2.7

Wind Erosion

Roger Funk[1] and Hannes Isaak Reuter[2]

[1]*Leibniz-Centre for Agricultural Landscape Research, Institute of Soil Landscape Research, D-15374 Müncheberg, Germany*
[2]*Joint Research Centre, Institute for Environment and Sustainability, I-21020 Ispra, Italy*

2.7.1 INTRODUCTION

Wind erosion occurs in many arid, semi-arid and agriculturally used areas around the world and is influenced by geological and climatic factors in addition to human activities. Wind erosion leads to land degradation in agricultural areas and has a negative impact on air quality. Dust emission by wind erosion is the largest source of aerosols, which directly or indirectly influences the atmospheric radiation balance and hence global climatic variations (Shao, 2000). Wind carries more fine sediment than any other geological agent. It has been estimated that windblown dust from soil erosion contributes about 500×10^6 t of particulates to the atmosphere each year (Greeley and Iversen, 1985). In view of this fact, it can be concluded that dust is an active factor in the climate system. Model calculations indicate that about 50 % (±20 %) of the total atmospheric dust originates from disturbed soils, i.e. soils affected by cultivation, deforestation or erosion (Tegen *et al*., 1996).

Wind erosion has been overlooked in the past in Europe as a land degradation process. However, it has received more attention as a process responsible for the creeping decrease in soil fertility and as a source of atmospheric pollution (Oldemann *et al*., 1990; Gobin *et al*., 2003; Warren and Bärring, 2003). This is mainly attributed to the removal of fine particles and organic material, which are the most fertile parts of the soil carrying the nutrients and other agents such as pesticides or herbicides. In addition to creeping degradation, single wind erosion events may result in soil losses of more than 100 t/ha^{-1} and cause considerable on- and off-site damages (Funk, 1995; Goossens, 2003). The main problem with wind erosion in Europe is its perception. Heavy sand storms attract attention by disturbing the public once in a while but, in general, the

Soil Erosion in Europe Edited by J. Boardman and J. Poesen
 ISBN: 0-470-85910-5

TABLE 2.7.1 Relationships between quantity of wind erosion, effects of erosion and annual soil loss. (Reproduced from Chepil WS, *Soil Sci. Soc. Am.* 1960, **24**: 143–145, with permission of the Soil Science Society of America)

Extent of erosion	Description of erosion	Annual soil loss (t ha^{-1})
None to insignificant	No distinct visible effects of soil movement	<40
Slight	Soil movement not sufficient to kill winter wheat in boot stage	40–125
Moderate	Removal and associated accumulations to about 2.5 cm depth, sufficient to kill wheat in boot stage	125–375
High	About 2.5–5 cm removal and associated accumulations	375–750
Very high	5–7.5 cm removal with small dune formations	750–1125
Exceedingly high	More than 7.5 cm removal with appreciable piling into drifts or dunes	>1125

processes mostly happen unnoticed. Chepil (1960) pointed out that annual average soil losses up to 40 t/ha^{-1} are possible without any visible sign of erosion (Table 2.7.1). Erosion and deposition processes both take place on large areas and are therefore difficult to identify. In contrast to water erosion, where the eroded material follows determined paths, wind-eroded material is widely dispersed over the landscape. Furthermore, the direction of transport is subject to changes and in some cases completely the opposite, and so are the areas of erosion and deposition.

Wind erosion is widespread in Europe, from Iceland in the north-west to the Russian plains in the far east. There are diverse sets of geographical extensions, which are affected by wind erosion. The affected areas are lowlands or exposed mountains, the climate is dry or humid, the soils are sand, loess or peat and they are fertile or wastelands. The continental dry conditions in eastern Europe favor wind erosion on large areas. In northern Europe, wind erosion is severe on light, sandy soils of the Pleistocene glacial outwash. Close to the coastlines, wind erosion is principally caused by high wind speeds despite humid climatic conditions. All these regions have in common that inappropriate farming practices have intensified the problem (Wilson and Cooke, 1980; Warren and Bärring, 2003).

The spatial extent of wind erosion has increased in recent decades, mainly caused by changes in agricultural practices. The first reason is the spectrum of growing crops, which has changed to greater proportions of arable land crops. Other factors include disturbances of the soil surface by ploughing and a multitude of tillage operations. The time of highest mechanical stress by tillage operations coincides with the time of highest climatic erosivity in spring. Some more factors influencing wind erosion have been identified:

- The higher level of mechanisation has led to larger fields and in consequence to the removal of hedges and other landscape structures.
- Drainage of arable land has caused faster drying of the soil surface, resulting in decomposition of organic matter and decreasing soil aggregate stability.
- Overgrazing is a significant causative factor in the semi-arid and arid regions, where no other type of land use is possible (Frielinghaus, 1990; van Lynden, 1995; Riksen *et al*., 2003).

The effects of wind erosion are soil deterioration, crop damage and pollution of adjacent areas. Soil deterioration includes the loss of fine material and organic matter, the degradation of soil structure and the loss or redistribution of fertilisers and herbicides. The loss of topsoil is the predominant impact of wind erosion in Europe (van Lynden, 1995). In the worst case, the productivity has declined so substantially that arable land has been removed from production, as reported in Sweden and Poland (Jönsson, 1992; Veen *et al*., 1997). The

TABLE 2.7.2 Some on-site and off-site effects of wind erosion. (Reproduced from Goosens D, On-site and off-site effects of wind erosion. In *Wind Erosion on Agricultural Land in Europe*, Warren A (ed.). EUR 20370. European Commission, Brussels, 2003; 29–38, with permission of D. Goossens)

On-site effects	Off-site effects
Soil degradation	**Short-term effects**
1. Fine material may be removed by sorting, leaving a coarse lag	1. Reduced visibility, affecting traffic safety
2. Evacuation of organic matter	2. Deposition of sediment on roads, in ditches, hedges, etc.
3. Evacuation of soil nutrients	3. Deposition of dust in houses, on cars, washing, etc.
4. Degrading water economy in the topsoil	4. Penetration of dust in machinery
5. Degrading soil structure	5. Deposition of dust on agricultural and industrial crops, ruining their quality
6. Stimulated acidification of the topsoil	**Long-term effects**
Abrasion damage	1. Penetration of dust and its constituents in the lungs, causing lung diseases and other respiratory problems
1. Direct abrasion of crop tissue, resulting in lower yields and lower quality	2. Absorption of airborne particulates by plants and animals, leading to a general poisening of the food chain
2. Infection of crops due to the penetration of pathogens	3. Deposition of heavy metals and other eroded chemical substances to the ground, infecting the soil
3. Stimulated dust emission due to sandblasting of the surface top layer	4. Contamination of surface and ground water via deposition of airborne particles
Other damage	5. Increased eutrophication of surface and ground water
1. Infection, with pathogens or soil constituents, of adjacent uncontaminated fields and crops	6. Infection of remote uncontaminated areas, transforming these into new potential sources
2. Accumulation of low-quality wind-blown deposits on the fields	
3. Building of sand accumulations at the field borders, covering of drainage ditches	
4. Burial of plants	
5. Loss of seeds and seedlings	

most severe damage is reported in the former Soviet Union, where wind erosion removes an estimated 1.5×10^6 ha of cropland from cultivation each year and a much larger area is damaged to some extent (Schroeder and Kort, 1989).

Crop damage is caused by the abrasion of seedlings, the excavation of the roots or the burying of the young plants. The consequences often include additional costs for re-sowing. An overview about on- and off-site effects of wind erosion is given in Table 2.7.2 (Goossens, 2003).

An additional factor is the emission of soil particles due to agricultural practice (tillage emission). This kind of emission takes place even under non-critical conditions (calm or low wind speed) and affects especially the finest soil particles. The entire area of a field can emit this fine material several times depending on tillage frequency. Tillage emission depends on the management practices (type and speed of operation), the soil type and the soil moisture conditions. Research related to this kind of emission is scarce at the moment and there is little knowledge about how agronomic practices affect the generation and composition of dust throughout the year. Studies performed in the USA, Sweden, Germany and the UK were focused on human respiratory problems caused by fine particulate matter and measured dust concentrations only at tractor driver level and in the air near the agricultural operation (Darke, 1976; Batel, 1979; Noren, 1985; Clausnitzer and Singer, 1996).

