FUNDAMENTOS TEÓRICOS PARA A DETERMINAÇÃO DO POTENCIAL DE EROÇÃO DO SOLO (CONCEITO DE ENERGIA) PARTE 1

THEORETICAL FUNDAMENTALS FOR DETERMINING SOIL EROSION POTENTIAL (ENERGY CONCEPT) PART 1

ТЕОРЕТИЧЕСКИЕ ПРЕДПОСЫЛКИ ОПРЕДЕЛЕНИЯ ПОТЕНЦИАЛА ЭРОЗИОННОЙ СТО- ЙКОСТИ ПОЧВОВ ГРУНТОВ (ЭНЕРГЕТИЧЕСКАЯ КОНЦЕПЦИЯ) ЧАСТЬ 1

Ivan I. Maksimov¹, Nail R. Adigamov², Anas A. Mustafin², Damir T. Khaliullin², Ilmar N. Gayaziev^{2*}, Alexandr V. Matyashin², Ruslan R. Lukmanov²

¹Federal State Budget Educational Institution of Higher Education "Chuvash State Agricultural Academy", ul. Karla Marksa 29, Cheboksary, 428003, Chuvash Republic, Russia

²Kazan State Agrarian University, ul. Karla Marksa 65, Kazan, 420015, Republic of Tatarstan, Russia

* Corresponding author
e-mail: gaze1.81@mail.ru

Received 12 June 2018; received in revised form 30 November 2018; accepted 14 December 2018

RESUMO

A solução de problemas de tecnologia anti-erosão e projeto de maquinário está associada à descoberta de um parâmetro objetivo integral que atenda aos requisitos pirateados e desenvolva um método de determinação aceitável. Portanto, o objetivo da pesquisa é desenvolver um critério integral para determinar o potencial de erosão do solo (PES). Como um valor para caracterizar PES, é proposto usar energia ΔA gasta para erodir e remover o peso unitário de uma massa de solo Δm sob condições específicas de sua ocorrência. Consequentemente, o desenvolvimento de pré-requisitos teóricos para o problema em questão, bem como a determinação experimental de ΔA e Δm , podem revelar as principais regularidades no mecanismo físico dos processos de erosão. Propomos usar o método do potencial termodinâmico para descrever os processos de erosão do solo. Tem sido demonstrado que o potencial termodinâmico de Gibbs é o mais adequado para este propósito, uma vez que é possível projetar dispositivos para a medição experimental de variáveis independentes. O PES é um valor aditivo e constante para um determinado solo e expressa as suas propriedades, isto é, pode ser adotado como um critério integral único para avaliar a resistência dos solos à erosão.

Palavras-chave: solo, fluxo de água, erosão, método do potencial termodinâmico.

ABSTRACT

Solving problems of anti-erosion technology and machinery design is associated with finding an integral objective parameter that would meet piratical requirements and develop an acceptable determination method. Therefore, the research objective is to develop an integral criterion to determine soil erosion potential (SEP). As a value to characterize SEP it is prosed to use energy ΔA expended to erode and remove the unit weight of a soil mass Δm under specific conditions of its occurrence. Consequently, the development of theoretical prerequisites for the problem under consideration, as well as the experimental determination of ΔA and Δm , can reveal the main regularities in the physical mechanism of erosion processes. We propose to use the thermodynamic potential method to describe soil erosion processes. It has been shown that the Gibbs thermodynamic potential is the most suitable for this purpose since it is possible to design devices for the experimental measurement of independent variables. SEP is an additive and constant value for a given soil and expresses its properties, i.e. it can be adopted as a single integral criterion for assessing the resistance of soils to erosion.

Keywords: soil, water flow, erosion, thermodynamic potential method.

АННОТАЦИЯ

Решение задач при проектировании противозерозионных технологий и технических средств связано с отысканием интегрального объективного параметра, который удовлетворял бы требованиям практики и разработкой приемлемого способа его определения. Поэтому цель исследования - разработка интегрального критерия оценки эрозионной стойкости почвогрунтов. В качестве величины, характеризующей эрозионную стойкость почв, предлагается использовать энергию ΔA , затраченную на разрушение и вынос единицы массы Δm почвы в конкретных условиях ее залегания. В такой постановке вопрос о взаимодействии водного потока с почвогрунтом будет сводиться к определению или к измерению величин ΔA и Δm . Следовательно, разработка теоретических предпосылок решаемой проблемы и экспериментальное определение величин ΔA и Δm , могут выявить основные закономерности в физическом механизме эрозионных процессов. Для описания эрозионных процессов предложено использовать метод термодинамических потенциалов. Показано, что термодинамический потенциал Гиббса наиболее подходит для этих целей, поскольку можно разработать устройства для экспериментального измерения независимых переменных. Потенциал эрозионной стойкости представляет собой величину аддитивную и постоянную для конкретного почвогрунта и выражает его свойства, т.е. может быть принят в качестве единого интегрального критерия оценки эрозионной стойкости почв.

Ключевые слова: почва, водный поток, эрозия, метод термодинамических потенциалов.

INTRODUCTION

Methods for predicting soil losses due to erosion have been developed for many years and considered as a consequence of erosion processes [1, 2, 3, 4, 5, 6, 7]. Early calculation models of soil losses were mainly of a qualitative nature, demonstrating the different effectiveness of specific agronomic methods of erosion control [2, 3, 4, 5, 7, 23, 24, 25, 26, 27]. Initially, to describe soil loss were obtained equations with a single independent variable. Such single-factor equations were used for particular situations with practical invariance of the remaining factors. As a result of the accumulation of more extensive experimental materials and further knowledge of the mechanism of erosion processes, multi-factor equations [8, 9, 10, 11] were obtained. A fairly large number of computational models of sheet wash erosion are presently known; the detailed information on this issue can be found in the monographs [23], [45], etc. All existing calculation methods can be conditionally divided into empirical [11, 22, 23, 27, 28, 29, 30, 31, etc.], hydromechanical [8, 9, 10, 11, 32, 33], and logical-mathematical (imitational) modelling [34, 35, 36, 37, 38, 39, 40, 41, 42].

Summarizing the above brief overview of methods for calculating and predicting erosion processes, it can be noted that empirical erosion models for solving prediction problems are applicable only for specific catchment and drainage areas and can not be interpreted for other soil-climatic conditions. Application of

hydromechanical and logical-mathematical models for the same purposes is predetermined by their sensitivity to erosion properties of soil, such as K factor of soil erosion susceptibility, which is determined experimentally; non-eroding flow velocity v_{ne} or permissible bottom velocity V_{Δ} per; soil aggregate stability used in all foreign simulation models.

Research objective. Development of an integral criterion to assess the resistance of soil to erosion.

MATERIALS AND METHODS

The soil is a natural body formed by the accumulation of organic and inorganic material on the Earth's surface as a result of the mass and energy exchange (physical, chemical, biological, mechanical, etc.) processes, being an object of thermodynamics. Processes of soil formation, erosion, accumulation of sediments, dehumidification, biomass harvesting, etc., showing receipt or loss of material or energy indicate that soil can be considered as an open thermodynamic system.

The above processes proceed at a finite rate, changes in the state of a system are unbalanced and non-static. Therefore, these processes should be described by methods of non-linear non-equilibrium thermodynamics [43]. However, the considered soil medium, is self-regulating, being an agroecosystem, and like any other biological object, it aims to:

- maintain a dynamic equilibrium or a stable state

in which the rate of material and energy inputs is equal to the rate of their loss, and accumulated energy and matter remain relatively constant;

- seeks to achieve a new dynamic equilibrium, when the rate of material and energy inputs or losses of the system vary; accumulation of energy and matter increases along with an increase in the rate of the energy and material flow of the system; the higher the accumulation (loss) capacity within the system at a given input, the lower (higher) the sensitivity of the system.

Such processes in thermodynamics are usually called equilibrium and quasi-static [44]. Therefore, very tentatively, it is possible to describe erosion processes using classical laws of equilibrium thermodynamics.

To research highly complex and multifactorial erosion processes, it is proposed to use the thermodynamic potential method. The essence of this method is described below. Using the basic equation of thermodynamics for the system under consideration, some state functions are introduced – thermodynamic potentials. An increase/decrease of thermodynamic potentials seen while changing the state of the considered system is a total differential, which makes it possible to compose corresponding (thermal, caloric) equations for the process in question. Solution and analysis of the results is the most correct research approach in the context of physics, since the method of thermodynamic potentials derives from more general ideas about the object and, what is more important, this method makes it possible to experimentally measure a number of parameters included in these characteristic functions.

As a quantity characterizing SEP, we [42,45] propose to use the energy input for the destruction and removal of the unit weight of a soil sample.

$$\psi = \frac{\Delta A}{\Delta m}, \quad (\text{Eq. 1})$$

where ΔA is the energy to erode and remove the unit weight of soil sample Δm .

