ISSN 1064-2293, Eurasian Soil Science, 2018, Vol. 51, No. 3, pp. 336–344. © Pleiades Publishing, Ltd., 2018.
Original Russian Text © G.A. Larionov, O.G. Bushueva, A.V. Gorobets, N.G. Dobrovolskaya, Z.P. Kiryukhina, S.F. Krasnov, L.F. Litvin, I.A. Maksimova, I.I. Sudnitsyn, 2018, published in Pochvovedenie, 2018, No. 3, pp. 347–356.

DEGRADATION, REHABILITATION, AND CONSERVATION OF SOILS

Experimental Study of Factors Affecting Soil Erodibility

G. A. Larionov*, O. G. Bushueva, A. V. Gorobets, N. G. Dobrovolskaya, Z. P. Kiryukhina, S. F. Krasnov, L. F. Litvin, I. A. Maksimova, and I. I. Sudnitsyn

Lomonosov Moscow State University, Moscow, 119991 Russia

**e-mail: larionov425@mail.ru*

Received July 27, 2016

Abstract—The effect of different factors and preparation conditions of monofraction samples from the arable horizon of leached chernozem on soil erodibility and its relationship with soil tensile strength (STS) has been studied. The exposure of samples at 38°C reduces their erodibility by two orders of magnitude. The drying of samples, on the contrary, increases their erodibility. It has been shown that erodibility decreases during the experiment. It has been found that the inoculation of soil with yeast cultures (*Naganishia albida, Lipomyces tetrasporus*) reliably increases the STS value in 1.5–1.9 times. The sterile soil is eroded more intensively than the unsterile soil: at 4.9 and 0.3 g/(m^2 s), respectively. The drying of soil followed by wetting to the initial water content (30%) has no significant effect on the STS value in almost all experimental treatments.

Keywords: soil erosion, soil loss rate, interaggregate bonds, soil density, soil tensile strength, yeast cultures (*Naganishia albida*, *Lipomyces tetrasporus*), leached chernozem (Luvic Chernozem (Pachic))

DOI: 10.1134/S1064229318030067

INTRODUCTION

Soil erodibility was initially considered as a constant function of stable soil parameters [2, 27–29, 39, 43], but there was no consensus among scientists about these parameters. Bennett [28] concluded from his observations of two soils differing in erosion rate that well-structured loose and R_2O_3-enriched soils with high water permeability are less susceptible to erosion than dense structureless soils with heterogeneous texture and low water permeability. According to Middleton [39], soil erodibility depends on a number of soil parameters: particle size distribution, content of colloids, density, water capacity, swelling, shrinkage, and plasticity. The author assigned the leading part to the dispersion factor (ratio between the contents of silt and clay in the soil determined by water dispersion and particle size analysis). Bouyoucos [29] concluded that the erosion rate is determined by the contents of the silt and clay fractions. After Bennett and Middleton, Baver [27] and Pelle [43] also thought that erodibility depends on such soil parameters as water permeability and degree of dispersion. Vilenskii [2] agreed with the conclusions of the above authors and emphasized the effect of such physical properties as the water stability of structural aggregates and the shear, crushing, and tensile strength of soil on soil erodibility.

Voznesenskii and Artsruni [3] studied the erosion of krasnozemic soils and concluded that the soil properties characterizing the dispersion behavior of soil in water—the degree of dispersion, the degree of aggregation, the content of colloids, the equivalent water content, and the silica-to-R_2O_3 ratio—are the most informative parameters for assessing the erodibility of soils. Voronin and Kuznetsov [5] studied the erosion of undisturbed light-chestnut soil samples and proposed to estimate the erosion stability from the ratio of the potential soil structureness (ratio between the total fractions ≤0.001 mm and the fraction ≥0.001 mm) to the soil dispersion factor.

The above approaches gave relative estimates of erodibility, but the development of empirical erosion models required the quantification of this parameter. At first, erodibility was determined within the empirical USLE model as the quotient of the long-term soil loss from a standard plot of 22.1 m long with a slope of 9% on a permanent fallow by the long-term erosion potential of rainfall [49]. Later on, an empirical relationship between soil erodibility and 15 (and then 24) soil parameters was developed from quantitative data on soil erodibility on runoff plots [47, 48]. For ease of use, a nomogram was created to determine erodibility from three quantitative and two qualitative parameters.

However, it soon surfaced that soil erodibility undergoes significant seasonal changes. It was found [34] that in southwestern Canada, soil erodibility reaches a maximum value in February and decreases to a minimum in August. Six-year-long field studies of soil erosion on several catchment areas confirmed the results of the above study [40]. It was revealed later on that, along with the seasonal dynamics of erodibility, the soil tensile strength (STS) and the mechanical strength of soil aggregates also change [30]. Later studies showed that

the seasonal dynamics of erodibility is typical for different geographical conditions [44–46].

The variability of soil erodibility is probably related to specific features of clay minerals, which ensure the cohesion of soils. Grissinger and Asmussen [33] studied the effect of the preceding wetting and exposure duration of wetted samples on their erosion resistance; they revealed a significant dependence of erodibility on these factors and concluded that the effects are related to water film on the surface of clay particles. Later on, Grissinger [32] studied samples with disturbed structure and showed that the erosion rate decreases with increasing density and content of clay particles and decreasing their sizes. The effect of preliminary wetting can both increase and decrease the erosion rate depending on the orientation of clay particles. When the water temperature rises, the erosion rate increases, which was confirmed by our experimental data [13].