Data on the extent of wind erosion in Europe are very limited and, if available, are the result of different methodologies of assessment on the national scale, which are not comparable. Consequently, there is no

TABLE 2.7.3 Extent of wind erosion in some European countries

Country[a]	Total (1000 ha)	Light	Medium	Severe
Bulgaria[1]	13			13
Czech Republic[2]	963	397	475	91
Denmark[3]	1000			
Estonia[4]	77			
France[5]	500			
Germany	4154			
Lower Saxony[6]	2000			
Mecklenburg[7]	1004	67	543	394
Brandenburg[8]	1150	533	327	290
Hungary[9]	1400			
Latvia[10]	230			
The Netherlands[11]	97			
Poland[12]	8843	5447	3077	318
Russia[13]	7138	2014	4599	525
Slovakia[14]	154			
Sweden[15]	35			
UK[11]	260			
Ukraine[16]	2200			

[a]References: [1]Ivanov, 1997; [2]Janecek *et al*., 2000; [3]Gross and Bärring, 2003; [4]Reintam *et al*., 2001; [5]Montanarella, 2002; [6]Thiermann *et al*., 2000; [7]Funk *et al*., 1996; [8]Funk *et al*., 2001; [9]Fenyö, 1997; [10]FAO, 1997; [11]Riksen and de Graaff., 2001; [12]CNT, 2000; [13]Larionov *et al*., 1997; [14]SSCRI, 2003; [15]Bärring *et al*., 2003, [16]Dolgilevich, 1997.

uniform map of the occurrence of wind erosion in Europe. Nevertheless, there are two regions where European-wide studies show good agreement: severe wind erosion in the north Caucasus and moderate wind erosion along the glacial outwash from the UK to Poland. The wind erosion risk in the Mediterranean region is regarded as high and over a large extent in an assessment by the USDA (USDA, 2003), whereas a European study showed no risk in this area (EEA, 1998). According to the Global Assessment of Soil Degradation, 42×10^6 ha or 4 % of the European territory are affected by wind erosion (Gobin *et al*., 2003). Unfortunately, most European-wide studies exclude the Russian Federation, where very large areas suffer from severe wind erosion. Soils in the north Caucasus region, for example, have lost about 20–60 % of their upper horizons within 15 years (Larionov *et al*., 1997). Studies on a regional scale produce much higher percentages of the affected area, but the comparability of these data is limited by the differences in methodology and definitions. A summary of the available data on the extent of wind erosion in Europe based on national assessments is given in Table 2.7.3.

Regional wind erosion assessments without a spatial reference to the territory of countries have also been made in Austria (Klik *et al*., 2000), Serbia (Letic and Savic, 2002), Lithuania (Racinskas, 1997) and Spain (Gomes *et al*., 2003). Attendant problems of wind erosion have been appraised in Finland (aeolian reactivation of dune fields) (Kotilainen, 2002) and Portugal (coastal wind erosion).

2.7.2 PROCESSES OF WIND EROSION

Wind erosion occurs when three conditions coincide: high wind velocity, a susceptible surface with loose particles, which can be picked up, and insufficient surface protection by crops or plant residues. The two main categories determining the extent of wind erosion are the *erosivity* of the climate and the *erodibility* of the soil.

Both are influenced by the interactions of various other components, resulting in a high temporal variability of the actual erodibility of a particular site (Lyles, 1988; Warren and Bärring, 2003).

2.7.2.1 Erosivity

The *erosivity* of the climate depends on the wind velocity, the amount and distribution of precipitation and evaporation. Their interactions determine the intensity, frequency and duration of wind erosion events on susceptible surfaces.

The wind regime in Europe is determined by three factors: the large temperature difference between the Polar air in the north and the Subtropical air in the south; the land–sea distribution with the Atlantic Ocean to the west, Asia to the east and the Mediterranean Sea and Africa to the south; and the main orographic barriers – the Alps, Pyrenees and Scandinavian mountain chain. Most parts of Europe are influenced by eastward-moving weather systems and so western surface winds dominate there. The average wind velocity in Europe is highest in the north-west (Ireland, Scotland) and decreases in a south-eastern direction. Areas of the highest wind velocities in Eastern Europe are located between the Urals and the coasts of the Caspian, Azow and Black Sea. Local wind systems are prevalent around the Mediterranean Sea. They are characterised by a constancy of strength and direction over long periods, such as the mistral or the scirocco (Troen and Petersen, 1989).

The wind velocity has a daily course caused by thermal effects, with a maximum in the early afternoon and a minimum at night. The yearly variation of wind velocity shows a maximum during the winter in north-western Europe but, owing to a positive climatic water balance, the moist soil surfaces can resist the wind forces in most instances at that time. The climatically highest erosivity is reached in spring, when high wind velocities appear and the evaporation increases owing to rising temperatures. In summer, the average wind velocity is at its minimum in northern Europe whereas in southern Europe local wind systems gain influence (Figure 2.7.1). Precipitation is highest in northern Europe, decreasing in a south to south-eastern

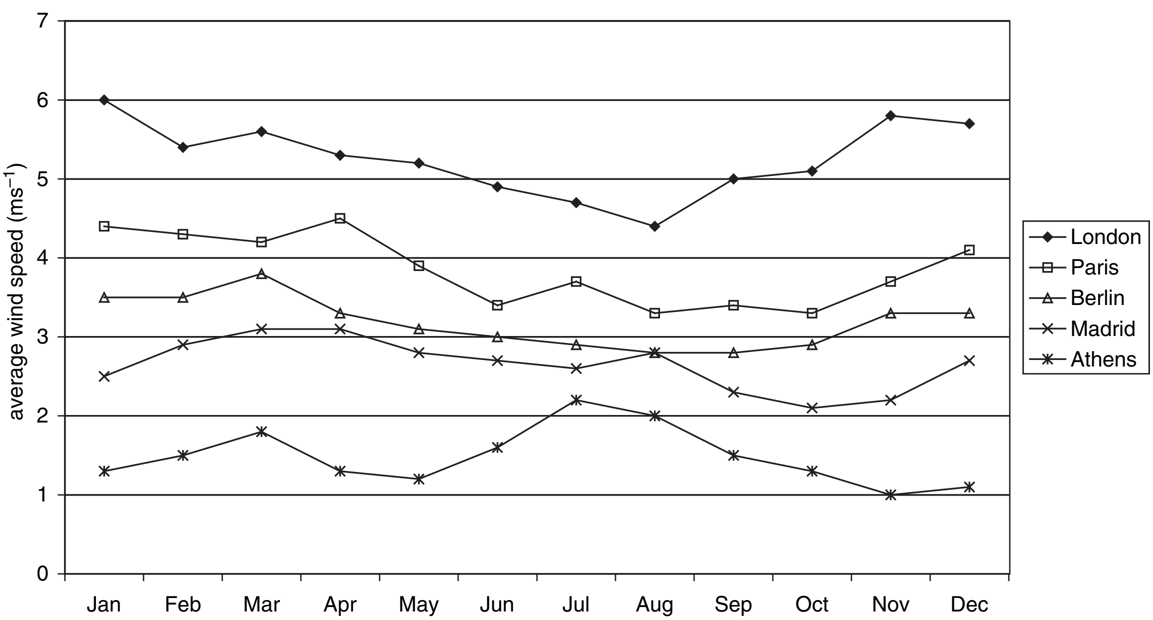

Figure 2.7.1 *Average monthly wind speed for some European cities.* (Reproduced from *World-wide Agroclimatic Database*. FAOCLIM 2. FAO Agrometeorology Group, with permission of the FAO)

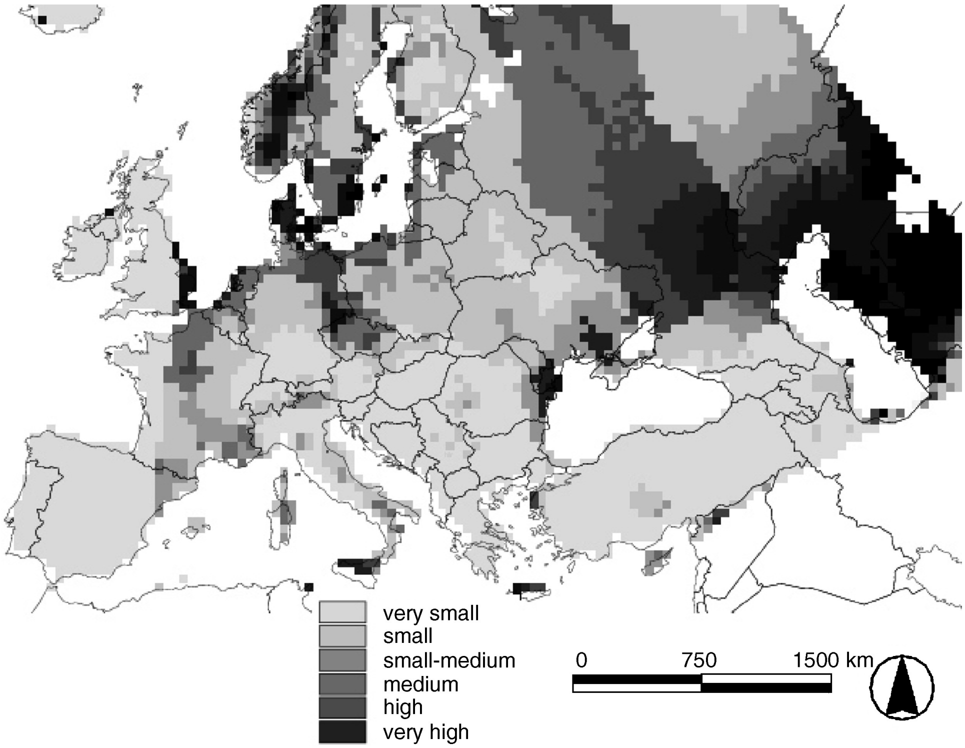

Figure 2.7.2 Wind erosivity of the climate in Europe based on CRU data

direction. The driest regions in Europe are in the south of the Iberian Peninsula and between the Black Sea and the Caspian Sea. The distribution of evaporation is just the reverse of the precipitation pattern, so that two favourable factors for wind erosion coincide in large parts of southern Europe.

