Infiltration of snowmelt water into frozen soil

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ABSTRACT: This report deals with the interaction between meltwater and frozen soil. To forecast the volume of snowmelt runoff, it is necessary to compute the volume of infiltration into frozen soil. This problem is dealt with in the first part of the report. Formulae are given for computing the depth of formation of impermeable layers and infiltration losses.

The second part of the report deals with computing infiltration into frozen soil. Formulae are given uniting infiltration into frozen top soil and groundmelt. These formulae are compared with the results of field experiments.

RESUME: On considère l'interaction entre l'eau de fonte et le sol gelé. Pour faire la prévision du débit total de la crue de printemps, il est nécessaire de calculer les pertes d'eau de fonte par infiltration dans le sol gelé. Ce problème est considéré dans la première partie du rapport. On déduit les formules pour calculer les conditions de la formation de la couche imperméable et les pertes d'eau de fonte par infiltration dans le sol gelé.

La seconde partie du rapport traite des moyens de calculer l'infiltration de l'eau dans le sol gelé, en tenant compte du gel partiel de l'eau d'infiltration. On donne les formules pour les cas les plus simples d'infiltration dans les sols gelés et dégelés. Ces formules donnent des résultats qui sont comparés avec les résultats expérimentaux.

SYMBOLS

C - parameter, depending on the ratio of diameters of ice crystals to soil particles, cm/cm;

C_b, C_η, C_n - heat capacity of water, ice, and soil, cal/g°C;

K - permeability of ground saturated with ice, mm/min;

K_0 - coefficient of filtration if $\eta = 0$, $T = 0°C$, mm/min;

$\overline{K}$ - effective permeability of the whole moistened zone, mm/min;

$K_T(U)$ - moisture conductivity of melting soil, mm/min;

η - ice saturation of ground in the moistened zone, cm^3 ice/cm^3 soil;

η_H - initial ice saturation of ground, cm^3 ice/cm^3 soil;

L - specific heat of melting, cal/g;

n - exponent;

P - total porosity, cm^3/cm^3;

P_{KT}, P_K - capillary head of melt and icy ground, cm of water column;

Q_X - cold content in a unit of soil volume, cal/cm^3;

Q_Z - heat content penetrating into a unit of soil volume during infiltration and released at soil freezing, cal/cm^3;

q_o - rate of infiltration (seepage), mm/min;

T_H - initial temperature of soil, °C;

T_{Hb} - initial temperature of infiltrating water, °C;

T_{KP} - critical temperature of soil when blocking layer is formed, °C;

V - layer of absorbed water;

v - volume of infiltrating water contained in a unit of volume of frozen soil, cm^3/cm^3;

$U(T_H)$ - liquid water content at the initial temperature of soil, cm^3/cm^3;

W - total moistening in the wet zone when infiltration occurs into melting and frozen ground, $W = U + \eta\rho_\eta\rho_b^{-1}$, cm^3/cm^3;

W_H - total initial moistening, cm^3/cm^3;

$\overline{W}$ - maximum bound water content, cm^3/cm^3;

$W_{Hz}(T)$ - bound water content not frozen at the negative temperature T, accounted for 1 g of dry soil, g/g;

$W_{H,KP}$ - initial critical volumetric soil moisture, cm^3/cm^3;

y - depth of wet soil layer, cm;

δ_T, δ - free porosity (capacity) of melting and frozen soil, cm^3/cm^3;

ρ_η, ρ_b - density of ice and water, g/cm^3;

σ - volumetric weight of ground, g/cm^3;

δ_y - volume of free pores above a given depth y, mm;

HB - minimum soil moisture, cm^3/cm^3.

COMPUTATION OF TOTAL LOSSES OF SNOW MELT WATER BY INFILTRATION INTO FROZEN SOIL

The rate and volume of spring snowmelt runoff depends not only on the amount of water in the snowpack and the snowmelt intensity, but also on the initial (flood-preceding) distribution of moisture and temperature within the soil profile, the depth and nature of freezing, and variations in infiltration. This problem is discussed in the next section of the paper. However, to forecast total spring snowmelt runoff it is sufficient to compute total losses of snowmelt water by infiltration.