Voronin and Kuznetsov [5] also studied the effect of the preliminary wetting and exposure duration of wetted samples. Their experiments showed the following. First, the erosion rate depends on the degree of wetting: the air-dry soil was eroded more rapidly than the wetted soil; the soils wetted to 0.75 of field water capacity showed the minimum erosion rate, and the soils of other treatments, including wetting to total water capacity, occupied intermediate positions. Second, the exposure duration of wetted samples is of great importance. When the exposure duration increased, the erosion rate decreased in 4–5 times and was stabilized only after exposure for 6–12 h.

Seasonal changes in soil erodibility can also be related to the biochemical factor, which plays an important role in the formation and strengthening of soil aggregates [9]. The effect of microorganisms on the formation and consolidation of soil structure is confirmed in a number of experimental studies [26, 37, 42]. The inoculation of soil aggregates with cultures of the yeasts *Cryptococcus albidus* and *Lipomyces tetrasporus* decelerated their degradation in streams on slopes [11], and the inhibition of microbial activity in soils by antibiotics accelerated erosion [10].

The universal seasonal variation of soil erodibility requires the prediction of this parameter for using in erosion models. Nearing et al. [41] proposed a method for predicting soil erodibility, which is based on the concept that the compaction (consolidation) of soil after the treatment occurs under the effect of effective pressure, which implies the difference between the total soil skeleton pressure, including bounded water, and the pressure of water in pores. In the absence of external load and under incomplete saturation of soil with water, the effective pressure (α'_p) in the surface soil layer is taken proportional to the soil water potential (ψ):

$$\alpha'_p = [-\chi\psi] = \max, \quad (1)$$

where χ is an empirical coefficient. The maximum effective pressure is observed when 50–58% of pores are filled with water. The maximum consolidation of soil occurs at the maximum effective pressure. According to Nearing et al. [41], "no matter how low *S* (degree of soil saturation) may go during subsequent drying, no greater value of α' (or therefore, α'_p) can be obtained" and, hence, the further consolidation of soil is impossible. However, it does nor ensure that soil changes under further drying will not affect the soil erodibility. After the soil reached the maximum effective pressure, which ensures the maximum consolidation of soil, the further drying of soil causes a volume shrinkage of soil particles, which inevitably results in the appearance of cracks between aggregates and, hence, the disturbance of bonds between them. Studies of the effect of alternate wetting–drying of soil samples showed that their shear strength decreases in this case [36]. Thus, it may be supposed that the erodibility of consolidated soil increases under further drying.

The main goal of this work was to test of whether the erodibility of soil does not remain constant after reaching the maximum consolidation but increases with drying. To test this supposition, the erodibility of model chernozem samples was determined in the water content range from 30 wt %, which corresponds to the maximum consolidation of the studied soil, to 9–11 wt % of oven-dry soil (dried at 105°C). In addition, the effect of some other factors on soil erodibility was studied, including the methods of sample preparation and the dynamics of erosion rate during the experiment, as well as the contribution of inoculation with yeast cultures (*Naganishia albida* (former *Cryptococcus albidus*) [35] and *Lipomyces tetrasporus*) and drying to the shear strength of soil, which is closely correlated with erodibility.

OBJECTS AND METHODS

The erodibility of soil samples from the plow horizon of light-clay leached chernozem (Luvic Chernozem (Pachic)) collected in the Volovo district, Tula oblast, were studied. The content of organic matter in the soil was 5.79%, and the content of physical clay was 50.9%. The total share of calcium and magnesium in the soil exchange complex was 97.2% of total exchangeable bases.

Experiments were performed with the particle-size fraction of 1–2 mm separated by dry sieving. The sample weight was calculated by multiplying the working volume of the cartridge by the preset density of the model sample (1.3 and 1.5 g/cm^3). Before the beginning of the experiments, air-dry soil samples were put in aluminum cups and wetted to 24 wt % of air-dry soil, which corresponded to 30.1% of oven-dry soil. This water content corresponded to 55% of capillary water capacity, which was experimentally determined

for this fraction by the capillary saturation method. Hereafter, the water content is expressed in weight percentage of oven-dry soil.

Soil samples were maintained in closed aluminum cups for 18–20 h for uniform wetting, thoroughly mixed, and transferred onto a parchment sheet. The cartridge with a wood insert in its bottom part covered with a parchment sheet to more easily fix the moment of total soil erosion was placed into a hand screw press. An extension 2.5 cm high (with the cross section equal to that of the cartridge) was installed onto the cartridge, and the soil sample was transferred by small portions into the cartridge, where it was smoothed and slightly compacted with a tight-fit pestle (rectangular in cross section). After the last portion of soil was packed into the cartridge and leveled, a plunger, the height of which was equal to the height of the extension, was inserted into the extension and compressed with a screw press so that the surface of the soil sample was adjusted to the level of the cartridge border. The mean soil density reached the specified value in this case (1.3 or 1.5 g/cm^3).

