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### Physical phenomena and processes in freezing, frozen and thawing soils

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## PREFACE

This translation is the eighth arranged by the Permafrost Subcommittee of the Associate Committee on Soil and Snow Mechanics of the National Research Council of the Russian permafrost publication, "Principles of Geocryology".

The first translation in this group was Chapter VI entitled "Heat and Moisture Transfer in Freezing and Thawing Soils" by G.A. Martynov (TT-1065). The second was Chapter IV "General Mechanisms of the Formation and Development of Permafrost" by P.F. Shvetsov (TT-1117). The third was Chapter VII "Geographical Distribution of Seasonally Frozen Ground and Permafrost" by I.Ya. Baranov (TT-1121). The fourth was Chapter IX "Ground (Subsurface) Ice" by P.A. Shumskii (TT-1130). The fifth was Chapter X "Ground Water in Permafrost" by V.M. Ponomarev and N.I. Tolstikhin (TT-1138). The sixth was Chapter XI "Cryogenic Physico-geological Phenomena in Permafrost Regions" by S.P. Kachurin (TT-1157), and the seventh was Chapter XII "Perennially Frozen Ground and Vegetation" by A.P. Tyrtikov (TT-1163).

This translation of Chapter V by N.A. Tsyтовich and several of his colleagues deals with physical phenomena and processes in freezing, frozen and thawing soils. It includes discussion of the principle components of frozen soils, the classification of moisture in frozen soils, ice content and liquid phase, equilibrium of water and ice, thermal and physical chemical characteristics, mechanical processes, and the alteration of frozen soils during freezing and thawing.

The Division is grateful to Mr. G. Belkov, Translations Section, National Research Council, for translating this chapter.

Ottawa  
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R.F. Legget  
Director

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Frozen Soils as Multiphase Systems

The Basic Components of Frozen Soils\*\*

Frozen soils are complex, multiphase and multicomponent systems. From the state of aggregation one can identify the following basic components:

(1) Solid which can be subdivided further: a) matrix of frozen soils (mineral or organic-mineral) and b) cryogenic minerals (ice, cryohydrate and crystal hydrate which crystallize at temperatures below freezing);

(2) Liquid;

(3) Gaseous.

Solid component. The skeleton of frozen soils is usually mineral or occasionally organic-mineral. The total of the specific and active surface, the chemical and mineralogical composition of the skeleton and particularly its colloidal fraction have a significant influence on the properties of frozen soils and the processes that occur in them. Not only is the quantitative relationship between ice and the soil skeleton of importance but also their mutual distribution.

The way ice is formed and its properties in frozen soils are of prime importance. Ice is subdivided into three basic types: ice as an independent monomineral deposit, ice as a component of the soil and sublimation ice that is not a component of the soil (Shumskii, 1955<sub>2</sub>)\*\*\*. The formation of crystal hydrate and cryohydrate in frozen soils is connected with physico-chemical processes occurring in them and has not been well elucidated.

Liquid component. The quantity and composition of the liquid phase determine to a substantial degree the direction and intensity of physico-chemical processes in frozen soils, and also the physical and physico-mechanical properties of the frozen soils particularly at temperatures just below 0°C.

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\* This chapter was previously published in Collected Papers No. 3, "Data on the laboratory investigation of frozen soils" edited by N.A. Tsytovich, 1957. In the present publication it has been abridged and revised. A number of changes and additions have been introduced based on recent investigations of frozen soils.

\*\* Written by Z.A. Nersesova.

\*\*\* Ground ice is considered in detail in Chapter IX.

Frozen soils may contain substantial quantities of liquid phase water depending basically on the nature of the mineral skeleton, i.e. the total of the specific and active surface, the mineralogical composition of the fine-grained fractions and their physico-chemical features; it also depends on the content and composition of soluble compounds (primarily salts) and also on temperature and pressure. The quantity of the liquid phase is subject to the equation of state which finds its expression in the principle of the equilibrium state of water in frozen soils (Tsyтовich, 1945).

Gaseous component. Gases in frozen soils may be free, adsorbed or trapped. The quantity of free gases is determined by the gas porosity. The quantity of adsorbed gases depends on the total surface and composition of the soil skeleton and porosity and increases greatly with the presence of organic substances in the soil.

#### Types of Moisture in Frozen Soils\*

The majority of present-day classifications of types of water in soils supplements and develops the classification of A.F. Lebedev (1936) who identified the following basic types of water: gaseous, hygroscopic, capillary, gravitational, and finally chemically bonded water.

Subsequently ground water was subdivided into free water that is not affected by soil particle surfaces and bound water which is affected by soil particle surfaces. The latter includes firmly bound and loosely bound water (Dolgov, 1948; Priklonskii, 1949; Rode, 1952). The categories of firmly bound and loosely bound water vary greatly with different authors.

It should be noted that the terms "free water" and "bound water" are quite arbitrary. For example, in the classification of geologists D.F. Belyankin (1933) and V.I. Vernadskii (1934), the prime consideration is the participation of water in the construction of minerals. Here bound water includes chemically combined water, water of crystallization and to some extent zeolithic water. Hygroscopic water is considered as being free.

In the classifications of agricultural and engineering soil scientists and hydrogeologists (Dolgov, 1948; Priklonskii, 1949; Rode, 1952; Sergeev, 1952; and others) hygroscopic water is regarded as being firmly bound.

A sounder classification of water is that based on the nature of the bond between the water and the soil skeleton (Popov, 1941; Rebinder, 1950).

P.A. Rebinder\*\* identifies the following basic types of bonds between moisture and other substances:

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\* Written by Z.A. Nersesova.

\*\* Cited from the paper by A.V. Lykov "Theory of Drying". 1950.

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|--|---|
| 1. Chemical (a bond in precise quantitative respect)                     | Ionic and molecular bonds.  |
| 2. Physico-chemical (a bond in various not very specific considerations) | Adsorption, osmotic and structural bonds.   |
| 3. Physico-chemical (the retention of water in undefined respects)       | Bond in microcapillaries ( $r$ less than $10^{-5}$ cm), bond in microcapillaries ( $r$ greater than $10^{-5}$ cm) and the wetting bond. |

In frozen soils part of the moisture may be in the liquid state because of the action of ions of soluble compounds.

The dependence of the quantity of liquid phase water in frozen soils on thermodynamic parameters determining the state of systems and partially on the content of soluble compounds differentiate it from the corresponding category of bound moisture defined by chemists, soil scientists and hydrogeologists for the region of temperature above  $0^{\circ}\text{C}$ .

Therefore in addition to the classification of moisture based on the nature of the bond between water and the skeleton, for frozen soils one should identify categories of moisture depending on the physical state of the water.

V.I. Vernadskii (1934), in considering the possible classifications of natural water, suggested a classification based on the physical state of the water: (1) gaseous water (water vapour), (2) liquid water (solutions), (3) solid water (ice).

Basing his classification on the physical state of the water and its transformation at freezing temperatures, Nersesova (1957) makes the following basic categorization of moisture in frozen soils: (1) gaseous; (2) alternating physical state (solid or liquid, freezing or thawing at temperatures below  $0^{\circ}$ ); (3) not changing its physical state with changes in temperature and pressure; (4) solid (ordinary ice formed at the temperature of  $0^{\circ}\text{C}$  and not thawing at temperatures below  $0^{\circ}$ ).

I.A. Tyutyunov (1955) made the following classification for frozen soils; (1) solid phase water - ice; (2) liquid phase water - boundary phase\*, near the boundary and volume phases; (3) gaseous phase.

The relation between the solid, boundary, near the boundary, volume and gaseous phases of water is determined on the one hand by the nature of the

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\* The term "boundary phase" was introduced by B.V. Deryagin in 1950 (Transactions of the All-Union Conference on Colloidal Chemistry. Izd. Akad. Nauk Ukr. SSR, 1952), however I.A. Tyutyunov ascribes a different meaning to this term (see page 11).

mineral skeleton and the content of soluble compounds and on the other hand by external factors, temperature and pressure.

#### Content of Ice and Liquid Water in Frozen Soils\*

As indicated above, the quantity of liquid phase water in frozen soil at a given temperature and pressure depends on the total of the specific and active surface, chemical and mineralogical composition and also on the content and composition of water-soluble compounds.

It would be impossible to isolate the effect of dispersion, physico-chemical features and mineralogical composition of the soil since variation in the ratio of exchange cations, for example the exchange of  $\text{Na}^+$  ions for  $\text{Ca}^+$  or  $\text{Fe}^{+++}$  ions which have different hydration energies, is accompanied by the corresponding change in dispersion - coagulation.

The effect of exchange cations on the water composition of soils is evident particularly when the water content is quite large.

The effect of exchange cations on the phase composition of water in frozen soils is also evident basically when the moisture content is high, exceeding the quantity of film water.

In turn the total of the specific and active surface and the exchange properties of soil depends on the mineralogical composition of the clay fraction which increases sharply in the presence of minerals of the montmorillonite group. The effect of salinization on the phase composition of water in frozen soils can be calculated if one knows the content and composition of water-soluble compounds. However, it is simpler to obtain experimental curves showing the variation in the quantity of liquid phase water depending on temperature for soil containing soluble compounds after the soluble compounds have been removed.

The variations in quantity of liquid water determine the extent of cementing of frozen soils by ice which determines a number of the basic properties of frozen soils. Moreover, all thermal calculations connected with the freezing and thawing of soils will not be reliable if one does not take into account the liquid water and the changes it undergoes under the influence of external forces.

Curves showing the dependence of the content of liquid water on below-freezing temperatures in combination with other data (unit weight, moisture content, specific gravity) permit one to calculate the quantity of the separate components of frozen soil (solid mineral particles, ice, liquid water, vapour and gases), ice content, effective heat capacity, coefficient of heat

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\* Written by N.A. Tsyrovich and Z.A. Nersesova.

conductivity in the region of substantial phase transformation of water, the total settlement due to temperature changes, etc.

Table II and Fig. 5 show the results of determining the content of liquid water for some typical soils, obtained by the Central Laboratory of the Permafrost Institute, from which it follows that the most important factor determining the phase composition of water in frozen soils is the below-freezing temperature.

Table II and Fig. 5 give the results of determining the content of liquid water depending on temperature for typical soils permitting one to use these data for practical purposes and accepted for standards and engineering specifications (NITU - Standards and Engineering Specifications).

#### The Principle of Equilibrium States of Water and Ice\*

The above-mentioned data as well as many other data on the content of liquid water and ice in frozen soils indicate that any change in the below-freezing temperature brings about a corresponding change in the quantity of liquid water which is expressed by a regular relationship expressed graphically by a curve showing the content of liquid phase at various below-freezing temperatures or curves for the melting of pore ice.

The quantity of liquid water in frozen soils depends also on the magnitude of the external pressure exerting enormous local stresses at points of contact of the mineral particles (Tsytovich and Sumgin, 1937) which bring about a corresponding melting of ice at below-freezing temperatures. This has been confirmed by special experiments (Table IV).

On the basis of the data cited above and other data published previously (Tsytovich, 1945, 1947; Nersesova, 1950, 1951; Boiko, 1957; Jung, 1943; and others), one can establish the following mechanism which is called the principle of equilibrium state of water and ice in frozen soils (Tsytovich, 1940, 1941, 2, 1945): the quantity, composition and properties of the liquid phase contained in frozen soils does not remain constant but varies with the parameter of the state of the system being in dynamic equilibrium with the latter.

The principle formulated is well supported by direct experiments and has a number of applications in problems of general and engineering permafrost studies.

The largest changes in the phase composition of water in frozen soils occur at the beginning of the freezing process when the temperature is not far below the freezing point, but when the temperature drops well below the freezing point these changes become so small that they can be neglected for practical purposes.

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\* Written by N.A. Tsytovich.



One can differentiate three temperature regions in the phase transformations of water in frozen soils: (1) the region of substantial phase transformations; (2) the region of transitional phase transformations and (3) the region of the practically frozen state.

As was shown earlier (Tsyтович, 1954) the region of substantial phase transformations of water in frozen soils can be limited to changes in the quantity of liquid water approximately equal (or greater) than one percent (of the weight of dry soil) per degree centigrade, and the region of practically frozen state - less than one-tenth of one percent per degree centigrade.

Experimental data show that for sandy soils the region of substantial phase transformations is limited to temperatures from 0 to  $-0.2^{\circ}\text{C}$  and for heavy clay it is at times extended to  $-7^{\circ}\text{C}$  and lower; the region of the practically frozen state for heavy clay should be considered established only at temperatures of the order of several tens of degrees below  $0^{\circ}\text{C}$ .

#### Physical Phenomena and Processes in Freezing and Frozen Soils

##### Temperature Characteristic of the Freezing of Water in Soil\*

The freezing of water in soil begins with the generation of crystallization factors characterized by greater or lesser supercooling of the water. The second stage - crystallization of water - is accompanied by the liberation of heat and a sudden increase in temperature.

The highest and most stable temperature in the ground following the jump in temperature is the temperature following the beginning of crystallization of water. The extent and duration of supercooling of water in the ground increase as the volume of water decreases, particularly when the amount of water is very small, and also when the rate of cooling of the ground decreases and the amount of water decreases.

Observations show that at temperatures near  $0^{\circ}\text{C}$  supercooling of the water extends over a prolonged time, and when the temperature of the cooling medium decreases, the water rapidly crystallizes but does not reach the temperature of the medium.

The formation rate of crystallization centres has a great effect on the stability of the supercooled state of the water and ice crystals are the most active primer. Ice crystallization in the soil takes place gradually over a certain temperature range.

Calculations and experiments show that in the temperature range from 0 to  $-0.2^{\circ}\text{C}$  free water crystallizes (capillary-gravitation water); water bound by

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\* Written by A.P. Bozhenova.

the surface of particles or by ions in solution crystallizes at a temperature that decreases as the effect of fields counteracting crystallization increases.

Data available at the present time on the temperature of the supercooling and beginning of crystallization of water in soils were obtained on small specimens with disturbed structure. The maximum supercooling established for water under these conditions is  $-5^{\circ}\text{C}$  and the temperature of the beginning of crystallization when the moisture content is high and soluble substance absent was close to  $0^{\circ}\text{C}$  ( $-0.1^{\circ}\text{C}$  and  $-0.2^{\circ}\text{C}$ ).