The whole process of snowmelt water infiltration into frozen soil may be generally divided into four phases:

1. Phase of decreasing infiltration rate,
2. phase free from infiltration (at complete saturation above an impervious layer),
3. phase of increasing infiltration rate (when frozen soil melts),
4. phase of steady infiltration.

The very first portion of water percolating into frozen soil with the temperature below zero freezes in the soil pores. Upon freezing, water releases a considerable amount of heat (about 80 cal/g). As the heat capacity of the frozen soil is small (about 0.5-0.7 cal/cm^3), the soil temperature rises to 0°C and subsequent portions of water percolate through the soil without freezing. This process corresponds to the first phase of infiltration during which the infiltration rate tends to decrease due to several factors: (a) decrease of piezometric gradient due to the extension of the infiltration way at the constant capillary head, (b) decrease of capillary head due to the combination of menisci of percolated water and the menisci of soil moisture, and (c) decrease in their cross-section due to the blocking of soil pores by ice crystals.

If the cold content in the frozen soil is not great, the ice crystals only partly fill the pores and complete cessation of infiltration does not occur; thus the infiltration immediately progresses to the third phase associated with gradual melting.

In case of a sufficient cold content ice fills all the empty pores, resulting in a blocking layer. Where there is complete soil saturation above the blocking layer, infiltration ceases and the second phase of infiltration begins. Up to this time some volume of water V, equal to the volume of pores free from ice and water (before infiltration) in the blocking layer and in all the layers above it may be absorbed. Subsequent percolation (the third phase) may start only after melting of the blocking layer usually after complete disappearance of the snow cover.

The negative soil temperature T_{KP} at which the cold content is sufficient for the ice to block all the free pores is called the "critical temperature", and the initial soil moisture $W_{H,KP}$, corresponding to this temperature, is called the "critical initial moisture".

Critical temperature T_{KP} depends on the volumetric weight of soil σ and the amount of bound water $W_{HZ}(T)$ not freezing at the negative temperature T. This is stipulated by the fact that at the increase of soil temperature to 0°C, the water frozen within the range of negative temperatures from 0°C to T_{KP} starts to melt, resulting in great heat losses. It may seem that, as some portion of frozen water melts at the freezing of infiltrated water, a complete water impermeability should not occur. But it is not so: the water in the larger pores freezes, whereas water absorbed on the surface of soil particles melts.

The amount of absorbed water $W_{HZ}(T)$ not freezing at different negative temperatures, T, was determined in the laboratory for four soils (Table 1).

Appropriate critical temperatures T_{KP} and initial soil moisture $W_{H,KP}$ when a blocking layer is formed in soil, are determined if the following sum equals zero:

$$Q_X + Q_Z = 0$$

in which

Q_x - is the cold content in a unit of frozen soil volume:

$$Q_x = [C_n\sigma + C_\eta \rho_b W_{H,KP} + \sigma W_{Hz}(T_{KP})(C_b - C_\eta)]T_{KP} - L\sigma[W_{Hz}(0) - W_{Hz}(T_{KP})] \quad (1)$$

Q_z - is the heat content in a unit of soil volume or originating from percolating water and liberated upon freezing.

Table 1

Soil type (the Don River Basin)	Volumetric weight g/cm^3	Liquid water content (in parts from the dry soil weight) at temperature					
		0°	-1°	-3°	-5°	-10°	-20°
Dark grey forest soil	0.8	0.27	0.20	0.17	0.17	0.17	0.16
Leached chernozem	1.10	0.34	0.18	0.14	0.13	0.13	0.12
Ordinary chernozem at Buturlinovka	1.10	0.27	0.21	0.18	0.17	0.16	0.16
Typical chernozem	1.10	0.23	0.22	0.18	0.15	0.15	0.15

$$Q_z = P\rho_\eta(L + C_b T_{Hb}) - W_{H,KP}\rho_b(L + C_b T_{Hb}) + \sigma W_{Hz}(0)(L + C_b T_{Hb})(1 - \frac{\rho_\eta}{\rho_b}) \quad (2)$$

