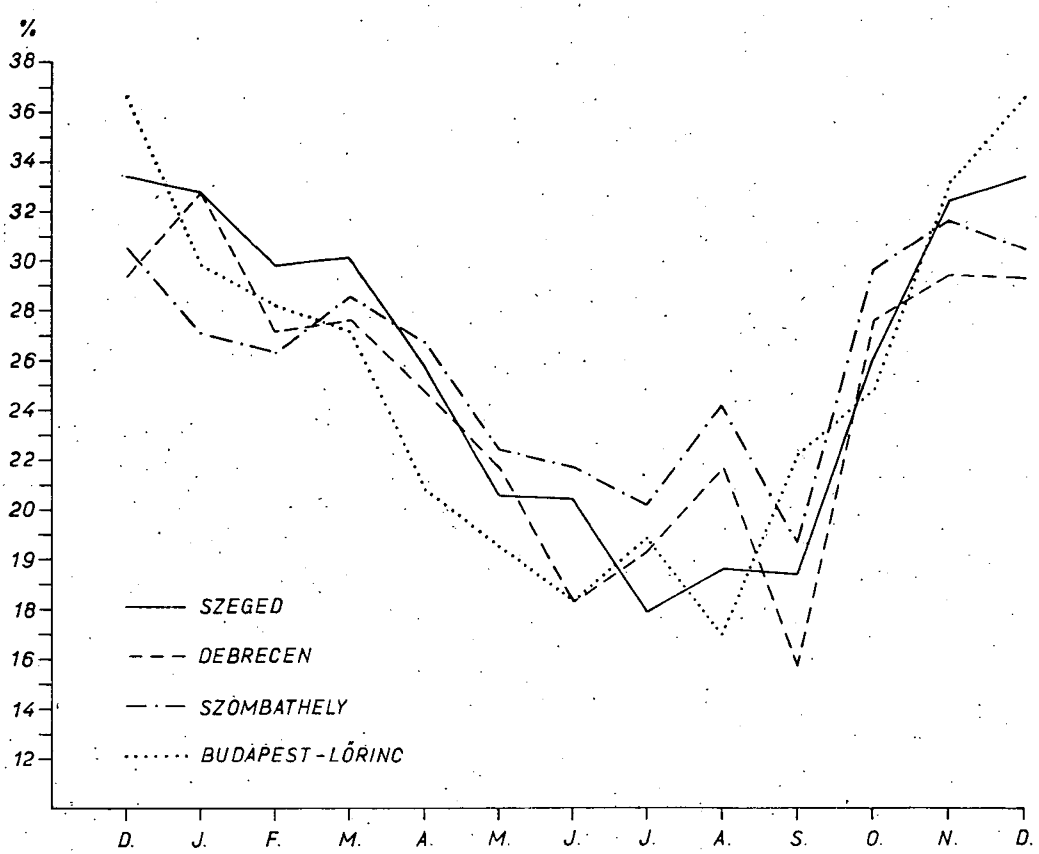

Fig. 6. Values of the residual scatter around the regression line between nebulosity and diurnal amplitude expressed in percents of the amplitude computed for days with 4 octas of nebulosity on the stations Budapest-Lőrinc, Szeged, Szombathely and Debrecen, 1970—74

6. ábra. A borultság és a napi ingás kapcsolatának regressziós egyenesétől számított reziduális szórás a 4 okta borultságú napokra számított napi ingás értékeinek százalékaiban, Budapest-Lőrinc, Szeged, Szombathely és Debrecen állomásokról, 1970—74 residual scatter in per cents of the amplitude computed under the assumption of a nebulosity of 4 octas. The monthly values of this parameter for the four stations are shown on *Fig. 6*. It is seen in this figure that the values of this parameter, called the "variation coefficient of residual scatter", are slowly decreasing from the winter months toward July, August and September, while they are, from September on, rapidly increasing. The curve is not a symmetrical one. Its amplitude is reaching on no station 20 per cents, and a Szombathely, it is not even reaching the value of 12 per cents. It is rather surprising, that the annual averages of these "coefficients" are only very slightly differing one from the other. The annual average is for Budapest—Lőrinc 24,7 per cent; for Szeged 25,5 per cent; for Szombathely, 25,6 per cent and for Debrecen 24,6 per cent. In spite of nearly identical annual averages, there are, among the various monthly values, some more important differences.

Summary

In this investigation, we used 5-year data series from four representative meteorological stations situated respectively in the northern, southern, western and eastern parts of this country.

We found that there exists a close correlation between the monthly average values of the diurnal temperature amplitude on the one hand and nebulosity on the other hand. The values of the correlation coefficient, the regressional coefficient, and the residual scatter are on all of the four observing stations nearly identical ones. The values of the correlation coefficient are fluctuating in the range of

$$0{,}8 \text{ to } 0{,}9$$

or, respectively, in the range of

$$-0{,}8 \text{ to } -0{,}9$$

and the magnitude of the residual scatters is fluctuating around 1°. The relation of three variables is yielding a common correlation coefficient of about 0,95 and it is characterized by a residual scatter of 0,5° to 0,8°.

On the basis of daily values, we investigated only the relation existing between the diurnal amplitude and the nebulosity. This correlation is not so close as the relation found among the monthly average values, and, among the correlation coefficients computed for various months, thre are experienced differences of about 0,3. The values of residual scatter are higher than 2°.

While the monthly averages of the really observed diurnal amplitude have their maximum in the summer months and in September, the values computed under the assumption of a uniform nebulosity of 4 octas or 0 octas have their highest values, on all four of the observing stations, in May.

Expressing in per cents the values of the scatter in the case of a nebulosity of 4 octas, a definite annual variation is found, with a minimum at the end of the summer, in September. Among the annual average values of the "variation coefficient" of the residual scatter as computed for the four observing stations, no difference exceeding 1 per cent could be found. This circumstance is again indicating that the relation existing between diurnal temperature amplitude and nebulosity has a similar structure in the different areas of this country.

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Data Concerning the Soil Temperature Conditions of Rice Stand with Flooding Water Cover of Different Depths

by

J. Juhász — Cs. Károssy and Á. Kiss

Adatok különböző magasságú árasztóvízzel művelt rizsállomány talajhőmérsékleti viszonyaihoz. A hazai rizstermesztés terméseredményei a rizs klímaigényessége és betegségekre való fogékonysága következtében nagyon változóak. A különböző okok folytán fellépő termésvisszaesések megfelelő agrotechnikai módszerek alkalmazásával elkerülhetők lehetnek, ha a klímatényezők és a rizsnövény fejlődése közötti szoros összefüggés természetét és mennyiségi vonatkozásait megismerjük.

Kutatási területünk szerves része a Dél-Alföld természeti tájainak. Tanulmányunkban rövid talajtani jellemzést adunk Szarvas környékének talajairól; kitérünk továbbá a felszínt alkotó talajrétegek fizikai és kémiai sajátságainak ismertetésére is.

A genetikus talajtípusok fejlődését kis területegységeken mutatjuk be, mivel a részletek közötti összefüggések csak kis területegységeken belül mutathatók ki. A Dél-Alföld természeti tájainak földrajzi osztályozásánál különös figyelemmel vagyunk a Körös-vidék és a Dél-Tisztántúl löszfennsíkjának talajviszonyaira.

A rizs fejlődéséhez nagyon fontos az optimális, vagy közel optimális talajhőmérséklet. Az árasztóvízzel művelt rizsállomány talajhőmérséklete — az árasztóvíz jó hőszigetelő képessége és nagy hőkapacitása következtében — a növény jó fejlődéséhez szükséges optimum közelében van. A különböző magasságú árasztóvízzel borított, különböző mélységű talajrétegek hőmérsékletének napi ingása júniustól szeptemberig különböző mértékben csökken.

Ez a csökkenés összefügg a növényállomány növekedésével és zártabbá válásával. Az összefüggő bugaszint fejlődésével a fényenergia egy része a bugaszintben használódik fel és így kevesebb energia éri a vízfelszínt és a talajfelszínt. Szeptember elején az árasztóvíz leeresztése után a bugaszinten átjutó besugárzás közvetlenül a talajfelszínt éri és a talaj felső szintje közvetlenül érintkezik a levegővel. Ennek következtében a száradó talajban a hőmérséklet napi amplitudója növekszik. Az árasztóvíz leeresztése után a talaj nedvességtartalmában és hőmérsékletében beálló változások elősegítik a növény teljes kifejlődését és a magvak érését.

Angaben zu den Bodentemperaturverhältnissen des Reisbestandes unter verschieden tiefer Wasserbedeckung. Die Ertragsdurchschnitte des heimatlichen Reisbaus sind sehr veränderlich wegen der Klimaanspruchsvollheit und Krankheitsempfindlichkeit des Reises. Die aus verschiedenen Ursachen erfolgenden Ertragsrückfälle können mit geeigneten agrotechnischen Methoden vermieden werden, wenn die Art und das Mass des engen Zusammenhangs zwischen gegebenen Klimaeinflüssen und der Entwicklung des Reises uns bekannt sind.

Das untersuchte Gebiet ist ein organischer Teil der Naturlandschaft der südlichen ungarischen Tiefebene. In einer kurzen Bewertung in der Abhandlung wird eine bodenkundige Klassifizierung der Gegend von Szarvas gegeben; daneben werden die physischen und chemischen Eigenschaften der die Oberfläche bildenben Bodenschichten beschrieben.

Die Entwicklung der genetischen Bodentypen werden in kleineren Gebietseinheiten angegeben, weil die Zusammenhänge der Einzelheiten nur innerhalb der kleinen Einheiten nachweisbar sind. Bei der geographischen Naturlandschaftseinteilung der südlichen Grossebene wird besondere Aufmerksamkeit den Bodenverhältnissen der Körös—Gebiet und der Lössebene des südlichen Tiszántúl (der Ebene östlich von der Theiss) gegeben.

Zur guten Entwicklung des Reises ist die Versicherung optimaler oder beinahe optimaler Bodentemperatur sehr wichtig. Die Bodentemperatur des mit Überschwemmungswasser kultivierten Reisbestandes ist — infolge der guten Wärmeisolierung und der grossen Wärmekapazität der Was

sermenge — in der Nähe des zum guten Wachstum der Pflanze nötigen Optimums. Die Tagesschwankung der Temperatur verschieden tiefer Bodenschichten unter verschieden tiefer Wasserbedeckung vermindert sich, in verschiedenem Masse, von Juni bis September.

Diese Verminderung hängt mit dem Wachstum und Geschlossenerwerden des Pflanzenbestandes zusammen. Mit der Entwicklung der zusammenhängenden Rispenzone wird ein grosser Teil der zum Reifen der Körner nötigen Lichtenergie in der Rispenzone benutzt und dadurch erreicht viel weniger Energie die Wasser- und die Bodenoberfläche. Anfang September nach dem Ablassen des Überschwemmungswassers erreicht die durch die Rispenzone durchdringende Einstrahlung den Boden unmittelbar, und die obere Schicht des Bodens kommt in direkte Berührung mit der Luft. Infolgedessen nimmt die Amplitude des Tagesganges der Temperatur im trocknenden Boden wieder zu. Nach dem Ablassen des Überschwemmungswassers begünstigen die Effekte der Veränderungen des Feuchtigkeitsinhalts der Temperatur des Bodens die volle Entwicklung der Pflanze und das Reifen der Körner.

The average rice crop yields in this country are very variable owing to the particular demands and susceptibility to disease of rice. Crop failures due to different causes can be avoided by suitable agrotechnical methods if we know the nature and extent of the relationship between the development of rice and the climatic conditions connected with the way of raising the crops.

The investigated area is an organic part of the natural landscape of the Southern Lowland. In a short analysis the pedological classification and the physical and chemical properties of the soil layers forming the surface in the region of Szarvas are described.

The development of the genetic soil types is described for smaller area units, because the connections between the parts can be demonstrated only within small units. In classifying the natural landscape of the Southern Lowland we are chiefly concerned, according to the purpose of our investigation, with analyzing the soil conditions of the Körös region and the loess table of the southern Trans—Tisza region.

For suitable growth of rice it is essential to ensure optimal or near-optimal soil temperatures. The soil temperature of the flooded rice field — owing to the good heat insulation and great heat capacity of the water mass — is about the optimum required for the development of the plant.