One can suppose that under natural conditions the supercooling of water takes place only in the surface layer of soil since the formation of ice in the upper layer excludes the possibility of supercooling the water in the lower strata.

#### Physico-chemical Processes in Frozen Soils\*

Until recently it was maintained in the literature that perennially frozen soils usually represent a zone of "chemical dormancy" (Saukov, 1951, p.296).

This opinion is based, first on incorrect extensions to frozen soils of the general rule in chemistry that the rate of chemical reaction depends on temperature and, second on a representation of frozen soils as systems in which water appears only as ice and vapour.

At the present time, when not only the presence of liquid water has been established in frozen soil but also the mechanisms determining the quantity of liquid water has been established, the question of physico-chemical processes in soils at temperatures below freezing is approached in a different way.

The interaction of water and dry soil is accompanied by the liberation of heat of wetting. Some scientists consider that this water interacting with the surface of soil particles converts into ice VI or ice VII (Marinesko, 1931; Parkhomenko, 1942; Shumskii, 1955<sub>2</sub>), others consider that it results in the formation of "solid hydration" (Priklonskii, 1949), an insoluble volume of moisture. The surface of soil particles is attributed the property of taking heat from the water and converting it into ice when they are both at the same temperature. However, crystallization of water on the surface of a solid particle is impossible since this contradicts the second principle of thermodynamics. What in such a case would be the heat of wetting? To answer this question one must proceed from the following principles established in interdisciplinary sciences.

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\* Written by I.A. Tyutyunov.

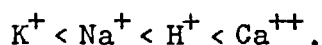
It is known that ions participating in the formation of a solid body, including crystals, introduce a specific energy quantity which is approximately equal to their hydration energy in dilute solutions.

Ions in a free solution give up all of their energy to the formation of a crystal lattice. Those that form the surface retain some of this energy. This unspent energy of the ions participating in the formation of a solid body is the surface energy. Since ions under a crystal face are arranged unevenly, the surface of a solid body is not uniform in energy. It has been established (Kuznetsov, 1952, 1954), that the quantity of surface energy is proportional to the area but also dependent on geometry.

The dissolving of a given substance and hydration of ions is possible because ions possess a great chemical affinity for water and a large electric field of force. Complex associations of water (Sokolov, 1955) falling into the sphere of influence of an ion lose their hydrogen bonds and convert to single molecules. The translational motion of such molecules falls abruptly owing to the orientation and increase in density around the ion (Kapustinski and Samoilov, 1950; Samoilov, 1952).

From these principles one can consider that the hydration of ions occurs also when the ions, on forming the surface of a solid body, cannot under the given thermodynamic state of the system abandon their position in the crystal lattice. However, even in this case ions manifest their potential surface energy in the form of the energy of chemical affinity and in the form of the energy of an electric field of force. They also disrupt hydrogen bonds, orient and compact water molecules in the sphere of influence of the field of force formed by their surface. With this action of the surface the water absorbs energy in the form of the energy of chemical affinity. Therefore, the heat of wetting is not the heat of crystallization of water on the surface of a solid particle but the residual of the heat of hydration of ions forming the surface of a solid particle.

Methods have not yet been developed for measuring the surface energy directly. Therefore, the importance of ions in determining the value of surface energy can be estimated by the heat of wetting which, as noted above, we call the residual heat of hydration. Data obtained for various clays saturated with cations  $K^+$ ,  $Na^+$ ,  $H^+$  and  $Ca^{++}$  (Ovcharenko and Bykov, 1954) indicate that with respect to the value of the heat of wetting, clay can be arranged in the same series as given by the heats of hydration of these ions in aqueous solutions:



Consequently soils saturated with sodium have a higher dispersion than soils saturated with hydrogen and multivalent cations. Having a greater specific

surface, they are characterized by a smaller specific surface energy in comparison with soils saturated with hydrogen and calcium.

Thus the liquid phase in non-saline frozen soils exist because it absorbs a large quantity of energy in addition to its heat of crystallization as the result of interaction with the surface of solid particles.

From the data of Jura and Harkins (1944) it follows that for one mole of adsorbed water to convert to ice one has to remove not 1,440 calories (latent heat) but at least 11,440 calories, since during crystallization there is the evolution of the heat of formation of two hydrogen bonds equal to ~10,000 cal/mole. Thus the latent heat of crystallization of adsorbed water is 635 cal/g and not 80 cal/g.

Consequently the absorbed water is not ice VI or ice VII but is a solution as was considered by B.I. Vernadskii. This solution we call the boundary phase of the soil system.

The boundary phase with respect to the layer of ions forming the surface of a solid particle plays the role of an electric field of force whose energy is equal to the energy of the rupture of hydrogen bonds of the hydrate. If there are any ions in the boundary phase then its effect as a field of force increases. The boundary phase in conjunction with the surface layer of a mineral particle is a so-called double surface layer, which is an exceedingly active and mobile, constantly varying part of the soil system.

The concentration of surface energy and the effect of the boundary phase as a field of force in the final analysis are the reasons for the rupture of the surface layer of mineral particles and its chemical decomposition which are intensified in a cryogenic system by the following factors:

- (1) the increase in the degree of concentration of surface energy with decrease in temperature,
- (2) the increase in the concentration of hydrogen ions in the boundary phase and
- (3) pressure due to the crystallization of water which increases as the temperature decreases.

Under the influence of these factors the ions forming the surface layer and particularly portions of the active surface (i.e. surfaces with the maximum concentration of surface energy), under the influence of displacement and the highly stressed state, come out abruptly from the "interstice" of the crystalline lattice. The place vacated by an ion is immediately filled by another, previously located in the boundary phase, thus producing an exchange between the solid and the boundary phase.

The presence of cation exchange in frozen soils was first shown in 1947 by I.A. Tyutyunov (1951) from data on measurements of decomposition of exchange ions in a number of soils during the winter.

These data necessitate a new approach to the problem of chemical erosion at freezing temperatures. In addition to cation exchange at below-freezing temperatures, there is chemical erosion of mineral matter of the soils similar to that occurring at higher temperatures.

Special experiments were set up to establish the reaction rate of cation exchange as related to temperature. The data of these experiments are given in Tables V and VI.

The data in Table V show that below-freezing temperature did not retard the reaction rate of exchange and did not result in a decrease in the quantity of substituted calcium as compared with that obtained for a relatively high temperature.

Under natural conditions under the effect of migration there is a continuous disruption of equilibrium in the exchange reaction. Ions coming out into the boundary and preboundary phases shift to the zone of maximum concentration of surface energy and there enter into chemical reaction with the solid phase and substances in solution.

The effect of the concentration of reacting solution on the yield of Ca ions was checked also on North Ural perennially frozen soils (Table VI).

The data in Tables V and VI show that below-freezing temperatures do not inhibit chemical processes, including the exchange of cations, and confirm the hypothesis that as the temperature decreases and the concentration of the solution increases exchange reactions are intensified. Field observations show that the chemical properties of soils in winter and summer are quite different (Tyutyunov, 1951).

In frozen peat-gley soil of the tundra region (Table VII) the degree of base saturation decreases with depth, whereas the degree of H-ion saturation varies inversely. The lowest degree of base saturation in the stratum at a depth of 47 - 57 cm is explained by the fact that during freezing the soil solutions migrate not only upwards but also downwards, i.e. towards the permafrost table. Its depth may vary during the winter and depends on temperature, which controls the gradient of chemical potential in the soil profile.

In the thawed state the absorption capacity and degree of base saturation increase with depth and the degree of H-ion saturation decreases.

The direction of ion exchange in the soil depends on the thermodynamic state of the soil system governed by heat exchange. This means that during the winter monovalent cations will enter the absorbing complex with greater energy and this process will be accompanied by heat liberation, whereas in the

summer multivalent cations will have the greater energy and exchange will take place with the absorption of heat (Tyutyunov, 1955). The consecutive substitution of cations at equal concentrations occurs in the direction of least thermal effects. This is a particular case of the principle of the transition of a system from the unstable state to the stable state as known in geochemistry.

Tundra soils develop under the influence of two opposing types of soil processes: one of these is characterized by the extrusion of the decomposition products of organic and mineral compounds and the other is the return of these to their initial zone. The effect of this annual cycle of soil chemicals and their interaction with the matrix determines the direction and particular features of soil formation in the tundra and arctic zones.

Prolonged investigations of the qualitative and quantitative changes of water-soluble substances in the soil made it possible to elucidate the unique mobility of individual components of the soil solution and seasonal changes in the colour of soil strata and also the upper layer of perennially frozen soil - from pale yellow and dark brown to blue-gray and gray-green. From this one can assume processes of oxidation and reduction of soil components at below-freezing temperatures.

Oxidation-reduction processes in soil and in permafrost were investigated by the compensation method using C and Zn as electrodes, comprising a sort of galvanic cell. These investigations, carried out with soils from the tundra showed that as the temperature decreased the energy of oxidation-reduction increased and when the temperature is increase from below-freezing to above-freezing the energy decreases.

For comparison similar investigations were performed on podzol soils of the Moscow region. Tests were made during the cold period when the microbiological activity in these soils were somewhat equivalent. It was found that with respect to oxidation-reduction energy the tundra and podzol soils were more or less equal, but in calcium exchange, according to the data of Z.A. Nersesova, there is more calcium exchange in the tundra soil than in the soil from the Moscow region.

The above-mentioned experimental investigations confirm the rapid rate of chemical reactions at below-freezing temperatures resulting from the increase in the concentration of surface energy as the temperature decreases. This can be supplemented by the following data.

For example, according to A. Puri (1949) the heat of wetting of soil increases as the temperature decreases.

The investigation of the adsorption of water vapour (Guany, 1954), ammonia (Morland, 1955) and oxygen (Dubinin, 1955) also indicates an intensification of

these processes when the temperature decreases and confirms the increase in concentration of surface energy with decrease in temperature. According to the data of Morland (1955) the adsorption of ammonia by Ca bentonite is greater than that of Na bentonite.

This dependence of the quantity of surface energy on the type of ions forming the surface is indicated by data obtained in the determination of heat of wetting for clay with varying exchange cations (Ovcharenko and Bykov, 1954).

From these considerations and from experiments it follows that clay and clay-loam soils that are hydrophobic at above-freezing temperatures become to some extent hydrophilic at below-freezing temperatures. For this reason they heave during freezing.

Thus the increase in the concentration of surface energy with decrease in temperature, leading to more intense adsorption, exchange, oxidation-reduction, and the formation of new interfaces resulting, on the one hand, in the physical decomposition of the initial substance, and on the other hand in physico-chemical aggregation.

The intensity of physico-chemical processes in soils is to some extent determined by the total of the specific surface which increases sharply in the presence of a colloid fraction. The most active part of soils is their fine fractions - clay and particularly colloid (with particle diameters less than 0.0002 mm). Therefore great importance is attached to coagulation and peptization and associated variations in the surface of the colloid fraction.

For hydrophilic colloids, as are most of the colloids of soils, one of the basic factors determining coagulation is the increase in concentration of electrolytes in a disperse medium. In freezing soils, as indicated above, the increase in concentration of electrolytes occurs because of changes in ice content or migration of moisture containing water-soluble compounds. The coagulation of colloids occurring at below-freezing temperatures is characterized usually by the rather inconvenient term "frost coagulation". Here one notes the reversible character of this type of coagulation. However, the siltyness typical of soils of northern regions, particularly of perennially frozen ground, is connected with the irreversible coagulation of the fine fractions as a result of variations in the concentration of the liquid phase.

Experiments carried out in the Central Laboratories of the Permafrost Institute with soils taken from below the boundary of seasonal freezing and thawing fully confirm what was said above. The determination of the grain size distribution of these soils showed a marked increase in particle size after freezing. This was particularly evident for soils containing soluble salts.

What has been stated above indicates that on the earth's surface the extensive distribution of processes involving changes in the soil (supergeneses) unquestionably also applies to permafrost. At below-freezing temperatures and during periodic changes in soil pressure due to seasonal freezing and thawing, the supergenic processes are extremely active since the liquid phase usually contains large quantities of free oxygen and carbonic acid. The supergenic processes at below-freezing temperatures are specific processes and they should be identified as a separate type - cryogenic processes. Consequently cryogenesis is the sum total of physico-chemical mineralogical changes and the transformation of the lithosphere and hydrosphere at temperatures below freezing. The particular features of this type of process is the formation of cryogenic minerals and the general accumulation of products of cryogenesis and extremely low losses due to erosion.

#### Physical and Physico-mechanical Processes in Freezing and Frozen Soils

The freezing of moist soils is accompanied by a large number of physical and physico-mechanical processes and phenomena. When the pore water freezes there is first of all a disturbance in the moisture equilibrium and moisture migration begins primarily in the direction of the freezing front. The distribution of moisture in fine-grained soils varies greatly during freezing and in the frozen state. There is a variation in the general distribution of moisture as well as in local distribution (in the freezing zone) where there is differentiation forming ice strata, lenses and inclusions as well as mineral strata from which, as a rule, part of the water is removed during migration.

This results in compression of the mineral strata of freezing soils (particularly clay soils) which in turn has a great effect on the mechanical properties.

Moreover, during the freezing of fine-grained soils a number of physico-chemical phenomena take place connected with the migration of moisture accompanying cation exchange between the mineral skeleton and the liquid phase. Of considerable importance here is the changes in concentration of dissolved salts which disrupt the equilibrium of cation exchange, coagulation and microaggregation.

The migration of moisture in freezing and frozen soils increases their volume and results in the heaving of the soil which is frequently uneven. This causes damage to roads and airport runways.

Physico-mechanical processes in freezing and frozen soils and the accompanying physico-chemical phenomena result in a particular texture of the soil which determines its mechanical properties.



As the temperature of frozen soil increases to the thawing point, changes in volume occur which for soils with a high ice content result in substantial settlement and at times collapse (rapid local settlement accompanied by the extrusion of the thawing mass). Substantial settlement and collapse of thawing soils with high ice content seem to occur under their own weight without the application of additional load.

We note also that there is a great deal of importance attached to the temperature deformation of frozen soils, particularly crack formation resulting from pressures owing to the uneven variation in volume.