An erosivity index for Europe considering the climate factors is shown in Figure 2.7.2. It is based on 30-year average global climate data of the Climate Research Unit on a $0.5 \times 0.5°$ grid. This is assumed to represent the 20th century space–time climate variability (New *et al.*, 1999, 2000). The erosivity is mainly determined by the wind velocity and modified by the moisture conditions. This is considered in the erosivity index by the cube of the annual averages of wind velocity (m s^{-1}] and the ratio of potential evaporation to precipitation.

2.7.2.2 Erodibility

The *erodibility* describes the potential of a soil to erode or, the reverse, the ability to resist the acting wind forces. This is mainly attributed to the texture and organic matter content, which influence the water-holding capacity and the ability of the soil to produce aggregates or crusts (Chepil, 1955). In general, sandy soils are highly erodible because they dry quickly, form only few, weak aggregates and contain a large proportion of particles in the most erodible fraction between 80 and 200 μm. Loamy soils are more resistant against wind erosion but have a greater potential for dust production if they erode.

Natural soils consist of a large variety of grain sizes, which also differ in shape and density. In most cases soil particles are aggregated, with aggregate sizes ranging from decimetres to micrometres. The average

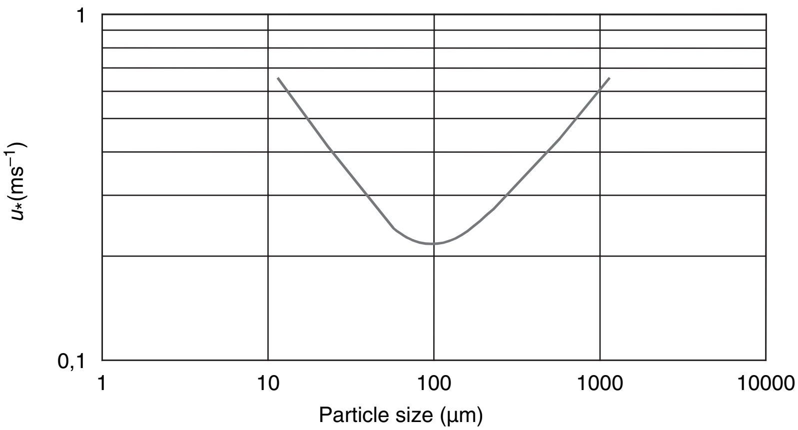

Figure 2.7.3 Dependence of critical friction velocity on particle size (after Bagnold, 1941)

diameter of single grains or aggregates that can be regarded as nonerodible is >840 μm (Chepil, 1942; Woodruff and Siddoway, 1965). These grains are usually too heavy to become airborne by the wind stress. The threshold of motion for uniform grain sizes, expressed in terms of friction velocity of the wind, is shown in Figure 2.7.3. The lowest friction velocity is needed to move particles of size between 80 and 100 μm. The threshold friction velocity increases to the greater diameters caused by weight and to the smaller diameters caused by cohesive forces. The size distribution and stability of aggregates vary within the year and are mainly influenced by soil management practices, hence the erodibility can vary in time and space.

Regarding textural composition, soils with the highest erodibility are located on the sandy glacial outwashes in the northern European lowlands and also on the loess soils of the large aeolian deposits in eastern Europe. The FAO Digital Soil Map of the world has been used to derive an erodibility index by combining soil texture with the organic carbon content of the soils (Figure 2.7.4).

2.7.3 FACTORS INFLUENCING WIND EROSION IN DETAIL

2.7.3.1 Wind

Wind is moving air and is caused by pressure differences in the atmosphere, which in turn result from temperature differences at the Earth's surface. Wind consists of a steady mean part and a superimposed turbulent part. The transport of moisture, heat, momentum and pollutants in the atmospheric boundary layer is dominated in the horizontal direction by the mean wind and in the vertical direction by turbulence. Mean wind is responsible for rapid horizontal transport and can exceed velocities of 100 km/h^{-1}. Turbulence is generated by frictional drag on the air moving over rough surfaces. It results in an irregular swirling motion with turbulent eddies moving up and down. Owing to the increase in wind velocity with height, the net effect of the turbulent motion is always a downward flux of momentum and an upward flux of constituents. Thus, detached soil particles are passed to higher layers of the atmosphere. The magnitude of the vertical wind is about one-tenth of the horizontal velocity (Stull, 1988).

Wind is the driving force of wind erosion if it exceeds a given threshold wind or friction velocity. The latter is better suited to express the momentum flux that the wind exerts on the soil surface and is influenced by the wind and also by surface roughness, as indicated in Equation (2.7.1). Wind velocity increases with height,

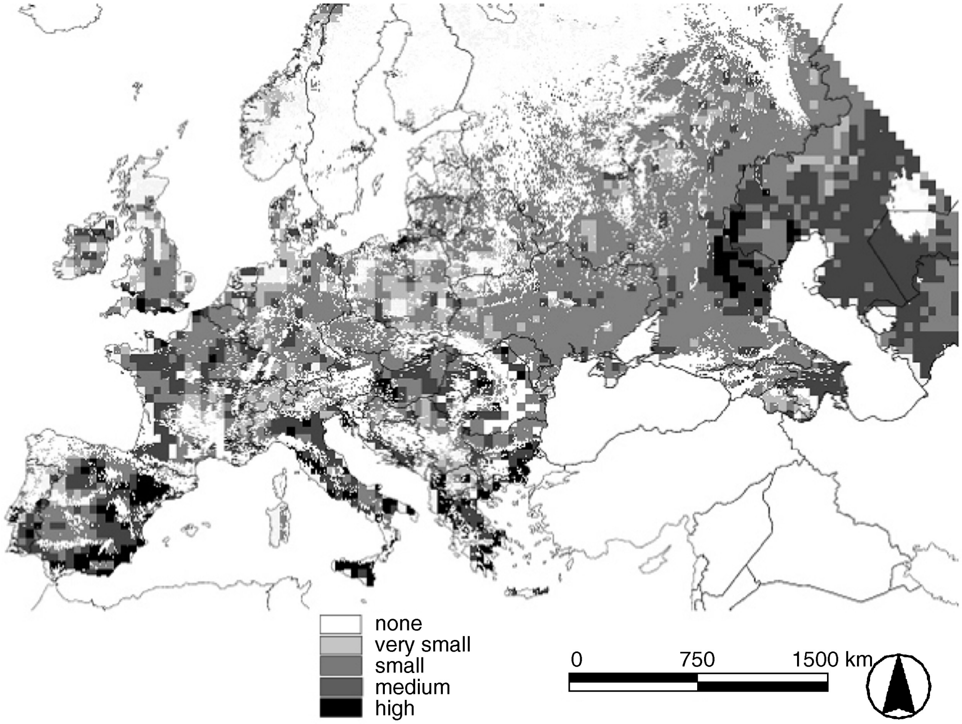

Figure 2.7.4 Soil erodibility in Europe, derived from texture classes and organic matter content, covered with the forest mask (Batjes, 1996; FAO, 2000b, 2003)

whereas the friction velocity is constant within the boundary layer (also called the ‘constant flux layer’). The correlation can be described with the logarithmic wind profile as

$$\frac{u_z}{u_*} = \frac{1}{\kappa} \ln\left(\frac{z}{z_0}\right) \tag{2.7.1}$$

where

u_z = velocity at height $z\,(\mathrm{m\,s^{-1}})$
u_* = friction velocity $(\mathrm{m\,s^{-1}})$
κ = Karman constant for turbulent flow (0.4)
z_0 = roughness length, height at which the velocity is zero (m).