Several model samples were subjected to erosion without additional treatment, as was made in our earlier studies [12–14].

Then, the samples thus prepared were placed in a thermostat at 38°C and maintained until water contents of 24, 20, 15, and 10% were reached. Then, 2 h before the beginning of the experiment, soil samples were wetted again to the initial water content (30%) by the gradual addition of water into the cartridge. To control this part of the experiment, samples were placed in a sealed container to prevent the evaporation of their water and maintained in the thermostat at 38°C for 2.5 h. Sterile soil samples were prepared in a similar way.

The samples to be used for studying the dynamics of erosion rate in the course of the experiment were dried to a water content of 14–15%; 2 h before the beginning of the experiment, their water content was adjusted to the initial level (30%), as in the preceding series of experiments. The duration of erosion was changed from 5 to 40 min with intervals of 5 min. The amount of soil lost during the second and following 5-min intervals was calculated as the difference between its values obtained in the present and preceding experiments.

Soil erosion was performed in a recirculation hydraulic flume 2.5 m long and 7.1 cm wide. The bottom and walls of the flume were smooth (organic glass, aluminum) to increase the flow velocity. The cartridge with soil was installed in an opening in the lower third of the flume and secured with a screw. The soil sample surface was adjusted on the level of the flume bottom. All the experiments were performed at a water temperature of 18–20°C, a mean flow velocity of 0.98–1.03 m/s, and a flow depth of 1 cm. This is a standard flow depth for the determination of soil erodibility according to the hydrophysical erosion model [17].

The sample surface in the cartridge was maintained on the level of the flume bottom by the rotation of the lead screw throughout the experiment. If the experiment was stopped before the total erosion of the sample, the remaining soil was removed from the cartridge and dried to a constant weight. The amount of lost soil was determined as the difference between the initial weight of the sample and that of the remaining soil. The erosion rate (q, g/(m^2 s) was calculated from the formula $q = \frac{P}{tS}$, where P is the weight of the lost soil, g; t is the time of soil erosion, s; and S is the eroded surface area (soil surface area), m^2.

Soil erodibility (k, s^2/m^2) was calculated according to the hydrophysical erosion model [17]

$$k = \frac{q}{v^3 \rho_0},$$

where q is the erosion rate, g/(m^2 s); v^3 is the cubic flow velocity, m^3/s^3; and ρ_0 is the water density, g/m^3.

To remove the effect of suspended and bottom sediments on the soil erosion rate [15, 31], water from the flume passed onto the sieve (with holes of 0.25 mm in diameter) and then into the reservoir. After each experiment, the sieve was cleaned of coarse sediments. Suspended sediments settled on the filter made of nonwoven material, which was installed in the reservoir before the tube leading to the pump.

The preparation of soil samples for the determination of STS was analogous to the preparation of samples for the study of erosion. Wetted and thoroughly mixed soil was transferred to cylindrical cartridges composed of two similar perforated tubes (25 mm in length and 17.1 mm in diameter) fastened together. Perforation ensured a uniform evaporation of water throughout the cartridge length. The original soil with a water content of 30% was used as a control. Soil samples in cartridges were dried at 40–43°C to water contents of 25, 20, 15, and 10%. When the target water content was reached, the soil was wetted again to the initial state (30%). Cartridges with rewetted soil were put in aluminum cups and a plastic container and left for 18–20 h; the STS value was then determined. The detailed description of procedure and the schematic diagram of the device for the determination of STS were reported earlier [16].

The effect of soil inoculation with cultures of the yeasts *Naganishia albida* (KBP Y-4574) and *Lipomyces tetrasporus* (KBP Y-2747) on the STS was studied analogously to the effect of soil drying. These yeast species were used, because *Naganishia albida* is a euribiont species common in different natural habitats, including top soil horizons, and *Lipomyces tetrasporus* is an autochthonic soil species occurring in chernozems and chestnut soils of the steppe zone.

Table 1. Effect of soil sample preparation on erosion rate and erodibility (soil density 1.3 g/cm^3)

Experimental treatment	Soil water content, %	n	v, m/s	Erosion rate	s	Cv, %	Erodibility, s^2/m^2
				g/(m^2 s)			
Traditional sample preparation*	30.1	5	0.98	60.0	13.8	23.1	63.7
Heating**	30.1	3	1.03	0.29	0.03	8.8	0.27
Sterile soil**	30.1	5	1.02	4.91	1.05	21.5	4.63
Drying	24.3	5	1.01	1.27	0.36	28.7	1.25
	20.2	5	1.00	1.80	0.45	24.8	1.80
	14.0	5	1.03	2.70	0.45	16.8	2.62
	8.7	5	1.06	2.80	0.38	13.5	2.64

* Samples were tested immediately after preparation. ** Heating in a sealed container for 2.5 h. Here and in Table 2, v is the flow velocity; s is the standard deviation; Cv is the coefficient of variation; and n is the number of measurements.