Migration of water in freezing soils.\* The initial accumulation of ice in freezing soils was explained by the majority of authors by the condensation of water vapour in the freezing stratum (Blizin, 1890; Izmail'skii, 1894; Bogdanov, 1912; Chirikov and Malyugin, 1926; Lebedev, 1913; Tyuremnov, 1924; Sambikin, 1926; Bastamov and others, 1933; Belokrylov and others, 1931; and others). Nevertheless many considered that the movement of moisture in freezing soils occurs in a liquid form and among the authors who held this point of view some considered that migration occurred under pressure, since on freezing a closed system is formed, others explained the redistribution of waters by suction in the form of thin films towards the freezing front.

The theory of migration under pressure was developed by S.A. Pod'yakonov (1903), V.N. Sukchaev (1911), K. Nikiforov (1912), R.I. Abolin (1913), D.A. Daranitsin (1914) and M.I. Sumgin (1927, 1929, 1931, 1940). The hypothesis of the suction of film moisture towards the freezing stratum from lower unfrozen strata was first advanced by V.I. Shtukenberg (1885, 1894). Later the hypothesis of film transfer of moisture received particular attention which is reflected by the papers of A.F. Lebedev (1930), Zh. Bouyoucos (1923), S. Taber (1930), G. Beskow, (1939), M.N. Gol'dshtein (1940-1948).

However, in spite of the many investigations the physical nature of water migration cannot be considered elucidated at the present time. Below we give the foundations for a physico-chemical hypothesis to explain the migration of moisture in freezing soils (Tyutyunov, 1955).

At the moment of contact between soil particles and water the energy of solid particles, which manifests itself in the form of the energy of chemical affinity and the energy of an electric field of force, is spent on the rupture of hydrogen bonds and on the packing of water molecules. Water finding itself on a portion of the surface of a particle and having undergone the appropriate orientation and packing becomes more mobile with respect to the specific strata of the water and under the influence of even weak fields of force occupies the

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\* Written by A.I. Tyutyunov.

entire surface of the solid particle. Immediately on formation this film of water comes under the influence of the field of force of a neighbouring particle and under its influence undergoes chemical affinity towards this particle and shifts along the surface of the particle.

The interaction of the surface of a mineral particle with water is, in fact, the process of hydration accompanied by the formation of a system whose chemical potential is less than the sum of chemical potentials of the initial substances.

If from the arrangement of particles capillaries of various cross-sections are formed, the films covering the particles shift faster, the smaller the capillaries (Kin, 1933).

The size and arrangement of capillaries under specific thermodynamic conditions depend not only on the shape of the surface of mineral particles but also on the structure of the hydrates. Here the difference in chemical potentials of the water and the total surface of the mineral particles is the reason for the movement of water in a soil system.

If there is no interaction between the water and mineral particles possessing a maximum chemical potential of the total surface for the given thermodynamic conditions, the water will not rise in the capillaries; it will only be retained. Thus capillaries are only migration paths and it is incorrect to say that capillary phenomena act in the opposite direction to film transfer of water. It is likewise incorrect to consider them as motive forces in the distribution of water in the soil (Bozhenova, 1957). The process of water migration in freezing soils occurs in two steps which are interrelated but differ in thermodynamic and physico-chemical features.

The first step is characterized by the fact that at the instant of interaction with mineral particles and with subsequent space displacement the water passes to a higher energy level since the energy of chemical affinity, manifested by the surface in the form of heat, is basically absorbed by the water and only some part of it is liberated as residual heat, called in the literature "heat of wetting". The absorption of energy is accompanied by disruption of hydrogen bonds, the orientation of single molecules and their compression in the electric field of force of the surface. As a result the free water is transformed into the boundary phase.

The second step results from the transition of the water of the soil system to a lower energy level and the formation of ice.

Under the influence of pressures due to the formation of ice, as a new solid phase of the system, transformation of the water of the boundary phase proceeds, accompanied by a reduction in its internal energy. First of all the initial hydrogen bonds are reconstituted and the resultant evolution of energy

goes to reconstitute the chemical potential of the surface of the mineral particles. In the formation of ice grains the water of the boundary phase passes to a lower energy level since an additional atomic bond is formed in the lattice of the growing ice crystal.

At the instant of formation and crystallization of the ice at the expense of the boundary phase, there is an increase in internal pressure of the soil system which results in the transformation of the peripheral part of the boundary phase into triple hydrate and then into microcrystals of ice. After reduction in pressure and the transfer of the heat evolved, the boundary phase again thickens and along with it the preboundary phase is reconstituted at the expense of water from the homogeneous phase. If the precipitation of ice in a cryogenic system proceeds for a long time and the reconstitution of the boundary and preboundary phases takes place at the expense of the homogeneous phase, thick layers of ice can accumulate in the soil system. When the homogeneous phase is absent the ice forms as thin veins and ice cement. Of course ice cement frequently forms also in the presence of the homogeneous phase but this occurs only when there is no equilibrium between it and the zone supplying water during the period of ice formation.

Thus the motive force of migrating water in the soil system, when there are changes in the external medium, is the difference in chemical potential of the interface between the solid and liquid phases of the system. The effect of the temperature gradient on migration occurs because the chemical potential of the solid and liquid phase increases as the temperature decreases, thus bringing about their interaction\*. This motive force is the most important property of the system itself but it manifests itself in different ways depending on external conditions and primarily on heat exchange. The chemical potential of any cryogenic system is a factor of the intensity of its chemical and energy transformations; it determines the direction and the limits of these transformations.

The migration process continues and manifests itself when the soil system is characterized, first by the presence of a difference in chemical potential between the mineral part and the water, second, by the formation and separation of the new phase, and third, by the exchange of matter and energy.

The heaving of soil during freezing\*\*. Heaving of soil is the name given to uneven increase in volume during freezing which takes place owing to the

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\* There are other points of view regarding the reason for the movement of ground water during freezing (Editor's note).

\*\* Written by Z.A. Nersesova and I.A. Tyutyunov.

expansion (~ by 9%) of the volume of the water contained in the soil, as well as because of the freezing of new volumes of water advancing towards the freezing front due to moisture migration. Because the composition of the soil is not uniform, the distribution of moisture is not even and there is a difference in the hydrological conditions, one hardly ever sees uniform increase in volume of soils during freezing.

If heaving of soils is only due to pore moisture without the inflow of water from the outside, even in the presence of moisture redistribution, it is called heaving in a closed (or locked) system (Taber, Sumgin et al.). This heaving is, as a rule, not large and is measured in values of the order of tenths of a percent up to percentage units of the depth to which the ground has frozen.

As the data given show, the value of the relative heaving (the ratio of the height to which the soil rises to the depth of freezing), when there is uniform freezing only from the top downwards, is less than when there is freezing from all sides; as the temperature decreases the value of relative heaving decreases. This is explained by the fact that at low temperatures there is not only migration of moisture but also its fixation which at very low temperatures will have the prevailing effect.

We note that according to the laboratory experiments of scientists studying the physics and mechanics of frozen soils at the Permafrost Institute, in compact clay soils the migration of moisture and heaving begins only when there is a rather high initial temperature gradient. Heaving in sand soils when there is free drainage for the water is very close to zero.

When there is an inflow of water from the outside, i.e. when the soil freezes under conditions of an open system, vertical heaving of the ground connected with the migration of moisture towards the freezing front and its fixation in the form of ice in the freezing layer increases substantially, reaching a high percentage of the depth of freezing.

The accumulation of ice in soil and resultant heaving depends on the conditions of migration in which the most important role is played by the chemical potential gradient, as shown by experiments. When the chemical potential gradient is zero in the soil system there is no segregation of ice nor is there any exchange of matter with exchange of energy.

Since the chemical potential of the surface of mineral particles is determined by the chemical potential of ions forming the surface, for soils containing different cations the intensity of migration during freezing should increase in correspondence with the increase in hydration energy of these cations. This is fully confirmed by experiments, the results of which are

shown in Table IX where the distribution of moisture and the magnitude of heaving of specimens of surface clay loam saturated with  $\text{Fe}^{+++}$ ,  $\text{Ca}^{++}$ ,  $\text{Na}^+$ , and  $\text{K}^+$  are given.

Analogous data were obtained earlier by A. Dücker (1939), who showed that during the freezing of a mixture of bentonite clay saturated with Ca and Na cations and powdered quartz there was more heaving with the Ca specimens than with the Na specimens. However, he gave no explanation for this phenomenon.

Specimens of surface clay loam saturated with  $\text{Fe}^{+++}$ ,  $\text{Ca}^{++}$ ,  $\text{Na}^+$  and  $\text{K}^+$  cations were frozen at an air temperature of  $-3$  to  $-5^\circ\text{C}$  with the possibility of moisture flow to the area of freezing. This provided conditions favourable for the migration of moisture and ice formation.

Data on the total moisture content of specimens after freezing and the amount of heaving are given in Table IX. Figure 6 shows photographs of specimens of  $\text{Fe}^{+++}$ ,  $\text{Ca}^{++}$ ,  $\text{Na}^+$  and  $\text{K}^+$  containing clay loam at the end of the experiment. The cryogenic texture of the specimens clearly shows the effect of the type of exchange cations on moisture migration and heaving; it would be sufficient to indicate that the ice veins formed in the Fe specimen comprised 35.3% of the height of the specimen whereas in the Ca specimen it was 3.2%, and in the Na and K specimens there are practically no horizontal veins and practically no heaving.

Variations in the volume of soils on freezing and the formation of frost fissures\*. Changes in volume of frozen soil may be due to various processes and they may be subdivided into three types: temperature, diagenetic and migration. The first type includes changes in volume of frozen soils when the temperature varies within the range of below-freezing temperatures, the second includes changes in volume during crystallization of water in soil and also during drying of the soil and the third includes changes in volume due to the accumulation of ice connected with migration of water.

Temperature variations in the volume of soil. Assuming a linear relationship between temperature pressures and the resultant diurnal shifts in the surface, and assuming that the deformation of the soil due to cooling is proportional to the temperature gradient and the shear stresses are distributed uniformly along the plane parallel to the freezing plane for a soil mass limited by horizontal and vertical surfaces (Dostovalov, 1957), it was shown that the value of the shear stress  $\tau_x$  in the daily surface of the soil is described by the expression:

$$\tau_x = \frac{1}{2} \alpha \alpha C x \text{ grad}_z(t), \quad (1)$$

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\* Written by B.N. Dostovalov.

where  $n$  - a coefficient less than unity, and for small deformations it is very close to unity;  
 $\alpha$  - the temperature coefficient for linear expansion of the soil;  
 $C$  - the shear modulus for frozen soil;  
 $x$  - the distance from the free vertical surface to the cross-section under consideration;  
 $\text{grad}_z(t)$  - the temperature gradient at a given depth, in this particular case with a linear variation in temperature at a depth equal to  $\frac{\Delta t}{h}$  (Fig. 7).

Comparing  $\tau_x$  to the temporary resistance during displacement, taken from expression (1), one can determine the distance corresponding to the appearance of fissures.

This distance is equal to:

$$x_B = \alpha \frac{2\tau_B}{C \text{ grad}_z(t)}. \quad (2)$$

Calculations show that the distance between parallel fissures obtained with formula (2), agree well with those observed under field conditions.

From formula (1) the following mechanisms are derived:

1. Since the stresses  $\tau_t$  are proportional to the distances from the free vertical surface, in a homogeneous soil mass a system of parallel fissures should form which would divide the soil mass into a number of strips and the strips should develop transverse fissures dividing the mass into rectangular blocks.

2. The transverse fissures must be perpendicular to the longitudinal fissures by virtue of the temperature gradient stresses being proportional.

3. Since the rupturing stresses are proportional to the products  $(x \text{ grad}_z t)$  it follows that, in the first place, as the gradients increase the rectangular block formed must subsequently divide in half by fissures of increasing order, secondly, at a given gradient masses larger than a specific size cannot exist (Fig. 8).

4. In the union of fissures in the form of a base and a perpendicular to it, the base is the first to form resulting in a rather long fissure of a lower order whereas the perpendicular forms later as a relatively short fissure of a higher order.

5. In heterogeneous soil masses there are variations in individual fissures and they are no longer parallel but the united fissures remain perpendicular.

6. The formation of fissures of higher order are less probable than the formation of fissures of lower order, since high values for the temperature gradient are less probable than are low values.

Diagenetic variations in the volume of soils are due to various physico-chemical processes including the drying and swelling of fine-grained soils.

In the north, fissures due to drying develop extensively each summer on silt-clay loam portions of flood plains which are exposed after the flood waters have receded.

The approach to an approximate calculation of stresses arising in soils because of non-uniform drying can be analogous to temperature stresses since the difference lies only in quantitative characteristics. Instead of a linear contraction coefficient on cooling ( $\alpha$ ) one must take a linear contraction coefficient due to drying ( $k$ ) and instead of a temperature gradient we have a moisture gradient along the vertical ( $\text{grad}_z W$ ). We then have an expression analogous to formula (1):

$$\tau_x = \frac{1}{2} nk \cdot C \text{ grad}_z W. \quad (3)$$

Since the coefficient of linear settlement ( $k$ ) is much larger in absolute value than the coefficient of linear expansion ( $\alpha$ ) parallel diagenetic fissures form at smaller distances than in the case of temperature stresses.

They are also observed in frozen clay soils but in this case they are frequently separated by horizontal and vertical ice veins forming the so-called "frost texture". Here the formation of the horizontal layers is explained by the migration of water under the influence of the temperature gradient towards the freezing front and the vertical layer can be hypothetically explained by the formation of a network of vertical fissures resulting from the moisture gradient arising from the formation of horizontal ice layers which "dehydrate" the adjacent volume of soil. Moisture migrates to these fissures and freezes forming vertical layers.

Variations in volume of soils during ice crystallization and cracking of frost mounds are also explained by expansion stresses arising because of the non-uniform crystallization of ice in freezing fine-grained soils because of water migration or heaving. Moreover the formation of frost mounds in clay soils under natural conditions changes the course of isothermic surfaces and directs the migration of moisture under the mounds thus ensuring their continued growth; here the uppermost layer of soil covering the growing ice inclusion is continually bent upwards resulting in stretching and cracking.