The diurnial fluctuation of the temperature of soil layers with different water covers and different depths decreases, in varying degrees, from June to September. This decrease is connected with the development and closing of the plant stand. With the formation of the panicle zone a large part of the light energy needed for the ripening of the grains is absorbed and utilized in the panicle zone, and thus less energy reaches the surface of the water or the soil. At the beginning of September, after the flooding water drained, the radiation penetrating through the panicle zone reaches the soil directly, and the surface of the soil is also in direct contact with the air. As a result of all this the amplitude of the diurnial variation of the temperature begins to grow again in the drying soil. After draining the flooding water, the effects of the changes in the moisture content and the temperature of the soil favour the full development of the plant and the ripening of the grains.

The region under examination is an organic unit within the natural land of the Southern Lowland of Hungary. After a short analysis a pedological classification of the region of Szarvas is presented together with the physical and chemical properties of the material of soil lyers of the surface.

The formation of the genetic soil types is best described according to smaller territorial units because the connections between the parts and details can be shown only within such smaller units. In this respect we are in a fortunate position since a detailed description of the soil-geographical conditions is to be found in the work of Stefanovits on "The soils of Hungary" [1].

In building up the regional natural geographic classification of the Hungarian Southern Lowland, our main objective is — in accordance with the aim of our investigations — to give an analysis of the soil conditions of the loessial table of the Körös region and thet of the southern region east of the river Tisza.

The region of the river Körös

The affluents of the river Tisza often changed their bedsand so their network covers a great part of thet area. In the lowland, a continually sinking region, river control has left several reedy and swampy territories in the old river bede. The areas freed from the water slowly dried out, the groundwater level sank down and thus it could not very much influence the formation of the soils of this land. Almost one half of the meadow clay soil of Hungary is to be found in this region. The subsoil of the petchblack clayey territories is often alkaline, it contains sodium and magnesium. According to Máté [2] the humus layer of the soil is of minimal thickness, the proportion between clay and humus is inverse.

Opinions differas to the formation of meadow clay. The development of meadow sols is attributed abundant moisture (*Sigmond*) acidity of theclay deposited by the rivers (*Csiky*), the effect of water (*Endrédy*) and to the marshy soil (*Ballenegger*). In many cases however, meadow soils formed also on loessial ground [3]: such a region is the loess table of the southern region east of the Tisza. Meadow soils are also found on acid alluvial soils. But let us accept any one of the theories, in every one of them water is of decisive importance in the formation of meadow soils and clays.

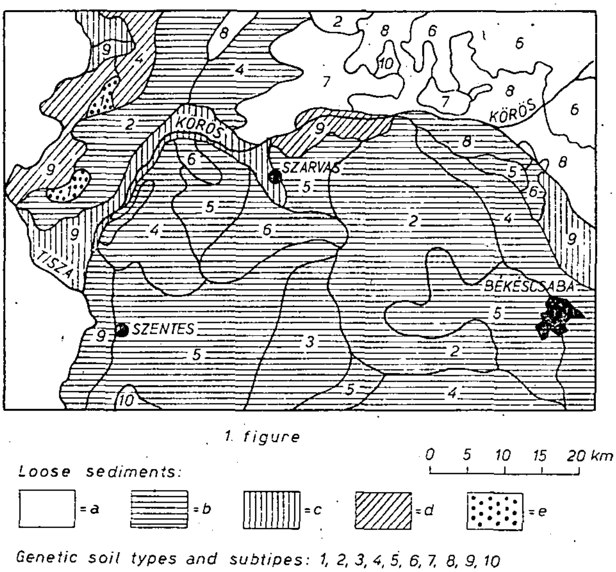

Fig. 1. Genetic soil map of Szarvas and its region according to Pál Stefanovits and László Szűcs

Symbols:

a = clay and clayey loam, loessial sediments; b = meadium compact clay, loessial sediments; c = clay and clayey loam, glacial and lacusstrine or alluvial sediments; d = medium compact loam, glacial and lacustrine or alluvial sediments; e = sandy glacial lacustrine sediments.

Genetic soil types and subtypes

1 = drift sand; 2 = lime-covered lowland chernozem; 3 = lime-covered lowland chernozem, alkaline in deep layers; 4 = meadow chernozem; 5 = meadow chernozem, alkaline in deep layers; 6 = steppifying meadow soloñetz; 7 = solonetz meadow soil; 8 = meadow soil; 9 = flood-deposit meadow soil; 10 = meadow solonetz

Table 1

(After Ibolya Kiss)
Values found in solonetz meadow soil at Galambos, 1965
a) The mechanical composition of solonetz meadow soil in %

Depth, cm	*Grain size, mm* 1,00—0,25	0,25—0,05	0,05—0,01	0,01—0,005	0,005—0,001	0,001	*Loss by treatment with hydrochloric acid*	*sand*	*clay*
0— 20	0,08	6,54	28,23	10,30	14,37	36,34	4,14	34,85	61,01
20— 40	0,83	1,96	23,75	14,13	8,20	47,14	3,99	26,54	69,47
40— 60	0,15	3,27	21,59	8,58	10,70	46,12	9,59	25,01	65,40
60— 80	0,07	0,98	25,68	9,09	11,86	45,25	7,08	26,73	66,20
80—100	0,09	2,76	32,42	12,20	6,95	37,36	8,11	35,27	56,51
100—120	0,20	9,24	37,13	6,15	10,49	30,83	5,96	46,57	47,47
120—140	0,14	14,48	33,62	9,11	7,79	29,25	5,61	48,24	46,15

Besides meadow soils, alkali soils of the solonetz type and their calciferous — sodic varieties the solontchak solonetzes play an important role (*Table 1*). Their amelioration was done by spreading yellow soil, limestone dust and caustic line [4]. The tempo and extent of the melioration is to be accelerated because the water amount of the existing irrigation canals is utilisable only after the melioration of the sodic soils.

The sodic and meadow soils of that land ara in some places characterized by a process of steppifying. The preliminary condition of that process is the change of water economy in the different soils, and a consequence of that is a quantitative and qualitative change in the fertile humus layer. Airlessness in the soil disappears, the acidity of the meadow soil decreases and thus an improvement, from the economical aspect, take splace.

The soils in the northern and eastern parts of this area show more favourable characteristics: meadow chernozem, salty meadow chernozem and chernozem with chalky sediment have developed (*Fig 1*).

The loess table of the Southern Trans-Tisza region — being a border-land — is of subdominant character since the region of Szarvas cannot be separated from this land.

The region is bordered by the rivers Körös, Tisza and Maros. The loessiel table extends also beyand the Rumanian border. In the formation of that land a great part was played by the river Maros which used to flow in other regions. Before the formation of the present river bed three periods may be distinguished [5]: other authors [6] make a distinction between several periods of stream deposit, silt transportation and periods of filling up, silting.

The Maros flows- with different periodical characteristich in a NW direction, then SW laying down its deposits. Reference to filling um in this region was already made by *Mihāltz* (in connection with sedimentation). In connection with the filling-upprocess a series of deposits formed gaining in refinement from below urward. In some cases this deposit series became repeated. These deposits were modified also by climatological changes and thus the accumulated layers became covered by layers of sand, mud and clay.

Table 1

(After Ibolya Kiss)

Values found in solonetz meadow soil at Galambos, 1965

b) Basic data of meadow soil and analysis of its 1:5 watery extract

Mark of genetic depth, cm		Total salt	Phenolphtalein alkalinity as soda	$CaCO_3$	Humus	pH	Watery solution (1:5) Ca^{2+}	Mg^{2+}	Na^+	K^+	CO_3^{2-}	HCO_3^-	C^-	SO_4^{2-}
		%					mg eq/100 g							
A_{level}	0— 18	0,07	—	1,07	2,78	7,2	0,62	0,50	1,20	0,05	0,40	1,50	0,30	0,48
B_1	18— 46	0,17	—	—	1,57	7,5	0,14	0,40	3,94	0,07	0,40	1,90	0,81	0,67
B_2	46— 68	0,15	—	—	1,24	7,4	0,14	0,13	4,00	0,09	—	2,70	0,51	0,73
B_3	68—100	0,15	0,08	9,61	0,95	7,6	0,27	0,43	1,52	0,03	0,40	1,85	0,14	0,23
BC	100—130	0,16	0,06	14,50	0,46	7,7	0,10	0,20	4,56	0,11	0,60	3,10	0,31	0,77
C_1	130—180	0,19	0,05	8,50	—	7,9	0,10	0,92	4,90	0,19	0,80	2,75	0,41	1,35
C_2	180—210	0,23	0,02	5,34	—	7,6	0,22	0,67	6,12	0,07	0,40	1,90	0,46	4,05
Ground water						7,9	mg eq/litre							
							1,32	7,92	35,93*		0,10	12,50	0,56	32,01

* Calculated value

Table 1

(After Ibolya Kiss)

Values found in solonetz meadow soil in Galambos, 1965

c) Exchangeable cations of solonetz meadow soil

Mark of genetic level, depth, cm		Ca^{2+}	Mg^{2+}	Na^+	K^+	Weight	Ca^{2+}	Mg^{2+}	Na^+	K^+
		mg eq/100 g					percentage of the weight			
A_{level}	0—18	25,70	7,00	1,09	0,46	34,25	75,04	20,44	3,18	1,34
B_1	18—46	13,50	17,70	5,87	0,56	37,63	35,87	47,04	15,60	1,49
B_2	46—68	9,30	15,40	7,87	0,46	33,03	28,15	46,62	23,82	1,40

d) Hrygroscopicalness, volume weight and compactness of solonetz meadow soil

		Hygroscopicalness	Volume weight	Compactness
A_{level}	0—30	5,05	1,49	56

Summarizing what has been said: the lowland lime-deposit chernozem can be found in the loess table of the southern Trans-Tisza region in a NE-SW direction connected with meadow chernozem and salty meadow chernozem and in the deeper regions meadow solonetz and steppifying meadow solonetz.

The utilization of the soils of that land is determined by the salt content, and the way of soil cultivation is to be made dependent on the distribution of salt in the different depths. Nevertheless the climatic and soil conditions are favourable even for the growth of more "demanding" plants. Melioration is motivated first of all in the meadow solonetzes and steppifying meadow solonetz soils.

Characterization of the material of the surface layers

The graine of river sand are sharp and angular while those of wind deposits rather round. The two kinds of sand can be separated according to their percentual composition [7]. The determination of the origin of some sandy layers can be made by the sedimented mud. The presence of mud alw fluvial origin, while there is no mud in windborn sand. Its other characteristic feature is the appearance of gastеropods, since in windborn sank pméy cpmtomemtaé sőecoes pr still-water species can be found. The water permeability of wind deposit sand is $k=10^{-3}$ cm/sec while that of river water is $k=10^{-2}$ cm/sec. The difference comes from the difference in composition. In the present case the characterization of the different kinds of loess is omitted: only a rough characterization of tide land loess and alkali-clayey loess is presented.

Tide land loess developed in the flood basins of rivers raising the clay-mud content of the area, because after the recession of the flood a large amount of suspended material was left behind. Its water permeability is smaller — 10^{-7} cm/sec than that of the other loesses.

In flood areas without an outlet alkalic clayey loess developed with a high clay content. Its development is partly parallel with loess-formation, partly a consequence of the present period of alkalization 7. Their origin can be traced back to water transport.

The results of the borings of *Mihåltz* enable us to give a rough outline of the fluvial deposits of the Valley of the Tisza and the Trans-Tisza region. 10 According to these, in the Tisza Valley (at a depth of some metres) and in the Trans-Tisza region fluvial deposits can be found.

The fluvial sand the in Pleistocene and Holocene layers finies-grained and mediumgrained respectively, (0,1—0,4 mm∅) with a water permeability of 10^{-4} cm/sec. The medium-grain layers are to be found below and the fine-grained ones above. The deposits of the rivers Körös and of Maros are characterised generally by medium-grained sand. Into the category of the loose sand-water sand belong the rough-grained, medium-grained and fine-grained sand. The differentiation of the diameters of grains entails categorization of the different kinds of sand and differentiation of the leakage factors according to value, thus i.e.:

The factor of rough sand (grain: 0,5 mm) is 10^{-2} cm/sec;
The factor of fine-grained sand 0,1 mm) is 10^{-5} cm/sec;
The factor of mud with grains of 0,02—0,002 mm is 10^{-7} cm/sec.

The leakage factor shows an almost direct proportion with the diameter of the grains. If the value of the leakage factor attains 10^{-8} cm/sec the sediment is sludge, clay, i.e. practically an impermeable layer. The leakage factor of the flood-mud of the investigated areas is not much smaller than that of clay: it is 10^{-7} cm/sec. Where the flood-mud covers grassy areas the leakage factor increases:

Among grain crops the role of rice is an important and particular one: it is a plant requiring warmth and a water-cover as it is of tropical origin. The rice grown in this country — under conditions very different from those of its place of origin — responds most sensitively to the extreme weather conditions characteristic of this country.

