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"Wind Erosion Climatic Erosivity"

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WIND EROSION CLIMATIC EROSIVITY

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Abstract. A physically based wind-erosion climatic factor has been derived:

$$CE = \rho \int \left[u^2 - (u_T^2 + \gamma^1/\rho a^2)\right]^{3/2} f(u)du$$

where ρ is the air density, a is a constant made up of other constants (von Karman, height of wind speed observation, roughness parameter), u is the horizontal wind speed, u_T is threshold wind speed, $f(u)$ is a wind speed probability density function, and γ^1 is the cohesive resistance caused by water on the soil particles. Cohesive resistance is proportional to the square of water content relative to water content at -1500 J kg^{-1}. Relative water content is approximated from the Budyko dryness ratio and the Thornthwaite PE index with similar results. CE is calculable from wind speed and other generally available meteorological data, and is usable in the wind erosion equation without some of the limitations of a previously used wind erosion climatic factor.

1. Introduction

Wind erosion climatic erosivity is a measure of the climatic tendency to produce conditions conducive to wind erosion. Wind erosion occurs when the shear stress exerted on the surface by the wind exceeds the ability of the surface materials to resist detachment and transport. Strong winds erode, and dryness increases the susceptibility of the surface to erosion.

The aridity of an environment is often evaluated by the Budyko dryness ratio (Budyko, 1958; Hare, 1983). The dryness ratio at a given site indicates the number of times the net radiative energy could evaporate the mean annual precipitation. Semi-arid zones where wind erosion is likely to be a serious problem have a dryness ratio between 2 and 7 (Hare, 1983). Areas with dryness ratios larger than 7 are in the desert and desert margin zones. Most of the Great Plains of the USA has dryness ratios between 2 and 5. The Sahara Desert in North Africa has a maximum dryness ratio as high as 200 (Henning and Flohn, 1977).

Chepil *et al.*, (1962) proposed a climatic factor to estimate average annual soil loss by wind for a range of climatic conditions. This factor, an index of wind erosion, is a function of soil moisture and average wind speed. The wind speed term was based

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on the rate of soil movement being proportional to the cube of average wind speed (Bagnold, 1943; Chepil, 1945; Zingg, 1953). The soil moisture term was developed on the basis that soil erodibility varied inversely with the square of water content in the upper few millimeters of soil which was assumed to vary as the Thornthwaite effective precipitation index (Chepil, 1956).

The climatic factor as proposed by Chepil (1962) was one of the five independent variables of the wind erosion equation which has been used widely during the past 20 years (Woodruff and Siddoway, 1965). Other variables of the wind erosion equation are identified in a companion paper (Skidmore, 1986, 'Wind Erosion Control', in this issue).

This research develops a method to characterise the climate's tendency to cause wind erosion based on the mechanics of the wind erosion process. This procedure is usable as a climatic factor in the wind erosion equation (Woodruff and Siddoway, 1965; Skidmore and Woodruff, 1968) for long and short term and event soil loss estimates. It can be used at various levels of sophistication and availability of climatic data and it provides a framework for research to better understand wind erosion variables.

2. Model

In the first part of this section (Equations 1 through 11), I review wind erosion climatic indices and present some fundamentals of wind erosion process. In the second part, I derive some new relationships aimed at accomplishing the objective of this paper.

The climatic factor as proposed by Chepil (1962) was expressed as:

$$C = 386\frac{\bar{u}^3}{(PE)^2} \tag{1}$$

where u is the mean wind speed and PE is the Thornthwaite (1931) index. A summary of notation is given in Annex. The term 386 indexes the factors to the conditions at Garden City, Kansas. Thornthwaite's index to evaluate precipitation effectiveness expressed the P/E ratio to temperature and precipitation as:

$$P/E = 0.316\left(\frac{P}{1.8T + 22}\right)^{10/9} \tag{2}$$

where P is the mean monthly precipitation, in mm; E is the monthly evaporation, in mm; and T is the temperature, in °C. Monthly values were added to obtain an annual value and multiplied by 10 to avoid fractions to give:

$$PE \text{ index} = 3.16\sum_{i=1}^{12}\left(\frac{P_i}{1.8T_i + 22}\right)^{10/9} \tag{3}$$

Equation (3) was used in Equation (1) to determine wind erosion climatic factors for many locations in the U.S. (Chepil *et al.*, 1962; Lyles, 1983; Skidmore and Woodruff, 1968).

