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DEGRADATION, REHABILITATION, AND CONSERVATION OF SOILS

Erodibility of a Model Soil in a Wide Range of Water Flow Velocities

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Abstract—It has been experimentally shown that there are two ranges of water flow velocities, at which the erodibility of a monofractional soil (of aggregates 1–2 mm) sharply differs. In the low-velocity range, the erodibility varies from 171.53 to 3.17 $m^{-2} s^2$ at an increase in the soil density from 1.2 to 1.5 g/cm^3. In the range of high velocities, it varies from 36.88 to 0.88 $m^{-2} s^2$. The simultaneous solution of equations for the two velocity ranges enables us to obtain the boundary values of the flow velocity. Above them, other erodibility values should be taken into account at calculations. The boundary velocities for the model soil are within 1.6–1.7 m/s. This is explained by the fact that at a slow flow, water removes aggregates, which have lost the contact with the main soil as a result of its peptization by water. At high-velocity water flow, aggregates are detached under the effect of hydrodynamic forces.

Keywords: erosion intensity, boundary flow velocity, monofractional soil, interaggregate bonds, wedging impact of film moisture, leached chernozem (Luvic Chernozem (Pachic))

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INTRODUCTION

Erosion intensity and soil erodibility are usually studied at water flow rates no higher than 1–1.5 m/s. Such rates are typical for slope and gully water flows [1]. There is a direct dependence of erosion intensity on the flow velocity cube in the range to 1.5 m/s. The exception is represented by low velocities, when only some part of water flow of particular pulsation velocity is capable to detach soil particles [10]. There is a sharp decrease in the erosion intensity, when the flow of only particular velocity pulsations is capable to remove soil particles [4].

It is known that adhesion forces in soil are much greater (by three orders of magnitude) than the hydraulic forces of small water flows, but they still erode soil [15, 16]. This is related to the fact that water, being a substance of the dipole structure, is capable to penetrate into the disperse soil system and to cause its peptization. When there is water on the soil surface, capillary forces, which pull particles together, disappear as soon as the soil becomes completely water-saturated. After that, the phenomena developed at the molecular level according to the theory of Deryagin–Landau–Verwey–Overbeek result in the weakening and disruption of bonds between soil particles [2].

Thus, water flow can detach particles in the top soil layer, when adhesion between aggregates and particles of the underlying layer disappears, and gravity is reduced by the hydrostatic weighing. Nevertheless, the flow velocity increases parallel to the effect of hydrodynamic forces on the soil. Therefore, at significant flow rates, a new boundary velocity may appear. When it is exceeded, the erosion processes are significantly changed.

The results of long-term researches at the Coshocton Erosion Experimental Station, United States, also show the need to study soil erosion at a wider range of velocities. According to these data, the role of rare rains in the total long-term erosion of catchment areas under fields is 60–70% [14].

The aim of this work is to assess the dynamics of erosion intensity and erodibility and to find the boundary erosion velocity of monofractional samples of chernozems of different compactness in a wide range of flow rates (0.5–6.7 m/s).

OBJECTS AND METHODS

The study object is represented by samples of the plow horizon of light-clay leached chernozem (Luvic Chernozem (Pachic)) taken in Volovo district of Tula oblast. The main physical and chemical properties and particle-size composition of the plow horizon of this chernozem are given in [8]. We used aggregates 1–2 mm obtained by dry sieving. Different weighted samples were taken from this soil fraction to obtain the required values of soil density from 1.2 to 1.5 g/cm^3 with the interval of 0.1 g/cm^3. The weighted samples were saturated with distilled water to the moisture content of 24% of the weight of air-dry soil and kept in sealed aluminum bottles for 18–20 hours. This moisture value was chosen because at these values, the ero-