Thes problems of the growing and acclimatization of rice, the selection of the most resistant sorts, experimentation with a view to find the most appropriate agro-technical methods, and also the investigation of the connections between the stand-climate of the different sorts of rice and the climatic influences, are debated questions to this day.

Within the programme of complex investigation on rice breeding and cultivation, climatological investigations were carried out by us — joining with the oecological and physiological investigation carried out by the Research Institute for Irrigation in Szarvas — in 1975 at the experimental station of the above Institute in Galambos; the investigations were carried out in a rice stand covered with flooding water of various depths.

As is generally known, rice grown with flooding gives the best crop results in warm and dry years abundant in sunshine [8., 9., 10.].

Since the wather of the summer months of 1975 was far from that, it seems necessary to give a short characteristic of the weather of the above period 11.

The summer months of 1975 were unusually rainy and moderately cool. The precipitation amount of the months June, July and August surpassed in the greater part of the Hungarian Lowland 175% of the average of many years, and in the southern part of the region between the Danube and Tisza even 200% of the normal value. According to the data of the climatological station of Szarvas the precipitation amount of the summer months was 295 mm, i.e. 189% of 50 years' normal value. Correspondingly to the abundant precipitation the number of sunshine hours was very small; 704 hours of summer sunshine duration is 83% of the normal. Accordingly the summer amount of the short-wave global radiation was considerably less than the average radiation values: 47,42 cal cm^{-2}.

The mean temperature of the summer months was by 0,5—1 °C lower than 50 years' normal value. The moderately cool summer weather gave less balanced extreme temperature values. The average daily maximum temperatures remained below 33 °C (as against the normal 35—36 °C). The number of heat-days was only 10 during the whole summer period, while their usual number is 25 in Szarvas 12.

Within our research programme the temperature of the soil lying under the cover of flooding water of different depthes was registered with thermo-electric thermometers. The daily energy-amounts of short-wave global radiation were measured with a Kipp-type radiation-integrator and the daily values of the sunshine duration were recorded with a Campbell—Stokes sunshine recorder. On the basis of the collected data connections were sought between the depth of the flooding water and the soil temperature, and also between the sunshine duration and the short-wave global radiation.

As the first step of the investigations, the daily values of global radiation and sunshine duration, the factors primarily influencing the soil temperature, were compared month for month by drawing their regression curves (*Fig. 2*).

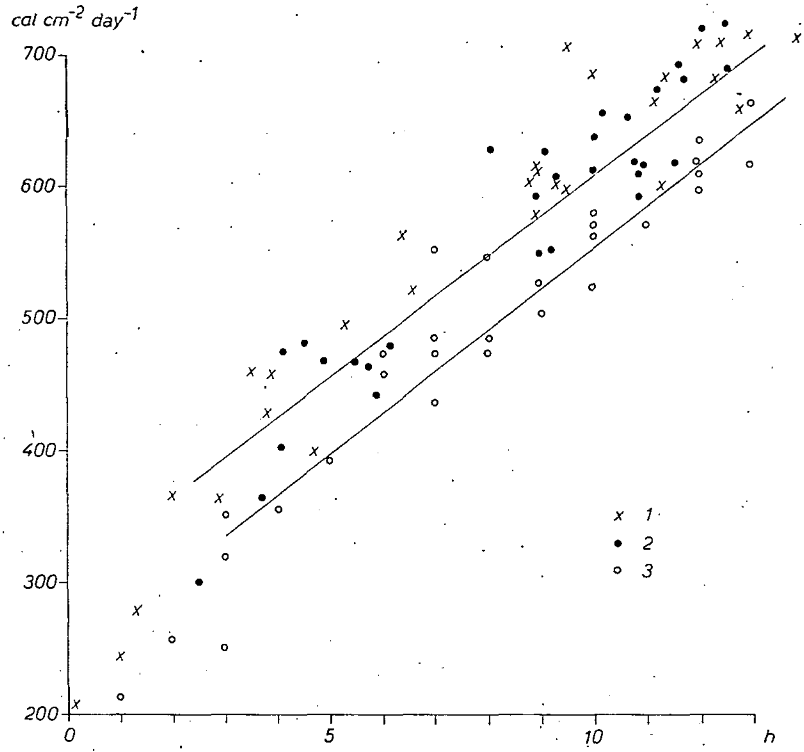

Fig. 2. Diurnal sums of global solar radiation and the empiric regression lines of the diurnal sums of sunshine at Szarvas, (June—July, 1975)
Key: 1 = June, 2 = July, 3 = August

The points in the above *Fig. 2* show a marked scattering around the empirical regression curves show a positive correlation but in the smaller value-ranges an obviously curvilinear correlation was found. Neglecting the overcast days (with values of global radiation lower than 350 cal cm^{-2} day^{-1} [13], we found a regression line instead of the empirical regression curves. The empirical regression lines can be considered approximately identical in June and July, while in August it is a line almost parallel with the lines of the previous two months shifted towards the smaller calory-values. The considerable deviation of the regression lines can be explained with the lower values of the sun's altitude in August and with the radiationweakening effect of the oceanic air masses streaming in more frequently in August [14].

In the region studied here soil temperature was measured at three measuring places — in accordance with the depth of the flooding water. These depths were: "shallow water" (S): a water layer of 5—10 cm; "medium deep water" (K): 10—15 cm and "deep water" (M), a water layer of 20—25 cm. The thermoelectric thermometers were fixid at two layers each measuring point. The thermometers situated in the highest layer of the soil below the water ara designated with "f" (surface), those placed deeper in the soil with "t" (in the soil). Thus at the three measuring places the following measuring points can be distinguished: S_f, S_t, K_f, K_t, M_f and M_t.

The data containing the temperature trend of the soil covered by flooding water of different depths were analysed by hourly division and from the hourly temperature

Table 2

Monthly mean values of the temperature of the soil layers under flooding water of various depths Szarvas, June, July, August and September, 1975

		Mean			Maximum			Minimum			Amplitude		
VI.	surface-near soil	24,2	22,9	19,6	28,2	27,7	23,2	20,6	27,7	23,2	7,7	8,8	6,9
	deeper soil	20,3	23,0	20,9	21,7	26,9	24,7	15,3	19,7	17,2	6,4	7,2	7,2
VII.	surface near soil	23,8	22,9	21,8	25,5	24,8	24,3	22,0	20,9	20,7	3,5	3,9	3,6
	deeper soil	23,0	23,5	24,3	25,0	25,0	25,7	22,3	21,9	22,8	2,7	3,1	2,9
VIII.	surface-near soil	21,1	20,3	19,4	21,8	21,1	20,0	20,2	19,3	18,5	1,6	1,8	1,5
	deeper soil	22,0	21,2	21,4	22,6	21,8	21,9	21,4	20,4	20,9	1,2	1,6	1,0
IX.	surface-near soil	17,5	16,4	17,4	19,0	17,3	21,0	12,1	10,3	11,2	4,1	4,3	6,0
	deeper soil	18,9	17,7	19,5	18,1	22,8	14,1	14,1	12,1	14,1	3,1	3,7	4,8

values thus obtained daily and monthly mean values were calculated. We determined also the monthly average values of the maxima and minima of soil temperatures and also their monthly average amplitudes. *Table 2* contains the monthly mean values of the temperatures of the soil layers under flooding water of different depths.

From this *Table* it can be seen that in June, from among the monthly mean temperatures of the six measuring points S_f (upper soil layer of the shallow water cover) is the lowest: the difference is about 4,6 °C.

Similarly, a large temperature difference appeared at the measuring station S between two points of different depths (3,9 °C). At the other two measuring places and at measuring places covered by deeper flooding water only smaller differences were found between the monthly mean temperatures of the levels of different depths. In the soil layers covered with deeper flooding water the deeper soil layers are warmer by 0,1—1,3 °C. In July already the monthly mean temperature of M_t of the deeper soil layers of the "deep water" is the highest, while in the upper soil layer of the same measuring place the monthly average of M_f is the smallest among the monthly mean temperature values of the six measuring places. Their difference is 2,5 °C, i.e. by 1,4 °C smaller than the difference of the previous month. At the measuring place "shallow water" the upper soil layer is still warmer than the deeper layer: only by 0,8 °C (as against 3,9 °C of the previous month). In August the deeper layer is warmer at all the three measuring places: their difference is 0,9—2,0 °C in favour of the deeper soil layers.

Between the individual measuring pints the deeper soil layer is somewhat warmer. The smallest mean value can be found — like in the previous months too — at the point M_f.

The monthly average of the temperature of the upper soil layers decreases at the measuring place "shallow water" from June till September; it is of the same value at "medium deep water" in June and July, and than it decreases until September. At the measuring place "deep water" the mean value of soil temperature still increases from June to July, while from July to September it decreases. The monthly mean temperature value of the deeper soil layer is, at each of the three measuring places, higher in July than in June, while it decreases from July to September. According to the above, the monthly average of the soil temperatures is the highest in July at all measuring places with the exception of the measuring points S_f [15].

The decrease of the monthly average values of the soil temperatures (beginning from July) is obviously in connection with the decrease of global radiation, but here also the role of the development os the rice stand is to be analysed [16].

To that end the closeness of the connection between the daily temperature of the soil of the rice stand and the daily global radiation amount was analysed by the sid of regression and correlation computations (concerning the months July and August).

For the comparison the soil temperature of the upper layers of the soil of the measuring place "shallow water" (S_f) and that of the deeper layer of the soil of the measuring place "deep water" was chosen. By graphic procedure it was found that the connection can be considered linear.

The results of the computations were the following:

July

	Equation of the regression line	*Residual scattering*	*Correlation coefficient*
S_f	$y = 12{,}4\ °C + 0{,}024\ x$	2,0 °C	0.5178 ($n = 22$)
M_t	$y = 20{,}5\ °C + 0{,}0068\ x$	2,1 °C	0,3655 ($n = 22$)

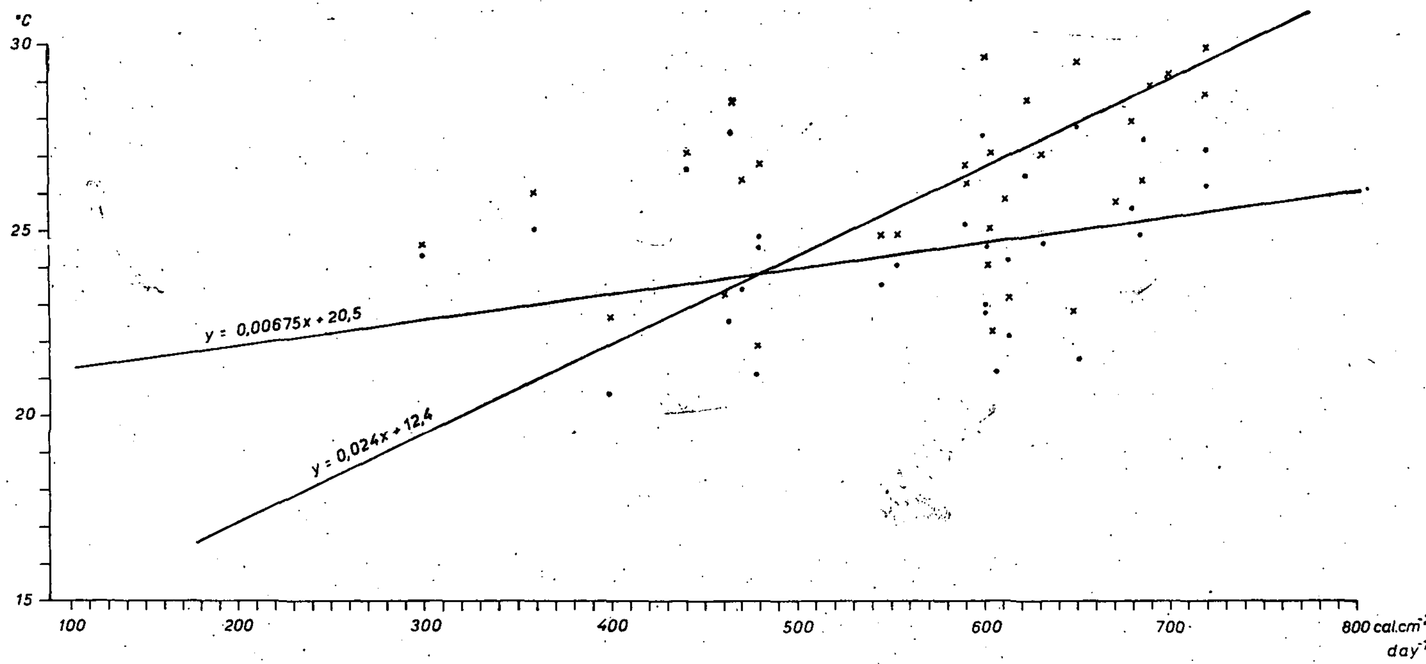

Fig. 3. Correlation of the diurnal mean temperature of the upper soil layer under „shallow water". and that of the deeper soil layer under deep water with the diurnal sum of global radiation in July. The S_f layer is the surface-near layer of the soil of „shallow water", the M_t layer is the deeper layer of the soil of „deep water". The values of the diurnal mean temperature of the S_f layer are marked by x, the layers of M_t by dots. The equation $y=0{,}024x+$ $+12{,}4°$ is the equation of the S_f layer, and the equation $y=0{,}00675x+20{,}5°$ is the equation of the regression line of the M_t layer (x=cal cm^{-2} day^{-1}).

