IDŐJÁRÁS

QUARTERLY JOURNAL
OF THE HUNGARIAN METEOROLOGICAL SERVICE

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VOL. 104 * NO. 3 * JULY–SEPTEMBER 2000

IDŐJÁRÁS

Quarterly Journal of the Hungarian Meteorological Service

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IDŐJÁRÁS
Quarterly Journal of the Hungarian Meteorological Service
Vol. 104, No. 3, July–September 2000, pp. 143–159

Microscale bare soil evaporation characteristics: A numerical study

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(Manuscript submitted for publication 17 January 2000; in final form 11 May 2000)

Abstract—Soil evaporation characteristics on spot (few hundred meters × few hundred meters) and patch scale (few kilometers × few kilometers) are analysed comparing a *deterministic* and a *statistical - deterministic* model. Soil evaporation characteristics are considered in terms of analysing evaporation curve (evaporation/soil moisture content dependence), areal distribution characteristics of soil evaporation and the aggregation algorithms for their estimation. The analyses are performed for different soil surface resistance parameterizations and atmospheric forcing conditions. The main results obtained can be summarized as follows: (1) The scale-invariance of soil evaporation depends upon both the parameterization of soil surface resistance and the soil surface wetness regime. (2) The relative frequency distribution of soil evaporation depends upon both the parameterization of soil surface resistance and the areal mean soil moisture content θ_m. Finally, the relationship between the aggregated soil moisture content θ_{ag} and the θ_m is linear and it does not depend upon atmospheric forcing conditions. This latter result plus the *deterministic* model can be applied for estimating patch scale soil evaporation.

Key-words: soil evaporation characteristics, spot scale, deterministic model, patch scale, statistical-deterministic model, areal evaporation.

1. Inroduction

Areal evapotranspiration is one of the most relevant factors in hydrology and meteorology. Its basic features are as follows: it shows high spatial and time variability at all scales and it is non-linearly related to environmental conditions. Concerning these features the basic questions are: Is evapotranspiration a scale-invariant quantity? Under which conditions is it scale-invariant? To study these questions it necessary to analyse all basic aspects of areal evapotranspiration among other things like:

(1) the course of $E(\theta)$ curve which expresses the dependence of evapotranspiration upon soil moisture content,

(2) the areal distribution characteristics of $E(\theta)$, e.g., the relative frequency of $E(\theta)$ and

(3) proposing aggregation algorithms for its estimation.

Such and similar aspects are investigated mainly by modeling, performing comparative numerical studies (e.g., *Rodriguez-Camino* and *Avissar*, 1999; *Li* and *Avissar*, 1994; *Mölders* and *Raabe*, 1996; *Kabat et al.* 1997; *Ács*, 1996; *Ács* and *Hantel* 1997). Many studies (e.g., *Rodriguez-Camino* and *Avissar*, 1999; *Sellers et al.*, 1997; *Wood*, 1997) refer to the landscape scale (about 10 km × 10 km). The attention is payed mainly to the aggregation algorithm problems. Two aggregation algorithm types are used: the so called "deterministic" one (e.g., *Lhomme*, 1992; *Lhomme et al.*, 1994; *Chehbounni et al.*, 1995; *Shuttleworth,* 1997) where the surface heterogeneity is considered via interpatch variability using weighting rules and the so called "statistical" one (e.g., *Famiglietti* and *Wood,* 1991, 1994; *Entekhabi* and *Eagleson,* 1989, 1991) where the surface heterogeneity is considered via intrapatch variability. In these approaches patch is not homogeneous; the patterns of variability are represented statistically via *probability density functions*. In some studies it is also noted (e.g., *Wood*, 1997) that the scale invariance of $E(\theta)$ depends upon the "state of the system". Further, the analyses refer mainly to the surface covered by vegetation.

The quantification of fine scale heterogeneity remains a clear and high priority (*Shuttleworth,* 1996). In spite of this, there are only a few studies dealing with small-scale heterogeneity (e.g., *Wood*, 1997). They use "statistical" approach applying either time-consuming numerical integrations (*Famiglietti* and *Wood*, 1994) or more analytical solutions (*Wood*, 1997). It is also recognized that the scale invariance problem is closely related to both the parameterizations used and the environmental conditions. Nevertheless, there is no any study with a detailed analysis of such type especially for bare soil surface.

The objective of this study is to analyse the scale invariance issues of bare soil evaporation at patch scale in terms of $E(\theta)$, relative frequency of $E(\theta)$ and the aggregation algorithm for $E(\theta)$. In the study we assumed that there are no advective effects accompanied by occasionally observed internal boundary layers (e.g., *Garratt,* 1992; *Hupfer* and *Raabe*, 1994) and there are also no mesoscale circulation patterns induced by surface discontinuities. Then the atmosphere can be assumed to be horizontally homogeneous with constant meteorological boundary conditions above a certain level (*Shuttleworth,* 1988).

2. Models

Two model types are used: on the spot scale a *deterministic* model (DM) characterizing the soil evaporation $E(\theta)$, and on the patch scale a *statistical - deterministic* model (SDM) coupling the deterministic model with a statistical one for generating θ as a random variable. The $E(\theta)$ relationship is simulated using two different parameterizations for bare soil surface resistance.

2.1 Deterministic model

The instantaneous value of $E(\theta)$ on spot can be determined by a deterministic model using atmospheric forcing data (net radiation, air temperature and humidity and wind speed) and surface parameters (roughness length, soil texture or soil hydrophysical parameters). The core of model is the Penman-Monteith's concept.

2.1.1 Turbulent fluxes

The latent heat flux is parameterized using *Penman-Monteith's* equation (*Monteith*, 1965) as follows:

$$L \cdot E = \frac{\Delta \cdot A + \rho \cdot c_p [e_s(T_r) - e_r] / r_a}{\Delta + \gamma \cdot (1 + r_b / r_a)} , \tag{1}$$

where L is the latent heat of vaporization, ρ is the air density, c_p is the specific heat of air at constant pressure, $\Delta(T_r)=L\cdot e_s(T_r)/R_{vap}\cdot T_r^2$ is the slope of saturated vapor pressure curve at reference temperature T_r, R_{vap} is the gas constant for water vapor, $e_s(T_r)$ is the saturation vapor pressure at T_r and e_r is the vapor pressure at reference level z_r. r_b is the bare soil surface resistance, r_a is the aerodynamic resistance and A is the available energy of bare soil surface. A is parameterized as

$$A=R - G, \tag{2}$$

where R is the net radiation and G is the ground heat flux at the surface of bare soil. G is parameterized after *Nickerson* and *Smiley* (1975):

$$G=0.15 \cdot R . \tag{3}$$

The sensible heat flux is estimated as residual,

$$H = R - G - L \cdot E \,. \tag{4}$$

2.1.2 Surface resistance

The non-linear relationship between E and θ is expressed using resistance concept. The soil surface resistance r_b is estimated by *Sun*'s (1982) and *Dolman*'s (1993) empirical formulae:

$$\text{Sun:} \qquad r_b^{\text{Sun}} = c_1 + c_2 \left(\frac{\theta_s}{\theta} \right)^{c_3} , \tag{5}$$

$$\text{Dolman:} \qquad r_b^{\text{Dol}} = c_4 \cdot \theta^{c_5} , \tag{6}$$

where θ and θ_s is the actual and saturated soil moisture content. The empirical constants: c_1=30 s m^{-1}, c_2=3.5 s m^{-1}, c_3=2.3, c_4=3.5 s m^{-1}, and c_5=-2.3. Both parameterizations are known and commonly used in the scientific literature (see *Bastiaanssen,* 1995). They refer to all soil textures. Sun's parameterization uses both θ and θ_s parameters while Dolman's parameterization only one, the θ. Note that there is a direct relationship between the θ_s and the soil texture.

2.1.3 Aerodynamic transfer

The aerodynamic transfer is parameterized via resistance concept using Monin-Obukov's similarity theory taking into account the atmospheric stability conditions. We assume that there is no difference between the transfer mechanisms of heat, water vapor and momentum. The set of equations used is as follows:

- *Basic equations*

The aerodynamic resistance can be calculated from wind velocity at reference level U_r and friction velocity u_* by

$$r_a = \frac{U_r}{u_*^2} \,. \tag{7}$$

The friction velocity is calculated by

$$u_* = \frac{k \cdot U_r}{\log(z_r / z_0) - \Psi_m}, \tag{8}$$

where k is the von Kármán constant, z_r and z_0 is the height of reference level and the roughness length, respectively, and Ψ_m is the stability function. Stability function depends upon the stratification of atmospheric surface layer. In neutral stratification

$$\Psi_m = 0. \tag{9}$$

In stable stratification we use two different entries for the argument of Ψ_m. When $z_r/L_{mon} \leq 0.5$ we enter the empirical expression of *Dyer* (1974):

$$\Psi_m = -4.7 \cdot \frac{z_r}{L_{mon}}. \tag{10}$$

For $z_r/L_{mon} > 0.5$ we use formula of *Holtslag* and *de Bruin* (1988):

$$\Psi_m = -A \cdot \frac{z_r}{L_{mon}} - B \cdot \left(\frac{z_r}{L_{mon}} - \frac{C}{D} \right) \cdot e^{-(D \cdot z_r)/L_{mon}} - \frac{B \cdot C}{D}, \tag{11}$$

with A=0.7, B=0.75, C=5 and D=0.35.

For unstable stratification

$$\Psi_m = 2 \log \frac{1+\mathrm{x}}{2} + \ln \frac{1+\mathrm{x}^2}{2} - 2 \operatorname{arc\,tg}(\mathrm{x}) + \frac{\pi}{2}, \tag{12}$$

with the function

$$x = \left(1 - 16 \frac{z_r}{L_{mon}} \right)^{1/4}. \tag{13}$$

The Monin-Obukhov length L_{mon} reflects the stability conditions. It is defined by

$$L_{mon} = -\frac{\rho \cdot T_r \cdot u_*^3}{g \cdot k \cdot \left(H/c_p + 0.61 \cdot T_r \cdot E\right)}, \tag{14}$$

where g is the acceleration of gravity (m s^{-2}), and E is the vapor flux (kg m^{-2} s^{-1}). Neutral stratification is supposed for $|L_{mon}| > 800$ m. If the stratification is not neutral, it is stable ($L_{mon} > 0$) or unstable ($L_{mon} < 0$).

- *Calculation procedure*

The turbulent flux/atmospheric stability relationship is implicitly defined, that is H and $L \cdot E$ depend upon L_{mon} and vice versa. To solve L_{mon}, u_*, r_a, and the turbulent fluxes H and $L \cdot E$ we applied the iterative procedure to their implicitly coupled equation system. For given soil moisture content value the computation starts from neutral stratification where $|L_{mon}| > 800$ m and $\Psi_m = 0$. Then via calculating u_* and r_a an estimate of H and $L \cdot E$ is obtained. This estimate of friction velocity and turbulent fluxes is used again to improve the estimate for L_{mon}, and so on. It appears that not more than five iterations are needed to achieve a deviation of 5 per cent in successive values of L_{mon}. In moderately unstable or stable stratification the convergence is quicker; it is usually achieved after three or four iterations. The iteration procedure does not converge in some physically unreal situations (for instance for strong radiation and weak wind; for details see *Czúcz* and *Ács* (1999) and in extremely strong unstable stratification close to free convection conditions (L_{mon} is negative and close to 0).

2.2 *Statistical - deterministic model*

The *statistical - deterministic* model estimates soil evaporation at patch scale (few kilometers × few kilometers). It consist of a *deterministic* submodel for estimating soil evaporation (see section 2.1), a *statistical* submodel for generating θ as a random variable (*Wetzel* and *Chang*, 1987) and a submodel for calculating the areal of $E(\theta)$. In the following the two latter submodels will be briefly considered.

2.2.1 *Modeling soil moisture variability*

According to observations of *Bell et al.* (1980) and *Hawley et al.* (1983), areal variations of θ on patch scale can be characterized by a normal distribution. According to *Wetzel* and *Chang* (1987), the corresponding standard deviation

$$\sigma_\theta = \min\left(0.08,\ \theta_m/2\right), \tag{16}$$

where θ_m is the areal mean value of θ. The areal variations of θ on patch scale are generated with Monte-Carlo runs applying a standard random number generation algorithm (see *Dévényi* and *Gulyás,* 1988) and using θ_m and σ_θ as input.

2.2.2 *Calculation of areally averaged soil evaporation*

Since θ is statistical variable, the turbulent heat fluxes H and $L \cdot E$ are also those. The statistical distribution of $L \cdot E$ or E is analyzed via their relative frequency distribution. The areal mean of E is estimated by numerical integration of its relative frequency distribution function $RF(E_j)$ as follows:

$$\langle E \rangle = \sum_{j=1}^{n} RF\,(E_j) \cdot E_j\,, \tag{16}$$

where $\langle E \rangle$ is the areal mean of E, j is the interval number and E_j is the corresponding E-value for the jth interval. The length of E-interval is choosen as 25 W m^{-2}. The submodel is applied in each step for $0 < \theta_m < \theta_s$ cycle to obtain the $E(\theta_m, \sigma_\theta)$ curve. Note that there is no lower boundary condition for θ_m.

3. *Numerical experiments*

The numerical experiments are performed by comparing the *deterministic* and the *statistical – deterministic* models. The simulations are made for different soil textures, atmospheric conditions and bare soil surface resistance parameterizations.

- *Soil texture*

The simulations are performed for sand, loam and clay. In this paper only the loam-referred results are presented. The results referring to sand and clay are presented in *Molnár* (1998).

- *Atmospheric forcing*

Atmospheric forcing is considered as the sum of all state and flux variables which determine the evaporation flux. In this study we distinguished strong and weak atmospheric forcing.

Strong atmospheric forcing

The strong atmospheric forcing is characterized by great net radiation flux, wind velocity and humidity deficit. We defined it by

Net radiaton flux, $R_n = 700$ W m^{-2},
Air temperature at reference level, $T_r = 25.8$°C,
Vapor pressure at reference level, $e_r = 18.0$ hPa, and
Wind velocity at reference level $U_r = 6.0$ m s^{-1}.

Weak atmospheric forcing

In weak atmospheric forcing conditions the net radiation flux, wind velocity and humidity deficit are small or moderate. We used

$R_n = 300$ W m^{-2},
$T_r = 25.8$°C,
$e_r = 32.0$ hPa, and
$U_r = 2.0$ m s^{-1}.

- *Parameterization of soil surface resistance*

Soil surface resistance is parameterized by Sun's and Dolman's formulae. They are introduced in section 2.1.2.

During numerical experiments $3 \times 2 \times 2$ runnings have been performed by both the *deterministic* and the *statistical - deterministic* models. Of course the computation time of the latter model is much longer with respect to the former one because of the generation of statistical variables.

3.1 Analysis of soil evaporation

Soil evaporation is analysed calculating evaporation curves $E(\theta)$ for different parameterizations and atmospheric forcing on spot and patch scale. $E(\theta)$ shows the evaporation/soil moisture content dependence. On spot scale θ does not show areal variations; $E(\theta)$ is obtained by running the *deterministic model* so that $E(\theta) \equiv E^{spot}(\theta)$. On patch scale an areal variations of θ is assumed defined via θ_m and $\sigma_\theta \cdot E(\theta)$ is obtained by running the *statistical - deterministic* model so that $\langle E(\theta_m, \sigma_\theta) \rangle \equiv E^{patch}(\theta)$.

$E^{spot}(\theta)$ and $E^{patch}(\theta)$ (including the factor L) obtained by Sun's and Dolman's parameterizations of r_b for strong and weak atmospheric forcing are presented in *Fig. 1* and *2*, respectively. All curves can be separated into three regions. In dry regime (low values of θ) $E(\theta)$ is controlled by soil surface; in wet regime (well-watered surface) $E(\theta)$ is controlled by the atmospheric conditions. Between the two regimes is the transition region where $E(\theta)$ is controlled by both the surface and the atmosphere. The curves can be characterized by two basic parameters: the slope $S = \partial E(\theta)/\partial\theta$ in the transition region and the saturation value $E(\theta_s)$. S is mainly determined by the parameterization of r_b while $E(\theta)$ by the atmospheric conditions, mainly by the radiation.

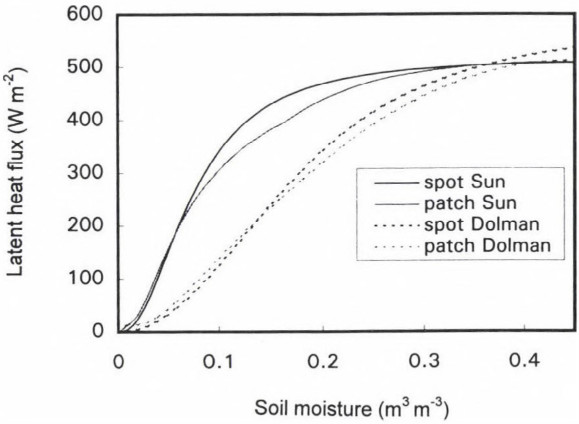

Fig. 1. Bare soil evaporation versus soil moisture changes for Sun and Dolman parameterizations on spot (black solid and dashed lines) and patch (grey solid and dashed lines) scale for strong atmospheric forcing. The curves refer to loam soil texture.

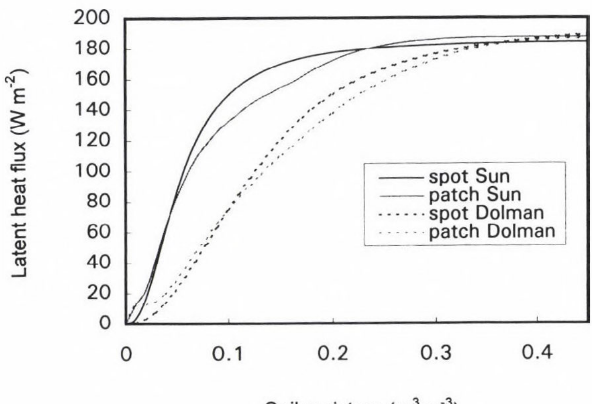

Fig. 2. As in Fig. 1 but for weak atmospheric forcing.

The scale-invariance of $E(\theta)$ depends mainly upon both the parameterization used and the surface wetness regime. In general the $E^{spot}(\theta)-E^{patch}(\theta)$ differences of Dolman's parameterization are smaller than those of Sun's parameterization; the greatest relative difference $E^{spot}(\theta)-E^{patch}(\theta)/E^{spot}(\theta)$ for Sun's parameterization achieves 13 per cent while for Dolman's parameterization between 5 and 8 per cent. These facts show that $E(\theta)$ obtained by Dolman's parameterization is more scale-invariant than the one for Sun's parameterization. Moreover we can say: If we neglect differences under 10 per cent, $E(\theta)$ obtained by Dolman's parameterization can be treated as scale-invariant. In most cases the $E^{spot}(\theta)-E^{patch}(\theta)$ differences are negligible in wet and dry regimes especially for strong atmospheric forcing conditions. This result is in accordance with the statement of *Sellers' et al.* (1997). The greatest differences appear always in the transition zone. It is very interesting to note that the $E^{spot}(\theta)-E^{patch}(\theta)/E^{spot}(\theta)$ relative difference is not negligible in extreme dry regime (θ is about 0.03 $m^3\ m^{-3}$) when the atmospheric forcing is weak. *Sellers et al.* (1997) does not note this interesting behavior.

3.2 Areal variation of soil evaporation

Areal variation of $E(\theta)$ is examined analysing relative frequency RF of $E(\theta)$. Relative frequency of $E(\theta)$ is studied for different soil textures and surface wetness states. The estimates are performed for strong atmospheric forcing conditions. In this study the analyses refer only to loam; the results referring to sand and clay are presented in *Molnár* (1998). The questions we want to consider are as follows:

(1) Does frequency distribution of $E(\theta)$ depend upon the surface wetness state? With which distribution can RF of $E(\theta)$ be approached?
(2) Is RF of $E(\theta)$ determined by the parameterization of r_b?