As the PE index approaches zero when precipitation approaches zero, as in arid regions, the climatic factor in Equation (1) approaches infinity. This high sensitivity to low precipitation illustrates why users were expressing concern when the index was used in areas more arid than those for which the index was developed and extremely high climatic factors were noted. In application, an upper limit is established by restricting minimum monthly precipitation to 13 mm (Lyles, 1983). Monthly climatic factors were also calculated using an annual PE index with monthly mean wind speed (Woodruff and Armbrust, 1968).

FAO (1979) solved the problem of the climatic factor approaching large values in arid conditions differently. They modified the Chepil *et al.*, (1962) index to

$$C^1 = 1/100\sum_{i=1}^{12}\bar{u}^3\left(\frac{ETP - P}{ETP}\right)d \tag{4}$$

where $\bar{u}$ is the mean monthly wind speed at 2 m height, ETP is the potential evapotranspiration, P is the precipitation and d is the total number of days in the month. In this case, as precipitation approaches zero, wind speed dominates the climatic factor. Conversely, as precipitation approaches ETP, the climatic factor approaches zero. The influence of soil moisture in the FAO version is less than the squared influence of soil water demonstrated by Chepil (1956).

Particle-movement rate of dry, erodible particles is directly proportional to friction velocity cubed as expressed by Bagnold (1943):

$$q = Ku_*^3, \tag{5}$$

Kawamura (1951, cited by Lettau and Lettau, 1978):

$$q = K(u_* - u_{*T})(u_* + u_{*T})^2, \tag{6}$$

or Lettau and Lettau (1978):

$$q = K(u_* - u_{*T})u_*^2, \tag{7}$$

where q is the mass flow rate, K is a proportionality constant, u_* and u_{*T} are the friction velocity and threshold friction velocity, respectively. Equation (7) most closely fits the transport of sand as measured in wind tunnels (Lettau and Lettau, 1978).

Friction velocity is defined by

$$u_* = (\tau/\rho)^{1/2} \tag{8}$$

where τ is the surface shear stress and ρ is the air density. Substitution of Equation (8) into Equation (5) gives

$$q = K(\tau/\rho)^{3/2}. \tag{9}$$

Then, to express the rate of erosion of damp material composed of all erodible particles, Chepil (1956) proposed

$$q = K\left[(\tau - \gamma)/\rho\right]^{3/2} \tag{10}$$

where γ is a threshold shear stress, which is a function of the water content at the soil surface. Chepil (1956) experimentally determined that

$$\gamma = 0.6(\theta/\theta_{15})^2 \tag{11}$$

where θ is the volume fraction of water in the soil and θ_{15} is the volume fraction of water in the same soil at -1500 J/kg potential. The ratio θ/θ_{15} was referred to as equivalent soil water content, ω.

Chepil (1956) measured erosion from four different soils with a wide textural range at various water contents and compared it to the erosion prediction by Equation (10) with a fitted exponent that varied around 3/2. The equation fit the data well.

To incorporate both the threshold shear stress as expressed by Equation (6) and (7) and the resistance due to the cohesion of adsorbed water as separate parameters, Equation (10) becomes

$$q = K\rho^{-3/2}(\tau - \tau_T - \gamma^1)^{3/2} \tag{12}$$

where τ_T is the threshold shear stress and γ^1 is the added resistance for soil moisture.

I used Equation (12) and Chepil's (1956) data in an effort to determine γ^1 as a function of ω and found γ^1 similar to γ and was approximated by

$$\gamma^1 = 0.5\omega^2 \tag{13}$$

I also assumed, as Chepil *et al.*, (1962) did, that on a long term basis equivalent surface water content was approximated by the ratio of precipitation to potential evaporation.