August

S_f	$y = 20{,}0\,°C + 0{,}0025\,x$	0,9 °C	0,4150 ($n = 30$)
M_t	$y = 20{,}6\,°C + 0{,}0018\,x$	0,8 °C	0,3333 ($n = 30$)

y = the temperature of the individual layers of the soil,
x = the daily amount of global radiation in cal cm^{-2}.

From the above it can be stated that the daily mean temperature of the deeper soil of the measuring place "deep water" shows a weaker correlation with the daily amount of global radiation than the mean temperature of the layer S_f of the measuring place "shallow water", and the value of the regression coefficient too, is smaller than the regression coefficient of the soil of the deep water.

In August the correlation between the daily mean temperatures of the two measuring points and the global radiation is considerably smaller than in July and even the regression coefficients are much smaller. In the case of the soil near the surface below the shallow water the coefficient of the regression line decreased by the tenfold order of magnitude from July to August, and by more than a tenth the value of the correlation coefficient. The rest of scattering around the line also decreased to more than one half. In the case of the deeper soil layer below the deep water an almost threefold decrease of the regression coefficient can be found from July to August; the value of the correlation coefficient changes accordingly, similarly to the decrease of the residual scattering.

It is remarkable that the regression lines of August run almost parallelly, showing thus the similarity of the correlation of the two soil layers with the daily amount of global radiation (*Fig. 3—4*).

Although the degree of reliability of the correlation coefficients are, on account of the small number of cases, unsatisfactory, they do not exclude and even confirm the presumption that in August the strengthening and closing rice stand disturbes and reduces the effect of the global radiation on the trend of the soil temperature. In other words: the plant stand more or less overshades the soil and the water layer from the short wave global radiation [17].

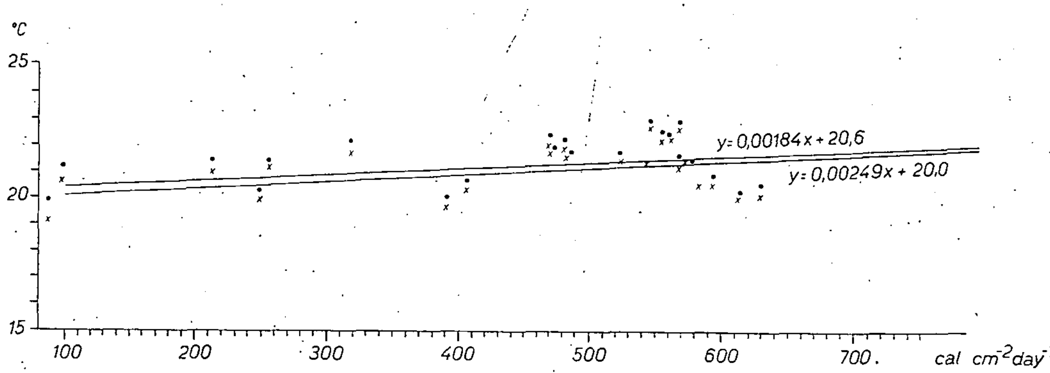

Fig. 4. Correlation of the diurnal mean temperature of the upper soil layer under shallow water with the diurnal sum of global radiation in August.
The S_f layer is the surface-near layer of the soil of „shallow water", the M_t layer is the deeper layer of the soil of „deep water". The values of the diurnal mean temperature of the S_f layer are marked, by x, those of the layers of M_t by dots. The equation $y=0{,}00249x+20{,}0°$ is the equation of the regression line of the S_f layer, and the equation $y=0{,}00184x+20{,}6°$ is the equation of the regression line of the M_t layer (x=cal cm^{-2} day^{-1}).

The monthly averages of the daily maxima of the soil temperature decrease from June to September at all the three measuring places at both measuring levels, with the exception that the temperature of both levels of the measuring place "deep water" increases from June to July and decreases only beginning from July. This is in connection with the well-known temperature phase-delay to be found in the annual temperature trend of the deeper soil layers [18].

The monthly trend of maximum temperature differs in June — July from the monthly trend of the daily mean temperature, as it to be expected, since the value of global radiation in June is somewhat higher than that of July, and the effect of it can appear already in the extreme values of the soil temperatures [19, 20].

Similarly to the trend of the monthly averages of maximum temperatures the monthly variations of the monthly averages of daily minima (the other extreme temperature value) is also obvious. In the case of soil layers near the surface the monthly averages of daily minima of the soil temperatures gradually decrease from June to September (with the exception of the layer S_f).

On the other hand, the average monthly minima of the soil temperature of the deeper soil layers and of the measuring place S_f increase from June to July, but decrease from July to September with a gradually increasing tendency.

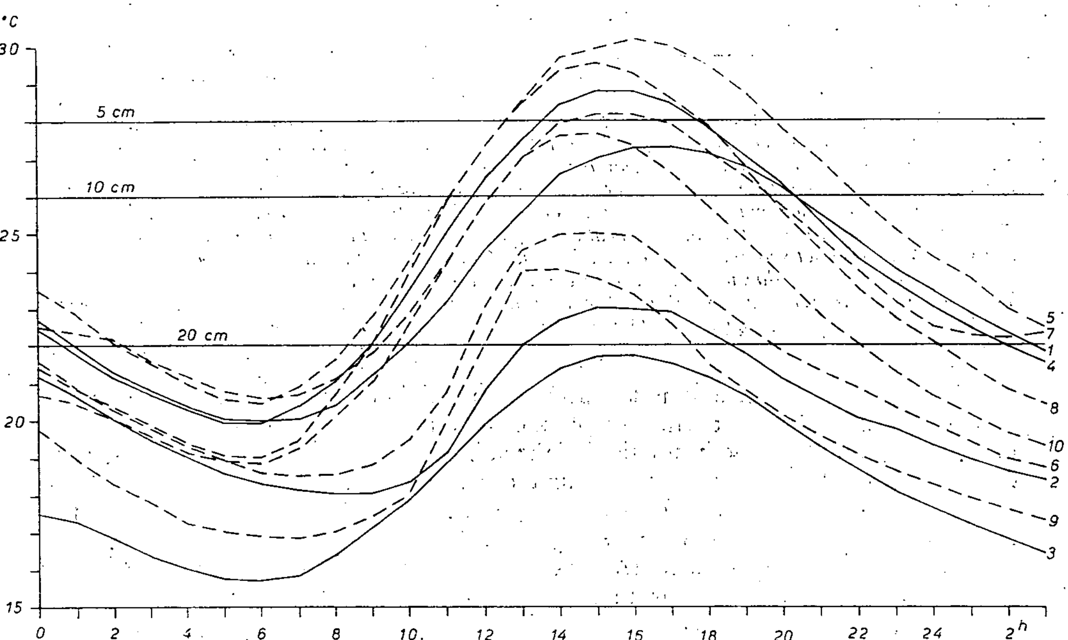

Fig. 5. The diurnal variation of the temperature of the soil of a rice field covered by flooding water of various depths in surface-near and deeper (10—15 cm deep) layers at Szarvas in June, 1975. Key: 1 = deeper layer of the soil of „shallow water", S_t on a cloudless day (June 27); 2 = S_t layer on a cloudy day (June 30); 3 = mean monthly sunshine hours; 4 = deeper layer of the soil of „moderately deep water"; K_t on a cloudless day; 5 = surface-near of layer the soil of „shallow water", S_f on a cloudless day; 6 = S_f layer on a cloudy day; 7 = mean of monthly sunshine hours of the S_f layer; 8 = surface-near layer of the soil of „moderately deep water", K_f on a cloudless day; 9 = K_f layer on a cloudy day; 10 = mean sunshine hour values of the K_f layer.
The horizontal lines are mean values of the soil temperature at 5, 10 and 20 cm depths measured at the agrometeorological station of Szarvas in the third ten days of July.

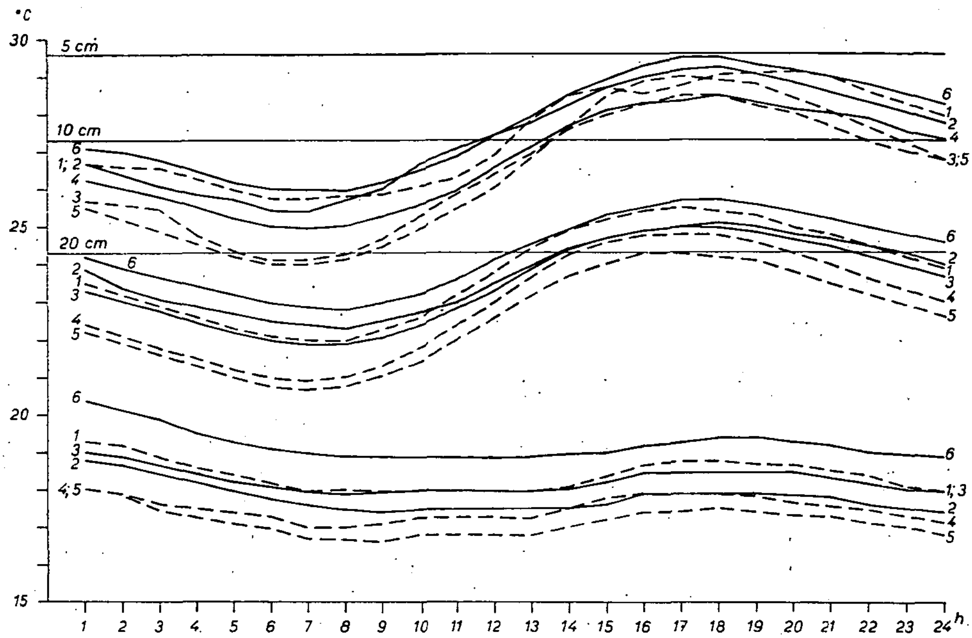

Fig. 6. Diurnal variation of the temperature of the soil of a rice field covered by flooding water of various depths in surface-near and deeper (10—15 cm deep) layers at Szarvas in July 1975.
Key: 1 = surface-near layer of the soil of „shallow water", S_f; 2 = the deeper layer of the soil of „shallow water", S_t; 3 = the deeper layer of the soil of „moderately deep water", K_t; 5 = the surface-near layer of the soil of „deep water", M_f; 6 = the deeper layer of the soil of „deep water", M_t.
The upper set of curves show the diurnal variations of the soil temperature observed on a cloudless day (July 15), the lower set show the same on a cloudy day (July 1), and the middle set of curves show the diurnal variations of the mean sunshine hour values.
The horizontal lines represent the mean soil temperatures observed at 5, 10 and 20 cm depths at the agrometeorological station of Szarvas in July.

Particularly considerable is the decrease of minimum temperature in the first half of September when, by raining off the flooding water the surface of the soil becomes dry and thus the irradiation of the soil may become more intensive. When draining off the flooding water the minimum values of August (about 20 °C on the average) show in the first half of September, both in the layers near the surface and in the deeper ones, lower values (by 8—10 °C).

The effect of the flooding water and of the rice stand influencing the soil temperature, appears most clearly in the monthly variations of the daily amplitude of the soil temperature. Parallelly with the growth of the plant stand the amplitude will decrease in both layers of the soils covered by "shallow" "medium" and "deep" flooding water. The decrease of the amplitude is about 4 °C from June to July and 2 °C from July to August. On the other hand, from August to September the average monthly amplitude of soil temperatures increase by 3—5 °C. The increase of the amplitude is a consequence of the draining of the flooding water and is particularly considerable in the soil layers covered earlier by deep water [21[.

