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Migration of Moisture in the Thermal Regime

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Table of Contents

	<u>Page</u>
Preface	iii
Introduction	iv
Section 1. Thermodynamic Considerations Related to Moisture Movement at Low Temperatures	1
I. Some General Definitions and Experimental Approaches	1
II. Thermodynamic Considerations	6
III. Freezing Point and Its Relation to Pressure	14
Section 2. Capillarity of Soil Related to Moisture Migration	18
I. Liquid Migration	19
II. Measurement of Capillarity Conductivity	21
III. Capillarity and Soil Deformation	27
IV. Vapor Migration	35
V. Compression Forces in Capillary Migration	36
Section 3. Moisture Migration Through Snow	44
Section 4. Thermal Gradients and Thermal Translocation of Moisture	48
Section 5. Frost Penetration as Influenced by Soil Type, Soil Moisture, and Soil Cover	58
I. Measurement of Frost Penetration	58
II. Factors Affecting Frost Penetrations	60
III. Freezing Index Related to Frost Penetration	62
IV. The Prediction of Depth of Frost Penetration	63
Section 6. Migration of Saline Solutions	74
I. Migration of Sodium Chloride, Calcium Chloride and Calcium Magnesium Chloride Solutions at Low Temperatures	74

Table of Contents
(Continued)

	<u>Page</u>
II. Migration of Naturally Occurring Sodium Sulfate in Soil Related to Humidity	77
III. The Effect of Water Vapor on the Migration of Salts	78
Section 7. Barometric Pressure as a Function of Moisture Migration	81
Section 8. Deterioration of Building Materials by Moisture Penetration and Frost Action	83
I. Permeability in Porous Solids	85
II. Suction	86
III. Capillary Rise in Porous Medium	89
IV. Mechanical Effects Produced by Crystallization of Solids in Porous Bodies	90
Section 9. Conclusions and Recommendations	94
Summary	99
Bibliography	103

Preface

This report is the result of a critical literature survey of the published information on migration of moisture in the thermal regime. This has been prepared by Dr. Robert C. Brasted, Associate Professor of Chemistry at the University of Minnesota, with the aid of others on the staff prosecuting associated researches in the field of snow, ice, and permafrost. In particular, Dr. Homer T. Mantis, Ass't. Professor, Mechanical Engineering, aided in the development and editing of the manuscript and Dr. R. G. Mokadam, Research Assistant, Department of Mechanical Engineering, provided extensive help in the development and enlargement of the bibliography, as well as specific contributions in such areas as frost penetration, soil deformation and moisture migration through snow cover.

This work has been prosecuted under Contract DA-21-018-ENG-256 established at the University of Minnesota by the Snow, Ice and Permafrost Research Establishment, Wilmette, Illinois, for the purposes of accomplishing literature research in the fields of snow, ice, and permanently frozen ground.

University of Minnesota
June, 1954

Richard C. Jordan
Project Director

Introduction

The purpose of this survey is to summarize the information available in the open literature on moisture migration. Although the emphasis is placed on studies reported at low temperatures, it is necessary to include certain data made at temperatures above freezing to serve as a foundation for later topics.

Certain of the areas, including those of soil-deformation saline solution migration, and migration of moisture through building materials are not intended to be exhaustive treatments. Only material related to or at least shedding some light upon the thermal regime are considered. Thus, the great mass of material allied to frost-boils, pavement failure, their cure and/or prevention are included only when data pertinent to moisture migration are included.

An attempt has been made to be critical in reporting experimental observations. Thus, certain conclusions may be drawn by an investigator without regard to variables which might well affect experimental data.

Areas which need additional experimentation are noted, as are experimental techniques (used at temperatures above 0°C) which might be applicable and useful for further studies of moisture migration in the thermal regime.

A great deal of pertinent information is to be found in the Russian literature. The writer was obliged to depend upon abstracts provided either by the usual abstracting services (such as Chemical Abstracts, British Abstracts and Zentralblatt) or from SIPRE file

abstracts for material printed in the Russian language. This survey may not be considered as complete (or critical) as would be desired without original reference translations of articles originating within the U.S.S.R.

Bibliographical notations are included within the body of the manuscript by author and journal reference. References are also arranged alphabetically (by authors) at the end of each section. A supplementary bibliography is included for certain of the sections.

Section 1.

Thermodynamic Considerations Related to Moisture Movement at Low Temperatures

Section I. Thermodynamic Considerations Related to Moisture Movement at Low Temperatures

I. Some General Definitions and Experimental Approaches

There is no complete agreement among authors [R. K. Frevert and D. Kirkham, Highway Research Board, Proc., 28, 433 (1948)] in the definition of terms related to the rate of movement of water in soil. Generally speaking the ability of the soil to transmit water is referred to as the permeability of soil. Certainly, moisture (vapor or liquid) migration in soils is a direct function of the permeability of the soil. Permeability has been defined exactly by Muskat [M. Muskat, "The Flow of Homogeneous Fluids Through Porous Media", McGraw-Hill, Publishers, New York, p. 71, (1937)] as the volume of a fluid of unit viscosity passing through a unit cross section of the medium in unit time under the action of unit pressure gradient. Since in a soil structure the fluid under consideration is almost without exception water, and gravity is the driving force, Muskat has suggested the use of an "effective permeability unit": the quantity of water passing in unit time through a unit cross section under a unit hydraulic gradient.

Frevert and Kirkham [loc. cit.] differentiate between percolation, permeability and infiltration. Percolation is considered to be the general movement of water through a soil profile. Such movement is, for the purpose of this discussion, not necessarily dependent upon the capillary nature of the soil system. As the pores in the soil become irregular or smaller, the frictional resistance to flow increases. When the pores become so small in combination with other physical characteristics that capillary tension becomes greater

than the force of gravity, subsurface drainage is not effective. Soils containing more than 70 per cent clay are practically non-percolating. Infiltration has been defined by Baver [L. D. Baver, "Soil Physics", John Wiley and Sons, Publishers, New York, p. 341 (1940)] as a process whereby, water enters the environment of the soil through the immediate surface. Horton [R. E. Horton, Proc. Central Snow Conference, 1, 5-21 (1941)] has elaborated considerably upon the phenomenon of infiltration and defined the term infiltration capacity, f , as the maximum rate at which the soil surface can absorb water (as from rain or melting snow). If the rain intensity or rate of melting is less than the infiltration-capacity, the water will all be absorbed. If greater, the excess of the supply over infiltration-capacity will build up surface detention and produce runoff.

On unfrozen soil infiltration-capacity is usually initiated when rain begins at a soil infiltration capacity value f dependent on dryness, structure and texture of the soil surface. As a rule f decreases rapidly for a time, attaining a constant value (fc) after one to three hours. Between rains the infiltration-capacity tends to recover its original higher value as the soil dries out and its natural structure is restored.

One of the most important factors governing the reduction of infiltration-capacity during rain is the effect of energy of falling rain in breaking down the crumb structure of the soil surface. The soil surface is packed and washed free of fine material into sun-cracks and into the large soil pores. In summer, vegetal cover if present breaks the force of falling rain and the infiltration-capacity therefore is quite high. Root perforations, earthworm and insect burrows and

other biologic structures play an important role. In general, snow cover acts in much the same way as vegetal cover and maintains high infiltration-capacity of the soil.

Swelling of soil colloids with wetting affects infiltration-capacity. Altogether, infiltration-capacity is controlled chiefly at the soil surface, and water can usually pass down through the soil, after it once has entered, faster than it can enter the soil surface. Hence the great soil surface is rarely saturated during rain. Another reason for this lack of saturation is that air must escape upward in an equal volume and through the same system of soil pores through which water enters the soil.

If the ground is unfrozen when snow falls, the soil surface immediately becomes wetted by melting at the contact surface between snow and soil. The soil surface remains wet and, as a rule, the infiltration-capacity remains at a constant minimum value when the snow disappears.

When the ground surface becomes frozen, part of the void space otherwise available for downward flow of infiltration is replaced by ice, decreasing the infiltration-capacity. This reduction is, however, not proportional to the reduction of void space, for various reasons, one being that the frozen soil-moisture chiefly occupies the smaller pores, while the larger pores are more effective in the transmission of water into and air out of the soil.

Since the soil does not become saturated appreciably above its capillary-capacity during rain, it is seldom saturated when frozen, although it may become saturated later on at the surface or a wholly impervious glazed surface may be formed by rain falling on soil the

temperature of which is below freezing. Swelling of water in freezing and temperature effects are also involved. On the other hand, the soil may freeze "dry", or with little initial moisture, and then receive a snow cover and thereafter maintain a relatively high infiltration-capacity until melting begins. Excluding heavy clay soils, most soils maintain an appreciable residual infiltration-capacity when frozen wet or dry.

According to the definitions noted for percolation and permeability they should have the same numerical values. It is the assumption of the writer that percolation, permeability and infiltration are of secondary importance in low temperature studies of moisture migration. Such a statement is not made to minimize the importance of such processes of migration in engineering problems; however, the condition of very low temperatures assumed at least over a portion of the year precludes free migration of liquid water in large quantities as by percolation. It is highly probable that the process is migration by capillarity associated with vapor migration.

The experimental technique and results of Frevert and Kirkham in determining water permeability below the level of the water table by two experimental procedures are described in some detail since their work appears to be almost unique in the field.

In one method Kirkham [D. Kirkham, Proc., Soil Sci. Soc of America, 10, 58-69 (1945)] advises the insertion of a pipe the same size as a hole bored into the water table. The pipe is filled with water to a measured height above the water table and fall in this water level is recorded over an interval of time. The permeability is then

calculated from the expression

$$K = \frac{\pi R^2 \ln(h_1/h_2)}{A(t_2 - t_1)}$$

where K is the permeability in inches per hour for unit hydraulic gradient; h_1 and h_2 are the distances of water table level in the pipe from the water table at times t_1 and t_2 respectively; R is the radius of the pipe. A is the coefficient determined by an electric analogue

$$A = \frac{R}{R_m \sigma \omega}$$

where R is the radius of the pipe as before; R_m is the radius of a model; σ is the specific conductivity of electrolyte used in the analogue; and ω is the resistance between electrode and tank bottom. Frevert [R. K. Frevert, Thesis, Library, Iowa State College, Ames, Iowa (1948)] has modified Kirkham's procedure, in that instead of pouring water into the pipe to create a higher hydraulic head inside than out, water is removed so that the soil water (from a water table) moves in. The water levels h_1 and h_2 are then distances from the water table to the initial (lower) and final (higher) levels respectively. The value of h_1 is then greater than h_2 as is the case for Kirkham's equation. The theory for either method is the same. The value of the A -function can be calculated analytically for simple cases of flow in porous media,

$$K = \frac{\ln(h_1/h_2)}{A(t_2 - t_1)} s$$

where A is $\pi R^2 / s$ and s is the depth of the actual soil column. The A -function from a physical standpoint takes into account the flow

pattern of the water in the soil. Frevert has listed four advantages in the use of his procedure for measuring true permeability of soils below the water table by the use of flow of water into a pipe rather than flow of water out of the pipe.

1. Natural structures of soil are undisturbed.
2. Temperature corrections to a standard are possible. Frevert has corrected all permeabilities to 20.2°C.
3. Effects of trapped air are eliminated when the soil has been saturated long enough for the air to have been absorbed in the soil water.
4. Irregularities in the soil are minimized by using a fairly large area of the cross-section of sample.

The permeability data by the methods of either Kirkham or Frevert would be of interest if a temperature gradient approaching 0°C were included. The influence of vapor migration in permeability below the water table would, of course, be minimized or even eliminated.

II. Thermodynamic Considerations

Anderson and his co-workers [A. B. C. Anderson, J. F. Fletcher and N. E. Edlefsen, Trans. Am. Geophys. Union, 23 (2), 356-71 (1942)] have considered the basic thermodynamic theory necessary to interpret the observations related to the movement of moisture associated with frozen soils. The freezing of soil samples with various soil moisture contents was studied by two standard techniques, the Beckmann and Dilatometric. Consistent results were obtained by both methods as proven by the fact that plots of freezing point depressions vs. moisture content (per cent by weight, dry basis) agree in their general trend.

Since Anderson's derivation interrelating the thermodynamic variables of state under different conditions of freezing soil-moisture is unusually complete and is one of the few found in existing literature (1953), it is described in some detail.

The interrelations among the thermodynamic quantities describing the states of moisture for frozen soils have also been expressed by Schofield [R. K. Schofield, Trans., Soil Sci., Third International Cong., 2, 27-48 (1935)]. Anderson [loc. cit.] and Schofield have independently arrived at the same general expression, noted below as (3). Schofield has attributed the depression of the freezing point of soil solution entirely to dissolved material. Also it must be assumed that the water at the water-ice interface in the soil-moisture is under a different pressure than that in the ice. Otherwise stated, as water under tension freezes out, it separates under the higher pressure of one atmosphere. There is presently insufficient experimental data to prove the latter point, however. A surprisingly high percentage of water in certain soils remains unfrozen even at temperatures well below freezing. Tsytovich [N. A. Tsytovich, Izvestiia Akademii Nauk SSSR, Seriia geologicheskaya, No. 3, 39-48 (1947); see also SIPRE File 1841] reports data on Moscow loam containing 21 per cent of fine clay particles and 21 per cent of water. On being subjected to a temperature of -1.6°C , 74 per cent of the water remained unfrozen; at -17°C , 57 per cent of the water was unfrozen; and at -24°C , 50 per cent was unfrozen. Mineralization of ground water, of course, contributes to a certain extent to the resistance to freezing.

According to Anderson et al [loc. cit.] the total free energy Δf of the moisture at any point B in the soil moisture may be considered

as the sum total of a number of component free energies:

$$\Delta f = \Delta f_p + \Delta f_o + \Delta f_F + \Delta f_K + \Delta f_\sigma \quad (1)$$

Each of the component free energies is herein defined.

Δf_p represents the free energy of the component at a point B due to hydrostatic pressure. This hydrostatic pressure is assumed to be in excess of that normally present in free, pure water under one atmosphere of pressure. If ΔT_p is the freezing point depression due to the hydrostatic pressure, ΔP_p , equation (2) describes Δf_p .

$$\Delta f_p = V \Delta P_p = (\lambda \Delta T_p / T \Delta V) \quad (2)$$

where λ is the latent heat of fusion of ice, ΔV is the change in specific volume when ice melts, and V is the specific volume of water (unity in cgs system). A necessary assumption in the derivation of (2) is that the hydrostatic pressure, ΔP_p , is applied equally on both the water and ice during freezing.

Δf_o represents the component of total free energy at the same point B due to the presence of dissolved material in the soil moisture (cf. Schofield, above). If ΔT_o is the freezing point depression due to the osmotic pressure ΔP_o , then Δf_o is defined by equation (3).

$$\Delta f_o = -V \Delta P_o = -(\lambda \Delta T_o / T). \quad (3)$$

The thermal component, Δf_F , represents the force fields, including gravitational, and absorption surrounding the soil-particle. A line integral is needed to describe Δf_F .

$$\Delta f_F = \int_A^B F \cos \phi \, dx \quad (4)$$

where A is considered to be the datum (located in a free, pure body of water at a given point in the gravitational field) and B is the point

location of the water particle under the influence of the field F surrounding the soil-particle. The angle between the element of the path dx between A and B and the direction of the force F is ϕ . The $\cos \phi$ is positive if F and dx are in opposite directions.