The surface shear stress is a function of wind speed, which can be approximated by the logarithmic law (Panofsky and Dutton, 1984)

$$u = k^{-1}(\tau/\rho)^{1/2}\ln(z/z_0) \tag{14}$$

where u is the wind speed at height z, k is the von Karman constant, and z_0 is a roughness parameter. Let

$$a = k/\ln(z/z_0) \tag{15}$$

which has the value 0.0774 for $k = 0.41$, $z = 10$ m and $z_0 = 0.05$ m. Standard observation height is 10 m, and $z_0 = 0.05$ m was the value the U.S. Department of Commerce used for reducing their data to an elevation of 10 m.

Then, from Equation (14) and (15)

$$\tau = \rho(au)^2 \tag{16}$$

and when substituted into Equation (12)

$$q = K\rho^{-3/2}\left[\rho a^2(u_i^2 - u_T^2) - \gamma^1\right]^{3/2} \tag{17}$$

where u_i and u_T are observed and threshold wind speeds, respectively. The threshold wind speed varies with the size and density of material, but generally for an erodible surface, threshold wind speed is about 6 m/s.

Equation (17) becomes

$$q = Ka^3\left[u^2 - (u_T^2 + \gamma^1/\rho a^2)\right]^{3/2} \tag{18}$$

The bracketed portion of Equation (18) raised to the 3/2 power and multiplied by ρ becomes wind power density (W m^{-2}) after overcoming the threshold shear stress of particles on the surface. Then, by multiplying Equation (18) by the wind speed probability density function and integrating over the range of wind speeds, we get an expression for wind-erosion climatic erosivity, CE, which is directly proportional to q

$$CE = \rho\int_R^\infty \left[u^2 - (u_T^2 + \gamma^1/\rho a^2)\right]^{3/2} f(u)du \tag{19}$$

where R is defined by

$$R = u_T^2 + \gamma^1/\rho a^2 \tag{20}$$

and where $f(u)$ is the wind speed probability density function. Equation (19) indicates the tendency of time-average values for meteorological elements to cause wind erosion. It accounts for the meteorological influence of both wind speed and wetness of the surface soil particles, as well as overcoming a threshold wind speed for surface particles.

The wind speed probability density function may be expressed as Weibull distribution (Justus *et al.*, 1976; Apt, 1976):

$$f(u) = (k/c)(u/c)^{k-1}\exp\left[-(u/c)^k\right] \tag{21}$$

where k and c are the shape and scale parameters, respectively. Parameter c has units of velocity, and k is dimensionless. Weibull parameters can be determined from wind speed distribution summaries and have been for many locations in the Great Plains (Hagen *et al.*, 1980).

The summation procedure for evaluating Equation (19) can be written

$$CE = \rho\sum_{u_{i+0.5}^2 > R}^{n} (u_{i+0.5}^2 - R)^{3/2}\left|F(u_{i+1}) - F(u_i)\right| \tag{22}$$

where $F(u_i)$ is the cumulative distribution function

$$F(u_i) = 1 - \exp\left[-(u_i/c)^k\right]. \tag{23}$$

Choose n large enough so that $F(u_{n+1}) \approx 1.0$. The notation $u_{i+0.5}$ refers to a windspeed midway between u_{i+1} and u_i.

Equation (19) with $f(u)$ defined by Equation (21) can be integrated when $k = 2$ to give

$$CE = 1.33\rho c^3\exp\left[-(R/c^2)\right] \tag{24}$$

where R is as defined by Equation (20). The author gratefully acknowledges help from Prof. Mohamed Hassan (University of Khartoum) to integrate Equation (19). Details of the integration are available from the author upon request.

3. Methods

The essence of this presentation is the model represented by Equation (19). Therefore, rather than give methods of experimentation, this section gives procedures for: (1) comparing the exact solution to the approximation; (2) estimating the wind speed distribution from the mean wind speed data; (3) showing the sensitivity to the surface dryness at various wind speeds; and (4) comparing the results with other wind-erosion climatic indices.

Equation (22) and (24) were evaluated for a range of values for R (30, 50, 70, and 90 $m^2\ s^{-2}$) with Weibull parameters k and c equal to 2.0 and 6.43 m s^{-1}, respectively. The wind speed interval for the summation of Equation (22) was 1 m s^{-1}.