according to formula:

$$W_{H,KP} = \frac{(L + C_b T_{Hb})[P\rho_\eta + \sigma W_{Hz}(0)(1 - \frac{\rho_\eta}{\rho_b})]}{\rho_b(L + C_b T_{Hb} - C_\eta T_{KP})} + \frac{[C_\eta + W_{Hz}(T_{KP})(C_b - C_\eta)]\sigma T_{KP} - L\sigma[W_{Hz}(0) - W_{Hz}(T_{KP})]}{\rho_b(L + C_b T_{Hb} - C_\eta T_{KP})} \quad (3)$$

in which $W_{H,KP}$ is in cm^3/cm^3;

$W_{Hz}(T_{KP})$ is in g/g;

$W_{Hz}(0)$ is the amount of unfrozen water if the negative temperature T tends to zero, g/g;

The initial temperature T_{Hb} should be taken into account only for the computation of $W_{H,KP}$ for upper soil layers (1-2 cm) as water reaching other soil layers has the temperature of 0°C.

If using formula (3) we obtain $W_{H,KP} < W_{Hz}(0) \frac{\sigma}{\rho_b}$, then appropriate critical values of T_{KP} and $W_{H,KP}$ are computed according to the formula:

$$W_{H,KP} = \frac{(L + C_b T_{Hb})P\rho_\eta + [C_\eta + W_{Hz}(T_{KP})(C_b - C_\eta)]\sigma T_{KP} + L\sigma W_{Hz}(T_{KP})}{\rho_b(2L + C_b T_{Hb}) - C_\eta \rho_b T_{KP} - (\rho_b - \rho_\eta)(L + C_b T_{KP})} \quad (4)$$

which is obtained from equation (3) where $W_{Hz}(0) = W_{H,KP} \frac{\rho_b}{\sigma}$.

By formulae (3) or (4) a number of appropriate values T_{KP} and $W_{H,KP}$ (Table 2) are computed and used for plotting a graph, $W_{H,KP} = f(T_{KP})$.

Table 2

Soil type (the Don River Basin)	Volumetric weight g/cm³	Initial critical soil moisture (volumetric) at critical temperature T°C				
		0	-1	-3	-5	-10
Dark grey forest soil	1.0	0.59	0.55	0.51	0.49	0.46
	1.4	0.46	0.40	0.35	0.33	0.31
Leached chernozem	1.0	0.60	0.44	0.37	0.35	0.33
	1.4	0.46	0.36	0.30	0.29	0.27
Ordinary chernozem at Buturlinovka	1.0	0.59	0.51	0.48	0.46	0.42
	1.4	0.45	0.36	0.34	0.33	0.31
Typical chernozem	1.0	0.59	0.56	0.52	0.49	0.45
	1.4	0.45	0.42	0.36	0.32	0.30

Infiltrating water volume, which may be contained in a unit of frozen soil volume (above the blocking layer), is computed by the formula:

$$v = [\rho_\eta P - W_H\rho_b + \sigma W_{Hz}(0)(1-\frac{\rho_\eta}{\rho_b})]\rho_b^{-1} \qquad (5)$$

Total snowmelt losses during the first phase of infiltration is determined by integrating v over depth from the surface y = 0 to the upper boundary of the blocking layer y_z

$$V_1 = \int_0^{y_z} v\,dy \qquad (6)$$

The blocking layer is formed at the soil depth $y = y_z$ where its initial temperature T_H is equal to, or below, the critical soil temperature $T_{KP} = f(W_H)$ at the initial water content W_H, ie.

$$y = y_z \qquad T_H \leqslant T_{KP} = f(W_H) \qquad (7)$$

The result of the laboratory heat physical research with monolith of different chernozem types (taken in the Don basin) verified computations of the depth of the blocking layer (y_z) and the amount of snowmelt water losses (V_1) for the first infiltration phase made according to formulae (3) - (7).