#### Variations in the Properties of Soils on the Freezing and the Thawing

With the freezing of water in soil, as was considered above, a unique cryogenic texture of the soil forms, there is a decrease in water permeability

and an increase in the cementing of mineral particles by ice. The physico-mechanical properties of soils undergo considerable changes and the greatest changes occur in the region of intense phase transformation of the water.

Since for moist sandy soils this region extends only from 0 to  $-0.2^{\circ}\text{C}$  their properties change to the greatest extent within this temperature range, i.e. during the process of freezing; for clay soils, particularly compact soils, the temperature region of intense phase transformations is extended to  $-5$ ,  $-7^{\circ}\text{C}$  and even lower and these soils undergo a less rapid change in physico-mechanical properties, remaining plastic over a rather large range of below-freezing temperatures.

With thawing of frozen soil the cementing bonds of the ice are weakened which has an evident effect on the strength of the soil; however, the cryogenic texture of soils even after complete thawing does not disappear completely but residual effects remain.

Changes in some physical and physico-mechanical properties of soils as a result of freezing and subsequent thawing are considered below.

Electric properties\*. The freezing of soils greatly affect their specific resistance ( ); and dielectric permeability ( $\epsilon$ ) of soils varies insignificantly during freezing and the magnetic permeability ( $\mu$ ) is practically negligible and therefore will not be considered further. The conductivity ( $\sigma = \frac{1}{\rho}$ ) of frozen soils is basically ionic and is determined mainly by the conductivity of the liquid phase and by its quantity, i.e. it depends on temperature, pressure, ion concentration and the total of the specific active surface of the soil skeleton and also on the conductivity of the skeleton and the distribution of ice inclusions.

Under natural conditions all the above-mentioned factors affecting the electric conductivity of frozen soils have interrelated variations and therefore the few laboratory experiments carried out up to the present time on some twofold relationships can be considered as being only particular cases.

The dielectric permeability ( $\epsilon$ ) of soils depends basically on the chemico-physical state of the moisture, on the quantity of the liquid phase in the frozen soil and on the frequency of the electric field ( $f$ ). The dielectric permeability of water is approximately 80 CGSE whereas that of ice is approximately 2 CGSE. The dielectric permeability of frozen soils varies within these limits depending on all the above-mentioned factors affecting the quantity of the liquid phase.

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\* Written by B.N. Dostovalov; for more details see No. 3 "Materialy po Labor. Issled.....", 1957.



The effect of variations in temperature ( $t$ ) and current frequency ( $f$ ) on the dielectric permeability of ice are shown in Fig. 9.

Figure 10 shows the dielectric permeability of Jurassic clay and fine-grained sand with various moisture contents.

Physico-mechanical properties\*. The physico-mechanical properties of soil undergo substantial changes during freezing. The nature of these changes depends on the properties of the soil and the conditions of freezing.

If freezing is accompanied by moisture migration such indices as total moisture content, unit weight, porosity\*\*, always vary. Here one can have two cases:

(1) freezing is accompanied by the additional inflow of moisture as a result of which the total moisture content and porosity in the freezing portion of the soil change substantially and the unit weight decreases;

(2) freezing is accompanied only by the redistribution of moisture.

The increase in total moisture content and porosity and decrease in unit weight is observed only in the upper strata of the soil adjacent to the cooling surface; in the strata removed from the cooling surface, moisture content and porosity decrease and the unit weight correspondingly increases.

However, for water-saturated sand the redistribution of moisture will be quite different; in soil strata adjacent to the cooling surface the moisture content decreases; in more distant strata, depending on the conditions under which the water is squeezed out, there may be either an increase or a decrease in moisture content (Khakimov, 1957; Tsytovich, 1954).

If the freezing of soil occurs only with the fixation of moisture without redistribution the total moisture content varies very little; the increase in unit weight and porosity is not large (of the order of 3 - 4%).

With freezing there are substantial changes in the water properties of soils.

Soils that are completely saturated with water become practically impervious to water. This is particularly valid for sands; clay soils, however, do not become absolutely impervious to water because of the presence of liquid water in them. However, before there can be any filtration in such soils, as shown by V.I. Ivanov (Materialy....., 1957), a very high pressure gradient is necessary.

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\* Written by N.A. Tsytovich, S.S. Vyalov and E.P. Shusherina.

\*\* Total moisture content is understood to include moisture due to the liquid phase, ice cement and ice inclusions; porosity is understood to mean the ratio of the volume occupied by air, liquid phase and ice to the total volume of the soil.

With the freezing of soils and partial filling of the pores with water their water impermeability decreases substantially although frozen soils always retain some water impermeability. Of greater importance is the effect of freezing on the strength of soil. Frozen soils have much higher resistance to mechanical action than thawed soils.

The strength of soils depends on the strength of the internal bonds.

With freezing a new type of bond arises, i.e. the cementing of mineral components by ice. This type of bonding is predominant and is much greater than bonding in thawed soils. For example, the bonding of unfrozen clay loam has a value of the order of  $0.2 - 0.7 \text{ kg/cm}^2$ . After freezing, the instantaneous bonding increases to  $5 - 7 \text{ kg/cm}^2$  (at a temperature of  $-1$  to  $-2^\circ\text{C}$ ).

There is also a substantial increase in shear resistance. For rapid shear (at a normal pressure of  $4 \text{ kg/cm}^2$ ) the resistance of unfrozen clay loam is  $1.0 \text{ kg/cm}^2$ , whereas the resistance to rapid shear of the same soil in the frozen state was determined at  $\tau = 6.5 \text{ kg/cm}^2$  (when  $t = -0.8^\circ\text{C}$ ) and  $9.0 \text{ kg/cm}^2$  (when  $t = -2^\circ\text{C}$ ).

The long-term strength of frozen ground is also as a rule higher than that of unfrozen ground. For example, the long-term strength of frozen sandy loam is  $1.0 - 2.5 \text{ kg/cm}^2$  (at a temperature of  $-0.4$  to  $-4^\circ\text{C}$ ), whereas in the unfrozen state the same soil has a bonding strength of the order of several hundredths to several tenths of a  $\text{kg/cm}^2$ . In correspondence with their greater strength frozen soils deform much less than unfrozen soils.

The compaction of frozen soils under prolonged load is insignificant; it comprises up to a few percentage points of the total thickness of the stratum under compression; the compression of unfrozen soil may be as high as 20 - 30%.

It should be noted that the lower the temperature of frozen soil the greater its strength and the less its deformability.

#### The Structure of Frozen Soils\*

In addition to its composition frozen soil is characterized also by specific structure, i.e. the space arrangement of its component parts. In the structure of various strata of the earth's crust it is customary to differentiate between two aspects - texture and structure - which unite the space characteristics of a different scale\*\*.

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\* Written by P.A. Shumskii. Published in No. 3, "Materialy po Labor. Issled.....", 1957. In this chapter the report is published in a somewhat abbreviated form.

\*\* There is no complete unity in the concept of these terms as used by various authors in various branches of petrography, not to mention such related sciences as soil science and soil mechanics.

The particular features resulting from the size and shape of component parts of soils, and for some soils from the relative quantity and orientation of the component parts, is called structure. Texture is usually defined as the space distribution of the soil component and the extent to which the space is filled by the soil materials; it includes characteristics of a comparatively larger scale, particularly stratification, schistosity and character of differentiation. In essence the presence of texture is always connected with a specific space differentiation of component parts of the soil and the absence of such differentiation corresponds to lack of texture (the so-called "massive texture").

The component parts of the soil whose space arrangement determines its structure, are the mineral crystals and grains, liquid and gaseous inclusions, and in the case of organic-mineral soils, also the organic formations. In frozen soils crystalline grains of ice become attached to the high temperature minerals which form the distinguishing characteristics of frozen soils.

As a rule the ice, as a component of soil, is not formed simultaneously with the other solid component parts ("the skeleton") but is formed subsequently upon freezing of soil moisture. The appearance of ice changes the primary structure of soil. New specific structural features appear which are frequently called cryogenic structure and texture. In using these terms one should keep in mind that cryogenic structure, and texture is not independent and isolated from the primary structure but arises from the latter and in conjunction with the primary features forms a single interrelated system of structure and texture of frozen soil.

The structure along with the material components gives frozen soil its mechanical and physical properties and at the same time is a direct visible expression of the results of soil formation and transformation. The study of structure is necessary to understand the genesis of frozen soils and to be able to control processes of formation and changes in the properties of frozen soil.

The structure of frozen soils has not been investigated extensively. The study of the texture of frozen soils, if one excludes a few random early papers, began only in the 1920's whereas the structure of fine-grained soils has not been studied at all.

The formation of frozen soils. In the complexity of processes involved in the formation of frozen soils we need to deal here only with those that are directly connected with the cryogenic features of structure - the

crystallization of water within freezing soil\*, or in other words, the formation of component ground ice (see Chapter IX).

The type of cryogenic structure and texture depends primarily on the effect of crystallizing water on the skeleton of the soil. This effect in turn results from the primary composition and structure of the soil (including of course moisture in the concept of composition) and the conditions of freezing.

The crystallization of water - the regrouping of its molecules into a crystalline lattice of ice - is accompanied by an increase of volume by 9.07% and a decrease of density by 8.31%. These changes disrupt the previously existing distribution of moisture in the soil. This question is discussed in more detail in a paper published earlier (Shumskii, 1957).

The flow of film water toward the freezing front when there is an insufficient reserve of free water in fine-grained soil results in dehydration of particles, coagulation of colloids, internal compression, aggregation and the formation of cracks in front of the freezing front which subsequently become filled with ice. All of these changes have an effect on the structure of frozen soil.

If the water more or less fills the pores of the soil skeleton and is limited by the walls of the pores but not from all sides, then the nature of crystallization will depend on the resistance of the skeleton to displacement. When the resistance is large the ice crystals are limited in growth by the free space of the pores. The excess of free water resulting from the increase in the volume during freezing is squeezed along the pores of the soil. If the pores are fine and all the water is bonded by the surface of the skeleton the squeezing effect is counteracted by the process of film migration.

If a closed system arises - some volume of water or moist soil is limited on all sides by a water-impervious shell in which the free space of the pores is less than the increase in volume of freezing water - a pressure may develop reaching 2.5 thousand atmospheres at a temperature of  $-22^{\circ}\text{C}$  and lower (Tamman, 1922), so that in practice it is limited only by the resistance of the rigid shell which must unavoidably break. The same effect may occur with the flow of ground water under hydraulic pressure in the path of which there may be an

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\* The crystallization of water vapour in the formation of structure of frozen soils is of very limited significance - only for coarse-grained soils containing very little water - but the formation of ice from liquid water and vapour penetrating from the outside into cavities of frozen soil leads primarily to the formation of ice lenses, i.e. single-mineral ice inclusions but not the formation of structure in polymineral frozen soils.

obstruction formed by impervious frozen soil. In such a case the displacement of the soil skeleton is brought about by the water squeezed out by the ice and subsequently freezing in the resultant cavities. Such a process of injection into a closed system can overcome exceedingly high resistances of the surrounding soil.

When the resistance of a soil skeleton is not large, the ice crystals in an open system under some conditions may not be limited in growth to the free space of the pores but displace the obstacles in the form of soil particles and continue to grow in the direction of freezing owing to the attraction of film water. In a primary sedimentary soil there may be imposed differentiation due to crystallization with the formation of ice and the suction of film water to the place where ice is being formed. In order to have crystal growth accompanied by displacement of soil particles one must have the combination of the following conditions:

- (1) the soil skeleton must be hydrophilic\*;
- (2) the presence of reserve water capable of filling the liquid films between ice crystals and mineral particles;
- (3) not too large a resistance to displacement of soil particles;
- (4) small size particles in the soil skeleton and the slow rate of freezing (since an increase in either one of these factors requires a higher rate of water flow over a given cross-section of film).

Thus ice formation is possible only in uncemented fine-grained soils with a relatively high moisture content freezing slowly and lying not too far from the surface.

Ice formation can take place only in clays, clay loams, sand loams, and at times in fine sands and also in peat and peat-containing soils. The upper limit in size of mineral particles permitting ice formation is 0.05 mm and in the most favourable conditions 0.1 mm (Beskow, 1935; Dücker, 1939). The moisture content of the soil must be above the maximum molecular moisture

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\* If the material of the soil skeleton is hydrophobic, which in natural fine-grained soils is very rare, a small amount of ice formation under otherwise favourable conditions can take place because of the hydrophilic properties of ice. The supply of water to a growing ice crystal takes place by means of film water adsorbed by the ice crystals, and the attraction of moisture towards the region of ice formation is provided for by the internal molecular bonding of water resisting a break in continuity when the water reserve is exhausted in the adjacent pores of the soil. As was shown by investigations of the V.A. Obruchev Permafrost Institute ("Materialy po Labor. Issled....", 1957), this mechanism of ice formation and water migration, considered by F. Taber (1930, 1943) to be the only one and rejected without foundation by a number of other scientists (Gol'dshtein, 1948), in fact does exist but plays an insignificant secondary role.

capacity below the limit of plasticity (Fedosov, 1940; Razumova, 1951; and others). Very large ice formation can occur when the given section of soil is supersaturated or when adjacent strata have large reserves of free moisture capable of migrating to the freezing layer. The soil should not be firmly cemented or very compact; these factors increase internal bonding and decrease moisture content. In fine clay loams and clays ice formation can frequently overcome rather high external pressures, up to 15 - 16 kg/cm<sup>2</sup>, which corresponds to a soil depth of 100 m (Taber, 1943), whereas in coarser soils a much lower load is required to inhibit ice formation. The rate of freezing is of great importance. Very rapid freezing is not accompanied by ice formation even in supersaturated clays as was shown in experiments using liquid air as a coolant (Jung, 1932). Under natural conditions, however, the rate of freezing for soils of 3 - 1 mm/hr is usually less than that required to inhibit ice formation in clay loams and clays by a factor of 10 (Beskow, 1935).

Thus, depending on the properties of the soil and conditions of freezing, ice formation can take place only inside the pores or it is accompanied by displacement of the soil skeleton and the formation of inclusions or ice schlieren\*, which are larger than the cavities that existed previously. In the first case the water is fixed in the pores of the soil giving rise to ice cement and in cracks - ice lenses. In the second case, water migrates to the region of ice crystallization. The formation of ice inclusions can occur in two ways: (1) by injection into the soil, i.e. by the intrusion of water under pressure along the boundaries of frozen layers in closed systems and (2) by means of crystallization differentiation, i.e. by ice crystallization with sucking in of film water in open systems. Ice formed by the first method is called injected ice and that formed by the second method segregated ice (see Chapter IX).