The daily trend of the soil temperature of rice grown with flooding water of different depth is determined in the first place by the development and closeness of the plant stand. The depth of the flooding water influences depending on the agro-

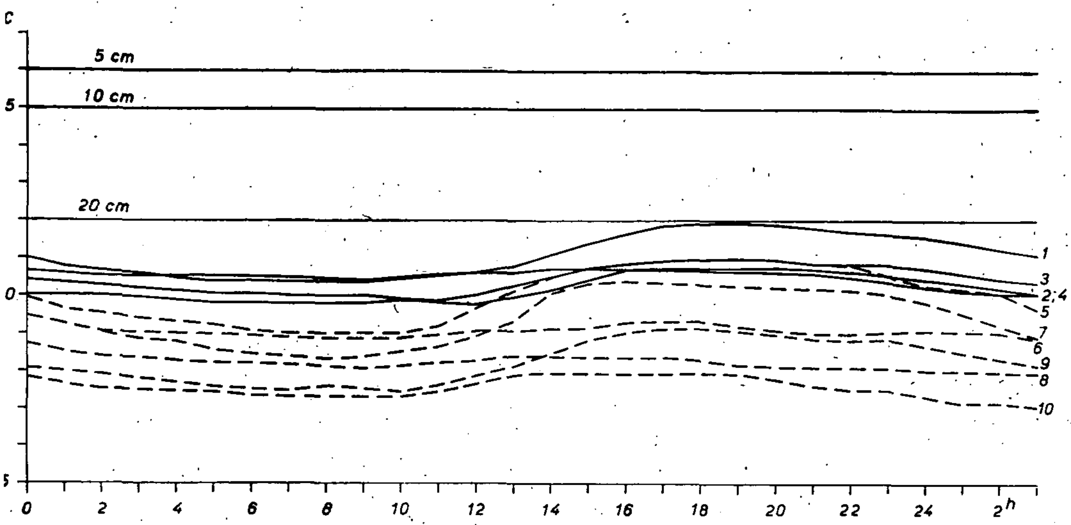

Fig. 7. The diurnal variations of the temperature of the soil of a rice field covered by flooding water of various depths in the surface-near and deeper (10—15 cm) layers at Szarvas in August, 1975.
Key: 1 = S_t layer on a cloudless day (August 21); 2 = S_t layer on a cloudy day (August 29); 3 = M_t layer on a cloudless day; 4 = M_t layer on a cloudy day; 5 = S_f layer on a cloudless day; 6 = S_f layer on a cloudy day; 7 = K_f layer on a cloudless day; 8 = K_f layer on a cloudy day; 9 = M_f layer on a cloudless day; 10 = M_t layer on a cloudy day.
The horizontal lines represent the mean soil temperatures observed at 5, 10 and 20 cm depths at the agrometeorological station of Szarvas in the third ten days of August.

technical methods applied according to the development of the stand, — only secondarily the tendency of the daily temperature of the different soil layers. The primary influencing role of the complex climatological ensemble (soil layer — rice stand — flooding water) is proved by the daily soil-temperature trends shown in *Figures 5, 6, 7, 8.* In our Figures we have shown (by lines parallel with the horizontal axis) the mean values of soil temperatures falling to the respective decades (measured at the Agrometeorological Station of Szarvas) at the depths of 5, 10 and 20 cm. The daily trends of soil temperature values (referred to average hour means) and also those of clear and overcast days of months following each other from July to August show a gradually increasing deviation from the mean values of soil temperatures of 5, 10 and 20 cm falling to the individual decades. From the Figures it can be clearly seen that it is in connection with the progressive closing of the plant stand and the development of a connected cluster level. The stand growing more and more dense gradually closes the soil surface and also the deeper soil layers from radiation. In the first half of September the flooding water is — according to the applied growing methods — drained off from the rice stand. As a consequence the gradually drying soil comes into direct contact with the air (having been warmed up during the day). Thus the daily trend of the soil temperature of the different soil layers becomes at the beginning of September — despite the considerably decreased daily global radiation values — similar to the daily trends of soil temperatures in June with markedly wide amplitudes. While in June the soil temperature values of the afternoon hours considerably suppass the ten-year average values of the layers of 5, 10 and 20 cm of the soil surface without water cover, in September the daily soil temperature values remain below these values even in the midday hours.

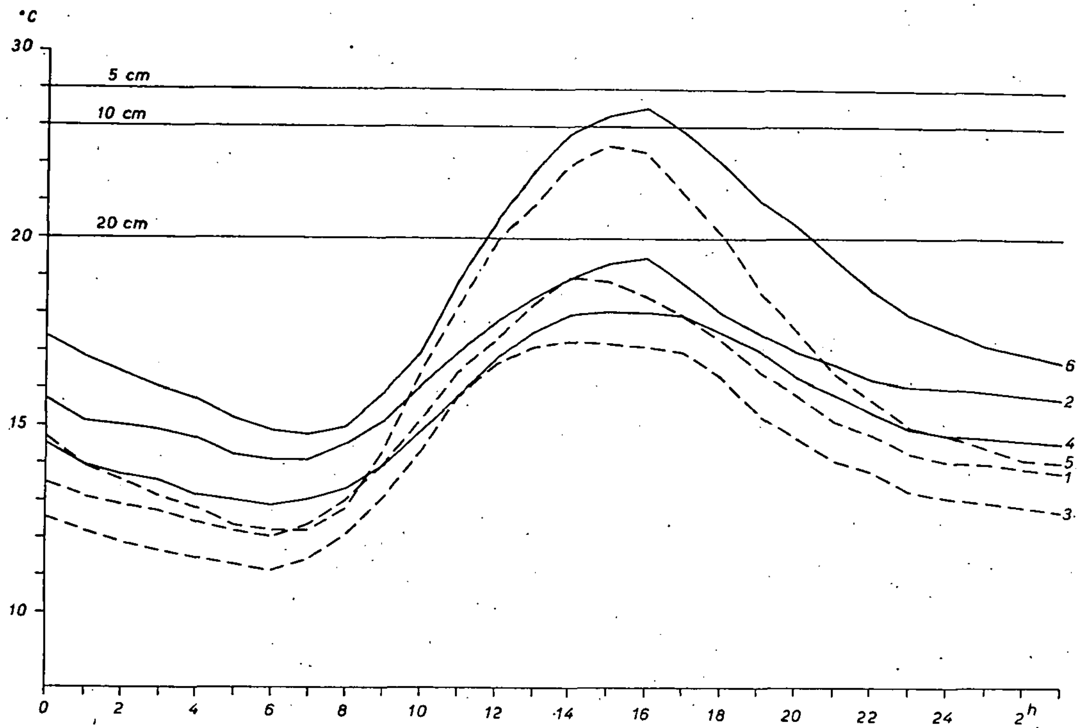

Fig. 8. Diurnal variation of the temperature of the soil of a rice field covered by flooding water of various depths in the surface-near and deeper (5—10 cm deep) layers at Szarvas on a cloudless day in September (Sept. 11) after draining of the flooding water (1975).
Key: 1 = S_f layer; 2 = S_t layer; 3 = K_f layer; 4 = K_t layer; 5 = M_f layer; 6 = M_t layer.
The horizontal lines represent mean soil temperatures measured at 5, 10, and 20 cm depths at the agrometeorological station in the first ten days of September.

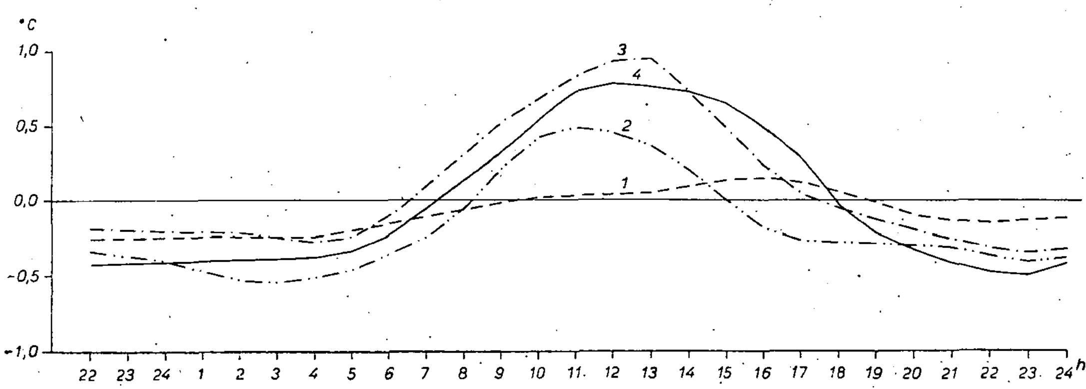

Fig. 9. The hourly variations of the soil temperature in the M_t layer in the case of global radiation of various intensity. Szarvas, July 1975.
Key: 1 = 130 cal cm^{-2} day^{-1}; 2 = 362 cal cm^{-2} day^{-1};
3 = 611 cal cm^{-2} day^{-1}; 4 = 719 cal cm^{-2} day^{-1}.

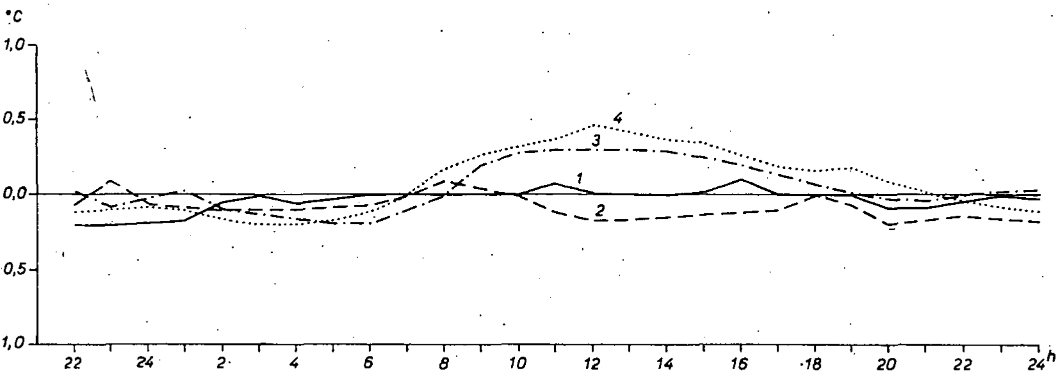

Fig. 10. Hourly variations of the soil temperature in the S_f layer in the case of global radiation of various intensity. Szarvas, August 1975.
Key: 1 = 87 cal cm^{-2} day^{-1}
2 = 319 cal cm^{-2} day^{-1}
3 = 568 cal cm^{-2} day^{-1}
4 = 614 cal cm^{-2} day^{-1}

The marked daily trend of soil temperature in June and September is the consequence of more sparse plant stand and the lack of water cover, while the considerable deviation of the temperature values follows from the decreased global radiation.

In contrast to the previous situations, in July and August the difference of the soil temperature of the night- and day-hours gradually decreases. This can be seen particularly well from the daily trends of the soil temperatures of July. In our Fig. we compared the soil temperatures of a perfectly clear day (15. VII), an overcast day (1. VII) and the soil temperatures of average hourly mean values. In July — as a consequence of the dense plant stand (overshading effect) the temperature values of even the perfectly clear day remain below the average decade mean value of the 5 cm layer of the dry soil.

The soil temperature values of the clear day and the average overcast day run almost parallelly with a difference of about 2—3 °C. On the other hand, the daily soil temperature values of overcast days progress with larger deviations almost completely smoothed deeply below the decade mean values of the dry 20 cm layer.

In order to take into consideration the effect of the global radiation influencing the trend of daily soil temperature we determined the daily trend of the soil temperature values appearing hourly in the soil temperature trend of days with different global radiation (for the case of the soil layer covered by shallow flooding water in July and August) (*Fig.-s 9 and 10*).

It can be stated that the variation of the hourly values of soil temperature is the highest in July during the day. The variation of the hourly values of soil temperature is during the day from 7-8 h to 12-13 h gradually increasing and from 13 to 17-18 h still increasing although in a gradually decreasing measure. After attaining the daily temperature maximum at about 17-18 h [22] an hourly temperature decrease (of equal measure almost during the evening and the night) in the trend of the temperature of the soil layer near the surface covered by shallow flooding water. The largest amplitudes do not appear on the days with maximum global radiation but in the case of a somewhat smaller radiation amount. At the same time, the amplitude is very insignificant on overcast days obtaining less energy (*Fig. 9*).