Figure 1 gives the results of one of the experiments with the chernozem soil. By using data on the volumetric weight and moistening (curves 1 and 3), critical temperatures for the different soil layers were calculated (curve 4). The figure shows that initial soil temperatures (curve 2) were below critical at depths greater than 30 cm. Consequently, the blocking layer should form at that depth.

Indeed, at the end of percolation, ie., at the 33rd minute from the start of the test, the increase of soil temperature was quite evident only above this depth.

The volume of free pores (curve 5) to the depth of 30-32 cm was 90-94 mm, whereas 89.3 mm was percolated by the moment that infiltration ceased, ie., practically the same value.

Figure 2 illustrates the role of moisture distribution. In both the cases cited the temperature distribution is similar (curve 1); total water storage in the surface 50 cm is also the same (161 mm). In the first case (curve 2) soil moisture contents tend to increase with depth, in the second they decrease. In the first case a blocking layer is formed at a depth of 34.5 cm whereas in the second it is formed at a depth of 9 cm. The amounts of water percolated are 96 mm and 21 mm respectively. Thus, the amount of absorbed water depends on both the temperature and water content distributions.

Verification of the suggested computation method by field observation data provided positive results as well: on the basis of data 1 and 2 and experimental data from E.N. Tsykin [5] Table 3 has been prepared in which values of initial critical moistening, $W_{H,KP} = f(T_H)$, corresponding to the initial soil temperature T_H are given. Table 3 shows that on stubble and fallow land at the high initial moisture contents W_H exceeded the critical $W_{H,KP}$ in the spring of 1954 and a blocking layer should have been formed directly on the soil surface. No percolation into soil was observed. On autumn-ploughed soil with low initial water contents W_H was less than $W_{H,KP}$. A blocking layer was not formed and complete absorption of snowmelt water occurred.

DYNAMICS OF INFILTRATION INTO FROZEN SOIL

In some cases for the computation of infiltration into frozen soil under a certain head, a simple model may be applied [1, 6], described by equations:

$$\delta\frac{dy}{dt} = q_o \quad (8)$$

$$q_o = \overline{K}(1+P_K y^{-1}) \quad (9)$$

The application of equations (8) and (9) infers a scheme usual for such a model. A wet zone is selected where water movement occurs, and a dry zone where movement is not observed. These zones are separated by a wetting front. It is assumed that freezing occurs at the wetting front. The temperature in the wet zone is assumed to be 0°C, whereas in the dry zone it is equal to the initial temperature.

In case of infiltration under head into a semi-finite column of homogeneous ground with uniform initial moisture contents system (1) - (2) has the following analytical solution:

$$t = \delta P_K K^{-1}\{K(q_o-K)^{-1}-\ln[q_o(q_o-K)^{-1}]\} \quad (10)$$

$$t = \delta P_K K^{-1}\{yP_K^{-1}-\ln(1+yP_K^{-1})\} \quad (11)$$

$$t = \delta P_K K^{-1}\{VP_K^{-1}\delta^{-1}-\ln(1+VP_K^{-1}\delta^{-1})\} \quad (12)$$

An approximation of equations (10) - (12) is known:

$$q_o = K+\sqrt{KP_K\cdot 0.5\delta t^{-1}} \quad (13)$$

$$y = Kt\delta^{-1}+\sqrt{2KP_K\delta^{-1}t} \quad (14)$$

Table 3

Observations of water infiltration into frozen soil
(Soil: chestnut fine loamy (Saratov district, vil. Ershovo))