The fourth component Δf_K represents the free energy due to the kinetic energy of the water at point B due to mass movement.

$$\Delta f_K = (V^2/2) \quad (5)$$

where V is the streaming velocity.

Δf_σ represents the increase in total free energy of a liquid surface A when its area is increased by the amount ΔA , or

$$\Delta f_\sigma = \sigma \Delta A. \quad (6)$$

Equation (6) must be altered to be consistent in units. If T represents the thickness of the surface layer of water in which the properties are different from those of the body of the fluid, then the average free energy per gram throughout the surface layer may be represented by

$$\Delta f_\sigma = \frac{\sigma}{T\rho}$$

where ρ represents the average density of water in the layer.

Δf_s is used by Anderson to represent the total free energy of the soil-moisture. This quantity is expressed in terms of the vapor pressure of the soil-moisture p and the vapor pressure of the datum.

$$\Delta f_s = RT \log_e (p/p_\sigma). \quad (7)$$

The constancy of Δf_s depends upon the equilibrium conditions of the soil or moisture. If the latter is in motion (as during moisture migration) Δf_s is not the same at all points. Even at equilibrium where Δf_s is the same at all points the components Δf_F and Δf_p of total free energy Δf (equation 1) may vary considerably from point to point in the soil moisture.

Anderson points out an important relation between freezing-points and free-energies of soil-moisture, not obvious without the two expressions (2) and (3). Although a positive hydrostatic pressure and dissolved material lower the freezing point of a solution, these two factors are not additive in producing changes in total free energy. From equations (2) and (3) it may be seen that an increase of hydrostatic pressure ΔP_p causes an increase, whereas a similar change in the osmotic pressure ΔP_o causes a decrease in the total free energy of the soil moisture. It is entirely possible that the component free energies produced by P_p and P_o may be of equal magnitudes but opposite in sign. The above treatment does not permit an explanation of whether ice formed, when moist soil is chilled, is under the original tension, or is separated out from the main body of the liquid soil-moisture at only atmospheric pressure. The latter pressure is assumed to be standard with respect to the pressure of Anderson's datum.

Consider the system represented schematically by Figure 1. In this system P_I represents the pressure exerted uniformly on ice by a piston that is permeable to water-vapor only and P_W represents the pressure exerted uniformly on water by a piston that is permeable likewise to water-vapor only. P_W and P_I are always so related that there is a common vapor pressure, p , throughout. There must be a

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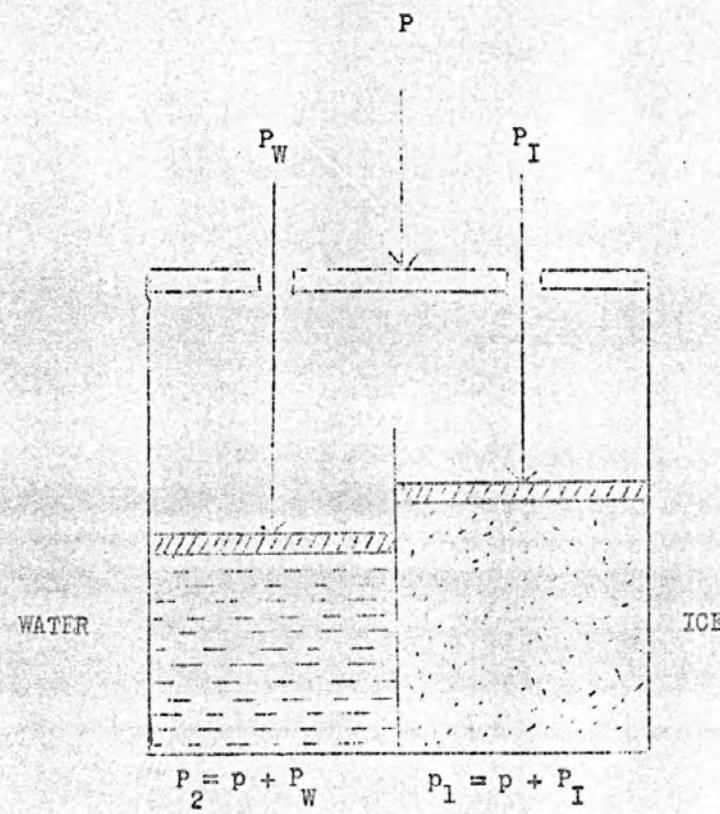


FIG. 1

constant equilibrium between the liquid, vapor, and solid phases. The total hydrostatic pressures in the water and in the ice are then

$$P_2 = p + P_W$$

$$\text{and } P_1 = p + P_I$$

respectively. Such a condition is necessary since the vapor pressure is acting on both the solid and liquid phases.

Assuming T to be the freezing-point of water under the conditions described (that is the temperature at which all three phases are at equilibrium), then the following differential expressions relating dT (change of freezing point) to a change in one or all of the pressures may be derived.

At equilibrium it is a necessary thermodynamic condition that the free energies f_1 and f_2 be equal.

$$f_1 = f_2 \quad (8)$$

and any changes in f_1 must be accompanied by a change in f_2 or

$$df_1 = df_2 \quad (9)$$

The free energies of water and ice may be expressed in terms of any two variables, for example T and F. Thus

$$df_1 = \left(\frac{\partial f_1}{\partial P_1} \right) dP_1 + \left(\frac{\partial f_1}{\partial T} \right) dT \quad (10)$$

$$\text{and } df_2 = \left(\frac{\partial f_2}{\partial P_2} \right) dP_2 + \left(\frac{\partial f_2}{\partial T} \right) dT \quad (11)$$

From elementary thermodynamic considerations

$$\left(\frac{\partial f}{\partial P} \right) = v \quad \text{and} \quad \left(\frac{\partial f}{\partial T} \right) = -S \quad (12)$$

Therefore equations (10) and (11) become

$$df_1 = v_1 dP_1 - S_1 dT \quad (13)$$

$$\text{and } df_2 = v_2 dP_2 - S_2 dT \quad (14)$$

Substituting (13) and (14) into (9)

$$v_1 dP_1 - S_1 dT = v_2 dP_2 - S_2 dT \quad (15)$$

where dP_1 and dP_2 are the total changes in pressure on the solid and liquid states respectively. Whereas, it is customary to assume an equality in pressures, in the soil-system considered these two pressures are not, in general, equal to each other. By substituting dP_1 and dP_2 (from the original statements $P_1 = p + P_I$) (and $P_2 = p + P_W$) into equation (15) and dividing through by the freezing temperature differential dT , the result is

$$v_1 \left(\frac{dp}{dT} + \frac{dP_I}{dT} \right) - S_1 = v_2 \left(\frac{dp}{dT} + \frac{dP_W}{dT} \right) - S_2 \quad (16)$$

On rearrangement of equation (16) the result is

$$(v_2 dP_W/dT) - (v_1 dP_I/dT) + (dp/dT)(v_2 - v_1) = (S_2 - S_1) \quad (17)$$

The term $(S_2 - S_1)$ in equation (17) represents ΔS or the difference in specific entropies of ice and water at equilibrium, which statement may also be represented by

$$\Delta S = S_2 - S_1 = (\ell/T) \quad (18)$$

where ℓ is the latent heat of fusion of ice. The final equation is then given in the form

$$\left(\frac{v_2 dP_W}{dT} \right) - \left(\frac{v_1 dP_I}{dT} \right) + \left(\frac{dp}{dT} \right) (v_2 - v_1) = (\ell/T) \quad (19)$$

where dP_I and dP_W are (as described previously) the total changes in pressure on the solid and liquid respectively, v_1 and v_2 are specific volumes of solid and liquid respectively and dT is the change in freezing point.

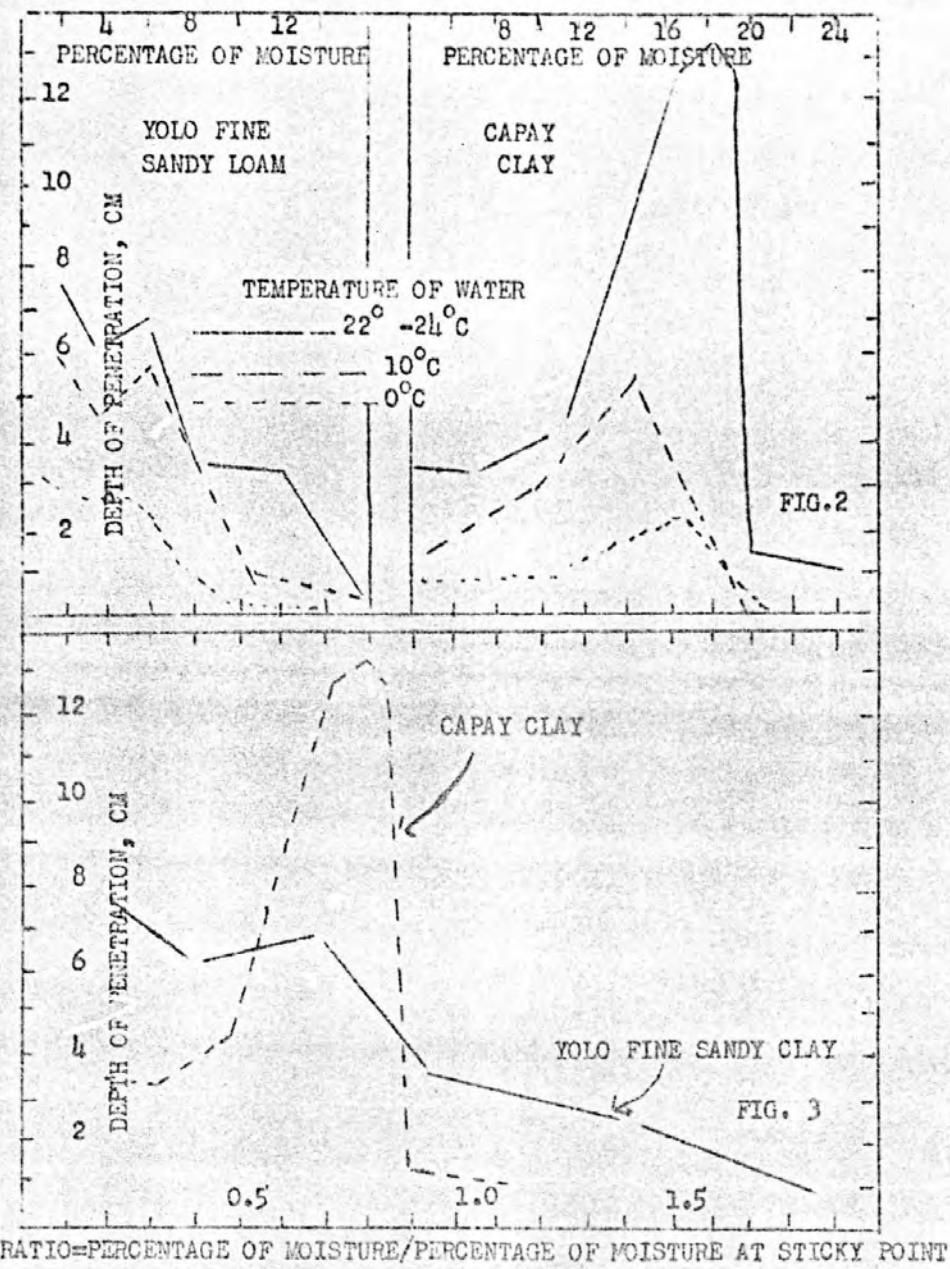
Anderson and his co-workers [loc. cit.] have made observations pertinent to the movement of soil-moisture at low temperatures using

the above thermodynamic considerations as a basis for their reasoning. If soil-moisture is subjected to a temperature lower than its freezing point, at least a portion of the moisture freezes. It is likely that the portion of water frozen will be that which is farther from the surface of the soil-particle, or that which in the liquid state possesses the greatest mobility. "This fact coupled with the consideration that the space formerly occupied by fluid is then occupied by a solid requiring slightly more volume makes it evident that the motion of water under a given force through a frozen soil is markedly curtailed." Two types of water movement related to freezing soils are considered.

One movement is that downward under gravitational force. Such movement has particular bearing on soil-erosion and watershed run-off. Only a few tests are available on moisture migration downward in frozen soils; however, it appears as if little moisture migrates through frozen soils, the depth of penetration is likewise slight.

Tests were conducted on Yolo fine sandy loam having a moisture-equivalent of 28.7, permanent-wilting percentage of 7.9 and a sticky-point of 8.5 as well as on Capay clay having a moisture-equivalent of 28.7, permanent-wilting percentage of 13.7 and a sticky-point of 23.1. The soils were frozen at -10°C . The data plotted in Figure 2 show the depth of penetration as a function of moisture-content of the particular soil. The location of the maxima indicate a structure factor which must be of importance. In Figure 3 the average depth of penetration is plotted as a function of the ratio of percentage moisture to percentage of moisture at the sticky-point. The soil

-13a-



structure factor is again emphasized by the maximum noted for the Capay clay.

It is likely that moisture falling on frozen soil will be nearly all lost by run-off. Further, it is logical that the soil below the frost line will be quite dry.

A thermodynamic consideration is, that as water penetrates the soil, the heat-content of the down-migrating water decreases until the water freezes. At the point of freezing an ice layer may form preventing further downward migration. The water content of the soil above the ice layer increases. The soil is thus made very susceptible to erosion.

A second type of moisture migration of importance in the thermal regime is that due to thermal gradients. Probably little moisture moves through frozen soil, but the movement is closely associated with some phase of soil-freezing phenomenon (See page 44).

III. Freezing Point and Its Relation to Pressure

Although using an entirely different line of reasoning, Hudson [C. S. Hudson, Phys. Revs. 22, 257-64 (1906)] arrived at the same final equation (19) as that described by Anderson. Four ways are described by Hudson in which the freezing point is changed by changes in pressure. These four situations are outlined briefly with actual values listed for dP and dT .

- (a) The total change in pressure on the ice is always equal to the total change in pressure on the water with $dP = 1 \text{ atm.}$, then dT is 0.00748°C.

- (b) The pressure on the water remains constant while the pressure on the ice is changed. $dP_{ice} = 1 \text{ atm.}$, at then is equal to 0.0899°C [note that dT for (b) is some 12 times greater than that for (a)].
- (c) The pressure on the ice remains constant while the pressure on the water is changed. $dP_w = 1 \text{ atm.}$, then dT is equal to 0.0824 . A comparison is to be made with Case (b).
- (d) The total changes in pressure on the ice and water are always so related that the freezing-temperature remains constant. Under such conditions the value of dT must be 0.

Case (c) above appears to be useful in describing the freezing of water brought up by migration through capillaries. Anderson makes the interesting supposition of an ice particle in a soil cavity some distance above the water table, above which table, capillaries are extending. The water in the capillaries is under tension and will, therefore, at 0°C have a lower vapor pressure than the particle of ice. Since the system is not at equilibrium the moisture system in the cavity will distill over to the water in the capillaries. If the water table is assumed to be $76 \times 13.6 \text{ cm.}$ below the cavity, then the hydrostatic pressure in the water (reaching the cavity) is $-76 \times 13.6 = -1 \text{ atm.}$ From Case (c) the ice-particle-soil-moisture system in the cavity will be in equilibrium at -0.0824°C . At this temperature that part of the soil-moisture adjacent to the cavity which is under tension of one atmosphere will begin to freeze.

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Section 2.

Capillarity of Soil Related To Moisture Migration

Section 2. Capillarity of Soils Related to Moisture Migration

Investigations of the laws of soil water movement are divided into three broad groups by Schofield [R. K. Schofield, Trans., Soil Sci., 3rd International Cong., 2, 37-58 (1935)].