In this formulation, the question of water flow and soil interaction will be reduced to calculations or measurements of ΔA and Δm . Consequently, development of theoretical fundamentals for the problem under consideration, as well as the experimental determination of values included in (Eq. 1), can reveal the main regularities in the physical mechanism of erosion processes and establish the most significant links in this mechanism.

In view of such a presentation of the problem, it is necessary to consider the interaction of water flow with thawed soil with and without consideration for infiltration and frozen soil, which take place under different conditions; and to establish a position and relationship of SEP ψ among other thermodynamic potentials. Below are the theoretical fundamentals for the formulated problems.

RESULTS

1) Conditions for the erosion process in thawed soils without allowance for infiltration

Erosion processes in thawed soils can be considered independently of infiltration if: eroded soil is almost waterproof; the rate of destruction of soil layer by water flow considerably exceeds the velocity of waterfront infiltrated into soil; the rate of destruction of soil layer by water flow is equal to the velocity of the front of water infiltrated into soil.

Assume that dh_s are the height of soil eroded by water flow over time dt , and $dh_{w.inf}$ is the velocity of waterfront infiltrated into the soil over the same time, then the rate of destruction of soil layer is dh_s / dt , and the velocity of the waterfront is $dh_{w.inf} / dt$.

Eroded soil is waterproof if $dh_{w.inf} = 0$. Waterproof soils in real conditions are almost not found. For example, the approximate values of the filtration coefficients for clay soils are less than 0.00001 cm/s, for loams – from 0.00001 to 0.0001 cm/s [46]. Therefore, $dh_{w.inf} / dt \geq 0$.

If the erosion rate of the soil layer by water flow is greater or at least equal to the velocity of the waterfront, then

$$\frac{dh_n}{dt} \geq \frac{dh_{un}}{dt},$$

or

$$dh_n \geq dh_{un}. \quad (\text{Eq. 2})$$

Condition (2) shows that there is no exchange of moisture between the soil and the water flow, that is, a change in the energy potential of a system due to the intake of infiltrated water or salts into the soil can be neglected

$$\sum_{i=1}^e \mu_i dn_i \approx 0, \quad (\text{Eq. 3})$$

where n_i is the amount of the i -th

component, expressed in molar units; dn_i is the chemical potential characterizing an amount of change in the free energy of the soil, caused by a change in the content of the i -th component, provided that the rest of the components remain unchanged.

By expressing dh_s and $dh_{w.inf}$ as elementary volumes $dV_s = dh_s dB dl$ and $dV_{w.inf} = dh_{w.inf} dB dl$, dry unit weight of soil ρ_s and unit weight of water $\rho_w f(h_{w.inf}, t)$ and the elementary masses

$$dm_s = \rho_s dV_s = \rho_s dh_s dB dl, \\ dm_{w.inf} = \rho_s dV_{w.inf} + \rho_w dV_{w.inf} = \rho_w + \rho_s dh_{w.inf} dB dl,$$

it is obtained Equation 4:

$$\frac{dm_s}{\rho_s} \geq \frac{dm_{inf}}{\rho_n + f(h_{inf}, t)}, \quad (\text{Eq. 4})$$

where $\rho_w = f(h_{w.inf}, t)$ is the distribution function of water density in the soil layer over a period of infiltration time t ; dh_s ; dB and dl are the linear dimensions of the elementary volume; $\frac{dm_s}{\rho_s}$ is the elementary volume of soil with an initial volumetric moisture content of θ_0 ; $\frac{dm_{inf}}{\rho_s + f(h_{inf}, t)}$ is the elementary volume of soil with a volumetric moisture content of θ_i due to intake of water absorbed into the soil over a period of time dt .

It's obvious that

$$\frac{dm_{inf}}{\rho_s + f(h_{inf}, t)} - \frac{dm_s}{\rho_s} = \Delta V_w \leq 0, \quad (\text{Eq. 5})$$

is the volume of moisture absorbed over a period of time dt to the depth $dh_{w.inf}$, which confirms the validity of (3).

2) The basic thermodynamic equation for eroded thawed soil without allowance for infiltration

Interaction of water flow with thawed soil, without allowance for infiltration, is accompanied by a whole complex of interrelated processes, in accordance with (4) or (5). Assume that all the parameters of these processes vary infinitely slowly and the erosion process has stabilized, i.e. a quasi-static process is being realized. Therefore, the basic equation of

thermodynamics for the process under consideration can be written in the following form:

$$TdS = dU + \sum_{i=1}^n B_i db_i, \quad (\text{Eq. 6})$$

where T is temperature; dS is the function determined by the differential equation; S is the entropy; dU is the amount of change in the internal energy of the system, determined by the initial and final state of the system and representing the total differential; B_i is the generalized force that is, at equilibrium, a function of the external parameters b_i and temperature T (for non-static processes, the generalized force B_i is a function not only of external b_i and internal a_i parameters, but also of the derivatives db_i/dt and da_i/dt).

Below we present the use of the method of thermodynamic potentials. It should be noted that the method of thermodynamic potentials or the method of characteristic functions was developed by Gibbs. This analytical method is based on the use of the basic equation of thermodynamics for quasi-static processes. If the state of a system "eroded thawed soil – water flow" is determined by some generalized coordinates b_i and the entropy of the system S , then its thermodynamic potential is the internal energy $U(S, b_i)$.

$$dU = TdS - \sum_{i=1}^n B_i db_i, \quad (\text{Eq. 7})$$

The state of a system, determined by the external parameters b_i and temperature T , is characterized by the thermodynamic potential – the Helmholtz free energy $F(b_i, T)$

$$F = U - TS. \quad (\text{Eq. 8})$$

Having differentiated (8) and bearing in mind (6), we can write the following equation:

$$dF = -SdT - \sum_{i=1}^n B_i db_i, \quad (\text{Eq. 9})$$

At $T = \text{const}$, expression (9) can be written as:

$$dF = - \sum_{i=1}^n B_i db_i, \quad (\text{Eq. 10})$$

If the state of a system is determined

by the temperature T and the generalized forces B_i , which are conjugate of the external parameters b_i , then the thermodynamic potential of the system is the Gibbs free energy $G(T, B_i)$:

$$G = U - TS + pV, \quad (\text{Eq. 11})$$

where p is the hydrostatic pressure; V is the volume of a system.

Having differentiated (11), we obtain the following equation:

$$dG = dU - SdT - TdS + pdV + Vdp. \quad (\text{Eq. 12})$$

Taking into account (6), the resulting expression (12) can be written as:

$$dG = pdV + Vdp - SdT - \sum_{i=1}^n b_i dB_i. \quad (\text{Eq. 13})$$

Assume that the work δA_1 , produced by a water flow in an eroded soil is to shift its boundaries and overcome some uniformly distributed hydrostatic pressure p . Notice also that p is not necessarily equal to the pressure in the system. However, this equality of pressures occurs if the eroded soil is in a quasi-static equilibrium with the water flow. If dV is an infinitesimally small amount of change in the volume of the system, then we define it as:

$$\delta A_1 = b_1 dB_1 \equiv pdV, \quad (\text{Eq. 14})$$

independently of the shape or internal structure of the system. On the basis of (14), equation (13) takes the following form:

$$dG = Vdp - SdT - \sum_{i=2}^n b_i dB_i. \quad (\text{Eq. 15})$$

If the requirements $p = \text{const}$ and $T = \text{const}$ for the thawed soil are met, the Equation 15 is rewritten as follows:

$$dG = - \sum_{i=2}^n b_i dB_i. \quad (\text{Eq. 16})$$

If the system's independent parameters are the entropy S and the generalized forces B_i , then the thermodynamic potential is the enthalpy $H(S, B_i)$

$$H = U + pV, \quad (\text{Eq. 17})$$

the differential for which can be written as:

$$dH = dU + pdV + Vdp. \quad (\text{Eq. 18})$$

Based on (7), (14), and provided that $p = \text{const}$, the latter can be written as:

$$dH = TdS - \sum_{i=2}^n b_i dB_i. \quad (\text{Eq. 19})$$

Presented below is an analysis of the values of the obtained equations (7), (10), (16), (19) in the context of the possibility of their experimental measurement. Formula (7) is inconvenient from a practical point of view because one of its independent variables – S cannot be measured directly, like and V , p , and T . A similar conclusion can be made for equation (19). It is necessary to measure the external parameters b_i for equation (10) since in this case, the generalized force B_i is not only a function of b_i and T , but also of the internal parameters a_i , and their time derivatives.

In view of the foregoing, we consider that it possible to use equation (16) to describe erosion processes in thawed soils, since in this

case, the work of external forces $\sum_{i=2}^n b_i dB_i$ is

equal to an amount of the free energy change in the system. The negative sign before the equation (16) denotes the fact that an increase in the free energy of the system is associated with energy input. The point of the independent

variables of $\sum_{i=2}^n b_i dB_i$ is that they present elementary works performed by the system against forces of different nature, for example gravity force $\delta A_2 = b_2 dB_2$, surface tension forces at phase boundaries $\delta A_3 = b_3 dB_3$, osmotic forces $\delta A_4 = b_4 dB_4$, magnetic forces $\delta A_5 = b_5 dB_5$, electric forces $\delta A_6 = b_6 dB_6$, etc.

