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Gold, L. W.

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The Ice Factor in Frozen Ground

by L.W. Gold

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THE ICE FACTOR IN FROZEN GROUND

L. W. Gold

*Division of Building Research, The National Research Council
of Canada, Ottawa, Ontario K1A 0R6 Canada*

ABSTRACT

The nature of ice formations in frozen ground has been a major subject of investigation for Ross Mackay. In this paper, characteristics of the freezing of water in porous materials are reviewed and examples given of the effects of frost action. The thermodynamic basis for the explanation of ice segregation is given and the possible role of spreading and disjoining pressures is described. A criterion for the initiation of ice lenses is presented. Brief consideration is given to the question of whether equilibrium can ever be attained under natural conditions.

RÉSUMÉ
Le facteur glace dans le sol congélé

La nature des formations de glace dans le sol congélé est un thème de recherche majeur de Ross Mackay. Dans cet article, les caractéristiques de l'engel de l'eau dans des matériaux poreux sont précisés et des exemples sont donnés des effets de l'action du gel. La base thermodynamique pour expliquer la ségrégation de la glace est donnée et le rôle possible des pressions de dispersion et de disjonction est décrit. Une critèrium d'initiation des lentilles de glace est présenté. La question de savoir si l'équilibre est jamais atteint dans des conditions naturelles est brièvement examinée.

ФАКТОР ЛЬДИСТОСТИ В МЕРЗЛОМ ГРУНТЕ

Л. В. ГОЛД

РЕЗЮМЕ

Характер льдообразования в мерзлом грунте является главным предметом исследований проводимых Росс Макаем. В настоящем докладе рассматриваются характеристики замерзания воды в пористых породах и даются примеры эффекта морозобойного действия. Представлены термодинамические основания для выяснения сегрегации льда и описывается возможная роль распространяющегося и разъединяющего давлений. Дается критерий для зарождения ледяных линз. Коротко рассматривается возможность достижения устойчивого равновесия при обычных природных условиях.

INTRODUCTION

The forms of ice that occur in the ground can vary from massive bodies to minute inclusions; the average weight of ice in permafrost several metres thick can be more than five times that of the dry soil (Mackay, 1971). Ross Mackay has devoted considerable effort to delineating the various types of ground ice and to determining their origin and mode of formation (Mackay, 1972). From his studies of our northern regions he has added greatly to knowledge of ice formed in horizontal layers or lenses by forces associated with the freezing of water in some porous materials, a process called ice segregation or frost action (Mackay, 1972); of ice formed in permafrost due to the filling of cracks with water from the surface, for example, ice wedges (Mackay, 1974); of ice formed by the intrusion and subsequent freezing of water with or without segregation, for example, pingos (Mackay, 1973, 1979); of homogeneously distributed pore ice formed by water freezing *in situ* (Mackay, 1972); of ice crystals formed in underground cavities by sublimation (Mackay, 1972); or buried ice resulting from geological activity (Mackay, 1972).

These are some of the geologic manifestations of ground ice that Ross Mackay has investigated. There has been a growing appreciation of the importance of the insight and knowledge he has provided through these investigations. The performance of roads, buildings and other structures in the North has demonstrated the need to know the geological structure and thermal condition of the ground on which they are placed. The presence of ice, its form, its distribution and how it will respond to disturbance of the surface and to loads must be taken into consideration in the design, construction and operation of these structures. If one wants an outstanding

example of contributions of scientific activity to engineering and economic development, one need only look at the use made of Ross Mackay's work in current northern oil and gas development.

For several years scientists and engineers have had a very active interest in ice segregation or frost action because of the damage it can cause to roads, buildings and other structures. This phenomenon is of great concern at present because of the severe conditions it poses for the operation of gas pipelines in the North. It is also one that interests Ross Mackay, with respect both to its possible role in the formation of ground ice and to its effects during the freezing of the active layer.

Frost action in soils and building materials has been an important subject of research at the Division of Building Research, National Research Council of Canada, for several years. The effects of frost action and some of its characteristics, as demonstrated by this work and that of others, are briefly reviewed in this paper. This is followed by a statement of the thermodynamic basis for the ice segregation process. The concepts of spreading and disjoining pressures in liquid films are discussed with respect to their possible role in frost action, and brief consideration is given to the effects of gradients of temperature and pressure on the stability of ground ice.

EFFECTS OF FROST ACTION

Perhaps the most common evidence of frost action is the heaving that occurs in some road surfaces in winter. Such heaving demonstrates one of the characteristics of frost action; that is, it is a segregation of ice that causes an increase in volume greater than can be attributed to the expansion of water on freezing. In fact, the process involved does not depend on that property of water; a similar behaviour can be obtained with other liquids.

In his classical work published in 1935, Beskow stated that observations on ground ice and frost action were recorded as early as 1700. Both he and Tabor (1930) stated that not until the early 1900s was it recognized that during the freezing of fine-grain soils water can flow to the freezing zone and that the phenomenon of heaving is due to ice segregation rather than to expansion of water on freezing.

