Mechanism of Frost Heave and Its Relation to Heat Flow

Mécanisme du gonflement dû au gel et sa relation avec le flux de chaleur

S. FREDEN, *Research Engineer, Geological Department, National Swedish Road Research Institute, Stockholm, Sweden*

SUMMARY

A study of the relation between the frost heave and the thermal conditions in soils has led the author to propose a new theory of the physics of frost heaving. From this theory it is possible to deduce that the rate of frost heave is approximately proportional to the temperature gradient in the frost active zone. This has been confirmed experimentally for some soils.

SOMMAIRE

Une étude de la relation entre le gonflement dû au gel et les conditions thermiques dans les sols, a permis à l'auteur de présenter une nouvelle théorie de la physique de gonflement dû au gel. De cette théorie il est possible de déduire que la vitesse de gonflement dû au gel est approximativement proportionnelle à la chute de température dans la zone active du gel. Ceci a été confirmé par expérience avec quelques sols.

IT IS KNOWN that the rate of frost heave is a function of soil characteristics, such as particle size distribution, types of minerals, and so on. It is also a function of other variables, such as load and pore water pressure, and is dependent on the thermal conditions in the soil. As it is of great practical importance to know the relation between the rate of frost heave and the thermal conditions, this relation has been investigated in a study done at the National Swedish Road Research Institute, Stockholm. The first part of the investigation was devoted to a tentative hypothetic description of the mechanism which makes it possible for ice to grow at the boundary of the frozen zone while still exerting pressure on the soil. The second part comprised a number of experiments carried out in order to verify the correctness of this hypothesis.

The scope of this investigation has been limited in some important respects: (1) The types of soils included in this investigation were assumed to be water-saturated. (This is the only case that is of interest in this connection.) (2) The percentage of clay minerals contained in the soil samples under test was assumed to be low or nil. (This limitation is temporary.) Finally, it should be observed that the investigation herein described has not yet been completed. However, it was considered that some of the results obtained might be of practical value, and their partial publication at this stage was therefore regarded as justifiable.

This report deals mainly with those portions of the investigation which directly concern the mechanism of frost heave and its relation to heat flow, and does not touch upon the studies of the influence of such characteristics as particle size, pore water pressure, or compressive load.

WATER IN SOIL

The water present in a soil can to a certain extent be considered to be adsorbed, if this term is used in a broader sense than usual. No strict definition will be given here, but generally it can be stated that all water present in the soil above the ground water table in a soil profile is adsorbed, the term "water table" designating the surface at which the water pressure is equal to the atmospheric pressure as long as the water remains stationary.

The capillary potential is often used as a measure of the adsorptive forces. This potential is defined (by analogy with the electric potential) as the work per unit quantity of water done in bringing an infinitesimal quantity of water from any point under consideration to a reference surface. It is convenient to choose as a reference surface a pool of pure water at the same level as the point under consideration (in that case, the work required to overcome gravitation will be equal to zero) or a pool of pure water at the same level as the water table (defined in the above). Unless otherwise specified, the first-mentioned definition of the capillary potential will be employed. In practice, it is much too inconvenient to determine the capillary potential in terms of work, and its measurement can be simplified in conformity with the following reasoning. If the free water is allowed to come in contact with the adsorbed water in a soil, then the water will be sucked up in the soil. This absorption of water can be prevented if the adsorbed water is subjected to a higher pressure than the free water. The difference between the pressures of free and adsorbed water at the instant when the velocity of water flow is zero constitutes a measure of the adsorption of water in the soil. If the denominations are appropriately chosen, then this difference will be numerically equal to the capillary potential as defined above. The observed difference in pressure is often somewhat improperly denoted by ψ and is termed "capillary potential," or suction, as will also be done in the present report. Its common logarithm is denoted by pF. For the rest, the measuring procedure itself will not be dealt with in this paper (see Coleman and Marsh, 1961; Croney and Coleman, 1960).

The difference in the state of energy between the free water and the water adsorbed in a soil is reflected in its effects on the physical properties of the water, e.g., the vapour pressure and the freezing point. To be able to calculate these effects, we must convert the capillary potential into the chemical potential, or affinity (usually denoted by μ) which is commonly used in physical chemistry. The relation between these potentials has been discussed very thoroughly in the literature ever since Schofield and Botelho da Costa (1935) showed that it was possible to use the lowering of the freezing point of water to determine the capillary potential (Takagi, 1959).