Land	Depth cm	Volumetric weight g/cm^3	Observed temperature °C	Observed moistening: gravimetric g/g	Observed moistening: volumetric cm^3/cm^3	Initial critical moistening: leached chernozem	Initial critical moistening: ordinary chernozem	Observed infiltration
Period before spring snow melting, 1954								
Stubble	1	1.1	-0.2	0.52	0.57	0.50	0.55	absent
"	5	1.1	-0.8	0.36	0.40	0.42	0.52	absent
"	10	1.1	-1.0	0.25	0.28	0.40	0.52	absent
"	20	1.1	-1.2	0.20	0.22	0.38	0.52	absent
Fallow	0	1.2	-5.0	0.45	0.54	0.32	0.41	absent
"	5	1.2	-5.2	0.30	0.36	0.32	0.41	absent
"	10	1.2	-5.7	0.23	0.28	0.32	0.40	absent
"	20	1.2	-6.0	0.18	0.22	0.32	0.40	absent
Autumn	0	0.9	-5.0	0.26	0.23	0.40	0.53	present
ploughing	10	0.9	-7.0	0.22	0.20	0.39	0.51	present
"	20	0.9	-7.2	0.17	0.15	0.39	0.51	present
Start of snowmelt								
Fallow	1	1.2	-1.0	0.47	0.56	0.38	0.49	absent
"	5	1.2	-0.4	0.37	0.44	0.44	0.51	absent
"	20	1.2	-0.6	0.14	0.16	0.41	0.50	absent
Autumn	1	0.9	-0.1	0.44	0.40	0.59	0.62	present
ploughing	5	0.9	-0.13	0.29	0.26	0.59	0.62	present
"	10	0.9	-0.4	0.26	0.23	0.56	0.61	present
"	30	0.9	-1.8	0.18	0.17	0.44	0.58	present

$$V = Kt + \sqrt{2KP_K\delta t} \tag{15}$$

For infiltration into a horizontal column of the homogeneous soil (from (13) - (15)):

$$q_o = (0.5KP_K\delta t^{-1})^{1/2} \tag{16}$$

$$y = \sqrt{2KP_K\delta^{-1}t} \tag{17}$$

$$V = \sqrt{2KP_K\delta t} \tag{18}$$

Equations (8) - (18) are similar for infiltration into frozen ground if the temperature at $t \approx 0$ is constant for all depths. Hydrophysical parameters K, δ, P_K, however, not only have quite different numerical values, but different physical meanings. Therefore the problem of distribution of a simple infiltration model in case of frozen soil is the problem of the computation of parameters.

It is possible to derive theoretical formulae relationships between the hydrophysical parameters K, δ, P_K and similar characteristics for melting[1] soils:

$$K = (1+C\eta)^{-2}K_T(U) \tag{19}$$

where

$$U = W - \eta\rho_\eta\rho_b^{-1} \tag{20}$$

$$P_K = P_{KT}(1+C\eta)(1-\eta P^{-1})^{-1} \tag{21}$$

$$\delta = \delta_T - \eta(\rho_b - \rho_\eta)\rho_b^{-1} \tag{22}$$

For sand completely saturated with ice, formula (19) may be presented as follows [2]:

$$K_o = K_T(1+C\eta)^{-2}(1-\frac{\eta}{P})^3 \tag{23}$$

The analysis of experiments made by V.D. Komarov to determine the coefficient of filtration of ground saturated with ice revealed good agreement between formula (23) and the experiment (Fig. 3a, b) for $C \approx 8$. Moreover it was possible to show [2] that this value actually corresponds to the ratio of pores to ice crystals formed during the experiments of V.D. Komarov.

It follows from formulae (19) and (23) that the moisture conductivity of ground saturated with ice decreases with increase of ice saturation due to two factors. Firstly, the increase of ice saturation causes the decrease of pores saturated with liquid water. This is taken into account in formula (23) by the factor $(1-\eta P^{-1})^3$ just as is done in the well-known formula of S.F. Averianov, I.A. Budagovsky, L.S. Leibenzon, *et al.* [1, 4].

In formula (19) instead of the factor $K_o(1-\eta P^{-1})^3$ the dependence of moisture conductivity on the melting ground moistening has been used. This permits computation when the wet zone is not completely saturated with water and ice but contains some air as well. Thus, it is possible to take into account tht effect of air bubbles blocked

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[1] The additional index T is used to indicate appropriate values at the infiltration into melting ground.

under the influence of infiltration with head. It makes possible the computation of infiltration free from any head using methods described in [1]. To compare this computation with the experiments it is essential to have data on free from head infiltration into frozen ground which are not available at present.