Pressures developed by frost action can be quite large. Damage to improperly protected foundations on frost susceptible soils is not uncommon. A total upward force of about 130 kN (30,000 lb) was measured by Penner (1970) on a 30 cm diameter plate placed on clay ground and prevented from heaving upwards. This behaviour is not confined to soils. Figure 5-1(a) shows displacement of brick and coping due to the formation
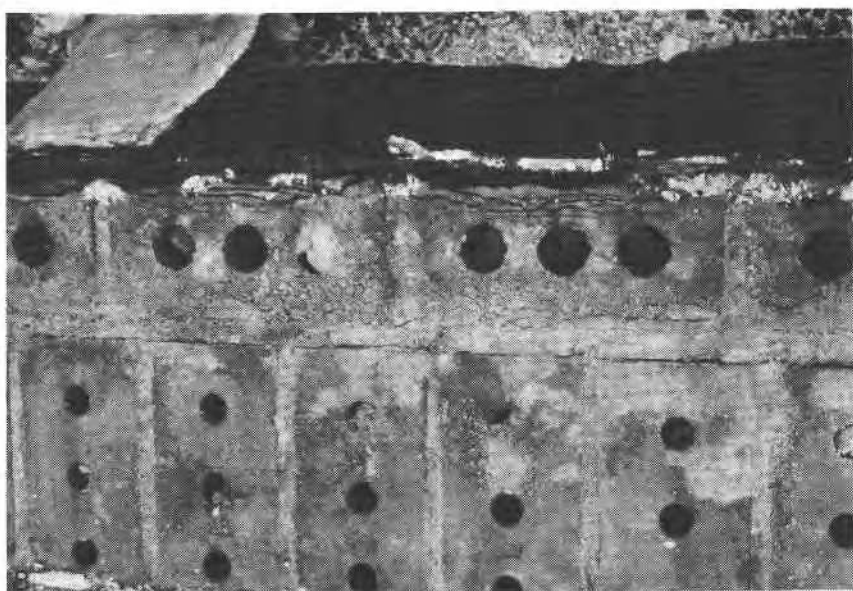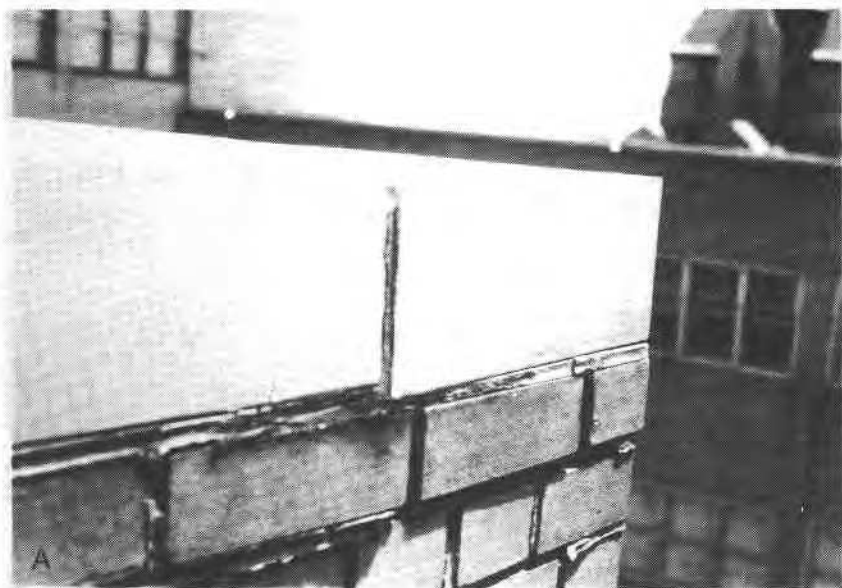
Fig. 5-1. (a) Damage to coping due to frost action.
(b) Damage to brick due to formation of ice lenses.

of ice, and figure 5-1(b) shows the splitting of brick considered to be due to ice lens formation.

Research on frost action soon demonstrated that the tendency to heave during freezing depends upon the characteristics of the soil, particularly grain size, the rate of heat extraction, pressure in the direction of freezing, water content and the pressure`or tension in the soil water. Figure 5-2
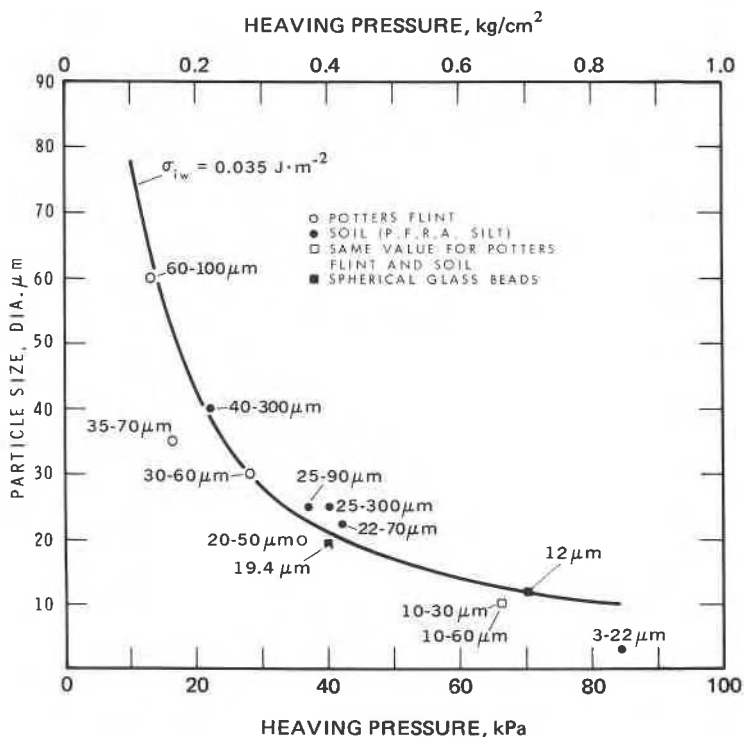
Fig. 5-2. Dependence of maximum pressure due to ice segregation on smallest particle size for various materials of given range in particle size (Penner, 1973) (Used with permission of the Organization for Economic Cooperation and Development.)

presents observations made in the laboratory on pressures developed during the freezing of particulate material of various grain size ranges (Penner, 1967, 1973). The specimens were confined and the frost action process allowed to proceed until the heaving pressure was maximum. Experiments such as these have demonstrated that the maximum force increases with decrease in the average grain size of the soil.