A certain wind or friction velocity has to be exceeded to initiate particle movement. This value is referred as the threshold velocity. For loose, unconsolidated grains there is a close relationship between particle size and this threshold value. Bagnold (1941) derived an expression by equating the acting and the holding forces on the uppermost grains of a surface:

$$u_{*\mathrm{t}} = A\sqrt{\frac{(\sigma - \rho)}{\rho}gd} \tag{2.7.2}$$

where

u_{*t} = threshold friction velocity $(\mathrm{m\,s^{-1}})$

σ = grain density $(\mathrm{kg\,m^{-3}})$

ρ = air density $(\mathrm{kg\,m^{-3}})$

g = gravitational acceleration $(\mathrm{m\,s^{-2}})$

d = grain diameter (m)

A = empirical coefficient which depends on the friction Reynolds number, $u_* d/v$, where v is the kinematic viscosity of the air.

After exceeding the threshold value, the transport capacity of the wind increases rapidly and follows an exponential function with exponents between 2 and 4. One of the basic transport equations that describes the relationship between transport rate and friction velocity was also given by Bagnold (1941):

$$q = C\rho u_*^3 \frac{\sqrt{\dfrac{d_p}{d_{250}}}}{g} \tag{2.7.3}$$

where

q = particle flux (mass per unit width per unit time)

C = a particle size distribution function with values ranging from 1.5 (homogeneous distribution) to 2.8 (heterogeneous distribution)

d_p = particle diameter ($d_{250} = 250\,\mu\mathrm{m}$).

As this equation was developed from investigations of dune sands, its application to soils can provide only a rough estimate. The difficulty in establishing an appropriate equation to fit all possible cases is illustrated by the large number of derived equations that are available in the literature. A compilation of mass transport equations was given by Greeley and Iversen (1985).

A comparative assessment of the local erosivity by wind was given by Beinhauer and Kruse (1994), where the daily wind forces were estimated by summarising the hourly wind integrals for chosen threshold wind velocities:

$$SFU_{6,7,8} = \sum_{n=1}^{24} (FF_n - FFT) {FF_n}^2 \tag{2.7.4}$$

where

$SFU_{6,7,8}$ = daily wind force by assumed threshold velocities of 6, 7 and $8\,\mathrm{m\,s^{-1}}$

FFT = threshold wind velocity

FF_n = hourly average wind velocity.

Further, the precondition is a dry soil surface, which has at least a 2.5-mm thick top layer. The advantage of this method is the representative comparability of the erosivity of the climate with a minimum of input parameters.

2.7.3.2 Other Climatic Factors

The other climatic factors influencing wind erosion are temperature, humidity, radiation, precipitation and evaporation. They cause temporal changes of the actual erodibility by affecting the soil water balance. In general, wet surfaces are stable enough to resist the wind forces, but for the initiation of wind erosion a very thin dry surface layer is sufficient. The water content of this layer is mainly dependent on climatic factors because the evaporation exceeds the hydraulic conductivity of sandy soils to a significant extent. An estimation of the surface water content can be derived by the comparison of the water content of the top layer (<5 mm) and the evaporation. Inclusion of the surface moisture to the transport equation results in (Skidmore 1986)

$$q = C\left(\frac{\rho}{g}\right)\left[{u_*}^2 - {u_{*t}}^2 - 0.5\left(\frac{SW}{WP}\right)^2\right]^{\frac{3}{2}} \tag{2.7.5}$$

where

q = erosion rate $(\mathrm{kg\,m^{-1}\,s^{-1}})$
C = parameter (2.5)
u_* = friction velocity $(\mathrm{m\,s^{-1}})$
u_{*t} = threshold friction velocity of the dry soil $(\mathrm{m\,s^{-1}})$
SW = actual water content of the top soil layer (10 mm) $(\mathrm{kg\,kg^{-1}})$
WP = water content at wilting point (1.5 MPa).

2.7.3.3 Roughness

The term *roughness* is used to describe properties of surfaces ranging from the micro to the macro scale, which represents the effects of single grains and of the topography. The roughness of a soil surface affects wind erosion in two ways. First, a rough surface increases the turbulence and therefore the dissipation of the kinetic energy of the wind at the surface, resulting in a slowdown of the wind velocity (Stull, 1988), and second, the leeward side of clods or furrows is sheltered against wind action and particle impact (Potter and Zobeck, 1988). In addition, the moving material can be trapped and the avalanching increase of transport is impeded.

According to Römkens and Wang (1986), the roughness of a field can be classified in four scales:

1. roughness caused by single grains or aggregates, <2 mm
2. roughness caused by aggregates or clods, <100 mm
3. oriented roughness caused by tillage operation, 100–300 mm
4. topography.

Roughness can be determined directly in the field by measuring height differences with pin meters or laser relief meters (Zobeck and Onstad, 1987; Huang and Bradford, 1990) and indirectly by comparing the foreshortened length of a chain lying on a rough surface with the full length of this chain (Saleh, 1993). Parameters to describe roughness are often the variance, the standard deviation (SD) or the root of the SD of the logarithm of the measured height values (Kuipers, 1957; Allmaras *et al.*, 1966). Geostatistical methods are also usable, such as spectral analysis or semivariograms (Linden and van Doren, 1986).

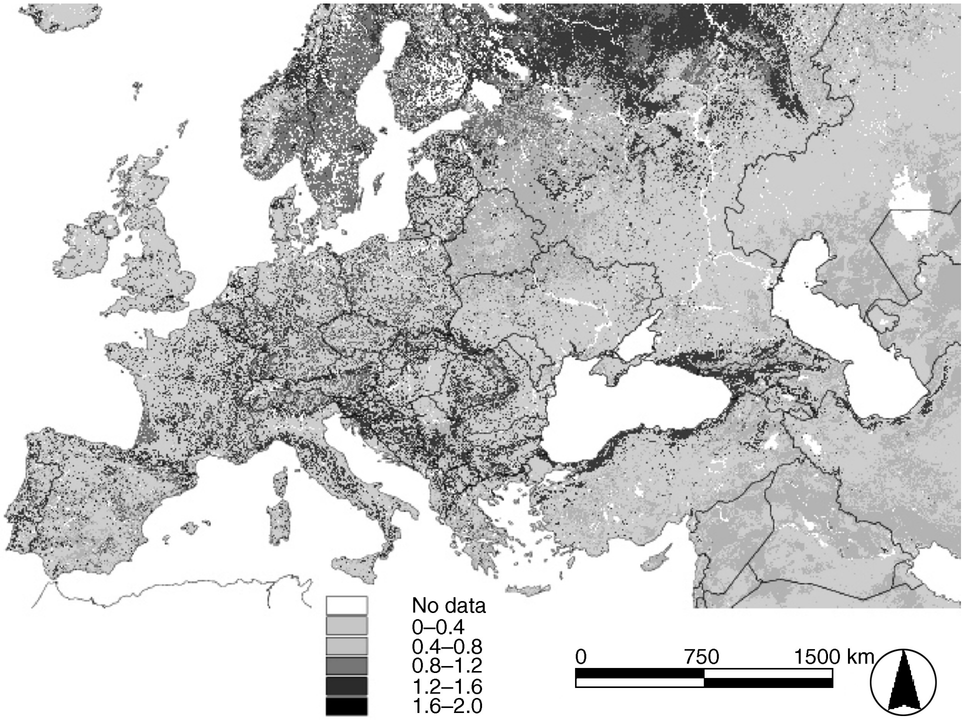

Figure 2.7.5 Aerodynamic roughness length ($z_0$in m) in Europe, derived from CORINE and land cover classification using parameters TA–LUFT (2001)

The roughness of a surface influences the wind profile [see Equation (2.7.1)], so conversely the wind profile can be used to derive information about the surface roughness. The parameter is the aerodynamic roughness length, z_0, which is a feature of the surface and can only be estimated by the wind profile. The wind velocity is considered to be zero at this height. In general, z_0 can be estimated as 0.03 times the average roughness height (Abtew *et al*., 1989). Aerodynamic roughness length classes for most terrain types have been derived by meteorology and can be found in meteorology handbooks.

On the large scale, the roughness of a landscape is determined by the size and distribution of the roughness elements that it contains. These are principally vegetation and built-up areas, which are dominant at the landscape scale. Commonly, the roughness length z_0 can be used to parameterise the roughness of a terrain. Roughness length values for Europe are shown in Figure 2.7.5. The highest values are generally in the mountainous parts. Western Europe is characterized by a heterogeneous distribution caused by a diversified landscape compared with eastern Europe, with more homogeneity and generally smaller roughness length values.

Another concept to determine roughness is the ‘shelter angle distribution’ (Potter *et al*., 1990), which is referred to the sheltering effect of roughness against particles in saltation and their impact angles.

2.7.3.4 Vegetation

2.7.3.4.1 In the Field

Permanent vegetation is the best measure of protecting soils from wind erosion and plant residues also serve this purpose. The effect of plants on wind erosion can be expressed by the percentage of the surface covered with non-erodible plant material or by morphometric parameters such as silhouette area or the leaf area index (LAI). The soil cover, in general, refers to flat residues lying on the surface. An equation which includes a wide variety of materials covering the soil surface was given by Fryrear (1985) and expresses the soil loss ratio (soil loss from covered soil relative to soil loss from bare soil) in terms of the soil cover:

$$SLR = 1.81\mathrm{e}^{-0.072\%\mathrm{SC}} \tag{2.7.6}$$

where

SLR = soil loss ratio
$\%SC$ = percentage soil cover.