3) SEP of thawed soils without allowance for infiltration

In an equilibrium system, with a stabilized erosion process, pressure and temperature of the system are constant, and

according to equations (15) and (16), the free energy dG is minimal and constant throughout the system's volume. Thus it follows that the constancy G is a necessary and sufficient condition for thermodynamic equilibrium in the isothermal system and, consequently, this equilibrium can be ensured by various combinations of $b_i dB_i$, defining G . After integrating equation (16)

$$\Delta G = -\Delta A, \quad (\text{Eq. 20})$$

where ΔA is the integral energy expended to erode thawed soil, without allowance for infiltration.

Then, based on (1), SEP of thawed soil, without allowance for infiltration can be formulated as:

$$\psi_{twd} = \frac{\Delta G}{\Delta m} = -\frac{\Delta A}{\Delta m}. \quad (\text{Eq. 21})$$

The obtained expression (21) shows that $\text{SEP}_{\psi_{twd}}$ is a constant value for a particular soil and expresses its properties. This statement is valid because, in equilibrium thermodynamics, stable states are recognized at least by a local minimum of free energy and at least by a local maximum of entropy due to the law of decrease of free energy and increase of entropy.

Suppose that the considered erosion process proceeds in an open thermodynamic system, which receives non-zero flows of arbitrary signs from an external medium, i.e. free energy does not necessarily has a minimum, and entropy – a maximum. Then, the generalized force is expressed as:

$$B_i = f\left(\frac{db_i}{dt}, \frac{da_i}{dt}\right);$$

$$\frac{db_i}{dt} \geq \frac{\Delta b_i}{\tau}; \frac{da_i}{dt} \geq \frac{\Delta a_i}{\tau},$$

where $\Delta b_i, \Delta a_i$ are the amounts of change in b_i and a_i , when the system is taken out of equilibrium state; τ is the relaxation time, i.e. the time interval during which the system returns to a state of equilibrium.

Other functions are necessary to distinguish stable non-equilibrium stationary states from other states. There are several criteria [43,47], sufficient for solving the problem

of the stability of a non-equilibrium stationary state. One of them is based on the analysis of stability by linear approximation, the other on the concept of the Lyapunov function. In addition, if an erosion process is linear or relatively weak non-linear in the context of thermodynamics, then for calculating statistical fluctuation characteristics of internal parameters, as well as for calculating flow parameters, it is possible to apply methods [44] based on Markov or non-Markovian fluctuation-dissipation relations. Without dwelling on the advantages of this or that stability criterion and methods for calculating statistical characteristics of fluctuations and flows, since they are the subject of separate studies, it is possible to formally propose the second derivative of the potential ψ_{twd} , as a function characterizing the stability of the erosive non-equilibrium process.

$$\frac{d^2 \psi_{twd}}{dt^2} \leq 0. \quad (\text{Eq. 22})$$

4) SEP of thawed soils with allowance for infiltration

When solving problems associated with erosion processes, it is necessary to differentiate the total amount of water mass interacting with soil, surface run-off and infiltration flow [12,13,14,15,16,17,18,19,20]. Usually, the works focused on calculating soil wash-off over a certain period of time distinguish the surface run-off as the main factor and do not take into account the soaking effects, or treat it as a certain averaged value, neglecting its spatial variability. While averaging absorption as a factor, the parameters influencing it are also averaged: hydraulic conductivity, diffusivity, sorptivity, etc. This significantly influences the results of calculations in solving the main task of erosion processes and leads to significant differences between the results of field experiments and calculated values.

5) Conditions for the erosion process in thawed soil with allowance for infiltration

Soil erosion with regard to infiltration is possible when the velocity of waterfront infiltrated into the soil is higher than the rate of destruction of the soil layer by water flow. If dh_s is the layer of soil eroded water flow over a time period dt , and $dh_{w.inf}$ is the depth of waterfront infiltrated into the soil over the same time, then

$$\frac{dh_{inf}}{dt} \geq \frac{dh_s}{dt},$$

and equation (4) can be written as:

$$\frac{dm_{inf}}{\rho_s + f(h_{inf}, t)} \geq \frac{dm_s}{\rho_s}. \quad (\text{Eq. 23})$$

Then, equation (5) can be written as:

$$\frac{dm_{inf}}{\rho_s + f(h_{inf}, t)} - \frac{dm_s}{\rho_s} = \Delta V_w \geq 0, \quad (\text{Eq. 24})$$

where ΔV_w is the volume of moisture absorbed over a time period dt to a depth $dh_{w.inf}$.

Volume ΔV_w can be expressed in terms of sorptivity, a function proposed by Philip [48], which is the volume of moisture ΔV_w , absorbed per unit area A within the first-second t

$$C = \frac{\Delta V_w}{A\sqrt{t}} = \left[(\theta_1 - \theta_0) \frac{F_1}{b_1} \right]^{1/2}, \quad (\text{Eq. 25})$$

where θ_0 is the volumetric soil moisture, m^3/m^3 ; θ_1 is the steady soil moisture content after passing through the wetting front at the flow potential F_1 .

According to Philip [48], $b_1 = 0,5$; according to Reynolds and Elric [49], for most soils $b_1 = 0,55$.

By expressing ΔV_w using (25), and by substituting it in (24), we obtain the following equation:

$$f(h_{inf}, t) = \rho_s \left(\frac{dm_{inf}}{dm_s + \rho_s C A \sqrt{t}} - 1 \right). \quad (\text{Eq. 26})$$

Thus, the obtained conditions (24) and (26) show that soil exchanges mass with water flow, that is, the system's energy changes, due to a change in the concentration or composition of its components, due to infiltration of water or salts by an amount

$$\sum_{i=1}^e \mu_i dn_i > 0, \quad (\text{Eq. 27})$$

$$\text{where } \mu_i = \left(\frac{\partial u}{\partial n_i} \right)_{S, b_i} \text{ is the chemical}$$

potential of the i -th component.

6) The basic thermodynamic equation of the erodible thawed soil, with allowance for infiltration

The difference between erosion processes with and without allowance for filtration characteristics of soils is that the former are accompanied by a change in the energy of a system by an additive quantity determined by expression (27). As applied to a quasi-static process, the basic equation of thermodynamics for systems with variable mass can be written as

$$TdS = dU + \sum_{i=1}^n B_i db_i - \sum_{i=1}^e \mu_i dn_i. \quad (\text{Eq. 28})$$

Considering the fact that the most commonly used characteristic functions are the thermodynamic potentials U , F , G , and H , then, on the basis of (28), we can obtain:

$$dU = TdS - \sum_{i=1}^n B_i db_i + \sum_{i=1}^e \mu_i dn_i, \quad (\text{Eq. 29})$$

$$dF = - \sum_{i=1}^n B_i db_i + \sum_{i=1}^e \mu_i dn_i, \quad (\text{Eq. 30})$$

$$dG = - \sum_{i=1}^n b_i dB_i + \sum_{i=1}^e \mu_i dn_i, \quad (31)$$

$$dH = TdS - \sum_{i=1}^n b_i dB_i + \sum_{i=1}^e \mu_i dn_i, \quad (\text{Eq. 32})$$

where

$$\mu_i = \left(\frac{\partial u}{\partial n_i} \right)_{S, V} = \left(\frac{\partial u}{\partial n_i} \right)_{T, V} = \left(\frac{\partial u}{\partial n_i} \right)_{T, P} = \left(\frac{\partial u}{\partial n_i} \right)_{S, P}$$

is the chemical potential of the i -th component.

An analysis of equations (29), (30), (31), and (32) in the context of possible experimental measurement of their independent variables leads to the conclusions obtained above.

Therefore, to describe erosion processes, taking into account the filtering properties of soils, it is possible to apply equation (31), since in this case the work of external forces

$\sum_{i=1}^n b_i dB_i$ and the change in the energy of a system due to a change in the concentration or composition of its components in mass-exchange

process $\sum_{i=1}^e \mu_i dn_i$ are equal to an amount of

change in the free energy of the system. The "-" sign before the first sum denotes the fact that an increase in the free energy of the eroded system is associated with energy inputs. The "+" sign before the second sum shows that infiltration reduces soil erosion.

The independent variables $b_i dB_i$ included in (31) denote the same as those in (16), that is, they represent the elementary works performed by the system against forces of different nature. The independent variables in

$\sum_{i=1}^e \mu_i dn_i$ denote the relative changes in the work

$\delta A_1 = \mu_1 dn_1, \delta A_2 = \mu_2 dn_2, \delta A_3 = \mu_3 dn_3, \dots$, done by the system against the same forces of a different nature in connection with the occurred mass-exchange processes during infiltration. It should also be noted that many variables in (31) necessarily change when other variables change. In addition, in a number of specific cases, small changes in such variables as the temperature T , the content of free substances dissolved in water, and others can be neglected, which may lead to further simplifications of expression (31) for dG .