Figure 5-3 shows the effect on maximum pressure of increasing the
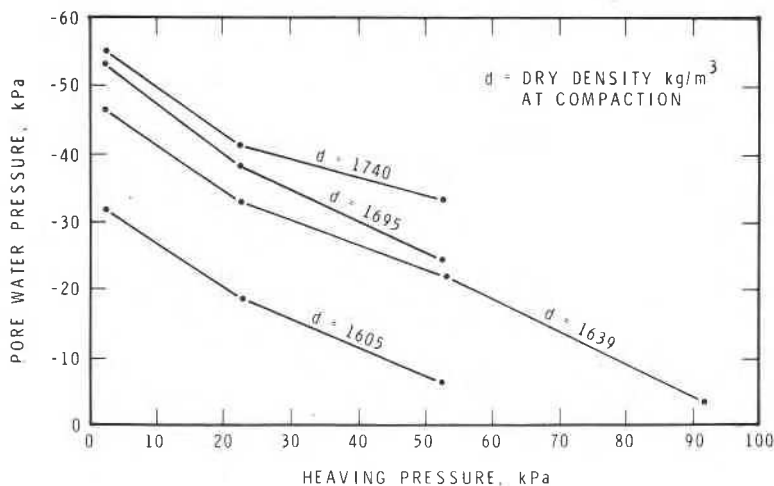
Fig. 5-3. Dependence of maximum pressure due to ice segregation on pore water pressure. From E. Penner (1959). The Mechanism of Frost Heaving in Soils, in Highway Research Board (Washington, D.C.) Bulletin *225*: 4. (Used with permission of Transportation Research Board.)

tension in water in the soil (Penner, 1959). It may be seen that the maximum pressure decreases almost linearly with increasing pore water tension.

One of the interesting features of water in porous materials is that not all of it freezes at the freezing temperature of bulk water. The amount of water that still remains unfrozen at a given temperature below freezing depends upon the material type and, in particular, upon its porosity (Williams, 1964; Litvan, 1978). In general, the amount of unfrozen water in a soil at a given temperature increases with decreasing grain size. For sand, significant amounts of unfrozen water are usually confined to temperatures above -1°C; for clay, there may still be significant unfrozen water a temperatures below -5°C. On thawing, the unfrozen water content curve is shifted to higher temperatures, indicating that on freezing the thermodynamic properties shift towards the values for bulk ice.

There has been growing evidence that the front of an actively growing ice lens does not coincide with the plane in which ice begins to form, but is on the cold side of it (Miller, 1973, 1978). Both fronts, called the ice front and the freezing front, respectively, are at temperatures less than 0°C, the degree depending on soil type, overburden pressure, pore water tension, and rate of freezing. An X-ray technique developed by Penner and Goodrich (1980) allows us to determine the position of ice while it is forming in soil in laboratory experiments. By spacing small temperature measuring

devices along the length of the specimens, they were able to determine the position of the ice relative to the 0°C isotherm. Figure 5-4 is an example of what can be seen by this method. The 0°C isotherm is indicated in the figure and the position of two ice lenses clearly defined at A and B.

Using this technique it has been demonstrated that changing pressure causes the ice front to move. If a lens has been established and the pressure is subsequently increased, a new lens is initiated on the cold side of the old one and the old one deteriorates. Conversely, decreasing the pressure after a lens has been initiated causes a new one to be established on the warm side of the old one. In figure 5-4 a lens was formed initially at A. The pressure was subsequently reduced and a new lens initiated at B.
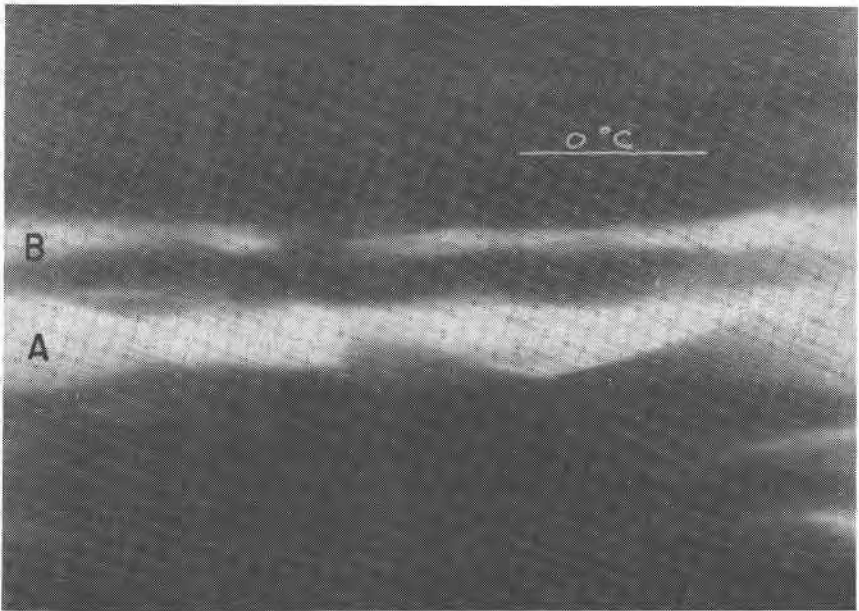
Fig. 5-4. X-ray photograph of ice lenses on cold side of 0°C isotherm (see Penner and Goodrich, 1980). See text for further explanation.