These yeasts are characterized by the formation of large polysaccharide capsules, the size of which can exceed that of cells in 2–3 times [24]. It was showed earlier that soil yeasts from the genus *Lipomyces* can participate in soil structuration due to the increase in the water stability of soil aggregates under the impact of extracellular polysaccharide [1, 20]. The procedure of soil preparation differed in wetting with yeast suspensions (titers of $5.3(8) \times 10^8$ and 1.8×10^8 cells/mL, respectively) rather than with water. The soil samples in cartridges with the added yeast cultures and the control samples were maintained in plastic containers at 13–15°C for 5 days.

RESULTS AND DISCUSSION

Results of experiments studying the effect of soil drying on erosion rate and erodibility are given in Table 1. First of all, noteworthy is a very strong effect of soil heating without drying (in a hermetic container) to 38°C for 2.5 h. The erosion rate is 60 g/(m^2 s) for samples eroded without preliminary heating and 0.29 g/(m^2 s) for preheated soil.

This phenomenon can be explained using two approaches. One of them is related to so-called squeezed air. Water arrived into the dry soil moves in capillaries inward aggregates and can compress their air to several atmospheres, if it has no free exit [4, 23]. Squeezed air can favor the breakage of soil aggregates. This approach would explain many experimental results; however, observations over erosion processes give no grounds for taking this hypothesis as a basis. The degradation of soil aggregates because of the evacuation of squeezed air can cause the appearance of turbidity clouds with air bubbles over the surface of the studied sample. However, observations show that the detachment of soil particles is usually not accompanied by the degradation of soil aggregates and the appearance of turbidity clouds. Groups of aggregates are frequently washed out and transported in the flow without appreciable water turbidity in the flume, as shown in the photo reported earlier [14]. However, this phenomenon should not be observed at the degradation of aggregates due to the compression of squeezed air. In addition, the removed soil entrapped on the fine sieve (holes of 0.25 mm in size) has aggregates similar to those in the soil fraction of 1–2 mm used for sample preparation. This also indicates no aggregate degradation by squeezed air.

Another mechanism of breaking interaggregate bonds is due to the wedging properties of water film included between soil particles [14]. According to the physicochemical theory of effective stresses in soils, the stability and instability of dispersed systems, including soils, are determined by the ratio between the attraction and repulsion of particles [21]. The attraction forces are due to intermolecular interactions (Van der Waals forces), sorption, and hydrophobic interactions. Hydrophobic components of humic substances directed toward the interlayer space of clay mineral particles favor the aggregation of particles and the decrease in the wedging pressure of soil water, which degrades soil aggregates in the absence of hydrophobic surface [19, 25].

The repulsion forces are due to the electrostatic interaction between the diffuse parts of double electric layers of adjacent particles. The increase of ion concentration in the space between particles due to the overlap of ionic atmospheres creates a local osmotic pressure, under the effect of which the liquid phase tends to fill the space and move the particles apart. This generates the electrostatic component of the wedging pressure [21].

Dispersed systems generally occur in thermodynamic equilibrium; however, in our case, the upper layer of the eroded soil sample experiences no external pressure except the own weight. The water film of diffuse genesis favors the separation of aggregates in the external layer of soil sample to a distance exceeding the action radius of intermolecular forces (Van der

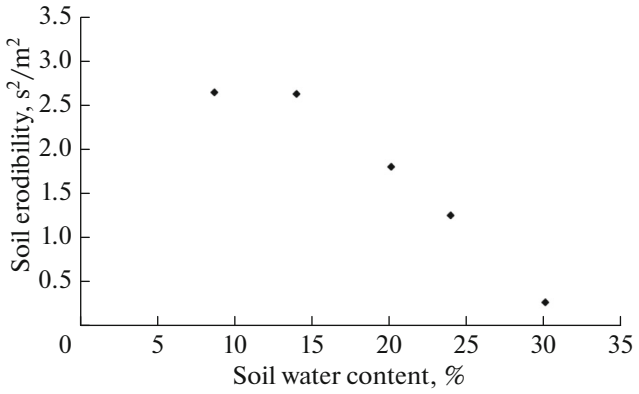

Fig. 1. Changes in soil erodibility under drying.

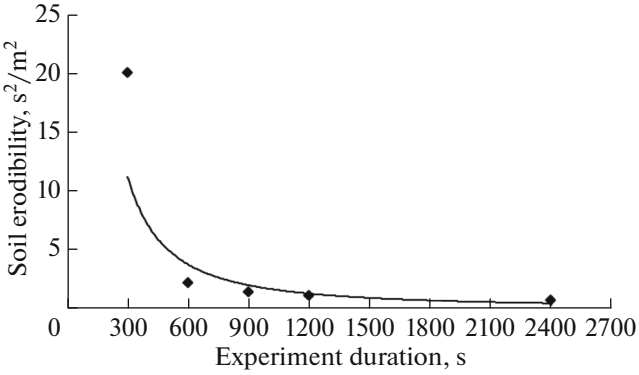

Fig. 2. Effect of experiment duration on soil erodibility.

Waals forces). Therefore, aggregates in the upper layer loose cohesion with the lower layer and are easily entrapped by the water flow as free particles.