In nature a very common type of ice formation is that which occurs in pores with cementing of the soil by ice to which, under favourable conditions, there is added ice crystallization and ice intrusion. Under the influence of these processes the texture and structure of frozen soils is formed.

Texture of frozen soils. The space distribution of mineral compounds (or organic-mineral compounds) cemented with ice and schlieren of constituted ice and also mineral compounds of varying composition and structure (from the primary texture of the soil) give the texture to frozen soil.

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\* Schlieren in petrography refers to mineral bodies crystallized in the process of differentiation during solidification of molten rock of complex chemical composition (constituted schlieren) and also the solidified products of the intrusion of fused rock into other rock (injected schlieren).

There is no accepted terminology at the present time for the texture of frozen soils. Because the formation of frozen soil is of a specific nature, none of the terms used in petrography can strictly be applied fully to frozen soils. However, we consider it expedient to retain the terminology current in petrography, giving preference to morphological indicators and at the same time trying to minimize any change in the genetic content of the concept\*. Proceeding from this, depending on the presence, form and distribution of ice streaks, one can identify three basic types of texture for frozen soils:

- (1) massive texture - ice schlieren absent;
- (2) stratified texture - ice schlieren in the form of parallel lens-shaped strata (Fig. 11);
- (3) net-like texture - ice schlieren of different shape and orientation forming more or less a continuous network or lattice\*\*.

The second and third types are represented by a large number of textures which can be subdivided as follows:

(a) depending on the size of the schlieren, into micro-stratified, finely stratified and coarsely stratified or micro-latticed, fine latticed and coarse latticed;

(b) depending on differences in density of distribution and in sizes, they can be subdivided into uniform and non-uniform stratification, uniform and non-uniform lattice (Fig. 12);

(c) depending on the space orientation of the schlieren, the subdivision may be into horizontal, diagonal and vertical stratification, rectangular, obtuse-angled and irregular lattice.

In between the above-mentioned types of texture there are also various intermediate textures forming a continuous series of transitions.

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\* The taxite group of textures is genetically the closest to that of frozen soils but the terminology is insufficiently expressive with respect to morphology and is difficult for non-petrographers.

\*\* Three types of frozen soils according to texture were first identified by Kokkonen who differentiated between massive, stratified and "porous" (hohlräumige) frozen soils (Kokkonen, 1926); the latter category corresponded morphologically to the lattice texture but genetically took into consideration only a particular case of filling cavities with ice during the freezing of soils. A.M. Pchelintsev called massive textures - "fused", stratified texture - "schistose", and lattice texture - "conglomerate-brecciated", and also identified a fourth type of texture - "sphenoid" which he subsequently rejected. K.A. Baranov (1946) used the term "cryomeshed" and N.A. Tsytoich called it "cellular". Most of these terms should be rejected because there are more common and more expressive terms. Regarding such terms as "schistose" and "conglomerate-brecciated" textures, they are completely inapplicable since the second term does not correspond to genesis and the first does not correspond to the morphology of the formations they are intended to represent.

The appropriate terms are not so much for classifying as for describing textures and their use where necessary should not replace a graphic description and a more precise quantitative characterization of dimension, shape and distribution of texture-forming elements.

Genesis of texture. The main role in the formation of texture of frozen soils is played by ice formation and the simultaneous formation of segregated ice. Injected and vein ice, if they exist, are of secondary importance.

To explain the reasons for the generation of a particular type of texture in frozen soils owing to striated ice segregation, a large number of factors have been considered: (1) the decrease in temperature of crystallization in fine pores and the generation of ice crystals primarily in large pores; (2) non-uniformity of the soil, the presence of discontinuities and layers of weak bonding; (3) the grain size composition of the soil; (4) the mineral composition; (5) the base exchange composition; (6) the compressibility of the soil; (7) the magnitude of settlement and tendency to the formation of fissures on freezing; (8) initial moisture content and migration rate of the water; (9) the orientation of isothermic surfaces; (10) the size of the temperature gradient; (11) the ratio of heat transfer rate to the evolution rate of crystallization heat of the water attracted to the freezing zone; (12) differences in the resistance of the soil to displacement in various directions, and finally (13) chaotic orientation of crystal nuclei, anisotropy of the growth rates of crystals and the segregation of crystals resulting from differences in growth (Bouyoucos, 1923; Kokkonen, 1926; Taber, 1929, 1930; Beskow, 1935; Bykov and Kapterev, 1938; Moroshkin, 1933; Rückli, 1943; Shumskii, 1955,).

Qualitatively there can be no doubt that each of the above-listed factors have some effect on the formation of soil texture during freezing. There can likewise be no doubt that their effect is interrelated, as a result of which it is exceedingly difficult to isolate and estimate quantitatively the role of each factor. The brief characteristics are given below for the fundamental and more elucidated factors.

Ice crystals are generated first of all in the larger pores of the soil where the bond between the particles and the water is relatively weak. The crystal nuclei are orientated in a disorderly manner and because of the internal properties of the crystal lattice of ice (because of the lesser reticular plane and greater surface energy of the faces of the prism) there is a tendency for the crystals to grow primarily in the base plane (Shumskii, 1955,). However, the conditions for the growth of various crystals are different depending on their space orientation, their distribution with respect to local non-uniformities in the soil and to neighbouring ice crystals; further, depending on the time the nucleus is generated in relation to its neighbours



and also on the properties of the soil including its initial moisture content, and on the rate of heat transfer away from the freezing zone. Under the influence of these factors during the freezing process there is a selection of ice crystals; some of them grow to substantial sizes displacing the mineral skeleton of the soil and connect together in groups in more or less large aggregations; others remain fine, cementing the particles of soil together.

The crystals that are first to be generated are usually the ones to grow; they are able to consume the local water reserves from the surrounding soils.

In supersaturated fluid soil the crystals grow at the expense of the surrounding water easily displacing the mineral particles in any direction. In such a case the orientation of the crystal nuclei is of no importance, all the nuclei generated in the early stages of the process grow along the base plane forming a disordered lattice of platelike single crystals with cells of soil between them and the faster the freezing the finer the mineral cells. Depending on the size of the temperature gradient the growth of the lattice of single crystal ice streaks takes place in a more or less narrow zone of freezing but in time this zone shifts into the bulk of the soil. Here the crystals, oriented with their base plane in the direction of freezing penetrate into the unfrozen soil before ice crystals are generated in it that would have a different orientation (which would require greater cooling) and have the advantage in their possibility of growth. Thus in the entire mass of soil a disordered lattice texture, layer after layer, is formed and the slower the freezing the greater will be the predominance of ice schlieren oriented in the direction of the temperature gradient. For this type of texture, apparently, the formation of segregated ice is characteristic mainly at the expense of local reserves of moisture in the soil with redistribution of moisture only over a very small range.

Because of the difference in resistance of the soil to displacement the selection of crystals extending beyond the limits of the pore takes place in various directions; further, only those can grow which are oriented with their base planes more or less parallel to the surface of the soil mass and consequently in most cases parallel to the isothermic surface. Because of this a new significance is acquired by a number of other factors. The fact that the generation of crystals occurs simultaneously at approximately the same isothermic surface, with preferred growth of them in this plane, tends to produce the fusion of crystals into polycrystalline ice schlieren parallel to the surface of the soil mass.

The greater the temperature gradient, the more uniform and fine-grained the soil, the more likely the simultaneous generation of crystals on one isothermic surface and the formation of continuous ice layers of considerable

length. Under conditions of small temperature gradients the role of isothermic surfaces in limiting space within which ice crystals can generate and develop decreases. It is self-evident that an increase in the temperature gradient can act favourably on the formation of continuous ice strata only inasmuch as the increase in the rate of freezing does not generally inhibit ice formation.

Non-uniformity, discontinuity and planes of rather weak bonding in plastic soil acquire great importance as a place for the generation and advantageous growth of ice crystals oriented with the base planes parallel to the weak planes. Ice schlieren arise primarily in the fissures along planes of hidden incipient schistosity of clay loams and clays, along boundaries of strata, around stones, in the holes made by worms, etc., whereas in porous ploughed soil irregular ice inclusions are formed. Particularly regular stratified texture is formed in filament clays. The effect of the soil structure before freezing is so great that if the direction of stratification differs greatly from the direction of the isotherms, two systems of ice schlieren occur, oriented in both directions with gradual transitions between them.

With cleavage of the plastic soil by growing crystals accompanied by dehydration of mineral particles and settlement of the soil, fissures are formed which are oriented primarily in the direction of the predominant growth of platelike crystals - parallel to the freezing front, but frequently deviating from this direction. In front of the freezing zone fissures may also be formed as a result of the extraction of moisture by the ice crystals. Such fissures are formed at right angle to the other fissures which are in the direction of freezing. The finer the soil, the more extensive the fissuring. In the fissures new crystals develop but if this does not take place the initial crystals which gave rise to the fissures can change the direction of growth and follow the fissures, resulting in lack of correspondence between the crystallographic and geometric orientation of the ice schlieren. In front of the freezing zone one frequently observes a discontinuous system of cracks penetrating at times up to several centimetres into the soil where the temperature is below  $0^{\circ}\text{C}$  and these cracks are subsequently filled with ice. These phenomena break up the parallel formation of stratified texture.

The generation of stratified texture parallel to the isotherms may be accompanied by a substantial flow of water to the freezing zone from lower-lying strata forming relatively large ice inclusions. The necessary condition for this is equilibrium between the quantity of heat flowing towards the source of cold that evolved during the process of crystallization of the water migrating to the zone of freezing. The more rapid the freezing, the more intense must be the flow of water. If equilibrium between them is disturbed because of the acceleration of freezing or decrease in the flux of water due

to the exhaustion of moisture in the adjacent strata, freezing extends to the pores in the soil below the ice layer where a new ice layer may begin to form, likewise growing in thickness due to the attraction of water if the necessary equilibrium is again established.

In the majority of cases, apparently, the process of freezing of plastic soil constitutes a series of stops of the crystallization front on the level where thermal equilibrium is established where the ice schlieren are formed, with spasmodic displacement of the front after exhaustion of the moisture from adjacent strata and disturbance of the equilibrium (Taber, 1930; Beskow, 1935; Tsyтовich and Sumgin, 1937). In general, the slower the process of freezing, the more favourable the conditions for maintaining thermal balance and the formation of large ice schlieren.

Conditions for the formation of texture and complex formations in frozen soils. The textures of frozen soils, as far as is known, are of primary character in the overwhelming majority of the cases, i.e. they are formed during the freezing of soil. The thawing of ice, migration and crystallization of water, resulting in the reconstruction of texture of freezing soils, are technically limited by the range of the temperature region of substantial phase transitions of water; at least at relatively low temperatures these processes do not give any noticeable results even during the course of prolonged geological time.

In nature one sees an infinite variety of textures in frozen soils corresponding to the variation in properties of the initial material and conditions of freezing. Nevertheless natural complexes of frozen soils can be combined into two types or groups. The first group includes strata of frozen soil which have been frozen epigenetically, i.e. after the stratum has been deposited, and the second includes strata which freeze syngenetically in the process of their deposition.

1. Epigenetically frozen strata of soil are generally characterized by comparatively low ice content and the predominance of massive and laminar cryogenic texture. The least favourable conditions for ice formation occur during the freezing of excessively dehydrated and compacted or even cemented sedimentary soils. When the soil is firmly cemented it differs very little from intrusive and metamorphic rock as regards ice formation. In such cases there is neither segregated ice nor ice cement but only vein ice in fissures whose formations are in no way connected with freezing.

When sedimentary soils retain sufficient moisture and plasticity for ice formation, epigenetic freezing apparently leads to the following characteristic distribution of cryogenic textures along with vertical cross-section resulting from a decreasing rate of freezing and an increasing pressure with depth owing

to the strata lying above. The upper 15 - 20 m of soil have a horizontally stratified texture with fine stratification at the top and thicker stratification towards the lower limit with occasional ice strata up to 10 - 20 cm thick. Less frequently one observes stratified lattice texture with well-developed horizontal ice schlieren which also become fewer and thicker towards the lower limit. Below 20 m ice inclusions are absent and the texture is of the massive type. Cross-sections of this type of texture are described by K.K. Markov, A.M. Pchelintsev and F.G. Bakulin.

2. Syngenetic frozen strata of soil differ from epigenetic strata by having considerably larger ice content, a predominance of lattice and stratified textures with a different arrangement of these two types in the vertical profile. With syngenetic freezing of subaerial and shallow-water sediments the frozen stratum undergoes a series of transitions layer after layer, with the lower part of the seasonally thawing layer being displaced upwards owing to the displacement of the surface as new deposits are laid down. The migration of water under these conditions is limited by the seasonally thawing layer which is frequently supersaturated owing to the infiltration of surface water down to the boundary of water-impervious frozen soils. During fall freezing in the lower part of the seasonally thawing layer, a closed system is formed resulting in lateral migration under pressure and the formation of lenses due to injected ice. Before it becomes part of the over-all frozen stratum the lower part of the seasonally thawing layer undergoes repeated freezing and thawing, and in the zone with low temperatures freezing also occurs from below due to the cooling effect of the frozen stratum.

Because of these particular conditions syngenetic frozen soils frequently have a high ice content (up to 60 - 80% and higher) and they have a lattice texture. The lattice texture can alternate with continuous ice strata or strata containing a large quantity of ice up to 3 - 19 cm thick, which are formed when there is a very high excess of moisture in the stratum above the permafrost and are frequently of injected origin. In regions where there is less moisture, laminar texture is formed. In the vertical cross-section and along the horizontal the type of texture may alternate repeatedly, corresponding to changing conditions of moisture and deposition of solid matter.

The structure of frozen soils. Information on the structure of frozen soils is exceedingly scant mainly because of the great technical difficulties in studying it (Shumskii, 1954). In the present state of knowledge one can advance only some general considerations concerning the structure of frozen soils based primarily on data concerning ice content and on studies of the structure of coarse-grained frozen soils.