The relative frequencies of $E(\theta)$ obtained by Sun and Dolman parameterizations for extreme dry (θ=0.03 $m^3\ m^{-3}$), dry (θ=0.07 $m^3\ m^{-3}$), moderate wet (θ=0.13 $m^3\ m^{-3}$) and extreme wet (θ=0.25 $m^3\ m^{-3}$) conditions are presented in *Figs. 3a, 3b, 3c* and *3d*, respectively. Note that the "extreme dry", "dry", "moderate wet" and "extreme wet" descriptions are not exactly defined wetness categories. RF of $E(\theta)$ obtained by Sun's formula is qualitatively in agreement with RF of $E(\theta)$ obtained by Dolman's formula only in extreme dry and extreme wet conditions. In extreme dry conditions the normally distributed soil moisture variations produce an exponential relative frequency distribution of soil evaporation. This is in accordance with simulation results of *Wood*

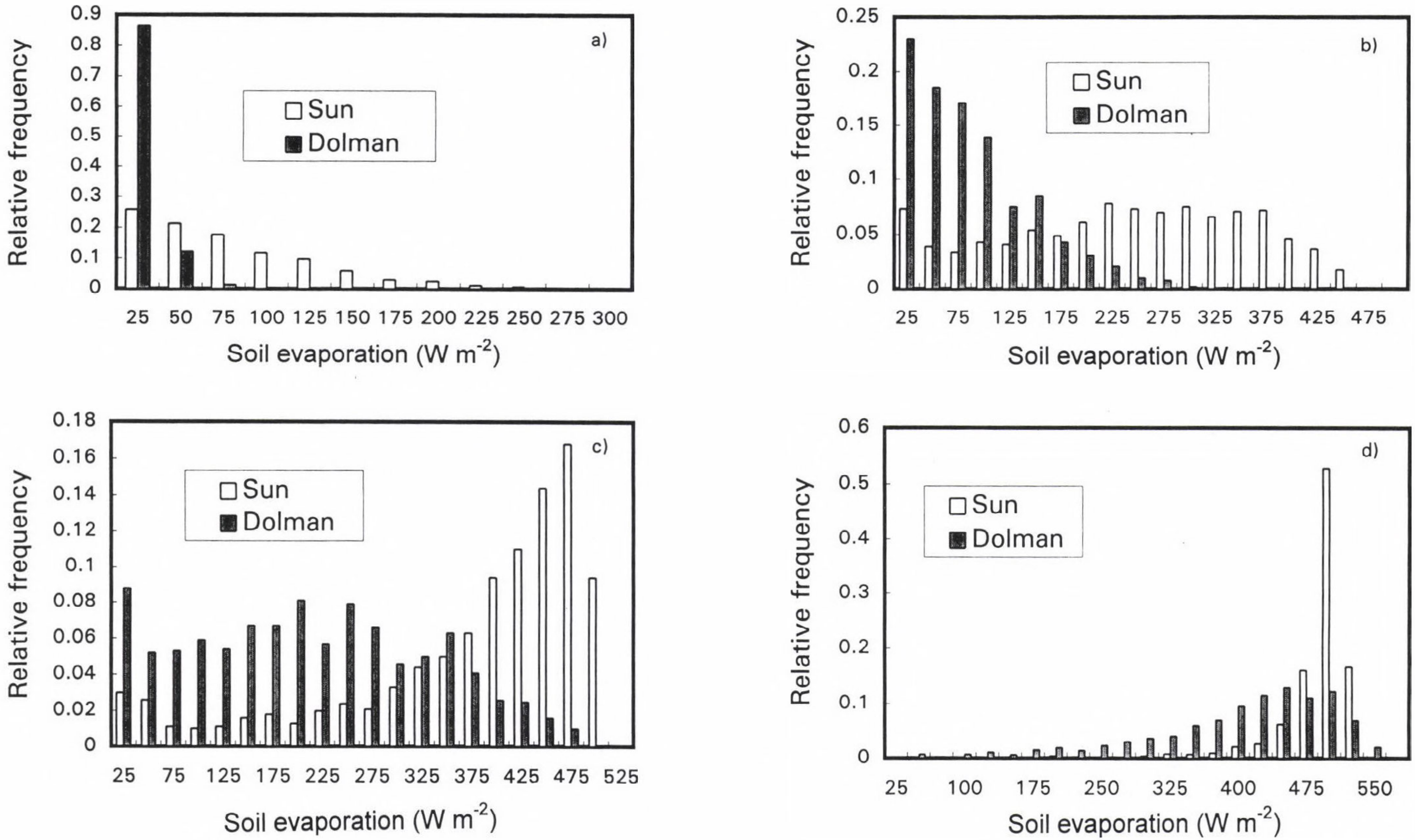

Fig. 3. Histogram of bare soil evaporation for Sun (light columns) and Dolman (dark columns) parameterizations and strong atmospheric forcing conditions in (*a*) extreme dry, (*b*) dry, (*c*) moderate wet and (*d*) extreme wet soil surface wetness conditions.

(1997) and analytical calculations of *Hantel* and *Ács* (1998). In extreme wet conditions the shape of relative frequency distributions is like a lognormal distribution as reflected by mirror. For moisture regimes between the extrem dry and wet regimes the shapes of RF distributions are qualitatively different. Thus for instance in dry conditions (see Fig. 3b) the RF distribution of $E(\theta)$ obtained by Dolman's formula can be characterized by an exponential distribution, while RF distribution of $E(\theta)$ obtained by Sun's formula by a uniform distribution. In moderate wet conditions the deviations between RF distributions are similarly as great as in dry conditions (see Fig. 3b).

3.3 Aggregated soil moisture content

The aggregated soil moisture content θ_{ag} is that soil moisture content value by which the deterministic model yields the areal mean patch scale evaporation. So

$$E(\theta_{ag}) = \langle E(\theta_m, \sigma_\theta) \rangle \,, \qquad (17)$$

where $E(\theta_{ag})$ is the areal mean soil evaporation calculated by *deterministic model* using θ_{ag} and $\langle E(\theta_m, \sigma_\theta) \rangle$ is the areal mean soil evaporation calculated by *statistical - deterministic* model using θ_m and σ_θ . θ_{ag}/θ_m relationship is possible to get comparing $E(\theta)$ curves refering to spot and patch scale (see Figs. 1 and 2). Obviously the relationship is well defined in the transition zone of $E(\theta)$ curves. In dry and wet regimes of $E(\theta)$ curves there is no unequivocal relationship between θ_{ag} and θ_m . The θ_{ag}/θ_m relationship for strong and weak atmospheric forcing conditions using Sun's and Dolman's parameterizations is presented in *Figs. 4* and *5*, respectively. According to the plots we can say:

(1) The relationship between θ_{ag} and θ_m extremly weakly depends upon atmospheric forcing conditions. There is practically no difference between the slope and the intercept of regression lines with respect to the forcing conditions (see Eqs. (18) to (21)),

Sun, strong forcing: $\theta_{ag} = 0.811\theta_m - 0.00002,$ (18)

weak forcing: $\theta_{ag} = 0.795\theta_m - 0.0015$ (19)

and

Dolman, strong forcing: $\theta_{ag} = 0.866\theta_m + 0.011,$ (20)

weak forcing: $\theta_{ag} = 0.834\theta_m + 0.0103.$ (21)

Hence this dependence can be neglected for both parameterizations.

(2) The relationship between θ_{ag} and θ_m are linear; the correlation coefficients in all cases are greater than 0.98.

Similar considerations are also valid for sand and clay (*Molnár*, 1998). Finally, it is obvious that neglecting the effect of atmospheric forcing conditions upon θ_{ag}/θ_m relationship (which is a reasonable assumption), it is possible to calculate areal mean soil evaporation on patch scale using the deterministic model in combination with θ_{ag}/θ_m relationship.

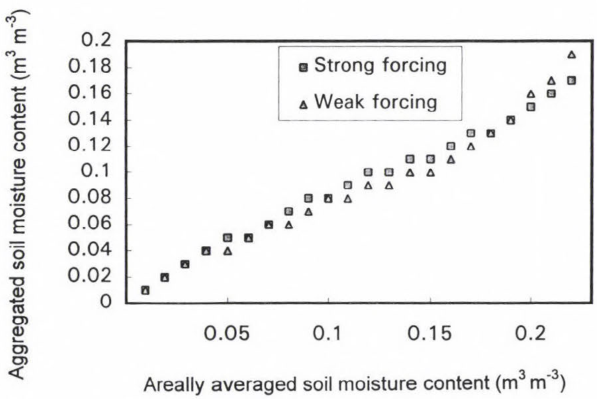

Fig. 4. Aggregated versus areally averaged soil moisture content for Sun parameterization under strong (square) and weak (triangle) atmospheric forcing conditions.

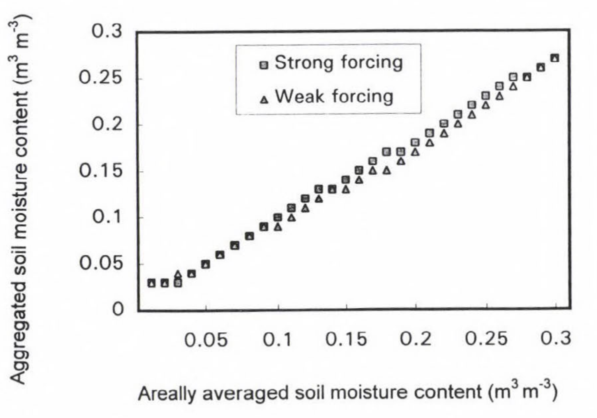

Fig. 5. As in Fig. 4 but for Dolman parameterization.

4. Conclusions

Microscale soil evaporation E characteristics are analysed comparing a *deterministic* and a *statistical – deterministic* evaporation model. Evaporation characteristics are considered in terms of evaporation curves $E(\theta)$, relative frequency distribution of $E^{spot}(\theta)$ and the aggregation algorithms for their estimation. The *deterministic* spot scale E^{spot}-model is based on the *Penman-Monteith* concept (*Ács* and *Hantel,* 1999), that is the evaporation flux is determined by the *Penman-Monteith*'s equation, the ground heat flux at soil surface is parameterized via net radiation and the sensible heat flux is obtained as residual component from the energy balance equation. The *statistical – deterministic* patch scale $E^{patch}(\theta)$ model consist of the $E^{spot}(\theta)$-submodel, a statistical submodel for generating θ as random variable and a submodel for calculating the areal mean $E^{patch}(\theta)$.

The numerical experiments are made for different soil textures, atmospheric conditions and bare soil surface resistance r_b parameterizations. In this study only the loam-referred results are analysed. The results can be briefly summarized as follows:

(1) The scale-invariance of $E(\theta)$ depends upon both the parameterization of r_b and the soil surface wetness regime. In general the normally distributed soil moisture variations on the patch scale moderate the nonlinear relationship between E and θ with respect to that on spot scale. The greatest $E^{spot}(\theta)-E^{patch}(\theta)$ differences appear for Sun's parameterization of r_b in transition region of $E(\theta)$ curves. The $E^{spot}(\theta)-E^{patch}(\theta)$ differences are negligible in wet and dry soil wetness regimes and also in the transition region but for Dolman's parameterization of r_b. Summarizing, if we neglect the differences under 10 per cent, $E(\theta)$ obtained by Dolman's parameterization can be treated as scale-invariant.

(2) The relative frequencies of $E(\theta)$ obtained by Sun's and Dolman's formulae are qualitatively in approximate agreement with each other only in extreme dry and extreme wet conditions. In extreme dry conditions the normally distributed soil moisture variations produce an exponential RF distribution of $E(\theta)$, while extreme wet conditions lead to a distribution like a lognormal distribution as reflected by mirror. For dry and the moderate wet soil moisture regimes the RF distributions obtained by Sun's and Dolman's parameterizations are qualitatively different and they can hardly be described.

(3) For calculating areal mean patch scale evaporation $E^{patch}(\theta)$ it is possible to introduce the aggregated soil moisture content θ_{ag} (see Eq. (17)). The relationship between θ_{ag} and the areal mean soil moisture content θ_m extremly weakly depends upon atmospheric forcing conditions that is the differences between the slope and the intercept of θ_{ag}/θ_m regression lines for both parameterizations are minor with respect to the forcing conditions. Further the relationship between θ_{ag} and θ_m is linear.

The results suggest that $E(\theta)$ characteristics depend strongly upon the parameterization of r_b. In this study we only want to show this fact and not more, that is we do not intend to make a decision which parameterization would be preferred. Furthermore, the θ_{ag}/θ_m relationship can be use for $E(\theta)$'s upscaling from spot to patch scale; that is knowing the relation between θ_{ag} and $\theta_{m,}$ it is possible to calculate $E^{patch}(\theta)$ not only by the *statistical - deterministic* model but also by the *deterministic* one. Of course, these results are valid only —as noticed in introduction—when there are no advective effects and meso-scale circulation patterns. In the latter cases the evaporation characteristics and its upscaling strategy are much more complex.

Acknowledgements—This study is partly financially supported by the *Hungarian Ministry* for *Culture* and *Education* via OTKA Foundation, project number T-02958.

References

Ács, F., 1996: Control of transpiration by soil moisture and atmosphere. *Proc. of the Int. Conf. on Water Resources & Environmental Research: Towards the 21st Century (Vol. I)*, 101-107. Oct 29-31, 1996, Kyoto, Japan.

Ács, F. and *Hantel, M.*, 1997: Upscaling of transpiration from spot to patch scale. *Proc. of the Joint Assemblies of the International Association of Meteorology and Atmospheric Sciences (IAMAS) and the International Association for the Physical Sciences of the Oceans (IAPSO)*, CMH4 Global and Regional Energy and Water Cycles, CMH4PP1-9 July, 1997, Melbourne.

Ács, F. and *Hantel, M.*, 1999: The Penman-Monteith concept based land-surface model PMSURF. *Időjárás 103*, 19-36.

Bastiaanssen, W.G.M., 1995: Regionalization of surface flux densities and moisture indicators in composite terrain. A remote sensing approach under clear skies in Mediterranean climates. *PhD Thesis*, *271* pp. Department of Meteorology, Eötvös Loránd University, Budapest, Hungary.

Bell, K.R., Blanchard, B.J., Schmugge, T.J. and *Witzak, M.W.*, 1980: Analysis of the surface moisture variations within large field sites. *Water Resour. Res. 61*, 796-810.

Chehbouni, A., Njoku, E. G., Lhomme, J.-P. and *Kerr, Y. H.*, 1995: Approaches for averaging surface *parameters* and fluxes over heterogeneous terrain. *J. Climate 8*, 1386-1393.

Czúcz, B. and *Ács, F.*, 1999: Parameterization of unstable stratification in land-surface model PMSURF: An *examination* of the convergence by empirical methods (in Hungarian). *Légkör* Vol. XLIV, No. 2, 2-6.

Dévényi, D. and *Gulyás, O.*, 1988: *Mathematical Statistical Methods in Meteorology* (in *Hungarian*). *Tankönyvkiadó*, Budapest, pp. 443.

Dolman, A. J., 1993: A multiple-source land surface energy balance model for use in general *circulation models*. *Agric. Forest Meteorol.* *65*, 21-45.

Dyer, A. J., 1974: A review of flux-profile Relationships. *Bound. Layer Meteorol.* *7*, 363-372.

Entekhabi, D. and *Eagleson, P. S.*, 1989: Land surface hydrology parameterization for atmospheric *general* circulation *models* including subgrid scale spatial variability. *J. Climate* 2, 816-831.

Entekhabi, D. and *Eaglesson, P. S.*, 1991: Climate and equilibrium state of land surface hydrology *parametrizations*. *Surveys in Geophysics 12*, 205-220.

Famiglietti, J.S. and *Wood, E., F.*, 1991: Evapotranspiration and runoff from large land areas: Land surface hydrology for atmospheric *general* circulation models. *Surveys in Geophysics 12*, 179-204.

Famiglietti,. J. S. and *Wood, E. F.*, 1994: Multi-scale modeling of spatially-variable water and *energy* balance *processes*. *Water Resour. Res.* *30*, 3061-3078.

Garratt, J. R., 1992: The internal boundary layer-A review. *Bound.-Layer Meteorol.* *50*, 171-203.

Hantel, M. and *Ács, F.*, 1998: Physical aspects of the weather generator. *J. Hydrol.* *212-213*, 393-*411*.

Hawley, M. E., Jackson, T. J. and *McCuen, R.H.*, 1983: Surface soil moisture variation on small *agricultural* watersheds. *J. Hydrol.* *62*, 170-200.

Holtslag, A. A. M. and *de Bruin, H.A.R.*, 1988: Applied modelling of the nightime surface energy *balance* over land. *J. Appl. Meteor.* *27*, 689-704.

Hupfer, P. and *Raabe, A.*, 1994: Meteorogical transition between land and see in the microscale. *Meteor.* Z. 3, 100-103.

Kabat, P., Hutjes, R.W.A. and *Feddes, R.A.*, 1997: The scaling characteristics of soil parameters: From plot scale *heterogeneity* to subgrid parametrization. *J. Hydrol.* *190*, 363-396.

Lhomme, J.-P., 1992: Energy balance of heterogeneous terrain. Averaging the controlling *parameters*. *Agric. Forest Meteorol.* *61*, 11-21.

Lhomme, J.-P., Chehbouni, A. and *Monteny, B.*, 1994: Effective parameters of surface energy *balance* in heterogeneous landscape. *Bound.- Layer Meteorol.* *71*, 297-309.

Li, B. and Avissar, R., 1994: The impart of spatial variability of land-surface characteristics on land-surface heat fluxes. *J. Climate 7*, 527-535.

Molnár, I. 1998: Relationship between the bare soil evaporation and the environmental conditions at *microscale*. *Dipl. Thesis*, 37 pp+Appendix. Department of Meteorology, Eötvös Loránd University, Budapest.

Monteith, J., 1965: Evaporation and environment. In *The State and Movement of Water in Living Organism. Proc. of 19^th^ Symp. Soc. Exp. Biol.*, (ed.: *G. Fogg*). Cambride University Press, Cambridge, pp. 105-236.

Mölders, N. and *Raabe, A.*, 1996: Numerical investigations on the influence of subgrid-scale surface *heterogeneity* on *evapotranspiration* and cloud processes. *J. Appl. Meteorol.* *35*, 782-795.

Nickerson, E. and *Smiley, I.* 1975: Surface energy budget parametrizations for urban models. *J. Appl. Meteorol.* *14*, 197-*300*.

Rodriguez-Camino, E. and *Avissar, R.*, 1999: Effective parameters for surface heat fluxes in heterogeneous *terrain*. *Tellus 51A*, 387-399.

Sellers, P.J., Heiser, M.D., Hall, F.G., Verma, S.B., Desjardins, R.L., Schuepp, P.M. and *MacPherson, J.I.*, 1997: The *impact* of using area-averaged land surface properties-topography, vegetation condition, soil wetness-in calculations of intermediate scale (approximately 10 km^2) surface-atmosphere heat and moisture fluxes. *J. Hydrol.* *190*, 269-301.

Shuttleworth, J.W., 1988: Macrohydrology-the new challenge for process hydrology. *J. Hydrol.* *100*, 31-56.

Shuttleworth, J.W., 1996: BAHC Focus 2-Future directions in aggregation. In *Bach News*. Edited by BAHC Core Project Office, Potsdam Institute for Climate Impact Research, P.O. Box. 601203, 14412 Potsdam, Germany.

Shuttleworth, J.W., 1997: Hydrological models, regional evaporation, and remote sensing: Let's start simple and maintane perspective. In *Land Surface Process in Hydrology. Trials and Tribulations of Modelling and Measuring* (eds.: *S. Sorooshian, H.V. Gupta* and *S.C. Rodda*). Springer Verlag, Berlin, Heidelberg.

Sun, S.F., 1982: Moisture and heat transport in a soil layer forced by atmosperic conditions. *M.S. Thesis*, Dept. of Civil Engineering, University of Connecticut, 72. pp.

Wetzel, J.P. and *Chang, J.T.*, 1987: Concerning the relationship between evapotranspiration and soil moisture. *J. Climate and Appl. Meteorol. 26*, 18-27.

Wood, E.F., 1997: Effects of soil moisture aggregation on surface evaporative fluxes. *J. Hydrol. 190*, 397-412.