Since wind speed is often reported as means only, a procedure was developed to estimate the Weibull parameters from the mean wind speed. Weibull parameters were selected from several Great Plains locations for analysis (Hagen *et al.*, 1980). The coefficients of determination for the fit of the Weibull distribution to the wind speed data were ≥ 0.98. Scale factors, c, ranged from 2 to 9 and the corresponding shape factors, k, varied from approximately 1.0 to 2.6 with a mean of 1.77. Equation (25) was used to find the relationship between the scale parameter and the mean wind speed

$$\bar{u} = c\Gamma(1 + 1/k) \tag{25}$$

where $\Gamma(1 + 1/k)$ is the gamma function.

Values for the Weibull shape parameter, k, was regressed on values for the scale parameter, c, to obtain the relation between c and k.

To demonstrate the sensitivity of climatic erosivity to surface dryness, Equation (22) was evaluated for 33 dryness ratios between 1 and 10 at mean wind speeds of 4, 5, 6, 7, and 8 m s^{-1}, where ρ and u_T were 1.2 kg m^{-3} and 6 m s^{-1}, respectively; c and k were evaluated for each mean wind speed by Equation (26) and (27), respectively.

$$c = 1.12\bar{u} \tag{26}$$

$$k = 0.52 + 0.23c \tag{27}$$

The summation index n was set at 25 with a Δu of 1 m s^{-1}, which was more than adequate to include the distribution range of wind speeds. Equation (13) was used to evaluate γ^1 where the equivalent soil water content was approximated by the inverse of the dryness ratio, D (Budyko, 1958; Hare, 1983).

$$D = R_n L^{-1} P^{-1} \tag{28}$$

where R_n is the net solar radiation and L and P are the latent heat of evaporation and precipitation, respectively.

Since the climatic factor of the wind erosion equation (Woodruff and Siddoway, 1965; Chepil *et al.*, 1962) is referenced to Garden City, Kansas, Equation (22) was

evaluated by month for the climate of Garden City. Wind speed distribution parameters were obtained from Hagen *et al.*, (1980), solar radiation at nearby Dodge City from Shaw (1982), and temperature and precipitation from NOAA (1982). Monthly net relationship was calculated from mean monthly solar radiation, R_s, from the relationship between solar and net radiation given by Rosenberg *et al.*, (1983).

$$R_n = 0.69R_s - 0.34d \tag{29}$$

where d is number of days in the month.

The Thornthwaite (1931) index (Equation (2)) was also used in addition to dryness ratio for approximating equivalent soil water content.

Wind-erosion climatic indices were calculated by using Equation (1), Chepil *et al.*, (1962); Equation (4), FAO (1979); and Equation (22) for Garden City, Kansas. Indices were calculated based on mean annual temperature and radiation for a precipitation range between 10 and 80 mm per month. The least and greatest monthly precipitation at Garden City were 9 mm and 78 mm per month for December and June, respectively.

4. Results and discussion

The results show that the approximate solution, Equation (22), is essentially identical with the exact solution, Equation (24). The difference was always less than 0.4%. However, Equation (24) is the solution only when $k = 2$.

The ratio of scale parameter c to mean wind speed was found to be 1.0, 1.13 and 1.12 for smallest (1.0), largest (2.6) and mean (1.77) values of the shape parameters from the sample data, respectively, which agrees well with Johnson's (1978) results. He found that for most wind regimes $1.3 \leq k \leq 3.0$ and c calculated from Equation (25) gave $1.11\bar{u} \leq c \leq 1.3\bar{u}$. In both examples, c was, on the average, 12% larger than $\bar{u}$ and was relatively insensitive to variation in wind speed except at very low wind speeds.

The Weibull shape parameter regressed on scale parameter gave the result shown in Equation (27) with a coefficient of determination of 0.87. Thus, if only mean wind speed is known, reasonable estimates of Weibull distributions can be obtained from Equations (26) and (27).