In strong porous materials such as rock and concrete, lens formation and associated heaving may not occur during ice segregation. Figure 5-5 presents a differential thermal analysis scan and associated length changes for small specimens of hardened cement paste. The freezing process began at about -8°C. As freezing progressed the specimen expanded, apparently
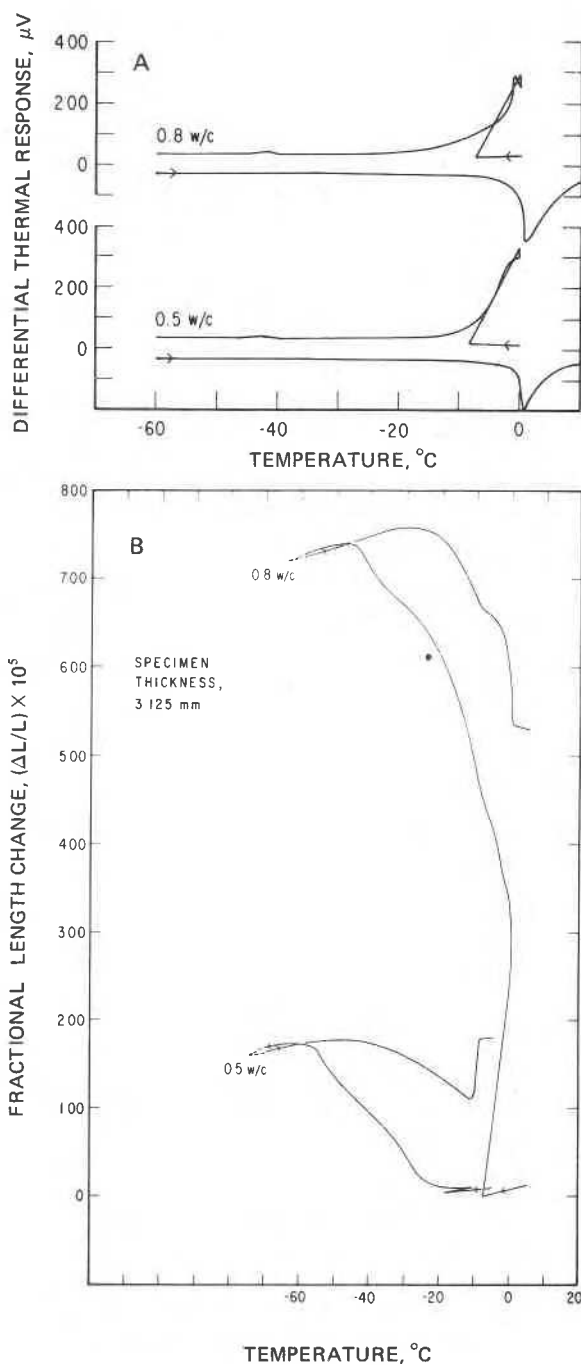
Fig. 5-5. Thermogram (A) and fractional length change (B) of vacuum saturated cement specimens determined simultaneously. The 3.2 mm thick x 32 mm diameter specimens were cooled to -60°C and rewarmed: w/c indicates the water/cement ratio in each sample. The apparent thawing at temperature above 0°C indicated in (A) is due to a temperature sensor (Litvan, 1972). (Used with permission of the American Ceramic Society.)

because the water moved out to the exterior surface. The expansion was small, in the range of strain that would be imposed on the material by a stress of about 50 MPa (7000 psi). As with soils on warming, the ice melted at a temperature higher than that at which it formed.

In summary, some of the characteristics of the freezing of water in porous materials that have been demonstrated through experiment are as follows:

1. Segregated ice may form as lenses. This implies a mobile film between the ice and the solid surfaces on which it is supported.
2. Ice segregation can be stopped by the application of pressure in the direction of freezing, by increasing the tension in the pore water, or by a combination of the two.
3. The applied pressure or water tension required to stop ice segregation depends, in general, on grain size distribution; the smaller the average grain size, the larger the pressure or tension.
4. Not all the water freezes at 0°C; the dependence of the unfrozen water content on temperature is determined by porosity, pore size distribution and nature of the material.
5. The freezing process occurs over a zone, the depth of which depends on the properties of the material, applied pressure, pore water tension and temperature gradient.
6. The location of the front of a growing ice lens may not coincide with the plane in which the ice first begins to form in the pores. Neither the ice front nor the freezing front need be at 0°C.

A model of the freezing process in porous materials must account for these observations. The six characteristics were clearly stated or implied in the combined contributions of Tabor (1930) and Beskow (1935).

THERMODYNAMIC MODEL

When Beskow stated in 1935 that a water film must exist between ice and soil particles for frost heaving to occur, a knowledgeable thermodynamicist probably could have stated the essence of the theoretical basis of the explanation of ice segregation that is now accepted. Fortunately for the experimentalist, the development and validation of a model usually requires a significant body of knowledge based on observation. This is certainly the case for ice segregation because of the variability of porous materials and the number of other independent factors that affect it. From the theoretical point of view, the greatest attention has been given to
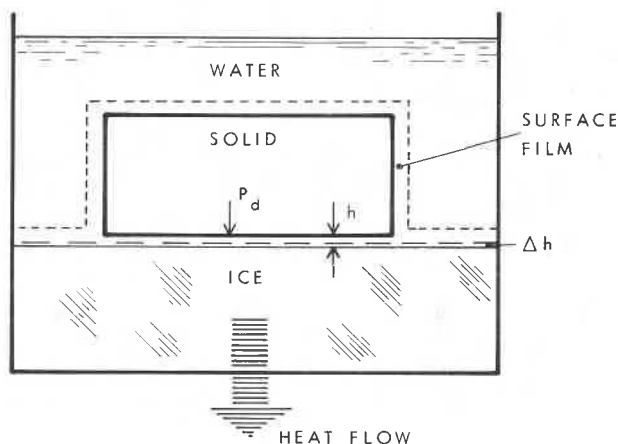
Fig. 5-6. Lifting of a solid with a plane ice-solid interface due to upward freezing of water.

describing the equilibrium condition for the water-porous solid system in terms of classical, steady-state thermodynamics.

Figure 5-6 is a simple representation of a flat solid resting on an ice surface in water that is freezing from the bottom upward. The ice is separated from the surface by a water film of thickness h. The thickness of this film is considered to be about 10^{-7} to 10^{-9}m (Hoekstra and Miller, 1967; Gilpin, 1978). A transition film also exists between the ice and water and the solid surface and water.

In principle, the thermodynamic properties of the ice, film and water can be determined. It is possible, therefore, to discuss the conditions governing equilibrium between them in terms of these properties. If the ice, film and water contain no impurities, a knowledge of the temperature and pressure is sufficient to define thermodynamic equilibrium.

Two phases are in equilibrium when their chemical potentials are equal (Guggenheim, 1950, p. 31). If they are not equal, there will be a transfer of material from the phase of higher chemical potential to the phase of lower chemical potential. For the single chemical species system under consideration, it is usual to use the Gibbs-Duhem relation for the dependence of change in chemical potential on change in pressure and temperature, i.e.,

$$d\mu = VdP - SdT \tag{1}$$

where V is the molar volume of the phase, S is the entropy per mole of

phase, $d\mu$ is the change in chemical potential for a change of pressure, dP, and of temperature, dT.

