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Chemical factors in soil freezing and frost heave

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ABSTRACT. Chemical factors that are essential in frost heaving of soils are examined through consideration of the process of ice formation in soils and the role of temperature gradients in generating water potential gradients in freezing soils. Unfrozen films are maintained around soil particles in frozen soils. The osmotic potentials at the ice–water interface of the unfrozen films and in the frozen fringe, the thin zone between the frozen and unfrozen soil, generated by dissolved salts and exchangeable cations that satisfy soil particle surface charge, are controlled by the local temperature. The coldest location and the most negative osmotic potentials at the ice–water interface are located immediately below the base of the ice lens, in the unfrozen films that separate the underlying soil particles from the ice lens. An osmotic potential gradient is generated because the osmotic potential at the water–ice interface in the frozen fringe becomes less negative with increasing temperature and distance from the ice lens. As water freezes onto the ice lens, re-supply of water to the unfrozen film along the osmotic potential gradient is the temperature-gradient-induced mechanism that generates the force that lifts the overlying frozen soil. Models that recognize this driving mechanism should improve predictions of soil freezing and frost heave, analysis of contaminant transport in freezing and frozen soils, and other aspects of the soil-freezing and frost-heave processes.

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Introduction

Frost heave is the upward or outward movement of the ground surface caused by the formation of ice in the soil (National Research Council of Canada 1988). The essential conditions for the occurrence of frost heave in soils are sub-freezing temperatures, the presence of water, and a frost-heave-susceptible soil. Frost heave has been studied since the early 1900s (for example, Taber 1930; Beskow 1935), and a large amount of knowledge has been gained as to the damage caused, the processes involved, and ways to alleviate resulting problems. Macroscopic mechanistic models of soil freezing have been proposed using heat and mass transfer approaches to address and model the hydraulic and thermodynamic aspects of soil freezing and the origin of the water potential gradients (see literature reviews by Konrad 1994 and Henry 2000). Chemical factors in the frost-heave process are largely neglected. In this paper, the essential roles of chemical factors in the frost-heaving process are discussed through the development of model scenarios that emphasize these roles. A conceptual model of the driving force of the frost-heave process is proposed with support provided by theoretical arguments and experimental observations reported in the literature on frost heave (especially numerous papers by R.D. Miller, with various colleagues, and the osmotic model of frost heave proposed by Horiguchi (1987)), and some by deductive reasoning. Finally, the probable significance of these factors for various aspects of frost heaving is briefly addressed.

Soil freezing

A soil that is undergoing downward freezing can be divided into three zones (Fig. 1): 1) A frozen zone in which segregated pure or almost pure ice (commonly in layers or lenses perpendicular to the direction of heat extraction) and pore ice are observed. While containing small amounts of unfrozen water as films around soil particles, this zone contains a continuous ice matrix and behaves as an ice-cemented material. Its lower boundary is the base of the warmest ice lens. 2) An unfrozen zone into which the freezing front, the most forward position of ice in the soil, has not yet advanced. 3) A thin, partially frozen zone, called the ‘frozen fringe’ (Miller 1972), which is located between the frozen and unfrozen zones.

Ice formation on the soil surface

First the freezing process from the time of imposition of a sub-freezing temperature at the surface of a thin layer of water overlying a saturated, unfrozen soil is examined. The assumption of saturation avoids some complications

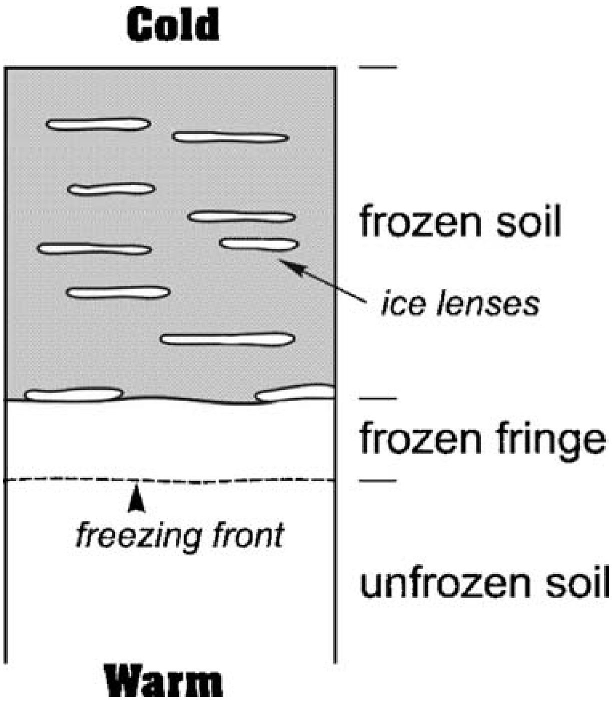

Fig. 1. Schematic diagram of the frozen, frozen fringe, and unfrozen zones within a freezing soil.