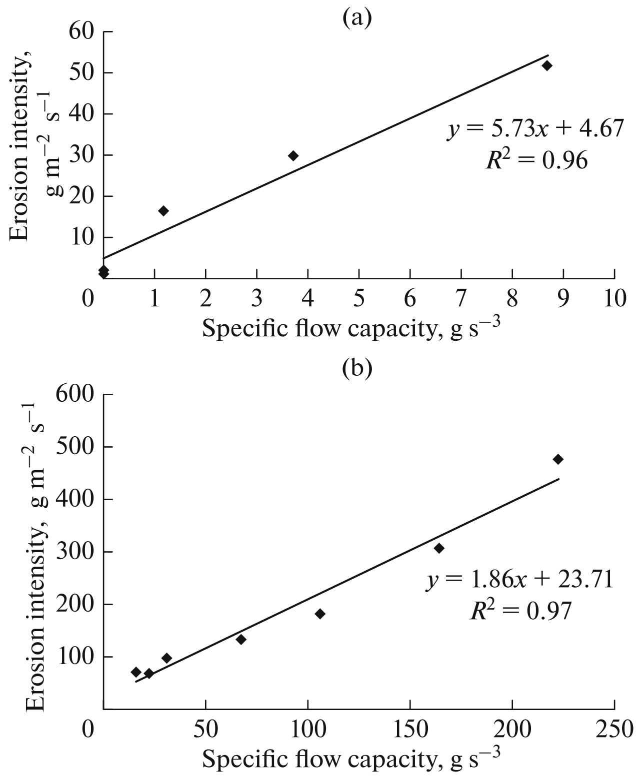

Fig. 1. The dependence of erosion intensity of soil with a density of 1.4 g cm^{-3} on the specific flow capacity in the range (a) below and (b) above the boundary velocity.

sion intensity of monofractional samples of the plow horizon of chernozem [6] is the lowest.

After that, the sample was divided into four parts, which were placed in turn into a metal cartridge (1.7 × 1.7 × 6.0 cm) with a moveable cassette. Each portion of the weighted sample was leveled and compacted to a needed density by a manual press. The method of preparing soil samples of a given density is described in detail in [7–9].

High flow velocities were obtained with the use of a water jet from a nozzle with a section of 2 × 2 cm inclined at about 2° to the surface of the soil sample. This nozzle inclination was chosen to wash the sample by only tangential forces. The soil was extracted from the cartridge by a lifting screw. The experiments were performed until the sample was completely washed out. In some variants of the experiment with the density of 1.5 g/cm^3 at low flow velocities (0.65, 0.79, and 1.06 m/s), the samples were eroded for 1 h, and then, the experiment was stopped. The remains of the soil sample were removed from the cassette, dried to a constant weight, and weighed. The amount of washed soil was determined as the difference between the initial weight of the sample and the weight of the residue. Water temperature at the experiments was 18–20°C, because it has been shown that the flow temperature affects the intensity of soil erosion [6]. The water jet velocity was changed from 0.5 to 6.7 m/s.

The erosion intensity (q, g m^{-2} s^{-1}) was calculated with the consideration of the mass of the washed soil, the erosion period, and the area of eroded surface. Soil erodibility was evaluated by the division of the erosion intensity by the specific flow capacity.

According to the hydrophysical erosion model [10], soil erodibility (k, m^{-2} s^2) is equal to:

$$k = \frac{q}{P},$$

where q is erosion intensity, g m^{-2} s^{-1} and P is specific flow capacity, g s^{-3}. The specific flow capacity is equal to the product of the shear stress on the flow velocity [15].

The erodibility may be graphically determined as the coefficient of the inclination angle on the plots of the dependence of the erosion intensity on the specific flow capacity. The methodology of the analysis of the experimental data consists in the obtaining such dependencies for each series of the experiment, including all tests at one density in the entire range of flow velocities. The experiments were performed for each soil density in the range of flow velocities from 0.5 to 6.7 m/s in 4–8 replications (about 300 samples in total).