According to the concepts of the stable thermodynamic state of clayey soils, the soil samples used in our experiments were mixtures of closely sized aggregates (1–2 mm) without any bonds between them. In the course of sample preparation, the mixture of aggregates was wetted and left for 18–20 h. During this time, bonds formed between aggregates, which broke again during the preparation of model soil samples, but new and stronger contacts formed between particles thereby. However, thermodynamic equilibrium corresponding to the maximum stability of samples could not establish during the short time period from sample preparation to tests, because the movement of water in the soil, primarily as diffusion and/or osmosis, is a relatively slow process. Therefore, the breaking rate of incompletely formed bonds between aggregates under the impact of free water and, hence, the erosion rate of samples was high. The exposure of samples at 38°C without drying in a thermostat for 2.5 h ensured a decrease of erodibility in more than 200 times (Table 1), which indicates the formation of bonds between particles and the acquisition of thermodynamic equilibrium by them [7, 8, 22].

As for the drying of model samples below the water content corresponding to the maximum consolidation of soil, it is noted above that the entire soil sample undergoes shrinkage, as well as separate aggregates, which results in the formation of cracks between them. In this case, the distances between adjacent soil particles can increase to the point that they exceed the action radius of intermolecular forces (Van der Waals forces). It is also obvious that when the drying penetrates deeper into the soil, soil shrinkage is enhanced, and the number of particles, the distance between which exceeds the action radius of the Van der Waals forces, increases respectively. A linear relationship between soil erodibility and water content is noted in the range from 30 to 14% (Fig. 1). However, when the soil samples are dried further, down to 8.7%, erodibility does not change and remains on the level of 2.64 s^2/m^2. This suggests that far from all interaggregate bonds in the consolidated soil are broken under drying to this water content. If the amount of stable interaggregate bonds in the soil samples with the initial water content (30%) exposed at 38°C is taken equal to 100%, the erodibility change indicates that about 10% of them remain at a water content of 8.7–14.0%. The partial retention of bonds on drying of soil samples can be explained as follows. The drying of soil is accompanied by a shrinkage of aggregates, which results in the formation of cracks. The size of cracks and, hence, the distances between aggregates are different and obey a normal distribution according to the central limit theorem [6]; therefore, some cracks between aggregates can fall within the range of intermolecular interactions. The amount of such cracks between aggregates due to soil drying probably depends on soil density. This hypothesis should obviously be tested experimentally.

Thus, contrary to the opinion of Nearing et al. [41], it was shown that when the maximum consolidation of soil is reached, the drying of soil below 30% water does result in a significant increase in erodibility. Nonetheless, results show that the erodibility of soil dried to a water content of 14–15%, which ensures the maximum erosion rate of soil consolidated and then adjusted to the initial water content (30%), also varies during the experiment from 20.0 to 0.66 s^2/m^2 (Fig. 2).

The erodibility of cultivated soils undergoes permanent changes in time. The most radical change in erodibility is due to basic cultivation, when the soil crumbles and loosens, which results in the degradation of soil aggregates [38]. With time, the erodibility of soil reaches the minimum values due to mechanical shrinkage and consolidation under wetting to total water capacity, but it increases again when the soil dries. These changes in erodibility are typical for relatively long time periods.

Table 2. Effect of soil inoculation with yeast cultures (*Naganishia albida* and *Lipomyces tetrasporus*) on the soil tensile strength (STS)

Soil density, g/cm^3	Experimental treatment	*n*	STS	*s*	*Cv*, %
			kPa		
1.3	Natural soil	4	6.59 ± 0.80	0.83	12.6
	Sterile soil	6	5.20 ± 0.65	0.53	10.2
	N. albida	4	7.73* ± 0.80	0.64	8.3
	L. tetrasporus	5	9.89* ± 0.71	0.97	9.8
1.5	Natural soil	6	10.5* ± 1.10	1.20	11.5
	Sterile soil	5	8.10 ± 1.20	1.51	18.7
	N. albida	5	13.4* ± 1.20	0.73	5.5
	L. tetrasporus	5	15.1* ± 1.20	1.51	10.0

* Reliable difference ($P = 0.95$) from the treatment with sterile soil.

However, the erodibility of soil varies abruptly during the interaction of water streams with the soil (i.e., during erosion). So, erodibility decreased during the second and third 5-min intervals of the experiment in 9 and 15 times, respectively, compared to the first interval (Fig. 2).

Consequently, the erodibility of soil varies not only during large time intervals, but also during relatively short periods of showers. This should be obviously taken into consideration in the development of physically based models of soil erosion.

The soil, as a bio-abiotic system, undergoes permanent changes in time under the effect of not only the anthropogenic factor and meteorological conditions, but also the biogenic impact. The growth and development of field crops strongly affect the erosion resistance of soil due to the arming effect of root systems. This aspect of the effect of crops on the erodibility of soil is usually considered within the erosion models, while the estimation of microbial contribution remains insufficiently substantiated.

In this context, of interest is the study of the effect of microbial activity on the STS, which is an important characteristic for erosion estimation. The physicochemical relationships between soil aggregates and particles include biochemical components: microorganisms and products of their metabolism play a peculiar role in the formation of aggregates [9]. The effect of microorganisms on the formation and strengthening of soil structure was confirmed in many studies and experiments [26, 37, 42]. Results of experiments on the abrasion of aggregates during their transport by sloped water flows indicate a significant effect of bacteria and yeasts on the mechanical strengthening of soil aggregates [11]. The inoculation of soil aggregates with yeast cultures (*Naganishia albida, Lipomyces tetrasporus*) decreased the loss of aggregates in 1.5–2 times compared to sterile soils. An efficient loosening of interaggregate bonds due to the inhibition of mycelial fungi and actinomycetes by an antibiotic was revealed earlier in experiments on the erosion of light-chestnut soils in a hydraulic flume [10].