The plant stand becomes increasingly dense in August, and the almost connected unbroken panicle level strongly overshades the water surface decreasing thus the energy coming to the flooding water and the soil, so the daily trend of soil temperature shows — even on clear days — an almost straightened curve (*Fig. 7*).

In Fig. 9 (showing the variation of the hourly values of soil temperature) we can see also the shading effect of the plant stand shutting out the radiation. On a clear day with an energy income of more than daily 600 cal $cm^{-2}d^{-1}$ the increase of the temperature of the soil in only 0,2—0,3 °C hourly. On account of the overshading effect of the plant stand in August even the maximum values of the soil temperature remain below the August average values (22 °C) measured at the level of 20 cm in the soil without water cover (*Fig. 7*).

But in the first half of September, when the flooding water, is drained off again a certain marked daily trend (although of a smaller amplitude) can be seen (*Fig. 8*) because the dried up soil surface comes into direct contact with the air warming up to 25—30 °C during the day. Thus the agrotechnical method applied influences the growth rate of the plants by changing the climate of the rice stand, stimulating accelerating the full development of the plant and the ripening of its seeds.

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Data Concerning the Influence of Climate and Human Activity on the Dynamics of Salts in the Region East of the River Tisza

by

M. Dzubay and J. Juhász

Adatok a klíma és az emberi tevékenység befolyásáról a talajok sódinamikájára a Tiszántúlon. Az emberi tevékenység és a talajok sódinamikája közötti kapcsolat vizsgálatát tiszántúli kötött szikes talajokon természetes (öntözés nélküli) viszonyok között folytattuk.

A mintegy 2000 talajminta analízise alapján a következőket állapítottuk meg:

1. A talajokban történő sóvándorlást a csapadék, a párolgás és a talajvízszint a talaj mechanikai és kémiai összetételétől, valamint a mélységtől függően egymással összhangban irányítják.

2. A „kritikus mélység" körüli talajvízállásnál a mélyművelés, a 60 cm mélységre történő talajlazítás sómobilizáló befolyása nagyobb, mint a sekély talajművelés, a 15 cm-es szántás sómozgató képessége.

3. Hatásban az előbbieket a kémiai és fizikai (műanyaghab) javítás követi. Ezek befolyása elsősorban arra a talajrétegre szorítkozik, amelynek megjavítására szolgált.

4. A kémiai és fizikai javítás a befolyásuk alatt levő talajrétegekben a sók mozgását struktúra javító hatásuk következtében fokozzák, míg a javítatlan talaj csökkenti.

5. A fizikai javító anyagok, a poli-stirol, a poli-uretán habok a kémiai javító-anyagok hatását még növelik.

6. A "kritikus mélység„ feletti zónában levő talajvízszint a feltalaj sómozgását helyzetének megfelelően — a csapadékkal és a párolgással összhangban — irányítja. Magas talajvízállásnál a sómaximum feljebb, alacsony talajvízállásnál pedig lejjebb található.

7. A szikes talajok tulajdonságától függően javításukat talajlazításnak kell megelőznie. Az oldható sók kilúgozása ezáltal fokozódik, s ez a *talaj javulását nagy mértékben segíti.*

Angaben zum Einfluss des Klimas und der menschlichen Tätigkeit auf die Salzdynamik der Böden in dem Gebiet östlich von der Theiss. Von den Verfassern wurde eine Untersuchung bezüglich des Zusammenhanges zwischen der menschlichen Tätigkeit und der Salzdynamik der Böden durchgeführt, undzwar in der Region jenseits des Flusses Tisza in sodahaltigen Böden unter natürlichen (unberieselt) Umständen.

Auf Grund der Analyse der etwa 3000 Bodenproben kann folgendes festgestellt werden:

1. Die in den Böden vor sich gehende Salzmigration wird vom Niederschlag, Verdunstung und Grundwasserniveau, in Abhängigkeit von der mechanischen und chemischen Zusammensetzung des Bodens, sowie von der Tiefe im Zusammenhang miteinander determiniert.

2. Bei einem um die „kritische Tiefe" liegenden Grundwasserstand ist der salzmobilisierende Effekt der tiefen Bodenbearbeitung, die bis zur Tiefe von 60 cm hinabreichende Bodenauflockerung grösser als die salzbewegende Fähigkeit der seichten (bis zu 15 cm reichenden) Pflugarbeit.

3. Bezüglich der Effektivität folgt nach den obigen die chemische und physikalische (Kunststoffschaum) Amelioration. Ihr Einfluss beschränkt sich vor allem auf die zur Amelioration vorgesehene Grundschicht.

4. Von der chemischen und physikalischen Amelioration wird in den von ihnen beeinflussten Grundhschichten die Salzbewegung, infolge ihrer Strukturverbessernden Auswirkungen gesteigert, während sie vom nicht ameliorierten Boden wegen dem Mangel der obenerwähnten Einflüsse vermindert wird.

5. Von den physikalischen Ameliorationsmaterien — Poly-Stirol- und Poly-Uretanschaum — wird die Wirkung der chemischen Mittel noch gesteigert.

6. Das sich in der über „kritischen Tiefe„ liegenden Zone befindliche Grundwasserniveau determiniert gemäss seiner Lage und im Einklang mit dem Niederschlage und Verdunstung die

Salzmigration des Obergrundes. Bei einem höheren Grundwasserstand liegt das Maximum höher, bei einem niedrigen niedriger.

7. Bei natronhaltigen Böden muss, in Abhängigkeit von ihrer Beschaffenheit, bevor ihrer Amelioration eine Bodenlockerung vorgenommen werden. Die Auslaugung der löslichen Salze wird hierdurch gesteigert, was zu Amelioration des Bodens sehr bedeutend beiträgt.

In the regions beyond the river Tisza investigations were carried out on connections between human activity and the salt dynamics of unirrigated alkali soils.

On the basis of the analysis of about 2000 soil samples the following could be stated:

1. The salt migration in the soils are determined by the precipitation, evaporation and groundwater level in concordance with each other and depending on the mechanical and chemical composition of the soil and also on the depth.

2. With a groundwater level about the „critical depth" the salt mobilizing effect of deep tillage, the subsoiling to the depth of 60 cm is more intensive than salt mobilizing effect of shallow tilling, the ploughing to 15 cm depth.

3. As to the effectivity the above factors are followed by chemical and physical amelioration (plastic foam). Their influence is restricted mainly on the layer envisaged to be ameliorated.

4. Chemical and physical amelioration increase the salt migration in the respective soil layer (on account of their structure-ameliorating effect) while the unameliorated soil will, in want of such effects exert a decreasing influence.

5. The physical ameliorating material — poly-stirol and poly uretan foams — will even contribute to the effect of the chemical factors.

6. The salt migration of the surface soil is determined — in accordance with its position — by the groundwater level situated in the zone above the „critical level" (in conformity with the precipitation and evaporation). With a high groundwater level the salt maximum is to be found higher and with a lower one lower.

7. Depending on the qualities of alkali soils their ameliorations is to be begun with subsoil work. The leaching of soulbe salt will be thus increased which greatly contribute to the amelioration of the soil.

The accumulation of water-soluble salts in the soil results in crop failure. The main cause of the decrease of crop is the excessive amount of sodium ions in the soil solutions and chemical bond. This is, unfortunately a rather frequent phenomenon in this country, owing to its basin-character and other conditions. Alkali soil amounts to about 1/2 million hectares in Hungary and this is why steps must be taken against the alkalisation of additional territories and ameliorations have to be undertaken in the existing soils.

Under average climatic conditions the migration of soil salts takes place periodically also in this country. In winter the soil salts are washed out from the soil profile while in summer they will accumulate there.

The dynamics of soil salts are determined, under unirrigated conditions — by three factors:

1. precipitation,
2. evaporation,
3. ground water level.

With the progress of civilization the soil dynamics is more and more influenced by man as a soil forming factor. Such human activity is e.g.:

a) the cultivation of soil,
b) chemical and physical soil amelioration.

Introduction

In the formation of alkali soils of the region beyond the river Tisza the main role, from among the soil forming factors was played by the climate and geological factors. The first one includes the amount and distribution of precipitation, the water vapour content of the air, wind conditions and temperature. To the second group belong the conditions of the soil, its mechanical texture, the salt conditions and hydrological conditions.

Thus from the aspect of salt migration and alkalisation the groundwater level is a criterion.

Groundwater tables under 4 metres are practically of no effect whatever on the upper soil layer.

On the other hand, — depending on the texture of soil and the salt content of the groundwater the surface evaporation may already be nourished by the capillarity zone lying at a water-level of about 1—3,5 m. Thus in this case the soil may be alkalinisated by the concentration of dissolved salts and an accumulation of salt may arise [9, 15, 8].

The depth from which the water soluble salts may get to the surface or near to the surface by capillary rising is called "critical ground water level".

The salt dynamics of the soil has been in detail investigated by *Darab* [4] and *Szabolcs* [12] concerning the region beyond the river Tisza, while that of the region between the rivers Danube and Tisza by *Várallyay* [14].

Methodics

Determination of the total salt percentage of soil samples taken by 10 centimetres from parallelly drilled profiles of soil. Statistical evaluation of them according to natural factors, — precipitation, ground water level, and also according to human intervention, — soil-amelioration, subsoiling [5].

Investigating of connections between the fluctuations of water table and the migration of soil, on the basis of reading of (by 14 days) the groundwater registrating wells.

The percentage of total salt has been determined on the basis of the electric conductivity of soil paste. The investigations were carried on in heavy sodic soil in parcels of 50—60 m^2 in the region beyond Tisza.

In our observation series of 180 years no precedent has been found to the repetition of the value of any climatic element in any one of the years. However, certain laws can be found if comparing their connections with each other and the factors forming the climate.

In the course of the past decades a considerable development has been achieved in climatology through the scientific results in the field of singularity, macrosynoptic processes, statistics made on the basis of the air-masses' calendars and the isoplethe representation of the annual fluctuations of the different elements. [3, 10, 2, 1].

In the present short work the authors do not intend to deal in detail with the individual climatological elements since that would lead to a deviation from their task, so that only precipitation and evaporation are investigated being factors of the first order from the aspect of salt migration in the soil.

In this country — like in many other ones the climatic elements are generally satisfactory for the requirements of plant cultivation. It is proved by many years' experience that the production results of agriculture is decided mainly by the conditions of precipitation. However, for getting a complete survey of precipitation conditions the conditions of the other climatic elements too, must be clearly seen, and, in addition, also the connection between these climatic elements and the precipitation. The evaluation and application of the results to be found in older references is not undertaken here.

1. Charts representing the territorial distribution of precipitation, both for the whole year and for the summer half year, are generally known. From them one may read off the distribution of precipitation — as determined from observations up to the present. On the basis of the measurements of the stations it has been found that the dryest part of Hungary (with a precipitation amount of less than 500 mm) is Hortobágy, further a region of the length of about 50 km and the width of 20 km extending from Szolnok to Szarvas, and also the environment of Kunszentmárton. In the rest of the country the precipitation amount is more favourable. At the borders of plains a rise of the precipitation amount is to be observed and so the isohyethe can be found mainly along the bordering lines of the lowland and the mountains or hilly regions respectively. In the Great Hungarian Lowland, along the upper reach of the river Tisza and eastwards from there amounts of more than yearly 600 mm are to be found, obviously as an effect of the proximity of the Carpathian Mountains. In the regions of the mountains Börzsöny, Mátra, Bükk and Zemplén the annual amount is exceeding 700 mm. A similar situation is to be found in the south counties of Transdanubia with even more than annual 800—900 mm in some places like Farkasgyepű, Bakonybél, Borzavár, Sopron and Kőszeg.

The air-lifting effect of the hills and mountains results in an increase of the precipitation amount at the higher level since the horizontal air currents are forced into the height. Ascending air masses cool down at a rate of 1 ° per 100 m and so, depending also on other circumstances, the condensation process will take place more rapidly, the cloud formation will increase and so of course the precipitation. Similar conditions do not occur in lowlands. The configuration of the terrain is of such an importance that in some territories the precipitation conditions will be generally a function of the height above sea level. The connection between precipitation and the height above sea level has been proved, on the basis of 50 years' average by Hajósy (Table 1.)