The phenomena of interest occur over a relatively narrow temperature range near 0°C, and it is reasonable to assume for discussion purposes that V and S are constants.

Integrating equation (1) gives

$$\mu = \mu_o + V\Delta P - S\Delta T \qquad (2)$$

where μ_o is the chemical potential for the pure phase at a reference state (e.g., 0°C and 1 atm pressure), and μ = chemical potential at temperatures and pressures that differ from their values at the reference state by ΔT and ΔP, respectively. The definition of equilibrium requires that μ_o for bulk ice be identical to that for bulk water, film and saturated water vapour. The temperature, pressure and other thermodynamic variables are relatively easy to measure for the bulk ice, water and vapour phases. This is not the case, however, for the film phase and often its properties must be implied indirectly.

It is useful to consider some of the observations on the behaviour of the film phase in order to develop an appreciation of its nature. It has been found for wettable materials that heat is evolved during the adsorption of an amount of water that would form a film about 7 to 10 molecules thick over all the surfaces; that is, a reaction occurs that binds the water to the surface.

A common characterization test for porous materials is to determine how their water content at a constant temperature depends on the pressure of the vapour with which they are in equilibrium. The change in chemical potential of water vapour, $\Delta\mu_v$, due to change in vapour pressure from the saturation value, P_o, to a lesser value, P, at constant temperature is given by

$$\Delta\mu_v = RT \ln(P/P_o) \qquad (3)$$

where R is the gas constant and T is the temperature in °K. As P is less than P_o, the right hand side of equation (3) is negative, indicating a decrease in chemical potential.

Decreasing the vapour pressure also results in a reduction of water content. This correlation shows that under conditions of constant temperature the chemical potential of the adsorbed water film decreases with decreasing thickness. For equilibrium between the film and vapour phases after a change in vapour pressure at constant temperature, equation (2) requires that there be a change in the film pressure equal to

$$P_1 = (RT/V_1) \ln (P/P_o) \qquad (4)$$

where V_1 = molar volume of the film fluid. Equation (4) implies that the decrease in film thickness is accompanied by a decrease in the fluid pressure in the film.

The film is the transition zone between the state of molecular order that exists at the solid-film interface and at the film-air interface, or in adjacent

bulk water. The attraction between water molecules and the solid surface causes a pressure to exist in the film normal to the film-surface interface. This intrinsic pressure is a body force similar to the force due to gravity, but acting over a much smaller distance (Gilpin, 1978). It has been termed the spreading pressure, and like the depth dependence of the pressure in fluids due to gravity its value at the film-solid interface increases with increasing thickness of film.

Spreading pressure is not a common concept in geotechnical practice. By implication it is assigned the value 0 when the film is in the reference state, i.e., when the chemical potential μ in equation (2) equals μ_o. This can be appreciated from a consideration of the following common test procedure. The water content of a soil specimen can be controlled by placing the specimen on a porous plate connected to a water reservoir. When the pressure of the water in the reservoir is such that the pore water pressure at the soil-porous plate interface is atmospheric, the specimen will be saturated, neglecting the effect of gravity, and its pore water in equilibrium with vapour at pressure P_o. The water content can be reduced to the value in equilibrium with vapour at pressure P by reducing the pressure of the water in the reservoir by the amount P_l in equation (4). As P_l is negative and water is withdrawn from the specimen, the process is interpreted as applying a suction. What has actually occurred within the specimen, however, is that the film pressure at the solid-water interface has been reduced by P_l.

If two surfaces with adsorbed water films are brought together, it is found that the films are not easily displaced and that a repulsive force is mobilized. This force has been called the disjoining pressure (Derjaguin and Melnikova, 1958). The disjoining pressure P_d is defined as the negative of the pressure given by equation (4) (Padday, 1970). It is the pressure that would have to be applied to a film in equilibrium with vapour at the saturated pressure P_o to reduce its thickness to that which would be associated with vapour at pressure P. It is another manifestation of the forces responsible for the spreading pressure. Sources of these forces are considered to be van der Waals' attraction, the electrical double layer and orientation effects.

A similar transition region must exist on the surface of ice, and evidence has been found for it at temperatures as low as -10°C (Fletcher, 1972). The very fact that ice lenses can grow in fine grained soils when subjected to pressures as high as 10^6 N m^{-2} shows that the films can mobilize very significant disjoining pressures and still remain sufficiently fluid to sustain the growth process. Clearly, the film phase must have the ability to transfer a compressive stress between ice and a solid surface while transporting water in the freezing zone.

Theoretical and experimental studies have shown that the thickness of

the film between ice and other solids depends on temperature, pressure between the ice and solid, and the pressure in the water (i.e., pore water pressure). As the film is in order of 10^{-8} m thick, it is clearly a most difficult task to establish the temperature and pressure dependence of the thickness other than in an indirect way.

Assume that heat flow from the ice-solid interface in figure 5-6 increases the thickness of the ice by Δh and, therefore, decreases the thickness of the film by the same amount. If the process occurs at constant temperature, there will be a corresponding decrease in the spreading pressure and in the chemical potential of the film. If the film is connected to adjacent films that have suffered no change, there will exist a difference in chemical potential and spreading pressures causing water to flow to the film and increase its thickness to the original value, displacing the solid upward.

If the solid is constrained from moving relative to the ice, the pressure must increase in the film. This results in an increase in its chemical potential toward the higher value of adjacent regions.

One of the controlling factors in the process is the resistance to the flow of water to the ice front. This resistance depends on the thickness of the film, the viscosity of water, the nature of the surfaces and the distance over which the water must flow. If the flow of water becomes a limiting condition, flow of heat away from the interface cannot be compensated by latent heat and the temperature of the interface must frop, thereby increasing the chemical potential and advancing the freezing zone.