in the discussion, and applies for many (but probably not most) situations of soil freezing. Imposition of a sub-freezing temperature at the surface of the thin layer of water establishes a temperature gradient along which heat is extracted. Until ice formation commences, this heat removal from the saturated, unfrozen soil occurs mostly by conduction. When the rate of heat conduction to the surface becomes inadequate to maintain the film of water at a temperature above its freezing-point, ice formation becomes possible. Initiation of ice formation requires the development of ice nuclei. Pure water at atmospheric pressure is, by definition, in thermodynamic equilibrium with ice at 0°C. While this implies a freezing-point of 0°C, a small degree of supercooling normally occurs before ice nuclei form. Once ice nuclei form, freezing occurs spontaneously and continues until the release of latent heat increases the temperature of the ice–water system to 0°C. If the water contains dissolved material or is confined in pore spaces, the temperature rises to the point at which ice and water are in equilibrium. Supercooling is not essential to the freezing process; a snowflake falling on a water surface at the instant that the water temperature reaches its freezing-point would initiate freezing without supercooling.

Initiation of ice formation on top of the soil surface does not mean that ice will also form in the soil pore space. At 0°C, ice is not thermodynamically stable in the pores of a soil, or other porous medium. Two factors depress the freezing-point of water within soil, and together determine the sub-zero °C temperature at which ice penetrates into and through soil pores. The first is instability of ice nuclei in small spaces. Ice nuclei do not form at 0°C in small spaces because the surface energy associated with their small radius of curvature decreases their free energy (that is, it increases the water pressure on the inside of the curved surface). For ice to be stable within the pore, the temperature must be below the normal freezing-point of water by an amount that is dependent on the diameter of the pore. The smaller the pore, the greater the depression of the freezing-point (Koopmans and Miller 1966).

The second factor is that soluble salts in the pore water and the counter-ions that satisfy the surface charges of the soil particles lead to an osmotic depression of the freezing-point. It should be noted that even if no dissolved salts were present in the soil water, the exchangeable cations satisfying the surface charge of soil particles would depress the freezing-point adjacent to the particle surfaces. This occurs regardless of soil mineralogy because, effectively, all common soil minerals bear a net surface charge. The charge consists of the permanent charge sites associated with isomorphous substitution within the mineral structure (as is the case with the 2:1 clay minerals), and a pH-dependent charge arising from dissociation of H^+ from OH groups on particle edges or surfaces (in some clay minerals and hydrous oxides) and from unsatisfied bonds at the surfaces of mineral particles (for all soil minerals). This net surface charge is negative for most minerals at the pH common in soils (Yariv and Cross 1979). Even when ice has formed in the centre of the pore, the exchangeable cations that satisfy the electrically charged soil particle surfaces maintain an osmotic depression of the freezing-point of the water immediately adjacent to the soil particle surfaces and maintain an unfrozen film of water between soil particles and pore ice. The consequence of osmotic freezing-point depression, augmented by adsorption forces between the water and hydrophilic locations on the mineral particle surfaces, is that throughout the frozen soil and frozen fringe, ice and soil particles are never in direct contact. This unfrozen water film decreases in thickness as the temperature at the specific location decreases, but water and ice coexist in thermodynamic equilibrium in the frozen soil, even at temperatures many degrees below the freezing-point of pure, free water. This coexistence of water and ice in freezing soil and the presence of unfrozen films throughout the frozen soil, and especially at the base of ice lenses, are essential to the process of frost heave.

Growth of convex ice projections into soil pores

Figure 2 shows the situation in which ice has formed at the soil surface but not yet penetrated through soil pores even though heat continues to be extracted from the soil–ice–water system along the temperature gradient. If the rate of heat extraction exceeds the rate of heat delivery to the soil surface, the temperature of the ice, water, and soil decreases and ice commences to grow into the soil pores. Convex ice projections develop downward from the lower boundary of the ice body into the pores, with the radius of curvature of the projections being