Mass flow rate (from Belly's 1964 wind tunnel data) plotted against the argument of Equation (19) shows an excellent linear relationship in Figure 1, where the coefficient of determination is 0.997. The threshold wind speed at height of observation in the wind tunnel was 5.9 m s^{-1}. Since the material was dry, γ^1 was zero.

Values calculated by Equation (22) and compared to the Garden City reference are shown in Figure 2 as a function of dryness ratio for several wind speeds. As the dryness ratio increases, climatic erosivity increases, but progressively at a slower rate until the dryness ratio reaches approximately 10. After that, a further increase does not further increase the wind erosion hazard because of dryness of particles.

Dryness ratios calculated on a monthly basis for Garden City, Kansas, are shown in column 8 of Table 1. They vary from a high in December of 8.94 to a low in May of

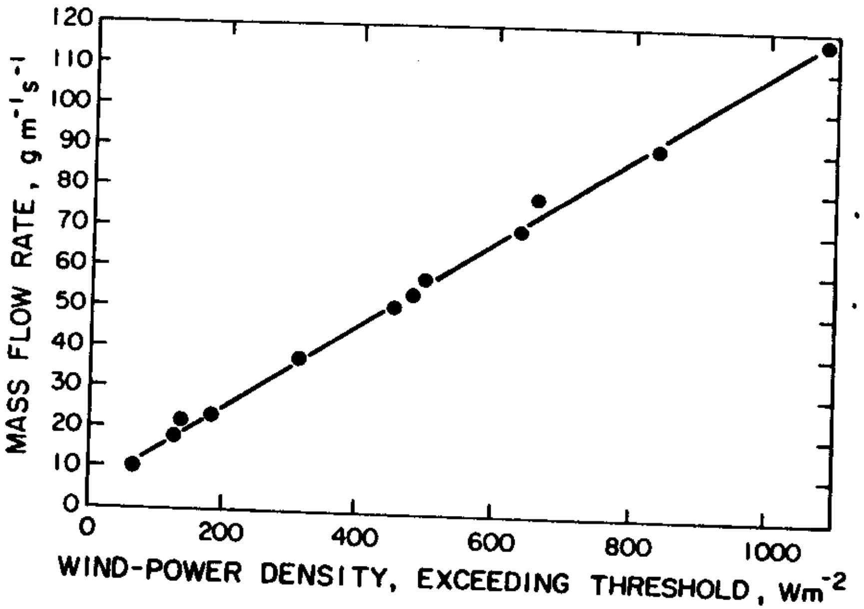

Figure 1. Wind tunnel erosion as influenced by calcuated erosivity. Data are from Belly (1964).

2.73, with an annual value of 3.90. December is the month with the least precipitation and solar radiation. The Thornthwaite P/E ratio, column 9, follows closely the same pattern as the dryness ratio. Comparisons of values in columns 10 through 13, Table 1, illustrate that for this data set the difference between using the dryness ratio and the P/E ratio in Equation (22) for calculating climatic erosivity was relatively small.

Henning and Flohn (1977) reason that because net radiation can be more clearly defined than potential evapotranspiration, ETP, for each spot on the surface of the Earth, the use of the net radiation is preferred to ETP as a climatological aridity index. Hare (1977, 1983) prefers the dryness ratio as an index of aridity and reported (1983, p. 121) that $\bar{u}^3(PE)^{-2}$ was a useful index of meteorological parameters for evaluating the intensity of wind erosion as a desert-forming process.

I assumed that the dryness ratio or its equivalent approximated the equivalent water content of the surface particles. This assumption is reasonable and should be sufficient for a climatological index. However, for a more detailed analysis or a flux equation, more research is needed to determine the relation of soil drying to wind-erodible dryness as a function of meteorological variables and soil hydraulic properties (Skidmore and Dahl, 1978).