Table 1

Connection of the annual amount of precipitation with the height above sea level (after Hajósy)

Height m	*100*	*150*	*200*	*300*	*400*
Great Hungarian Plain	545	560	—	—	—
Kisalföld	580	620	—	—	—
Transdanubia	650	670	690	700	720
Northern Mountains	545	575	590	650	700
	560	600	650	680	710

The difference may be explained by the different properties of the air of the individual regions and by the special variations of the air masses streaming there. The interaction of the local and arriving air masses varies from region to region. In the eastern territories continentality is prevailing.

The seasonal variations, the yearly tendency of precipitation, storms and hails, the daily march of precipitation and also the evaluation of snow-falls and passages of front have been neglected.

The number of days with precipitation — the averages of many decades — area, as a result of the work of Hajósy, at our disposal concerning the whole territory of the country and thus the frequencies of daily precipitation amounts attaining and going beyond 1 mm, 5 mm, 10 mm and 20 mm are also known. On the basis of the data concerning the frequencies of days with precipitation it can be established, according to 50 years' average that in the course of a year 120—160 days with precipitation can be expected. The most days with precipitation occur in December and the least in July to September. Generally at the end of winter a second minimum and in spring a second maximum can be observed while the autumn maximum appears sometimes only in December. When analysing the singular precipitation tendency of the calendar days it may be seen that the precipitation-probability of the individual days shows values in the beginning of summer (June) rivalling those of December. In the group of precipitation up to 1 mm in the spring maximum even that forges ahead; the frequency distribution of days with abundant precipitation shows yearly one important wave with a May—August maximum and a January—February minimum, which follows from the smaller vapour capacity of the cold air and partly also from the development of summer and winter monsoons.

As to the variation of the conditions of soil surface only some commentaries are given. Dry soil surface can be expected in yearly averages in 180 days, wet in 100—110 days and surface covered with snow and ice in 50—60 days. Dry surface occurs the most frequently in the second half of summer, wet one in autumn and those with snow and ice in January. The autumn frequency of wet surface indicates that the energy of insolation and the snow stored in the soil are not sufficient for the evaporation of the moisture of the surface while in summer even more than the disponible amount could be evaporated from the surface.

2. Within the theme of evaporation the circumstances and factors are to be shortly analysed causing a decrease (in mm/time unit) of the level of a water layer with a free surface.

The instruments of *Wild* and *Piche*, used up to the present in climatology are not suitable for determining the water amount actually evaporated from the surface: the values obtained by the instruments express the evaporating disposition of the existing atmospheric conditions and atmospheric processes respectively.

The evaporation, if considered as an amount, is of microclimatological character because it depends on several circumstances the complex of which show differences almost at every cm² of the surface of the soil. Influencing factors in the evaporation are: water temperature, air temperature, humidity, radiation and wind. In addition to the above, there is one more factor in the actual evaporation: the presence or absence of the evaporating water, thus giving a proof of its microclimatological character. The actual evaporation is measured with lysemeters but its determination, climatological evaluation, is more complicated than that of evaporation. Thus instead of measurements, attempts were undertaken to determine the actual evaporation by the aid of calculations and estimates.

Measurements of the evaporation are carried on in this country since 1875, by using *Wild's* evaporimeter. However, on account of the different microclimates of the meteorological stations the series of measurements are not homogeneous and thus unsuitable to give a general survey for the country. The analysis of such series enable us only to investigate the annual march (Table 2).

From the data of evaporation doubtless air temperature and relative air humidity are the influencing factors.

When comparing the effect of wind and humidity it can be seen that the evapotranspiration of April is almost the double of that of October although the mean temperatures of those two months are almost identical. From this it appears that the devia-

Table 2
Evaporation amounts in mm-s in the average of 16 years (1929—1944).

	I	II	III	IV	V	VI	VII	VII	IX	X	XI	XII	Year
Kecskemét	11	19	48	70	86	95	103	82	61	41	23	14	653
Túrkeve	11	15	35	55	67	71	83	70	51	34	15	14	521
Debrecen	11	17	37	65	81	83	90	77	54	38	21	14	588

tion of the evaporation capacity may be caused by the difference of air humidity and wind.

Considerable efforts have been made to determine the evaporation by means of computations. So we have the equations of *Bacsó* and *Ubel* but they are valid only for a given place and for the given instrument. For another places and instruments empirical constants are required.

A most simple and practical method for the measurement of the actual evapo transpiration and for the determination of the water amount of the soil has been elaborated by *Dunai, Posza,* and *Varga.*

The factors influencing evaporation are: the disponible water (w), the porperties of the evaporating surface (water, soil, plant) (F) and the evaporation capacity of the air (E_0).

$$E = f(w, F, E_0)$$

Under conditions when two of the above three factors do not change, the third one can be determined. Thus E_A the evaporation of the pan "A":

$$E_A = f(E_0)$$

In measurements the evaporation pans type "A" are used all over the world so that comparisons can be made between the results obtained by different countries.

In the course of the investigations carried out in Szarvas the authors have shown that the evaporation values of pan "A" can be computed on the basis of air temperature and saturation deficit.

The air temperature has been substituted by the saturation vapour pressure (E) and the saturation deficit ($E-e$) expressed as follows:

$$E_A = \frac{E(E-e)}{E+(E-e)} \text{ mm/day}$$

Thus the evaporation capacity may be computed:

$$E = \frac{1-f}{2-f} t$$

$f = \frac{e}{E}$ ("e" is the actual -, "E" the saturation vapour pressure)

E_0 evaporation capacity of the air,
E_A quantity of the evaporated water
E evaporation value

All of the above three factors are practically of the same value. Having at our disposal the percentile values of relative humidity and the values of mean temperature the evaporation can be determined in the form of mm/day.

3. With a view to evaluate the investigations concerning salt-dynamics wells have been built for the registration of ground water level. These wells can be divided into two groups. The first three ones (No-s 1, 4 and 7) are situated at a distance of about 500, resp. 850 metres from an artificial fish pond. The other two wells (No-s 12 and 13) are at a distance of 50, resp. 100 m from the abovementioned pond.

The water-level variations of these belonging to the first groups are more equal, they show less oscillations. At the time of the observations they were about 2 metres. The variations of the water level were directly not influenced by the pond.

In the case of the nearer registration wells the case is different. The mean water level of them was about 1/2 m higher and that was to be ascribed to the presence of the fish pond (*Fig. 1*).

The water level of the well being nearest to the pond (See *Fig. 1.* No. 12, within 50 m); line marked with triangles) depends on the water level of the fish pond. That manifested itself in an increase of the water level of the registration well after filling up the artificial pond with water of after abundant rains (*Fig. 1*).

In contrast to well No. 12 the water level variations of No. 13 (situated somewhat farther: within 100 m) were influenced first of all and more considerably by precipitations above 10 mm. This may be seen in *Fig. 1* (line marked with circles) where the sudden rise of the precipitation diagram is exactly followed by the rise of the groundwater level.

In the oscillations of the registrating wells No. 1, 4 and 7 the effect of the abovementioned large rains too, can be observed, apart from the seasonal variations determined by their position. Of course in a smaller measure than in wells with a high water level.

The seasonal water level oscillations are more characteristic in the wells situated farther from the pond (No-s 1, 4 and 7). In this case the phenomenon could not have been disturbed by the water in the pond and the precipitation [7].

The somewhat differring water level of wells No. 1, 4 and 7 comes from the different mechanical composition of the respective soil layers.

In order to examine the effect of the abovementioned human activity the investigations of salt dynamics in alcalic soil were divided into two parts:

Observations A) *concerning the groundwater level about the "critical depth"*;
Observation B) *observation of salt migration in depthshigher than the critical one.*

A. Analyses with groundwater levels near the critical depth

The salt migration examinations (shown in *Fig. 2*) were carried out in the surroundings of wells No. 4 and 7. Between their water level oscillations and the salt migration taking place in the soil profile the following connections could be found:

The salt amount of the soil samples (taken to the depth of 60 cm), i.e. the salt migration taking place in the soil profile, far from the fish pond was influenced above all by the soil cultivation. By the loosening of the soil executed every 4 years to the depth of 60 cm a better desalinization was achieved (*Fig. 2.* See graph-series below) than by the shallow loosening, the annual ploughing of 15 cm (*Fig. 2.* See: graph series above).

After the soil amelioration the next important factor of salt mobilization (from among the three main factors of salt migration, i.e. precipitation, evaporation, groundwater) was the precipitation. This statement is corroborated by the situation to be read out jointly from *Fig.-s 1* and *2*) 19 July 1973. Because of the season (summer) the salt maximum should have been nearer to the soil surface. Its situation is however, nearly identical with the conditions of 15 March 1974 (*Fig. 2.* Comparison of the drillings No.-s 130—134 and 190—194). The equality of salt content was brought about in this case by the leaching effect of the large precipitation amount (14—18 mm, and 130 mm in all) fallen during the previous 11/2 years on seven occasions (*Fig. 1.*). Thus the precipitation had in the case of the situation of 19 July 1973 a different role which may be separated from the groundwater of about 2 metres and from the soil amelioration to be discussed later.

In the investigated period the role of the groundwater level came only after those of the abovementioned factors from the aspect of the formation of salt profiles.

When comparing the salt-curves of the samples taken at the same time (See: *Fig. 2*) it becomes clear that the profiles of the soil ploughed yearly (to 15 cm) contain, apart from chemical amelioration, generally more water soluble salts than the soil loosened to the depth of 60 cm, resp. the soil having been ameliorated by plastering even at the bottom of the furrow. Thus the chemical amelioration had on the soil layer investigated to 60 cm a smaller effect, from the aspect of salt migration, than the deep tillage, the soil-loosening to the depth of 60 cm. As to effectivity it comes only after the latter. In the further investigations it will become even clearer that the desalinizating effect of the chemical soil amelioration is limited only to the upper layer of 20 cm and to the bottom of the furrow respectively, i.e. mainly to the layer obtaining the most part of the chemical materials. This is proved also by the analyses of exchangeable sodium.

The demonstrated salt-migration investigations are not significant, they show only tendencies, because the heterogeneity of the soil, the statistical deviation of the analyses is larger than the difference of the deviations between the treatments. This is a consequence of, a deficiency in the method of taking the samples, since the samples were not taken at the starting of the experiment but only 10 years later.

However the data of salt migration taking place as an effect of soil cultivation, their trends are nevertheless considered as reliable since the analyses corroborate the following statement of Sigmond [11]: "If by some reason, and if even temporarily a decrease of the groundwater level is observed (or the surface salt water is drained) a strong leaching process will take place, the first phase of which is a decrease of the amount of water soluble salts (below 0,10—0,15 %) while the soil absorption complex will not yet loose from its Na content.

The abovementioned leaching process may be even promoted by loosening the soil. In this case we open the way to, and accelerate the process of leaching, the removal of the soluble salts from the upper soil layer.

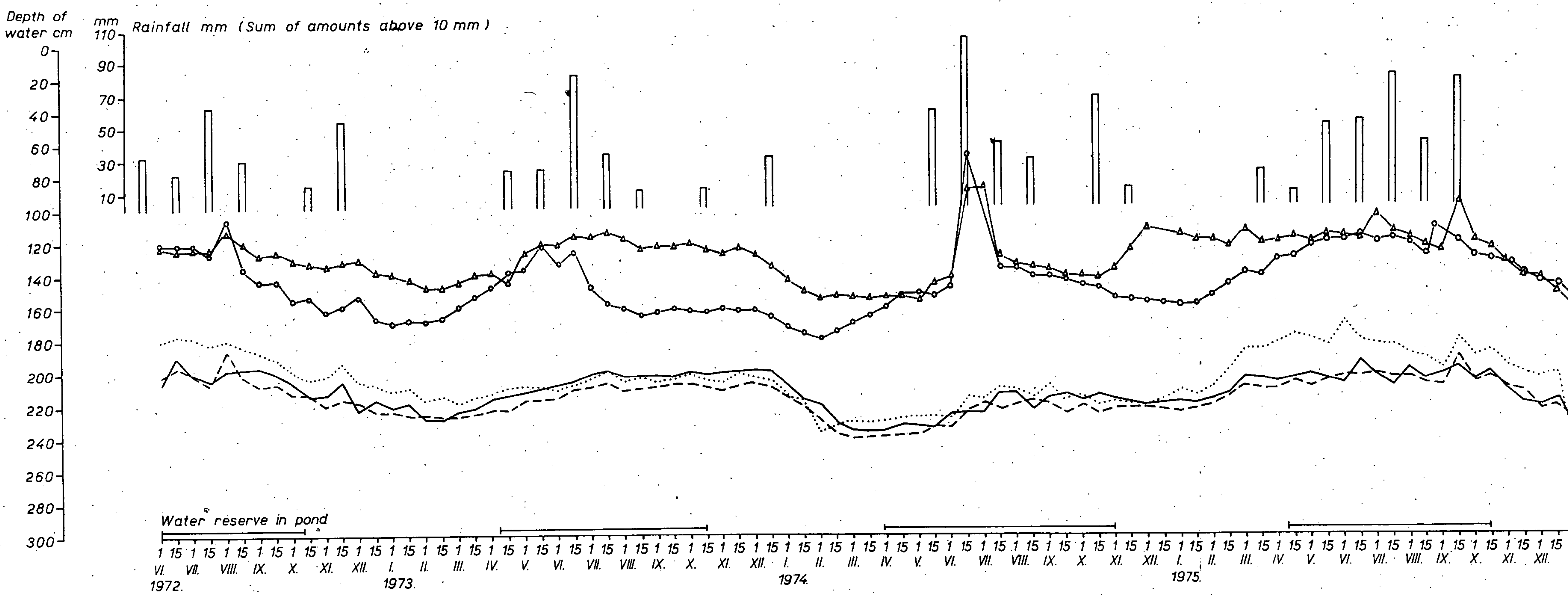
When carrying out chemical soil amelioration the first step is (depending on the characteristics of the soil) the subsoiling of the alcalic soil opening the way in this way to the disappearance of the salt. On the basis of the law of mass-effect the amelioration is better and more effective in this case.