Mackay and Burrous (1979) demonstrated that the ice segregation process can occur at relatively large flat surfaces. In one experiment they displaced a penny and a 2 cm³ polished granite cube upwards by an advancing ice front for a distance of some 2 to 3 cm, and stated that "many other solid objects were readily uplifted, particularly when the freezing rate was slow and the objects were raised on small supports in order to permit a thin layer of ice to form on the bottom before uplift commenced." They pointed out that this mode of frost action is of some importance during freeze-back of the active layer from below. Connell and Tombs (1971) measured directly a disjoining pressure of 20 kN m⁻² between an ice surface with the relatively large radius of curvature of 3 mm and a flat glass plate.

The shorter the distance water has to flow along the ice front the more readily can ice segregation be maintained. Freezing in porous solids offers such possibilities to reduce this flow distance. The porous structure also allows the mobilization of another important mechanism.

EFFECT OF CURVED INTERFACES

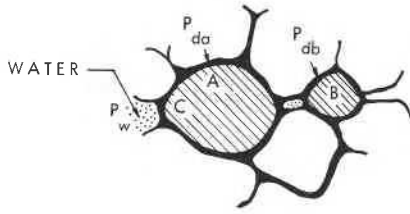One of the characteristics of porous solids is that if the ice phase is to

Fig. 5-7. Schematic representation of ice in pores; the heavy black lines represent the surface films. The effective radius of curvature of the ice at A = r_a, at B = r_b and at C = r_c.

grow it must propagate into and through pore spaces. Consider the ice-filled pores depicted in figure 5-7. Let the pore have an average radius of $\bar{r}_a$ at A and $\bar{r}_c$ at the opening C. Surface tension forces in the film on the curved surfaces cause pressure in adjacent ice. The dependence of this pressure, P_r, on the average radius of curvature, $\bar{r}_a$, of the ice surface at A is given by (Everett, 1961)

$$P_r = 2e_{iw}/\bar{r}_a \tag{5}$$

where e_{iw} is the surface energy of the ice-water interface, and P_r is positive for a concave ice surface. If the disjoining pressure between the ice and the adjacent solid surface is P_{da}, the total change in pressure from the reference state in the ice at the curved interface is

$$\Delta P_i = P_{da} + 2e_{iw}/\bar{r}_a \tag{6}$$

Substituting this value for ΔP in equation (2) gives for the chemical potential of the ice at the interface at A

$$\mu_{ia} = \mu_o + V_i(P_{da} + 2e_{iw}/\bar{r}_a) - S_i\Delta T_{ia} \tag{7}$$

If the pore water pressure at C is P_w, the pressure on the ice at C is

$$P_i = P_w + (2e_{iw}/\bar{r}_c) \tag{8}$$

The chemical potential of the water at C is

$$\mu_w = \mu_o + V_w P_w - S_w \Delta T_w \tag{9}$$

and of the ice

$$\mu_{ic} = \mu_o + V_i(P_w\ 2e_{iw}/\bar{r}_c) - S_i\Delta T_{ic} \tag{10}$$

For equilibrium $\mu_w = \mu_{ia} = \mu_{ic}$. Equating equations (7) and (10) and assuming $\Delta T_w = \Delta T_{ia} = \Delta T_{ic} = \Delta T$ gives

$$P_{da} + (2e_{iw}/\bar{r}_a) = P_w + (2e_{iw}/\bar{r}_c) \tag{11}$$

Equating equations (9) and (10) gives

$$V_i(P_w + 2e_{iw}/r_c) - V_w P_w = (S_i - S_w)\Delta T \tag{12}$$

For T small (e.g., less than -5°C) it is usually assumed that

$$S_i - S_w = - L/T_o \tag{13}$$

where L = latent heat of fusion for water; T_o = the melting point of bulk ice = 273.18 °K.

Therefore, at equilibrium

$$V_i(P_w + 2e_{iw}/\bar{r}_c) - V_w P_w = - L\Delta T/T_o \tag{14}$$

Equations (8), (11) and (14) define the conditions for which ice will propagate out of a pore of effective radius $\bar{r}_a$ through effective radius $\bar{r}_c$ (i.e., ice will propagate through all pore openings of average radius greater than or equal to $\bar{r}_c$). They indicate that a reciprocal relation should exist between the applied load and the pore water pressure, as observed in laboratory experiments.

Koopmans and Miller (1966) utilized equations (5) and (14) in a comparison of the pressure dependence of the water content of a saturated soil free of colloidal material (determined in a pressure plate apparatus) and the temperature dependence of the unfrozen water content in the same soil. Williams (1966) used equation (8) as the basis for a comparison of the air pressure required to initiate air intrusion into an unfrozen saturated soil and the conditions for the penetration of ice into pores at the freezing front for the same soil. The theoretical relation that applies when the mean curvatures of the respective air-water or ice-water interfaces are the same, i.e.,

$$(P_a - P_w)/(P_i - P_w) = e_{aw}/e_{iw} \tag{15}$$

was confirmed by experiment. In equation (15) P_a = air pressure applied in the pressure plate test or in the air entry test, P_i = pressure exerted by the ice, P_w = pore water pressure at the ice- or air-water interface, e_{aw} is the surface energy of the water-air interface. These investigations, particularly that of Koopmans and Miller (1966), demonstrate the equivalent effect of pore air pressure and pore ice on the water films with which they are in contact.
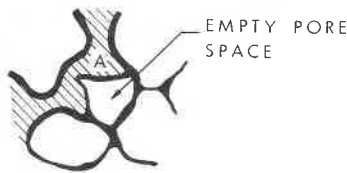
Fig. 5-8. Schematic representation of ice growing into a pore; the heavy black lines represent the surface films. If the temperature is sufficiently low, the pore will spontaneously fill with ice, even if it is not initially filled with water.