Equation (2.7.6) is limited to soil cover between 8 and 80 %.

On arable land, the growing crops change every year and therefore there are temporal variations in the protective effect of vegetation according to the time of the year. Measurements have shown that soil covers <10 % increase wind erosion (Morgan and Finney, 1987; Funk, 1995; Sterk, 2000). Soil cover >10 % reduces wind erosion rapidly. The soil loss is reduced to about 50 % at a soil cover of 20 % compared with a bare surface and is prevented completely at soil cover >40 %.

The effect of vegetation on the surface wind can be explained with the principles of fluid mechanics, as roughness elements absorb part of or the entire shear forces from the airflow. As long as the vegetation is sparse and the wind can reach the soil surface, the total shear stress can be split into the stress on the vegetation, τ_v, and the stress on the underlying surface, τ_s (Marshall 1971):

$$\tau = pu_*{}^2 = \tau_v + \tau_s \tag{2.7.7}$$

Within a plant canopy, the drag resistance is proportional to the dynamic pressure $0.5\rho_A u^2$ (where ρ_A = air density and u = wind velocity) and the frontal (silhouette) area, L_A. The shape is taken into consideration by a drag coefficient, C_v. It can be calculated with Equation (2.7.8) by measuring the drag forces, W_v:

$$W_v/F = W_v/NA = C_v \rho_A u^2 L_A/2A \tag{2.7.8}$$

The plant density is given by $A = F/N$, where A is the uniform ground area per plant, F the total floor area and N the number of plants. For row crops, a special case exists (regular arrays) with $A = aw$, where a is the distance between the rows and w the mean distance of the plants within a row. Drag coefficients for plants have been evaluated by Morgan and Finney (1987) and Funk and Frielinghaus (1998). There is another possibility for calculating the drag coefficient from measurements of the wind profile (Morgan, 1989):

$$C_v = \frac{2u_*}{\int_{z_1}^{z_2} u^2 L_A(z)\mathrm{d}z} \tag{2.7.9}$$

Raupach (1992) developed equations to predict the stress partition on rough surfaces for practical applications. The equations were derived for solid roughness elements but they are also applicable to

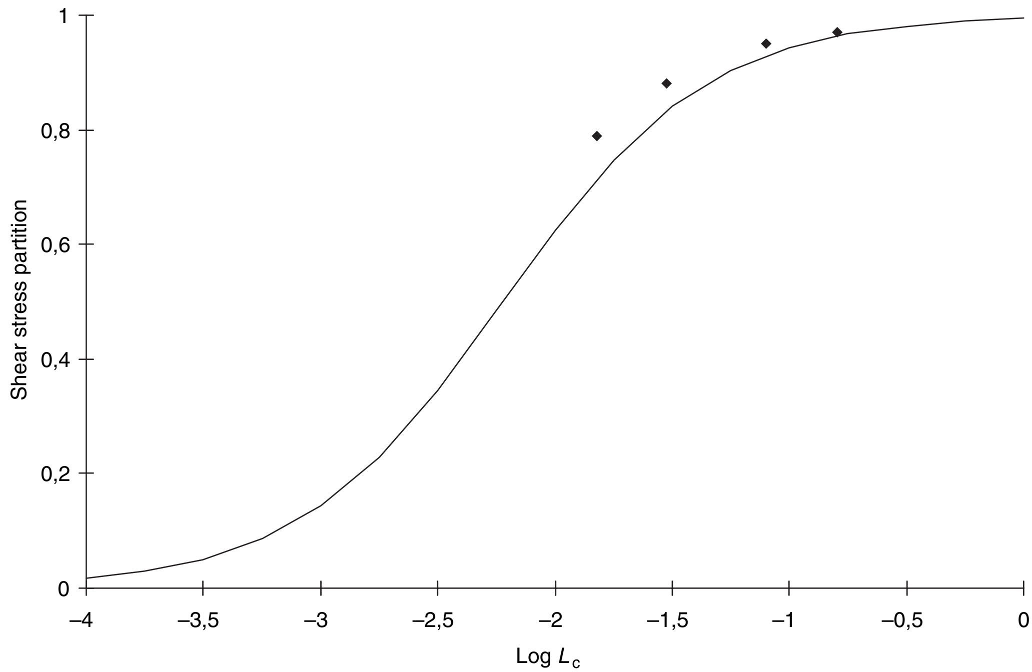

Figure 2.7.6 Dependence of shear stress partition of vegetation (symbols) and solid roughness elements (line) on the lateral cover (plant silhouette area per unit total floor area)

vegetation if its special features are considered. Increasing wind velocity causes changes of the drag coefficients and silhouette areas (Funk and Frielinghaus, 1998). The stress partition given by Raupach (1992) (see Figure 2.7.6) is

$$\frac{\tau_v}{\tau} = \frac{\beta L_c}{1 + \beta L_c} \text{ and } \frac{\tau_s}{\tau} = \frac{1}{1 + \beta L_c} \tag{2.7.10}$$

where

$\tau, \tau_v, \tau_s =$ shear stress (total, vegetation, surface)
$\beta =$ ratio of drag coefficients C_v/C_s (vegetation/surface)
$L_c =$ lateral cover.

2.7.3.4.2 *Around the Field*

Shelterbelts are another kind of vegetation used to prevent wind erosion. Especially in the northern European regions with high wind speeds there are many traditional systems of hedges, which protect the fields. Hedges have influences on the local wind field and on many other components of the micro- and macroclimate (Figure 2.7.7). They should best be arranged perpendicularly to the prevailing wind direction. Shelterbelts give protection downwind for a distance of about 25 times of their height, depending on the porosity, kind of trees and number of rows (Nägeli, 1943). Most effective are triple rows, with a tree

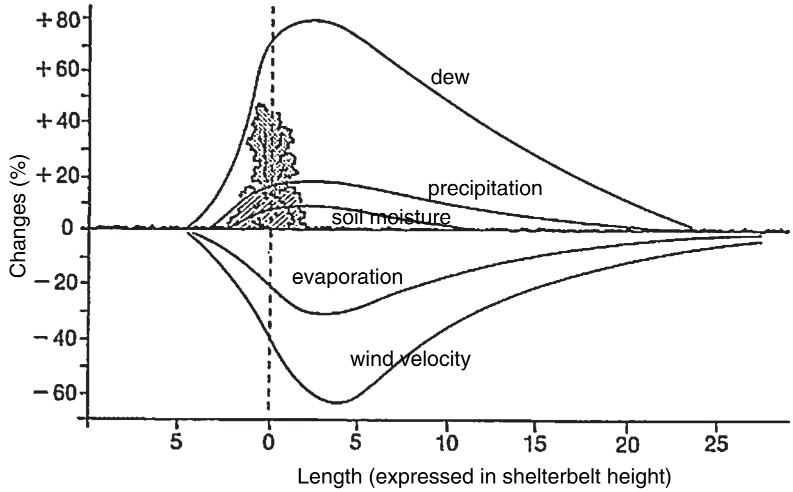

Figure 2.7.7 Effects of shelterbelt height on wind velocity, evaporation, soil moisture precipitation and dew

row in the centre flanked by shrubs, with a triangular cross-section (Chepil and Woodruff, 1963). Increasing permeability with height prevents ‘wall effects’ on the leeward side. The distances between shelterbelts depend on the erodibility of the soil. Highly erodible soils need a dense network of hedges, which is contrary to an effective work rate of field machinery (Riksen *et al.*, 2003). The installation of shelterbelts is fairly expensive, needs a long period of support and becomes effective only after a number of years. Therefore, shelterbelts can be only a supporting measure to prevent wind erosion in combination with measures in the field.

2.7.4 MODES OF PARTICLE MOTION

The modes of particle motion are closely related to the particle size. The density and the shape of particles are also important, if the consideration is extended to agriculturally used soils, which comprise all textural classes and contain organic material with a much lower density than mineral particles of the same size. The first distinction of particle motion was given by Bagnold (1941), who distinguished creep, saltation and suspension (Figure 2.7.8).

Creep. Mineral particles larger than 500 μm are too heavy to be lifted from the surface by wind. They roll or are pushed along the surface by wind or the impact of saltating particles.