The specific features of the structure of frozen soil are due to the presence of ice playing the role of cement. Cementing by ice, as with any other substance, does not alter the main primary structural features of soil or the size and shape of the component part of its solid skeleton. Nevertheless the structure of soil, after the formation of ice cement in a quantity which brings about a noticeable effect on the properties of the soil, undergoes substantial changes. Therefore the basic characteristics and consequently the name of the structure of frozen soil will not differ from the primary structure but it must be supplemented with the characteristic of the type of ice cement and its structure, and also the relationship between the structure of the cement and the skeleton of the soil.

All monolithic or rock formations regardless of their genesis (magmatic, metamorphic, chemical precipitate, etc.) undergo only a change in water content since they do not contain ice cement. The structure of such formations remains the same after freezing, and the ice filling the fissures (see Chapter IX) should be looked upon as independent mineral matter. The unique cryogenic structure belongs only to fine-grained sedimentary soils (which are either bonded or not bonded before freezing), fragmented and organic-mineral soils. Soils with massive texture are the only ones that have uniform structure, but when there is laminar or lattice texture the mineral or organic-mineral aggregates cemented with ice and ice schlieren have different structures. Here we will deal only with the first examples (a description of the structure of ice schlieren is given in Chapter IX).

As is known, fine-grained sedimentary soils have an agglomerate structure which corresponds to the grain size composition of the soil. Thus clays have pelite structure, unbonded soils - aleurite, sand - psammite, coarse fragmented soil - psephite structures, which are subdivided into brecciated and conglomerate structures. The same structural terminology applies also to the frozen state with the one difference that now one must also include the specific type and structure of the ice cement.

Depending on the extent to which the pores are filled, one can distinguish the types of cement:

- (1) contact - located only in the places of contact between the particles of the skeleton;
- (2) film - enveloping the surface of the particles without filling a substantial part of the pores;
- (3) pore - filling the pores completely;
- (4) basal - forms the basic mass of the soil in which individual particles of the skeleton are immersed.

As is evident from a consideration of the formation of frozen soils, the type of ice cement depends on the initial moisture content of the soil and on

the conditions of moisture migration during freezing. Contact and film cement are characteristics primarily of unsaturated sand and coarse fragmented soil, whereas pore cement is characteristic of such soils but when they are saturated with water. Fine-grained unbonded and particularly frozen clay soils, apparently in the majority of cases, have a basal ice cement consisting of ultra-microscopic crystals.

The structure of ice cement, i.e. its shape, size, orientation and distribution of crystals in the frozen soil, have received very little attention from scientists up to the present time. One can only suppose that they will turn out to be no less varied than the structure of single mineral ice formations and will reflect the conditions of the formation of the ice cement.

The structure of ice cement is of the crystalline-granular type. In this group, depending on the size of the crystalline grains, one can differentiate clearly crystalline structure (coarse, medium and fine) and hidden crystalline structure (microcrystals and ultra-microcrystals), as well as uniform grained and porphyritic (non-uniform grained) structures. Depending on the shape and crystallographic orientation of grains one can distinguish the following structures:

(a) Irregular-grained (allotrimorphogranular) - the shape of the grains is irregular owing to the effect of other grains surrounding the given grain; the crystallographic orientation is disordered;

(b) Prismatic (prismatic-granular or panidiomorphogranular) - the crystals are of more or less regular form having the given class of symmetry (prismatic); the orientation is ordered and linear (the main axes are parallel to each other having long geometrical axes of prismatic crystals);

(c) Structures that are intermediate between the above two (hypodiomorphogranular). In ice the intermediate structures can be of two types: plate (grains flattened along the main axis) and columnar (grains extended along the main axis).

Depending on the relationship between the cement grains and the particles of the soil skeleton the following structures are identified:

(a) interparticle (intersertal) - cement grains located in single spaces between skeleton particles and do not exceed them in size (polycrystalline cement);

(b) enveloping\* (poikilitic) - cement grains larger than the particles of the skeleton and including the latter within themselves (the so-called "single crystal" cement).

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\* The terms "irregular-grained", "prismatic", "interparticle" and "enveloping" structures were introduced instead of those used in petrography on the suggestion of the editor, N.A. Tsytovich.

Thus a full definition of the structure of frozen soil will be as follows: aleurite enveloping structure with irregular-grained basal ice cement (for unbonded clay loam), psammite interparticle structure with platelike pore ice cement (water-saturated sand), conglomerate interparticle structure with columnar film-ice cement (unsaturated coarse gravel), etc. Naturally these definitions must be supplemented with concrete descriptions with quantitative characteristics of shape and sizes of skeleton particles and grains of ice cement, the extent of development of optical anomalies in the ice, with diagrams of crystallographic orientation and space distribution of crystals of ice cement, photographs of structure, etc. (for more details see Shumskii, 1954).

Detailed investigations of texture and structure of frozen soil should in future make it possible to analyze the process and conditions of their formation and explain all aspects of their properties. Such studies are necessary in investigating the history of frozen soils as well as for practical engineering purposes.

#### Physical Phenomena and Processes in the Thawing of Soils\*

When the frozen soils begin to thaw part of the cement bond between individual mineral particles and aggregations of particles brought about by the formation of ice begins to weaken or is ruptured, which, in most cases, results in a decrease in the strength of the soil and deformation resulting from compaction and settlement of the soil. The moisture resulting from the thawing of ice becomes mobile thus creating the possibility of redistribution as well as hydration of mineral particles and aggregation of particles, dispersion and swelling. All of these phenomena and processes occur differently depending on the composition, structure and the properties of the soil but the direction of the processes resulting from thawing are in most cases opposite to those which are observed during freezing. Part of the phenomena and processes of freezing and thawing are reversible, for example, the freezing of water - thawing of ice; others are not fully reversible or are completely irreversible: coagulation - dispersion of mineral particles, compression - swelling, etc. In thawing coarse-grained soils many of these processes are relatively insignificant.

The thawing of ice contained in the frozen soils. With melting of ice the molecular crystalline lattice of pure ice is disrupted (Ousten and Lonsdale, 1948); the energy required is equivalent to 79.69 cal/g.

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\* Written by F.G. Bakulin, B.A. Savel'ev and V.F. Zhukov.

The melting of ice in frozen soils, in comparison with the melting of pure ice, is complicated by the adsorption properties of the surface of mineral particles in the skeleton, supplementary physico-chemical phenomena (resulting from the dissolving of part of the mineral skeleton in the water) and chemical effects (the formation of crystal hydrates and interphase chemical reactions). Changes in temperature, concentration of solutions and pressure or volume of the system disturbs the equilibrium state between the ice and liquid phase.

When the temperature increases and the frozen soil begins to thaw at a specific range of below-freezing temperatures partial thawing occurs and in a narrow temperature range near  $0^{\circ}\text{C}$  there is complete transition from ice to water (Tsytovich, 1945, 1947; Nersesova, 1950, 1951; Jung, 1932). In this latter temperature range the greatest physico-mechanical changes in the characteristics of the soil occur.

The melting of ice in soil varies depending on the texture of the frozen soil. Laminar ice melts at  $0^{\circ}\text{C}$  whereas ice cement in pores thaws gradually within a specific temperature range.

In addition to ice and liquid water the pores of frozen soils always contain a certain quantity of water vapour which likewise plays a part in the melting of ice and migration of moisture. Since the melting of ice results in a decrease in volume of the system it is accompanied by the formation of pores and fissures resulting in a more rapid redistribution of moisture in thawing soils.

A partial melting of ice in frozen soil occurs when there is a continuous increase in soil temperature, however the melting of the basic massive ice is characterized by a relatively slow increase in temperature stopping at the level approximately corresponding to the transition between ice and free water.

The average temperature of frozen soil corresponding to the state of the basic transition of ice into water differs from the freezing point - it is as a rule somewhat lower. The continuity of a phase transition depends on the density of heat flow and also on the composition of the frozen soil - on the relation between the mineral parts of the skeleton, ice and liquid water. The more ice there is in frozen soil, the closer will be the temperature of the phase transition to the melting-point and conversely the less ice, the lower will be the temperature of phase transitions.

The redistribution of moisture in fine-grained thawing soils. The redistribution of moisture occurs in the direction of heat flow - from warmer regions of the soil to cooler regions. Gravitational forces have an influence on the movement of water, the direction of which may not correspond with that of the heat flow.



In the thawing of coarse-grained soils (gravel, broken stone, sand) the change in the quantity of moisture, as occurs during freezing, is in most cases negligible. In sand all the moisture resulting from the melting of ice cement as a rule is retained in the form of film or capillary water. In some cases when some of the water has been squeezed out during the freezing process there may be an increase in the water content upon thawing due to the flow of water from neighbouring unfrozen strata. If the amount of water in the soil after thawing is greater than that retained in the form of film or capillary water the excess disappears under the influence of hydrostatic pressure and gravitational forces.

All of the above-mentioned types of redistribution of moisture in coarse-grained thawing soils occurs rather quickly by virtue of the high filtration ability of the soils. The rate of redistribution of moisture is determined basically by the rate of thawing.

In fine-grained soils, particularly in unbound soils and also soils having laminar and lattice structure with substantial quantities of ice inclusions, the process of redistribution of moisture on thawing is much more complex. In these soils the quantity of ice on thawing is in most cases considerably greater than that contained before freezing. Therefore on thawing there is a specific excess of water found mainly in macro-pores which is subject to redistribution in the thawing soil mass.

As the ice melts, part of the liberated moisture (in some cases up to 15% of the weight of the soil skeleton) first of all goes to hydrate the mineral particles and aggregation of particles which have become dehydrated during freezing. Hydration is greatly influenced by the mineral composition and the composition of exchange cations in the soil. Kaolin clay, for example, binds less moisture than clay containing a great deal of montmorillonite. However, any clay will bind more water when it is saturated with exchange complex of  $\text{Na}^+$  than, for example, when it is saturated with  $\text{Ca}^{++}$ ,  $\text{Mg}^{++}$  and  $\text{H}^+$  cations.

Part of the water remains in the thawed soil in the form of capillary water and part is retained in the intra-aggregate macro-pores and remains in them in the form of the so-called immobilized water.

The remaining water formed from the ice usually cannot be retained in the thawed stratum and is displaced under the influence of gravitational forces. Thawed fine-grained soils having laminar and lattice structure frequently retain the cavities and fissures that were formed during freezing for a long period of time, and as a result have a higher water permeability than they had before freezing.

Under natural conditions in regions where permafrost occurs, the thawing stratum is usually underlain with a frozen stratum which interferes with the

free filtration of water downwards and therefore the water is retained in the thawed stratum. The redistribution of moisture due to the difference in temperature between the thawed layer and the underlying frozen stratum does take place but only to a very small extent. This frequently results in super-saturation of the thawed layer, particularly when it is in contact with permafrost over a long period of time. This leads to a great decrease in its bearing capacity, to a decrease in strength and to such phenomena as soil creep and slides (on slopes), etc.

When frozen clay soils of monolithic composition containing relatively little ice thaw out there is no soil creep. The moisture formed in them as the result of the thawing of ice cement is almost completely used up in the hydration of mineral particles and their micro-aggregation and also in the filling of capillaries.

The deformation due to compaction of frozen fine-grained soils on thawing occurs because of the disappearance of ice inclusions and also because the mineral particles which were moved apart during freezing now come together (Tsytoich, 1940, 1941<sub>2</sub>; Lapkin, 1937; and others).

It has been established that on the freezing of silty and clay soils there is frequently a differentiation of moisture and mineral particles and simultaneously with a general increase in the volume of the frozen soil there is compaction of its mineral components.

This type of formation of compact mineral units during the freezing of soil was mentioned during the last century by I.A. Lopatin (1873) and later by other scientists (Holmquist, 1898; Abolin, 1913; Given, 1915; Fedosov, 1935, 1940; Beskow, 1935).

The moisture content of compacted mineral concentrations is usually 2 - 4% higher than the moisture content of soil with a lower plastic limit. Such moisture is usually observed in unfrozen soils that have been compacted by a load of 6 - 8 kg/cm<sup>2</sup>.

As a result the freezing of fine-grained soils is accompanied by compacting of separate mineral concentrations as well as general loosening of the soil because of the separation of mineral concentrations by ice inclusions.

When this type of soil thaws its former homogeneous structure is not reconstituted. The thawed soil retains compact mineral concentrations which do not become more compact but rather frequently swell and increase in volume because of hydration of the mineral particles and because of the splitting action of fine films of water as was first established by V.V. Deryagin (1932) and subsequently reported by Bakulin and Zhukov (1955).

The decrease in volume of thawing soil takes place because of the melting of ice inclusions and the squeezing out of the water with simultaneous increase

in volume due to swelling of mineral concentrations. If the frozen soil contains relatively little water one cannot exclude the possibility of a general increase in volume on thawing.

The deformation due to compaction that is observed during the thawing of soil is the total effect of complex deformation of the soil.

Ice layers in frozen soil, however large they may be, are discontinuous and of different direction, shape and size. Frozen soil is in general a mineral body with micro- and macro-inclusions of ice, liquid water and air. On thawing, even under conditions of free movement of water, such an anisotropic body possesses the ability to resist disturbance of its structure. For example, in a specimen of clay soil of disturbed structure an ice plate was inserted before freezing, after thawing under a load of  $1 \text{ kg/cm}^2$  a free cavity was retained in the place of the ice plate. This resistance varies, depending on the strength of the bond between mineral particles and their aggregations. The strength of the bonds between mineral particles in thawing soil is determined by the mineralogical and chemical composition, the quantity of water soluble substances, genesis, conditions of formation during the freezing process, changes undergone during the frozen state, etc. The formation and strength of bonds within the mineral skeleton depends on the rate of freezing, the formation of frozen soils (epigenetic or syngenetic formation) which gives the particular cryogenic structure to the ground. This has been indicated by A.M. Pchelintsev (1948-1954), F.G. Bakulin and V.F. Zhukov (1955).

The disturbance of structure of thawing plastic soil under the influence of its intrinsic weight or load is accompanied first of all by deformations of bridges formed between aggregations of mineral particles. As these bridges are deformed their area increases and the size of the macro-pores decrease and consequently for further disturbance of structure one must apply a greater load. Complete compaction of thawing soil occurs only when substantial loads are applied approaching the loads of compaction of the mineral concentrations.