Consider figure 5-8, which portrays ice advancing into a pore. Miller (1973), and Bresler and Miller (1975) presented the conditions for which ice will exist in reentrants and similar cavities associated with pores. As the ice phase grows, a metastable condition is attained at which the pore will spontaneously fill with ice, whether it was previously filled with water (saturated) or not (unsaturated). This implies that a state is reached for which the chemical potential of ice in a filled pore is less than that in a

partially filled pore. Such a condition occurs when ice advancing from adjacent crevices coalesces or when the ice front from a single source has advanced to a particular position. In both cases the ice surface in the pore reaches a position at which, with further growth, its curvature decreases and may even reverse, as illustrated at A in figure 5-8, resulting in a decrease in P_r in equation (5) and a lowering of the chemical potential.

This possibility of spontaneous filling of space by ice has very significant implications for porous materials. As the zero degree isotherm surface advances through the material, the position of the freezing front will be at the temperature at which ice will propagate through the maximum size openings (pores or cracks). Behind this front pores will spontaneously fill with ice when their temperature reaches a value determined by the pore geometry and pore water pressure. It can be expected that if water is available to a porous material on the cold side of the 0°C isotherm, the material will become saturated with ice (Bresler and Miller, 1975). This agrees with field and laboratory observations of the ice-saturated state of soils in which water can flow to the freezing zone during the freezing process.

INITIATION OF ICE LENSES

A question of some interest is: what are the conditions necessary to initiate an ice lens? Consider again the schematic representation of pores in figure 5-7. Assume that in the thawed state the soil is subjected to an effective stress of $P_e = P - P_w$ where P is the total stress on a horizontal plane. For the purpose of this discussion, assume that ice is present in pores A and B having effective radii $\bar{r}_a$ and $\bar{r}_b$, respectively. Let $\bar{r}_a$ be larger than $\bar{r}_b$. If the pores are interconnected and in thermodynamic equilibrium at the same temperature, then from equation (7)

$$P_{da} + (2e_{iw}/\bar{r}_a) = P_{db} + (2e_{iw}/\bar{r}_b) \qquad (16)$$

where P_{da} and P_{db} are the disjoining pressures imposed on the ice through the film by the adjacent solid surfaces. P_{da} must be larger than P_{db}.

Consider now the film phase. The total pressure in the film, P_t, is the sum of the local intrinsic or spreading pressure, P_s, and the imposed or disjoining pressure, P_d, i.e.,

$$P_t = P_s + P_d \qquad (17)$$

Since the film at A is in equilibrium with the films at B and C (no flow), $P_{ta} = P_{tb} = P_{tc}$. For this to occur, any change in disjoining pressure must be compensated by an equal but opposite change in spreading pressure. It should be re-emphasized that the spreading pressure is felt only within the film.

The change from the reference state in the film pressure at C, where the disjoining pressure is zero, is equal to the pore water pressure P_w. For the

equilibrium condition, therefore, the change in the total film pressure from the standard state at A and B is also P_w and the equivalent of equation (12) is

$$V_i(P_d + 2e_{iw}/\overline{r}_c) - V_lP_w = (S_i - S_l)\Delta T \qquad (18)$$

If it is assumed that $(S_i - S_l) = -L_l/T$, where L_l is the latent heat released in the freezing of film water at temperature T°K, the equivalent of equation (14) is

$$V_i(P_d + 2e_{iw}/\overline{r}_c) - V_lP_w = -L_l\Delta T/T \qquad (19)$$

Equation (19) shows that for a surface of constant temperature the disjoining pressure that exists between the ice and adjacent solid surface for each pore is determined by the temperature, the effective radius of the pore, and the pore water pressures. At equilibrium, where there is no flow, the total pressure in the film must equal that of the films in the unfrozen zone at the freezing front corrected for the effect of gravity.

When the average of the disjoining pressures over a constant temperature surface equals the effective stress, P_e, the condition is established for separation and the initiation of a lens. Equation (19) shows that it depends on the pore water pressure and the applied or overburden pressure through their effect on the required value for P_d

This model for ice segregation has some interesting implications. Equation (19) indicates that the smaller the pore size the smaller is the disjoining pressure required for equilibrium and, therefore, the thicker is the film. This, along with the increasing surface area associated with decreasing pore size, should increase the possibilities for water to move through the freezing zone. Once a lens has been initiated and as long as permeability is not a limiting factor, it is found that the finer the pore system the more readily can water move to regions of the ice front where the local effective radius of curvature is large and, therefore, the disjoining pressure is large. A negative radius of curvature such as that over a soil particle would require an even larger disjoining pressure for equilibrium.

An extreme situation would be a flat surface with many fine holes over which essentially the full disjoining pressure of which the water film is capable can be mobilized. Such a situation could be approached in clays for which the plate-like particles have a preferred orientation. Assuming $\Delta T = -2°C$, $L_l = 335 \times 10^{-3}$ J kg^{-1}, $\overline{r} = \infty$, $P_w = 0$ (saturated soil), T = 270°K, and $V_i = 1.1 \times 10^{-3}$ m³ kg^{-1}, equation (19) gives $P_d = 2.26 \times 10^6$ N m^{-2} (330 psi). This is in the same range as the maximum heaving pressures observed for clay soils.

Increasing the pore water tension is a more severe limiting condition for ice segregation. In soils, the openings to the ice front vary in size. As the pore water tension increases, pores begin to empty of water and ice, beginning with the largest (Bresler and Miller, 1975). This restricts their ability to

deliver water to the ice front, if it exists, and to contribute to the total disjoining pressure.

For a growing ice lens the total film pressure at the ice front is less than that in the unfrozen water at the freezing front, and this pressure difference along with the permeability of the freezing zone will determine the rate of flow to the ice lens. The lens will continue to grow until the total film pressure at the ice front becomes equal to that in the unfrozen water through either an increase in the disjoining pressure or a decrease in the pore water pressure, or until a new lens is initiated at another location.