In this connection, it is not possible to deal with these questions in detail. However, the discussion in Schofield and Botelho da Costa (1935) can be adumbrated as follows. If the water is present in more than one phase, and if these phases are in equilibrium (e.g., water and water vapour or water and ice), then both these phases must have the same chemical potential. The water vapour which is in equilibrium with the water that has a lower chemical potential than the free water must also have a lower chemical potential than the water vapour which is present above the free water. The chemical potential of water vapour varies with the pressure in such a way that a lower potential corresponds to a lower pressure. Consequently, the vapour pressure above the adsorbed water is lower than that above the free water. Moreover a lower potential corresponds to a lower freezing point (defined as the temperature at which ice and water are in equilibrium). The last-mentioned relation is important in this connection. If the lowering of the freezing point does not exceed about one degree centigrade, then the relation between the lowering of the freezing point, $\Delta\Theta$, and the capillary potential can be expressed with a sufficient accuracy by the simple formula $\Delta\Theta = \psi/C_{ice}$. The constant C_{ice} in the above formula is dependent on the denominations of the temperature and the capillary potential. If the temperature is expressed in degrees centigrade and the capillary potential, ψ, in $gf \cdot cm^{-2}$, then $C_{ice} \approx 1250$. In this connection, it is necessary to emphasize that this equilibrium of the liquid phase and the vapour pressure or the temperature of ice formation is referred to the interface between the phases. This implies that the possible curvature of the interface must be taken into account, because this curvature changes the chemical (and capillary) potential of the interface.

FROST HEAVE

Frost-susceptible soils are characterized by the fact that their water content can increase in connection with freezing under certain conditions, and that this increase is so great (or the primarily available, not water-filled space in the voids is so small) as to cause an increase in the total volume of the soil. In practice, for mechanical reasons, this increase in volume takes place in a vertical direction.

Let us consider a sheet of ice which is pressed not too tightly against a mineral surface. Experience shows that the ice can grow in such a manner as to move away from the mineral surface, and then water molecules must be added to the crystal surfaces which constitute the boundary between the ice and the mineral (diffuse, internal growth is unknown in crystallography, and may be ignored in this connection). Furthermore, we can assume that the water molecules are transferred in the boundary layer between the ice and the mineral to the points where they are finally incorporated in the crystal lattice. Accordingly, between the ice and the mineral, there exists a boundary layer which possesses the following properties. (1) It has a relatively high degree of mobility. (2) It exists at temperatures below 0 C. (3) The ice growth can also take place at a pressure between the ice and the mineral surface. Therefore, the boundary layer must be capable of resisting this pressure without losing its properties. We can assume that this boundary layer has a structure which corresponds approximately to that of water.

It has previously been mentioned that the capillary potential at the interface or in the boundary layer between the water and the ice is determined by the expression $\psi = -\Theta \times C_{ice}$, where Θ is the temperature, and that the thickness of this layer decreases as the temperature becomes lower. If Z denotes a vertical co-ordinate axis, and if we assume a vertical temperature gradient which is directed downwards (i.e., the temperature increases in a downward direction) and which is constant at least above the zero isotherm, that is the boundary of the frozen zone, then the above equation can be transformed into

$$-d\psi/dZ = C_{ice} \times d\Theta/dZ.$$

At the distance Z_1 from the zero isotherm, the capillary potential is

$$\psi_1 = Z_1 \times C_{ice} \times d\Theta/dZ.$$

If we suppose that the capillary potential corresponds to, and denotes, a pressure of an equal numerical magnitude, then the water film enclosed between the mineral surface and the ice surface will exert on these surfaces a pressure which can be calculated from the above formula. If the counterpressure due to the mineral and the ice is lower than the water pressure, then the water film will increase in thickness. In that case, the interface between the water and the ice will be displaced from its previous position of equilibrium, and the capillary potential of the ice surface will no longer correspond to the temperature at the instant under consideration. This will cause the ice to grow as it tends to recover the former position of equilibrium. Since the rate of ice growth is considerable even if the unbalance in capillary potential is very small (Kost, 1953), no perceptible displacement of the interface will take place, and the ice will grow while the position of equilibrium will on the whole remain unchanged. It follows that the force exerted by a frost-susceptible soil is caused by the differences in the capillary potential, and that the ice, considered from a mechanical point of view, acts only as an inactive matrix.