7) SEP with allowance for infiltration

In an equilibrium system, with a stabilized erosion process, and the system's pressure and temperature are constant, and according to equations (31), the free energy dG is minimal and constant throughout the system's volume. Consequently, the constancy G , which is a necessary and sufficient condition for thermodynamic equilibrium in an isothermal system, can be provided by a different combination of the quantities $b_i dB_i, \mu_i dn_i$, defining G .

After integrating equation (31)

$$\Delta G = -\Delta A = -\Delta A_1 + \Delta A_2, \quad (31)$$

where ΔA_1 is the integral energy expended to erode soil with allowance for infiltration; ΔA_2 is the integral energy expended in energy-mass exchange processes due to absorption of water

into the soil.

Such a separation of ΔA into components ΔA_1 and ΔA_2 is rather conditional, but there is fundamental importance: it is possible to develop a device for experimental determination of both ΔA_1 , and ΔA_2 and thereby assess the contribution of these components to the erosion process.

By dividing the resulting equation by Δm , we obtain

$$-\frac{\Delta A}{\Delta m} = -\frac{\Delta A_1}{\Delta m} + \frac{\Delta A_2}{\Delta m}, \quad (32)$$

which, bearing in mind (1), can be written as:

$$\psi_{twd} = \psi_{twd.inf} - \psi_{inf}, \quad (\text{Eq. 33})$$

where $\psi_{twd.inf}$ is the SEP with allowance for infiltration; ψ_{inf} is infiltration potential with allowance for the influence of energy-mass exchange processes due to infiltration of water into the soil.

It follows that

$$\psi_{twd.inf} = \psi_{twd} + \psi_{inf}, \quad (\text{Eq. 34})$$

that is, the actual SEP ψ_{twd} is a value that is additive and constant for a particular soil and expresses its properties. Naturally, spatial variability of ψ_{inf} also affects the value $\psi_{twd.inf}$, nevertheless, the constancy of ψ_{twd} for a particular soil is preserved.

To evaluate the stability of erosive non-equilibrium processes with allowance for infiltration, by analogy with (22), one can propose the following function:

$$\frac{d^2 \psi_{twd}}{dt^2} = \frac{d^2 \psi_{twd.inf}}{dt^2} - \frac{d^2 \psi_{inf}}{dt^2} \leq 0. \quad (\text{Eq. 35})$$

8) SEP for frozen and thawing soils

Preliminary remarks

The process of soil erosion that occurs in the spring period is the result of a complex interaction of the system "atmosphere – the underlying terrain". Below are the processes resulting from the exchange processes: soil freezing, accumulation of precipitation, melting of snow, thawing of soil, infiltration of meltwater into the soil, the formation of overland run-off, and soil erosion. The listed processes are studied in depth and presented in [21, 50, 51]. However, the issues of freezing and thawing of

soil, migration of soil moisture to the freezing front, are currently the subject of scientific discussions. We shall dwell on these issues in more detail since their formulation close to the problem of interaction of slope run-off of meltwater with frozen and thawing soil. It should be noted that the following brief review cannot claim to cover the entire volume of literature related to this thermo- and mass-exchange processes.

Apparently, the migration of water in the freezing soil was for the first time noted by V.I. Stuckenberg [52] while studying the frost-heaving phenomenon. He suggested that this effect arises from a change in capillarity due to the appearance of cracks during freezing and increase of water volume during its transition to ice.

Later this approach found both its followers and opponents. Thoroughly set up experiments by A. Jumikis [53], showed that the capillary meniscus disappears quickly enough upon contact ice and water, and therefore, the capillary forces can not cause migration of soil moisture to the freezing front.

M.N. Goldstein [54] suggested that the migration of soil moisture is mainly due to osmotic forces arising as a result of the difference in pressures between frozen water in hydrate shells and unfrozen water from the thawed horizons, where the concentration of the solution is lower. However, subsequent studies [55, 56] have shown that the effect of osmotic forces on the migration process is insignificant.

According to N.A. Puzakov [57], the main role in the migration of soil moisture belongs to the mineral composition of clay components and the presence of exchangeable cations. Their influence is significant when the content of clay particles has an insignificant effect on water permeability of soils.

The migration phenomena have been considered in the paper [58] in terms of crystal chemical concepts and it is believed that the migration of moisture occurs as a result of a decrease in the free energy of the system or its isobaric potential. Moreover, with increasing concentration of soluble salts, the coagulation process and aggregation of the finely dispersed material is observed.

As early as 1936, A.F. Lebedev [59] came to the conclusion that the migration of soil moisture is a result of the interaction of film water

adsorbed on the surface of solid particles and brought into motion due to the action of growing ice crystals. The film mechanism of mass transfer has been further developed in [60, 61, 62], according to which, as the temperature falls, free and slightly distorted water turns into the ice first. In this case, the thickness of water films decreases and the process of pumping from thick films to thin films occurs, and the adsorption forces on the ice surface bringing water molecules from thin films into motion with subsequent crystallization.

In the work [63], the diffusion theory was used to describe the movement of moisture in freezing soil. However, experiments have shown that the satisfactory correlation between calculations and observations was achieved only after decreasing the diffusivity by an order of magnitude. According to V.E. Goryaev [64], the transfer of soil moisture in the winter period is a thermodiffusion process.

Thus, the complexity of the processes in freezing soils and the lack of unified approaches to the migration of soil moisture do not allow to create a unified calculation scheme for a quantitative accounting of the forces acting at the same time, and for the quantitative assessment of migrating moisture. The application of systems of heat and mass transfer equations and their modifications for these purposes [65, 66] was unsuccessful since an analytical solution of the mentioned systems of equations required a number of simplifying hypotheses or special experiments for determining newly introduced coefficients.

Therefore, in our opinion, the experimentally established fact [52] of fundamental significance is that the driving force of moisture migration in freezing soil is the gradient of the thermodynamic potential μ_w of soil moisture, arising from the difference in specific energies in thawed and frozen soil. Detailed experimental studies have revealed that at the initial moment of cooling of soil, there is practically no change in μ_w . With appearance of intensive phase transitions of water and formation of frozen zone, the binding energy of water remaining in the unfrozen state increases and, consequently, the potential μ_w begins to increase in absolute value and reach a maximum, and then decreases again due to a decrease in the temperature gradient in the frozen zone by the time of establishment of thermal equilibrium of

soil with the external environment. That is, the potential of soil moisture is based on the concept of chemical potential ($\mu=\mu_w$), representing in physical context the work to be done to isothermally and reversibly move the unit mass of pure water from the volume to a given point of soil. It is obvious that the potential μ_w consists of individual potentials, each of which is associated with the action of specific forces, for example, capillary, adsorption, osmotic, gravitational, electric, etc. Then, the above approaches to the study of migration phenomena in disperse environments are quite understandable and, what is especially important, the structural and texture changes that occur during thawing and cyclically recurring freezing-thawing processes in soil observed in the spring periods on fields during the snowmelt, are explainable.

Indeed, many works [58, 60, 67, etc.] have shown that the freezing of dispersing environments is accompanied by ice formation, an increase in the concentration of dissolved substances in the pore space, partial aggregation of the mineral skeleton, partial fragmentation and reorientation of micro- and macroaggregates (as a result of which particles of dust fraction aggregate to the size of 0.1 ... 0.5 mm, while the large aggregates are partially crushed to smaller fractions).

Two variants of thawing are considered in [60]: without and with the migration of moisture from the thawing to the frozen part of the soil. Thawing without migration occurs with the rapid advance of the thawing front and at low humidity, i.e. ice content, and thawing with migration is observed most often with the slow thawing of sufficiently moist (icy) rocks. Almost the same physicochemical processes develop in both types of thawing: increase in the dispersity of thawing rocks as a result of dispersion and peptization of soil aggregates and blocks, as well as crushing of primary particles. The difference is that in the frozen part of thawing rocks with a massive cryogenic texture, due to *grad* μ_w , there is a movement of unfrozen water in the direction of thawing front into frozen water. As moisture enters the region of lower negative temperatures, it partly freezes due to disturbance of the equilibrium between unfrozen water and ice. As a result, there is an increase in ice content, which is almost always observed in the spring in the catchment areas during the snowmelt.

It is noted in [68] that the freeze-thaw cycles lead to the dispersion of the mineral part and the coalescence of structural soil elements.

The enlargement of structural elements occurs due to coagulation and aggregation during the phase transition of the pore solution to ice, where the authors observed a noticeable coarsening after the first freeze-thaw cycle.

Yu.V. Gurikov and colleagues [69] established that the processes of seasonal freezing-thawing of soils lead to mobilization of water-soluble calcium in A horizon.

Thus, considering that the soil is a multiphase multicomponent system, the elements of which, despite the different location conditions, are always united by a structural and functional relation, and are in a state of dynamic equilibrium with the environment, then to describe the interaction of the slope run-off of thawed water with frozen and thawing soil, we consider it possible in the first approximation to apply the laws of thermodynamics, and in particular, its third law.