Of particular interest are porous materials such as porous rock or concrete that are too strong to allow an ice lens to be initiated. If water-saturated material is subjected at one face to freezing conditions, the absorbed water will experience a force that moves it to regions where ice has formed (e.g., on the surface or in large pores). Assume that a temperature gradient exists such that not too far from the ice front the material is saturated at 0°C and that initially there is space for ice to form without imposing a force on surrounding walls. As freezing in such fine pore materials may not begin until the temperature is in the range of -10°C, it is easy to imagine a situation for which ΔT in equation (19) could be -5°C or lower. A difference in film pressure of the order of 10^7 Nm^{-2} could conceivably be introduced between the ice front and the saturated region at 0°C. As the ice would come into contact with the surrounding cavity walls, this difference in spreading pressures would gradually be reduced by the development of disjoining pressures.

If cavities are spaced sufficiently closely and are large enough to provide storage for the water present in the solid, damaging spreading pressures can be dissipated and disjoining pressures avoided. This is the basis for air entrainment of concrete and a recent invention by Litvan and Sereda (1978) that involves the mixing of coarse pored paticles with normal aggregate.

NON-EQUILIBRIUM EFFECTS

A most difficult question to answer is whether thermodynamic equilibrium is ever established in the field or in the laboratory in porous ice-filled materials subjected to a gradient of temperature and pressure. Certainly experiments can be run in the laboratory for which dimensional changes due to ice freezing or thawing are either zero or less than the resolution of the measuring method. Such experiments are run usually for a period of days at the most and cannot show, for example, whether frozen soil and rock formations subjected to natural or imposed conditions will thaw or heave significantly or experience redistribution of ice over a period of years or centuries. This question is of great interest with respect to the stability of

the ground ice that has absorbed so much of Ross Mackay's attention, but it
has not been studied widely (Harlan, 1974).

Consider the relatively simple case of a spherical particle embedded in
ice (figure 5-9a). Assume that it and the ice are initially at a constant
negative temperature T_a, that a gradient is imposed such that the tempera-
ture at a remains the same, and that at b on the opposite side becomes $T_b <$
T_a. The chemical potential of the film at b is now greater than that of the
adjacent ice, and some of the water is converted to ice. If the temperature
change is maintained, according to equation (19) the pressure in the film at b
and the adjacent ice must increase to re-establish equilibrium. This pres-
sure is felt at a, resulting in the melting there of some of the ice. The net
effect is a transfer of water in the cold direction and of the particle in the
warm direction. Experiments by Hoekstra and Miller (1967) and by
Romkens and Miller (1973) have demonstrated that this indeed occurs. The
rate of movement is affected by several variables, amongst them the tem-
perature gradient, the nature, shape and size of the particle, and its average
temperature. Significant movement occurs only for small particles and
average temperature greater than -2°C.

A better known phenomenon is regelation (Nye, 1967; Drake and
Shreve, 1973). Consider a wire drawn through ice at constant temperature
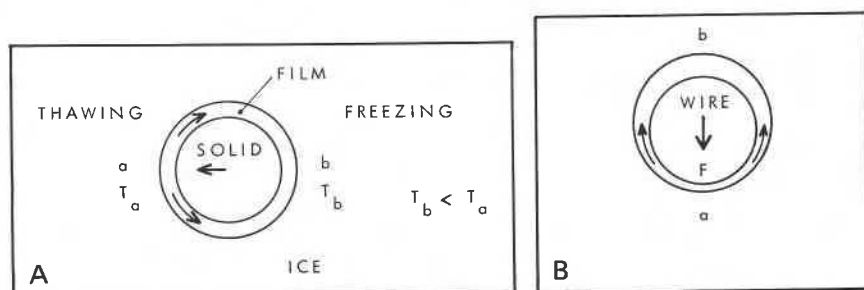as shown in figure 5-9b. Because of the pressure between the ice and the

Fig. 5-9. (a) Displacement of a solid particle by temperature induced flow of water in the film
from the warm to the cold side.
(b) Movement of a wire through ice by regelation due to a force, F. Ice melts at a, flows in the
film to b and refreezes.

wire the chemical potential of the ice at a is greater than that of the adjacent
water film, and melting occurs. The wire moves downward, reducing the
pressure at b, and the pressure gradient that has now been imposed on the
film forces water to the back of the wire. As the film at b now has a larger
chemical potential than the adjacent ice, water from it freezes. Factors that

control this effect are the type and size of wire, the applied pressure, the average temperature of the wire, the viscosity of water and the resistance to flow around the wire. Significant rates of regelation under laboratory conditions are also confined to temperatures close to the melting point. Although the relative rate of movement of ice and particles due to temperature gradients and regelation may be small, the total movement may be significant over geological periods of time.

Pressure gradients may also result in thermodynamic instability, tending to move ice from positions of high stress to positions of low stress, resulting in significant redistribution. Although this effect is of potential importance for many field situations, it has received little attention.

CONCLUSIONS

In this presentation I have tried to give an appreciation of the interesting behaviour of ice and water in porous materials. This behaviour is confined primarily to the temperature range of 0 to -1°C for coarse grained soils such as sand; 0 to -5°C for fine grained soils such as clay; and perhaps to temperatures as low as -10°C for rocks and similar fine pore materials. Below -10°C ice in porous materials is more inert, more rock-like in its behaviour, and from Ross Mackay's point of view probably less interesting except with respect to distribution and quantity.

The thickness of frozen ground in the temperature ranges of 0 to -1°C and 0 to -5°C depends on the temperature gradient. For deposits subjected to the geothermal gradient of 3×10^{-2}°C m^{-1} it can extend to over 150 m. These are relatively extensive thicknesses of thermodynamically sensitive deposits.

A major challenge now is to establish the dependence of the rate of segregation and of the redistribution of ice on the factors that determine them. Another interesting question is: if the combined overburden and pore water pressures are such as to not totally suppress ice segregation, can there ever be true thermodynamic equilibrium in the extensive thickness of ice-saturated soils that exists naturally? If not, what is the significance of this on a geologic time scale, particularly for deposits subjected to long-period temperature and stress changes? These are questions of some geologic and engineering interest and ones that Ross Mackay may wish to tackle.

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