In order to calculate the quantity of water which flows in an upward direction in the water film, it is convenient to use Darcy's law, which may be assumed to be applicable in the interval that is of interest in this connection: $v = k \times d\psi/dZ$ where v is the velocity of water flow, that is the volume of water flowing per unit area per unit time, k is the hydraulic conductivity, and the co-ordinate axis Z is coincident with the direction of water flow. From this law, we obtain the velocity of water flow as a function of the temperature gradient $- v = k \times C_{ice} \times d\Theta/dZ$.

The process thus described is bound up in many ways with the concept of osmotic pressure, and it is quite possible to carry out the above reasoning on the basis of an imagined distribution of ions at the mineral surface and by applying the theories of osmosis. However, such a procedure would be complicated, and it appears to be most convenient to connect our reasoning as closely as possible with the capillary potential and the lowering of the freezing point, which are directly measurable quantities. All the same, in certain cases—perhaps, primarily, in studying the effects produced on the frost heave by the ions which are present in the water—a representation of the process of frost heave based on "osmotic" considerations can be most advantageous. Be that as it may, since only the basic processes involved in frost heave are to be dealt with in what follows, such a discussion lies beyond the scope of this report.

MOTION OF INDIVIDUAL MINERAL PARTICLES IN ICE UNDER ACTION OF TEMPERATURE GRADIENT

From the hypothesis which has been adopted in order to explain the mechanism of frost heave, it may be inferred that individual mineral particles in water which are situated close to the zero isotherm can perform an upward motion under the action of a temperature gradient.

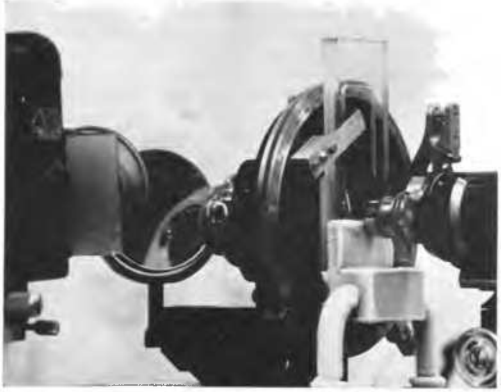

FIG. 1. Test set-up for studying frost heave of individual mineral particles and thin particle layers. This photograph shows, from left to right, lamp, mirror, polariser and condenser, stage with freezing cell, objective, and microscope tube with analyser.

In order that the process which takes place at the surface of contact between the ice and a mineral particle might be studied by microscopic examination, a special freezing cell, which can be mounted on the stage of a polarizing microscope, has been designed (Fig. 1). This cell consists of a glass container, whose approximate dimensions are 0.1 by 20 by 50 mm. The bottom of the cell is made of metal and is provided with a freezing coil. For experimental engineering reasons, the cell was designed for a vertical, upward temperature gradient. Accordingly, the ice, which is in this case fixed to the bottom of the cell, will be in the lowermost position, and the mineral particles to be studied will rest on the surface of the ice owing to gravity and will exert a slight pressure on this surface. If the temperature in the lower part of the cell is slowly decreased, at the same time as the upper part is kept at room temperature, then a horizontal ice surface can be caused to move slowly in an upward direction in the apparatus. If the sample consists of mineral particles in water, then it is found that they are first pushed upwards by the advancing ice surface, then gradually caught in the ice, immobilized, and finally completely embedded in the ice. If the ice surface does not move, then the particles which are not entirely embedded in the ice will move upwards, out of the ice. If the particles accumulate in the path of the advancing ice surface, then it can be observed how the particles are, as it were, compressed together to the highest possible degree of compaction, and how they at the same time move with the ice surface. The freezing cell is shown in Fig. 1, and the microphotographs (Figs. 2 to 4) represent the behaviour of quartz particles in contact with the ice. These particles are on an average 100 microns in size. Those particles which seem to be floating over the ice surface are in reality very loosely fixed to the glass wall of the cell by adsorption between the three components, that is, mineral, glass, and water.