In one, the known laws of viscosity and surface tension are used. The principles of viscous flow can be used successfully in the case of movements of water in porous soils below the water-table. In the second group which he classes as inductive, are the experiments indicated in the field.

The few laws that form the third group, to which Schofield contributes, have endeavored to establish energy relationships which are valid whatever the mechanism at work.

Schofield recommends the use of a pF scale. This scale is the logarithm of the height in centimeters of an equivalent water column which denotes the "suction" or Buckingham's "capillary potential" (see following section). A transfer of this scale to any liquid defining its pF can also be made.

The pF can be calculated directly or may be estimated from freezing point depressions and relative humidities.

$$H = (Tg/L_j)xt \quad (20)$$

$$pF = \log_{10} H = \log_{10} (Tg/L_j) + \log_{10} t.$$

$$= 4.1 + \log_{10} t \text{ (for water at freezing point)} \quad (21)$$

Also $H = -(RT/Mg) \log_e (h/100)$

$$= 2.303 (RT/Mg) \log_{10} (100/h) \quad (22)$$

$$pF = \log_{10} H = \log_{10} 2.303 (RT/Mg) + \log_{10} (2 - \log_{10} h)$$

$$= 6.5 + \log_{10} (2 - \log_{10} h) \text{ (for water at } 20^{\circ}\text{C}) \quad (23)$$

where

H = height of liquid column to give equivalent suction, cm.,

T = absolute temperature, $^{\circ}\text{C}$,

g = 981 ergs. $\text{gm}^{-1} \cdot \text{cm}^{-1}$,

l_f = latent heat of fusion, 3.336×10^9 ergs. gm^{-1} ,
for water,

t = freezing point depression, $^{\circ}\text{C}.$,

R = Universal gas constant 8.315×10^7 ergs. $\text{moles}^{-1} \cdot \text{C}^{-1}$.

M = Molecular weight in vapor phase 18.02 gm. moles^{-1}
(for water),

and

h = relative humidity per cent.

It is observed that the moisture content to which an ordinary agricultural soil can be dried by a given suction is considerably greater than that to which it can be wetted against the same suction. It is therefore assumed that for each moisture content there are two pF values, one for drying and the other for wetting, the one in the former case being greater than the one in the latter case.

Making due allowance for gravity it may be said that water will not move except under the conditions of a pF gradient. Even for small pF gradients movement may be slow due to low permeability. It is important to note that a moisture gradient does not necessarily involve a pF gradient.

I. Liquid Migration.

Lewis [M. R. Lewis, U. S. Dept. Ag. Tech. Bull. 579, Oct. 1937] has made a study of the rate of flow of capillary moisture with irrigation problems in mind. The greater importance of root movement or root growth to water sources than capillary movement of water to

the roots has been emphasized. The elucidation of the problem of moisture migration necessitates an understanding of capillary potential and capillary conductivity of different soils. The former is an expression for the force with which moisture is held at a given time and at a particular point in the soil by capillary action. It is further defined and used by Lewis as the work performed by the capillary force in moving a unit mass of water from a free water surface to the point in question. A capillary potential gradient would then be the rate of change from point to point of the capillary potential.

The second term defined by Lewis, capillary conductivity, is a measure of the soils' ability to transmit water when in an unsaturated condition; thus it is the ratio of the rate of flow of moisture in an unsaturated soil to the capillary potential gradient.

Lewis has made a number of measurements on the actual rate of flow of moisture under the influence of capillary forces acting in soil columns of different moisture contents. Since the approach made by Lewis was from a practical agricultural view point, over-dried soils were not considered nor were low temperatures.

The reports of various investigators on the volume, distance and quantity of water raised from a water table by capillary action are at such odds that it is difficult to correlate the data. Lewis has reviewed the work of others through the period 1937 and includes some 57 references in his paper. Fundamental information is reported on: (1) the relationship between moisture content of the soil and the capillary potential; (2) the relation between the capillary potential gradient and the rate of flow of capillary moisture; and (3) the specific conductivity of soil for moisture.

From a large amount of tabular and graphic data, it is concluded that finer textured soils will not transmit capillary water as readily as will coarse textured soils. For steady flow in any particular case the moisture content gradient is always steeper as the moisture content becomes less. The variable of temperature, especially as might concern subfreezing temperatures was of no concern to Lewis. For this reason only the primary conclusions reached by him are summarized herein.

- (1) With differences in moisture content between the field capacity and the wilting point, water can be raised by only a few inches by capillary action from a moist subsoil (eg. water in quantities sufficient to support plant life).
- (2) Most of the water movement takes place at moisture contents well above the wilting point.
- (3) At high relative moisture contents water may move through considerable distances whereas at low relative moisture contents the distance which moisture will move is very small.
- (4) The variable factors as gravitational field, texture, initial moisture content and temperature are of such importance that capillary movement becomes almost an individual problem with each set of conditions (geographically).

II. Measurement of Capillary Conductivity

Richards and Moor. [L. A. Richards and D. C. Moore, Trans., Am. Geophys. Union, 33 (4), 531-40 (1952)] have devised a pressure-type apparatus for measuring capillary conductivity involving the use of a soil column mounted between two porous plates. Pressure control of

water is utilized rather than that of a vacuum [L. A. Richards, Physics, 1, 318-33 (1931)]. Accurate measurements of water input and outflow are possible. Pressure of water input is noted as H_1 and the hydraulic head of the outflow as H_0 . A reference datum for pressure, H , is taken at the center of the soil column. The net hydraulic head ΔH will then be represented by $\Delta H = H_2 - H_1$ where the subscripts 1 and 2 refer to head differentials from the soil-column center datum to the bottom of the column and the soil-column center to the top of the column. If H_1 and H_0 are made equal to zero, then H_2 and H_1 will also approach zero. The equivalent soil-moisture tension of water in the soil at the center of the column is then controlled by and becomes equal to the air pressure (H) maintained in the pressure chamber. The soil-moisture tension and the corresponding moisture content of the soil can be varied at will. With the application of an overall driving head equal to $H_1 - H_0$, water is moved through the porous plate at the top of the soil-column and out through the porous plate at the bottom where its volume is measured.

Results are reported by Richards and Moore on six soil types with the following particle size characteristics (Table I).

Table I. Particle Size Analysis of Soils Subjected to
Conductivity Measurements

Soil Type	Sand	Silt	Clay
Superstition sand	89.6	4.8	5.6
Coachella Loamy fine sand	85.1	10.5	4.4
Pachappa fine sandy loam	59.4	32.9	7.7
Millville silt loam	24.9	60.3	14.8
China silty clay loam	5.8	58.1	36.1
Clay (Preston, Idaho)	1.5	52.9	45.6

Tabulation of conductivity (k) in relationship to air pressure clearly shows a progressive decrease in the conductivity with increasing silt and clay content. The sandy soils have higher conductivity at low tension and lower conductivity at high tension (0.2 atm.) than medium textured soils.

A relation between soil-moisture tension and the moisture content of the soil is expressed as

$$P_W = aT^b$$

where P_W represents the moisture percentage on a dry weight basis, a and b are constants of the system, and T is the soil moisture tension.

Data on the capillary-conductivity of the different soils studied show that it may vary from 0.2 to 20 cc per hour. As the soil-moisture tension increases up to about 250 cm of water, the capillary conductivity drops rapidly.

Evidence for a hysteresis effect is reported in the data by Richards and Moore. The conductivity is much lower at corresponding tension values when the tension is decreasing than when the tension is increasing. This hysteresis in the relation of tension to both moisture content and capillary conductivity is a very real effect in soils and is oppositely directed in drying and wetting processes.

During the process of wetting of a dry soil, the transfer of water across the wetting front will progress at a finite rate. The hydraulic gradient at this front is high because of the rapid change of soil-moisture tension with distance, whereas within the wetted mass of soil the conductivity is relatively higher and the moisture tension and hydraulic gradient will be lower. With a given rate of water flow

near the wetting front, it is obvious that the average rate of change of the moisture content in the wetted zone will vary inversely with the volume of relatively mobile water in the wetted zone.

The wetting and moisture-retaining processes can be explained in terms of capillary conductivity. In the wetting zone, the conductivity is low, but the hydraulic gradient is high so that the flow velocity is definitely appreciable and prevents the maintenance of moisture at the field capacity if the wetting zone extends to the soil surface. However, as the wetting front proceeds downward through the soil, there comes a time when the moisture content at the wet side of the wetting zone is high enough to be on the rapidly rising part of a capillary conductivity-curve. Moisture contents higher than this value cannot persist in the overlying soil.

It is generally recognized that field capacity represents a moisture content in field soil that occurs and changes slowly with time because of the dynamic moisture-transmitting properties of the soil. Further, the field capacity does not in any sense represent an equilibrium condition. If the factors that tend to control the rate of transfer of water through the wetting zone tend to make the rate of transfer somewhat constant, then the thicker the zone of wet soil is, the more slowly does its moisture content change with time.

There is strong indication that field capacity corresponds to that tension and moisture content at which conductivity becomes negligible. This latter supposition is not yet completely verifiable.

In the work of Richards and Moore, the Darcy relation^{*} in the form $\tilde{v} = k \tilde{f}$ is assumed to be valid for the type of flow expected in the six soils studied. In this equation \tilde{v} is the velocity vector (volume per unit area per unit time), and \tilde{f} is a vector sum of the force due to gravity and the pressure gradient. The term k is sometimes referred to as the Darcy coefficient of permeability. If the term "permeability" is reserved for reference to a specific physical property related only to the pore structure and independent of the nature of the fluid, then k may be termed as the hydraulic conductivity. When the soil pores are not completely filled with water, k depends on the moisture content of the soil and may be called unsaturated permeability or capillary conductivity.

The temperature maintained throughout the measurements made by Richards and Moore was 20°C. Considerable merit would be found in continuing the study initiated by these workers in the region of freezing temperatures. The apparatus appears admirably suited to

*It may be mentioned that Hubbert [M. K. Hubbert, Jour. of Geology, 58, 655 -59 (1950)] has criticized the Darcy relationship in the above form. Hubbert describes the relationship in the form

$$\tilde{v} = -K \operatorname{grad} \tilde{h} \quad (24)$$

where the term $(-\operatorname{grad} h)$ is the hydraulic gradient and v the velocity vector (volume per unit time per unit area). A correctly modified form of the Darcy relationship is written by Hubbert as

$$\tilde{v} = \frac{R}{u} (\sqrt{\rho g} - \operatorname{grad} p)_1$$

where R is equal to Kd^2 (a constant), d is the mean grain diameter, ρ is the density of the fluid, u is the viscosity of the fluid, p is the pressure and g is the acceleration due to gravity. Hubbert points out that it is incorrect to assume that \tilde{v} is capable of being derived from a potential, especially if the fluid density is not constant.

maintaining temperature gradients of a magnitude to give data of interest to migration studies in the thermal regime. Initially frozen soils might for instance be studied using water at 0°C for permeability studies.

Benkelman and Olmstead [A. C. Benkelman and F. R. Olmstead, Proc. Eleventh Ann. Meeting, Highway Res. Board, 11, 152 (1931)] have applied the migration of moisture from a water table (or other source) at higher temperatures than the "near surface" soil to explain ice lens formation. The nature of the soil is without question an important factor in the degree of moisture migration from a water table upward. In relatively cohesionless silt-soils it is claimed that the moisture may move upward 30 feet by capillarity. There is disagreement, however, among authors on the height to which moisture migrates by capillarity. Too frequently investigators report migration heights without suitable identification of soil type or initial moisture content.

The effect of coarse particle layers on the migration of water by capillarity has been reported by Mamanina [L. Mamanina, Pochvovedenie, No. 2-3, 101-05 (1944). See also SIPRE File U 2041]. Laboratory experiments on a test tube scale of soil columns showed that two layers of loam separated by 2 cm. interlayer of coarse sand particles, 1 to 2.5 mm in diameter, completely cuts off the capillary rise of water. The results are less favorable when the interlayer is gravel (2 to 12 mm in diameter) or medium sand (0.5 to 1 mm in diameter). Interlayers of packed loam retard the capillary rise of water. Application of the above results are useful in halting capillary rise under surface structures.

III. Capillarity and Soil Deformation

Emery [D. J. Emery, Canadian Eng., 72, 5-9 (1937) March] has expressed the importance of water migration via capillary action in explaining the formation of ice lenses. The minute droplets within a soil subgrade resist freezing even at temperatures well below 0°C (values down to -70°C have been reported, however, there is some reason to doubt the excessively low figures). An increase in the droplet size through moisture migration via capillaries leads to a water body of such dimensions that freezing is possible at 0°C. Migration is again noted in the drawing of minute droplets from capillaries by the larger freezable drops. An ever increasing frozen body or lens is favored by this trapping process.

A step-by-step investigation of water freezing and ice-crystal growth has been carried out by Kolesnikov [A. G. Kolesnikov, Doklady Akademii Nauk SSSR, Novaia, 82 (6) 899-91 (1952)]. Soil waters in the free state freeze at temperatures near 0°C after which bound (capillary) waters begin to freeze. The amount of ice formed from bound water increases with decrease in soil temperature. The variations in the ratio of water to ice produces a change in the thermal properties of the frozen soil because the thermal characteristics of ice and water are dissimilar.

The total amount of water raised from a water table to a frozen zone depends upon several factors; the force of capillarity, the area of pore space, and frictional resistance. Capillary pressure varies inversely with the diameter of the pores. The frictional resistance to flow through a soil is a function of the surface area of the soil particles and as a result must increase with a decrease in

grain size at a much greater rate than does the capillary pressure. If sufficient water is to be raised from a ground table by capillary pressure, the pore size must be small enough to furnish appreciable pressure, yet large enough to prevent high frictional resistance to flow.

With sufficient information on the capillarity properties of a soil it should be possible to make reasonably accurate predictions as to the magnitude of ice lens formation since moisture migration will be predictable. Cohesionless soils with little capillarity should freeze with little expansion since the occluded water will be in drop sizes large enough to freeze independently with little migration and coalescence of drops. Little or no moisture will be expected to rise from a water table in such a soil system. It is reasonable to expect that a soil may have low cohesion but high capillarity. The frictional resistance due to small pore size will not favor moisture migration. Unless the water table is abnormally close to the surface, little water should rise.

Krynine [D. P. Krynine, Highway Res. Board, Proc., 26, 474-83 (1946)] has reported in some detail an investigation of capillary migration in gravels and silts. The equation developed by him contains essentially two variables; time (t) and capillary height (h) and the constant, c .

$$h^n = ct \quad (25)$$

The value of n may or may not be constant. The shape of the curves reported are of parabolic order, or $n = 2$. In soil depths (tube lengths in actual experimental measurements) of 4 ft., it was found not uncommon to have uniform, but less than saturation, moisture

content throughout. The moisture content was usually lower at the top. Krynnine has concluded that under conditions of very tight packing as well as very loose packing capillary rise is retarded.

Osterberg [J. O. Osterberg, Civil Engineering, 10(2), 100-2 (1940)] in a survey of the frost heaving problem has pointed out the general agreement among several investigators with regard to the factor of soil particle size and its relation to capillary migration. The latter phenomenon, as already discussed, is an important consideration on the extent of soil deformation. On the basis of tests on pure fractions of soil, a particle diameter of 0.1 mm is maximum for any appreciable deformation (ice lens formation). [G. Beskow, Sveriges Geologiska Undersokning, Stockholm, 1935, Series, Cv, No. 375 (242 pp)]. Tests on actual soils (rather than fractions) give a critical size of 0.02 mm. [A. Casagrande and S. Taber, Proc. Eleventh Annual Meeting of Highway Res. Board, 11, 168 (1931)]. With ground quartz only the faintest indication of lens formation is observed at a particle size of 0.07 mm. At a size of 0.01 migration takes place readily. [S. Taber, J. Geology, 38, 303-17 (1930)].