Secondly, with increase in ice saturation the specific surface of soil tends to increase because ice crystals have the same effect as additional particles of ground in the pores. This mechanism is taken into account by factor $(1+C\eta)^{-2}$ of formulae (19) and (23). As shown in [2], if this mechanism is not taken into account an overestimation of moisture conductivity of the frozen soil results.

The effective ice saturation of soil in the wet zone may be computed from the heat balance equation. In this case two variants are possible.

If we accept that ice at the wetting front is formed due to the cold content in the skeleton and ice in the frozen soil, then

$$\eta = \eta_H - T_H(C_\eta \rho_\eta \eta_H + C_n \sigma)(\rho_\eta L)^{-1} \tag{24}$$

This equation gives good results (Fig. 3a, b) for infiltration into frozen sandy soil. In this case soil at 5-10°C below zero has a cold content capable of forming an amount of ice sufficient to fill only a small portion of the available pore space. This explains the high permeability of frozen sand during infiltration.

The second variant is based on the assumption that besides the cold content introduced for the computation in formula (24) the cold content stipulated by phase transitions at negative temperatures also affects infiltration. This hypothesis was put forward by V.D. Komarov, who assumed, that during infiltration into frozen ground at each layer, ice melting at the surface of mineral particles and the formation of big ice crystals in free large pores could be simultaneously observed. Then the effective ice saturation may be written as follows

$$\eta = \frac{-T_H[C_n\sigma + C_\eta\rho_\eta\eta_H + C_b\rho_b U(T_H)] + L\rho_b[\overline{W} - U(T_H)] + \rho_\eta L\eta_H}{\rho_\eta L} \tag{25}$$

Though this hypothesis was proved neither by detailed microscopic investigations nor by special macroscopic tests, and therefore could not be taken for granted, nevertheless it is the only well-known explanation for differences in the absorbing capacity of loose and fine heavy frozen particles discovered during the tests by V.D. Komarov and tests mentioned above. For loose (sandy) soils practically free from bound water, cold content, stipulated by phase transitions at the surface of mineral particles, is very small. In heavy (argillaceous) soils the role of bound water is much greater, and consequently their absorbing capacity is much less at negative temperatures than at positive ones.

Equations (8) - (25) provide a mathematical processing of the most simple field and laboratory tests of percolation into frozen ground. Figure 3c and d compares computed values of infiltration into frozen ground with field tests made by N.A. Mosienko [3].

Formulae (10) - (25) also show that the usual method of calculating the rate of infiltration from the corresponding values of nonfrozen soil (method of coefficients) is not acceptable in the general case. Coefficients obtained in the general case depend not only upon the properties different at positive and negative temperatures (for example, ice saturation), but also upon similar properties (porosity, diameter of ground particles, etc.) and even upon the conditions of the test (water supply rate, time since the beginning of the test, etc.). Therefore such coefficients are not comparable. More representative results may be obtained when equations (19) - (25) defining the hydrophysical characteristics for frozen and melting ground are used. Such equations should be especially useful if it is essential to compute infiltration into soil for a basin with little observational and experimental data available.

The transition from computation of infiltration at a point to computation of total losses by infiltration over a whole basin is a serious problem. In this case it is necessary to take into account not only the portion of areas with different infiltration capacity but sub-basin areas as well, relating to the areas of intensive infiltration. This problem may be solved by means of surveys of hydrophysical characteristics of the melting ground taking into account landscape features and subsequent application of design methods of the determination of frozen ground parameters and the computation of total infiltration losses into frozen soil according to physical and statistical equations.

The methods of computation presented cannot be used for forecasting purposes. In operational practice for prediction of snowmelt water flow, use is made of different empirical correlations between the depth of snowmelt, snow storage, autumn moistening, temperature, and the depth of soil freezing. The intensity of snowmelt is computed by heat balance methods taking into account the depth and density distribution of the snow cover over the basin.