FIG. 2. Microphotograph taken by means of the test set-up shown in Fig. 1. The ice is growing at its upper boundary surface, but the rate of heave of the mineral particles, i.e., the velocity of their upward motion with reference to the ice, is greater than the rate of ice growth, with the result that these particles are accumulated at the ice surface. Magnification about 100 ×.

FIG. 3. Microphotograph taken under the same conditions as that shown in Fig. 2, but representing only a few individual mineral particles. Magnification about 50 ×.

FIG. 4. Microphotograph taken in the same test as that shown in Fig. 3, but about five minutes later. By comparing Figs. 3 and 4, the motions of the mineral particles lying on the ice surface can be observed and measured. The particles which are completely embedded in the ice can be used as fixed reference points. Magnification about 50 ×.

From the microphotographs which show the growth of ice in the freezing cell described in the above, it can also be seen that the particles move with reference to the ice only if they are less than half embedded in the ice. However, if heat flows through a particle embedded in the ice, that is to say, if there is a temperature gradient in this particle, then we can suppose on purely theoretical grounds that the particle in question will be displaced in the ice for following reasons. If we assume that two opposite surfaces of the particle differ in temperature, then the capillary potentials of the water film which surrounds the particle will also be different at these two surfaces, wholly in accordance with the conditions which have been described. This leads to water flow and to growth of ice at the colder surface. The state at the warmer surface will be unbalanced in the opposite direction, and the ice will melt at this surface. This process should cause the particle to be displaced along, and in the direction of, the temperature gradient. Under the conditions stated in the above, the ice melts very slowly, and, to the author's knowledge, no such displacement has ever been mentioned in the literature. Nevertheless, this—negative—observation corroborates the hypothesis that it is only those mineral particles which are not immobilized by frost that actively participate in frost heaving.

RELATION BETWEEN RATE OF FROST HEAVE AND TEMPERATURE GRADIENT

The part played by the temperature gradient in the process of frost heave has already been dealt with at some length. However, we have assumed that this gradient is parallel to the direction of water flow. In the actual soils, all directions of mineral surfaces will obviously be represented. If the direction of a mineral surface deviates by an angle u from the direction of the temperature gradient which is in a normal way macroscopically measurable in a soil sample (grad Θ), then the temperature gradient in the direction of this surface (and of the associated water film) is given by the relation (grad Θ) $\cdot \cos u$. Since this angle is a constant, the above relation implies that the effective temperature gradient is lower than the corresponding macroscopic gradient. If this effect is summed over all surfaces, then it may be included as a correction in the hydraulic conductivity, k, because all calculations must be based on the vertical macroscopic temperature gradient.

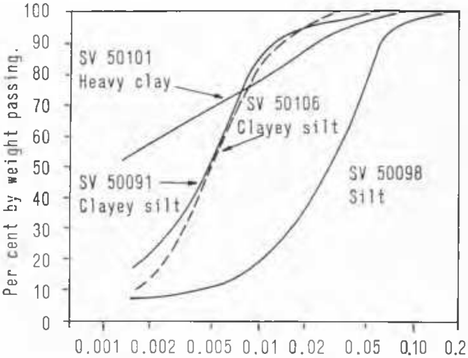

FIG. 5. Particle size distribution curves of the soils used in the frost heave tests.

To verify the hypothesis that the rate of frost heave is proportional to the temperature gradient in the layer in and above the frozen zone, frost heave tests have been carried out at the National Swedish Road Research Institute. Up to now (1963), three types of soils have been systematically studied in these tests, viz., a silty fine sand (SV 50098), a clayey silt (SV 50091), and a heavy clay (SV 50101). The particle size distributions of these soils are shown in Fig. 5. The tests in question were performed in an apparatus whose design is similar in principle to that which has been devised by Beskow (1935). In the apparatus used in these tests, the temperature at each end surface of the soil sample was regulated individually as a linear function of the time. The temperature of the sample was measured by resistance thermometers. The amount of frost heave was measured by the aid of a potentiometer circuit which was connected to a multipoint recorder. In these tests, the rate of penetration of the boundary of the frozen zone was varied from 10 to 0.5 mm × h^{-1}. Within this range, it was not possible to demonstrate any effect of the rate of frost penetration on the rate of frost heave. The test results are represented in Fig. 6.