From a practical standpoint a lack of uniformity in particle size is to be expected in natural soils. For well graded soils only 3 per cent of grains finer than 0.02 mm is required to produce deformation. For very uniform soils at least 10 per cent of grains finer than 0.02 mm is required.

The capillary movement in moist soils (as previously described) may be simply measured by determining the height to which water will rise above a free water surface in a "soil tube". However,

for soils with fine particles (large capillarities) the determination is more difficult. The limit of capillarity between soils that do not exhibit deformation (by lens formation) and soils that do is given by the simple relationship $C_L = 1$ meter and $C_D = 1 \frac{1}{4}$ meters where C_L is the capillarity for loose packing and C_D the capillarity for the densest possible packing. The limit for soils which normally do not heave is $C_L = 1.5$ meters and $C_D = 2.0$ meters.

It has been found empirically that the rate of lens formation varies inversely as the square of the pressure and inversely as the cube of the particle diameter. Pressure, here, refers to the total pressure at the frost line and is dependent on (among other factors) the capillary pressure. For clays the capillary pressure is very large and changes in surface pressure affect lens formation only slightly since the total pressure is but little changed. For coarser soils the surface pressure may reduce lens formation materially. As a general rule, then, as grain or particle size decreases, lens formation increases, however, at a certain critical size, the capillary pressure increases greatly with decreasing grain size and the combined effect is to decrease lens formation and subsequent soil deformation. In saturated clays shrinkage (as evidenced by cracks) may occur at subfreezing temperatures due to the withdrawal of water to form ice layers. S. Tabor, [loc. cit.]

Benkelman and Olmstead [loc. cit.] have shown by carefully controlled laboratory observations that ice formation, as lenses, (leading to soil deformation) can only be accomplished in soils which are saturated either by capillarity or by seepage. As the air

temperature drops below freezing the frost line is shown to move downward into the soil. The interstitial water (either from capillary rise or seepage) is frozen. The expansion of the soil mass due to this ice is known to be only slight. With an increase in the freezing temperature the frost line moves upward. The ultimate effect of this migration is to leave a cumulative water void following the contact line of the frozen and unfrozen soil. Water will tend to fill this void, the amount of which is dependent on several factors: (1) interfacial tension between water and ice; (2) hydrostatic head; (3) atmospheric pressure; (4) other pressure effects existing below the frozen soil (e.g. capillarity properties of the soil). More than one of the voids may result in alternating freezing and thawing conditions. The expansion of the ice formed in the now water filled voids accounts for the frost heave. The importance of understanding moisture migration at low temperature cannot be underestimated. Capillary migration is of great importance since water brought into a soil by such a process cannot be drained.

Bykov and Kapterev [N. I. Bykov and P. N. Kapterev, Izvestiia Akademii Nauk SSSR, Seria geograficheskaiia i geofizicheskaiia, bull. no. 1, 93-99 (1938)] have determined the moisture content of some 200 soil samples before and after freezing in order to study the extent of water migration due to freezing and the subsequent formation of ice lenses. Their data indicate that moisture in sandy soils moves away from a freezing layer. The moisture in loamy soils with a moisture content of 40 per cent, accumulates near the freezing surface and carries with it minute particles of soil. No definite regularity

is observed in the distribution of moisture in a vertical direction under normal conditions. The force developed during the growth of ice crystals in an open soil is quite negligible and cannot account for the soil deformations in the course of freezing.

It has been claimed that the volume change in a soil (during freezing) due to lens formation is independent of the air temperature [A. Ducker, Die Strasse, 4, 138-40 (1937)]. Experiments were conducted on sandy soils at a moisture content of 20-23 per cent placed in 3.9 cm. diameter glass tubes so insulated as to permit cold penetration only from the top. Lens formation and deformation rates were measured with decreasing temperature. Results indicated that deformation increased continuously to a temperature of -16°C at which temperature a maximum deformation rate of 0.150 cm. per hour was noted. Below -16°C the deformation rate decreased. It must be concluded that air temperatures (below freezing) do influence the rate of soil deformation, at least in the particular type of soil used by Ducker.

No single theory of ice formation is adequate in scope to cover all types of soils and all methods of moisture migration. Watkins [Benkelman and Olmstead, ibid., see page 166] suggests that cool weather serves as a "primer" in favoring strong capillary action to achieve saturation of the upper subgrade soils. A supersaturated layer may form just below the frozen layer connected by capillary tubes to a water table. With rapid freezing of the upper supersaturated layer, voids might develop which possibly might again be filled with moisture through migration from the water table or even by seepage. Freezing of the water in this space (lens formation) might well account for the soil expansion.

It is notable that Benkelman and Olmstead suggest alternate freezing and thawing whereas Watkins and others feel the cyclic action is unnecessary. Regardless of the theory of lens formation, it is obvious that there is insufficient data on the processes of water migration under conditions of low temperature.

Beskow [G. Beskow, Sveriges Geol. Undersokn. Arsbok., 26. Ser. C. Avhandl. och Uppsat. No. 375, 242 pp (1935). See also SIPRE File U 667] has observed that the absorption film surrounding large particles is relatively thin. Pressure decreases the thickness of absorption films, and the rate of movement of water to the freezing areas is reduced. The distribution of the ice in a frozen soil is influenced by particle size as well as pressure and the nature of the water supply. Water must be present in the soil in excess of saturation, either at the beginning of freezing or by subsequent capillary transfer from the ground water supply. If ice lenses of such size as to cause appreciable soil deformation are to form, additional water must be supplied to the frost zone. From laboratory studies Beskow concludes that the rate of heaving (which in turn must be related to the rate of moisture migration to permit ice lens formation) is inversely proportional to the ground water dept. in highly capillary soils.

The hydrodynamic subpressure controlling the cohesion of water films surrounding the soil grains has been calculated from the rate of heaving in freezing tests as well as by direct manometric measurements [R. F. Ruckli, Proc. 2nd Internat'l. Conf. on Soil Mechanics, 2, 282-87 (1948)].

The formation of an ice lens and thence soil deformation has been shown possible in the absence of capillary pressure in soil sample overflowed with water. The cause of soil deformation is then attributed to the cohesion of water films surrounding the soil grains.

If there is no extended ground-level, the water needed to supply the ice-lenses is drawn from free water, accumulated in pores of the soil beside and under the lenses. The resulting plane (two-dimensional) and spatial (three-dimensional) ground water flow towards an ice lens was calculated. Ruckli agrees essentially with Beskow

[G. Beskow, Sveriges Geologiska Undersokning, Stockholm, 242pp, Series Cv, No. 375 (1935)] in the conclusion that the rate of flow and therefore the frost heaving in the center of the lens is inversely proportional to the diameter of the lens. For this reason small lenses may cause greater soil deformation than a single large lens.

Taber [S. Taber, J. of Geology, 37, 428-61 (1929); 38, 303-17 (1930)] on the basis of extensive laboratory tests, is in agreement with other investigators whose work has been reported herein, that the pressure effects accompanying the freezing of soils are due to the growth of ice lenses and crystals and not to any general soil expansion. The pressure and therefore deformation is developed in the direction of crystal growth (rather than direction of least resistance) which is in turn determined chiefly by the direction of cooling (or heat conduction).

Taber further concludes that soils, when subjected to freezing under normal conditions, usually behave as open systems. Thus, the pressure developed during heaving in an open system is limited by the tensile stress that can be developed in the water. Deformation

in a clay type soil is greater than in sandy soils even though only a part of the water in the former soil freezes. The boiling point of a liquid permeating the soil seems to be of greater importance than the freezing point in determining the pressure developed by freezing in an open system.

The actual pressure exerted by the expansion of gravitational and capillary water has been estimated by Tsytovich [N. A. Tsytovich, Akademia Nauk SSSR, Instituta Merzlotovedenia im. V. A. Obrucheva, p. 7-15 (1946)] as 2,115* kg. per sq. cm. on foundations of engineering structures.

IV. Vapor Migration

From condensation observations Krynine [loc. cit.] concludes that water vapor migration not only precedes the moving capillary moisture but also accompanies it. Moisture films appear to be formed on soil particles above the visible boundary of the liquid phase. No definite conclusions could be reached as to which (or both) of two factors cause condensation of the migrating moisture. Certainly a decrease in temperatures favors condensation; however, a decrease in relative humidity of the surrounding atmosphere is contributory when the temperature does not change. There is obviously a need for further experimental data to permit the inclusion of temperature as a third variable in the capillary equation $h^n = ct$.

Vapor migration and its effect on soil cohesion have been discussed by Winterkorn [H. F. Winterkorn, Highway Res. Board, 26, 421 (1948)]. In the waterproofing of soils it is necessary to maintain part of the cohesion-giving water-affinity of the soil particles and to

* Note that Wainberg gives 31 kg/cm² as a mean value for the crushing strength of ice.

make part of the internal soil surface water repellent. A thin undisturbed water film between individual soil particles contributes to the cohesive attraction. Water proofing agents are added in very small percentages to provide non-wettable barriers throughout the soil system. Although liquid water cannot readily penetrate the barriers, vapor can pass over them, enter the water proof area and add "water layers" to the previously thin surface coatings. The system then swells and cohesion is lost.

Lewis [M. R. Lewis, U. S. Dept. Ag. Tech. Bull., 579, Oct. 1937] has reported data on vapor migration at temperatures above freezing. Soil in a moistened condition was placed in tubes within which were screened spaces 1 mm. in width. The amount of moisture migrating across the screened gap was no more than 1-12 mg. per hour. Lewis concluded that if no (or very little) moisture vapor migrated across this gap, very little vapor should pass through pores of a moist soil. These conclusions of Lewis are not in complete accord with those of Krynnine [loc. cit.]

Additional information on vapor migration is included under the section heading 4, Thermal Gradients and Thermal Translocation of Moisture.

V. Compression Forces in Capillary Migration

Winterkorn [H. F. Winterkorn, Soil Science, 56 (1) 169 (1942)] has used the concept of compression forces on polymolecular absorption layers to describe the condition of water in soil pores. The compression forces decrease logarithmically with increasing distance from the solid absorbent [ibid. 41, 25-32 (1936)]. In a soil capillary

it would be expected that water should be held most strongly and compressed most highly at the pore wall. At the center of the pore the water might be more or less free (corresponding to gravitational water) to move under the force of gravity [G. J. Bouyoucos, Soil Science, 11, 33-48 (1921)]. At low temperatures water in a soil-system may vary from that which is already solid above 0°C due to pressure effects (corresponding to ice VI and V) through water with melting points decreasing to -22°C. There might also be water with melting point increasing to 0° or slightly above, (as can be understood from the phase diagram of water). Ice I may be obtained above 0° if the water is under tension. The following equation describes the condition

$$\frac{dt}{dp} = T(V - v)/S \quad (26)$$

where T = absolute melting temperature; dt = change in melting temperature; dp = change in pressure; V = specific volume of liquid water; v = specific volume of ice; and S = latent heat of fusion.

From the above equation (26) it follows that a decrease in pressure on normal water increases the melting point by $0.0075^{\circ}\text{C}/\text{kg/cm}^2$. Immediately below a capillary meniscus, water is under a pressure decrease of $P = 2 T/r$, where T is the surface tension, r is the radius of curvature of the meniscus, and P is the pressure decrease (or suction). For a droplet of colloidal dimensions ($r = 10^{-5} \text{ cm.}$) the pressure effect will be 15 kg/cm^2 thereby increasing the freezing point of water directly under the meniscus by 0.113°C . This point will thereby act as a normal center for ice formation. Between this point and the solidly absorbed water on the pore wall is a zone of water (liquid) possessing melting points down to -22°C . This zone serves as a passageway for the conduction of water to the freezing or crystallization

center [see also S. Taber, Jour. Geol. 37, 428-61 (1929); ibid., 38, 303 (1930); Public Roads, 11, 113-32; and Beskow, G., Sveriges Geol. Undersokn. Ser. C, No. 375].

The direction and the amount of volume change occurring upon cooling below the normal freezing point of water depends upon not only the temperature reached and the normal moisture content, but also on the distribution of this moisture in the different ranges of absorption pressures. The phenomenon of certain soils contracting upon cooling (rather than expanding which would be the expected behavior) becomes reasonable as well as predictable.

Winterkorn [H. F. Winterkorn, Trans., Am. Geophysical Union, 25, 405-11 (1944)] has further stated that the entrance of water into or for that matter through a soil is due to the affinity of water for the internal surface of the system. Without this affinity no entrance of water will occur, irrespective of the surface tension of the liquid. The importance of the affinity of the liquid for the surface of the solid (as in a soil system) must be considered in the understanding of any migration phenomenon. A second factor in water migration is the cohesive force binding the soil. The above two factors (their magnitude) determine the general reaction, however, the speed depends largely upon the permeability of the soil-system and upon the ease with which free and absorbed gases may escape from a pore-space. If vapor migration and thence escape is prevented in a soil, a failure may result (as in a road surface) not unlike an explosion. If the concept of driving force, being the affinity of the internal surface for water is accepted, it is easy to explain why moist cohesive systems may be

exposed to free water without observable change. In such a system the affinity for water is satisfied and no driving force for the water-attack is left. Without water attack general deterioration of a soil-system (surface flaws, erosion, slides, etc.) will not occur.

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SECTION 3.

Moisture Migration Through Snow

Section 3. Moisture Migration Through Snow

The rate of absorption of melt-water from snow cover by the underlying soil is limited by the lesser of two quantities: (1) the melting rate, (2) the infiltration capacity of the soil [L. D. Horton, Proc. Central Snow Conference, 1, 5-21 (1941)].

A snow cover is like a layer of soil. When in a given condition of structure and compactness it can absorb rain and melt-water from the surface and hold it as a sponge holds water, up to its capillary-capacity. This is generally less than its saturation capacity. New light snow with density 0.25 to 0.50 may hold in capillary suspension a volume of water of the same order of magnitude as the volume of snow.

When rain or melt-water enters a snow layer the moisture content of the snow is built up to capillary-capacity progressively from the surface downward. At any moment,

$$dp' = \frac{W}{\Theta d} \quad (27)$$

where dp' = the depth of penetration in inches,

W = water added, inches depth on surface,

and Θd = capillary deficiency as a fraction of the snow volume.

For dry snow the capillary-deficiency and capillary-capacity are identical or nearly identical and both relatively small; for wet soggy snow the capillary-deficiency may be much less than the capillary-capacity. Water up to capillary-capacity is held in the snow against the force of gravity and hence moves downward only slowly, by diffusion. Therefore, melt-water or rain is not added to the base of a snow layer at the ground surface until the full depth

of snow is saturated to capillary-capacity, and no runoff occurs from snow cover until the rain or melt-water has penetrated to the ground surface.