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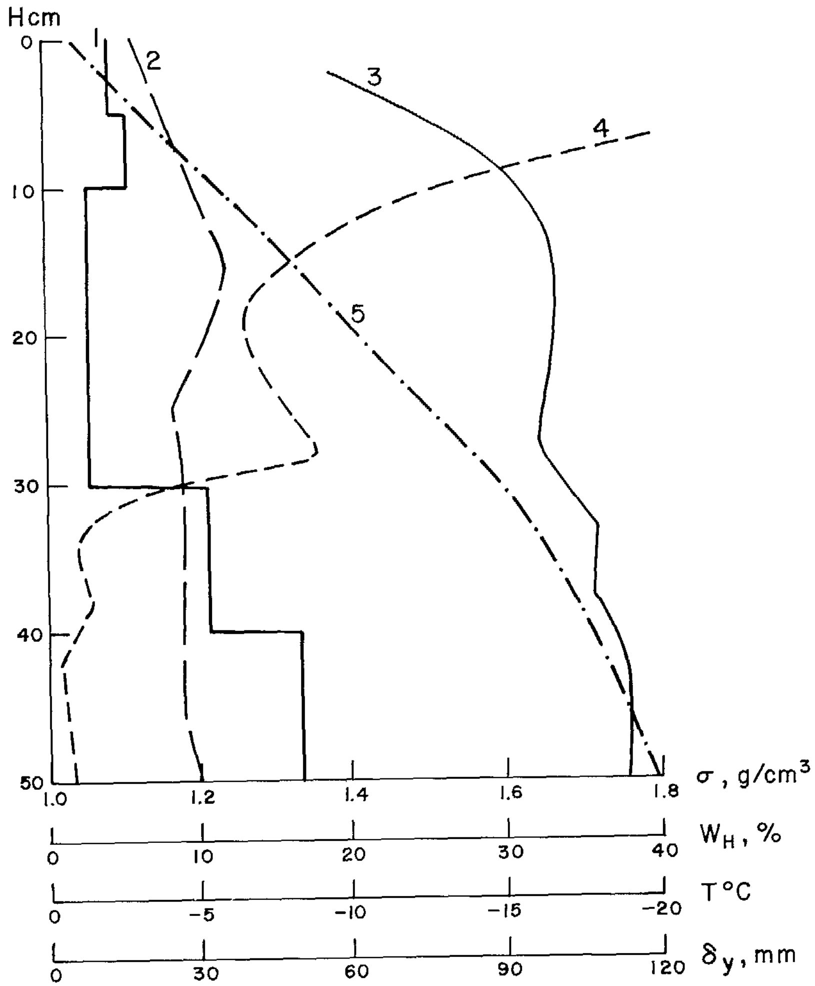

Fig. 1 Infiltration of water into frozen soil. Ordinary chernozem. Test 24.

1 - volumetric weight (σ);

2, 4 - initial (T_H) and critical (T_{KP}) temperatures;

3 - initial volumetric moistening of soil (W_H);