Saltation. The bouncing motion of particles across the surface is called saltation. Inactive particles are thrown steeply aloft, accelerate in the higher layers and return to the surface with a small impact angle. The typical lift-off angle is around 55° and typical impact angle is around 10°. The grain size is between approximately 70–500 μm. It is the principal mechanism of transport for large quantities of soil particles in the direction of the wind and amounts to 50–80 % of the total transport. Saltation causes an avalanching increase in transport, if there is sufficient erodible material, until saturation of the transport capacity of the wind is reached. Below saturation, the surplus energy of the bouncing particles works to abrade crusts or aggregates. The height of the saltation layer is often below 1 m (Lyles, 1988; Shao, 2000).

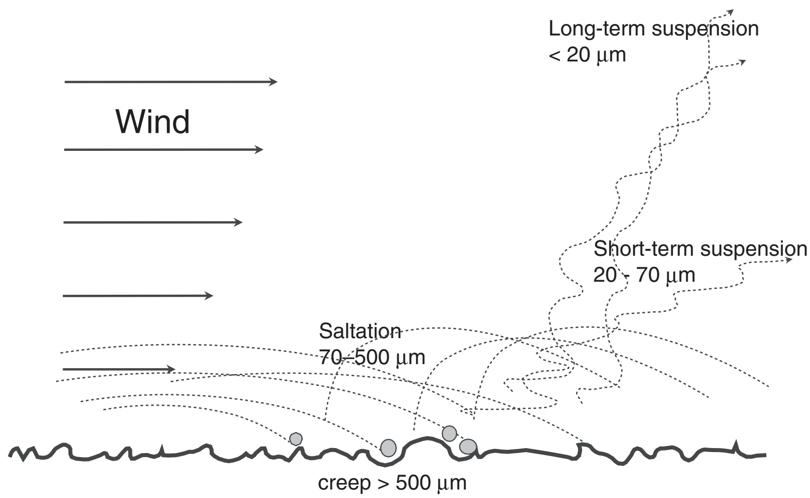

Figure 2.7.8 The three main transport modes of particles during wind erosion

Suspension. Particles with a smaller terminal velocity (u_F) than the vertical upward directed turbulent motions become suspended. The value of the vertical component of the wind in the boundary layer is approximately equal to the friction velocity u_*. Hence all grains with terminal velocities smaller than the actual friction velocity are transported upwards. For wind erosion events this is only relevant above the threshold friction velocity u_{*t}, so the boundary is (Greeley and Iverson 1985)

$$\frac{u_F}{u_{*t}} \bigg/ \frac{u_*}{u_{*t}} = 1 \tag{2.7.11}$$

Furthermore, suspension can be divided into long- and short-term suspension. Particles smaller than 20 µm are subjected to long-term suspension, whereby they can be transported for several days over several hundred kilometres. Particles with diameters between 20 and 70 µm remain suspended for only a few hours and cannot be transported for very large distances (Shao, 2000).

2.7.4.1 The Saltation and Suspension Link

Saltation can be referred to as the driving process of wind erosion. Soil particles, which were lifted by fluid dynamic forces, accelerate in the higher layers and return to the ground. Owing to the momentum gained, these particles rebound and continue their movement in saltation and/or strike other grains. This causes a rapid increase in the downwind transport rate, also known as 'avalanching'. As the transport rate increases, the surface wind velocity decreases owing to the extraction of momentum by the grains in motion. After certain distances, equilibrium or the maximum transport capacity of the wind is reached. The transport capacity is independent of soil type and about the same for all soils, but the distance from the point of initiation to saturation varies with soil erodibility (Chepil, 1959). These distances differ between a few metres on highly erodible surfaces such as sandy beaches, where erodible material is not limited, to a few hundred metres on arable land, as shown in Figure 2.7.9. Aggregates or crusts at the soil surface generally reduce wind erosion on arable land by reducing the amount of erodible material. Therefore, the

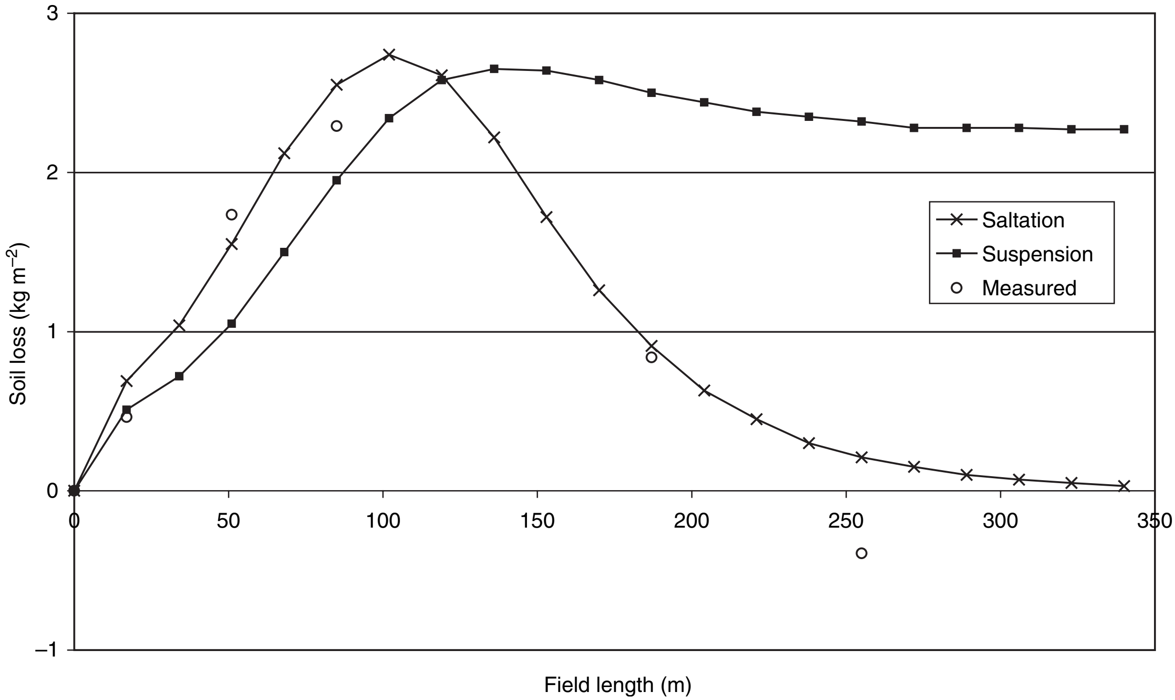

Figure 2.7.9 Soil loss over a field divided into the saltation and suspension part (measurements by Funk, 1995; saltation and suspension, calculated with WEPS)

transport is initially unsaturated. As long as the transport is unsaturated, the particles in the saltation layer acquire more energy by the acceleration in the higher layers than is absorbed, and this surplus energy in the saltation layer is responsible for the abrasion of aggregates or crusts and the production of more transportable material.

Another effect of the saltating grains is sandblasting. Very fine particles ($<20\,\mu m$) can be ejected from the soil surface or from the aggregates, where they are generally embedded (Alfaro *et al.*, 1997). Once detached, these particles stay in suspension even if the initial conditions of the threshold wind or friction velocity decrease (see Figure 2.7.3). Consequently, saltation is the determining process for the destruction of aggregates or crusts, the rapid increase in the transport intensity and the production of suspension-sized particles.

2.7.5 SUMMARY

Wind erosion is an important land surface process in Europe, which has acquired increasing relevance in the last decade. Especially the emission of the finest and most valuable soil particles has led to degradation processes affecting the agriculturally used areas and the atmosphere. The focus of consideration has changed from on-site (in most cases economic effects) to off-site effects (impacts on environment). Also, the discussions about the climate change and the associated effects on human society in general and agriculture in particular have set the problem of wind erosion in a broader perspective.

Significant progress has been made in the last 10 years. Computer and Geographic Information Systems (GIS) have improved the abilities for modelling and managing huge amounts of data. The political changes in eastern Europe and the Internet have facilitated and accelerated the exchange between scientists. Field and laboratory studies have been carried out at national and international levels, so that the knowledge about the mechanics of wind erosion has increased. Measuring devices have been developed or improved and reliable methods to quantify the windblown sediment applied. Wind tunnels are in use in Aarhus, Aberdeen, Bremen, Debrecen, Ghent, London, Loraine, Müncheberg and Paris to investigate the basic processes in the laboratory or at the field site.

Field studies have been carried out by Sterk (1997) in Niger and Funk (1995) in Germany to quantify the mass transport for single erosion events and to improve the understanding of the wind erosion processes. Studies with special emphasis on the dust fraction have been carried out by Herrmann (1996) in Niger and by Gross (2002) and Goossens and Gross (2002) in northern Germany. Aspects of erodibility, roughness and soil wetness to the erosion threshold and transport intensity have been investigated by Neemann (1991), Düwel (2000) and Cornelis *et al*. (2004). The combined effects of sand transport and dust production were investigated by Alfaro *et al*. (1997). Projects such as 'Wind Erosion on European Light Soils' (WEELS) and 'Wind Erosion and Loss of Soil Nutrients in Semiarid Spain' (WELSON) have promoted cooperation between European researchers.