RESULTS AND DISCUSSION

The analysis of plots of the dependence of soil erosion intensity on the specific flow capacity for all the series of the experiment has shown that there is a direct reliable correlation between these parameters. However, the values of the angular coefficients (erodibility) differ significantly for the areas below and above the boundary velocity (Fig. 1). For example, for low and high flow rates at soil density of 1.4 g cm^{-3}, the equations are, respectively:

$$q = 5.73P + 4.67, \quad (1)$$

$$q = 1.86P + 23.71, \quad (2)$$

where P is specific flow capacity.

The simultaneous solution of these equations enables to calculate the velocity, at which the erodibility significantly changes. In our case, it is equal to 1.70 m/s and should be taken as the boundary velocity (termed erosion velocity in published works). For soils, it is equal to the first tens of centimeters per second [3]. The boundary velocities for other soil densities were similarly calculated (Table 1).

The erodibility in the range below the boundary velocity is represented by the angular coefficient of the equation (the number prior to P in equation (1)), which characterizes the area of velocities lower than the boundary value. In equation (2), it is located after P. There are also two free members in the erosion equations. The free member in equation (1) is assigned to velocities below the boundary and should be used for this range. The free member in erosion equation (2) is assigned to the values above the boundary velocity of flow.

Table 1. Change in erodibility of the model soil of different density under a wide range of flow velocities

Soil density, $g\ cm^{-3}$	Boundary velocity, $m\ s^{-1}$	Erodibility, $m^{-2}\ s^2$		Free member of equations	
		below the boundary velocity	higher the boundary velocity	below the boundary velocity	above the boundary velocity
Fraction 1–2 mm					
1.2	1.58	171.53	36.88	–23.21	510.12
1.3	1.68	49.24	3.48	2.34	221.31
1.4	1.70	5.73	1.86	4.67	23.71
1.5	1.72	3.17	0.88	2.57	14.18
Fraction <1 mm					
1.4	1.03	27.05	1.58	3.48	31.84

It is interesting that the boundary velocity slightly increases from 1.58 to 1.72 m/s parallel to soil density (Table 1). When soil aggregates <1 mm are eroded, the boundary velocity is slightly smaller (1.03 m/s) in comparison with the erosion of aggregates 1–2 mm at the similar density (1.4 g/cm^3).

Soil erodibility in the range of velocities below the boundary decreases from 171.53 to 3.17 $m^{-2}\ s^2$ with the rise in soil density from 1.2 to 1.5 g/cm^3. For flow velocities above the boundary, the erodibility is smaller. In this range, it varies from 36.88 to 0.88 $m^{-2}\ s^2$. In case of erosion of soil fraction <1 mm in the range below the boundary, the free member in the equation is small similarly to the soil with large aggregates (1–2 mm): 3.48 and 4.67, respectively. At the boundary of the two velocity ranges, the change in the erosion intensity is not instant, but it sharply decreases contrary to the velocity at this interval (1.6–1.7 m/s) (for the fraction of aggregates of 1–2 mm).

The existence of two areas of the dynamics of erosion intensity may be probably explained by the following reasons. As it is known, soil mainly consists of aggregates, which are in turn composed of elementary clay particles bound by cohesion and by amphiphilic molecules of humus acids [12, 13]. Amphiphilicity is related to the capability of humus acids to bind elementary soil particles by hydrophobic and hydrophilic parts. Since hydrophobic parts of molecules are not dissolved in water, such aggregates are more water-resistant. As a result, the bonds between aggregates are weaker in comparison with bonds inside the aggregates. In addition, the bonds between aggregates appear at the final formation of sample, when its mean density is brought to the required value. When there is water on the sample surface, it penetrates into the underlying layer and first breaks bonds between aggregates.

In this case, the flow takes almost free particles on the soil surface. This is proved by our research [7]: after a pause, during which the soil is under the water layer, water flow sometimes detaches a group of aggregates of the surface layer, which are carried by the flow, maintaining bonds between them. This may be explained by the fact that the interaggregate component of bonds can be preserved between horizontally adjacent aggregates, because the wedging force of water films is mitigated by the initial compression of aggregates at the sample formation.