5 - volume of free pores to a given depth y,mm $\delta_y = 10\int_0^y \delta dy$

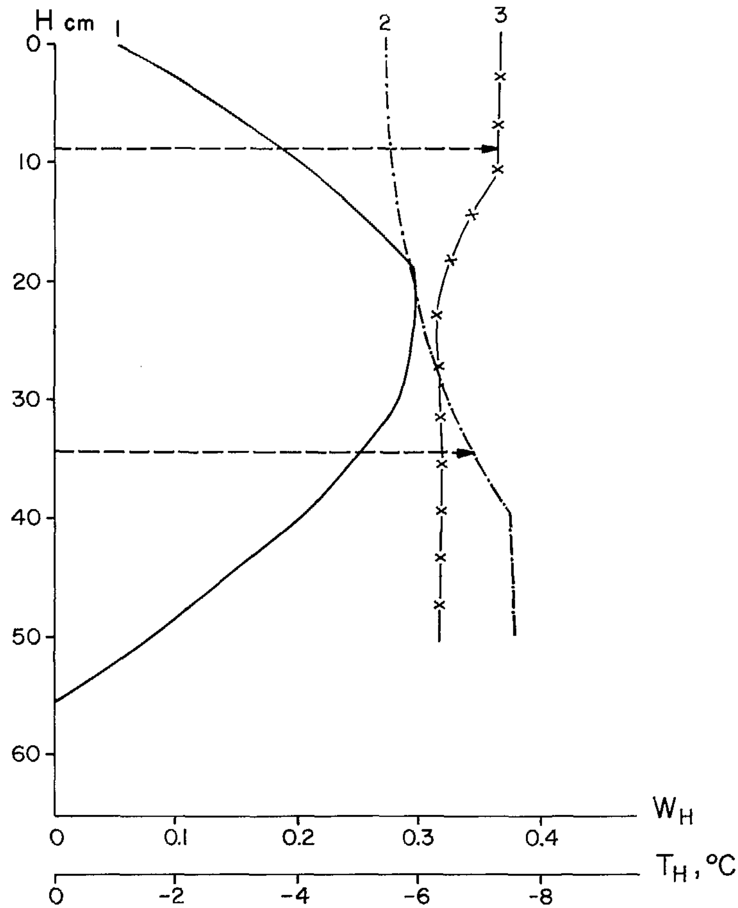

Fig. 2 Distribution of initial temperature (T_H) and soil moistening (W_H) in leached chernozem.
1 - soil temperature (T_H);
2, 3 - soil moistening (W_H), the first and second cases

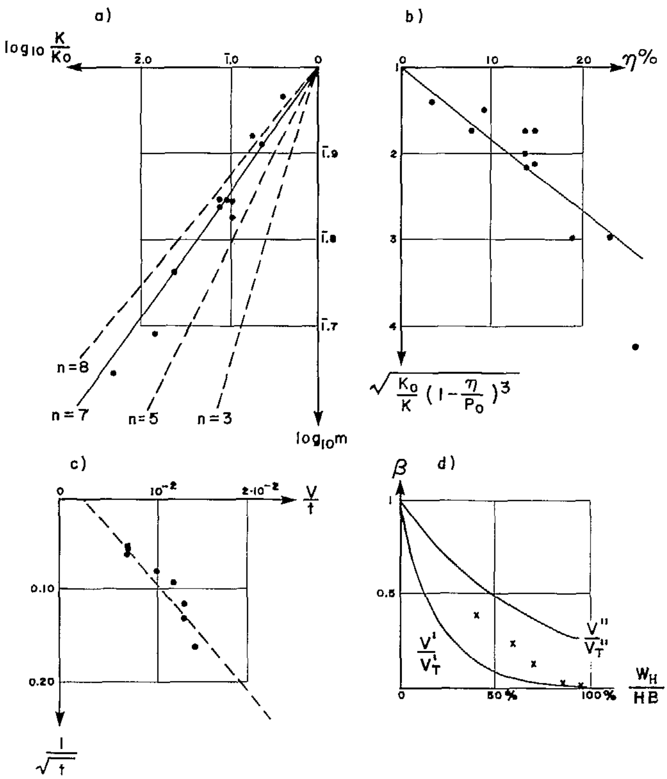

Fig. 3 The comparison of theory with the tests of V.D. Komarov, S.A. Mosienko [3].

a) The dependence of iced sand filtration coefficient on the saturation of soil with water in the form of

$$K/K_0 = \left(\frac{P_0 - \eta - W_H}{P_0 - W_H}\right)^n = (m)^n$$

is shown by the degree indicator $n \approx 7$. It is, however, known [1] that for thawed soil $n \approx 3$-4.

b) It is caused by the fact that the specific surface of solid and liquid phases division in the iced soil is greater for the value of ice crystals surface. The consideration of this factor leads to the formula (16), the comparison of which with the experimental data is shown in Figure 3b.

c) The comparison of the data of Mosienko's field tests [3] with the formula [6]. V-mm, t-min.

d) If the experimental data are presented in the form of relations of water saturated into the frozen and thawed soil

$\beta = \frac{V}{V_T}$, then it is possible to show that $\frac{V'}{V_T'} < \beta < \frac{V''}{V_T''}$ where $V' = KT, V'' = \sqrt{2Kt\delta P_K}$

This paper was presented by R.L. Harlan.
There was no discussion of this paper.