The ascertainment of the soil losses by wind erosion and their detection seem closely related to developments of the measuring techniques. Especially laser technology has shifted the detection of particles to a new level in the last decade. Widely used methods also include real-time particle sizing in submicron dimensions and the vertical scanning of the atmosphere by Lidar. In spite of the progress in recent years, the transfer of the results to larger scales should be the next, European-wide challenge.

REFERENCES

Abtew W, Gregory JM, Borelli J. 1989. Wind profile: estimation of displacement height and aerodynamic roughness. *Trans. ASAE* **32**: 521–527.

Alfaro SC, Gaudichet A, Gomes L, Maille M. 1997. Modeling the size distribution of a soil aerosol produced by sandblasting. *J. Geophys. Res.* **102**: 11239–11249.

Allmaras RR, Burwell RE, Larson WE, Holt RF, Nelson WW. 1966. Total porosity and random roughness of the interrow zone as influenced by tillage. *USDA Cons. Res. Rep.* No. 7.

Bagnold RA. 1941. *The Physics of Blown Sand and Desert Dunes*. Methuen, London.

Bärring L, Jönsson P, Mattsson JO, Åhman R. 2003. Wind erosion on arable land in Scania, Sweden and the relation to the wind climate–a review. *Catena* **52**: 173–190.

Batel W. 1979. Staubbelastung und Staubzusammensetzung an Arbeitsplätzen der landwirtschaftlichen Produktion und daraus abzuleitende Belastungsgrenzen und Staubschutzmassnahmen. *Grundlagen Landtech.* **29**: 41–54.

Batjes NH. 1996. *Documentation to ISRIC–WISE Global Data Set of Derived Soil Properties on a 0.5 by 0.5 Degree Grid*. Working Paper and Preprint 96/05. International Soil Reference and Information Centre (ISRIC), Wageningen.

Beinhauer R, Kruse B. 1994. Soil erosivity by wind in moderate climates. *Ecol. Model.* **75/76**: 279–287.

CNT. 2000. *Environmental Indicators Study*. Center for Nations in Transition. http://www.hhh.umn.edu/centers/cnt/indicators/poland/poleros.htm

Chepil WS. 1942. Measurement of wind erosiveness of soils by dry sieving procedure. *Sci. Agric.* **23**: 154–160.

Chepil WS. 1955. Factors that influence clod structure and erodibility of soil by wind. IV. Sand, silt and clay. *Soil Sci*. **80**: 155–162.

Chepil WS. 1959. Wind erodibility on farm fields. *J. Soil Water Conserv.* **14**: 214–219.

Chepil WS. 1960. Conversion of relative field erodibility to annual soil loss by wind. *Soil Sci. Soc. Proc*. **24**: 143–145.

Chepil WS, Woodruff NP. 1963. The physics of wind erosion and its control. *Adv. Agron*. **15**: 211–302.

Clausnitzer H, Singer MJ. 1996. Respirable-dust production from agricultural operations in the Sacramento Valley, California. *J. Environ. Qual.* **25**: 877–884.

Cornelis WM, Gabriels D, Hartmann R. 2004. A parametrisation for the threshold shear velocity to initiate deflation of dry and wet sediment. *Geomorphology* **59**: 43–51.

Darke CS. 1976. Respiratory disease of agricultural workers harvesting grain. *Thorax* **31**: 294–302.

Dolgilevich MJ. 1997. Extent and severity of wind erosion in the Ukraine. http://www.weru.ksu.edu/symposium/proceedings/dolgilev.pdf

Düwel O. 2000. Die Bedeutung der Bodenrauhigkeit für die Bodenerosion durch Wind – Ein Beitrag zur Quantifizierung der Bodenverluste. *Geol. Jb., Reihe F*, SF1.

EEA. 1998. *Europe's Environment: the Second Assessment*. European Environment Agency. Elsevier, Amsterdam.

FAO. 1997. *Irrigation in the Countries of the Former Soviet Union in Figures*. FAO, Rome, 1997.

FAO. 2000a. *World-wide Agroclimatic Database*. FAOCLIM 2. FAO Agrometeorology Group, Rome.

FAO. 2000b. *Global Forest Resource Assessment 2000. Main Report*. FAO Forestry Paper 140. FAO, Rome.

FAO. 2003. *The Digital Soil Map of the World*. Version 3.6. FAO, Rome.

Fenyö G. 1997. *State of Land Ownership and Land Utilization in Hungary*. FAO, Rome. http://www.fao.org/sd/LTdirect/LTforum/LTfo0014.htm

Frielinghaus M. 1990. *Stand der Erosionsforschung in der DDR*. Sonderheft, Berichte über Landwirtschaft, Bodennutzung und Bodenfruchtbarkeit, Band 3, Bodenerosion.

Fryrear DW. 1985. Soil cover and wind erosion. *Trans. ASAE* **28**: 781–784.

Funk R. 1995. *Quantifizierung der Winderosion auf einem Sandstandort Brandenburgs unter besonderer Berücksichtigung der Vegetationswirkung*. ZALF Bericht Nr. 16. Zentrum für Agrarlandschafts- und Landnutzungsforschung, Müncheberg.

Funk R, Frielinghaus M. 1998. Influence of sugar beet and corn on wind erosion. In *Proceedings of the International Workshop on Technical Aspects and Use of Wind Tunnels for Wind-erosion Control; Combined Effect of Wind and Water on Erosion Processes*, November 17–18, Ghent, Belgium. ICE Special Report No. 1998/1. University of Ghent, Ghent; 53–64.

Funk R, Frielinghaus M, Thiere J. 1996. Risikoabschätzung der Winderosion für das Land Mecklenburg – Vorpommern als Grundlage für Schutzstrategien. *Mitt. Dtsch. Bodenkd. Ges.* **79**: 379–382.

Funk R, Deumlich D, Steidl J. 2001. GIS application to estimate the wind erosion risk in the Federal State of Brandenburg. In *Soil Erosion Research for the 2st Century, Proceedings of the International Symposium*, January 3–5, 2001, Honolulu, Hawaii; 400–403.

Gobin A, Govers G, Jones R, Kirkby M, Kosmas C. 2003. *Assessment and Reporting on Soil Erosion*. Technical Report 94. European Environment Agency, Copenhagen.

Gomes L, Arrue JL, Lopez MV, Steerk G, Richard D, Gracia R, Sabre M, Gaudichet A, Frangi JP. 2003. Wind erosion in a semiarid agricultural area of Spain: the WELSONS project. *Catena* **52**: 235–256.

Goossens D. 2003. On-site and off-site effects of wind erosion. In *Wind Erosion on Agricultural Land in Europe*, Warren A (ed.). EUR 20370. European Commission, Brussels; 29–38.

Goossens D, Gross J. 2002. Similarities and dissimilarities between the dynamics of sand and dust during wind erosion of loamy sandy soil. *Catena* **47**: 269–289.

Greeley R, Iversen JD. 1985. *Wind as a Geological Process on Earth, Mars, Venus and Titan*. Cambridge Planetary Science Series 4. Cambridge University Press, Cambridge.

Gross J. 2002. Quantifizierung winderosionsbedingter Staubausträge in Agrarlandschaften Niedersachsens. Geosynthesis 12. Universität Hannover, Hannover.

Gross J, Bärring L. 2003. Wind erosion in Europe: where and when. In *Wind Erosion on Agricultural Land in Europe*, Warren A. (ed.). EUR 20370. European Commission, Brussels; 13–28.

Herrmann L. 1996. Staubdepositionen auf Böden West-Afrikas. *Hohenheimer Bodenkundliche*, Hefte 36. Universität Hohenheim, Hohenheim.

Huang C.-H Bradford, J.M. 1990. Portable laser scanner for measuring soil surface roughness. *Soil Sci. Soc. Am. J.* **54**: 1402–1406.

Ivanov P. 1997. Meteorological and climatological aspects of wind erosion in Bulgaria. http://www.weru.ksu.edu/symposium/proceedings/ivanov.pdf

Janecek M, Tippl M, Pivcova J, Vetiskova D. 2000. Mapy potenciální ohroženosti zemědělských půd ěské republiky vodní a větrnou erozí. Výstup z projektu NAZV EP 7057 Způsoby omezení degradace půd erozí a systémy protierozní ochrany. VÚMOP Praha, Prague.

Jönsson P. 1992. Wind erosion on sugar beet fields in Scania, southern Sweden. *Agrie. Forest Meteorol.* **62**: 141–157.

Klik A, Madl W, Cepuder P. (2000). Assessment of soil erosion by wind using WEQ. In *Abstracts of ASA/CSSA/SSSA Meeting*, Minneapolis, MN, November 5–9, 2000; 326.