Ahlmann [H. W. Ahlmann, Union Geodesique at Geophysique Int. D'Hydrologie Scientifique, Commission des Neiges, 2 (1) Rapport 2, 4 pp. (1939)] investigated the capacity of snow to absorb water and store it. Shafts up to seven meters deep were dug into snow and firn layers to determine the depth and water content of the total winter accumulation. It was observed that rain generally did not cause ablation, but permeated the snow where it was absorbed in various quantities by different layers. Compaction of the snow mass occurs. Part of the water is retained at the surface of ice layers where freezing takes place. The recrystallization of firn and gradual change into glacier ice is the result of freezing, melting, and refreezing processes to which permeating water is exposed in a firn mass at temperatures below 0°C. Firn on glaciers with extensive accumulation possesses a large capacity for water absorption. The water may be converted into ice layers, retained in water-saturated layers, or it may drain off.

Seepage of water into snow cover has been reported by Oechslin [M. Oechslin, ibid., Quest. 1, Rapport 1, 4 pp. (1939)]. A known amount of colored water at 4°C was poured over snow surfaces ranging from ten sq. cm. to 100 cm. sq. in size while timing the rate of penetration. It was noted that the time of seepage generally increases with increasing specific weight of the snow. In cases of compact and old snow, uniform penetration rate may decrease because of the presence of open ducts. Seepage is not equally distributed in the snow profile

due to obstructions at layer limits. Retentivity was measured by means of a transparent snow tube 10 cm x 10 cm x 50 cm, containing a known quantity of snow. One dm.³ of colored water at 4°C is poured over the snow and the quantity of water caught in a container below the tube is measured over specified time limits. It was concluded that snow loses its ability to retain water when saturated and its retentivity increases with decreasing temperature.

The same general experimental approach has been used by Clyde [G. D. Clyde, Monthly Weather Review, 57, 328 (1929)]. Fluorescein dye was used to trace the course of rain water through snow. During a medium downpour water reached the ground first through small continuous channels and later through a general downward movement of water through the snow. The general downward movement was uniform; however, on clear warm days well defined channels were indicated by the dye. Color penetration during cold weather was only three inches in 12 hours. Migration in a sloping direction through snow cover located on downhill slopes was noted to be very slight. Clyde also makes the statement that rain water plays only a minor part in the ultimate melting of snow.

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SECTION 4

Thermal Gradients and Thermal Translocation of Moisture

tion 4. Thermal Gradients and Thermal Translocation of Moisture

One of the earliest comprehensive discussions of the effect of temperature on the movement of moisture is by Bouyoucos [G. J. Bouyoucos, J. Agric. Research 4 (V) 145 (1915)]. The variation of viscosity with temperature is noted as being more pronounced than the variation of surface tension over an identical range. Thus for the same range in temperature, 0 to 50°, the viscosity changes from 100 to 30.17 while the surface tension changes from 100 to 85.

The capillary action depends largely upon surface tension while the facility of moisture movement is a function of viscosity. Water translocation should, from a theoretical standpoint, be toward the cooler area from the warmer area. In hot climates, however, the migration may be in the reverse direction according to studies on subpavement soil structures.

Bouyoucos determined the moisture migration in a variety of soils with a large variation in initial moisture contents over a temperature gradient from 0 to 20°C as well as 0 to 40°C. He concluded that the maximum thermal water movement depends upon a definite condition of moisture for any particular soil. Any deviation from this definite percentage of moisture in either direction causes a decrease in thermal movement of water. The above described percentage appears to be a constant for a given soil, and the term "thermal critical moisture content" is given to this percentage. A thermal critical moisture in soil is defined by Bouyoucos as that percentage of moisture in soils which allows the greatest amount of water to move from a warm to cold soil at any amplitude of temperature.

Light sandy soils were observed as being extremely sensitive to the maximum thermal movement with slight variation in moisture content.

A silt loam soil (as Clyde silt loam) with 30-35 per cent moisture exhibits very low water movement. It appears that the inherent capacity of such a soil to hold water by attractive forces is high and free movement of water (as by gravitational force) may take place only until this capacity is satisfied.

Bouyoucos has made four assumptions to explain the thermal translocation of water which cannot be accounted for entirely or in part by viscosity and surface tension variations with changes in temperature.

1. The soil possesses an attractive power for water and holds it with great adhesive force, as noted from experimental observations.
2. The adhesive forces decrease with increase in temperature, a reasonable assumption if based on nothing more than kinetic principles.
3. The surface tension and cohesive powers of the liquid diminish with rise in temperature.
4. The forces arising from capillary attraction are decreased with increase in water content of the soil.

The validity of the third postulate has been questioned, however, Bouyoucos feels strongly that experimental evidence is overwhelming in its favor.

Using the four postulates as a basis, the thermal water movement in soils of various moisture contents may be explained as follows: (a) For soils of low moisture content, the water is held by the soil with great tenacity (the order to magnitude may be from 6000

to 25,000 atmospheres). With a temperature gradient impressed over a column of soil the equilibrium condition is disrupted. In a warm area the attractive and adhesive forces of the soil for water decrease and in the cold area the forces increase. The cold area thereby exerts a "pull" effect and draws whatever water is susceptible to movement into the cold area. Only a small quantity of water would be expected to move by this mechanism, however.

As the moisture contents of a soil increase, two factors (adhesive and attractive forces) tend to balance out. There is an increase in the ease with which water can be removed (a decrease in adhesive forces) from the warm area but a decrease in the capacity of the soil in the cold area to attract water. The ease of moisture loss in the warm area appears to be more influential, resulting in an increase of moisture translocation with increase in moisture percentage.

At the previously mentioned thermal critical moisture content the soil in the warm area yields water with such ease that a maximum thermal water translocation is observed. This maximum is reached in spite of the decrease in attractive force from the already moist cold area. Above the critical moisture value the attractive force of the soil in the cold area is too small to further moisture migration, and a decrease in migration tendency begins. Moisture migration may and does occur above this saturation point, however, the migration is not a function of the soil but rather the surface tension of the soil water.

The thermal movement of moisture according to Bouyoucos in soils is always upward and never downward which may under certain temperature conditions appear to be an anomaly. Even though the

temperature below is usually less than that above, the water-attractive forces of the soil below are more satisfied than those above. The temperature-differential, resulting in differences in attraction, adhesion, and surface tension, is not as important as the upward pull. During cold weather (or at night when a temperature drop is expected) all forces act in parallel and the moisture migration is upward.

Additional light might be shed on moisture migration if the thickness of the moisture shell around soil particles at the point of maximum thermal translocation were known. Due to the irregularities in shape of soil particles (lack of spherical shape) and air spaces the determination of such a film thickness would be extremely difficult and would vary tremendously from soil to soil.

Bouyoucos has reported further on the phase of moisture migration from a moist and warm area (or column) to a dry and cold area of soil assuming an air space between the two areas or columns. This vaporization and condensation of soil water from a point of high temperature to a point of low temperature is of particular interest to the primary subject under consideration. In his experimental approach, Bouyoucos used columns of moist soil (1 1/2 in. diameter) separated from dry soil by a 1/4 in. space. The moist soil was maintained at 20 and 40°C and the dry soil at 0°C. The duration of each separate measurement was 8 hours. Five types of soil were used each at three different moisture contents: Quartz sand at 3.90, 6.83, and 13.52 per cent; sandy loam at 7.23, 10.27 and 15.82 per cent; Miami silt loam at 9.16, 14.52, and 16.40 per cent; Clyde

silt loam at 9.85, 15.51 and 16.40 per cent; and Miami clay at 10.77, 15.36, and 20.75 per cent. The percentage of moisture vaporization migrating (and being absorbed) on the dry column was so small in every case as to be insignificant; a very surprising result in view of the observations reported by others (See also Section 2, IV, Vapor Migration). Even soils with over 20 per cent moisture showed only about 0.3 per cent migration and the increase from a 20 to a 40° gradient brought about only a few tenths of a percent increase in moisture migration.

Buckingham [E. Buckingham, U. S. Dept. Agric. Bur. Soils, Bull. 38, 61 p.] has noted very slight vapor migration in studying the loss of soil moisture by direct evaporation from points below the surface. That is, the actual mean rate of loss of water through diffusion of water vapor through soils in still air is very small.

The results of both Bouyoucos and Buckingham would belie the statement that ground dew is formed from vapor rising from warmer soil into a cold atmosphere.

The movement of moisture from a moist and warm area (column) to a dry and cold area of soil as well as from a moist and cold area to a dry and warm area is reported by Bouyoucos. From a practical agricultural standpoint the movement of moisture from moist ground to a surface mulch is involved. The same soils as previously described above by Bouyoucos were used in this latter study. Three different moisture contents were used and the temperature gradients 0 to 20° and 0 to 40°C were impressed as before about 8 hours. Thus, for each soil, at each of three moisture contents data were

obtained for water moved from a moist column at 20°C and 40°C to a dry column at 0°C; and water moved from a moist column at 0° to a dry column at 20° and 40°.

The results showed that the amount of water moving from the moist and warm column to the dry and cold column is: (1) nearly twice as great at 40° as at 20°C; also (2) somewhat greater in soils with higher than with lower colloidal content; (3) increases with a rise in moisture content. All of these observations are nothing more than would be expected from elementary kinetic behavior.

When dry soil is kept at either of the higher temperatures (20° or 40°) and the moist soil at 0°C, the dry soil was found to take up very little water from the moist soil. The quantity of water, small as it is, decreases with a wider temperature gradient. The percentage of water absorbed by the dry soil at 20°C in every case is greater than that absorbed at 40°C. As before the percentage taken up increases with the colloidal content in the soil. The insulating effect of a dry warm mulch plays an important part in the percentage of moisture migration from lower and cooler areas beneath the surface.

Edelfsen and Bodman [N. E. Edelfsen and G. B. Bodman, J. Amer. Soc. Agron., 33 (8) 713-31 (1941)] have reported on moisture migration based upon laboratory test only at temperatures above 0°C and, therefore, of somewhat limited use in the present study. Their results are important enough to warrant a brief discussion. During the winter in the top ten feet of soil the temperature gradient is such as to cause water in the vapor and liquid

phases to move upward. The reverse is found to be true in the summer, that is a downward migration of liquid and vapor water. The actual magnitude of this moisture movement in summer and winter is slight but may be great enough to explain some field conditions.

The very large subject of ice lenses and frost heaving is fundamentally a moisture migration problem. (See Section 2, III). Anderson and others have pointed out that when a soil is frozen from the surface down to a definite depth and a warmer moist layer exists below, the temperature-gradients are such as to move the water (liquid and vapor) up to the frozen layer, favoring the formation of lenses. There is not complete agreement, however, among authors in reports of this direction of migration. It is also reported that water should move in the direction of the lower temperature and should thus move away from the ice-strata. The development of large ice lenses or extensive ice stratification should not be favored. Field examination should favor the former explanation (movement to the frozen layer) of moisture migration.

Among the significant contributions from the U. S. S. R. to the theory of moisture migration under thermal gradients is that of Bozhenova [A. P. Bozhenova, Akademiia Nauk SSSR, Reraraty nauchnoissledovatel'skikh rabot za 1945 g. Moscow, 1947. Otdelenie geologo-geograficheskikh nauk, p. 156. See also SIRE File U. 1643] who reports on the extent of water migration in relation to soil saturation during freezing. The results indicate that water migration decreases as the moisture content is decreased. The presence of air spaces appears to be the main factor determining the velocity of

water migration when the moisture content is diminished, the porosity is constant, and the soil is in the three-phase system, soil-water-air. The velocity of water migration can be determined by the amount of free water in the soil when porosity and saturation are simultaneously decreased and the soil is in a two-phase system, water-soil.

W. O. Smith [W. O. Smith, Trans., Am. Geophys. Union, 24, (2) 523 (1943)] has reported some data on moisture transfer at temperatures above freezing. Previous work of Bouyoucos [G. J. Bouyoucos, Michigan Agric. Exp. Sta. Tech. Bull., 22, 1-63 (1915)] appears to be subject to criticism. Smith has observed a much greater magnitude of moisture transfer for thermal gradients which are even smaller than those impressed by Bouyoucos. Smith reports that moisture is transferred from a warm and moist soil to a cold and dry one only by the slow process of vapor-diffusion and possibly by convection. An exception is in the region of actual contact. When a temperature-gradient is applied, a large transfer occurs for a limited range of moisture-conditions and is confined entirely to the region in which the temperature gradient is applied. Smith explains the movement by capillary action induced by vapor-condensation. Thermal transfer of moisture, even under small gradients, begins at a water-content roughly just below that required for equilibrium with relative humidity of 99 per cent and ceases approximately at moisture-equivalent. Thermal transfer occurs chiefly with funicular water according to Smith. It reaches a maximum at about half the moisture equivalent water-content, a soil condition described by Bouyoucos as the thermal critical moisture-content. Thermal moisture

transfer is greatly influenced by the physical condition of the soil, for example, it is small in clays in a natural state, but large when the clays are fragmented. The need for more investigation and data is admitted by all authors, before accurate and comprehensive explanations can be made under varied conditions for thermal transfer of moisture.

A relationship between the ground moisture migration and the heat conduction has been given by Semiakin [F. M. Semiakin and P. F. Mikhalev , Compt. Rend. Acad. Sci. USSR, 17, 405 (1937). See also SIPRE File U 1930] . The above variables have been in turn related to the spacing between ice streaks formed during the process of soil freezing. An equation similar to the experimental relationship for the formation of Liesagang's rings exists and has been applied satisfactorily to photographs of frozen cylinders of soils.

From all the experimental observations it is more than ever obvious that the kinetic effect plays an important part in the absorptive power of soils for water. The possibility should not be overlooked, however, of greatly exalted migrations with cold areas being maintained at temperatures in the region of -10 to -30°C rather than 0°C. Under the condition of subfreezing temperatures the phenomenon of water trapping by freezing might well so disrupt the equilibrium conditions that moisture migration (both liquid and vapor) would be greatly facilitated.

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Section 5

Frost Penetration as Influenced by Soil Type, Soil Moisture, and Soil Cover

Section 5. Frost Penetration as Influenced by Soil Type, Soil Moisture
and Soil Cover

I. Measurement of Frost Penetration.

Bay and his co-workers [C. E. Bay, G. W. Wunnecke and O. E. Hays, Trans., Am. Geophys. Union, 33 (4) 541-46 (1952)] have studied several of the factors affecting frost penetration using the electrical resistance method. In this method blocks (as of plaster of Paris or fabric material) are placed in contact with the soil at desired depths. Because of the pore space and absorptive properties of the material of construction, an equilibrium between moisture in the soil and the block is quickly established. The electrical resistance (as measured on a Wheatstone bridge) varies accordingly. High moisture contents give low readings, and with decreasing moisture the resistance increases. Freezing of moisture is indicated by a sharp rise in resistance. Bouyoucos and Mick [G. J. Bouyoucos and A. H. Mick, Soil Sci., 66, 217-32 (1948)] state in this respect that the freezing of soil water produces a curve (temperature vs. resistance) similar to a drying curve. As long as unfrozen water remains in the unit, a fairly high conductivity is maintained. When no further liquid remains the resistance rises sharply. Fabric blocks have been used successfully as resistance units since freezing does not destroy the absorbent nor have any detrimental effects on the subsequent functioning of the fabric absorbent.

The conclusions of Bay regarding frost penetration into soil using resistance units are in general agreement with those using

the King soil-tube and hammer method made some 12 years earlier [H. B. Atkinson and C. E. Bay, Trans., Am. Geophys. Union, 21, 935-47 (1940)].