An analysis of the data on the influence of structure on the compressibility of frozen, highly saturated soils that have become differentiated into ice inclusions and mineral concentrations, and have subsequently undergone thawing, permits one to make a schematic representation (Fig. 13) of the change in the coefficient of porosity of the soil mass under load.

The initial coefficient of porosity of frozen soil  $\epsilon^0$  decreases on thawing under ever-increasing pressure and reaches a limit - the coefficient of porosity of mineral interstrata  $\epsilon_M$  - only when the load is very large. But in general the porosity coefficient remains somewhat greater than  $\epsilon_M$ .

The compression curves expressing the dependence of the porosity coefficient on pressure can be divided into three portions.

In the first portion (QN) the change in porosity coefficient takes place mainly because of decrease in the volume of the soil resulting from the melting of ice inclusions and the squeezing out of water followed by decrease in the macro-pores. During this process the initial structure is disturbed.

In the second portion (NM) the change in the porosity coefficient is due primarily to further disturbance of the cryogenic structure of the frozen soil during which the macro-pores disappear completely and there is plastic compression of the aggregates, particularly in places of contact between particles.

In the third portion (ML) compaction is due entirely to the reduction of distance between mineral particles. For this type of compaction one must apply considerable force.

The process of compacting thawing soil is in fact much more complex than presented above. Compaction develops simultaneously through destruction of the cryogenic structure and plastic compression; however, during the first part compaction due to destruction of structure predominates whereas in the second part plastic compression predominates.

For practical purposes it is admissible to simplify the relationship described without forgetting, however, the physical essence of the compaction process experienced by thawing fine-grained soils: the point N because of the small value of the load (Fig. 13) can be transferred to the ordinate axis and one can assume that the change in the porosity coefficient from  $\epsilon_0$  to  $\epsilon_N$  occurs spontaneously, without the action of external loads; the change in the porosity coefficient from  $\epsilon_N$  to  $\epsilon_M$  is proportional to the load and  $\epsilon_M$  (the porosity coefficient of mineral interstrata) corresponds to compacting  $p_M$ .

The value of the load corresponding to compaction up to  $\epsilon_M$ , i.e. to the density of the mineral interstratum, can be determined by using an odometer in testing an individual specimen with any quantity of ice inclusions in it.

The porosity coefficient ( $\epsilon_0$ ) of the initial state of the frozen soil should be determined for large soil masses, for example, in boreholes or test pits for a column or a monolithic mass of soil. The coefficient  $\epsilon_N$ , which depends on the structural stability of a frozen soil mass undergoing thawing, should be determined under natural conditions.

Thus the physical essence of the deformation of frozen soils on thawing can be expressed by the following principles:

- (1) The structure of frozen soils on thawing is to a certain extent stable and its destruction is not spontaneous but occurs under the influence of external load;

- (2) In the thawing of highly ice-saturated soils the compaction is due to decrease in the volume of macro-pores occupied by the ice inclusions. There

is no compacting of mineral aggregates when the load is not large; their volume is retained on thawing and may even increase because of swelling.

The change in the physico-mechanical properties of soils due to freezing and subsequent thawing.\* As a result of freezing and subsequent thawing soils change their properties substantially. Physico-mechanical properties such as strength, compressibility, filtration, etc., are particularly subject to change.

The basic reason for the change in the physico-mechanical properties of soils during the course of freezing and subsequent thawing\*\* is the change taking place in the structure which is the result of two processes occurring in sequence: first, the formation of a particular structure during freezing and second, the destruction of this structure on thawing. The changes in structure of the soil, depending on the grain-size composition, moisture content, texture and also the conditions of freezing and thawing (rate of cooling, external pressure etc.), may vary.

The greatest changes in the structure of soil under the influence of freezing-thawing is observed when there is ice separation with the formation of laminar and lattice texture\*\*\*. In case of freezing without ice segregation but with the formation of massive texture, the change in the structure of the soil is expressed only in changes in the distance between particles.

Changes in the physico-mechanical properties of the soil during the process of thawing have received little attention by scientists. Some data on settlement resulting from thawing, the rate of compaction and the strength of soil have been elucidated by some authors (Gol'dshtein, 1948; Blinov, 1948; Shusherina, 1955).

The greatest increase in settlement under otherwise equal conditions is observed when there is an excess of ice segregation (the increase may be in tens or even hundreds of percent). When ground freezes without the flow of moisture, the increase in settlement on thawing, as compared with the state of the soil before freezing, is much less (up to 3 - 4%).

The compaction rate of the soil on thawing is much greater than the compaction rate before freezing, which is explained basically by the increase in the filtration coefficient of thawed soil.

The change in strength of the soil on thawing at the boundary between thawed and frozen soil is determined by the nature of the thawing surface.

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\* Written by E.T. Shusherina. For more details see No. 3 "Materialy po Labor. Issled.....", 1957, page 99 and following.

\*\* For brevity we will henceforth write freezing-thawing.

\*\*\* The structure of frozen soil is contained in the third section of this chapter.

When the thawing surface is uneven with abrupt ice outcrops the strength at the interface between the thawed and the frozen soil increases; when the thawing surface is even and smooth the strength is either less than before freezing (in the presence of ice inclusions), or does not change (if the frozen soil is of massive texture).

For example, according to the data of E.T. Shusherina (1955), the shear strength of clay loam (rapid shear) before freezing was  $0.12 \text{ kg/cm}^2$ ; on the thawing boundary, however, when the surface was uneven, under otherwise equal conditions (moisture content, normal pressure), it increased to  $0.52 \text{ kg/cm}^2$ . The data of M.K. Zakharov (1951) showed that the shear strength at a smooth thawing boundary for soils frozen with the formation of ice inclusions decreased: for clay the decrease was from  $0.64$  to  $0.28 \text{ kg/cm}^2$ , for clay loam from  $0.55$  to  $0.22 \text{ kg/cm}^2$ ; for sand, however, in which there were no ice inclusions there was no change in shear strength.

An uneven thawing surface occurs with lattice texture of frozen soil and also with laminar texture but only when the thawing surface transverses the ice layers.

A smooth interface between thawing and frozen strata is characteristic of frozen soils having massive and laminar texture when the thawing surface is parallel to the ice layers.

Changes in the physico-mechanical properties of frozen soils after thawing has been investigated more extensively than during the process of thawing (Fedosov, 1940; Migin, 1947; Gol'dshtein, 1948; Pchelintsev, 1948<sub>2</sub>; Evdokimov and Zauerbrei, 1950; Pustovalov and Chukhrova, 1952; Tsytoovich and Shusherina, 1955).

Changes in the moisture content of soil during the freezing-thawing process is determined by changes undergone during thawing and also the ability of the soil to retain the moisture after thawing.

If during freezing the moisture content increases, after thawing it will be greater than before freezing; if the water resulting from thawing cannot be squeezed out and when there is excessive ice segregation it is possible that the thawed ground may be supersaturated. When the water can be freely drained or filtered away from the thawed ground then only the water that can be retained by the soil will remain.

If on freezing the moisture content decreases, then on thawing if there is no free flow of water to the thawed layer, its reduced moisture content will be retained; when there is free flow of water to the thawed layer the soil will swell and the moisture content of the soil may become higher than it was before freezing.

The unit weight of the soil after the freezing and thawing process changes primarily because of changes in moisture content. When the pores of the soil are completely filled with water before freezing, the increase in moisture content as a result of the freezing-thawing process results in a decrease in unit weight and consequently in an increase in the porosity of the soil.

The changes in compressibility of the soil is connected with the change in porosity of the soil after freezing-thawing. When there is an increase in the porosity coefficient the compressibility increases and when there is a decrease in porosity there is an inverse relationship.

In this case the freezing of the soil with formation of lattice texture results in an increase in compressibility due to the freezing-thawing process.

This follows not only from the increase in porosity but also from the formation of aggregations and macro-pores which increases the compressibility of the soil.

When the freezing-thawing process does not result in changes in porosity and a massive texture is formed during freezing, the compressibility coefficient of the soil remains practically unchanged.

The changes in the rate of compressibility of the soil resulting from freezing-thawing depends to a large extent on the texture formed during freezing.

When the soil freezes and a massive texture forms, the compaction rate of the thawing soil remains practically the same as it was before freezing. When, however, a lattice texture forms, the compaction rate after thawing increases substantially owing to the formation of aggregations and macro-pores, causing an increase in filtration and decrease in strength of the bond between the structural parts of the soil.

Filtration of the soil resulting from the freezing-thawing process as a rule increases because of the large pores that are retained in the soil after thawing.

In conclusion we will deal briefly with one of the most important properties of soil - its strength. Available data indicate that the strength of soil after freezing-thawing varies differentially, it may increase or decrease and the changes in strength are not uniform.

The changes in the strength of the soil after freezing-thawing depend on the initial composition and structure, and also on the texture after freezing and porosity after thawing. These factors are specifically interrelated and, as a rule, occur simultaneously; the changes in the strength of the soil are determined by their total effect.

Under otherwise equal conditions, a decrease in the strength of the soil resulting from freezing-thawing will be greater, the greater the density before

freezing, the greater the amount of ice segregation during freezing and the greater the porosity after thawing.

The total effect of the texture after freezing and porosity after thawing becomes evident in the following way.

(a) If laminar and lattice textures are formed during freezing and porosity increases after thawing, the strength of the thawed soil will always be less than it was before freezing. The greatest decrease in strength of the soil is observed when there is influx of water during freezing, since this results in the greatest ice segregation during freezing and greatest increase in porosity after thawing.

For example, according to the data of the Laboratory of the Permafrost Institute the shear strength of soil after freezing-thawing, when there is excess ice segregation, was only 20% of the shear strength before freezing, i.e. the decrease was by a factor of 5; in the absence of water inflow the reduction in shear strength was only 25%.

(b) If laminar and lattice texture are formed during freezing, on thawing the soil compacts but the decrease in strength is considerably less than when there was an increase in porosity. When there is a specific amount of compaction of the soil there is no decrease in strength after freezing-thawing and in the case of extensive compaction there is an increase in strength.

The decrease in strength of the soil due to freezing and subsequent thawing can be reduced by compacting the soil after thawing.

(c) If massive texture is formed on freezing the changes in strength following the freezing-thawing process may likewise be heterogeneous but to a lesser extent since changes in porosity are not as large. As shown by experiments, there is frequently an increase in porosity in this case.

Thus the most favourable conditions from a point of view of retaining soil strength after freezing-thawing is the formation of massive texture during freezing.

#### Methods of Improving the Properties of Freezing and Thawing Soils

Knowledge of the process occurring in soils during freezing and thawing make it possible to approach the question of controlling the changes in their properties.

The first thing is to consider the possibility of controlling the heat exchange in freezing and frozen soils. The more rapid the freezing process the less water has time to flow to the freezing front and there is less heaving of the soil. Further decrease in temperature of frozen soils increases their strength substantially, whereas an increase in temperature decreases the strength.



The application of external pressure can also alter the rate and magnitude of soil heaving on freezing, which also has an effect on the formation of texture.

The control of physico-chemical and electrochemical action in freezing and thawing of soils may also introduce a substantial contribution to the problem under discussion.

It is likewise important to develop methods of improving the properties of thawing soils, particularly taking into account the fact that they frequently form a liquefied mass. Here one may successfully use the methods developed for weak unfrozen soils such as: electric drying, chemical strengthening of the soil, the use of soil piles, compaction with heavy machines etc. However, the development of the above-mentioned methods for strengthening thawing soils successfully can only be done after detailed investigations of the properties of the soil and of the physical processes occurring in them.

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Table I

The effect of exchange cations on the quantity of liquid water in frozen soils

(from the data of Z.A. Nersesova)

Soil	Quantity of liquid water as a percentage of the weight of dry soil						
	-0,2°	-0,5-0,6°	-1,2-1,4°	-2,0-2,5°	-5,0-5,5°	-10-10,5°	-30,0-32,0°
Na-kaolin . . . . .	37,5	15,3	6,5	2,8	—	0,9	—
Ca-kaolin . . . . .	32,5	12,6	6,4	2,8	—	0,9	—
Na-askanite gel . .	114,0	60,4	51,5	45,6	40,1	37,4	22,0
Ca-askanite gel . .	51,4	45,8	43,2	39,5	35,2	32,5	18,0
Na-clay loam . . . .	21,5	15,3	13,0	12,6	—	9,9	—
Ca-clay loam . . . .	17,1	13,6	—	12,0	11,0	9,2	—

Table II

The quantity of liquid water in some typical soils depending on temperature

(from the data of Z.A. Nersesova)

Soil	Quantity of liquid water as a percentage of the weight of dry soil					
	-0,2-0,3°	-0,5-0,7°	-1,0-1,5°	-2,0-2,5°	-4,0-4,5°	-10,0-11,0°
Sand . . . . .	0,2	0,2	—	2,0	0,0	0,0
Sand loam . . . . .	—	5,0	4,5	—	4,0	3,5
Clay loam . . . . .	12,0	10,0	7,8	—	7,0	6,5
Clay . . . . .	17,5	15,0	13,0	12,5	—	9,3
Clay, containing montmorillonite . .	34,3	25,9	—	19,8	—	15,3

Table III

Content of liquid water as related to temperature for  
typical non-saline soils  
(according to the data of Z.A. Nersesova)

Soil*	Max. mol. moisture capacity	Quantity of liquid water as a percentage of the weight of dry soil				
		-0.2-0.5°	-1.0-2.0°	-4.5-5°	-9-10°	Below -10°
Sand.....	1-7	0.5-2	0.5	0.5	0.5	0.5
Sand loam .....	9-13	3-10	3-6	3-6	3-6	3-6
Clay loam .....	15-23	10-20	5-15	5-10	4-8	4-8
Clay.....	23-35	15-25	10-20	10-15	5-10	5-10
Clay, containing montmorillonite..	>30	30-40	20-30	15-25	15-20	15-20

\* The classification of soils is based on the maximum  
molecular moisture capacity

Table IV

The effect of external pressure on the quantity of liquid  
water in frozen soils