IDŐJÁRÁS
Quarterly Journal of the Hungarian Meteorological Service
Vol. 104, No. 3, July–September 2000, pp. 161–172

Estimation of historical atmospheric lead (Pb) deposition over Hungary

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(Manuscript submitted for publication 24 May 2000; in final form 11 August 2000)

Abstract—Lead is one of the most harmful toxic metals in our environment. It is emitted mainly by anthropogenic sources (combustion of gasoline, oil and coal burning, waste incineration etc.) whose source strengths varied significantly during the past several decades in all European countries. Even relatively low fluxes across the Earth's surface can result in accumulation of lead in various soils. From the soils it can be taken up by the plants and maybe leached out into the ground water. This paper presents estimations on the historical lead depositions and the atmospheric lead budgets over Hungary, for the period of 1955–2000. Computations are based on the results of long-range transport model simulations as well as on the regional background air pollution measurements carried out at K-puszta station in Hungary. Besides the estimations of the actual and cumulative atmospheric loads, the geographical origin of lead that is deposited is also calculated.

Key-words: atmospheric deposition, lead, long-range transport.

1. *Introduction*

Since the middle of this century, energy generation, industrial production and transportation have caused serious environmental contamination by trace elements including lead. The rate of contamination can vary from place to place as a function of source densities and intensities of lead flux as well as meteorological conditions. The pattern of pollution may be characterized not only by local, highly concentrated sites such as densely populated urban areas, but also by lower concentrations of pollution widely dispersed over the landscape including agricultural regions, forests and surface waters. Aerosol particles containing lead and other trace elements can be transported far away from their sources by advection before being deposited on the surface (e.g., *Mészáros*, 1999).

Consumption of leaded gasoline was the major source category in the total lead emission in Europe (*Pacyna*, 1996). Introduction of unleaded gasoline in the mid-80's in western Europe resulted in significant lead emission reduction as well as in re-ranking the relative contributions of source categories. This progress could be detected approximately a decade later in the eastern part of the European continent (*Olendrzynski et al.*, 1995). Long-range transport of atmospheric lead is not simply a clean air protection problem in Europe but a major part of environmental policy as well: considering its characteristic residence time in the atmosphere (a few days) and the sizes of countries in our continent, international co-operation is needed to control the transboundary flux and deposition of atmospheric lead.

2. *Atmospheric long-range transport model applied for the computations*

For the long-range transport computations a continental scale, climatological-type model (TRACE) was used. Input fields of emission inventory, meteorological data and deposition parameters are preprocessed for 150×150 km^2 spatial resolution grid system (EMEP). The concentration of lead at a certain receptor point is given by a simple loss function:

$$c(x_r, y_r; x_e, y_e) = \beta\, E\, (x_e, y_e)\, R^{-1}\, (1-a)\, e^{-(kd+kw)t}\,, \qquad (1)$$

where

- c is lead concentration in the air in a certain receptor point as a result of an individual source point (ng m^{-3});
- (x_r, y_r) and (x_e, y_e) are the spatial co-ordinates of source and receptor points;
- E is the emission at the source point (ng s^{-1});
- β is derived by setting the upward and downward atmospheric fluxes equal assuming mass conservation (s m^{-2})
- R is distance between the source and receptor (m);
- A is local emission coefficient (dimensionless);
- kd, kw are loss parameters for dry and wet depositions (s^{-1});
- t is atmospheric transport time between source and receptor (s);

The total concentration at the receptor point is computed from the sum of contributions coming from each emission source, weighted according to the frequency of 925 hPa backward trajectories, $F(s)$ coming from a particular sector, s:

$$c(x_r, y_r) = \sum_{s=1}^{8} F_{(s)} c_s (x_r, y_r). \tag{2}$$

As indicated in Eq. (2) there are 8 spatial sectors considered during trajectory analyses. In the next step of the calculations, wet and dry deposition of the pollutant at the receptor are computed from the atmospheric lead concentration:

$$d_w = c(x_r, y_r)\ W P \tag{3}$$

where
- W is the scavenging ratio (dimensionless);
- P is the precipitation intensity (m s^{-1}).

Dry deposition is expressed as:

$$d_d = c(x_r, y_r)\ v_d \tag{4}$$

where
- d_d is the dry deposition (ng m^{-2} s^{-1});
- $c(x_r, y_r)$ is the lead concentration at the receptor point (ng m^{-3});
- v_d is the dry deposition velocity (m s^{-1}).

Dry deposition velocities of lead are calculated separately for each grid element, depending on roughness length, friction velocity and size distribution of aerosol particles containing lead. Size distribution of particles were taken from *Mészáros et al.* (1997) whose research group carried out Berner-type cascade impactor sampling and trace metal measurements in Hungary.

For detailed model description, testing and verification see *Alcamo et al.* (1992), *Bozó* (1994) and *Bozó et al.* (1992).

3. Emission of lead in European countries during the past decades

The major source of lead during the past several decades was the consumption of leaded gasoline (*Pacyna,* 1996). In addition, coal and oil burning, as well as non-ferrous metal industry contribute to the total anthropogenic emission of lead. Based on a source category ranking study referring to lead in Hungary in the mid-80's, line sources (gasoline) represented approximately 85% of the total emission. Detailed historical (1955–) emission inventories for the European countries were compiled by *Pacyna* (1993, 1996), *Pacyna et al.*

(1991) and *Olendrzynski et al.* (1995). These emission inventories are based on the detailed and country-specific information on the:

- industrial and energy technology used,
- production and consumption rates in industrial, energy and transportation sectors,
- lead content of raw materials and fuels,
- environmental protection installations at the emitters.

During the 50's and 60's the lead content of European gasoline was approximately 0.4 g ℓ^{-1}. It decreased significantly down to 0.15 g ℓ^{-1} in Germany in the mid-70's: similar reductions were achieved in Scandinavia and Benelux countries in early 80's. In Hungary, lead content of gasoline was remarkably reduced in 1985 (< 0.25 g ℓ^{-1}). By that time unleaded gasoline was widely used in western European countries: by the mid-90's it became dominant on the fuel market in those countries. This progress could be detected in Hungary and other central-eastern European countries only approximately 8–10 years later. Temporal variation of lead emission from Hungary is shown in *Fig. 1.* It can be seen that the rate of emission increased rapidly during 1955–1975: highest mass of lead (1150 t a^{-1}) was emitted in the mid-70's. It is expected that by the end of this century the rate of lead emission will have been reduced down to 70-80 t a^{-1}

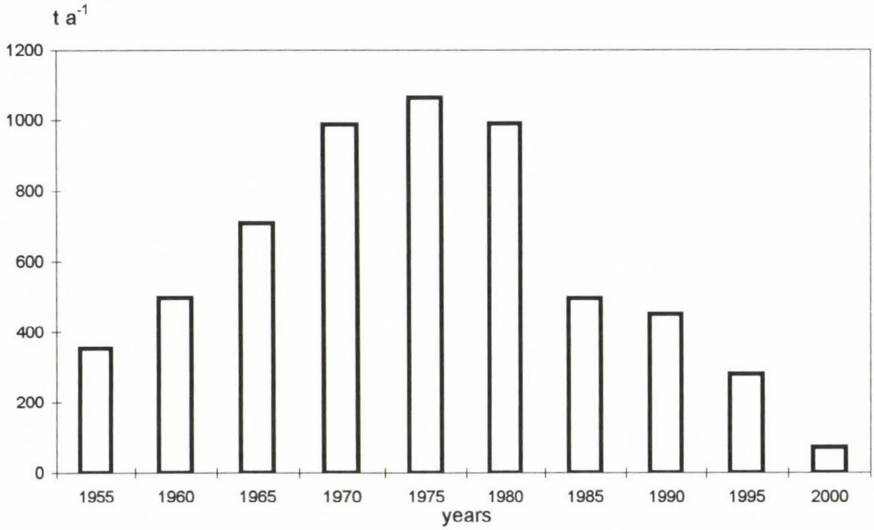

Fig. 1. Emission of lead in Hungary.

4. Origin of lead deposition in Hungary during the period 1955–2000

Model computations have been performed to estimate the temporal variation of the geographical origin of atmospheric lead deposition in Hungary during 1955–2000. Source-receptor relationships of grid elements were first aggregated to country-to-country matrices. Due to the fact that there were significant differences in industrial technologies, environmental installations and lead emission densities in different parts of Europe, for better representation of our results, three main areas, covering the whole European continent, were separated in computations: (i) Hungary; (ii) eastern part of Europe, i.e. the former socialist countries without Hungary and (iii) western part of Europe, i.e. all the remaining European countries. Results are shown in *Fig. 2.* It can be seen that total (wet+dry) lead deposition increased continuously from 510 t a^{-1} up to 1450 t a^{-1} during the period 1955–1975.

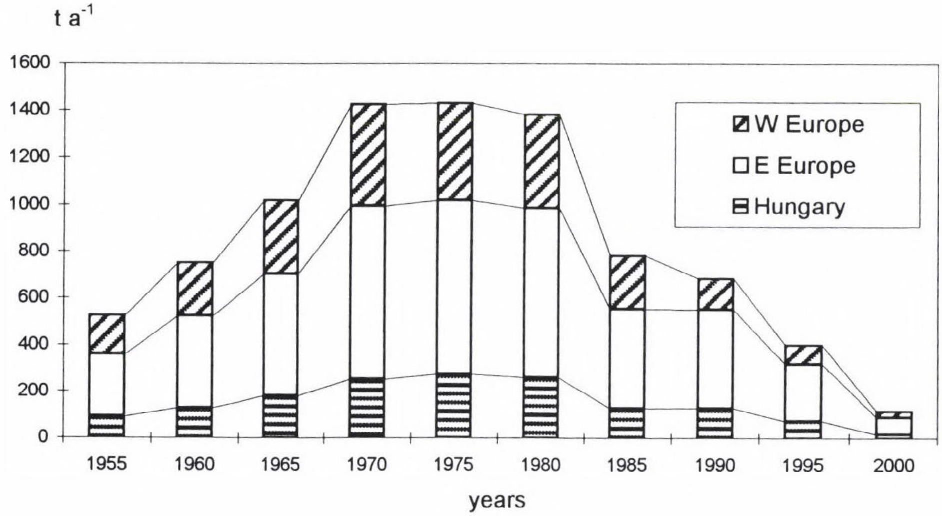

Fig. 2. Origin of total (wet+dry) lead deposition in Hungary.

Total lead deposition in Hungary started to decrease in the mid-80's. It is expected that due to the further reduction of European lead emission (*Berdowski et al.,* 1998), the rate of total lead deposition in Hungary will not exceed 100 t a^{-1} in 2000. Percentage contributions from the different source regions selected are presented in *Fig. 3.* It can be concluded from this figure that the relative share of regions contributing to Hungarian lead deposition were practically unchanged during the period 1955–1980: relative contributions of Hungarian sources, western European sources and eastern European sources were between 18-20%, 29-32% and 48–53%, respectively. Lowest Hungarian contribution to the total lead deposition was estimated for the mid-80's. During the 90's, the relative contribution of the western European sources decreased significantly: it can be explained by the intensive reduction of lead content in gasoline used in western European countries.

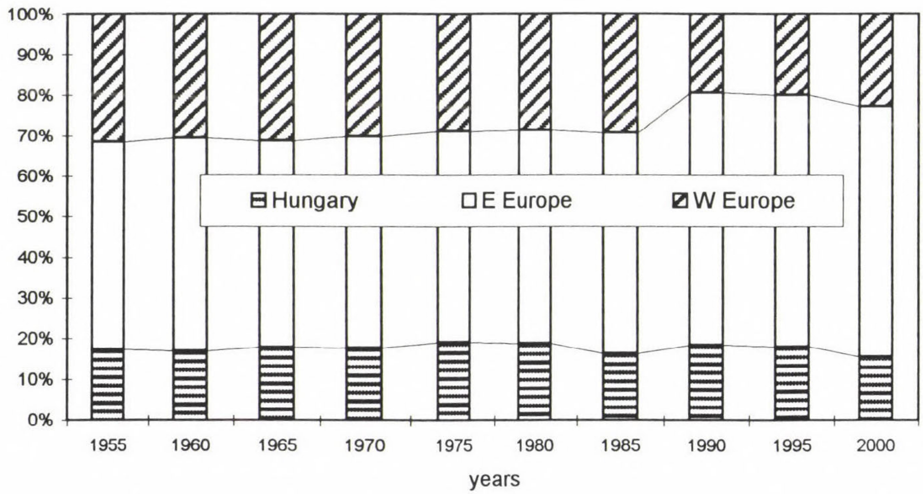

Fig. 3. Origin of total (wet+dry) lead deposition in Hungary (%).

Spatial variabilities of 850 hPa backward trajectories and lead concentrations were estimated at K-puszta station for the ten year period of 1988–1997. It can be seen in *Fig. 4*, that the prevailing backward trajectory directions are NW and W, while highest lead concentrations were observed when air masses came from SE, E and S. It is explained by the fact that lead emission densities were higher over the SE European countries than those of over western and northern part of the European continent during the period investigated.

As it was mentioned in the sections above, atmospheric lead transport and deposition is a long-range environmental problem. It means that Hungary not only "imports" lead through the atmosphere but it is a significant lead "exporter" as well. Based on TRACE model computations it was possible to estimate the quantitative distribution of total lead deposition in Europe, originating from Hungarian sources. Results of computations are plotted in

Fig. 5. It was concluded that 26% of lead emitted in Hungary is deposited in our country, while 57% is deposited in the eastern European region. The remaining 17% is transported and deposited in western Europe. In other words it means that only 260 t of 1063 t lead emitted in 1975 in Hungary was deposited in our country, the majority of lead emitted left Hungary through atmospheric long-range transport processes.

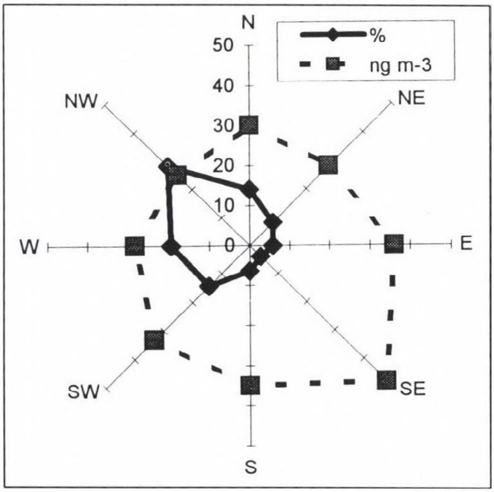

Fig. 4. Distribution of 850 hPa backward trajectories (%) and Pb concentrations (ng m^{-3}) by spatial sectors at K-puszta, during 1988–1997.

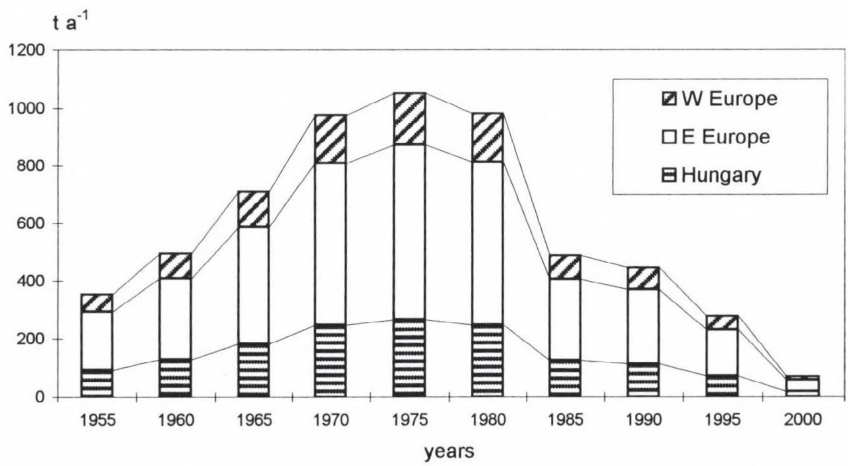

Fig. 5. Total deposition of lead emitted from Hungarian sources.

5. Comparison of model results with wet deposition measurements in Hungary

Lead concentration in atmospheric aerosol and precipitation has been monitored since the early-80's in Hungary at the regional background air pollution monitoring station, K-puszta (*Bozó*, 1996). Wet deposition is calculated by multiplying the lead concentration in precipitation water by the precipitation amount. Highest wet deposition rate was detected in 1987 (9.06 mg m^{-2} a^{-1}) while the lowest was detected in 1998 (2.21 mg m^{-2} a^{-1}). Based on historical European emission data model computations were performed for the receptor location of K-puszta, Hungary: their results and the wet deposition rates measured are plotted together in *Fig. 6*. For the period of 1955–1985 only model computations were available: it can be seen that highest wet deposition rates were calculated for the 70's (appr. 11 mg m^{-2} a^{-1}). Since the early 80's there has been a continuous decrease in the wet deposition rate of lead in Hungary: it is reflected both in model computations and regional background measurements. On the basis of model calculations one can expect that wet deposition rate of lead in 2000 could be around 1 mg m^{-2} a^{-1} in Hungary under regional background conditions. It can also be concluded from the figure that results of model simulations underestimate the wet deposition measurements by 10–20%. The reason for that could be the underestimations in lead emission data and the uncertainty in parameterization of wet scavenging processes in the atmosphere.

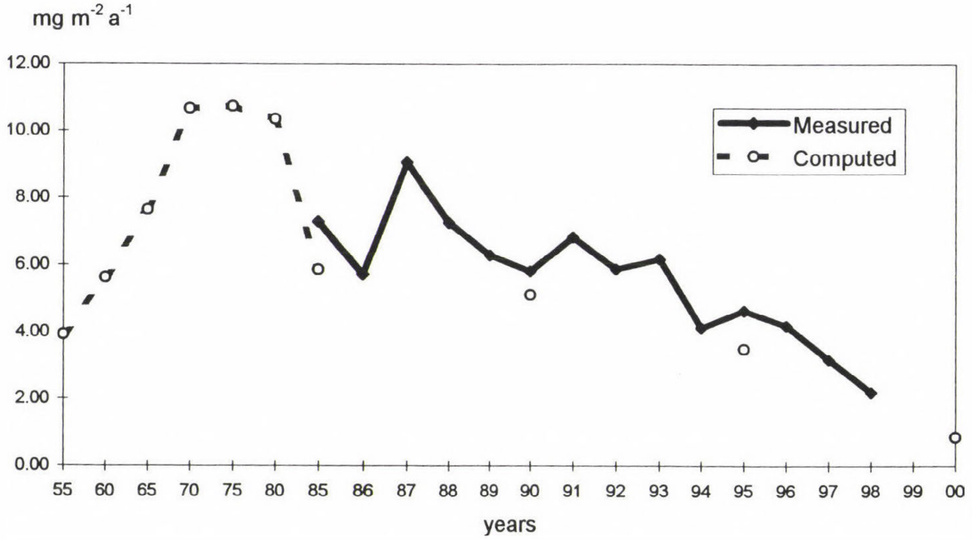

Fig. 6. Computed and measured rates of wet lead depositions.

6. Cumulative lead deposition

The rate of lead deposition varied significantly during the past decades. Due to the cumulative characteristics of lead in our environment, it is advisible to estimate the cumulative lead deposition in Hungary for the past 45 years and to provide some quantitative estimates for the next decade. This type of simulation was also done by means of TRACE model computations. Historical emission data were taken from *Olendrzynski et al.* (1995) while future scenarios are based on the calculations of *Berdowski et al.* (1998). For comparisons, the target of model simulations was not only Hungary but a few other countries in different regions of Europe — United Kingdom, The Netherlands, Spain, Austria, Romania and Poland (*Fig. 7*) It is not surprising that cumulative lead deposition was much higher during the 30 years of the period 1955-1985 than that of 1985–2015. Regarding Hungary, the rate of total lead deposition was 320 mg m^{-2} during 1955–1985, while on the basis of model computations it is expected that it will not exceed 95 mg m^{-2} during the consecutive 30 years period (1985–2015). It can also be stated that in some selected countries (e.g., The Netherlands or Austria) the cumulative lead deposition rate was higher than in Hungary, while in the case of Romania and Spain lower cumulative deposition rates were estimated.