Kotilainen M. 2002. Subarctic aeolian reactivity in northern Finnish Lapland: a 10 ka year record. In *Proceedings of ICAR5/ GCTE-SEN Joint Conference*, Lee JA, Zobeck TM. (eds). International Center for Arid and Semiarid Lands Studies, Texas Tech University, Lubbock, TX; 395–397.

Kuipers H. 1957. A reliefmeter for soil cultivation studies. *Neth. J. Agric. Sci.* 255–262.

Larionov G, Skidmore EL, Kiryukhina ZP, 1997. Wind erosion in Russia: spreading and quantitative assessment. http://www.weru.ksu.edu/symposium/proceedings/larionov.pdf

Letic L, Savic R. 2002. Wind Erosion in Voivodina. In *Proceedings of ICAR5/GCTE-SEN Joint Conference*, Lee JA, Zobeck TM. (eds). International Center for Arid and Semiarid Lands Studies, Texas Tech University, Lubbock, TX; 164–167.

Linden DR, van Doren DM. 1986. Parameters for characterization tillage induced soil surface roughness. *Soil Sci. Soc. Am. J.* **50**: 1560–1565.

Lyles L. 1988. Basic wind erosion processes. *Agric. Ecosyst. Environ.* **22/23**: 91–101.

Marshall JK. 1971. Drag measurements in roughness arrays of varying density and distribution. *Agric. Meteorol.* **8**: 269–292.

Montanarella L. 2002. Soil at the interface between agriculture and environment. http://europa.eu.int/comm/agriculture/envir/report/en/inter_en/report.htm

Morgan RPC. 1989. Design of in-field shelter systems for wind erosion control. In *Soil Erosion Protection Measures in Europe*, Schwertmann U, Rickson RJ, Auerswald K. (eds). Soil Technology Series 1. Catena Verlag, Cremlingen-Destedt; 15–23.

Morgan RPC, Finney HJ. 1987. Drag coefficients of single crop rows and their implications for wind erosion control. In *International Geomorphology 1986, Part II*, Gardiner V. (ed.). John Wiley & Sons, Ltd, Chichester; 449–460.

Nägeli W. 1943. Untersuchungen über die Windverhältnisse im Bereich der Windschutzstreifen. *Mitt. der Schweiz. Anst. für forstl. Versuchswesen*, Band XXIII.

Neemann W. 1991. Bestimmung des Bodenerodierbarkeitsfaktors für winderosionsgefährdete Böden Norddeutschlands. *Geol. Jb., Reihe F25*.

New M, Hulme M, Jones PD. 1999. Representing 20th century space–time climate variability. Part I: development of a 1961–90 mean monthly terrestrial climatology. *J. Climate* **12**: 829–856.

New MG, Hulme M, Jones PD. 2000. Representing 20th century space–time climate variability. Part II: development of 1901–1996 monthly terrestrial climate fields. *J. Climate* **13**: 2217–2238.

Noren O. 1985. *Dust Concentrations During Operations with Farm Machines*. ASAE Paper No. 85–1055. American Society of Agricultural Engineers, St Joseph, MI.

Oldemann LR, Hakkeling RTA, Sombroek WG. 1990. *World Map of Human-induced Soil Degradation: an Explanatory Note*. Global Assessment of Soil Degradation (GLASOD), ISRIC and United Nations Environment Program (UNEP), FAO-ITC. International Soil Reference and Information Centre (ISRIC), Wageningen. United Nations Environment Programme (UNEP), in cooperation with Winand Staring Centre, Food and Agriculture Organization of the United Nations (FAO). International Institute for Aerospace Survey and Earth Science (ITC), Enschede.

Potter KN, Zobeck TM. 1988. *Simulation of Soil Microrelief*. ASAE Paper No. 88–2557. American Society of Agricultural Engineers, St Joseph, MI.

Potter KN, Zobeck TM, Hagen LJ. 1990. A microrelief index to estimate soil erodibility by wind. *Trans. ASAE* **33**: 151–155.

Racinskas A. 1997. Assessment of natural conditions in Lithuania from the point of view of wind erosion. http://www.weru.ksu.edu/symposium/proceedings/racinska.pdf

Raupach MR. 1992. Drag and drag partition on rough surfaces. *Boundary Layer Meteorol.* **60**: 375–395.

Reintam L, Rooma I, Kull A. 2001. Map of soil vulnerability and degradation in Estonia. In *Sustaining the Global Farm*. Selected papers from the 10th International Soil Conservation Organization Meeting Held May 24–29, 1999 at Purdue University and the USDA-ARS National Soil Erosion Research Laboratory, Scott DE, Mohtar RH, Steinhardt GC. (eds). International Soil Conservation Organization in cooperation with the USDA and Purdue University, West Lafayette, In; 1068–1074.

Riksen MJP, de Graaff J. 2001. On-site and off-site effects of wind erosion on European light soils. *Land Degrad. Dev.* **12**: 1–11.

Riksen M, Brouwer F, Spaan W, Arrue JL, Lopez MV. 2003. What to do about wind erosion. In *Wind Erosion on Agricultural Land in Europe*. Warren A. (ed.). EUR 20370. European Commission, Brussels; 39–54.

Römkens MJM, Wang JY. 1986. Effect of tillage on surface roughness. *Trans. ASAE* **29**: 429–433.

Saleh A. 1993. Soil roughness measurement: chain method. *J. Soil Water Conserv.* **48**: 527–529.

Schroeder WR, Kort J. 1989. Shelterbelts in the Soviet Union. *J. Soil Water Conserv.* **44**: 130–134.

Shao Y. 2000. *Physics and Modelling of Wind Erosion*. Atmospheric and Oceanographic Sciences Library, Vol. 23. Kluwer, Dordrecht.

Skidmore EL. 1986. Wind-erosion climatic erosivity. *Climate Change* **9**: 195–208.

SSCRI. 2003. Soil Science and Conservation Research Institute, Bratislava, main activities. http://www.uvtip.sk/english/rezort/vupu/akti.html

Sterk G. 1997. *Wind Erosion in the Sahelian Zone of Niger: Processes, Models and Control Techniques*. Tropical Resource Management Papers No. 15. Wageningen Agricultural University, Wageningen.

Sterk G. 2000. Flattened residue effects on wind speed and sediment transport. *Soil Sci. Soc. Am. J.* **64**: 852–858.

Stull RB. 1988. *An Introduction to Boundary Layer Meteorology.* Atmospheric Science Library. Kluwer, Dordrecht.

TA-Luft (2001): *Erste Allgemeine Verwaltungsvorschrift zum Bundes-Immissionsschutzgesetz.* Technische Anleitung zur Reinhaltung der Luft – TA-Luft Stand 12.06.2001.

Tegen I, Lacis AA, Fung IY. 1996. The influence of mineral aerosol from disturbed soils on the global radiation budget. *Nature* **380**: 419–422.

Thiermann A, Sbresny J, Schäfer W. 2000. Ermittlung der Erosionsgefährdung durch Wind. *Mitt. Dtsch. Bodenkd. Ges.* **92**: 104–107.

Troen I, Petersen EL. 1989. *European Wind Atlas*. Published for the Commission of the European Communities Directorate-General for Science, Research and Development, Brussels, Belgium. Riso National Laboratori, Roskilde.

USDA. 2003. Maps of the wind erosion vulnerability and human induced wind erosion risk. http://www.nrcs.usda.gov/technical/worldsoils/mapindex/eroswind.html and http://www.nrcs.usda.gov/technical/worldsoils/mapindex/ewinrisk.html

van Lynden GWJ. (1995). *European Soil Resources. Current Status of Soil Degradation, Causes, Impacts and Need for Action*. Nature and Environment, No. 71. Council of Europe Publications, Strasbourg.

Veen PH, Kampf H, Liro A. 1997. *Nature Development on Former State farms in Poland*. Report within the framework of the Memorandum of Understanding for Nature Conservation between the Polish Ministry of Environmental Protection, Natural Resources and Forestry, and the Dutch Ministry of Agriculture, Nature Management and Fisheries. Polish Ministry of Environmental Protection, Natural Resources and Forestry, Warsaw.

Warren A, Bärring L. 2003. Introduction. In *Wind Erosion on Agricultural Land in Europe*, Warren A. (ed.). EUR 20370. European Commission, Brussels; 7–12.

Wilson SJ, Cooke RU. (1980). Wind Erosion. In *Soil Erosion*, Kirkby MJ, Morgan RPC. (eds). John Wiley & Sons, Ltd, Chichester; 217–251.

Woodruff NP, Siddoway FH. (1965). A wind erosion equation. *Soil Sci. Soc. Am. Proc.* **29**: 602–609.

Zobeck TM, Onstad CA. 1987. Tillage and rainfall effects on random roughness: a review. *Soil Till. Res.* **9**: 1–20.