Typical resistance observations (using fabric blocks) in terms of moisture and temperature are those of Colman and Hendrix [E. A. Colman, and T. M. Hendrix, Soil Sci., 67, 425-38 (1949)]. In a completely saturated soil the resistance rises from 120 to 150 ohms as the temperature drops from 60 to 28°F. Soil moisture freezes at 28°F and the resistance rises sharply to 250,000 ohms. A drop in temperature to 10°F brings the resistance up to 1.3 megohms. This 18° temperature differential with over a million ohms difference in resistance indicates the larger increase in resistance per degree change in temperature for ice than for liquid water. A soil at field capacity moisture freezes at 24.5°F. The resistance on freezing was found to rise from 1250 to 50,000 ohms. A saturated soil shows an even greater rise in resistance though the initial resistance is also quite high. The difference is quite sufficient to distinguish between the frozen and liquid states of water in this type of soil. A lower moisture (as for instance at the wilting point) content soil at freezing temperature, but with the water as yet unfrozen, was observed to have an initial resistance of about 200,000 ohms. Upon freezing the resistance rose to 600,000 ohms. The differential is thus not as great as for higher moisture-content soils.

The conclusion drawn from the above field and laboratory resistance observations is that as long as a soil contains more than wilting-point moisture the resistance method is an accurate guide for

detecting freezing and thawing in soils. Below wilting-point moisture the resistance differential is not enough to give conclusive results. The advantages of the resistance method over that of the King hammer and tube method are obvious.

II. Factors Affecting Frost Penetration.

Using gypsum resistance units Bay [loc. cit.] concluded that dense vegetative and snow cover insulate the soil and reduce the rate of soil freezing the thawing. Certain crops (as winter wheat) give very little protection to the soil. When very little snow has fallen, frost may penetrate to a depth of 36 in. with an average daily minimum temperature of about -13°F. Eighteen inches of snow provide insulation to an Almena soil preventing frost penetration further than 12 in. when the weekly average minimum temperatures were as low as -13°F. The same snow cover would not prevent frost penetration to this depth with an average weekly minimum temperature of -21°F. However, 24 in. of snow would prevent frost penetration below 12 in. at this lower average temperature (-21°F).

Surface slope has an important bearing on the depth of penetration as would be expected. Frost penetration is greater on a northwest slope than on a southern or eastern slope. Field data on such slopes with similar vegetative cover, however, show only a 4 to 6 in. differential in the northwest and eastern slopes. As air temperatures rise above thawing it has been observed that the frost line rises due to heat from the earth as well as heat from the surface. If sufficient insulation is provided on the soil surface, frost penetration may be prevented or frost may be drawn from the soil by the warmth of the earth.

An important observation of Bay [loc. cit.] is that concerning the lack of moisture penetration on frosted soils. The ultimate and very detrimental effect of this impervious layer below the surface is that of soil erosion and water run off.

Fuller [H. U. Fuller, J. New England Water Works Ass'n., 54, (3), 275-816 (1940)] has correlated data on frost penetration and mean temperatures through winter months for clays and gravel soils. The air temperature acts more rapidly in gravel than in clay. However, the ultimate penetration was found to be slightly greater in clay than gravel. A minimum temperature of 24°F was observed in gravel and 29°F in clay. The maximum depth of penetration was observed to be 48 in. in clay and 45 in. in gravel. Under the influence of increasing daily temperature the frost went out of the gravel in one-fourth the time required for the clay (one week as opposed to four). The surfaces of the test areas were kept free of snow and other insulating cover.

From the above discussion it is obvious that several factors affect the depth of frost penetration, assuming that soil freezes without increase in its water content during the freezing process.

Among the most important factors Shannon [W. L. Shannon, J. New England Water Works Ass'n., 59, 356-63 (1945)] has suggested:

- (a) The magnitude and duration of air temperatures below freezing.
- (b) The thermal properties of soil, including the surface.
- (c) The temperature conditions within the soil at the start of and during freezing.

III. Freezing Index Related to Frost Penetration.

Casagrande [A. Casagrande, Proc. Highway Res. Board, 2, 168-72 (1932)] has suggested a method of correlating the above noted factor (a) with frost penetration by assigning a numerical index (Freezing Index) to the magnitude and duration of air temperatures below freezing. This method consists of plotting the cumulative departures (differences) of temperatures from 32°F. These differences are summed consecutively commencing with the first day in the fall or early winter on which the temperature drops below freezing. When the average temperature is above freezing (32°F) the numerical difference is given a negative value. The difference between the highest negative value and the largest positive value during the winter months is termed the freezing index. The greater the magnitude of the freezing index, the greater will be the frost penetration, other factors of soil properties and insulation being the same.

An approximate value for the freezing index (FI) may be computed from monthly average temperatures from the following expression:

$$FI = (32y - x) \cdot 30.2 \quad (28)$$

in which y is the number of months during which the average temperature is less than 32°F; x is the sum of the average temperatures for these months; and 30.2 is the average number of days for the months of severest winter (December, January, February and March). The value computed will always be found to be less than the true index but will approach it with increasing magnitude of the actual index. The

formula is of little value when the actual freezing index is less than about 300.

The thermal properties of the soil influence greatly the depth to which frost penetrates. The thermal properties of frozen soil are not the same as those of unfrozen soil. The color of the surface soil (or vegetative cover) greatly influences the depth of frost penetration due to variations in the amount of heat reflected or absorbed. The temperature of the soil mass at the state of freezing will influence the depth of frost penetration. Soil temperatures near freezing initially will result in deeper penetration than soil temperatures that are warmer.

IV. The Prediction of Depth of Frost Penetration.

Berggren [W. P. Berggren, Trans., Am. Geophys. Union, 24, (3) 71-77 (1943)] has developed a rigorous solution to the problem of depth of freezing temperature penetration which includes certain of the properties of soil, as: thermal conductivity; heat capacity; temperature condition of the soil (as well as the air); and the latent heat of fusion of water in the soil. It is admitted that great difficulties are encountered in measuring thermal properties of soils subject to such complex variables as structure, packing, moisture content, and migration of moisture during freezing. The last mentioned phenomenon is likely to invalidate any determination using initial soil properties.

Berggren has developed his penetration relationships through the analysis of an idealized system. The mathematical analysis

may then be rigorous. Failure of the theoretical to agree with experimental results are charged to excessive idealization. The idealized equations at least will serve as: (a) a guide to interpolation or extrapolation of observed soil-temperature variations under freezing conditions; (b) an indication of the departure of soil-properties (including moisture-distribution during freezing) in situ from laboratory measurements on samples; (c) a means of predicting the rate of freezing, particularly as influenced by the initial and subsequently-imposed conditions of temperature and/or the presence of snow or vegetation.

The predictions of rate of freezing may be assumed to be fairly reliable since it has been found that the rate of increase of the depth of the frozen layer is relatively insensitive to departure of field conditions from the properties of idealized systems

[Berggren, loc. cit.]

Berggren has defined the ideal system as a semi-infinite slab, bounded only by a single plane (depth $x = 0$) of porous material through which a liquid, initially at uniform temperature t_0 (above the freezing point), is uniformly diffused. The surface temperature is then reduced to, and maintained at a sub-freezing value, t_s , at a certain time $\theta = 0$. A frozen layer is then assumed to form at the surface, $x = 0$. The increasing thickness of this layer is to be denoted by the term X . The freezing point of the permeating liquid is denoted by t_f . Its latent heat of fusion (per unit volume of matrix material) is L . It is assumed that no volume-change will occur during freezing. Other thermal properties of the frozen and unfrozen

material are identified as follows:

Property	Solid	Liquid
Thermal conductivity	k_1	k_2
Volumetric heat-capacity	C_1	C_2
Thermal diffusivity	$[= (k/C)a_1]$	a_2

The differential equation applicable to the solid zone is

$$\frac{\partial t}{\partial \theta} = a_1 \frac{\partial^2 t}{\partial x^2} \text{ for } 0 < x < X \quad (29)$$

For the liquid zone the differential

$$\frac{\partial t}{\partial \theta} = a_2 \frac{\partial^2 t}{\partial x^2} \text{ for } x > X \quad (30)$$

The initial conditions are

$$t = t_0 \text{ for } x > 0, \theta = 0 \quad (31)$$

$$X = 0 \text{ for } \theta = 0$$

The fixed boundary conditions are

$$t = t_s \text{ for } x = 0, \theta > 0 \quad (32)$$

$$t \rightarrow t_\infty \text{ for } x \rightarrow \infty, \theta > 0 \quad (33)$$

The conditions at the advancing boundary of the frozen layer will then be

$$t = t_f \text{ for } x = X, \theta > 0 \quad (34)$$

$$k_1 \left(\frac{\partial t}{\partial x} \right)_{X=0} - k_2 \left(\frac{\partial t}{\partial x} \right)_{X=0} = L \frac{dx}{d\theta} \quad (35)$$

The temperature-distribution at any time θ , in conformity with the foregoing conditions are of the form

$$t = A_1 + B_1 G \left(\frac{x}{2\sqrt{a_1 \theta}} \right) \quad \text{for } 0 < x \leq X \quad (36)$$

$$t = A_2 + B_2 G \left(\frac{x}{2\sqrt{a_2 \theta}} \right) \quad \text{for } x \geq X \quad (37)$$

where A_1 , B_1 , A_2 and B_2 are constants defined by

$$A_1 = -(t_f - t_s), \quad (38)$$

$$A_2 = \frac{-(t_o - t_f) G(\sqrt{a_1/a_2} \beta)}{1 - G(\sqrt{a_1/a_2} \beta)}, \quad (39)$$

$$B_1 = \frac{t_f - t_s}{G(\beta)}, \quad (40)$$

$$B_2 = \frac{t_o - t_f}{1 - G(\sqrt{a_1/a_2} \beta)} \quad (41)$$

G is the probability-integral (also known as the Gauss "error function").

$$G(y) = \frac{2}{\sqrt{\pi}} \int_0^y e^{-z^2} dz \quad (42)$$

$G(y)$ can be approximated fairly well by $G(y) = y$ up to $y = 0.6$.

The magnitude of the function then approaches unity.

The thickness of the frozen layer is found to be proportional to the square root of time

$$X = 2\beta \sqrt{a_1 \theta} \quad (43)$$

where a_1 is the thermal diffusivity in the frozen zone and β is a "growth coefficient". The magnitude of β in terms of the dimensionless parameters $\sqrt{\frac{k_2 C_2}{k_1 C_1}} \left(\frac{t_o - t_f}{t_f - t_s} \right)$ and $\frac{L}{C_1(t_f - t_s)}$ are noted in the following Table II.

Table II. Magnitudes of β in the equation $X = 2\beta\sqrt{a_1\theta}$

$\left(\frac{k_2C_2}{k_1C_1}\right)X$ $\sqrt{\frac{t_o - t_f}{t_f - t_s}}$	0	L / $[C_1(t_f - t_s)]$ =				
		0	1	5	10	20
β for $a_1/a_2 = 1.00$		β for $(a_1/a_2) = 1.82$				
0.10	1.195	1.130	0.578	0.299	0.216	0.155
0.20	0.978	0.915	0.540	0.291	0.213	0.153
0.50	0.648	0.635	0.455	0.271	0.203	0.148
1.00	0.477	0.442	0.361	0.241	0.187	0.141
2.00	0.305	0.285	0.257	0.197	0.162	0.127
5.00	0.1488	0.1425	0.138	0.124	0.112	0.0961
10.00	0.0807	0.0786	0.0779	0.0747	0.0714	0.0662

In computing the time required for a soil to freeze to a given depth the equation previously described in the form

$$X = 2\beta\sqrt{a_1\theta} \quad (43)$$

is rewritten as

$$\theta = \mu X^2 \quad (44)$$

where μ is the "freezing-time coefficient" and

$$\mu = (1/4a_1\beta^2)$$

The dimensions of μ are then time/distance² or an inverse thermal diffusivity. Table III includes certain data pertinent to fresh snow, dry soil, moist soil (frozen and unfrozen), wet soil (frozen and unfrozen), ice, and water. The designations as previously identified are: a , thermal diffusivity [$a = k/C$]; k , thermal conductivity; C , volumetric heat-capacity; L , volumetric heat of fusion and θ as defined above (for $t_s = 23^\circ\text{F}$, $t_f = 32^\circ\text{F}$, and $t_o = 50^\circ\text{F}$).

Table III. Representative Thermal Properties and Freezing Time
($\theta = \mu X^2$)

Substance	Per cent moisture by volume	a	k	C	L	μ
Fresh Snow	10	0.04	0.12	3	900	-
Dry Soil	0	0.012	0.19	16	0	220
Moist Soil, frozen	-	0.031	0.68	22	-	-
Moist Soil, unfrozen	20	0.017	0.48	29	1800	280
Wet Soil, frozen	-	0.042	1.21	29	-	-
Wet Soil, unfrozen	45	0.022	0.97	44	4000	330
Ice	-	0.048	1.34	28	-	-
Water	100	0.005	0.31	62	9000	510

Berggren has considered the temperature changes in dry soil. The problem reduces to that of a semi-infinite slab whose surface-temperature is suddenly changed. The rate at which a given temperature penetrates the dry soil is, as before, proportional to the square root of time. A computed plot of inches penetration to 32°F vs. hours indicates a progressive increase in penetration in the order: dry soil, moist soil, wet soil, and moist soil under 4 in. snow; with the first mentioned showing the most rapid penetration. After some 5½ hours a 3 1/2 in. penetration is noted in wet soil as contrasted to 18 hours for the same penetration for dry soil with a 50°F initial temperature cooled to a surface temperature of 23°F. It is recognized that the computations resulting in the curves described neglect the heat removed from the unfrozen soil, therefore, the actual depths frozen under a snow (or vegetal cover) layer at any given time is quite possibly lower by as much as 40 per cent than the estimated depth.

The expression for temperatures in the solid (t_1) and liquid (t_2) zones are given by Berggren as:

$$[t_1 = [23.0 + 54.1G(0.90x)] \quad (45)$$

$$\text{and } [t_2 = 26.8 + 23.2G(1.21x)] \quad (46)$$

where x is the depth in feet from the soil-surface.

Several adaptations of the above described ideal analyses may be considered. Stefan [J. Stefan, SitzBer. (Math-Natur.) der Wiener Ak., 98(IIa), 965-83 (1890)] obtains for the case of the initial soil-temperature coinciding with the freezing point the expression

$$x = \sqrt{(2k_1/L) (t_f - t_s) \theta} \quad (47)$$

assuming that the latent heat of the material is large compared to its heat capacity. The "growth coefficient" β is given for this case

$$\beta = \sqrt{C_1(t_f - t_s)/2L} \quad (48)$$

Good agreement between experimental and calculated penetrations have been reported using these simplifications [T. B. Franklin, Proc. Roy. Soc. Edinburgh, 40, 56-79 (1919-20)].

Stefan [loc. cit.] has further simplified the case of unfrozen material initially at the freezing point. A correction for the time-variation in surface-temperature is applied, replacing the product $[(t_f - t_s)\theta]$ in equation (47) above for x , by the integral of the varying temperature-difference $(t_f - t_s)$ over the time interval concerned.

Shannon [loc. cit.] has used the Berggren equation for determining the depth of frost-temperature penetration

$$x = 2\beta \sqrt{a_1 \theta} \quad (49)$$

in relation to the "freezing index" values. A log plot relating freezing index and frost penetration gives a flat curve approximating a straight line. Experimentally measured values of frost penetration, however, are less than those calculated by the Berggren formula by nearly 50 per cent. Thermal characteristics of the soil, temperature

conditions, and the effect of the surface boundary upon penetration differ so widely from one soil to the next that a direct application of Berggren's equation is useful only as an approximation. The specific heat of dry soil is nearly constant for all practical purposes, hence the value of the heat capacity of soil varies directly with its unit weight and water content. Also, the total latent heat of fusion of the water permeating into a soil varies directly with the water content of the soil. When soils of similar thermal properties are observed, the linearity of the plot of depth of freezing of soil vs. freezing is well defined. Gravels, for instance, show a much greater thermal conductivity than such soils as silt, clay, or peat.