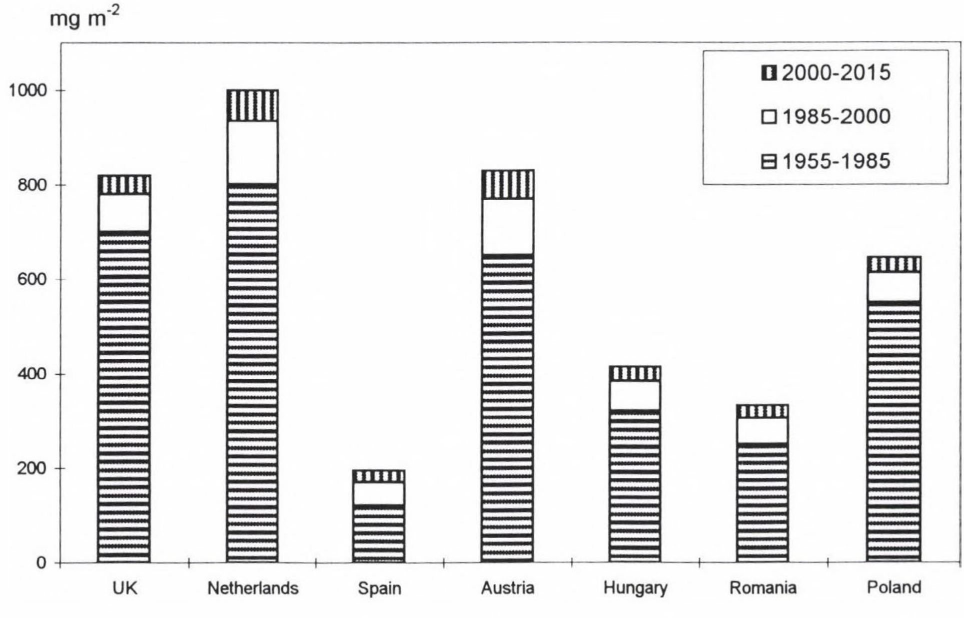

Fig. 7. Cumulative lead depositions in a few European countries.

7. Atmospheric budget of lead over Hungary during 1955–2000

On the basis of model computations performed and the historical lead emission data for Hungary, the atmospheric budget of lead can be estimated as the difference between total deposition and emission. Results are shown in *Fig. 8.* It can be concluded that the budget is positive during the whole period investigated so our country plays a net "importer" role in the European atmospheric lead budget. It was also calculated how the atmospheric budgets relates to the actual emission rates in a few selected European countries. Results for the period 1960–2000 are presented in *Fig. 9.* It was concluded that at the western boundaries of the European continent (UK, The Netherlands) this parameter is negative due to the fact that the relative contribution of the countries' own sources to the total lead deposition in these countries is significant: prevailing atmospheric transport from the Northwest reduces the influence on atmospheric lead transport from other European countries. In Hungary, the value of this ratio varied between 0.32–0.58 between 1960–2000.

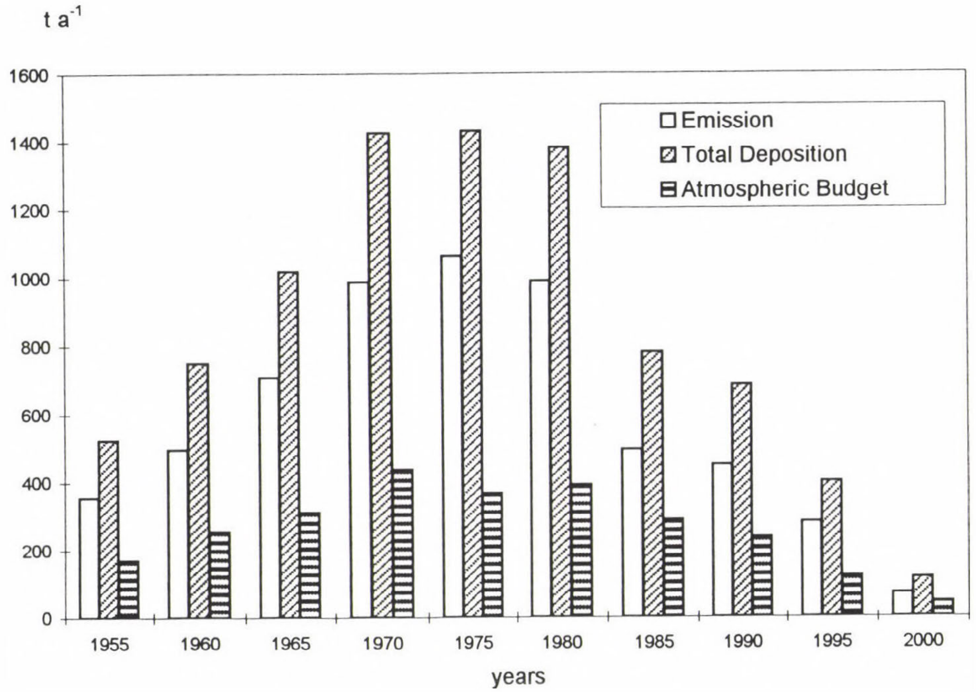

Fig. 8. Atmospheric budget of lead in Hungary.

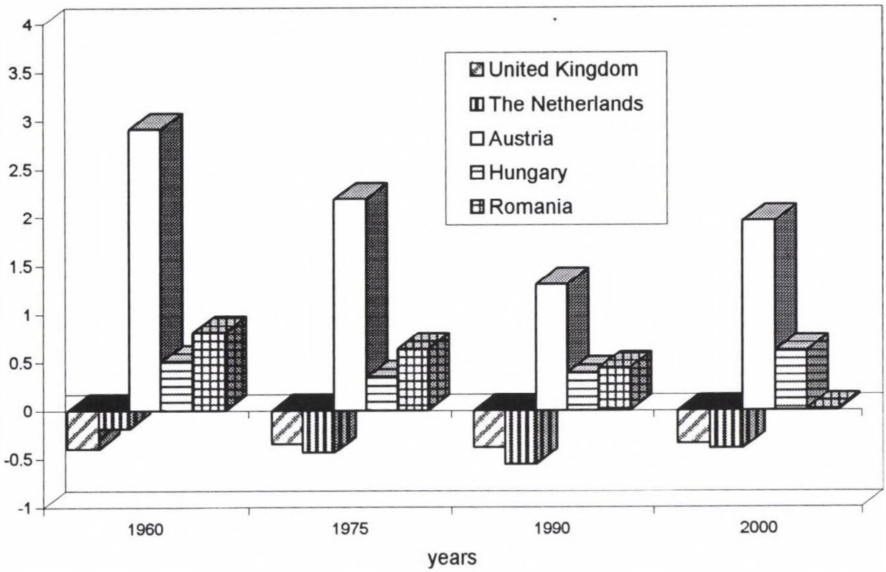

Fig. 9. Ratio of atmospheric lead budget to the lead emission in selected countries.

8. *Conclusions*

Historical lead deposition in Hungary was investigated in this paper by means of long-range transport model computations. To the extent that observations were available, modeling results were compared with the wet deposition rates measured at K-puszta station in Hungary. On the basis of these investigations the following main conclusions can be drawn:

(1) Transboundary sources of lead dominated the total (wet+dry) lead deposition in Hungary during the period 1955-2000: their contribution is estimated to be 81–84% of the total. Highest deposition rates were computed for the 70's (above 1400 t a^{-1} lead deposited);

(2) Cumulative deposition rate of lead in Hungary was computed to be 320 mg m^{-2} for the period 1955–1985, while it is expected not to exceed the rate of 95 mg m^{-2} during the period 1985-2015;

(3) Phasing out leaded gasoline from the market in most of the European countries will eliminate the majority of atmospheric lead pollution and deposition problems during the next decade. Annual emission rates of lead are going to decrease rapidly in SE European countries as well, while the relative contribution of industrial and energy source categories

to the total annual lead emission of countries will reach higher percentage shares than those during the past decades.

(4) Atmospheric budget of lead was positive in Hungary during 1955–2000, that is the rate of total lead deposition in the country exceeded the rate of emission from Hungarian sources: ratio between the atmospheric budget and emission varied in the range of 0.32–0.58.

References

Alcamo, J., Bartnicki, J., Olendrzynski, K. and *Pacyna, J.*, 1992: Computing heavy metals in Europe's atmosphere - I. Model development and testing. *Atmospheric Environment 26A*, 3355-3370.

Berdowski, J.J.M., Pulles, M.P.J. and *Visschedijk, A.J.H.*, 1998: Incremental cost and remaining emission in 2010 of Heavy Metals (HM) resulting from the implementation of the draft HM Protocol under UN ECE Convention on Long Range Transboundary Air Pollution. *TNO Report* - R 98/020.

Bozó, L., 1996: Long-range transport model estimations on the origin of atmospheric lead and cadmium deposition over Hungary. *Időjárás 100*, 43-49.

Bozó, L., Alcamo, J., Bartnicki, J. and *Olendrzynski, K.*, 1992: Total deposition and budgets of heavy metals over Eastern Europe. *Időjárás 96*, 61-80.

Mészáros, E., 1999: Fundamentals of Atmospheric Aerosol Chemistry. *Akadémiai Kiadó, Budapest*

Mészáros, E., Barcza, T., Gelencsér, A., Hlavay, J., Kiss, Gy., Krivácsy, Z., Molnár, A. and *Polyák, K.*, 1997: Size distributions of inorganic and organic species in the atmospheric aerosol in Hungary. *J. Aerosol Sci. 28, 1163-1175.*

Olendrzynski, K., Anderberg, S., Bartnicki, J., Pacyna, J.M. and *Stigliani, W.*, 1995: Atmospheric Emissions of Cadmium, Lead and Zink in Europe During the Period 1955-1987. *IIASA Working paper* WP-95-35.

Pacyna, J.M, 1993: Emission inventory for atmospheric lead, cadmium and copper in 1989 in Europe. *A report for the TNO Netherlands Organization for Applied Scientific Research*, the University of Michigan, Ann Arbor, MI.

Pacyna, J.M., 1996: Emission inventories for HMs and POPs in Europe. *WMO GAW Report* No. 117. 4-32.

Pacyna, J.M., Münch, J., Alcamo, J. and *Anderberg, S.*, 1991: Emission trends for heavy metals in Europe. *Proc. Int. Conference on Heavy Metals in the Environment*, Glasgow.

IDŐJÁRÁS

Quarterly Journal of the Hungarian Meteorological Service
Vol. 104, No. 3, July–September 2000, pp. 173–196

Synoptic-climatological investigation of weather systems causing heavy precipitation in winter in Hungary

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(Manuscript submitted for publication 19 January 2000; in final form 20 July 2000)

Abstract–In this paper, winter cases associated with heavy precipitation are examined. The aim of the investigation was to determine as many characteristics of these cases as possible with regard to the geographical and temporal differences. Both the amount and the state of precipitation were considered in order to improve the forecast of winter precipitation. In the first part of the paper the database and methods used, as well as the main features of heavy precipitation events in winter are presented. The most important results connected with these include the quite limited duration and spatial extent of the investigated events, considerable geographical (West vs. East) and seasonal differences, and quite low frequency of the cases with only snow. These results can be considered very useful because a good knowledge of the characteristics of heavy precipitation events is essential to their prediction. The second part of the paper contains the classification of heavy precipitation events (synoptic-climatological examination) and the research called parametrical investigation including the examination of the connection between some meteorological parameters and the state of precipitation. Results have shown that the connections between the macro-synoptic types and winter precipitation are quite strong. Nevertheless, these alone cannot be used for the considerable improvement of precipitation forecasts. Further investigations on smaller scales (sub-synoptic scale, mesoscale, microscale) are needed in order to reveal the important processes determining the development of precipitation. The first step into this direction has already been made in this research. Using a mathematical-statistic method (Bayes-decision) an accurate procedure for the objective forecast of the state of precipitation was developed, which can be successfully used in the operational work, too.

Key-words: winter precipitation, heavy precipitation events, precipitation forecast, synoptic-climatology, state of precipitation, Bayes-decision.

1. Background

Nowadays, numerical models are widely used in many areas of meteorology. They provide valuable information for the operational short-term weather forecast, too. The development of these models has led to a substantial improvement of short-term weather prediction, but there are some meteorological elements, which cannot be forecasted exactly enough using these models. One of them is precipitation.

There are many reasons for the inaccuracy of precipitation forecasts by numerical models. The main problem is associated with the parameterisation of precipitation processes in models with limited resolution in space and time. The only efficient way to improve precipitation forecasts is to examine as many past cases as possible and to try to determine what the dominant processes connected with the development of precipitation are. A proper method is based on synoptic-climatology, which means an investigation of meteorological parameters as a function of the synoptic situation. The rapid development of synoptic-climatology began in the 40-s, when *Baur* (1948) made the first classification of synoptic situations for Europe. Since then many new classifications have been made and much research has been carried out using these weather types.

Many synoptic-climatological studies have been made in connection with precipitation. One of these was performed by *Maddox et al.* (1979). They investigated 151 intense mesoscale weather systems resulting in heavy precipitation between 1973–1977 in the USA. It was found that these systems develop in one of 4 different synoptic situations. Another similar investigation was made by *Bartels* (1989), who studied 82 cases with heavy precipitation between 1931–1980 in Germany. He showed that orografic lifting by the Alps leads to a large difference between the northern and southern parts of Germany in the characteristic weather types related to heavy precipitation.

According to *Lauscher* (1985), Mediterranean cyclones play an important role in the weather of South- and East-Austria. One of his results was that a special kind of Mediterranean cyclone, which passes through the Carpathian Basin and touches the eastern part of Austria (these are also called Vb cyclones; their name created by *Van Bebber* dates back to the end of the 19th century), causes the highest mean daily precipitation amount in Vienna, namely 9.7 mm.

Similar research has been carried out in Hungary, too. *Péczely* (1961), who created the weather types for the Carpathian Basin, investigated some quantities in connection with precipitation. He determined the synoptic situations in which the mean daily precipitation amounts are the largest in Hungary. It was also shown that the situation with Hungary located in the warm sector of a Mediterranean cyclone is very favorable to the development of large precipitation amounts.

Bodolainé (1983) examined the synoptic conditions of floods on the basin of the Danube and the Tisza rivers. She classified the weather types that can lead to flooding using a period of over 100 years. The greatest merit of her classification is that the weather types are defined by relatively stable meteorological fields both on the surface and in the lower Troposphere. These fields always describe a process and not a momentary state of the atmosphere like many other classifications do.

Verifications have shown that the biggest errors of precipitation forecasts occur in case of large amounts of precipitation. Therefore, it is worth performing research in connection with weather systems causing heavy precipitation. Although much research on heavy precipitation during the warm season has occurred, there are only few investigations associated with large precipitation amounts in winter. The main reason for this dearth of research is that such situations are generally not characteristic of this season. But it also must be taken into account that winter precipitation in certain states can cause dangerous situations even if its amount is not too large. Nevertheless, only case studies have mostly been made in this topic. The work of the Spanish researchers *Sellés* and *Franch* (1994) can be considered as an example of the very few exceptions. They investigated the snowfalls in the south-eastern part of the Pyrenees using data from a 30-year period. Among others it was shown that on average 31% of the annual precipitation falls as snow and the biggest daily amounts of snow are caused by Mediterranean cyclones.

In this paper, the gap mentioned above is to be filled. Our aim is to examine winter situations with heavy precipitation in Hungary from as many points of view as possible with regard to the spatial and temporal differences. The biggest part of the research is made up by a synoptic-climatological investigation. The main purpose of this examination was to find out to what extent processes on the synoptic scale determine the amount and state of precipitation. Recent numerical models are able to forecast the macro-synoptic situation quite accurately for a few days. So if there is a very strong connection between the macro-synoptic types and precipitation, the forecast of the latter can be considerably improved. Nevertheless, investigations on smaller scales are absolutely necessary, because many of the important processes determining the development of precipitation are related to these. Accordingly, in the second part of the research, it was tried to carry out investigations on smaller scales. Both the amount and the state of precipitation were involved in this examination, but only the results associated with the latter are presented in the paper.

2. *Database and methods*

The weather systems causing large amounts of precipitation occurring in the winter periods of the years between 1986 and 1997 in Hungary were investigated. The number of the cases in these 11 winter periods was sufficient to carry out statistical investigations. The winter periods consisted of the months November, December, January, February and March. The months November and March were involved because winter precipitation (snow, sleet, freezing rain, etc.) quite often occur in both of these months in Hungary. In order to determine geographical differences, Hungary was divided into six parts each representing a characteristic region of the country: 3 areas in West- and 3 areas in East-Hungary (*Fig. 1*).

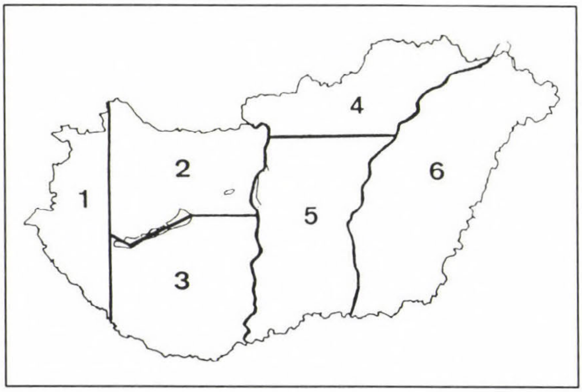

Fig. 1. The six areas of Hungary.

Those days of the mentioned periods were selected when the areal average of the daily precipitation amount was at least 5 mm in at least one of the areas. These cases were regarded as heavy precipitation events. The threshold value of 5 mm can be considered adequate because areal averages of the daily precipitation amount exceeding this value are quite rare in Hungary in winter. The areal average of the precipitation was calculated by the arithmetic mean of the daily precipitation data of the meteorological stations in the area. Data of all meteorological stations were used where precipitation amount is measured, so the calculation can be regarded as quite accurate in spite of using a relatively simple method.

The state of precipitation, the forecast of which is not an easy task during the winter period, was also considered in this research. It is also quite difficult, however, even to determine the state of precipitation occurring earlier because of the large spatial and temporal changes. Relatively detailed data on precipitation type are only available for synoptic stations, but at limited spatial resolution. If all the meteorological stations are considered, then there are sufficient data in space but not in time, because the stations measuring only precipitation report precipitation type just once a day, namely the one with the highest code. Therefore, it has to be accepted that the state of precipitation cannot be determined without errors.

The selected cases were also classified using two different methods, namely Péczely's and Bodolainé's weather types. Péczely's weather types (*Péczely*, 1957, 1983) are defined only by the surface pressure field, but contain all the synoptic situations. Their codes and explanation are shown in the Appendix. Bodolainé's weather types (*Bodolainé*, 1983) are defined by the fields of several meteorological parameters (not just pressure), namely absolute topography of the 500 mb level (AT500), relative topography of the 500/1000 mb level (RT500/1000), surface pressure and precipitable water, but contain only those synoptic situations that can lead to flooding (see the Appendix). In spite of that, they are more suitable for research connected with precipitation than Péczely's weather types.

In the second part of the study the connections between some meteorological parameters and the amount and state of precipitation were examined. In order to carry out these investigations for the weather systems causing heavy precipitation, the fields of the parameters are needed. Because of the sparsity of upper-air data, the data of only one station, Pestlőrinc (Budapest) was used to calculate the parameters for this point only. Those days of the winter periods mentioned earlier were selected when the daily precipitation amount was at least 5 mm at this station. This data is measured from 06 UTC to 06 UTC next day, so this 24-hour period can be divided into the following parts: 06 UTC–18 UTC and 18 UTC–06 UTC. Data from the 12 UTC sounding are taken to refer to the first period and those from the 00 UTC sounding to the second period. Only those 12-hour periods were considered, in which precipitation actually occurred. In this part of our research, emphasis was placed on the state of precipitation. A mathematical-statistic method called Bayes-decision was used to determine the threshold value that separates the values of a certain parameter characteristic for the liquid and the solid state in the most optimal way from a statistical point of view. That makes possible the objective forecast of the state of precipitation, too.

It must be emphasised that—even if it is not always mentioned—everything presented in the paper is relevant only to the cases with heavy precipitation, as defined above.