The chemical causes of this phenomenon have been analysed in our previous paper [6].

On the basis of the above the authors of the present paper are in disagreement with the principle followed by *Herke* [13] in determining the exchangeable ammonium carbonate Na, and approve the method of *Mados*. In our opinion the latter does not involve any "overdosing" of amendment given to the respective soil layer.

Fig. 1

Depth of water in groundwater tablecontrolling wells and rainfall



Symbols: —— well No. 1 (550 m from the pond) - - - - well No. 4 .... well No. 7 (850 m from the pond) —△—△— well No. 12 (within 50 m) —o—o— well No. 13 (within 100 m)

Fig. 2 Variation of the total percentage of the agricultural experiment

15 cm deep annual plowing

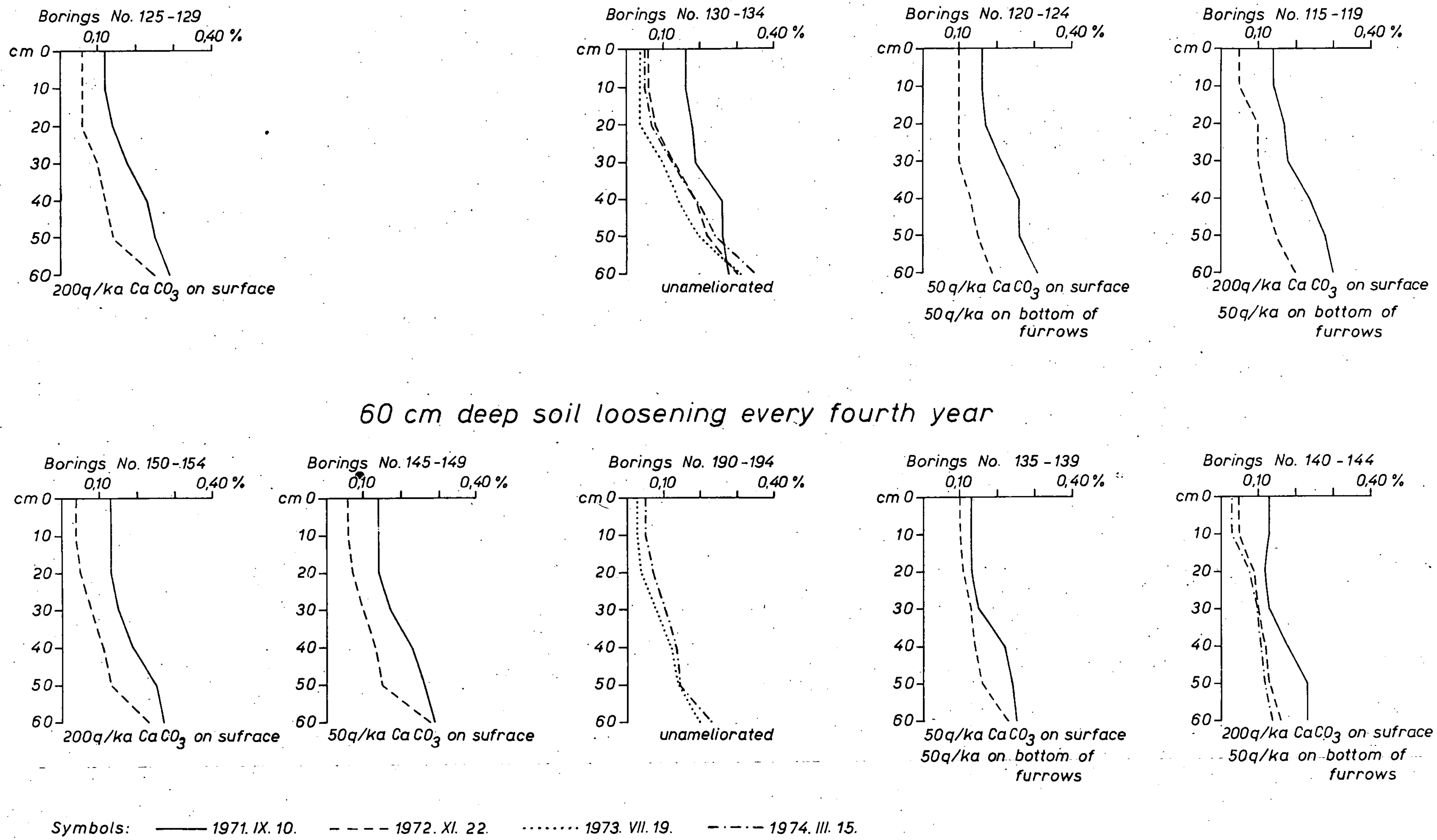
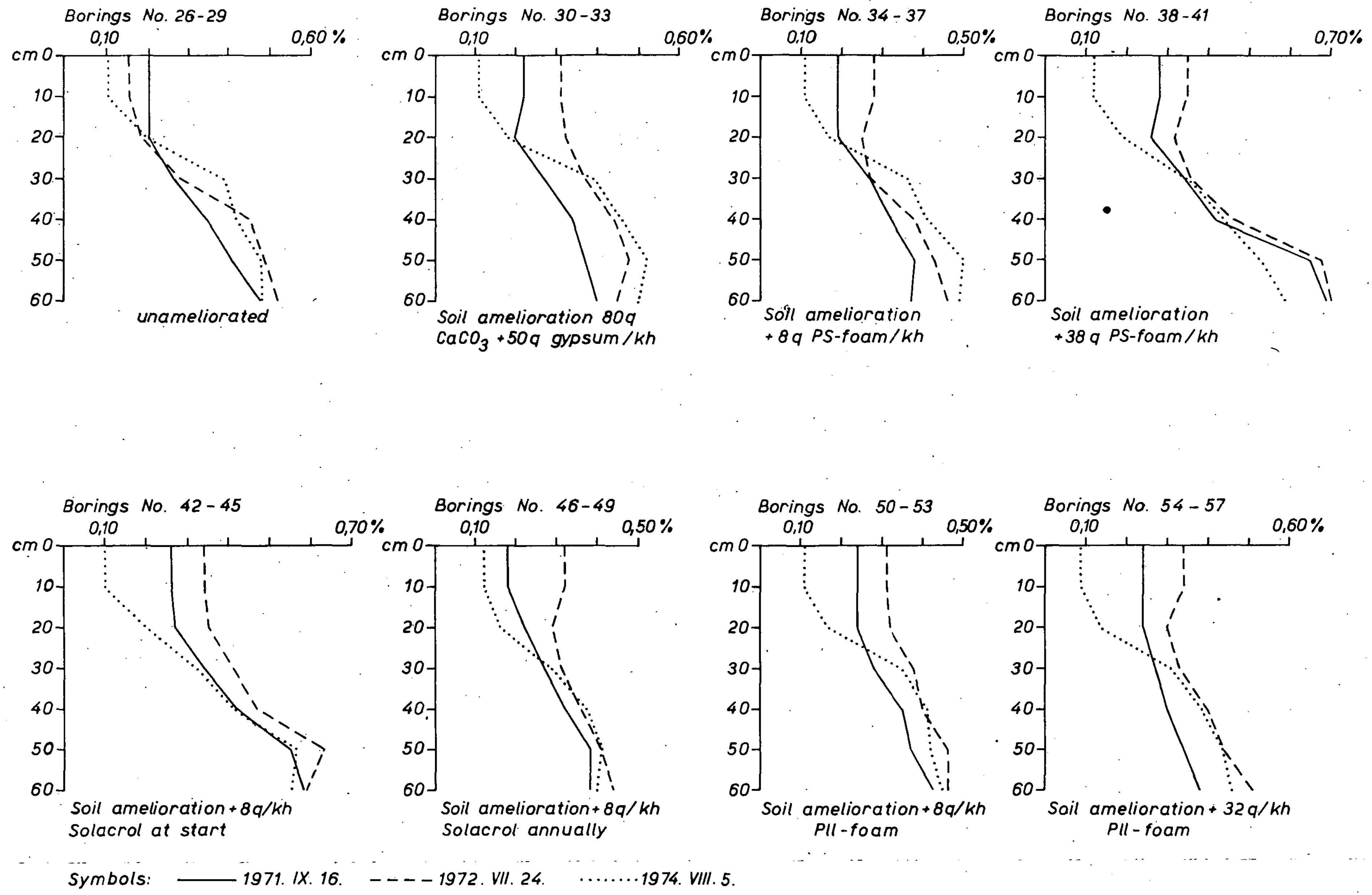


Fig. 3 Variation of the total percentage of the agricultural experiment



B. Examination of salt dynamics analysed under circumstances of a groundwater level above the „critical depth"

The examined soil amelioration experiment was carried out between the fish pond and the registration well No. 13.

From *Fig. 3* and the data contained therein the following determinations can be made:

The obvious difference between the ameliorated soil (drillings No. 30—57) and the non-ameliorated one is that in the ameliorated soil the movement of the water-soluble salts, their quantitative variation, is much more important than in the original unameliorated one. This difference is characteristic mainly for the layers between 0—10 and 10—28 cm. Thus it can be stated that in unameliorated soil the increase of total salt and its decrease respectively, is 0,1% while in the ameliorated one 0,17—0,25%. Between the two ways of tillage the difference is significant.

The cause of this phenomenon is the fact that the ameliorated soil structure furthers the movement of the water (and the salt in it) much better than the unfavourable structure of the untilled alkalic soil.

This is corroborated also by the significant difference of salt percentages to be found between cases of tillage where 8 quintals of poly-stirol (PS) or poly-uretan (PU) and 32 quintals of the same materials were applied (*Fig. 3*).

It can be stated also that this intensive salt movement extended in the unameliorated soil mainly over the layer of 0—10 cm while in the ameliorated one over the layer of 0—20 cm.

The investigated material was collected from the soil-profile of 0—60 cm. The movement of water soluble salts was influenced, in addition to the above, most considerably also by the precipitation (evaporation) and by the groundwater level: by the latter one for the simple reason that it is no rare case to find the groundwater at a depth of about 1 m (*Fig. 1*). In such a case the salt migration is, in contrast to the water level of the registration well No. 13, determined rather by the groundwater level than by precipitation. The correctness of this observation is proved by the larger salt content of the soil samples taken on 24 July 1972 (*Fig. 3*, broken line, and Fig. 1, water level of 105 cm) and the significantly smaller salt content of the samples collected on 5 August 1974 (*Fig. 3*, dotted line, and *Fig. 1*, water level of 138 cm).

In the investigated two cases considerable precipitation fell at that time on the territory: in the first case 65 mm and in the second one by 25 mm more than the quantity of 104 mm of the previous month (*Fig. 1*). In spite of that the salt content of the soil profiles is rather a proof of the influence of the groundwater level than of the leaching effect of the large rains. The salt content of the soil profiles is larger with a higher groundwater level (*Fig. 1*, on 24 July 1972) (See *Fig. 3*, broken line) and smaller with a lower water level (*Fig. 1*, 5 August 1974 and *Fig. 3*, dotted line). Obviously in the latter case the leaching effect too, will manifest itself as it can be readily seen on the upper stages of the salt curves.

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