The deeper layer of particles pressed by the above layer is not completely saturated with water (to the status, when the interaction between aggregates is absent). Therefore, the film moisture between the units is saturated some time after the detachment of the top layer of soil particles by flow, because water diffusion is slow. Only in the case, when cohesion becomes equal to zero and the buoyancy of water reduces the gravity, water flow can take particles of the exposed lower layer.

This process is developed until the flow velocity becomes so high that the inter aggregate interaction cannot withstand the flow. The erosion process is developed to the next stage, when the erodibility is decreased more than three–four times. At this stage, the flow already continuously breaks down the adhesion between soil particles, and the erodibility becomes stable at the flow velocity to at least 6–7 m/s.

To determine the effect of the size of aggregates on the intensity of soil erosion, the experiments with the fraction <1 mm at a density of 1.4 g cm^{-3} were performed. It has been revealed that the erodibility of this soil fraction is significantly higher as compared to the fractions of 1–2 mm in the range below the boundary: 27.05 and 5.73 $m^{-2}\ s^2$ and is slightly lower in the range above the boundary velocity: 1.58 and 1.86 $m^{-2}\ s^2$, respectively. Thus, it is shown that the monofractional soil with larger aggregates is more erosion-resistant at flow velocities below the boundary as compared to soil with fine structure. This is probably related to different thickness of the top layer of aggregates.

The erodibility of soil is strongly determined by its density [5]. Well pronounced dependencies were revealed by us at processing the experimental data:

$$k_1 = 0.12(\rho_s - 1)^{-4.62}, \quad (3)$$

$$k_2 = 0.046(\rho_s - 1)^{-4.00}, \quad (4)$$

where k_1 and k_2 are soil erodibilities in the ranges below and higher the boundary velocity, $m^{-2}\ s^2$; and ρ_s is soil density, g cm^{-3}. Determination coefficients for the equations (3) and (4) exceed 0.95.

This is related to the fact that soil particles approach each other with the increase in density, and the number of points of their mutual contacts consequently increases.

CONCLUSIONS

The experiments for estimating the dynamics of soil erosion intensity at a wide range of velocities have shown that there is a particular boundary velocity, above which, the erodibility significantly changes. The boundary velocity under the experimental conditions (fraction of aggregates 1–2 mm) varies within a small range of 1.6–1.7 m/s and slightly increases parallel to soil density. However, soil erodibility at the velocity above the boundary is much lower (three times or more) as compared to the erodibility at the flow velocity below the boundary, and is described by a gentler curve.

It is difficult to explain the above results by only mechanical laws of interaction between the flow and the underlying rough surface of the aggregated soil. It is obvious that wedging force of water films on the surface of clay particles plays a great role in erosion [11]. At small flow velocities below the boundary, the top layer of aggregates is quickly saturated with water. An outer film formed around them can result in complete disappearance of gravitation between aggregates. After that, the detachment and transportation of aggregates by water flow are only slowed down by gravity minus the lifting force of water.

In the range of velocities higher than the boundary, the hydrodynamic effect of flow on soil aggregates is so great that it can disrupt bonds between the particles even prior to the saturation of interaggregate films. In this case, the flow requires relatively more energy than at velocities below the boundary. As a result, the erodibility decreases three times and more in comparison with that in the range of velocities lower than the boundary.

Some theoretical considerations and experimental data enable us to suggest that the size of soil aggregates may affect the erosion efficiency of the wedging action of water films. The comparison of erodibilities of soils with aggregates <1 mm and 1–2 mm shows that they only differ at flow velocities below the boundary: 27.05 and 5.73 $m^{-2}\,s^2$. At flow with the velocity above the boundary, the differences for these soil fractions are leveled and become 1.58 and 1.86 $m^{-2}\,s^2$, respectively. Under terrain conditions, the velocity of slope flows rarely exceeds 1.5 m/s, so soils with larger aggregates better resist erosion.

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