Shannon's conclusion is in agreement with that of Berggren, that the depth of freezing is greatest in dry sand or gravel. If, however, ice segregation occurs in soil with the resulting increase in water content, the depth of frozen soil would be reduced due to the heat required to change the temperature and freeze the additional migrating water.

Berggren has recognized the transmigration of moisture under the influence of gravity after freezing has begun, a condition discussed herein [A. B. C. Anderson, et. al., Trans., Am. Geophys. Union, 23, 355-71 (1942)]; however, no corrections have been applied to the idealized equations. Likewise no corrections have been applied due to dissolved salts and internal pressure of small droplets. There is no question but that these factors may affect the idealized equations appreciably and may account in part for the differences between calculated and observed values reported by Shannon.

Kostin [S. I. Kostin, Csnovy meteorologii i klimatologii, Leningrad: Gidrometeorologicheskoe Izdatel'stvo, pp. 86-90, 93-95, (1951)] includes among the variables affecting frost penetration: climate, snow cover, surface relief, soil structure, moisture content, type of soil, vegetative ground cover, and ground-water level. Repeated freezing and thawing have been shown to change the physical properties of the soil mainly by increasing porosity and permeability. The moisture content of the soil under the frozen layer increases through condensation of vapor from deeper and warmer layers of soil.

A still different approach to the problem of frost formation and penetration into a soil mass is that of Keränen [J. Keränen, Fennia, 73, (1), 14 pp. (1950-51)]. Frost formation is considered as a negative heat phenomenon. The depth of frost penetration is calculated theoretically as a function of the cold stream, the accumulated amount of heat (dependent upon the specific heat of the subsoil), and the volume percentage of the water in the soil. Thawing of the frost layer is noted to be more rapid in fields where steps have been taken to drain moisture by ditching. The moisture content decreases in passing from the frost layer to the underlying unfrozen ground.

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Section 6.

Migration of Saline Solutions

Section 6. Migration of Saline Solutions

The literature is rich in publications concerning the use of various salts, lyes, sulfite liquors, and saline solutions as related to frost heave, lens, and frost boil prevention. Few of the articles, however, concern themselves with the migration of the salts after thawing. The effect of ions, liberated by base exchange phenomena, on the freezing of moisture has herein been discussed in part but little quantitative information could be found in this area.

The importance of dissolved substances which alter pH has been recognized by Clare [K. E. Clare, Proc. Second Intern. Conf. Soil Mech. Foundation, Eng., 1, 230-36 (1948)]. It was noted that water absorption is greater in alkaline soils than in soils having acidic reactions. Some very alkaline soils, however, may be waterproofed using sodium stearate or tall oil emulsions.

Dissolved salts may decrease substantially the rate of soil deformation. It has been shown by Osterberg [J. O. Osterberg, Civil Engineering, 10 (2), 100-02 (1940)] that solutions change the ion concentration through base exchange effects. The thickness of an adsorbed water film is thereby changed. Waste sulfite liquor bleaches have a particularly large effect.

I. Migration of Sodium Chloride, Calcium Chloride, and Calcium Magnesium Chloride Solutions at Low Temperatures.

The migration of certain salts as sodium and calcium chloride in different types of soils has been studied in detail by Slessor [C. Slessor, Engineering Bull. Purdue Univ. Res. Ser. 89, Highway Res. Series 11]. The study includes low temperature effects especially in

regard to freezing of soils and formation of ice lenses. Certain of the conclusions pertinent to moisture migration are herein reported.

1. Calcium chloride and sodium chloride migrate differently under similar conditions. The latter tends to crystallize and form a white surface which is susceptible to surface washing. The hygroscopicity of calcium chloride at relative humidities above 30 per cent provides for its redissolution and re-entrance into the soil profile.
2. During natural weathering there is a downward movement of the salt in subgrades under chemically treated base courses, without surface courses.
3. As determined by chloride analysis, there is no direct correlation between moisture content and salt content.
4. Sodium chloride and calcium chloride do not migrate identically. The amount of sodium chloride decreases more uniformly with depth than does calcium chloride.
5. The migration of salt solution (sodium chloride and calcium chloride) is affected markedly by a surface layer of gravel. The capillary movement of salt solution to the surface during dry periods is greatly retarded, thereby preventing sodium chloride runoff. The calcium chloride content in the upper horizon is diminished by reducing the effectiveness of the evaporation forces.
6. The downward movement of calcium chloride and sodium chloride is accelerated by a receding but still high ground-water table. The former salt migrates more readily than the latter. The experiments conducted to establish the validity of the above statement were conducted during a cold period of high snow and rainfall.

7. Fine textured soils are more capable of holding water soluble salts and in replacing the salt content after downward migration by re-entry (upward movement) by capillarity than are coarse-textured soils. The importance of defining soil texture in a study of vapor or liquid migration cannot be overemphasized.
8. A bituminous paving over a soil accelerates the downward migration of saline solution (both sodium and calcium chloride) from coarse-textured base courses into underlying fine-textured subgrades.
9. The base exchange binding capacity of clays play an important factor in the migration of salt solution. Cation migration is lower due to their attraction to soil colloids. The divalent calcium ion is more permanent than the chloride ion under comparable conditions.

Slesser has established such variables as soil texture, evaporation temperature, percolation rates and soil cover all of which affect the migration of soluble salts in soils. Laboratory studies to complement field studies are reported covering the above variables. The difficulty in segregating each of the variables was recognized since soil texture actually determines soil capillarity. The latter is closely associated with evaporation and leaching phenomenon. Certain of the experiments and conclusions reached by Slesser are enumerated below.

1. The upward migration of calcium chloride, sodium chloride, and calcium magnesium chloride (CaMgCl_4) is very pronounced under conditions of low humidity (rapid evaporation). The migration is a function of the soil texture. A more pronounced movement is found in fine textured soils.

2. Temperature as a variable was studied in the laboratory for four silt models with 2 per cent calcium chloride under impervious pavement conditions. The tests included a surface temperature of 210°C (410°F) for 28 days; 60°C (140°F) for 90 days; 25°C (77.1°F, room temperature) for 21 days; and 63 hours at -60°C (-76°F, using dry ice). Additional samples containing calcium magnesium chloride and sodium chloride were subjected to evaporation and heat tests which permitted additional conclusions to be drawn with regard to chemical migration. All tests showed that the effect of the temperature or migration (as indeed for any of the variables) of calcium chloride, sodium chloride, or calcium magnesium chloride is always in the same direction, although the degree may be different. If the magnitude of the temperature change is enough to bring about a change in phase of water, the variable is important in governing chemical migration in soil. If no change in phase occurs, temperature is of less importance.
3. As has been stated from field observations the salt content of a soil decreases with depth. The concentration of salt in the soil will be an influencing factor on the amount of upward chemical movement. The condition of impervious pavement is assumed here also.
4. Water migration by percolation depletes rapidly the chemical content of a soil whether the chemical be sodium chloride, calcium chloride or calcium magnesium chloride.

II. Migration of Naturally Occurring Sodium Sulfate in Soil Related to Humidity.

Purie [A. N. Purie, Punjab Irrigation Res. Inst., Rept. Year Ending April 1936, p. 3-8.] in a report on certain Indian Soils states

that salts tend to accumulate in the surface horizon during periods of low temperature. Such would be reasonable due to the low rate of evaporation of moisture permitting unbroken capillary movement of salt-laden water from lower horizons.

Highly saline soils of the area (sodium sulfate and sodium chloride) show white crystalline surface formation due to the migration of saline solution to the surface. The effect is more marked in the case of sodium sulfate than sodium chloride. The texture of the soil is changed to a loose fluffy consistency with the rise of the sodium chloride. Atmospheric humidity is an important factor in causing the migration of saline moisture. Low temperature and wide range in humidity are important factors in causing the migration of saline moisture. With low temperature and wide range in humidity variation during the day and night (30 to 100 per cent) sodium sulfate was noted by Purie as moving to the surface and efflorescing, that is, the alternate absorption of moisture from the atmosphere and its evaporation appears to be responsible for the salt efflorescence. The salt content of the surface soil is affected by the same moisture migration effects as the salt content of soils about 5 feet in depth. The soil below 5 feet acts as a receptical for saline moisture migrating downward from the surface layers. Lower depths supply saline moisture to upper levels by moisture migration.

III. The Effect of Water Vap on the Migration of Salts.

Laboratory experiments by Erlenmeyer [H. Erlenmeyer, Chem. Erde, 8, 317 (1933)] have shown that the migration of salts in soils

may be effected by water vapor in the absence of ground or even capillary water. This migration was established by allowing water vapor to diffuse through sand underlying a sodium nitrate layer on top of which was a filter paper. The sodium nitrate was found in appreciable quantities in the filter paper.

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Section 7

Barometric Pressure as a Function of Moisture Migration

Section 7. Barometric Pressure as a Function of Moisture Migration.

The influence of barometric pressure on the absorption and migration of moisture was proven by Krynine [D. D. Krynine, Highway Res. Board, Proc., 9, 66-74 (1929)]. Tubes filled with hygroscopically moist clay whose ends were immersed in water were weighed at intervals over a 35-day period. From the increase and decrease of weight with changes in barometric pressure it was found that with high pressures entrapped air bubbles contracted and moisture migrated into the soil, with decrease in pressure the entrapped air bubbles expanded, and water was driven from the tube (that is, a decrease noted in the weight of the tube).

An exception to the above general principle was reported which corresponds to a violent decrease in atmospheric pressure. With such a change, the air not only drives the water back but itself escapes from the system leading to an overall increase in weight.

The significance of entrapped air is, of course, more marked in recent earth fills than in ancient natural clays. It is expected that water would migrate more freely into artificial fills during a decrease in barometric pressure. Rising temperature would have the same effect. Fresh fills should show considerable settlement during a continual decrease of barometric pressure following precipitation. It is not completely certain that usual atmospheric pressure variations would be of sufficient magnitude to cause the described effects.

Section 8.

Deterioration of Building Materials by Moisture Migration and Frost Action

Reference

Krymire, D. D., "Barometric Pressure as a Function of Moisture Migration",
Proc. Highway Res. Board, 9, 66-74, 1929.

Section 8. Deterioration of Building Materials by Moisture Migration and Frost Action.

The causes of and factors likely to play a role in the deterioration of building materials by frost action have been studied by Wilsdon [B. H. Wilsdon, J. Soc. Chemical Industry, Trans., 53, 397T-402T (1934)]. The work of Wilsdon is unique and comprehensive and will be discussed in some detail.

Certain selected types of building stones and bricks were subjected to natural frost for two seasons under the test conditions described by Butterworth [Trans. Ceram. Soc., 33, 495 (1934)]. The conditions essentially demand that half the stone (or brick) be in the ground and half exposed to the atmosphere. Table IV includes strength, absorption, and resistance data for a selected group of bricks. The type EK is an exceptionally strong, dense stone. The types EG, EA and EE as noted in the "Dry" column are of about the same strength.

Table IV. Frost Resistance in Bricks

Type	Strength lbs/sq. in.		Water absorption weight %	Frost-resistance - Per cent Number of bricks damaged		
	Dry	Wet		Badly	Seriously	Little
EK	8960	8160	8.6	-	-	100
EG	4060	4050	17.7	-	-	100
EA	4090	4030	17.2	45	33	22
EE	4590	4110	16.9	22	51	27
EF	4290	3870	20.4	95	5	0
EB	1724	1140	25.1	80	-	20

It is obvious that neither strength nor absorption alone is a completely reliable guide to the frost-resistance of a building brick. The physical factors related to and governing porosity must be defined and studied since it is reasonable that the frost action is closely

dependent upon porosity. Porosity is more than a measure of the percentage of pore space. The size, shape, and distribution of pore spaces are factors, and govern capillary movement of moisture. They will, therefore, be related to frost action on building materials.

Failure of concrete always comes from one of two causes

[F. M. Lea and N. Dowey, J. Inst. Civil Engrs., 32, 248-95 (1949)]. Those which are internal to the concrete and those which determine its quality (dependent in turn upon mixing, placing, and curing). Factors such as shrinkage, creep and thermal also must be considered. The external causes of deterioration, broadly classified as physical, chemical, or mechanical, arise from natural weathering, exposure to extremes of temperature or mechanical overload, or from more unusual destructive agencies such as electrolysis.

The following factors are discussed in some detail as they relate to frost action in porous solids:

1. Permeability to liquids and gases.
2. Suction (a term though not highly descriptive perhaps is the best that can be devised to describe the behavior of a porous solid with respect to a liquid. Variables affecting capillary pressure are: the nature of the liquid, the characteristics of the porous structure, and the relative amount of pore space occupied by the liquid).
3. Mechanical effects produced by wetting or by moisture movement into and through the porous body.

I. Permeability in Porous Solids.

The movement of a liquid in a porous solid may be defined by the general equation

$$Q = -k \frac{\partial \phi}{\partial x} \quad (49)$$

where Q is the flow per unit cross section, normal to x , ϕ is the hydrostatic pressure, and k is the transmission constant.

When all pores are saturated with moisture and as long as the velocity is small, the value of k will be a function characteristic of the porous solid and the velocity of the liquid. Under such conditions the flow of liquid is described as percolation and is to be distinguished from permeation. The latter is descriptive of migration of moisture when the pores are not saturated. In permeation the flow is restricted to films and the pressure within the liquid is a function of film curvature. The value of Q is then given by the expression

$$Q = - [k_1 \phi / \partial x] (a_4 / \partial x) \quad (50)$$

The term a_4 represents the moisture concentration in the porous solid. The bracketed expression is a complex permeability function relating the flow to the moisture gradient.

Various attempts have been made, with little success, to describe ϕ explicitly in such geometrically simple cases as those presented by close-packed uniform spheres [B. H. Wilsdon, Mem. Dept. Agric. India, 6, 154 (1921); R. A. Fisher, J. Agric. Sci., 16, 492 (1926); E. Manegold, C. Stüber and W. Engelhardt, Kolloid-Z., 64, 12, 307 (1933)]. A statistical treatment of geometrical properties of a porous medium, including a grading of sizes and shapes of particles and

their compaction, has been somewhat successful in explaining ϕ and will be described. An empirical approach is used by Wilsdon [loc. cit.] to obtain as close a relationship as possible between experiment and theory.

II. Suction.

Laplace's equation for the hydrostatic pressure ϕ is assumed to be valid. The force responsible for suction is postulated as

$$\phi = 2T(1/r_1 + 1/r_2) \quad (51)$$

The term T is the air-liquid surface tension, and r_1 and r_2 are the principle radii of curvatures of the surface.

The water in a porous medium is assumed to be held at the points of contact of the constituent particles. If this assumption is true the curvatures may be positive or negative. For a pair of spheres which are wetted by a liquid up to a point where $r_1 = r_2$, the curvatures will be negative and from equation (51) the value of ϕ must be negative. This value of ϕ would represent the pressure deficiency measured in passing across the liquid surface from air to liquid. The drops of liquid can only be assumed as independent over a limited range of moisture contents for the uniform assemblages of spheres in the porous solid [Keen, J. Agric. Sci., 4, 170 (1924)]. For graded assemblages of spheres the limitation just described may not be too serious since there should be relatively few contact points when compared with the 12 per sphere for trigonal assemblages of uniform spheres.