To assess the effect of microorganisms on STS, i.e., on the strength of intra- and interaggregate bonds, the yeasts *Naganishia albida* and *Lipomyces tetrasporus*, which dominate in this soil type, were used. In the course of their vital activity, yeasts are capable to synthesize extracellular polysaccharides and thus create a viscous environment with high hygroscopicity, as well as to form acid polysaccharides, which stimulates aggregation and increases the water stability of aggregates. Sterile samples, soil samples in the natural state, and samples inoculated and incubated at 13–15°C for 5 days were used in our experiments (Table 2).

Experimental results point to a significant effect of the studied yeasts on the STS, which increases in 1.8–1.9 times for the samples inoculated with *L. tetrasporus* and in 1.5–1.7 times at the addition of *N. albida*, the soil density being 1.3 and 1.5 g/cm^3, respectively. The difference between the mean STS values in all experimental series are statistically significant for confidence levels no less than 0.90.

Thus, the experiments with samples held for 5 days after inoculation showed that sterile soils had the least STS values. The samples inoculated with the yeasts *N. albida* and *L. tetrasporus* had the maximum STS values, and natural (nonsterile) soils occupied intermediate positions. Experiments with sterile soil confirmed the unidirectional effects of microorganisms on STS and erosion resistance. The sterile soil samples were eroded an order of magnitude more rapidly (4.91 g/(m^2 s) than the nonsterile control samples (0.29 g/(m^2 s) (Table 1).

We earlier revealed a close correlation between soil erodibility and STS [18]. Data on the estimation of the effect of soil drying followed by wetting to the initial water content (30%) on STS are given in Table 3. The STS value little depended on soil water content in the

Table 3. Effect of drying of soil with a density of 1.3 g/cm3 on the soil tensile strength (kPa)

Statistical parameters	Soil water content, %				
	30	25	20	15	10
Mean (*M*)	5.35	8.75*	5.96	6.11	5.56
Median (*Med*)	5.42	8.73	5.33	6.27	5.89
Standard deviation (*s*)	0.75	0.64	1.33	1.48	1.16
Maximum (max)	6.72	10.1	8.43	8.38	7.46
Minimum (min)	3.96	7.37	4.49	4.21	3.75
Coefficient of variation (*Cv*, %)	13.9	7.3	22.4	24.2	20.8
Number of measurements (*n*)	13	12	13	12	12

* Reliable difference ($P = 0.95$) from the treatment with the initial water content (30%).

course of drying. Only in the experimental treatment with a water content of 25%, STS was reliably higher in 1.6 times ($P = 0.95$) than in the soil with the initial water content (30%), which calls for further experimental studies.

CONCLUSIONS

It is shown that the erodibility of the plow horizon of chernozem is in permanent dynamics mainly due to changes in the stability of dispersed systems, including soils enriched with clay minerals.

The maximum consolidation of soil is reached at the filling of 50–58% of pores with water [41]; this corresponds to the minimum erodibility, which is observed at a water content of 30% for the studied soil. Under drying, the soil begins to crack; therefore, the amount and strength of interaggregate bonds decreases, and the erodibility of soil increases. This phenomenon is traced in the range of soil water content from 30 to 9%. A linear relationship between erodibility and water content is observed in the range from 30 to 14% water. In this case, a short-term wetting of soil to the initial water content (30%) at the beginning of the experiment does not restore the initial erodibility values. According to Voronin and Kuznetsov [5], an appreciable decrease in erodibility is observed 5–6 days after the wetting of dry soil.

It is found that the erodibility of soil significantly varies depending on the duration of water impact, which obviously should be taken into consideration in the development of rainfall erosion models.

The inoculation of soil with yeast cultures (*Naganishia albida* and *Lipomyces tetrasporus*) reliably increases the STS value in 1.5–1.9 times.

It is shown that the drying of soil followed by wetting to the initial water content (30%) had no reliable effect on the STS value in almost all experimental treatments. This could be related to the fact that bonds between aggregates were broken and aggregates themselves were not touched in these experiments.

ACKNOWLEDGMENTS

The work was performed within the framework of the research plan of the Makkaveev Research Laboratory of Soil Erosion and Channel Processes and was supported in part by the Russian Foundation for Basic Research (project no. 16-05-00474a) and the microbiological part of the work was supported by the Russian Science Foundation (project no. 14-50-00029).