Soil	Temper- ature	Initial data		Quantity of liquid water in % of total moisture		Remarks
		Pressure kg/cm <sup>2</sup>	Moisture %	Freezing without pressure	Freezing under pressure	
Moscow region clay loam	Below freezing	2	22	50.2	61.5	Tempera- ture not measured precisely
" "	"	10	22	72.7	74.2	"
Bentonite clay	"	2	49	59.3	66.1	"
Moscow region clay loam	-1.7°	2	20	42.4	58.1	-
Bentonite clay (askanite gel)	-5.8°	5	46	42.5	45.6	-

Table V

Quantity of Ca ions displaced by Na ions owing to interaction of soil with NaCl solution

(from the data of I.A. Tyutyunov)

Temperature °C		Concentration of solution in %	Quantity of displaced Ca-ions in M. equiv. per 100 g soil	Remarks
Soil	Solution			
20	20	0.06	0.5	Moisture content of the soil was approximately 19%
20	20	0.06	0.5	
-0.3	0	0.06	0.6	
-0.3	0	0.06	0.6	
20	20	0.3	1.3	Interaction of the soil with the solution lasted for 5 minutes
-0.3	-0.2	0.3	1.4	
-0.3	-0.2	0.3	1.3	
20	20	2.0	1.8	
-0.5	-1.1	2.0	3.2	
20	20	8.0	8.1	
-4.5	-4.9	8.0	8.0	

Table VI

Quantity of Ca ions displaced by Na ions owing to interaction between soil and NaCl solution

(from the data of I.A. Tyutyunov)

Series no. of experiment	Temperature °C			Quantity of Ca, M. equiv. per 100 g dry soil	Concentration of solution in %
	Soil	Solution	Filtrate		
I	+20	+20	+20	2.4	2
	+20	+20	+20	2.3	2
	-1.5	-10	-1.5	2.5	2
	-1.5	-10	-1.5	2.5	2
II	+20	+20	+20	3.0	8
	-4.1	-4.1	-5.0	3.2	8
	-4.1	-4.1	-5.0	3.2	8
III	+20	+20	+20	3.3	14
	+20	+20	+20	3.0	14
	-7.9	-7.8	-6.5	3.7	14
	-7.9	-7.8	-6.5	3.6	14

Table VII

Absorption capacity, absorption of hydrogen and base saturation of peat-gley soils

(from the data of I.A. Tyutyunov)

Depth at which specimen taken in cm	Absorption capacity according to Bobko-Askinazi		Hydrogen absorption		Absorbed bases, M. equiv.	% of base saturation	% H ion saturation
	% Ca	M.equiv.	%	M.equiv.			
14 May 1945							
20-37	0.33	16.50	0.0029	2.90	13.60	82.42	17.58
37-47	0.30	15.00	0.0056	5.60	9.40	62.67	37.33
47-57	0.42	21.00	0.0099	9.90	11.10	52.86	47.14
57-67	0.39	19.50	0.0071	7.10	12.40	63.59	36.41
3 September 1945							
22-37	0.35	17.50	0.0042	4.20	13.30	74.86	25.14
37-57	0.38	19.00	0.0038	3.30	15.70	82.63	17.37
57-67	0.42	21.00	0.0034	3.40	17.60	83.81	16.19
67-77	0.46	23.00	0.0026	2.60	20.40	88.70	11.30

Table VIII

The magnitude of relative soil heaving depending on  
conditions of freezing  
(from the data of N.A. Tsytovich and I.S. Vologdina)

Soil	Total moisture content %	Magnitude of rela- tive heaving in %	
		freezing only from above	freezing from all sides
Sand (1-0.25 mm - 60.2%; 0.25-0.05 mm - 39.8%) .....	18.7	0.42	1.6
Structureless silty- clay loam (1-0.05 mm - 3.7%; 0.05-0.005 mm - 81.8%; less than 0.005 mm - 14.5%) .....	21.9	1.27	Untested
Clay (1-0.01 mm - 3.7%; 0.01-0.005 mm - 34.3%; less than 0.05 mm - 62.0%) .....	32.4	2.38	5.2

Table IX

Changes in total moisture content and heaving of surface  
clay loam during freezing  
(from the data of Z.A. Nersesova)

Depth in mm	Moisture (percentage of the weight of dry soil)			
	Fe-clay loam	Ca-clay loam	Na-clay loam	K-clay loam
0-3	157.8	98.1	83.0	71.9
3-25	90.0	45.6	42.4	25.8
25-60	90.5	40.5	40.2	25.9
60-85	28.3	27.5	37.2	21.6
85-115	24.5	--	--	--
Heaving, mm*...	30.0	7.0	--	--
" %...	35.3	8.2	--	--
Moisture before freezing, %...	29.0	29.0	38.2	26.9

\* Height of specimen before freezing was 85 mm.  
The specimens were subjected to a pressure  
of 0.5 kg/cm<sup>2</sup>



Table X

Changes in the shear strength of soil as a result of freezing  
and subsequent thawing  
(E.P. Shusharina, 1955)

Soil	Normal pressure, kg/cm <sup>2</sup>	Shear strength before freezing and ( $\tau_1$ ), kg/cm <sup>2</sup>	Maximum changes in shear strength after freezing- thawing pro- cess ( $\tau_2$ ), kg/cm <sup>2</sup>	$\frac{\tau_2}{\tau_1} \cdot 100$
Moscow region surface clay loam (soil paste)	0.1	0.13	0.06-0.22	46-170
	0.1	0.41	0.09-0.50	22-122
	0.1	0.50	0.17-0.32	34-64
Oka delta clay (soil paste)	0.1	0.06	0.04-0.14	67-234
	0.1	0.29	0.07-0.46	24-159
	0.1	0.54	0.11-0.22	20-41
Oka delta clay (undis- turbed structure)	0.1	0.50	0.10-0.64	20-128
Vorkuta surface clay loam (soil paste)	0.1	0.10	0.05-0.18	50-180
	0.1	0.20	0.08-0.47	40-235
	0.1	0.28	0.08-0.32	29-114
Kinel' clay (undisturbed structure)	0.1	0.9	0.2	22
	0.6	1.2	0.7	58
	1.1	1.4	0.7-1.7	80-122
	2.1	1.9	0.7-2.4	37-127
Kuibyshev alluvial clay (undisturbed structure)	0.1	0.7	0.4	28
	0.6	1.0	0.35	35
	1.1	1.4	0.6	36
	1.6	1.4	0.6	43

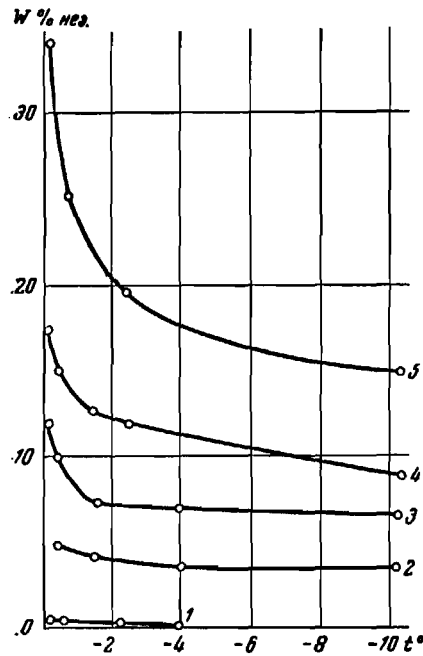


Fig. 5

Quantity of liquid water in typical soils depending on temperature

- 1 - quartz sand; 2 - sandy loam; 3 - clay loam;  
4 - clay; 5 - clay containing montmorillonite

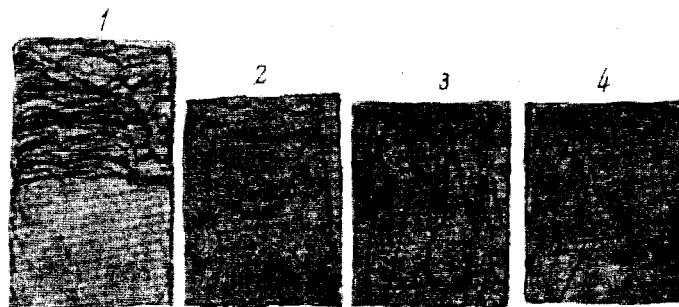


Fig. 6

Surface clay loam saturated with cations  $\text{Fe}^{+++}$ ,  $\text{Ca}^{++}$ ,  $\text{Na}^+$  and  $\text{K}^+$   
according to experiments of Z.A. Nersesova  
Photo by Yu.A. Shchiplitsev

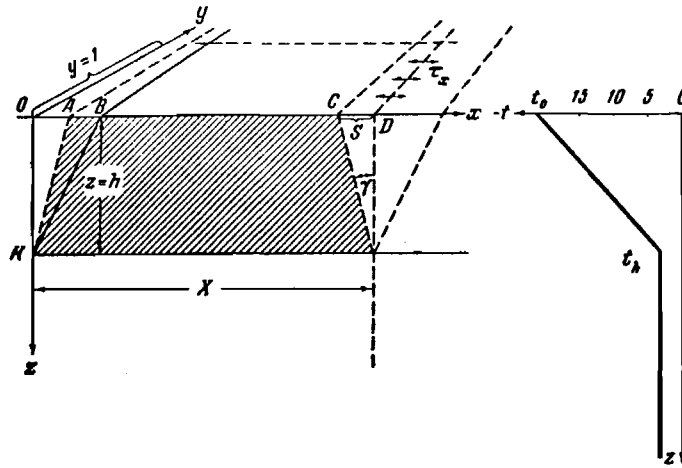


Fig. 7

Diagram for calculating rupturing stresses

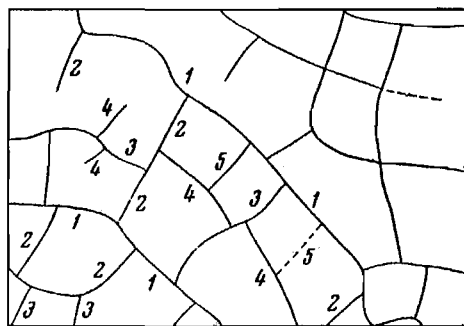


Fig. 8

### Sequence of fissure formation of different orders of genesis

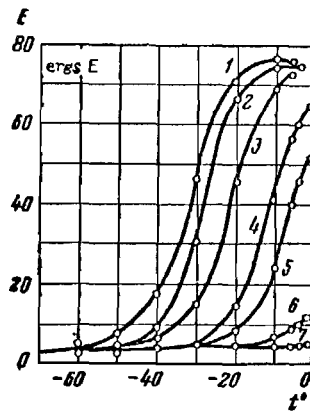


Fig. 9

Changes in dielectric constant  $\epsilon$  of ice depending on temperature  $t$  and frequency  $f$

- 1 -  $f = 300 H_L$ ; 2 -  $f = 500 H_L$ ; 3 -  $f = 1000 H_L$ ;  
 4 -  $f = 3000 H_L$ ; 5 -  $f = 5000 H_L$ ;  
 6 -  $f = 20000 H_L$ ; 7 -  $f = 60000 H_L$ ;

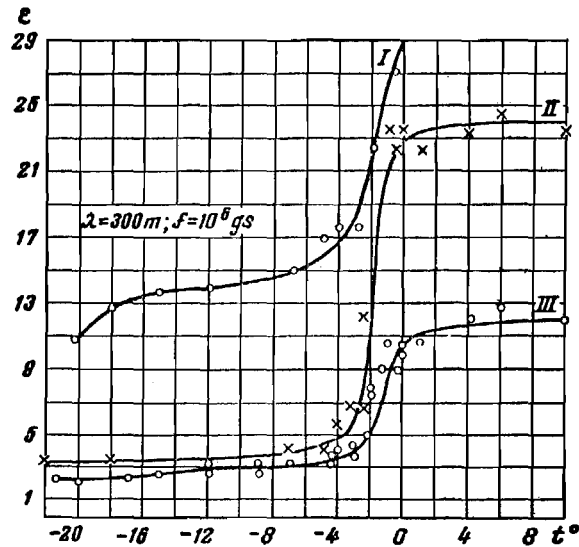


Fig. 10

Variations in the dielectric constant  $\epsilon$  of soils of different moisture contents depending on temperature  $t$

- I - Jurassic clay at moisture content of  $w = 35.5\%$  of the dry weight of the soil; II - fine sand  $w = 9\%$ ;  
 III - the same soil with  $w = 3\%$



Fig. 11

Laminar texture of structureless clay loam  
Photo by E.M. Katasonov

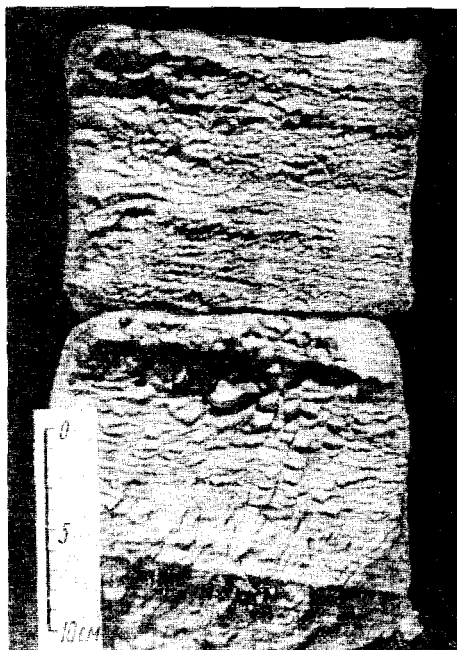


Fig. 12

Laminar-lattice texture of structureless clay loam  
Photo by E.M. Katasonov

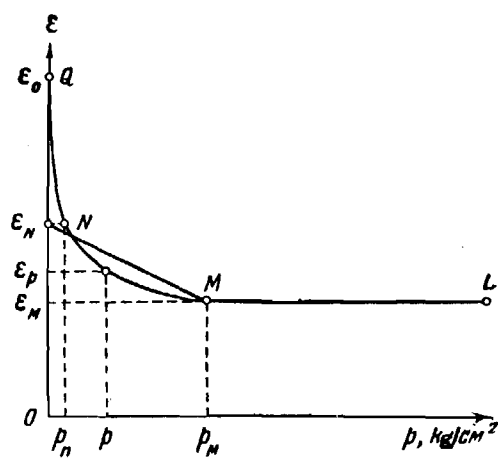


Fig. 13

Variation in the porosity coefficient  $\varepsilon$  of the soil during the thawing process under load