A second difficulty is the variation of curvatures of the solid surfaces from point to point. Such variations cause corresponding volume changes. It is not, nevertheless, unreasonable that the

pressure in the free water will follow some such law as already given by equation (50)

Efforts have been made to determine the range over which equation (51) is valid. Four experimental approaches have been used to estimate the value of ϕ (hydrostatic pressure). All involve the determination of moisture content at equilibrium; (a) under direct hydrostatic pull for pressure deficiencies less than one atmosphere, (b) under osmotic suction through semipermeable membranes, (c) under reduced vapor pressures, and (d) under centrifugal force.

Wilsdon [loc. cit.] reports data on three types of material (ball clay, pozzolana, and sand) and compares the pressure deficiencies in these porous materials at various moisture contents by hydrostatic (method a) and by osmotic suction (method b) methods. The estimates by hydrostatic were actually of a different order of magnitude to those of osmotic suction. The hydrostatic pressure required to reduce the water content of a saturated material to a given value was always found to be smaller by 1/10 to 1/100 than corresponding to the osmotic pressure method. The discrepancy is comparable when data for methods (a) and (c) are compared. Experimental values of ϕ differ from those calculated from the vapor pressure data of Szigeti [Kolloid-Beh., 38, 99 (1939)] by the Kelvin formula [Phil. Mag., 42, 448 (1871)]

$$\phi = (R'\theta/W) \log_e(p/p_0) \quad (52)$$

where R' is the gas constant per gram, W is the specific volume of water at temperature θ (in °A), and p and p_0 are vapor pressures in the porous material and a flat surface respectively. The discrepancies in pressure deficiencies for H-permutite and H-clay using Szigeti's

data (and the Kelvin equation) compared with actual vapor pressure measurements are large.

Scofield [H. H. Scofield, Am. Soc. Testing Materials, Proc., 37, 306 (1937)] has reported tests to show the effect of freezing on permeability, strength, and elasticity of concrete and mortars. Russian engineers have reported the practice of freezing concrete solid in the forms in cold months followed by the removal of the forms after the concrete has thawed and hardened in the spring. No injury was reported to the concrete. In Scofield's tests specimens to be frozen immediately were subjected to a temperature of 10°F for freezing periods of 1 to 7 days. Certain of the specimens were subjected to a delayed freezing cycle in which the concrete was allowed to harden for various periods before freezing. The freezing period for these samples was 2 days at 5°F. An average loss of strength for concrete immediately frozen was some 40-60 per cent of the normal unfrozen strength of the same age. These results are not in accord with those reported by the Russian engineers.

Moisture permeability tests were also reported. A large increase in leakage for frozen specimens was noted, especially for samples frozen over a 7-day period. The leakage was less for specimens tested at one year, however. Specimens cured for 24-27 hours in dry warm air before freezing were not injured.

It is a necessary conclusion that the water held by a porous medium depends in some manner upon the way in which equilibrium is approached. The liquid films which are formed in the porous structure

are not in the same equilibrium state under all the above described conditions. Wilsdon postulates that the films brought to equilibrium through vapor phase or through a semipermeable membrane may possess an oriented structure which reduces their vapor pressure sufficiently to invalidate the Kelvin equation for the systems. Other lines of evidence [Kerjaquin, Z. Physik, 84, 657 (1933); Hardy, Phil. Trans., 230A, 1 (1932)] confirm these abnormal pressure behaviors of thin water films. Methods based upon the centrifuge (method d) permitting very high pressure deficiency measurements may add to the fund of information.

III. Capillary Rise in Porous Medium.

Frost effects in porous solids should be closely related to capillary rise. Equation (52) may be modified to show the relationship between vapor and hydrostatic pressures at any point in a solid. Summing the constants of equation (53)

$$\phi = 3190 \log_{10} (p/p_0) \quad (53)$$

for values of p and p_0 in atmospheres. Assuming a collection of uniform spheres of 0.0001 cm. radius in a system 5 per cent by weight moisture, a pressure deficiency of 4.3 atmospheres exists. The vapor pressure lowering would be 0.51 per cent. The height at which this vapor pressure would be attained above a saturated surface may be calculated from Boyles law in the form

$$h = (R' \theta / g) \log_e (p/p_0)$$

or to 145 ft. No such heights have been observed experimentally (see page 26), however, the difference between fact and theory again points out the abnormal behavior of thin water films, as well as the inadequate knowledge of the nature of small particles and pore spaces in porous materials.

IV. Mechanical Effects Produced by Crystallization of Solids in Porous Bodies.

In addition to the percentage pore space and over all strength of building materials, the formation of solids (as ice crystals) greatly influences the mechanical resistance to disintegration. A stone material (as a cement) whose pores are in the effective diameter of 0.0005 cm. (highly microporous) is very susceptible to detrimental frost action. Convenient experimental observations on porous materials may be made involving the crystallization of sodium sulfate decahydrate, $\text{Na}_2\text{SO}_4 \cdot 10\text{H}_2\text{O}$, at temperatures above and below the transition temperature, 32.4° . The ultimate results closely resemble those derived when water is allowed to freeze and thaw in the porous medium.

The concept of crystal thrust is inadequate in explaining many of the rather violent disintegrations noted in porous bodies subjected to alternate freezing and thawing. There is doubt that it is thermodynamically possible for an incompletely confined saturated solution of a salt (as sodium sulfate) to do work under isothermal conditions by crystal growth only. It must be assumed, then, that energy due to transformation in a porous solid of a solid phase is available for work to be done. In the case of ice crystallization it is probable that no stresses could be produced if the transformation from liquid to solid could take place very slowly without confining the water. The greatest damage is noted when pores are sealed by freezing the outside surface and under conditions of the greatest supercooling. It has been observed that highly microporous solids are characterized by marked tendencies toward supercooling. Highly hydrated salts (in solution) have been known to cause nearly explosive disintegrations in microporous solids under conditions of strong supercooling.

The conclusion is obvious that building materials of high microporosity should only be used in dry regions of low frost action. Building materials may be examined for their resistance to frost action by relatively simple salt solution tests. The simple application of the Kelvin equation is inadequate in explaining the properties of water in microporous solids.

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Section 9.

Conclusions and Recommendations

Section 9. Conclusions and Recommendations

1. Considering the rather considerable coverage of the literature in this survey, it is obvious that there are large gaps in the published data on thermal translocation of water in soils. The few papers containing extensive data on the subject of transmigration of moisture do not use a low temperature zone in the thermal gradient lower than 0°C. It would appear to this writer that very sharp differences might occur in the quantity of water moved if a low temperature zone at -20°C were used. The very small amounts of water noted as being moved to or from a 0°C zone might well be greatly increased.
2. The experimental methods of following moisture migration have been in general limited to gravimetric procedures. The use of x-ray technique may be of interest. Indeed one paper published in the USSR [I. Sedletskii, M. Sumgin, and A. Malovichko, Doklady Akademii Nauk SSSR, 47, 296 (1945), see also SIFRE File U 1078] includes this method in the study of moisture migration at sub-freezing temperatures. The internal structures of soil specimens (as well as migration) were followed using air-dry clay and sand samples, moistened to the desired degree and x-rayed before and after freezing, after standing for 24 hours in a chamber at -10°C and after thawing for 24 hours at room temperature. By such a technique the very important and still little understood behavior of small droplets absorbed and frozen on soil surfaces might be better defined.

3. It is the understanding of this writer that some experiments have been performed on moisture migration using tracer techniques. Such techniques appear on the surface to be well-adapted to such studies. Since non-radioactive oxygen or hydrogen would of a necessity have to be used in the production of water for the studies, any data published should be examined critically for exchange phenomenon. Thus, tracer water appearing in a capillary column above the water source (as a water table) might be so located because of exchange of the isotope from one water molecule to another rather than by actual translocation. Such exchange would be dependent upon the initial moisture content of the soil and would be of minor importance in the migration of water in highly dried soils.
4. Such an assortment of figures were found on the quantity of water raised from a water table by capillarity that further data is badly needed on this type of migration. It is the feeling of this writer that the initial moisture content of the soil has not been sufficiently well-defined in certain of the capillary migration studies. The variations from 30 feet to a few inches as being the maximum rise of water might well be explained in terms of the initial wetness of the soil above a table.
5. It does not seem justified to make any general rules on the direction of migration (as from a warm to a cool zone) of water in soil. The variables as barometric pressure, initial moisture content of the soil, and humidity appear to be important factors which make general rules questionable.

6. The work of Tsytovich (see page 7 of survey manuscript) seems highly significant in the migration of water to very low temperature regions. The fact that some 50 per cent of soil moisture remains unfrozen at -24°C in a Moscow loam might account for inconsistencies in published data involving migration to ice lenses. Too little attention has been paid to the nature of the ground water. High mineral content of the water or ions passing into the migrating water by base exchange must surely have an effect on the freezing temperature.
7. The general statement has been made that only small amounts of moisture move downward in frozen soils. That is, only run-off should result. What the exact figures are, for water percolation when water stands on frozen top soil, should be determined. The porosity of frozen soil types needs clarification.
8. The brevity of the section in this literature survey on the subject of Vapor Migration is an indication of the need for further studies in the field. It is possible that the vapor pressure is sufficiently low at subfreezing temperatures to minimize this method of translocation; however, there are not enough experimental data to prove whether this is the case or not.
9. There is no complete agreement among various investigators on the exact definition of different terms used in the study of moisture migration. As an example the proportionality term k in the Darcy equation is not a characteristic constant of a given soil since it depends upon the moisture content of the soil and upon the temperature.

No measurements of k have been found at low temperatures. The validity of the Darcy relationship for low temperatures needs further investigation.

10. The resistance measurements of Richards and Moore (see page 21) appear admirably suited to migration studies at low temperatures and would add to the existing knowledge of moisture movements in the thermal regime. It is suggested that initially frozen soils might be studied using water at 0°C for permeability measurements.
11. Capillary and adsorption effects have been considered as a combined phenomenon. In the case of vapor transmission through a soil which is unsaturated, the adsorption effect is quite pronounced due to the inherent hydroscopicity of most soils. Capillarity seems to be effective only in the movement of water in the liquid phase. There is general agreement among investigators on the relationship between the grain size of a soil and the extent of capillarity movement. There is not complete agreement, however, on the critical size of soil particles favoring maximum migration. Information on this subject at freezing temperatures is particularly scant. The variable factors such as gravitational field, texture, initial moisture content and temperature all have such a large bearing on the problem that it would appear as if each locality presents its own unique situation.
12. Reports concerned with the direction of moisture migration under thermal gradients are not consistent. Although the general movement appears to be from the high temperature zone to the low, enough evidence is reported to the contrary to infer the importance of a soil texture variable.

13. The whole area of water migration through snow cover is not as completely investigated as might be expected.
14. Insulation of frozen ground by inexpensive and abundant materials for the purpose of constructing engineering or architectural structures appears to be possible. The search for suitable insulators would be desirable.
15. The susceptibility of highly microporous building stones to moisture permeation (and migration) and subsequent detrimental frost action has been definitely proven. Additional evidence should be presented for the limiting pore size permissible in concrete and other poured foundation materials to be used in cold regions.

SUMMARY

Throughout this literature search a lack of continuity in defined terms was evident. Definitions relating to soil characteristics were not universally accepted. The development of applicable thermodynamic theories to permit predicting moisture transmigration, therefore, becomes understandably difficult.

The initial moisture states of soils are usually ill defined in the published data treating moisture migration. The difficulties are therefore magnified when attempting to correlate the experimental works of several authors.

The scarcity of experimental (laboratory and field) data relating soil particle size to the energy of absorption of water droplets makes difficult the development of equations relating moisture flow to temperature and soil character. The forces of attraction associated with the surface of silt and other very fine particles must be extremely high.

General trends in moisture migration by capillarity are predictable from knowledge of soil particle size and initial moisture content at temperatures above freezing. The capillarity characteristics of frozen soils are so well understood.

The writer is of the opinion that a very important phase of moisture migration has been neglected: that of liquids standing (or flowing) over frozen soil. Such a condition would be prevalent in northern (arctic or subarctic) regions where glacial melt water (or water from ice and snow) either stands or flows over a permafrost area. Also the nature of the rise of a water level through a frozen soil layer might better be predicted.

Probably the richest source of information on low temperature moisture migration is found under the general heading of the formation and cure of frost heave (boils). Unfortunately the complete description of the nature of the soil, initial moisture content, particle size, etc., of the various soils for each report are often lacking. The lack in uniformity reported on the nature and direction of migration as well as on the nature of crystal growth and direction of soil expansion is undoubtedly due to the variety of soil conditions encountered.

The direction and the magnitude of soil expansion encountered upon cooling a soil mass below normal freezing point of water has been proven to depend upon the temperature reached and the normal moisture content of the soil. The distribution of the soil moisture through a soil profile also has been shown to be an influencing factor in the direction of moisture migration. This latter factor controls in a measure the phenomenon of soil contraction rather than the usual expansion on freezing.

Moisture migration through snow surfaces and into firn masses is reported in considerable detail and appears to be better understood and more predictable than migration into and through soils. Experimental methods providing visual observation are more readily available (by colored solutions) for snow systems than those of soil.

A considerable amount of data are published concerning the transmigration of moisture under controlled thermal gradients. The experiments are well planned and accurately reported. From the standpoint of interest to this particular survey (thermal regime), however, a low temperature zone of 0°C does not permit the reader to draw general conclusions as to the possible behavior in areas where a zone may well be at -20°C or lower. The extent of migration might well be

greatly exalted under such conditions, and the extent of vapor migration might be a more important factor than at higher temperatures.

Additional light might be shed on moisture migration (under the influence of thermal gradients) if the thickness of the moisture shell around soil particles at the point of maximum thermal translocation were known. The obvious difficulties in such a determination are in the irregularities in shape of soil particles (lack of spherical nature) and the prevalence of air spaces. Variations from one soil to another would be profound.

Some of the better experimental work is reported on frost penetration since electrical resistance methods adapt themselves well to frost studies as long as a soil contains more than the wilting moisture. The extent of protection to frost penetration by snow, foliage (crop), and slope direction are well-defined. Measurements of frost penetration at temperatures to -21°F are reported, as is the protection afforded by 24 inches of snow. Although admittedly empirical, knowledge of freezing index, permits an approximation of the depth of frost penetration. Berggren's detailed thermodynamic treatment of frost penetration is admirable but a comparison of predicted values for penetration with experimental observations are off by 50 per cent, due in a measure to different thermal characteristics of soils, temperature conditions (as in the subsurface), and the effects of the surface boundary upon penetration which differ widely from soil to soil and location to location.

The migration of saline solutions at low temperature is quite well-defined and the reader is referred to pages 74 and 75 of this report for a summary of conclusions pertinent to this type of migration.

For practical importance the effect of barometer pressure is negligible on moisture migration through soils.

The penetration of moisture and frost into building materials is highly dependent upon particle size as well as size, shape, and distribution of pores. The curing time of concrete materials poured during low temperature is a function of the extent of moisture migration and thence of strength. Inconsistencies are found in the literature originating from the U.S.S.R. since Russian engineers report higher strength of concrete poured at low temperatures allowed to freeze, and the molds removed with the return of warmer temperatures than concrete handled similarly in the United States.

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