REFERENCES

1. I. P. Bab'eva and Kh. Moavad, "Soil yeast of the genus *Lipomyces* as agents of soil structuring," Pochvovedenie, No. 8, 144–146 (1973).
2. D. G. Vilenskii, "Properties of soils determining their erodibility and their analysis methods," in *Soil Erosion Control in the Soviet Union* (Academy of Sciences of the Soviet Union, Moscow, 1938), pp. 111–130.
3. A. S. Voznesenskii and A. B. Artsruni, "Effect of physical-chemical properties of soils on surface outflow as the factor of surface wash-out," in *Soil Erosion Control in the Soviet Union* (Academy of Sciences of the Soviet Union, Moscow, 1938), pp. 131–154.
4. A. D. Voronin and V. G. Vityazev, "Characteristics of water energy in major soil types of the European part of the Soviet Union," in *Life of the Earth*, Sb. Tr. Muz. Zemleved. (Moscow State Univ., Moscow, 1974), No. 10, pp. 146–152.
5. A. D. Voronin and M. S. Kuznetsov, "Evaluation soil loss tolerance," in *Soil Erosion and Channel Processes* (Moscow State Univ., Moscow, 1970), No. 1, pp. 99–115
6. V. E. Gmurman, *The Theory of Probability and Mathematical Statistics* (Vysshaya Shkola, Moscow, 1998) [in Russian].
7. B. V. Deryagin, *Theory of Stability of Colloids and Thin Films* (Nauka, Moscow, 1986) [in Russian].
8. B. V. Deryagin and N. V. Churaev, *Wetting Films* (Nauka, Moscow, 1984) [in Russian].

9. D. G. Zvyagintsev, *Interaction of Microorganisms with Solid Surfaces* (Moscow State Univ., Moscow, 1973) [in Russian].
10. M. S. Kuznetsov, "Influence of preliminary moistening on soil loss tolerance of light chestnut soils in Yergeni as based on the study of undisturbed soil samples," Vestn. Mosk. Univ., Ser. 6: Biol., Pochvoved., No. 3, 91–99 (1967).
11. G. A. Larionov, O. G. Bushueva, N. G. Dobrovol'skaya, Z. P. Kiryukhina, S. F. Krasnov, and L. F. Litvin, "Effect of the water temperature and soil moisture on the erodibility of chernozem samples: a model experiment," Eurasian Soil Sci. **47**, 734–739 (2014).
12. G. A. Larionov, O. G. Bushueva, N. G. Dobrovol'skaya, Z. P. Kiryukhina, L. F. Litvin, and I. A. Maksimova, "Destruction of soil aggregates in slope flows," Eurasian Soil Sci. **40** (10), 1128–1134 (2007).
13. G. A. Larionov, O. G. Bushueva, N. G. Dobrovol'skaya, Z. P. Kiryukhina, and L. F. Litvin, "Erodibility of model soils with different densities," Eurasian Soil Sci. **44**, 914–918 (2011).
14. G. A. Larionov, O. G. Bushueva, N. G. Dobrovol'skaya, Z. P. Kiryukhina, L. F. Litvin, and S. F. Krasnov, "Assessing the contribution of nonhydraulic forces to the destruction of bonds between soil particles during water erosion," Eurasian Soil Sci. **49**, 546–550 (2016). doi 10.1134/S1064229316050100
15. G. A. Larionov, N. G. Dobrovol'skaya, Z. P. Kiryukhina, and L. F. Litvin, "Effect of suspended sediments on soil erodibility," Eurasian Soil Sci. **41**, 768–773 (2008).
16. G. A. Larionov, N. G. Dobrovol'skaya, Z. P. Kiryukhina, S. F. Krasnov, L. F. Litvin, A. V. Gorobets, and I. I. Sudnitsyn, "Effect of soil density, tensile strength, and water infiltration on the rupture rate of interaggregate bonds," Eurasian Soil Sci. **50**, 335–340 (2017). doi 10.1134/S1064229317010094
17. G. A. Larionov and S. F. Krasnov, "Probabilistic model of soil and cohesive ground erosion," Eurasian Soil Sci. **33**, 205–211 (2000).
18. G. A. Larionov, L. F. Litvin, N. G. Dobrovol'skaya, Z. P. Kiryukhina, S. F. Krasnov, A. V. Gorobets, and O. G. Bushueva, "Theoretical and experimental studies of surface erosion," in *Soil Erosion and Channel Processes* (Moscow State Univ., Moscow, 2015), No. 19, pp. 35–57.
19. E. Yu. Milanovskii, *Humic Substances of Soils as the Natural Hydrophobic-Hydrophilic Compounds* (GEOS, Moscow, 2009) [in Russian].
20. Kh. Moavad, I. P. Bab'eva, and S. E. Gorin, "Aggregation of soils affected by extracellular polysaccharide *Lipomyces lipofer*," Pochvovedenie, No. 9, 65–68 (1976).
21. V. I. Osipov, "Physicochemical theory of effective stresses in soils," Gruntovedenie, No. 2, 3–34 (2013).
22. V. D. Perevertaev and M. S. Metsik, "Adsorption of water vapor on the surface of mica crystals," Kolloid. Zh. **28** (2), 254–259 (1966).
23. I. I. Sudnitsyn, *Movement of Soil Moisture and Water Consumption by the Plants* (Moscow State Univ., Moscow, 1979) [in Russian].
24. I. Yu. Chernov, *Yeast in Nature* (KMK, Moscow, 2013) [in Russian].
25. E. V. Shein and E. Yu. Milanovskii, "The role of organic matter in the formation and stability of soil aggregates," Eurasian Soil Sci. **36**, 51–58 (2003).
26. I. P. Bab'eva and I. Yu. Chernov, "Geographical aspects of yeast ecology," Physiol. Gen. Biol. Rev. **9**, 1–54 (1995).
27. L. D. Baver, "Some factors effecting erosion," Agric. Eng. **14**, 51–52 (1933).
28. H. H. Bennett, "Some comparisons of the properties of humid-tropical and humid-temperate American soils, with special reference to indicated relations between chemical composition and physical properties," Soil Sci. **21**, 349–375 (1926).
29. G. J. Bouyoucos, "The clay ratio as a criterion of susceptibility of soils to erosion," J. Am. Soc. Agron. **27**, 738–741 (1935).
30. D. R. Coote, C. A. Malcolm-McGovern, G. J. Wall, W. T. Dickinson, and R. P. Rudra, "Seasonal variation of erodibility indices based on shear strength and aggregate stability in some Ontario soils," Can. J. Soil Sci. **68**, 405–416 (1988).
31. G. R. Foster, "Modeling the erosion process," in *Hydrologic Modeling of Small Watersheds*, Ed. by C. T. Haan, H. P. Johnson, and D. L. Brakenslek (American Society of Agricultural Engineers, St. Joseph, MI, 1982), No. 5, pp. 297–380.
32. E. H. Grissinger, "Resistance of selected clay systems to erosion by water," Water Resour. Res. **2** (1), 131–138 (1966).
33. E. H. Grissinger and L. E. Asmussen, "Channel stability in undisturbed cohesive soils," J. Hydraul. Div., Am. Soc. Civ. Eng. **89** (6), 259–264 (1963).
34. P. C. Kirkby and G. R. Mehuys, "The seasonal variation of soil erosion by winter in South-Western Quebec," Can. J. Soil Sci. **67**, 55–63 (1987).
35. X.-Z. Liu, Q.-M. Wang, M. Goker, M. Groenewald, A. V. Kachalkin, H. T. Lumbsch, A. M. Millanes, M. Wedin, A. M. Yurkov, T. Boekhout, and F.-Y. Bai, "Towards an integrated phylogenetic classification of the Tremellomycetes," Stud. Mycol. **81**, 85–147 (2015).
36. R. Ma, C. Cai, Z. Li, J. Wang, T. Xiao, G. Peng, and W. Yang, "Evaluation of soil aggregate microstructure and stability under wetting and drying cycles in two ultisols using synchrotron-based X-ray micro-computed tomography," Soil Tillage Res. **149**, 1–11 (2015).
37. J. P. Martin, J. O. Ervin, and R. A. Shepherd, "Decomposition and aggregating effect of fungus cell material in soil," Soil Sci. Soc. Am. Proc. **23**, 217–220 (1959).
38. *Stream Channel Stability*, Appendix G: *Soil Erosion and Sediment Characteristics of Typical Soils and Land Uses in the Goodwin Creek Catchment, Project Objectives 3 and 4*, Ed. by L. D. Meyer and W. C. Harmon (USDA Sedimentation Laboratory, Oxford, 1981), pp. G2–G64.
39. H. E. Middleton, "Properties of soils, which influence soil erosion," USDA Tech. Bull., No. 178, (1930).
40. C. K. Mutchler and C. E. Carter, "Soil erodibility variation during the year," Trans. ASAE **26**, 1102–1104 (1983).

