Actual soil is a heteroporous body, characterized by definite size distribution of pores. The liquid that fills the soil pores possesses anomalous properties near particle surfaces because of its viscoplastic state and osmotic phenomena during flow [1, 2]. Let us consider the effect of heteroporosity of a soil mass on its consolidation under load. First let us examine results of experiments on consolidation of Moscow (Podmoskovskii) loams and fuller's earth completely saturated with water [3]. Load was applied in equal stages, and each succeeding application was made after settlement from the preceding stage had stabilized. The dissipation of pore pressure observed in the experiment showed that the process of squeezing the liquid out takes place according to the theory of consolidation through filtration only for soils that have not been first compacted. Preliminary compaction leads to formation of structural bonds which affect the mechanism of transmitting external pressures to the pore liquid. The pore-pressure coefficient $\beta$ is an index that characterizes the effect of structural bonds on the consolidation process and on the pressure distribution in the pore liquid and in the soil skeleton. As follows from the discussion below, $\beta$ may be defined as the ratio of pore pressure to external pressure in compression in a closed system or at the instant load is applied in an open system.

During thawing in very icy uncompacted soil under load as well as in thawed ground, the entire load is transmitted to the water. In soil that has been first compacted, only part of the external load is transmitted to the water at the time of thawing. The denser the ground the lower the initial pore pressure. Change of pore pressure with time in dense thawing soils has the same character as in thawed soils: at first the pore pressure grows to some maximum, and then it is dissipated during consolidation through filtration. Experiments have shown that the coefficient $\beta$ does not depend on external load and is constant for height of specimen (measurements of pore pressure were made at three levels).

From Fig. 1, in which we show the change in the coefficient $\beta$ according to preliminary compaction of thawed and thawing ground, it is seen that in surficial loam, decrease in $\beta$ takes place more strongly than in fuller's earth because of the difference in mineralogy and grain size. For soil not completely saturated with water, the coefficient $\beta$ increases with increase in density (degree of preliminary compaction) (Fig. 2).

Experimental investigations of consolidation of clayey soils show that application of the Terzaghi-Gersevanov theory depends primarily on consistency of the soil. For clay soils with a consistency $B < 0.5$, conclusions drawn from the theory of consolidation through filtration find no support. For these soils, as Maslov has noted [4], the load is supported by the skeleton structure and is gradually transmitted to the pore liquid according to the degree of soil deformation. Experimental data support this view.

It should be noted that, during successive loading by stages, soil gradually changes from a "soil mass" characterized by the absence of structural bonds to a quasi-single-phase system. Current theories do not provide for transition of soil to a state in which pore pressure in the soil does not develop or, if it has developed, has no effect on the course of compaction. The structure of the equations of state is such that at any density the mechanism of pressure transmission to the water and the skeleton is the same and, whatever the degree of compaction, the pore liquid always supports the entire applied external load.
In [5] it was shown that on the basis of tensor properties of the stress fields, deformation, and pore pressure, it is possible to obtain general forms of the relations in the soil system satisfying the condition set above. However, the physical meaning of these relations has not been discovered. Consideration of the heteroporosity of the soil medium permits us to find some basis for the physical meaning of the relations.

During consolidation of previously compacted clayey soils, at the initial moment the liquid remains immobile between mineral grains in thin layers. Here the active porosity \( n_a \), corresponding to the relative volume of mobile fluid, is less than total porosity \( n \). Only at a certain load \( q \) may the liquid be pressed from thin layers between clay particles into pores of larger size, as a consequence of which the pore liquid migrates through the soil medium. The external pressure necessary for bringing the mineral particles together, particles that carry a surface charge and are separated by a thin layer of electrolyte solution, is determined from the following expression:

\[
q = k \frac{H}{L} \left( \tau_s + \tau_m + \tau_i \right).
\]

The first term on the right side of Eq. (1) corresponds to plastic resistance of a layer of liquid with thickness \( L \) having a shear strength \( \tau_s \), included in the space between mineral particles of characteristic dimension \( H \); \( \tau_m \) and \( \tau_i \) are determined by the presence of molecular attraction between particles and by the wedge effect of ionic-electrostatic origin, developed by the liquid throughout the interstitial space. Molecular attraction, computed for unit area of the interstitial space, is inversely proportional to the cube of the distance between particles. An expression similar to Eq. (1) may be found also for the bringing together of particles of different form, such as spherical [2].

Equation (1), solved relative to \( L \), defines the minimal size of pore \( L^* \) for which liquid may move at a given pressure \( q \). The migration of pore liquid in soil is possible if the distance between particles \( L \) is equal to or greater than \( L^* \); when \( L \) is less than \( L^* \), the liquid remains immobile.

Heteroporous soil possesses a definite size distribution of pores. If \( \phi(L) \) is the distribution function of pore volume by size, the active porosity may be expressed in the form [6]

\[
n_a = n \left[ 1 - \int_{L^*}^{L_{\text{max}}} \phi(L) \, dL \right].
\]

It is hence clear that the active porosity tends toward zero when \( L^* \to L_{\text{max}} \), and it is equal to total porosity when \( L^* \leq L_{\text{min}} \). In case of normal distribution of pores according to size, the active porosity will be

\[
n_a = n \left[ 1 - \Phi \left( \frac{L^* - \langle L \rangle}{b} \right) \right],
\]

where \( \Phi \left[ (L^* - \langle L \rangle)/b \right] \) is the probability integral, \( b \) is standard deviation, and \( \langle L \rangle \) is the average size of pores.