10) Conditions of the erosion process for frozen and thawing soils

Our laboratory-based reconnaissance experiments on thawing soils, using the specially developed device [70], as well as a number of studies [71, 72, 73, 74] have shown that the process of snowmelt passes through a series of phases that do not have clear time boundaries:

- melting and heating of snow within a temperature range from $T < 0^\circ\text{C}$ to $T = 0^\circ\text{C}$;
- wetting to saturation and partial infiltration through snow;
- filtration through the entire thickness of snow cover;
- continuous longitudinal filtration of water through the snow along the slope;
- formation of a sub-snow rill network;
- formation of an open rill network;
- formation of thawed patches and run-offs along the soil surface.

Thus, with continuous longitudinal infiltration of water into snow and the formation of a sub-snow rill network, the temperature of the surface layer of soil varies from $T < 0^\circ\text{C}$ to $T = 0^\circ\text{C}$ and the water flow interacts predominantly with the frozen soil. The formation of thawed patches leads to the interaction of the water stream mainly with the thawing soil.

Assume that dh_s is the layer of soil that is eroded by water flow over a period of time dt , and dh_{twg} is the thaw depth progression over the same time, then

$$\frac{dh_{inf}}{dt} \geq \frac{dh_s}{dt}, \quad (\text{Eq. 36})$$

By expressing dh_s and dh_{twg} through elementary volumes dV_s and dV_{twg} , the density of soil skeleton ρ_s , the density of thawed soil ρ_{twd} , the density of the migratory moisture $\rho_{mig} = f(h_{mig}, t, grad T_s)$ and the elementary masses dm_s and dm_{twg} , the following formula can be obtained:

$$\frac{dm_{twd}}{\rho_{twd} - f(h_{mig}, t, grad T_s)} \geq \frac{dm_s}{\rho_s}, \quad (\text{Eq. 37})$$

where $\rho_{mig} = f(h_{mig}, t, grad T_s)$ is the distribution function of the density of the migratory moisture from the thawed zone to the frozen soil layer h_{mig} over a time period t and with a temperature gradient $grad T_s$.

With some approximation, we can assume that $\rho_s \approx \rho_{twg}$, since the thawed soil receives water formed during the snowmelt. Therefore,

$$\frac{dm_{twd}}{\rho_s - f(h_{mig}, t, grad T_s)} \geq \frac{dm_s}{\rho_s}, \quad (\text{Eq. 38})$$

The obtained conditions (37) and (38) show that the thawed soil exchanges mass both with the water flow and with the frozen soil, that is, the amount of the system's energy changes due to a change in the concentration or composition of its components, due to migration of moisture from the thawed layer to the frozen and of infiltrated meltwater or salts by an amount

$$\sum_{i=1}^e \mu_{inf,i} dn_{inf,i} + \sum_{i=1}^e \mu_{mmi} dn_{mmi}, \quad (\text{Eq. 39})$$

where $n_{mm,i}$ is the amount of the i -th component in the migration of moisture from the thawed zone to the frozen layer, expressed by the number of moles; $n_{inf,i}$ is the amount of the i -th component in the infiltration of thawed water into thawed soil, expressed by the number of moles; $\mu_{inf,i}$ is the value characterizing the change in the free energy of soil during infiltration, caused by a change in the content of a given i -th component, at constant values of the remaining independent components; $\mu_{mm,i}$ is the value characterizing the change in the free energy of soil during the migration of moisture from the thawed zone to the frozen layer, caused by a change in the content of the i -th component, at constant values of the remaining independent ones.

Assume that dh_{fs} is the layer of frozen soil eroded by water flow over a time period dt , and dh_{twg} is the thaw depth of soil over the same period of time, while $dh_{fs} > dh_{twg}$, that is, the slope run-off of thawed water interacts with frozen soil. Then,

$$\frac{dh_{fs}}{dt} \geq \frac{dh_{twg}}{dt}, \quad (\text{Eq. 40})$$

$$\frac{dm_{fs}}{\rho_{fs} + f(h_{mig}, t, grad T_s)} \geq \frac{dm_{twg}}{\rho_{twg} - f(h_{mig}, t, grad T_s)}. \quad (\text{Eq. 41})$$

If, with some approximation, it is assumed that $\rho_{fs} \approx \rho_{twg}$, while dm_{fs} and dm_{twg} differ insignificantly, then inequality (41) changes its sign. Consequently, erosion of frozen soil with water flow is possible only if it is thawed, i.e. $dh_{fs} = dh_{twg}$.

The critical value dm_{fs}^* can be established after determining the function $\rho_{w.mig} = f(h_{mig}, t, grad T_s)$, then

$$\frac{dm_{fs}^*}{\rho_{fs} + f(h_{mig}, t, grad T_s)} \equiv \frac{dm_{fs}}{\rho_{twg} - f(h_{mig}, t, grad T_s)}. \quad (\text{Eq. 42})$$

Condition (41) is obviously satisfied when $dm_{fs} > dm_{twg}$, that is, in the interaction of frozen soil with a water flow of a sufficient power (rupture of soil and soil aggregates, hydraulic fracturing, destruction by injection, etc.). It also follows from (41) that the frozen soil exchanges mass with the water flow, that is, a change in the energy of a system occurs due to a change in the concentration or composition of its components due to moisture migration from the flow to the frozen layer by an amount

$$\sum_{i=1}^e \mu'_{mmi} dn'_{mmi}, \quad (\text{Eq. 43})$$

where $n'_{mm,i}$ is the amount of i -th component in the migration of moisture to the frozen layer, $\mu'_{mm,i}$ is the chemical potential of the migrating moisture.

11) The basic thermodynamic equation for eroded frozen and thawing soil

Interaction of water flow with frozen and thawing soil is accompanied by a whole complex of interconnected heat and mass exchange, physicochemical and physicomachanical processes. Considering that all the parameters of the interrelated heat and mass exchange, physicochemical and physicomachanical processes vary infinitely slowly and the erosion process has stabilized, i.e. a quasi-static process is carried out, the basic equation of thermodynamics for a system with variable mass can be written as:

$$TdS = dU + \sum_{i=1}^n B_i db_i - \sum_{i=1}^e \mu_{inf,i} dn_{inf,i} - \sum_{i=1}^e \mu_{mmi} dn_{mmi} \quad (\text{Eq. 44})$$

when the water stream predominantly interacts with the thawed soil and conditions (37) or (38) and equality (42) are satisfied, and the change in the energy of a system occurs by an additive quantity (39);

$$TdS = dU + \sum_{i=1}^n B_i db_i - \sum_{i=1}^e \mu'_{mmi} dn'_{mmi}, \quad (\text{Eq. 45})$$

when the water flow predominantly interacts with the frozen ground and condition (41) is satisfied, while the energy of a system is changing by the value (43).

Analogously to the above, based on (44) and (45) for the Gibbs free energy $G(T, B_i, n_i)$ in a differential form, can be obtained the following equations:

$$dG = -SdT + pdV + Vdp - \sum_{i=1}^n b_i dB_i + \sum_{i=1}^e \mu_{\text{inf } i} dn_{\text{inf } i} + \sum_{i=1}^e \mu_{mmi} dn_{mmi}, \quad (\text{Eq. 46})$$

$$dG = -SdT + pdV + Vdp - \sum_{i=1}^n b_i dB_i + \sum_{i=1}^e \mu'_{mmi} dn'_{mmi} \quad (\text{Eq. 47})$$

For $\delta A_1 = b_1 dB_1 = pdV$, $p = \text{const}$, equations (46) and (47) can be written as:

$$dG = -SdT - \sum_{i=2}^n b_i dB_i + \sum_{i=1}^e \mu_{\text{inf } i} dn_{\text{inf } i} + \sum_{i=1}^e \mu_{mmi} dn_{mmi} \quad (\text{Eq. 48})$$

$$dG = -SdT - \sum_{i=2}^n b_i dB_i + \sum_{i=1}^e \mu'_{mmi} dn'_{mmi} \quad (\text{Eq. 49})$$

We consider possible to apply equations (48) and (49) to describe erosion processes in frozen and thawing soil, since in this

case the work of external forces $\sum_{i=2}^n b_i dB_i$, the

change in the energy of a system due to a change in the concentration or composition of its components in mass-exchange processes

$$\sum_{i=1}^e \mu_{\text{inf } i} dn_{\text{inf } i}, \quad \sum_{i=1}^e \mu_{mmi} dn_{mmi},$$

$$\sum_{i=1}^e \mu'_{mmi} dn'_{mmi} \text{ and the accompanying thermo-}$$

physicochemical processes is equal to the change in the free energy of a system.

The sign "-" before the first term and the first sum of equations (48) and (49) reflects the fact that an increase in the free energy of an

eroded system is associated with thermo-physicochemical processes and energy inputs. The sign "+" before the second and third sums of equation (48), and also before the second sum of equation (49) shows that mass exchange between frozen and thawed soil and water flow leads to a decrease in the free energy, that is, to decreasing erosion processes.