3. *Characteristics of weather systems causing heavy precipitation in winter*

One of the most important features of heavy precipitation events is their frequency. In *Fig. 2* it is shown how often cases with an areal average of the daily precipitation amount of at least 5, 10 or 20 mm occurred during the investigated period in at least one of the 6 areas. There is a marked seasonal change in the number of the cases. Heavy precipitation events mostly occurred in November and the fewest cases were observed in January. Also, weather systems causing a daily precipitation amount of at least 20 mm on areal average are quite rare in Hungary during winter time. Their mean frequency is smaller than 1 even in November!

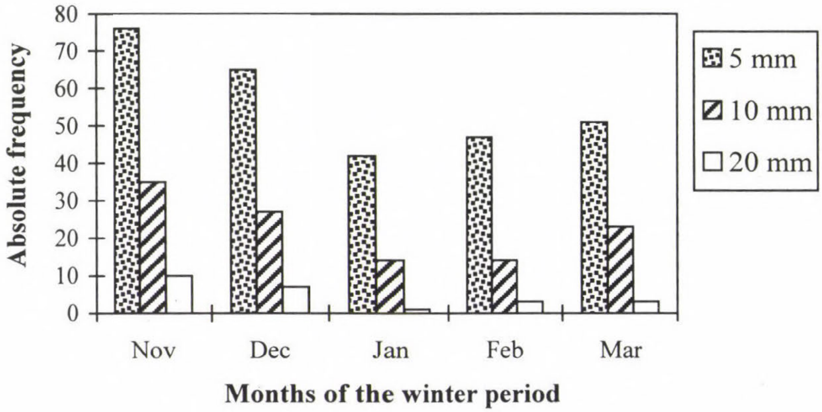

Fig. 2. Frequency of the days with areal average precipitation amount of at least 5, 10 or 20 mm.

The next step was to determine the spatial differences. It was found that in each area the monthly change of the frequency of heavy precipitation events is quite similar to the country-average mentioned earlier, except in East-Hungary, where the majority of the cases occurred in December and not in November as in the other parts of the country. *Fig. 3* compares the number of heavy precipitation events that occurred in the 6 areas during the investigated period. It shows that in Transdanubia (Area 1, 2 and 3) more cases occurred than in the eastern part of the country. It also can be seen that in two of the eastern areas (in the Northern-Mountains area and in East-Hungary) there were no days during the investigated period when the areal average of the precipitation amount was at least 20 mm. Meanwhile, there were days with a precipitation amount of more than 27 mm on areal average in all the three

Transdanubian areas, and the maximum value was 35 mm, which occurred in South-Transdanubia.

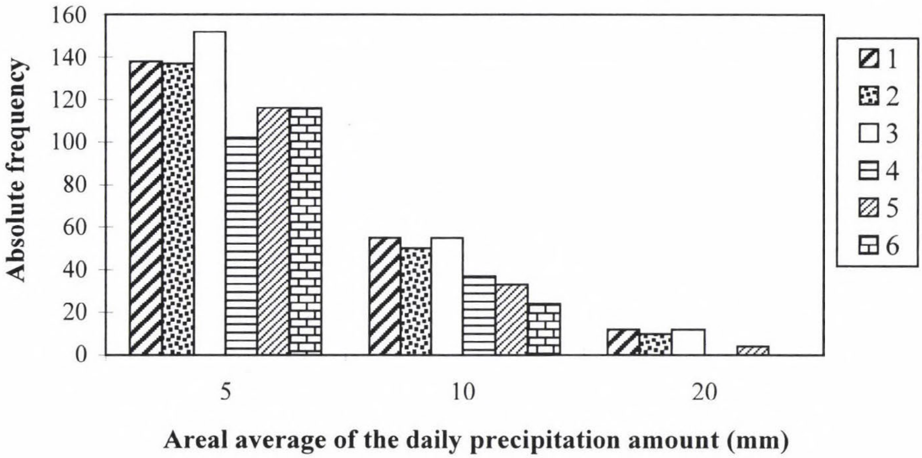

Fig. 3. Frequency of the days with areal average precipitation amount of at least 5, 10 or 20 mm in the six areas.

The maximum daily precipitation amounts were also investigated in our research. It was found that values over 40 mm are quite rare in Hungary during the winter period, but they occurred in the western part of the country much more frequently than in the eastern part. The absolute maximum daily precipitation amount was 75 mm and it was measured in North-Transdanubia. A very close connection was found between the areal average of the daily precipitation amount and the maximum daily precipitation amount measured in the same area. It can be stated that the maximum value is at least 1.5 times higher and on average twice as high as the areal average. Therefore, in case of an areal average precipitation exceeding 20 mm the maximum value must be more than 30 mm and on average more than 40 mm. The possibility that in East-Hungary the lack of cases with a daily precipitation average of at least 20 mm is due to the relatively large extent of the area compared to the other areas can be dismissed, because this was the region where the fewest cases with a maximum value over 40 mm occurred (only 2 cases).

The areal extent of heavy precipitation events was examined as well. Mostly only one or two areas are involved. The cases in which heavy precipitation occurs throughout the country are quite infrequent. It was also shown that heavy precipitation events including at least half of Hungary (3 areas or more) mostly occur in November and their number is decreasing during the winter period, whereas the frequency of the cases when much precipitation occurs in only one or two areas has a minimum in January and maxima in November, December and March.

Following the many spatial characteristics it is worth mentioning some temporal features, too. In *Fig. 4* the distribution of the length of those periods is shown when the areal average of the precipitation amount was at least 5, 10 or 20 mm in at least one of the areas every day of the period. Even in case of 5 mm the heavy precipitation period usually lasts only 1 day. Considering precipitation amounts of at least 20 mm, only one case during the investigated period lasted for 2 days. The maximum length of the period consisting of days with a precipitation average of at least 5 mm is 5 days and for 10 mm is only 3 days. The time intervals between heavy precipitation events vary quite irregularly. There were 3 winter periods without a daily precipitation amount of at least 20 mm on areal average, but there were also periods lasting more than 70 days without a daily precipitation average of at least 5 mm.

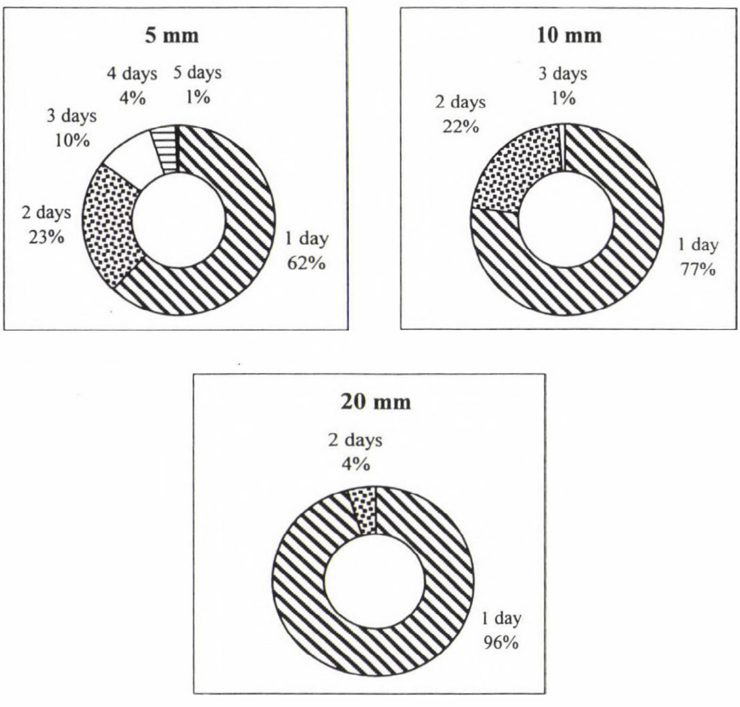

Fig. 4. Distribution of the length of heavy precipitation periods.

Next the state of precipitation was investigated. The type of precipitation was determined in the areas in case of heavy precipitation events. The types used are listed in *Table 1*. The type "only freezing rain" is absent from Table 1 because this case did not occur during the investigated period. Results from the cases with a precipitation amount of at least 10 mm on areal average can be seen in *Fig. 5*. Here the distribution of the precipitation types are shown with those areas where the areal average of the daily precipitation amount was at least 1, 5 or 10 mm. Therefore these results do not refer to one area but to that part of the country where a given amount (1, 5 or 10 mm) of precipitation occurred. In case of the biggest part of heavy precipitation events, only rain occurs. The frequency of cases with snow and rain is also high, but there are not many cases with only snow. Furthermore, if the areas with smaller amounts of precipitation are also considered, their frequency is even lower. Thus, cases with heavy precipitation and only snow in the whole country are very rare in Hungary and they require special conditions. The frequency of freezing rain is also quite low; it occurs more frequently when both rain and snow are falling.

Table 1. Precipitation types used in the research

Precipitation type	**Explanation**
Rain	Only rain in the whole area during the whole 24-hour-period
Snow	Only snow in the whole area during the whole 24-hour-period
Snow, rain	Both snow and rain in the area
Snow, rain, freezing rain	Besides snow and rain, also freezing rain in some parts of the area
Snow, freezing rain	Besides snow only freezing rain in some parts of the area
Rain, freezing rain	Precipitation in fluid state, part of it freezing rain

Spatial differences were also investigated. The relative frequency of cases with only rain is quite similar in the whole country. The same cannot be said for snow, because in the western part of Hungary more cases occurred (the relative frequency is about 25% for precipitation amounts of at least 10 mm occurring in the given area), than in the eastern part of the country. Surprisingly, it was the Northern Mountains area where the relative frequency of the cases with only snow was the lowest during the investigated period for the days with a precipitation amount of at least 10 mm. The most plausible explanation for that will be given in section 4. Finally, it was also found that the frequency of freezing rain during heavy precipitation events is much higher in the eastern areas than in Transdanubia. The change of the characteristic precipitation type during the winter period was investigated very thoroughly as well. It was found among others that freezing rain was the only type that did not occur in each month of the winter period. In March there were no big precipitation events with only snow in any of the eastern areas, but they did occur in all the western areas.

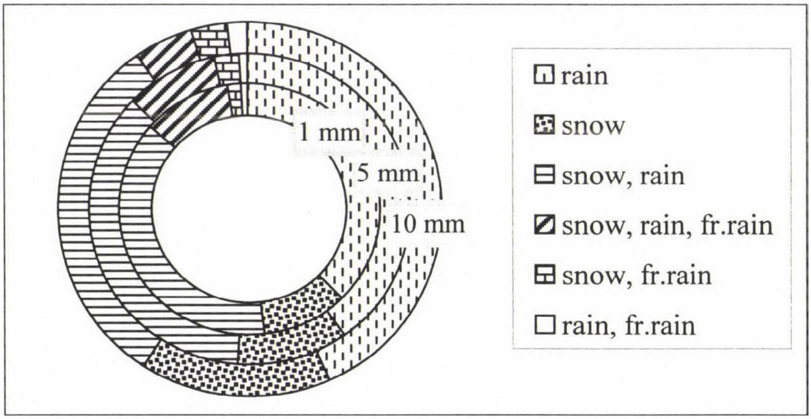

Fig. 5. Distribution of the precipitation types with daily precipitation amount of at least 10 mm considering the areas with a precipitation average of at least 1, 5 or 10 mm.

4. *Classification of heavy precipitation events*

As mentioned in section 2, Péczely's and Bodolainé's weather types were used to classify heavy precipitation events. Péczely's weather types were determined for every day of the investigated period, and since they describe only a momentary state of the atmosphere, it seemed to be expedient to do the classification twice for each day (00 and 12 UTC). It is important to mention that our way of classification does not exactly correspond to that of Péczely's. The main reason for that is that Péczely's method for distinguishing the anticyclonic from cyclonic types by determining whether the surface pressure is over or under 1015 hPa in the biggest part of Hungary was not strictly applied in our investigation. The relative value of pressure compared to the surrounding areas was considered to be much more important than its absolute value.

Fig. 6 shows the frequency of Péczely's weather types that occurred in case of large amount of precipitation during the investigated period. More than half of heavy precipitation events were caused by the weather type CMw (Hungary is situated in the warm sector of a Mediterranean cyclone). The frequency of the weather types CMc and C is also quite high, but considerably lower than that of CMw. According to Fig. 6, it is obvious that anticyclonic types do not play an important role in the development of heavy precipitation events in winter. In the following, only cyclonic types will be considered. The types zC and mCw are mostly only the antecedents of heavy precipitation events because their frequency is much higher immediately before the onset of heavy precipitation than during such events.

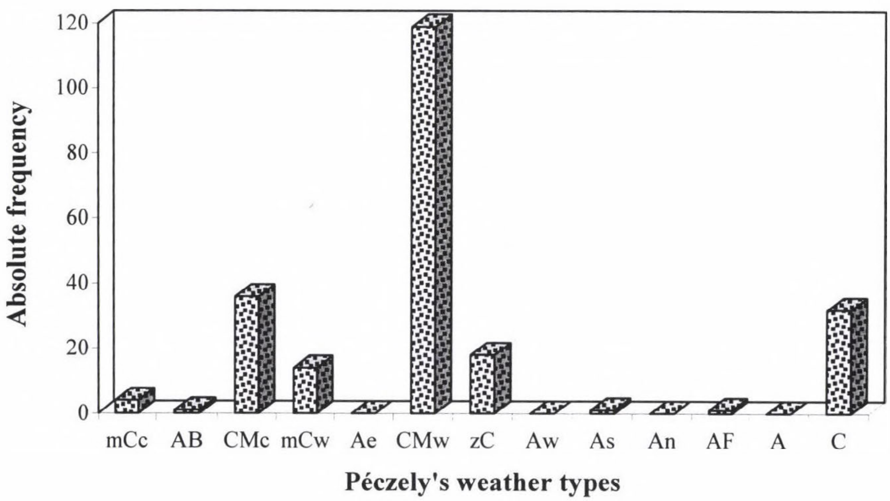

Fig. 6. Frequency of Péczely's weather types with daily precipitation amount of at least 10 mm on areal average.

As for spatial differences, it was found that in all areas of the country the type CMw has the highest relative frequency, but this value becomes gradually lower toward the East. Exactly the opposite can be said for the type CMc, which more often causes heavy precipitation in the eastern areas than in the western ones. The only cyclonic type that did not occur in all the areas in case of heavy precipitation is mCc. Valuable information can be obtained by determining the distribution of the areal average of the precipitation amount for each weather type. The most important result of this examination is that for all weather types the daily precipitation average was less than 5 mm in more than

60% of the cases. This means that even in case of CMw, which most frequently causes heavy precipitation events throughout the country, much precipitation on areal average is not at all guaranteed. That is, it is not enough to know the synoptic situation, it is also necessary to consider other processes and parameters connected with the development of large amounts of precipitation. The results for West-Transdanubia are shown in *Fig. 7*.

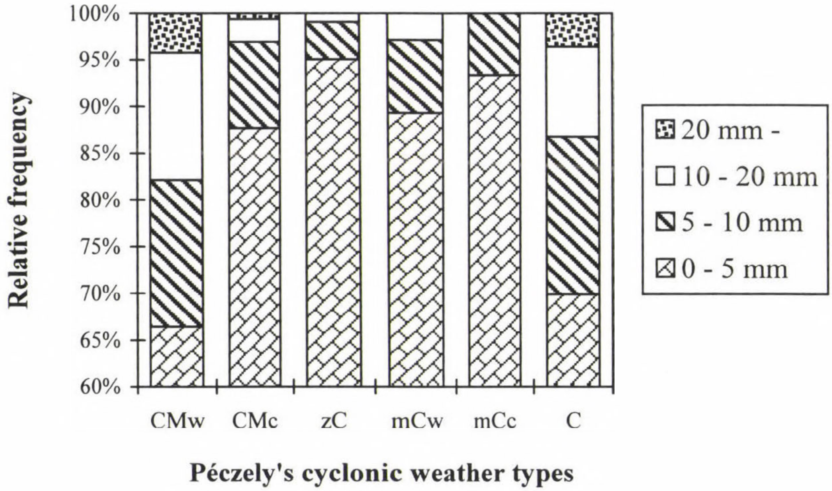

Fig. 7. Distribution of areal average daily precipitation amount in case of Péczely's cyclonic weather types in West-Hungary.

This is the only area where CMw is the weather type with the highest probability of developing daily precipitation averages of at least 5 mm (and also 10 mm). In all the other parts of the country heavy precipitation events are most probable in case of the cyclone centre type (C). The type with the least probability of heavy precipitation is zC in West- and North-Transdanubia, and mCc in the other parts of the country. It was also shown that in the winter period, daily precipitation amounts of at least 20 mm on areal average occur only in case of Mediterranean cyclones and cyclone centres over Hungary. Cases with heavy precipitation in at least half of the country are also mostly caused by these weather types.

One of the problems connected with Péczely's classification is that it is not able to separate the situation with a real cyclone centre from the situation with a temporary cyclone centre connected with the passage of a Mediterranean cyclone through Hungary. This was the primary reason for using

Bodolainé's weather types as well to classify heavy precipitation events. These types were determined only for the days with a precipitation amount of at least 10 mm on areal average in at least one of the areas. The results are shown in *Fig. 8*.

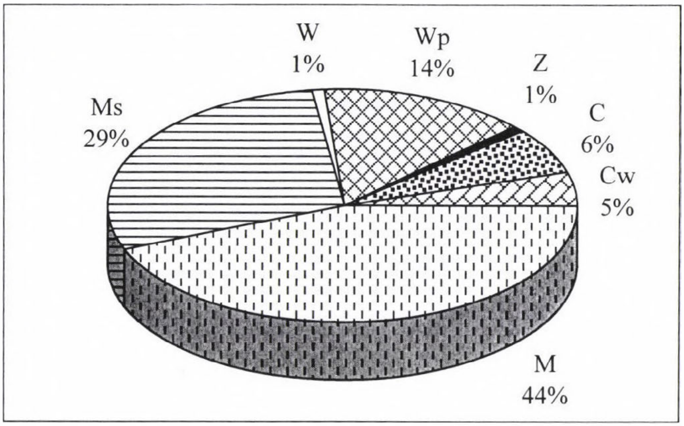

Fig. 8. Relative frequency of Bodolainé's weather types with daily precipitation amount of at least 10 mm on areal average.

A new type with the code Ms can be seen as well, which is not included in Bodolainé's weather types. It was necessary to create it because it quite often caused large amounts of precipitation on areal average, and it could not be classified as any of the Bodolainé's weather types. The type Ms represents a Mediterranean cyclone situated south from Hungary moving very slowly or not at all. This slow motion is caused by a huge blocking anticyclone with centre over Northeast-Europe or by the development of the cyclone also on higher levels of the atmosphere. In this case the deep trough characteristic for Bodolainé's transposing Mediterranean cyclone (type M) is missing. The fast motion of the type M is provided by the very strong south-westerly current in the east side of the trough. Cyclones belonging to the type Ms are mostly much deeper than those belonging to the type M and their genesis is also different in the most cases. Transposing Mediterranean cyclones usually form as frontal waves on cold fronts passing over the Alps. Cyclones belonging to the type Ms can develop directly in the Mediterranean (e.g. as a result of cold surges) or by the movement of cyclones formed over North-Africa or the Atlantic into this region. According to Fig. 8, in 73% of the cases the development of large amount of precipitation was connected with Mediterranean cyclones. From this point of view, Wp is relatively important as well, but the real cyclone centre situations (type C) caused only 6% of all cases. Type W and Z occurred only once and type H did not occur in case of heavy precipitation events during the investigated period.

Seasonal changes were investigated as well. The frequency of the type transposing Mediterranean cyclone (M) is the highest, except for the months January and February, when Ms is the type that most often causes heavy precipitation events. This is due to the higher frequency of blocking anti-cyclones over Northeast-Europe in these months. The real cyclone centre type (C) did not occur in December and January in case of heavy precipitation events, whereas Wp caused about 30% of the cases in these months. As for the spatial differences, the type M turned out to cause the most heavy precipitation events (40–55%) in every area. From this point of view, the type Ms is also very important in most of the areas, especially in West- and North-Transdanubia and in the middle part of Hungary. The relative frequency of the type Ms in case of a daily precipitation amount of at least 10 mm on areal average is lowest in the Northern Mountains area. This is also the area where the role of Mediterranean cyclones (M and Ms) is the least important in the country in the development of heavy precipitation, but their frequency exceeds 60 % even in this part of Hungary. Finally, the relative frequency of types C and Cw in such cases is much higher in the Northern Mountains area and in the middle part of Hungary than in the other regions of the country.