Values in columns of Table 1 headed by C^{11} can be used to calculate soil loss based on the wind erosion equation (Woodruff and Siddoway, 1965; Skidmore and Woodruff,

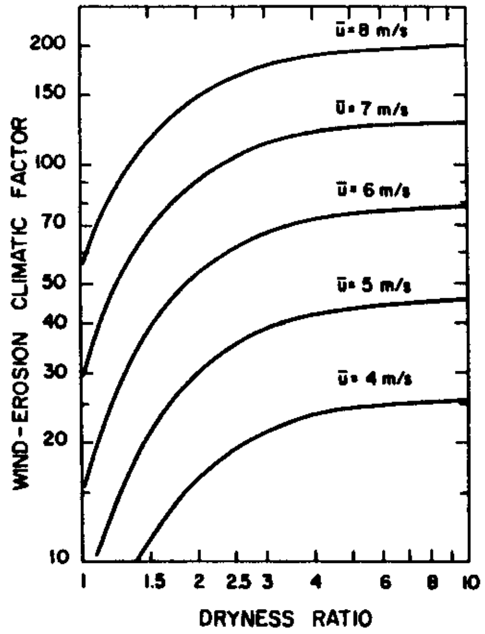

Figure 2. Climatic erosivity as influenced by dryness ratio and mean windspeed.

1968). When the erosion variables—equivalent vegetative cover, ridge roughness, soil erodibility—as well as climate are not constant throughout the year. These monthly climatic factors are especially useful.

The sensitivity of three wind erosion climatic indices to precipitation is illustrated in Figure 3. The FAO (1979) and my procedure are similar and vary much less with precipitation than does the index of Chepil *et al.*, (1962).

The assumption that precipitation < 13 mm does not decrease wind erosion potential is contrary to the behavior of the Chepil index as shown in Figure 3. A cap is placed on the index when it is showing the greatest sensitivity to precipitation. The index increases almost fivefold when precipitation decreases from 25 to 13 mm per month.

In Chepil's (1956) original research on the influence of soil moisture on erosion by wind, he found that the shear stress to initiate erosion was proportional to equivalent soil water content squared. He also demonstrated that mass flow rate was proportional to the difference between shear stress and cohesive resistance caused by water, all raised to the 3/2 power as expressed in Equation (10). However, instead of using the relationship of Equation (10) to develop a climatic index, he extrapolated the relationship to mean that erosion was inversely related to water content squared as expressed by Equation (1).

Table 1. Climatic information for Garden City, Kansas

Month	c	k	P	T	Solar	Net	$R_n/(LP)$	E/P	$EE^\dagger$	$EE^\ddagger$	C''¶	C''¶
	ms^{-1}		mm	°C	MJm^{-2}	MJm^{-2}			MJm^{-2}	MJm^{-2}	%	%
Jan	6.46	2.07	10	-1.8	331	218	8.82	6.58	448	440	66	67
Feb	7.11	1.99	12	1.3	374	248	8.36	7.14	678	673	100	102
Mar	7.96	2.03	34	5.2	543	364	4.33	2.88	1122	1041	166	159
Apr	8.23	2.16	38	12.0	664	448	4.77	3.68	1135	1097	168	167
May	7.82	2.04	74	17.6	738	499	2.73	2.20	938	862	139	131
Jun	8.48	2.23	78	23.6	817	554	2.87	2.56	1123	1084	166	165
Jul	7.25	2.48	68	26.7	834	564	3.36	3.25	476	471	71	72
Aug	7.32	2.69	61	25.4	769	520	3.45	3.55	438	442	65	67
Sep	7.56	2.61	42	20.2	620	418	4.03	4.57	551	564	82	86
Oct	7.08	2.60	32	13.6	494	330	4.17	4.83	410	423	61	64
Nov	6.16	1.84	19	5.1	358	237	5.05	5.65	437	443	65	67
Dec	6.27	2.17	9	0.3	304	199	8.94	8.62	343	342	51	52
Ann	7.23	2.29	477	12.4	6846	4599	3.90	3.58	8100	7882	100	100

† Calculated by Equation (22) using the dryness ratio for estimating the water content.

‡ Calculated by Equation (22) using the Thornthwaite PE index for estimating the water content.

¶ Monthly climatic factor based on erosive wind energy using dryness ratio and Thornthwaite PE index, respectively.

Data from several sources (Shaw, 1982; NOAA, 1982, Hagen *et al.*, 1980).