The meaning of the independent variables $\sum_{i=2}^n b_i dB_i$ is that they are elementary

works performed by the system against forces of a different nature. The independent variables in

$$\sum_{i=1}^e \mu_{\text{inf } i} dn_{\text{inf } i}, \quad \sum_{i=1}^e \mu_{mmi} dn_{mmi} \quad \text{and}$$

$$\sum_{i=1}^e \mu'_{mmi} dn'_{mmi} \text{ reflect the relative changes in}$$

the work, $\delta A_2 = \mu_{\text{inf } 2} dn_{\text{inf } 2}, \dots$, performed by a system against the same forces of different nature in the context of the occurred mass exchange processes (migration of soil moisture to the freezing front).

The component $S dT$ takes into account the total change in the energy of a system as a result of the transition $\delta A_1 \rightarrow \delta A_1'$, $\delta A_2 \rightarrow \delta A_2'$, $\delta A_3 \rightarrow \delta A_3'$, ..., and the accompanying thermo-physico-chemical processes. For example, during freezing and thawing of soils, hydration, peptization and other processes are observed. Thus, during hydration, ions, molecules of the dissolved substance and colloidal particles are combined with water molecules. The study of hydration of ions by various methods (in terms of electrical conductivity, osmotic properties, spectroscopic method, ion mobility) has shown that a large amount of energy is released during hydration, which is the basic energy necessary for the separation of ions bound in a crystal lattice upon its dissolution. Peptization or colloidal dissolution is the process of transition of a gel to a sol, which is associated with the formation of double electrical layers. As shown by numerous studies, the charging of gel particles can occur either due to the adsorption of ions into the inner coating of the double layer or by the increased diffusion of the outer layer of the double layer. These examples are given only to show that these or those transition processes are accompanied by various chemical reactions, physical phenomena, etc., that is, there are thermo-physicochemical

processes leading to a total change in the thermodynamic system by an amount of $S dT$.

It should also be noted that many variables in (48) and (49) necessarily change when other variables change. In addition, in a number of practical cases, small changes in such variables as the osmotic and the capillary pressure in frozen soil, the content of free substances dissolved in water, etc., can be neglected, which can lead to further simplifications of equations (48) and (49).

12) SEP for frozen and thawing soils

In an equilibrium system, with stabilized thawing, infiltration, and erosion processes, the temperature of all system's elements is of a finite value, and the free energy dG according to equations (48) and (49) is minimal and constant throughout the system's volume. This equilibrium can be ensured by a different combination of the following values:

SdT , $b_i dB_i$, $\mu_{inf i} dn_{inf i}$, $\mu_{mmi} dn_{mmi}$, $\mu'_{mmi} dn'_{mmi}$. After integrating (48) and (49), we obtain the following expressions:

$$\Delta G = -\Delta A = -\Delta A_f + \Delta A_{thg} = -\Delta A_1 - \Delta A_2 + \Delta A_3 + \Delta A_4 \quad (\text{Eq. 50})$$

$$\Delta G = -\Delta A = -\Delta A_f + \Delta A_{thg} = -\Delta A_1 - \Delta A'_2 + \Delta A'_3, \quad (\text{Eq. 51})$$

when ΔA_i is the integral energy expended to erode frozen soil; ΔA_{thg} is the integral energy expended to thaw frozen soil (thawing); ΔA_1 , ΔA_2 , $\Delta A'_2$, ΔA_3 , $\Delta A'_3$, ΔA_4 are the energies expended, respectively, to erode soil, for accompanying thermo-physicochemical processes in the transition of soil from the frozen to thawed state, for energy-mass exchange as a result of migratory processes taking place in frozen and thawed soils.

Based on (1), SEP for frozen and thawing soils can be written as:

$$\psi_{thd} = \frac{\Delta G}{\Delta m} = -\frac{\Delta A}{\Delta m} = -\frac{\Delta A_f - \Delta A_{thd}}{\Delta m} = \psi_{fs} - \psi_{twg}, \quad (\text{Eq. 52})$$

where ψ_{thd} is SEP for thawed soils; ψ_{twg} is the thaw potential for frozen soils.

It follows that

$$\psi_{fs} = \psi_{thd} + \psi_{thg}, \quad (\text{Eq. 53})$$

that is, SEP ψ is an additive and constant value for a specific soil.

To assess the stability of non-equilibrium erosion processes for frozen and

thawing soils, by analogy with (22) and (35), the following function can be proposed:

$$\frac{d^2 \psi_{TWD}}{dt^2} = \frac{d^2 \psi_{fs}}{dt^2} - \frac{d^2 \psi_{twg}}{dt^2} \leq 0. \quad (\text{Eq. 54})$$

Of course, initial soil temperature T_s , function $\rho_{mig} = f(h_{mig}, t, \text{grad } T_s)$, spatial variability of other parameters under which a water flow interacts with frozen and thawing soil influence on ψ_{thg} and, consequently, on ψ_{fs} , but, nevertheless, the constancy ψ for a particular soil is preserved. Therefore, we propose to determine the potential ψ_{fs} by formula (53).

Discussions

38 samples from A horizon of different fields were taken to determine SEP for the most common soil types of the Chuvash Republic. Komsomolskiy District - 4; Shemurshinsky District - 3; Batyrevsky District - 4; Yantikovsky District - 3; Urmarsky District - 3; Alikovsky District - 3; Porecki District - 6; Cheboksary District - 8; Krasnochetayskiy District - 4. In addition, soil samples were taken from the fields of Leninskaya Iskra collective farm of Yadrinsky District. After preparing soil samples according to a unified method (sifting air-dry soil through a sieve with a hole diameter of 1 mm and vacuum humidification to the complete saturation), and determination of PES was carried out using an experimental unit of our own design [42]. Table 1 shows the results of experimental research [42].

An analysis of the research results shows that the values of SEP for the main soil types of the Chuvash Republic vary within fairly wide limits: 0.082...0.276 J/kg for varying bulk densities of soils in solid phase within the range of 2.70...2.96 g/cm³. Nevertheless, the SEP for leached heavy loam black soils is almost 2 times higher than for alluvial sod soils, which is comparable with the data of M.S. Kuznetsov [21].

Let us compare the obtained values of SEP with the relevant values from previous studies. For this purpose, we use the most well-known formulas to determine the critical flow velocities in a hydraulic tray, derived from the principles of calculation by the limit state method:

Mirzkhulava Ts. E. [10]:

for cohesive soils with plasticity index $W_p \leq 15$:

$$V_{\Delta p.n.e.} = 1,25 \left(\frac{2m}{2,6\gamma_0 n} \left[(\gamma_2 - \gamma_0)dg + 1,25C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 55})$$

$$V_{p.n.e.} = 1g \frac{8,8H}{d} \left(\frac{2m}{2,6\gamma_0 n} \left[(\gamma_2 - \gamma_0)dg + 1,25C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 56})$$

15: for cohesive soils with plasticity index $W_p >$

$$V_{\Delta p.n.e.} = 1,25 \left(\frac{2m}{1,54\gamma_0 n} \left[(\gamma_2 - \gamma_0)dg + 1,25C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 57})$$

$$V_{p.n.e.} = 1g \frac{8,8H}{d} \left(\frac{2m}{1,54\gamma_0 n} \left[(\gamma_2 - \gamma_0)dg + 1,25C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 58})$$

for non-cohesive soils, by size of soil aggregates:

$$V_{\Delta p.n.e.} = 1,25 \left(\frac{2m}{0,44\gamma_0 n} \left[g(\gamma_2 - \gamma_0)D_{a.p.} + 2C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 59})$$

$$V_{p.n.e.} = 1g \frac{8,8H}{d_{95}} \left(\frac{2m}{0,44\gamma_0 n} \left[g(\gamma_2 - \gamma_0)D_{a.p.} + 2C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 60})$$

for cohesive soils, by size of soil aggregates:

$$V_{\Delta p.n.e.} = 1,25 \left(\frac{2m}{0,44\gamma_0 n} \left[g(\gamma_2 - \gamma_0)d + 2C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 61})$$

$$V_{p.n.e.} = 1g \frac{8,8H}{d_{95}} \left(\frac{2m}{0,44\gamma_0 n} \left[g(\gamma_2 - \gamma_0)d + 2C_d^n K \right] \right)^{0,5}, \quad (\text{Eq. 62})$$

Kuznetsov M.S. [21]: equation 63

where $V_{\Delta per}$ is the permissible (non-eroding) bottom flow velocity (i.e. velocity at the level of bed projections Δ), m/s; V_{per} is the bottom-averaged permissible (non-eroding) velocity, m/s; the logarithmic relationship between the permissible bottom velocities $V_{\Delta per}$ and the permissible averaged velocities from top to bottom for the above formulas (56), (58), (60), (62) have been taken based on Goncharov's model, assuming that the height of bed projections Δ is equal to $0.7 d$; $V_{\Delta pw}$ is the bottom eroding flow velocity for soil with the initial moisture content W , m/s; H is the flow depth, m; d is the average size of detached parts, m; $D_{a.p.}$ is the average particle size of erosion pavement, calculated as the weighted average over the area of erosion pavement occupied by large particles of the soil surface layer, m; d_{95} is the diameter of

particles that are larger than those making up 95% of soil mass, m; d_w is the weighted average diameter of water-resistant aggregates after wet sieving (Savvinov's method), with its initial moisture content W , m; m is the erosion index, depending on slope vegetation, presence of erosion pavement, and the initial soil moisture; m_1 is the erosion index, depending on the presence of erosion pavement and suspended sediments in the flow; m_2 is the index characterizing the binding action of plant root systems; n is the index characterizing the velocity pulsation of the flow; γ_2 , γ_0 is the bulk density of soil and water particles, respectively, t/m³; P is the porosity of aggregates, %; α is the angle of inclination of the flow channel, deg.; C_d^n is the nominal surface durability of soil, t/m²; C_w is the cohesive strength (Tsytoich's method) between soil and initial moisture content w after rapid flooding of its surface and saturation up to the moisture storage capacity, t/m²; K is the uniformity coefficient; I is the coefficient characterizing the soil consistency and depending on the bulk mass of soil.