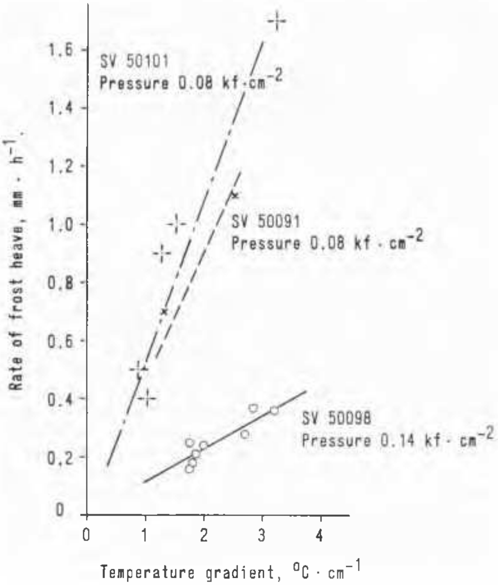

FIG. 6. Relation between the rate of frost heave and the temperature gradient for three different soils.

As may be seen from the graph in Fig. 6, a linear relation between the rate of frost heave and the temperature gradient exists. The dispersion in the test results may be attributed to the relatively large errors in measurements. In particular, the measurements of the temperature gradient are unreliable at low rates of frost heave because layers of stalky and porous ice are often formed under such conditions, and it is probable that the thermal conductivity of these layers is markedly lower than that of the other portions of the soil. Since the calculations are based on a difference in temperature which is measured over a distance of several centimetres, and on the necessary assumption that the soil is homogeneous in this region, the actual value of the temperature gradient in the frost-active layer can differ considerably from its calculated value.

Moreover, a small number of tests have been carried out with a new apparatus, which is equipped with a heat flow meter. As the temperature gradient is proportional to the rate of heat flow, it is to be expected that there also exists a linear relation between the rate of heat flow and the rate of frost heave. This assumption seems to have been confirmed by the tests in question, which were made on an

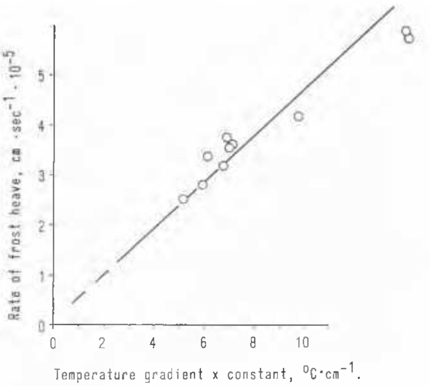

FIG. 7. Relation between the rate of frost heave and the temperature gradient for a silty sand.

artificial soil (SV 50106) consisting of powdered quartz produced by grinding (Fig. 7; its particle size distribution is shown in **Fig. 5**).

For the types of soils referred to, the hypothesis that the rate of frost heave is directly proportional to the temperature gradient in the zone of frost heave may be regarded as well verified by the tests described in this report.

REFERENCES

BESKOW, G. (1935). *Tjälbildningen och tjällyftningen* (Soil freezing and frost heaving). National Swedish Road Research Institute, Proc., No. 48, 242 p. (English translation published by Techn. Inst., Northwestern University, Ill., USA, 1947.)

COLEMAN, J. D., and A. D. MARSH (1961). An investigation of the pressure membrane method for measuring the suction properties of soil. *Jour. Soil Science* (London), Vol. 12, pp. 343–462.

CRONEY, D., and J. D. COLEMAN (1960). Pore pressure and suction in soils. *Conference on Pore Pressure and Suction in Soil* (London), pp. 31–7.

KOST, W. (1953). Vorgänge bei der Eisbildung. Dissertation, Technische Hochschule Karlsruhe.

SCHOFIELD, R. K., and J. V. BOTELHO DA COSTA (1935). The measurement of pF in soil by freezing point. *Jour. Agricultural Science*, Vol. 28, pp. 644–53.

TAKAGI, S. (1959). Theory of freezing-point depression of soil water, and a note on the extra-thermodynamics of soil moisture. *Soil Science*, Vol. 88, pp. 25–31.