One of the advantages of creating the type Ms can be seen in *Fig. 9*, where some of the results of our investigation concerning the connection between the synoptic situation and the state of precipitation are shown. Only the areas with a precipitation average of at least 5 mm were considered. About 80% of the cases with only snow were caused by the type Ms. The other Mediterranean cyclone type, M turned out to cause only 6–7% of such cases. Type Ms provides favourable conditions for the development of freezing rain as well, whereas the type M mostly causes cases with only rain or both rain and snow. Another important result is that the real cyclone centre type (C) did not cause any cases with only snow or cases with freezing rain during the investigated period.

At the end of this section, we attempt to give an explanation for our most surprising result mentioned in section 3, namely that it is the Northern Mountains area where the cases with only snow make up the smallest part of heavy precipitation events. It has been shown in this section that heavy precipitation events with only snow occurred in about 80% in case of the type Ms. It has also been mentioned that for weather type Ms it is the Northern Mountains area where the probability of heavy precipitation is the smallest in the country. Therefore, it follows that it is the Northern Mountains area where the relative frequency of heavy precipitation events with only snow is the lowest, because the weather type Ms, which is the most favourable for such cases, does not often provide suitable conditions for the development of large amount of precipitation in this area. Here the type M causes heavy precipitation events much more often. For this weather type, higher temperatures are characteristic in the lower Troposphere than for the type Ms, so precipitation mostly falls as rain or both snow and rain even in this mountainous area.

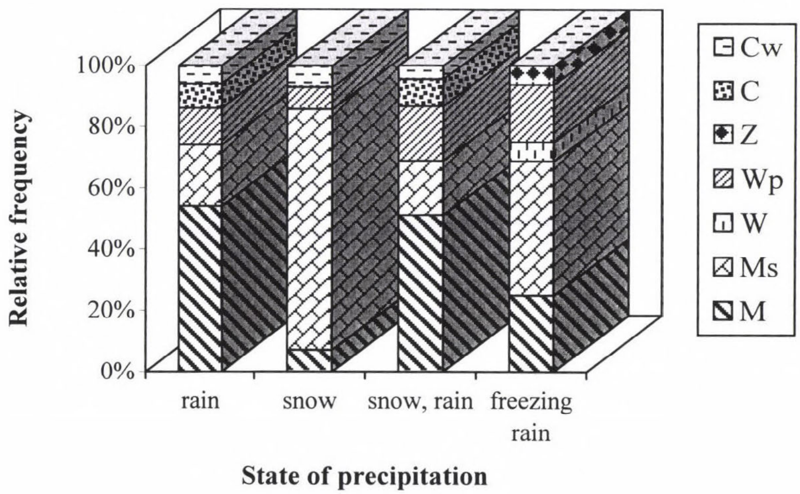

Fig. 9. Relative frequency of Bodolainé's weather types with daily precipitation amount of at least 10 mm on areal average as a function of the state of precipitation.

5. *Parametrical investigations*

In the last part of our research the connections between some meteorological parameters and the amount and state of precipitation were examined. In this section some of our results concerning the state of precipitation will be presented. As mentioned in section 2, the data of the station at Pestlőrinc (Budapest) were used and only those days of the investigated period were considered when the daily precipitation amount was at least 5 mm at this station. Data from the 12 UTC soundings were taken to refer to the first half of these days and those from the 00 UTC soundings to the second half of them.

119 days were found which met the criteria above. They were divided into 238 12-hour periods, but precipitation was observed only in 197 of these cases. First, the distribution of the state of precipitation for these periods was determined. In 65% of the cases only rain occurred and the percentage of the

cases with only snow was merely 20%. Both rain and snow occurred in 11% of the cases and the relative frequency with partly freezing rain made up only 4%. Next, it was examined which surface temperatures (Ts) are characteristic for the different precipitation types (*Fig. 10*). The precipitation types used are shown in *Table 2*. It can be seen that precipitation of liquid and solid state mostly occurred in case of positive and negative surface temperatures, respectively. Between -4°C and +4°C, however, both types were observed and also these are the most characteristic temperature values for the mixed type.

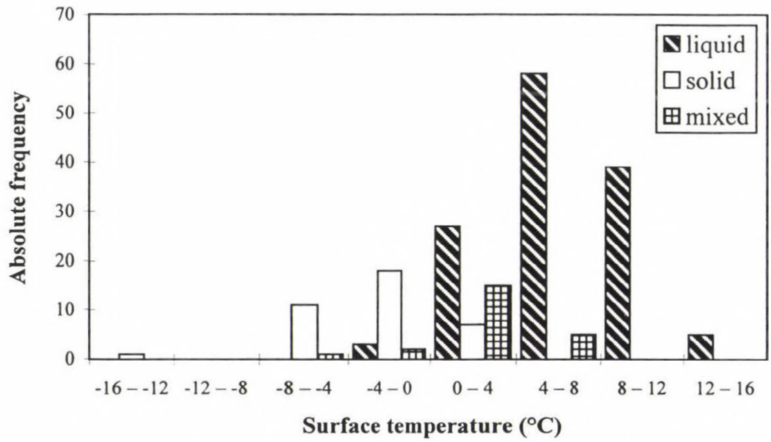

Fig. 10 Distribution of surface temperature for heavy precipitation events in winter as a function of the state of precipitation.

Table 2. Precipitation types used in the research

Precipitation type	Explanation
Liquid type	Only rain or mostly freezing rain during the whole period
Solid type	Only snow during the whole period
Mixed type	Both snow and rain or partly freezing rain

The state of precipitation observed on the surface is influenced by all the layers of the atmosphere that the precipitation elements were falling through. Therefore, the precipitation type reflects the meteorological conditions of the whole lower Troposphere. Although there is connection between the surface temperature and the state of precipitation (as shown in Fig. 10), in most cases the temperature measured near the surface (at a height of 2 m) is obviously not able to influence the precipitation type. The only exception is freezing rain that is caused by negative temperatures near and on the surface. Mostly it is just the other way around, that is the state of precipitation influences the surface temperature. This fact can be illustrated with the following example: if the surface temperature is positive and in spite of that the precipitation elements reach the surface as snow, then after a while the surface temperature is gradually decreased by the thawing of the snow. So other parameters had to be found that really influence the state of precipitation. The parameters chosen can be seen in *Table 3*. The last two parameters provide information about the temperature conditions of atmospheric layers and not only single levels like the other 4 parameters do. After determining the distribution of each parameter in case of the different precipitation types, it was found that there were significant differences in the extent of overlapping the values characteristic for the liquid and the solid state in case of the different parameters. It could be stated that the higher the level, the less the values were separated, resulting in larger overlapping. In case of RT850/1000 the values were quite separated (small overlapping).

Table 3. Meteorological parameters used in the research

Parameter	Explanation
T925	Temperature of the 925 hPa level
T850	Temperature of the 850 hPa level
T700	Temperature of the 700 hPa level
T500	Temperature of the 500 hPa level
RT850/1000	Thickness of the layer between 850 and 1000 hPa
RT500/1000	Thickness of the layer between 500 and 1000 hPa

In the following, a mathematical-statistic method based on the Bayes-decision (*Dévényi* and *Gulyás*, 1988) was used to determine the threshold value that separates the values of a certain parameter characteristic for the liquid and the solid state in the most optimal way from a statistical point of view. In this investigation only that part of the cases was considered in which only snow or only rain was observed. Two random variables were used: one of the above-mentioned parameters was regarded as the predictor and the state of precipitation (snow or rain) as the predictand. It is important to emphasise that the value of the predictor was assumed to be known. The function, with the help of which the predictand is estimated from the value of the predictor, is called decision function. The accuracy of the estimation is given by the so-called loss function, which is only a function of the difference between the real and estimated values of the predictand. The loss function is also a random variable and its expected value is called the risk of the estimation.

Our aim was to find the Bayes-decision, that is the decision function with the minimum risk in case of a certain loss function. Since our predictand variable is discrete and it can have only the values "1" and "2" (snow, rain), the decision function is equivalent to dividing the possible values of the predictor variable into two disjoint sets. In this case the decision function operates as follows: if the value of the predictor is in set1, the value of the predictand is "1" (that is the estimation of the state of precipitation is snow) and if the value of the predictor is in set2, the predictand has the value "2" (that is the estimation of the state of precipitation is rain). The predictor variable can be a vector (*Matyasovszky et al.*, 1993), but in our case it was chosen as a scalar, that is it always contains only one of the mentioned parameters. This means that the elements of the two sets are also scalar, so these sets can be simply created by determining the threshold value that separates the values of the predictor belonging to set_1 from those belonging to set2. It can be shown that in this case the sets equivalent to the Bayes-decision are the following:

$$c_1^* = \{x \mid W_{21} \cdot p_2 \cdot f_2(x) \le W_{12} \cdot p_1 \cdot f_1(x)\}, \tag{1}$$

$$c_2^* = \{x \mid W_{21} \cdot p_2 \cdot f_2(x) \ge W_{12} \cdot p_1 \cdot f_1(x)\}. \tag{2}$$

W_{21} and W_{12} are the elements of the loss matrix (equivalent of the loss function in case of a discrete predictand), and both were given the value "1", because both kinds of poor estimations (snow was estimated but rain occurred and the other way around) were regarded as equally important. In the formula above p_1 and p_2 mark the probability of snow and rain, whereas f_1 and f_2 are the conditional density functions of the predictor for snow and rain. Considering the value of W_{21} and W_{12}, it is obvious that the threshold value can be obtained from the following equation:

$$p_1 \cdot f_1(x) = p_2 \cdot f_2(x). \tag{3}$$

The four components of this equation were estimated using the cases that occurred in the investigated period. Probability of snow and rain (p_1 and p_2) were substituted with their relative frequency during the investigated period. The distribution of the predictor variable was assumed to be normal, so the estimation of the expected value by the mean of the sample and that of the variance by the empirical dispersion were sufficient to determine f_1 and f_2. The threshold value could be simply obtained by determining the value that satisfies the equation above. The products on both sides of the equation can also be interpreted as the conditional density functions of the predictor weighted with the probability of the given state of precipitation, and the curves belonging to the two functions intersect at the threshold value.

In order to demonstrate this, *Fig. 11* was made, where the weighted conditional density functions are represented in two cases, namely when RT850/1000 and when T500 was chosen as the predictor variable. These two parameters were selected intentionally, because RT850/1000 is the parameter with the smallest and T500 is the parameter with the largest overlapping the two density functions among the investigated parameters. It also can be seen that in case of T500 the value belonging to the intersection of the two curves (threshold value) is smaller than the expected value of T500 in case of snow, that is the probability of rain is higher than that of snow even if the predictor has the value, in case of which snow most often occurs. A similar statement can be made only in case of T700 (*Table 4*). Obviously, it is worth using only those parameters for estimating the state of precipitation, for which the threshold value is between the two expected values. In order to find out how exact the estimations are in case of the mentioned parameters, the risk of the estimation was determined in each case.

Fig. 11. Weighted conditional density functions of two meteorological parameters for heavy precipitation events in winter.

According to its definition, the risk corresponds to the area situated under the intersection of the two density functions bordered by the curves belonging to these functions (this area has been called overlapping so far). Using an approximation the risk was determined for all parameters investigated. It is made up of two parts, one contains the probability of estimating snow in case of rain (Errorl) and the other is the probability of the reversed situation (Error2). In *Table 5* all the three values are given as a percentage. It is important to emphasise that these values are only valid when the value of the predictor is exactly known. In order to forecast the state of precipitation the value of the predictor has to be forecasted as well, with the help of which the estimation is carried out. So the real value of the risk is typically higher than that given by Table 5 depending on how accurate the forecast of the predictor is. Although Ts is the third best predictor among the investigated parameters, there is no value in using it, because its forecast is quite inaccurate as mentioned earlier. The risk is in case of T500 and T700 the largest, which indicates that the temperature conditions over about 3000 m do not influence the state of precipitation too much.

Table 4 Conditional expected values of the investigated parameters and the threshold values

Parameter	Expected value for snow	Expected value for rain	Threshold value
Ts	-2.8°C	6.7°C	0.7°C
T925	-4.5°C	5.1°C	-1.0°C
T850	-5.4°C	2.0°C	-3.1°C
T700	-10.3°C	-6.3°C	-10.7°C
T500	-25.7°C	-22.7°C	-28.7°C
RT850/1000	1281.0 gpm	1326.9 gpm	1298.0 gpm
RT500/1000	5315.6 gpm	5426.3 gpm	5339.8 gpm

Table 5. Mean error of the estimation of the state of precipitation using different parameters as predictor variables

Parameter	Errorl (%)	Error2 (%)	Risk (%)
Ts	2.05	3.09	5.14
T925	1.90	2.20	4.10
T850	3.45	3.34	6.79
T700	4.49	12.56	7.05
T500	2.67	18.00	20.67
RT850/1000	*1.68*	*1.88*	*3.56*
RT500/1000	4.95	6.75	11.70

Cases with freezing rain were also investigated, but their number was too small to draw conclusions. Another problem was that in most of the cases freezing rain occurred between two soundings, so the data available can not be considered representative. The detailed investigation of freezing rain requires the extension of the examination to the cases with a daily precipitation amount less than 5 mm, because freezing rain very seldom occurs in case of heavy precipitation events investigated in this research.

6. Conclusion

First, it can be stated that many important features of heavy precipitation events in winter were determined. Probably the most important of these was that very special conditions are required for large amounts of precipitation in form of only snow in the whole country. Further results included:

- the marked change in the frequency of the cases and the characteristic precipitation type during the winter period;
- considerable geographical differences (West and East) in the number of the cases, precipitation amounts and the percentage of several precipitation types;
- quite limited duration (usually 1 or 2 days) and spatial extent (mostly only 1 or 2 areas, that is, less than half of the country was involved).

The most important results relevant to the connection of the synoptic situation and heavy precipitation events (HPE) in winter:

- HPEs are mostly caused by Mediterranean cyclones (Péczely's CMw, CMc and Bodolainé's M, "Ms" type), but the cyclone centre type has the highest probability of developing such cases (30–40%) except for West-Hungary.
- The new macro-synoptic type Ms, which was created in order to make Bodolainé's classification complete in terms of this research, caused about 80% of the cases with only snow and nearly 50% of the cases with freezing rain.
- A plausible explanation for the surprising finding about the Northern Mountains area (percentage of the cases with only snow) was given.

Finally, parametrical investigations have led to many valuable results as well including:

- The characteristics of the layer between the surface and about 1500 m are the most important concerning the state of precipitation.

- Using a mathematical-statistic method based on the Bayes-decision a procedure for the objective forecast of the state of precipitation was developed.
- RT850/1000 proved to be the best parameter investigated the mean error of the estimation being only 3.6% in case of using this parameter as the predictand.

To sum up the facts above, it can be stated that synoptic-climatological investigations are able to improve our understanding of the processes connected with the development of precipitation. A considerable improvement of precipitation forecasts, however, cannot be achieved without carrying out comprehensive examinations on smaller scales (sub-synoptic scale, mesoscale, microscale) as well. Nevertheless, our investigation can be considered very useful because it pointed out the utmost importance of the processes on these scales even in winter. Possible further research could include the examination of the applicability of the so-called composite chart method (*Maddox*, 1979; *Bonta*, 1991) to winter cases.

Furthermore, the state of precipitation can be considerably influenced by precipitation intensity as indicated by *Steinacker* (1983), among others. The two main processes connected with that are cooling of the air by the evaporation of precipitation and the melting of snowflakes before reaching the ground. In the future, both of these processes should be taken into account when estimating the state of precipitation. This, however, might only be successful when much better quantitative precipitation forecasts are available.

Acknowledgements–The author would like to thank the Hungarian Meteorological Service (HMS) for providing the data required for this study. Special thanks are due to *Dr. Imre Bonta* (HMS) and *Dr. Zsuzsanna Iványi* (ELTE) for supporting my work.

References

Bartels, H., 1989: *Koordinierte Starkregenregionalisierung, Auswertung.* DeutscherWetterdienst Referat Hydrometeorologie, 04. 1989.

Baur, F., 1948: *Einführung in die Großwetterkunde.* Dieterichsche Verlagsbuchhandlung, Wiesbaden.

Bodolainé, J. E., 1983: *Synoptical Conditions of Floods on the Basin of the Rivers Danube and Tisza* (in Hungarian*).* OMSZ Hivatalos Kiadványai, LVI. kötet. Országos Meteorológiai Szolgálat, Budapest.

Bonta, I., 1991: *Investigation of weather systems causing heavy precipitation on the basin of the rivers Danube and Tisza* (in Hungarian).*Doctoral Thesis*, ELTE TTK Meteorológiai Tanszék, Budapest.

Dévényi, D. and *Gulyás, O.*, 1988: *Mathematical-statistic Methods in Meteorology* (in Hungarian). Tankönyvkiadó, Budapest.

Lauscher, F., 1985: *Klimatologische Synoptik Österreichs mittels der ostalpinen Wetterlagen-klassifikation. Arbeiten aus der Zentralanstalt für Meteorologie und Geodynamik, Heft 64.* Zentralanstalt für Meteorologie und Geodynamik, Wien.

Maddox, R. A., 1979: A methodology for forecasting heavy convective precipitation and flash flooding. *National Weather Digest: Flood 4*, 30-42.

Maddox, R. A., Chappell, C. F. and *Hoxit, L. R.*, 1979: Synoptic and Meso-α scale Aspects of Flash Flood Events. *Bulletin of the American Meteorological Society 60*, 115-123.

Matyasovszky, I., Németh, L., Németh, P. and *Takács, Á.*, 1993: Forecasting the state of winter precipitation (in Hungarian). *Beszámolók az 1990-ben végzett tudományos kutatásokról*, Országos Meteorológiai Szolgálat, Budapest, 46-53.

Péczely, Gy., 1957: *Großwetterlagen in Ungarn. Kleinere Veröffentlichungen der Zentralanstalt für Meteorologie, No.30.* Országos Meteorológiai Szolgálat, Budapest.

Péczely, Gy., 1961: *Climatological Characteristics of the Macrosynoptic Types of Hungary* (in Hungarian). OMSZ Kisebb Kiadványai, No.32. Országos Meteorológiai Szolgálat, Budapest.

Péczely, Gy., 1983: *Catalogue of the macrosynoptic types of Hungary (1881-1983)* (in Hungarian). OMSZ Kisebb Kiadványai, No.53. Országos Meteorológiai Szolgálat, Budapest.

Sellés, C. G. and *Franch, F. S.*, 1994: Snowfall analysis in the Eastern Pyrenees. 23.Internationale Tagung für Alpine Meteorologie. 5-9 September 1994, Lindau. *Annalen der Meteorologie 30*, 303-306.

Steinacker, R., 1983: Diagnose und Prognose der Schneefallgrenze. *Wetter und Leben 35*, 81-90.

APPENDIX

Péczely's weather types

Code	Explanation
mCc	Hungary is located in the rear of an East-European cyclone
AB	Anticyclone over the British Isles
CMc	Hungary is located in the rear of a Mediterranean cyclone
mCw	Hungary is located in the warm sector of a West-European cyclone
Ae	Anticyclone east from Hungary
CMw	Hungary is located in the warm sector of a Mediterranean cyclone
zC	Zonal cyclonic situation
Aw	Anticyclone extending from the west
As	Anticyclone south from Hungary
An	Anticyclone north from Hungary
AF	Anticyclone over the Fennoscandinavian region
A	Anticyclone with centre over the Carpathian-Basin
C	Cyclone with centre over the Carpathian-Basin

Bodolainé's weather types

Code	Explanation
W	West type
Wp	West with secondary disturbation type
Z	Zonal type
M	Transposing Mediterranean cyclone type
C	Cyclone centre type
Cw	West cyclone type

IDŐJÁRÁS

Quarterly Journal of the Hungarian Meteorological Service
Vol. 104, No. 3, July–September 2000, pp. 197–211

Preparation of regional scale wind climatologies

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(Manuscript submitted for publication 21 December 1999; in final form 11 August 2000)

Abstract—This paper presents newmethod and preliminary regional scale wind patterns for Europe, which fill the gap between the global climatologies and more local scale surface wind climatologies like the European wind atlas. The aim of establishing regional scale wind maps is to provide readily available regionally representative flow statistics. Preliminary maps based on 2 years of measured 850 hPa wind direction and speed data from 120 sounding stations show characteristic patterns (see Figs. 1, 2 and 3). Statistical analysis shows close correlation between short and long series of wind statistics. Such maps could be eventually used for top-down extrapolation of yearly mean surface wind statistics for data sparse areas by considering effects of terrain, surface and obstacles.