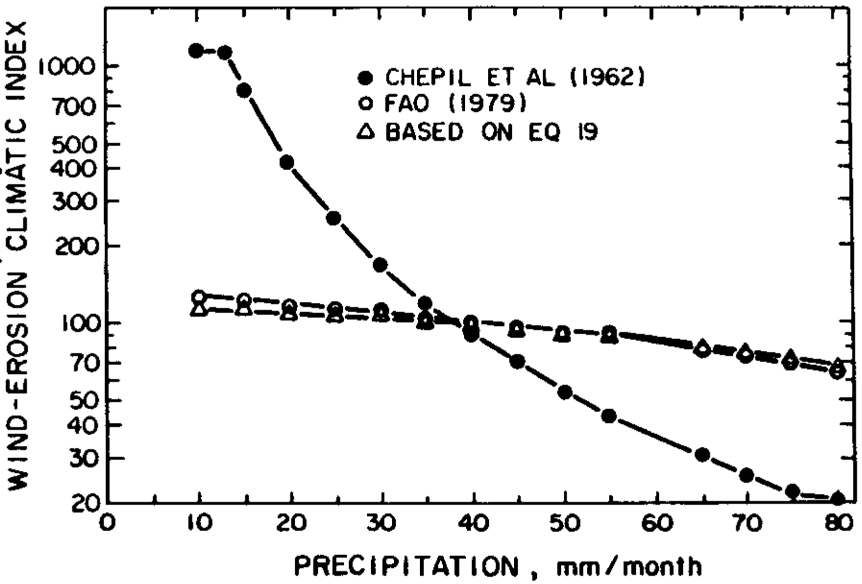

Figure 3. Comparison of wind erosion climatic indices as influenced by precipitation.

The wind erosion climatic index given by Equation (19) is based on the mechanics of wind erosion and accounts for the influence of surface soil water, wind speed, and wind speed probability distribution. It can be calculated from readily available meteorological data, is usable in the wind erosion equation, is adaptable to erosion events, and is a framework for additional research to better understand wind erosion.

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Annex Notations and Units

Symbol	Explanation	Units
a	Combination of constants: $k/\ln(z/z_0)$, for $k = 0.41$, $z = 10$ m, and $z_0 = 0.05$ m, $a = 0.0774$	1
C	Wind erosion climatic factor (Chepil *et al.*)	%
C^1	Wind erosion climatic factor (FAO)	%
C^{11}	Wind erosion climatic factor (Erosive wind energy)	%
c	Weibull distribution scale parameter	m s^{-1}
CE	Wind-erosion climatic erosivity	W m^{-2}
D	Dryness ratio or aridity index	J J^{-1}
d	Number of days in the month	days
E	Evaporation	mm
EE	Erosive wind energy, climatic erosivity × duration	J m^{-2}
ETP	Potential evapotranspiration	mm
K	Proportionality constant	1
k	von Karman constant, 0.41	1
k	Weibull distribution shape parameter	1
L	Latent heat of evaporation	J kg^{-1}
n	Upper limit of an index	–
P	Precipitation	mm
PE	Thornthwaite precipitation evaporation index	–
q	Mass flow rate of eroding material	g m^{-1} s^{-1}
R	Sum of erosion resistive elements (threshold & moisture)	m^2 s^{-2}
R_n	Net solar radiation energy	J m^{-2}
R_s	Incoming solar radiation	J m^{-2}
T	Temperature	°C
u	Wind speed	m s^{-1}
$\bar{u}$	Mean windspeed	m s^{-1}
u_*	Friction velocity	m s^{-1}
u_{*T}	Threshold friction velocity	m s^{-1}
z	Distance from ground reference to height of observation	m
z_0	Roughness parameter	m
γ	Cohesive resistances of adsorbed water + particle threshold	N m^{-2}
γ^1	Cohesive resistance of adsorbed water	N m^{-2}
θ	Water fraction	m^3 m^{-3}
θ_{15}	Water fraction at -1500 J kg^{-1}	m^3 m^{-3}
ρ	Air density	kg m^{-3}
τ	Shear stress	N m^{-2}
τ_T	Threshold shear stress	N m^{-2}
ω	Equivalent water content, θ/θ_{15}	m^3 m^{-3}