The critical flow velocities, determined by formulas (55...63), characterize the conditions for the erosion of soil samples in the hydraulic tray, while SEP ψ_m calculated using the experimental data and formulas [42] (64) and (65),

$$\psi_{TWD} = \frac{Q_{sp}^2}{2B^2 h_2^2} + \frac{Q_w \rho_w}{2B^2 \rho_s (Q_{sp} - Q_w)} \left(\frac{Q_{sp}^2}{h_2^2} - \frac{Q_w^2}{h_1^2} \right), \quad (\text{Eq. 64})$$

$$\psi_{TWD} = \frac{t}{2B^2 m_s} \left[\frac{Q_w^3 \rho_w}{h_1^2} - \left(Q_w + \frac{m_s}{\rho_s t} \right)^3 \frac{\rho_s (\rho_w Q_w t + m_s)}{h_2^2 (\rho_s Q_w t + m_s)} \right], \quad (\text{Eq. 65})$$

allows to determine the amount of energy required to destruct a unit mass of soil sample in the hydraulic tray at a certain "critical" flow velocity (where B is the width of the tray, h_1 , h_2 is the height of the water flow and suspension, respectively, $\rho_w, \rho_s, \rho_{sp}$ is the density of water, soil, and suspension, respectively; Q_w , Q_{sp} is the flow rate of water and suspension, respectively; m_s is the soil mass eroded by the flow over a time period t).

Consequently, the above equations describe the same process, but equations (55 ... 63) are expressed kinematically, and equations (64) and (65) – energetically. Indeed, the unit of the potential ψ is [J/kg] or [m²/s²], and the unit of critical velocity – [m/s]. Therefore, using the theory of dimensions:

$$\left. \begin{aligned} \psi_{TWD1} &= V_{p.n.e.}^2 \\ \psi_{TWD2} &= 1,25^{-2} V_{\Delta p.n.e.}^2 \lg^2 \left(\frac{8,8H}{d} \right) \\ \psi_{TWD3} &= 1,25^{-2} V_{\Delta p.w}^2 \lg^2 \left(\frac{8,8H}{d} \right) \end{aligned} \right\} \quad (\text{Eq. 66})$$

The above dependencies (66) take into account the permissible (non-eroding) averaged velocities from top to bottom, characterizing most fully the energy of the flow, but not the bottom flow permissible velocities that take into account the energy of the flow at the level of bed projections.

Similar results can be obtained by examining studies on determining the permissible (non-eroding) bottom velocity in the plunge pool and in centrifugal devices [42]. Table 2 shows some results of the determination of ψ_f potential, using (66) and based on the experimental data [10, 21, 42].

By comparing numerical values ψ_{TWD} , shown in Tables 1 and 2, it can be seen that the values ψ_{TWD} are of the same order of magnitude and, therefore, show the comparability of the considered studies.

CONCLUSIONS:

The developed theoretical fundamentals for determining the soil erosion potential enables us to draw the following conclusions.

1. We propose to use the thermodynamic potential method to describe soil erosion processes. It has been shown that the Gibbs thermodynamic potential is the most suitable for these purposes since it is possible to design devices for experimental measurement of independent variables included in the relevant differential equations (16), (31), (48) and (49).

2. SEP is an additive and constant value for a given soil and expresses its properties, i.e. it can be adopted as a single integral criterion for assessing the resistance of soils to erosion.

3. Functions (22), (35) and (54), representing the second derivatives of SEP ψ_{thd} over a time period t have been proposed to assess the stability of erosive non-equilibrium processes in the context of thermodynamics.

REFERENCES:

1. Kuznetsov M.S., Rozhkov A.G., Glazunov G.P. *Sovremennoye sostoyaniye i perspektivy razvitiya issledovaniy po zashchite pochv ot erozii v Rossii* [The current state and prospects for the development of research on soil erosion protection in Russia]. *Pochvovedeniye Journal*, 1994, No. №4, pp. 100-109.
2. Zaslavsky M.N. *Eroziya pochv* [Soil erosion]. Moscow, Mysl Publ., 1978, 245 p.
3. Surmach G.P. *Vodnaya eroziya i bor'ba s ney* [Soil erosion by water and countering it]. Leningrad, Gidrometeoizdat Publ., 1976, 254 p.
4. Sobolev S.S. *Razvitiye erozionnykh protsessov na territorii Yevropeyskoy chasti SSSR i bor'ba s nimi* [Development of erosion processes on the territory of the European part of the USSR and countering them]. Moscow, AN SSSR, v. 1, 1948, 307 p; v. 2, Moscow, AN SSSR, 1960, 248 p.
5. Bennett H.H. Some comparisons of the properties of humid tropical and humid-temperate American soils. *Soil Science*, 1926, v. 21, p. 249-275.
6. Ryabov E.I. *Dorozhezolota* [More valuable than gold]. *Zemledeliye Journal*, 1988, No. 9, pp. 25-29; 1988, No. 11, pp. 27-31.
7. Konke G., Bertran A. *Okhrana pochvy* [Soil protection] (In Russian), ed. by Sobolev S.S. Moscow, Selskokhozyaystvennaya literatura Publ., 1962, 343 p.
8. Magomedova A.V. *Prognoz erozionnykh protsessov i transporta nosov* : Dissertatsiya d.t.n. [Forecast of erosion processes and sediment transport: Thesis for a Candidate Degree in Technical Sciences]. Makhachkala, 1982, 426 p.
9. Mirzkhulava Ts. E. *Inzhenernyye metody rascheta i prognozavodnoy erozii* [Engineering methods for calculating and prediction of water erosion]. Moscow, Kolos Publ., 1970, 240 p.
10. Mirzkhulava Ts. E. *Osnovy fiziki i mekhaniki erozii rusel* [Physical and mechanical engineering fundamentals for riverbed erosion]. Leningrad, Gidrometeoizdat Publ., 1988, 303 p.