Key-words: regionally representative wind, environmental impact assessment, geostrophic wind, harmonization of meteorological data preprocessing, wind atlas.

1. Introduction

The paper describes preparatory work and preliminary results on a new method of wind analysis, which might serve towards the establishment of a regionally representative wind atlas. The research is part of efforts to harmonize meteorological data preprocessing. The atlas—an outcome of such work—might form a useful tool for top-down (considering locally representative surface, topographical and orographical conditions) wind inter- or extrapolation purposes or for checking climatological or territorial representativity of shorter term wind statistics.

In contrast to the wealth of meteorological data pouring from global measuring networks and data assimilation procedures, regionally and temporally representative flow data for environmental applications are, for numerous reasons, in some countries very scarce. In practice, this long standing input-data shortage often led to the situation where even the shortest term measurements of field data were used without their representativity being questioned.

Literature review

Efforts for mapping wind conditions have long history. Vector mean wind direction and speed maps based on 1880–1915 simple surface wind (Wild) observations were analysed for the territory of the former Austrian-Hungarian Monarchy by *Defant* (1920). Global distributions of seasonal standard vector deviations of 700 hPa winds were presented by *Meteorological Office* (1950). *Machta* (1979) presented global analyses of seasonal vector mean wind direction, speed and vector standard deviation at 850 and 700 hPa and presented a graphical methodology to estimate typical concentrations, depositions and flux rates for transport within 100–1000 km-s. *Peixoto and Oort* (1992) and *Landsberg* (1986) also presented global scale wind climatologies. *Jensen* (1984) produced an 850 hPa wind speed map from radiosonde data for a narrow band of the Western Europe coastline for wind energy applications.

Downward interpolation from geostrophic winds to predict surface winds was the original method adopted by the Riso group in producing the Danish Wind Atlas but was dropped in favor of a method based on surface wind data for their European Wind Atlas (*Taylor*, 1997). WAsP (Wind Atlas Analysis and Application Programme; *Troen et al.*, 1988) corrects wind data measured at a specific point and transforms these data into a data set describing the wind climate of a region, called a wind atlas. It utilizes such data sets to estimate the wind conditions at any particular site and height in the region, in principal by applying the same routines.

Szepesi and *Fekete* (1993) and *Szepesi et al.* (1995) reported preparation of regionally representative surface wind maps for Hungary. For the analysis of these maps all available surface wind data series (more than 200 between 1881–1980) and upper air ascents (22 long series between 1929–1989) in Hungary were considered.

Summarizing, it is concluded that while world climate centers are easily able to produce global scale assessments for hemispheric grid points based on stored data from the principal sounding stations, for regional scale assessments there is no substitute for conventional analysis of cleaned long-term data series obtained from national archives (e.g., KLIMA 90).

2. *Representativity of flow statistics*

One of the most important aspects of pollution transport is representativity of ensemble of flows at the height of the plume. To be representative the ensemble of flows has to meet territorial, temporal and height criteria.

2.1 Regional representativity

Locally measured winds are often biased and not representative of the height of plume. The smoothed out flow pattern is regionally more representative and thus reveals many inconsistencies in siting, measuring and data analysis.

2.2 Temporal representativity

Due to climatic variability of wind conditions temporal representativity is ensured mostly by using long-term (~30-year) historical data. If such data series are not available, short-term (1–5 year) series could be used if they meet temporal representativity criteria i.e., both the short-term and the long-term (10–30 year) periods have similar macrosynoptic weather type (MSWT) frequency distributions.

As far as the presently analysed 1980–81 period is concerned the MSWT analysis shows that cyclonic weather types were 5% more frequent than in the long-term normal period. By using such MSWT analysis and even forecasting, the effect of possible climate modification might also be accounted for by applying similar criteria.

3. *Assessment of results*

The goal of this research is to establish eventually a design 850 hPa wind concept to promote harmonization of meteorological data preprocessing for dispersion estimates. As a first step upper air wind analyses have to be completed taking into account the following principles: (a) They should be based on all the long-term (10–30 yr) 00 and 12 UTC 850 hPa wind data available, (b) Where only shorter-term (1–5 yr) data series are available they have should be checked against long-term macrosynoptic statistics before inclusion, and (c) Territorial analysis (smoothing) should be used to ensure representative flow data for the region concerned.

The preliminary wind atlas (see *Figs. 1, 2* and *3*) was prepared using 2 year (1980–1981) 850 hPa measured upper air wind direction and speed statistics. This rather old period was selected because the upper air sounding network in Europe was the densest at that time and nothing limited regional scale analyses. The evaluation and analysis was carried out manually in the late 80-ies when the idea of constructing design wind maps emerged. A map showing the station locations (see Fig. 1) allows the reader to make a better judgement of the spatial representativity of data. Sample gradient wind maps were presented at the Ostende Meeting and encountered considerable interest (*Szepesi* and *Fekete*, 1996).

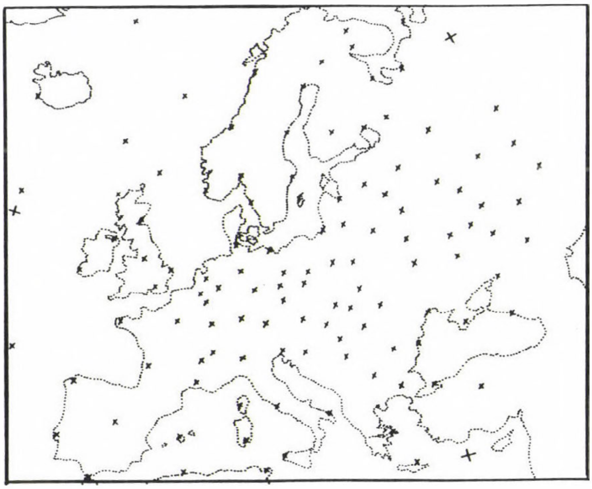

Fig. 1. Upper air station network.

To make these preliminary maps climatologically more representative, the following countries have already provided long-term wind statistics (see *Tables 1–4*). (a) For the UK 12 year (1976–1987) 13 station 850 and 700 hPa 00 and 12 UTC statistics (*D.J. Thomson*), (b) for Germany 30 year (1961–1990) 850 and 700 hPa 00 and 12 UTC statistics (*E. Dittmann*), (c) for Hungary 28 year (1962–1989) 2 sounding and 10 pilot balloon station 850 and 700 hPa 00 and 12 UTC statistics (*D.J. Szepesi*), (d) for Switzerland 32 year (1959–1990) 925, 850 and 700 hPa 00 and 12 UTC wind pattern (*A. Aschwanden*), (e) for Poland 20 year (1971–1990) 4 station statistics (*Z. Litynska*), (f) for Finland 31 year (1965–1995) 3 station statistics (*M. Lahti*). In addition a number of other researchers have indicated their intention to participate in this project.

The comparison of preliminary data from the wind atlas (Figs. 2 and 3) and longer-term data which are yet not to be part of the atlas are shown in *Figs. 4* and *5* and *Tables 5* and *6*. It can be seen in the RMS diagrams (Fig. 4 and 5) that all the data are near to the 45° trendline. It suggests that the data are in strong relationship.

The RMS of data is higher in case of wind directions (Table 5) than in windspeed (Table 6). This means that the windspeed is less changeable than the wind direction. There are only few data, which are relative far from the trendline. This is shown also by the numerical analysis.

The RMS of wind direction data is 3.54 for the long, and 3.98 for the short series. The covariance coefficient between the long, and short series is 11.23, the correlation coefficient is 0.80. The relative high value of the correlation coefficient shows a close relationship.

The RMS of windspeed data is much smaller, the points are very close to the trendline. This is shown by the numerical analysis, too. The RMS of the long series is 2.11, and 1.97 for the short series. The covariance coefficient is 3.62; the correlation coefficient is 0.88, which clearly shows a very close relationship between the two series.

Summarizing we conclude, that the interval of 1980–1981 represents for Europe macrosynoptically quite well the upper air conditions of a long (20 to 35 years) period.

The main results are as follows:

(a) Analyses based on just 2 years of measured upper air wind data show clearly the new method's ability for regional scale analysis of characteristic wind patterns. Even with 2 year long records of wind data, the statistics obtained already approximate long-term patterns provided the years are climatologically normal.

(b) Short series wind speed averages approximate long series averages even better than the wind direction data do.

(c) Sharp chanelling caused by the mountain ranges of e.g., the Alps and the Jura was detected in the short as well as in the long-term wind direction statistics. Inside the dominant wind sectors, however, considerable variation in directional frequencies occurred due to the different MSWT characteristics of the respective periods.

(d) Details of the finer scale structure of e.g. the E and NE wind speeds over level terrain west of Urals should not be considered artifacts. They are mostly due to regional differences in climate. It is expected, however, that 30-year data statistics will show less fine structure in the upper air windspeed pattern.

(e) For the preparation of wind statistics, the application of the conventional 16 meteorological sector distribution is recommended instead of the often used 12 sector standard.

(f) Based on these preliminary results, which were obtained after collecting 30 year wind statistics, finer areal as well as temporal analyses are justified. The final analysis might show new features, which are not apparent from this preliminary assessment.

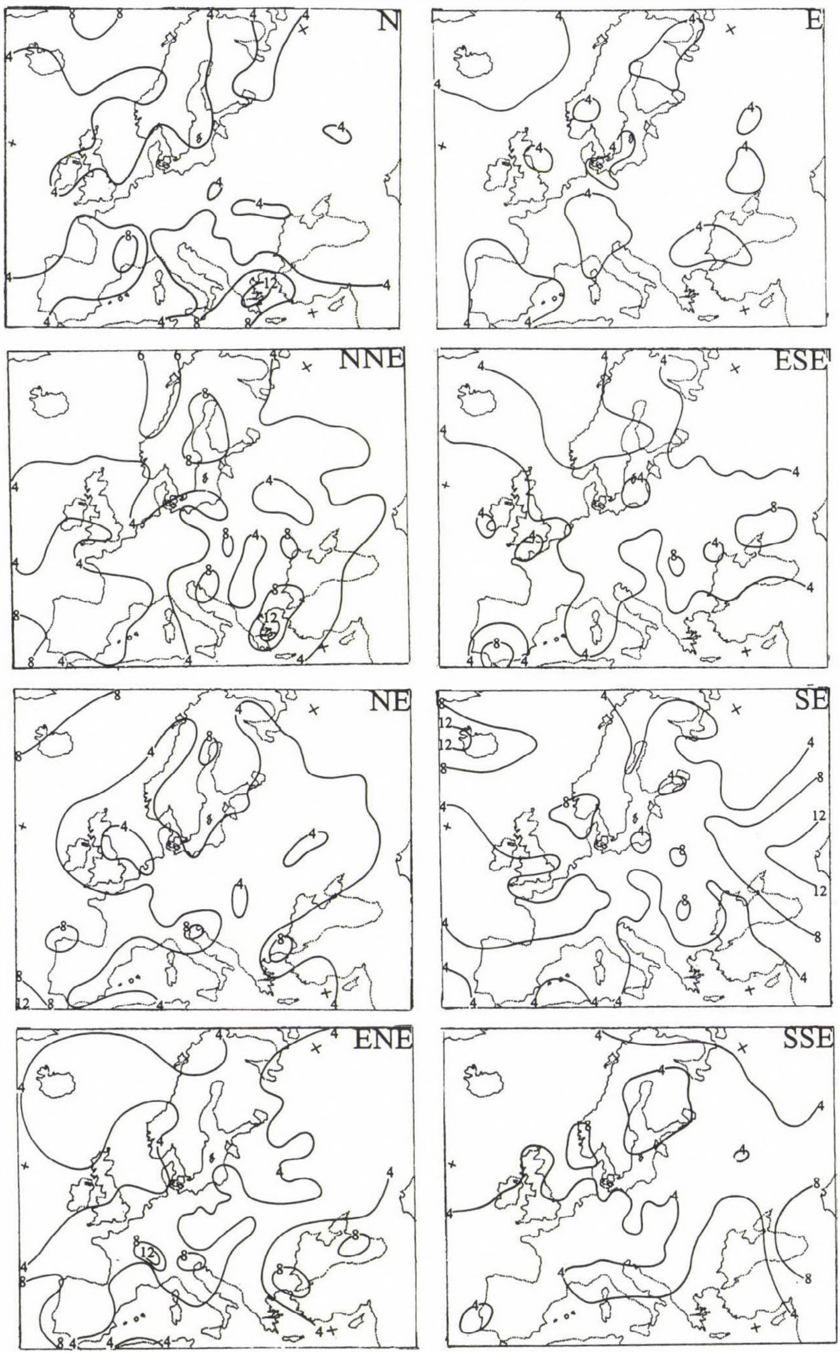

Fig. 2a. Realtive frequency of 850 hPa wind direction (N–SSE) in percent (1980–1981).

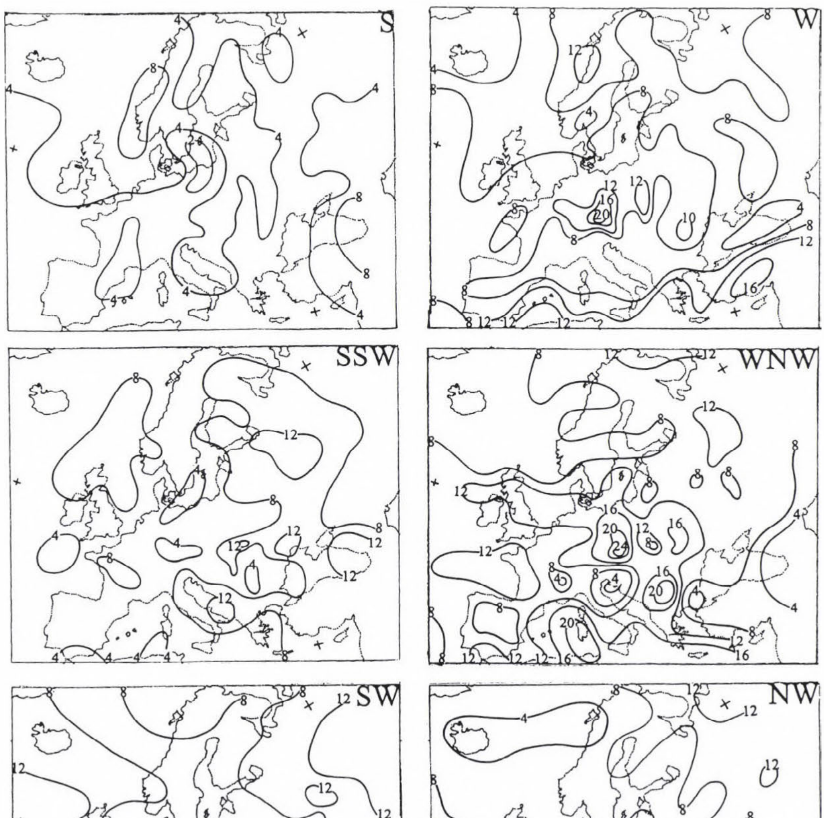

Fig. 2b. Relative frequency of 850 hPa wind direction (S–NNW) in percent (1980–1981).

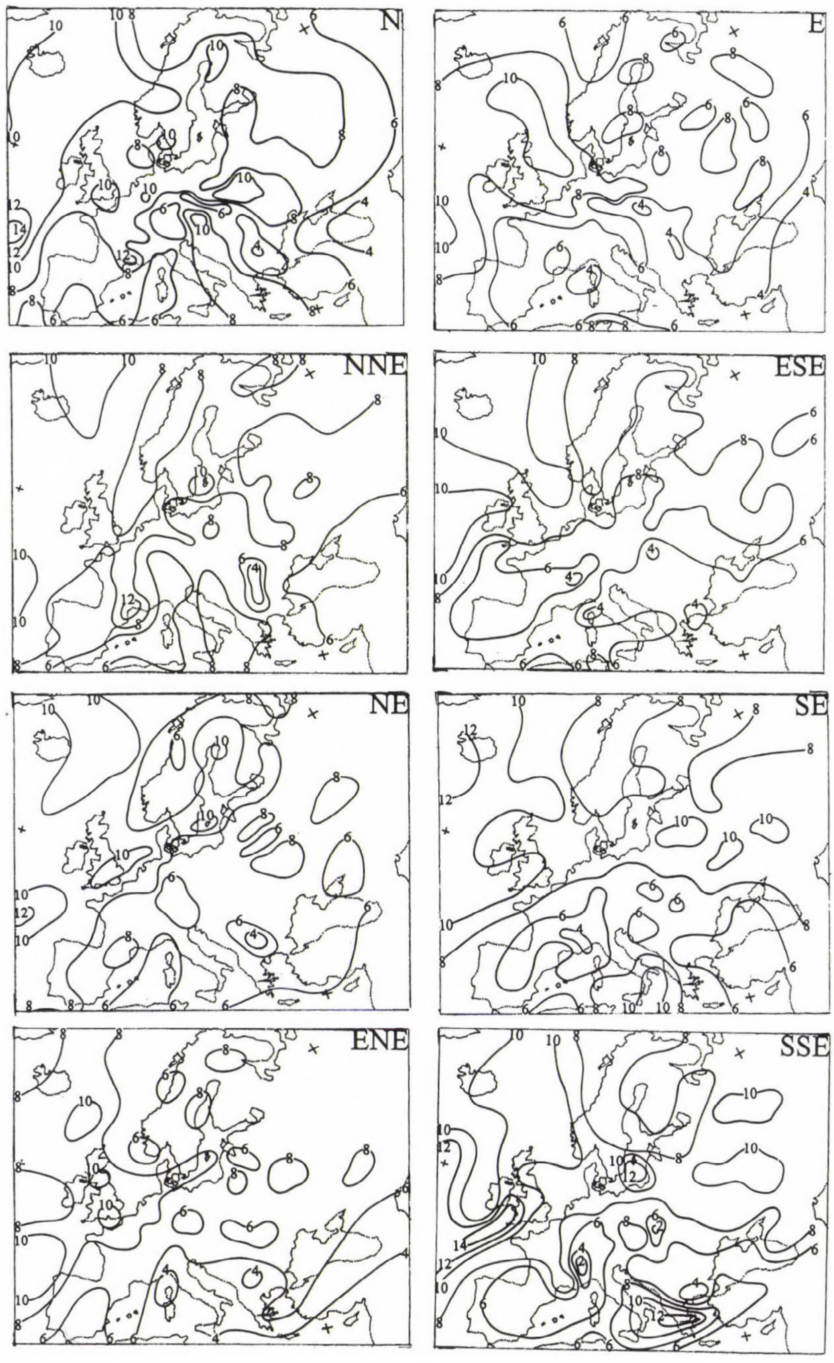

Fig. 3a. Yearly mean 850 hPa wind speed for directions N–SSE, in m sec^{-1}, (1980–1981).