41. M. A. Nearing, L. T. West, and L. C. Brown, "A consolidation model for estimating changes in rill erodibility," Trans. ASAE **31** (3), 696–700 (1988).

42. E.-J. Park, W. J. Sul, and A. J. Smucker, "Glucose additions to aggregates subjected to drying/wetting cycles promote carbon sequestration and aggregate stability," Soil Biol. Biochem. **39** (11), 2758–2768 (2007).

43. T. C. Pelle, "The relation of certain physical characteristics to the erodibility of soils," Soil Sci. Soc. Am. Proc. **2**, 97–100 (1937).

44. J. Rejman, R. Turski, and J. Paluszek, "Spatial and temporal variations in erodibility of loess soil," Soil Tillage Res. **46**, 61–68 (1998).

45. M. P. S. Sanchis, D. Torri, L. Borselli, J. Poesen, "Climate effects on soil erodibility," Earth Surface Processes and Landforms. **33** (7), 1082–1097 (2008).

46. D. Torri, J. Poesen, L. Borselli, "Predictability and uncertainty of the soil erodibility factor using a global dataset," Catena, **31**, 1–22 (1997).

47. W. H. Wischmeier, C. B. Johnson, B. V. Cross, "A soil erodibility nomograph for farmland and construction sites," Jour. Soil and Water Conservation **26**, 189–193 (1971).

48. W. H. Wischmeier, J. V. Mannering, "Relation of soil properties to its erodibility," Soil Sci. Soc. Amer. Jour. **33**, 131–137 (1969).

49. W. H. Wischmeier, D. D. Smith, "Predicting rainfall erosion losses from cropland east of the Rocky Mountains: Guide for selection of practices for soil and water conservation," Washington, USDA Agriculture Handbook, No. 282, (1965).

Translated by K. Pankratova