11. Erosiyapochvy [Soil erosion]. (Translation in Russian foreword by Pushkarev M.F.). Moscow, Kolos Publ., 1984, 415 p.
12. ELE International Limited 1988. Guelph Permeameter Operating Instructions
13. Elrick D.E., Reynolds W.D., Baymgartner N., Tan K.A. and Bradshaw K.L. In Situ Measurement of Hydraulic properties of soils using the Guelph Permeameter and the Guelph Infiltrometer. In Proc. Third International Workshop on Land Drainage, 1987, Columbus, OH, Dec, p. 7-11.
14. Reynolds W.D. and Elrick D.E. In Situ Measurement in the Vadose Zone of Field-Saturated Hydraulic Conductivity, Sorptivity and the Conductivity - Pressure Head Relationship, - 1986, Groundwater Monitoring Review, v. 6, p. 84-95.
15. Reynolds W.D. and Elrick D.E. In Situ Measurement of Field-Saturated Hydraulic Conductivity, Sorptivity and α - Parameter Using the Guelph Permeameter // Soil Science, - 1985, v. 140, p. 292-302.
16. Globus A.M. Eksperimental'naya gidrofizikapochv [Experimental hydrophysics of soils]. Leningrad, Gidrometeoizdat Publ., 1969, 355 p.
17. Globus A.M. Pochvenno-gidrofizicheskoye obespecheniye agroekologicheskikh matematicheskikh modeley. [Soil and hydrophysical data in agroecological mathematical modelling]. Leningrad, Gidrometeoizdat Publ., 1987. - 427 p.
18. Philip J. R. Teoriya infil'tratsii. Izotermicheskoye peredvizheniye vody v zone aeratsii [Theory of infiltration. Isothermal movement of water in the aeration zone.], ed. by Averyanov S.F. Leningrad, Gidrometeoizdat Publ., 1972, pp. 6-17.
19. Reynolds W.D., Elrick D.E., Pondered infiltration from a single ring: 1. Analysis of steady flow. S.S.S.A., - 1990, 1.58, p. 1233-1241.
20. Grigoryev V.Ya. Nekotoryye metodicheskiye voprosy laboratornogo izucheniya erozionnykh protsessov. Sovremennyye aspekty izucheniya erozionnykh protsessov [Some methodological issues of laboratory study of erosion processes. Bulletin: Modern aspects of the study of erosion processes]. Novosibirsk, Nauka Publ., 1980, pp. 63-65.
21. Kuznetsov M.S. Protivoerozionnaya stoykost' pochv [Anti-erosion resistance of soils]. Moscow, MGU Publ., 1981, 135 p.
22. Larionov G.A. Eroziya i deflyatsiya pochv. [Erosion and deflation of soils]. Moscow, MGU Publ., 1993, 200 p.
23. Shvebs G.I. Formirovaniye vodnoy erozii, stokananosovikhotsenka (naprimere Ukrainy i Moldavii) Formation of water erosion, sediment drainage and their assessment (on the example of Ukraine and Moldova). Leningrad, Gidrometeoizdat Publ., 1974, 183 p.
24. Vilensky D.G. Svoystva pochv, opredelyayushchiye podatlivost' ikherozii metody issledovaniya etikh svoystv. Bor'ba s eroziyey pochv v SSSR [Soil properties determining the compliance of their erosion and methods for studying these properties. Countering soil erosion in the USSR]. Moscow, AN SSSR Bulletin, 1938, pp. 111-129.
25. Gussak V.B. Izucheniye protsessov smyva i erozii v lotke. Pochvovedeniye [Studying soil loss and erosion processes in a tray], Pochvovedenie. No. 1, 1945, pp. 27-39.
26. Lutz V.F. The physicochemical properties of soils effecting erosion. Missouri Agr. Exp. Sta. Research, Bul., 1931, pp. 5-45.
27. Wischmeier W.H., Smith D.D. Predicting Rainfall erosion Losses // Agriculture Handbook, - 1978, United States Department of Agriculture, Washington, D.C.
28. Saupe G. Die Erosivitat die Niederschlage im Suden der DDR - ein Beitrag zur quantitativen Prognose der Bodenerosion // Arch. Naturschulz Landschaftsforsch., 1985, 25, 3, s. 155-169.
29. Mutchler C.K., Carter C.E. Soil erodibility variation during the year // Trans. ASAE, Joseph, Mich., 1983, 26, 4. p. 1102-1104.
30. Johnson R.R. Putting soil movement into perspective // Product. Agr., -1988, 1, 1, p. 5-12.
31. Wall G.J., Dickinson W.T., Rudra R.P., Coote D.R. Seasonal soil erodibility variation in southwestern Ontario // Can. J. Soil Sci., - 1988, 68, p. 417-424.
32. Makkaveev N.I. Ruslo reki i eroziya v yeye basseyne. [Riverbed and its erosion]. Moscow, AN SSSR Publ., 1955, 346 p.
33. Makkaveev N.I., Chalov R.S. Ruslovyeye

- protsessy [Riverbed processes]. Moscow, MGU Publ., 1986, 264 p.
34. Kondratiev S.A. Gidrodinamicheskoye modelirovaniye vodnoerozionnykh protsessov na sklonakh i malykh vodosborakh. Trudy 5 Vsesoyuznogo gidrologicheskogo syezda. [Hydrodynamic modeling of water erosion processes on slopes and small watersheds. Proceedings of the 5th All-Union Hydrological Congress]. Leningrad, Kniga 2, tom 10, "Ruslovyye protsessy i nanosy", 1986, pp. 134-141.
 35. Sukhanovsky Yu.P. Razrabotka imitatsionnykh modeley vodnoy erozii pochv. Pochvozashchitnaya tekhnologiya poliva i povysheniye nadezhnosti protivopavodkovoy zashchity [Development of simulation models for soil erosion by water. Soil-protective irrigation technology and increasing the reliability of flood protection]. Pushchino, Sbornic nauchnykh trudov, 1990, pp. 100-102.
 36. Shvebs G.I. et al. Raschet sklonovykh nanosov i ovrazhnoy erozii dlya obosnovaniya protivoroziionnykh melioratsiy. Trudy 5 Vsesoyuznogo gidrologicheskogo s"yezda. T. 10. - Ruslovyye protsessy i nanosy, kn. 2. [Calculation of slope sediments and gully erosion to justify erosion melioration. Proceedings of the 5th All-Union Hydrological Congress. V. 10. "Channel processes and sediments", Book 2.]. Leningrad, Gidrometeoizdat, 1986, pp.141-148.
 37. Franti T.G. et. al. Modeling the mechanical effects of incorporated residue on rill erosion // Amer. soc. of agric. engineers, 1987, Paper 87-2010, p. 15.
 38. Griffen M.L. et. al. Estimating soil loss on topographically nonuniform field and farm units // Soil and Water Conserv., - 1988, 43, p. 326-331.
 39. Hirschi Michael C. et. al. KYERMO - a physically based research erosion model. Part 2. Model sensitivity analysis and testing // Trans. ASEA, -1988, v. 31. p.814-820.
 40. Van Liew M.W., Saxton K.E. Dynamic Simulation of Sediment Discharge from Agricultural Watersheds // Trans. ASEA, - 1984, v. 27. p.1087-1093.
 41. Foster G.R. et al. User requirements: USDA - water erosion prediction project (WEPP) // Amer. soc. of agric. engineers, 1987, p. 6, Paper 87-2539.
 42. Maksimov I.I., Maksimov V.I. Energeticheskaya kontseptsiya erozionnoy ustoychivost antropogennykh agrolandshaftov [Energy concept of erosion resistance for anthropogenic agricultural landscapes]. Cheboksary: Chuvash State Agricultural Academy Publ., 2006, 304 p.
 43. Stratonovich R.L. Nelineynaya neravnovesnaya termodinamika [Non-linear non-equilibrium thermodynamics]. Moscow, Glavnaya redaktsiya fiziko-matematicheskoy literatury, 1985, 480 p.
 44. Bazarov I.P. Termodinamika: Uchebn. posobiye dlya un-tov. 2-ye izd. pererab. i dop. [Thermodynamics. A manual for graduate students, 2nd ed.]. Moscow, Vyschaya shkola Publ, 1976, 447p.
 45. Maksimov I.I., Sirotkin V.M. Energeticheskiy podkhod k izucheniyu erozionnykh protsessov // Povysheniye effektivnosti vuzovskoy nauki i uluchsheniye kachestva podgotovki spetsialistov s vysshim obrazovaniyem: Sbornik tezisov [An energy approach to studying erosion processes. Increasing the effectiveness of university science and improving the quality of postgraduate education. Collection of abstracts]. Cheboksary, 1990.
 46. Bogomolov A.I., Mikhailov K.A. Gidravlika [Hydraulics]. Moscow, Stroiizdat Publ., 1972, 648 p.
 47. Kaiser J., Statisticheskaya termodinamika neravnovesnykh protsessov. Perevod s angliyskogo, predislovie YU. L. Klimontovicha [Statistical thermodynamics of non-equilibrium processes. Translated from English. Foreword by Yu.L. Klimontovich]. Moscow, Mir Publ. 1990, 608 p.
 48. Philip I.R. The theory of infiltration: 4. Sorptivity and algebraic infiltration equations // Soil Sci. Soc. Amer. - 1957. - vol. 84, p. 257 - 264.
 49. Reynolds W.D., Elrick D.E., Ponded infiltration from a single ring: 1. Analysis of steady flow. S.S.S.A., - 1990, 1.58, p. 1233-1241.
 50. Bobrovitskaya N.N. Grafoanaliticheskaya model' formirovaniya stoka nanosov s raspakhannykh sklonov. Trudy 5 Vsesoyuznogo gidrologicheskogo

- s"yezda. t. 10. Ruslovyie protsessy i nanosy. Kniga 2 [A graph-analytical model for forming sediment runoff from plowed slopes. Proceedings of the 5th All-Union Hydrological Congress. Vol. 10. "Channel processes and deposits". Book. 2]. Leningrad, 1986, pp. 126-134.
51. Gussak V.B. Pribor dlya uskorenogo opredeleniya erodiruyemosti pochv i nekotoryye rezul'taty yego primeneniya [Soil erodibility accelerated assessment device and some results of its application]. Pochvovedenie Journal, 1946, No. 8, pp 481 - 491.
52. Mikrostroyeniye merzlykh porod. Pod red. E.D. Yershova [Microstructure of frozen rocks / Ed. by Ershov E.D.]. Moscow, MGU Publ., 1988, 188 p.

Equitation 63

$$V_{\Delta pw} = 1,55 \left(\frac{m_1 m_2 g}{\gamma_0 n} \left[\left(1 - \frac{P}{100} \right) d_w (\gamma_2 - \gamma_0) (\cos \alpha - \sin \alpha) + 1,25 K I C_w \right] \right)^{0,5},$$