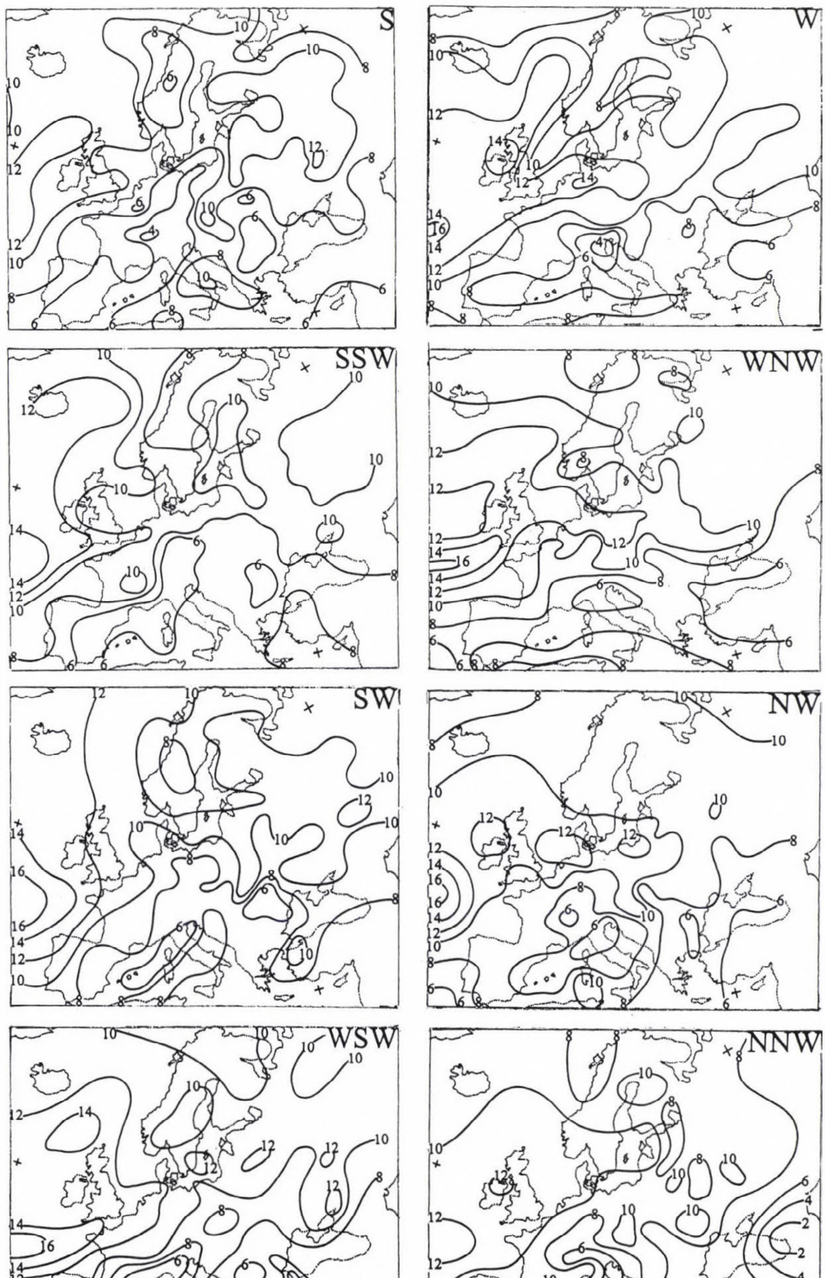

Fig. 3b. Yearly mean 850 hPa wind speed for directions S-NNW, in m sec^{-1}, (1980–1981).

Table 1. Relative frequency of wind direction at 850 hPa, in percent (available long series data)

Station	N	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW
1	6.6	5.0	4.2	3.1	3.1	2.7	3.3	4.0	5.3	9.1	11.3	10.2	9.4	7.6	7.9	7.2
2	5.3	4.6	3.7	3.1	3.1	3.0	3.4	4.4	6.6	8.0	10.1	10.1	9.4	8.3	9.8	7.1
3	6.3	4.7	3.8	3.0	3.2	3.0	3.4	4.0	6.4	8.0	10.0	9.8	10.0	9.2	8.3	6.9
4	5.7	3.9	3.2	2.5	2.3	3.1	4.3	5.4	8.3	8.6	10.1	10.6	10.5	7.8	7.3	6.4
5	4.1	3.3	3.1	2.8	3.0	3.1	3.6	4.2	7.3	10.2	12.5	12.9	11.9	7.2	6.0	4.8
6	5.1	3.9	3.3	2.7	3.4	3.2	3.6	3.9	4.8	6.8	10.0	12.1	13.5	9.7	8.0	6 0
7	4.5	3.1	3.1	3.3	3.9	2.6	3.3	4.1	5.5	8.3	9.7	10.1	12.5	10.7	9.0	6.3
8	3.8	3.0	3.2	3.4	3.5	3.5	4.0	4.3	5.6	7.6	10.0	11.4	13.8	10.0	7.8	5.1
9	5.2	3.7	3.7	3.6	3.6	3.2	3.3	3.6	5.5	7.3	9.8	11.3	12.8	9.4	7.9	6.1
10	4.7	3.7	3.7	3.4	2.6	2.7	3.2	3.7	5.7	8.0	10.8	11.9	11.7	8.8	8.3	7.1
11	4.1	4.8	5.1	4.5	3.7	3.1	2.8	2.8	5.1	7.4	9.9	11.4	13.5	9.8	7.3	4.7
12	4.4	4.3	4.5	4.3	4.2	3.4	3.4	3.4	4.9	7.2	9.6	10.9	12.8	9.6	7.7	5.4
13	4.7	4.5	4.8	4.5	4.3	3.1	3.0	3.0	3.7	6.3	9.8	12.2	13.4	9.3	7.7	5.7
14	5.4	3.9	3.8	3.6	3.5	2.9	3.3	3.7	5.5	7.2	9.8	11.2	11.9	9.9	8.3	6.1
15	4.7	3.0	2.8	2.7	3.3	3.3	3.8	4.2	6.2	7.3	9.7	11.3	13.4	10.1	8.3	5.9
16	5.2	3.5	3.1	2.7	3.2	3.1	3.8	4.4	6.6	8.5	10.3	10.7	11 0	9.4	8.2	6.3
17	3.2	2.9	9.6	14.6	3.8	1.3	1.2	1.1	2.1	5.4	22.6	17.2	7.8	3.1	2.2	1.9
18	2.7	1.9	2.1	2.7	5.4	6.2	5.1	3.4	3.3	3.3	4.4	7.2	20.4	16.3	10.9	4.7
19	4.7	3.8	3.7	2.1	3.1	2.8	4.1	4.1	5.5	5.6	9.5	10.8	16.9	10.7	8.2	4.4
20	4.7	3.1	3.2	2.4	3.5	3.0	4.9	4.4	5.5	4.5	7.8	10.0	15.7	12.8	9.6	4.9
21	4.9	3.1	3.1	2.4	3.3	3.2	5.0	4.4	6.1	5.6	8.1	9.2	14.7	11.8	9.7	5.4
22	5.2	3.0	3.3	2.4	3.0	3.2	4.2	3.7	5.6	5.2	6.9	7.1	13.9	14.3	13.0	6 0
23	5.8	4.2	4.2	3.4	3.2	3.0	3.8	5.3	6.1	7.3	8.7	7.3	6.3	9.7	12.3	9.4
24	6.1	5.6	4.0	2.8	2.8	3.1	4.8	6.2	6.1	6.5	7.9	8.7	7.6	7.6	11.5	8.7

Table 2. Mean wind speed at 850 hPa, in m s^{-1} (available long series data)

Station	N	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW
1	9.6	8.7	8.4	7.7	6.9	7.3	7.8	8.2	9.0	10.5	11.3	10.5	9.4	9.2	9.3	9.8
2	9.2	8.5	7.8	7.8	7.4	7.8	8.1	9.0	10.0	11.0	11.2	11.1	10.4	10.4	10.5	9.7
3	9.0	8.3	8.0	7.6	7.3	7.7	8.1	8.8	10.1	11.0	11.5	11.3	10.3	10.3	9.9	9.7
17	2.6	4.1	6.1	7.4	4.0	3.0	2.9	2.9	3.8	6.6	10.1	9.0	6.6	4.7	4.1	4.0
18	7.7	7.2	6.9	7.3	7.9	5.6	5.7	6.0	6.3	7.4	8.2	9.0	13.4	12.4	11.1	9.7
19	8.8	8.7	8.2	7.7	7.7	7.9	8.3	8.2	8.4	8.8	10.2	12.0	13.1	12.6	10.2	8.6
20	8.0	7.3	6.9	6.8	7.4	8.2	8.1	8.7	8.2	7.6	8.4	10.5	11.8	12.7	11.0	9.0
21	7.9	7.1	6.6	6.4	6.7	8.1	8.7	8.1	8.1	8.3	9.2	10.1	10.4	11.7	10.5	8.7
22	7.8	6.8	7.0	6.8	6.9	7.4	7.3	8.1	8.8	9.2	8.1	8.9	11.0	13.1	12.4	9.4
23	7.3	5.7	6.0	5.9	5.6	5.5	5.8	7.4	7.2	7.8	9.1	8.6	7.7	9.8	10.6	9.1
24	6.7	7.0	5.6	5.1	4.8	5.1	6.1	7.6	7.2	7.2	7.8	8.4	7.8	7.8	9.0	8.1

Table 3. Sources of data in Tables 1, 2 and 4

Station(s)	Source
1., 2. 3.	M. Lahti. Helsinki. Finland. 1996 COST 710 WG4
4. -16.	D. Thomson. Met.Office. 1996 COST 710 WG4
17.	Aschwanden A. et al.. 1996. Bereinigte Zeitreihen. Die Ergebnisse des Projekts KLIMA90. Band 1. Auswerungen. Klimatologie
18.	E. Dittman. 1996. DWD COST 710 WG4
19.-22.	Z. Lityinska. Legionowo. Poland. 1996 COST 710 WG4
23., 24.	D. Szepesi. 1996 COST 710 WG4

Table 4. Station numbers and WMO-codes with names of the upper air stations

Station number	WMO code	Country	Station name	φ-Coordinate	λ- Coordinate	Interval
1*	2836	Finland	Sodankyla	N 67°21'	E 26°39'	1961–1995
2*	2935	Finland	Jokioinen	N 60°48'	E 23°30'	1961–1995
3	2963	Finland	Jyvaskyla	N 62°23'	E 25°40'	1961–1995
4*	3005	UK	Lerwick**	N 60°06'	W 1°12'	1976–1993
5*	3026	UK	Stornowa**y	N 58°12'	W 6°18'	1976–1993
6*	3170	UK	Leuchars**	N 56°24'	W 2°54'	1976–1993
7	3213	UK	–**	N 54°18'	W 3°24'	1976–1984
8	3322	UK	Aughton**	N 53°36'	W 2°54'	1976–1987
9*	3496	UK	Hemsby**	N 52°42'	E 1°42'	1976–1987
10	3502	UK	Aberporth**	N 52°06'	W 4°36'	1976–1987
11	3693	UK	Shoeburyness**	N 51°36'	E 0°48'	1976–1987
12	3743	UK	Larkhill**	N 51°12'	W 1°48'	1976–1987
13*	3774	UK	Crawley**	N 51°06'	W 0°12'	1984–1992
14	3808	UK	Camborne**	N 50°12'	W 5°18'	1976–1987
15*	3920	UK	Hillsborough**	N 54°30'	W 6°06'	1976–1992
16*	3953	Ireland	Valentia**	N 51°54'	W 10°18'	1976–1987
17*	6610	Switzerland	Payern	N 46°48'	E 6°57'	1961–1990
18*	10866	Germany	München–Riem	N 48°07'	E 11°33'	1961–1990
19	12120	Poland	Leba	N 54°45'	E 17°31'	1971–1990
20*	12330	Poland	Poznan	N 52°24'	E 16°49'	1971–1990
21*	12374	Poland	Legionowo	N 52°09'	E 20°57'	1971–1990
22*	12425	Poland	Wroclaw	N 51°06'	E 16°52'	1971–1990
23*	12843	Hungary	Budapest	N 47°21'	E 19°11'	1962–1989
24*	12982	Hungary	Szeged	N 46°15'	E 20°06'	1962–1989

Note: * Station which appears in the analysis of both series
** Originally 12 directions transformed to 16 directions

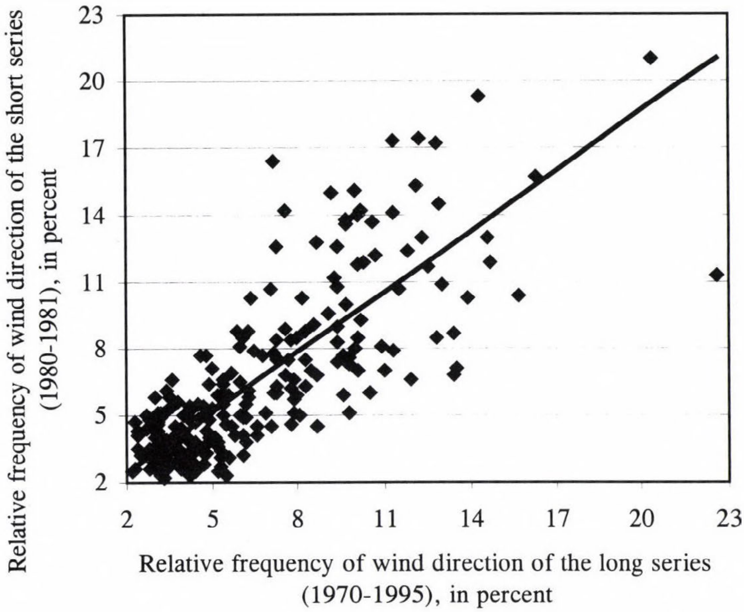

Fig. 4. Scatter diagram of wind direction.

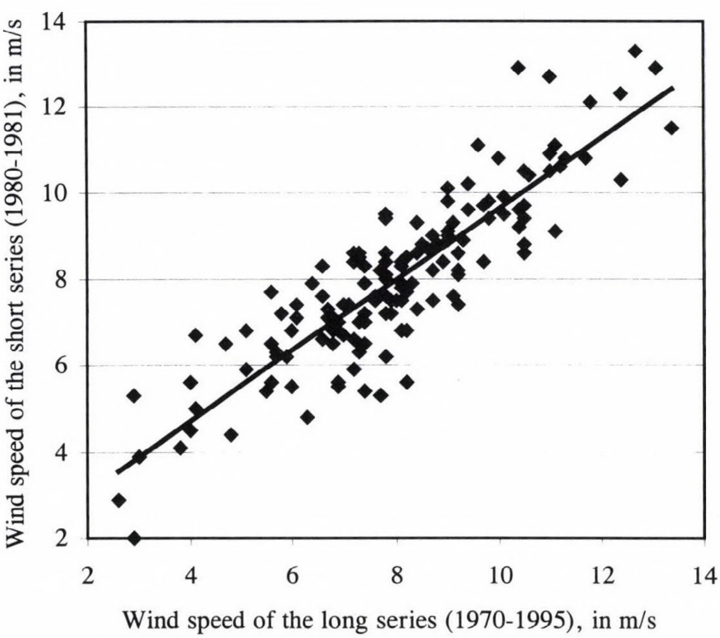

Fig. 5. Scatter diagram of wind speed.

Table 5. Statistics of wind direction

Statistics	**Long series**		**Short series**
RMS of the frequency of wind direction	3.54		3.98
Correlation		0.80	
Average	6.08		6.12
Covariance		11.23	

Table 6. Statistics of wind speed

Statistics	**Long series**		**Short series**
RMS of wind speed	2.11		1.97
Correlation		0.88	
Average	8.01		8.00
Covariance		3.62	

4. Follow-up work

To finalize this work it is necessary to involve the rest of the European countries in providing 10–30-year upper wind statistics. This program can be realized only after appearance of this paper and has to be based on the joint effort of more contributors. Recent experience in collecting data shows that 30-year series actually were cleaned and processed only after personal contact and request. This also reveals some of the benefits of such conventional approaches.

It is conceivable that authors in their analyses will be able to cope with orographic effects, in say, the Alps region, better than global climatological analyses do. This kind of regional scale wind atlas could be used as a starting point for further surface wind surveys. After completion of the geostrophic wind atlas, countries might construct regionally representative design surface wind charts for their territory (see *Szepesi et al.*, 1995 for Hungary). For environmental applications, the surface wind statistics at any site can be determined from these wind maps using a model correction (e.g., according to WAsP) for the effects of terrain, surface roughness and obstacles.

Acknowledgement—Critical remarks and careful editing of this paper by *David J. Thomson* of UK Meteorological Office, the former chairman of COST 710 project of the EC is gratefully appreciated.

References

Defant, A., 1920: *Windverhältnisse in Gebiete der ehemaligen Österr.-Ungar. Monarchie*. Anhang zum Jahrbuch der ZMG, Wien, *Neue Folge 57.*

Jensen, N.O, 1984: *Extrapolation of Mean Wind Statistics with Special Regards to Wind Energy Applications*. WCP-86, World Meteorological Organization, Geneva.

Landsberg, H.E., 1986: *World Surveys of Climatology*. Elsevier, Amsterdam.

Machta, L., 1979: *A Handbook to Estimate Climatological Concentration, Deposition and Horizontal Fluxes of Pollutants on a Regional Scale*. MARC Tech. Report 11, UNEP, Chelsea College, Univ. of London.

Meteorological Office, 1950: *Upper Winds over the World*. Geophysical Memoirs 85, London.

Peixoto, J.P. and *Oort, A.H.*, 1992: *Physics of Climate*. American Inst. of Physics.

Szepesi, D.J. and *Fekete, K.E*, 1993: *Design Wind Maps for Hungary* (in Hungarian), *Meteorológiai Tudományos Napok,* MTA Budapest.

Szepesi, D.J. and *Fekete, K.E.*, 1996: Inquiry on meteorological data pre-processing for complex terrain dispersion modelling: A european reference Benchmark exercise. *4th Workshop on Harmonization within Atm. Disp. Modelling for Regulatory Purposes*. May 1996, Ostende. Preprint.

Szepesi, D.J., Fekete, K.E. and *Gyenes, L.*, 1995: Regulatory models for environmental impact assessment *in* Hungary. *Int. J. Environment and Pollution* 5, 497-507.

Taylor, P., 1997: Personal communication.

Troen, I., Mortensen, N.G. and *Petersen E.L.*, 1988: *Wind Atlas Analysis and Application Program.* National *Laboratory* Riso, Denmark.

BOOK REVIEWS

Wilfried Schröder (editor): **Long and Short Term Variability in Sun's History and Global Change**. Science Edition, D-28777 Bremen-Roennebeck, 2000, 63 pages. Price: $20.

This book is the printed version of the 26 lectures presented during the IUGG (International Union of Geodesy and Geophysics) last General Assembly Meeting (1999, Birmingham, UK) by the members of the Interdivisional Commission on History (IDCH) of IAGA. Like such collections, the papers cover very different topics. However, most of them deal with the connection of solar phaenomenae and the Earth's climate. The investigation of this connection has long history. The recent results suggest a connection of about 20 years cycle, that is the magnetism might have some role. It is also interesting, that the terrestrial and solar atmospheres have some common features.

I think this book is very useful for those scientists and students who are interested in history of solar physics, geophysics and meteorology.

G. Major

Wilfried Schröder (editor): **Geschichte und Philosophie der Geophysik (History and Philosophy of Geophysics)** Science Edition, D-28777 Bremen-Roennebeck, 2000, 219 pages. Price: $30.

The material found in this book has already been published in *Beträge zur Geschichte der Geophysik und Kosmische Physik,* Band 2/2000 and at the same time in *IAGA IDC History Newsletter* No. 42. Most of it is written in German, the rest in English.

The title suggests that this is a complex study of every aspects of the geophysics, but the reader founds that it is wider, it relates to all geosciences. For example on page 40 the list contains the scientific journals founded before 1896 (the first year is 1665). From the 24 items exactly the half belongs to the discipline nowadays called meteorology. It has to be noted that IDŐJÁRÁS was founded in 1897 and since then it has been published continuously.

After the introduction and preface, the next 100 pages have bee written jointly by *Wilfried Schröder* and *Herbert Hörz*. Their work is a systematic study of all scientific aspects of geophysics/geosciences. The main value of this part is the opinion and evaluation of the authors expressed on the history, the methods and data of geophysics. The last 100 pages contain 9 papers from different authors mostly on special partial historical events. There are two exceptions: the first one deals with the scientific and practical use of history of geophysics in general, the last one is "an approximately complete list" of geophysical teaching books appeared between 1912 and 1998, it contains nearly 400 items.

It is suggested to all geoscientists stop for a small time in their rush and read this book and think on our loved disciplines.

G. Major

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Published by the Hungarian Meteorological Service

Budapest, Hungary

INDEX: 26 361

HU ISSN 0324-6329