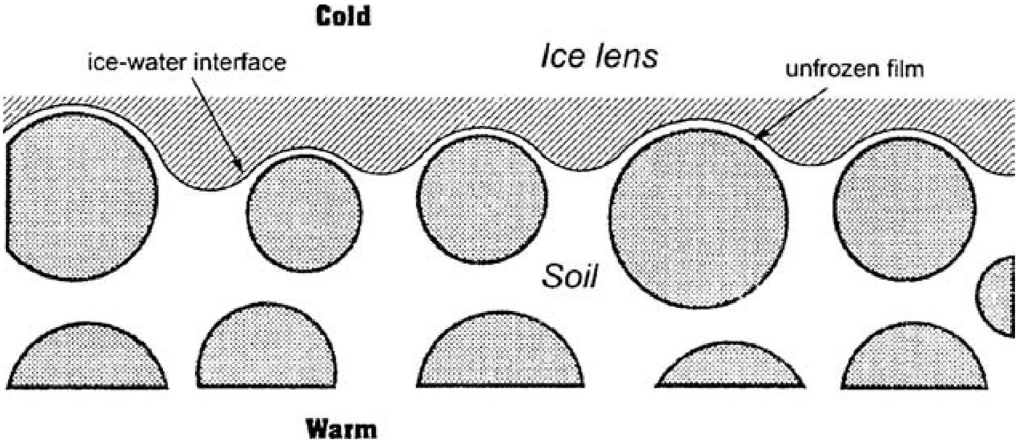

Fig. 2. Schematic diagram of an ice lens base in/on soil that lacks a frozen fringe.

consistent with that of the ice nucleus that would be stable at the pore temperature. If this curvature radius is greater than the pore radius, the ice surface does not grow completely through the pore. Rather, the ice surface over the pore reaches its limiting curvature for the temperature present, and the temperature at the tip of the ice projection over the pore will be warmer than that at the base of the ice lens over the particle. The temperature at each position along the ice–water interface determines the osmotic concentration of the water at each position. The colder the location, the more concentrated will be the water adjacent to the ice. Consequently, the temperature difference along this interface above the particle induces an osmotic pressure gradient that attracts water into and along the film above the particle to replace the water that was frozen. It must be emphasized that the freezing of water onto the ice lens lifts the ice and soil above only by the 9% that the volume of ice exceeds the volume of the water that froze. Freezing of re-supplied water to the unfrozen film in response to the osmotic pressure increase associated with salt exclusion when the water froze to form the new ice is responsible for the remaining 91% of the heave of the overlying ice and soil. This temperature-gradient-induced osmotic pressure gradient within the unfrozen film between the soil particle and the overlying ice body is fundamental to the process of frost heave. Without water re-supply to the film, the soil particles will be incorporated into the ice body.

The extra overlying load added by the growth of the ice lens is borne by the mineral skeleton of the soil, as opposed to the short-term increase of the local pore-water pressure as occurs with load application that eventually induces consolidation of fine-grained soils under drained conditions. Under continued heat extraction, water continues to be supplied to and freezes onto the ice surface that lies above the soil particles by flow through the osmotically maintained unfrozen films between the ice lens and the soil particles (Miller and others 1960). As the ice body is lifted, ice will in turn freeze onto the convex ice projection over the pore to maintain its position relative to the soil particles. The uplift force is not related to the temperature (and associated pressure across the ice–water interface) at which an ice protrusion can enter the pore space. This was the fundamental error in the capillary model of frost heave, upon which the first ideas of a frost heave ‘shut-off pressure’ (Arvidson and Morgenstern 1977; Hill and Morgenstern 1977) were based (Williams 1986). In the case that the sensible heat transfer by the arriving water, the heat transfer by conduction, and the latent heat release upon freezing exactly satisfy the heat extraction along the temperature gradient, the thickness of the ice on the soil surface will increase but the temperature at the base of the ice will remain unchanged. If the heat transfer associated with the water supplied from below is not sufficient to satisfy the heat extraction rate, or if the unfrozen film cannot conduct water fast enough to re-supply the portion of the ice surface above the soil particles, the temperature at the base of the ice body will decrease until the depression of the freezing-point of the water in pores will be satisfied and ice grows through pore necks into the underlying pore space.

The situation of no penetration of ice into the soil pores can occur in nature at least briefly during ice formation on the soil surface, and probably briefly below some parts of a newly initiated ice lens. An analogous scenario probably applies to water supply required for the formation of needle ice at the surface of wet, frost-heave-susceptible soil during a sustained and severe frost. It is not appropriate to call the development of needle ice ‘soil frost heave’ because the elevation of the soil surface may not change. In nature, it is common for needle ice to lift a few, or a thin layer of, soil particles (C.R. Burn, 2002, personal communication).

Ice penetration through the soil pores

Figure 3 shows the situation where growth of ice has occurred through soil pore necks below an ice lens when the temperature in the pore neck satisfies the depression of the freezing-point for that pore neck, including any osmotically induced freezing-point depression.

Water in the underlying pore may then freeze, if that water happens to be supercooled, or after enough heat

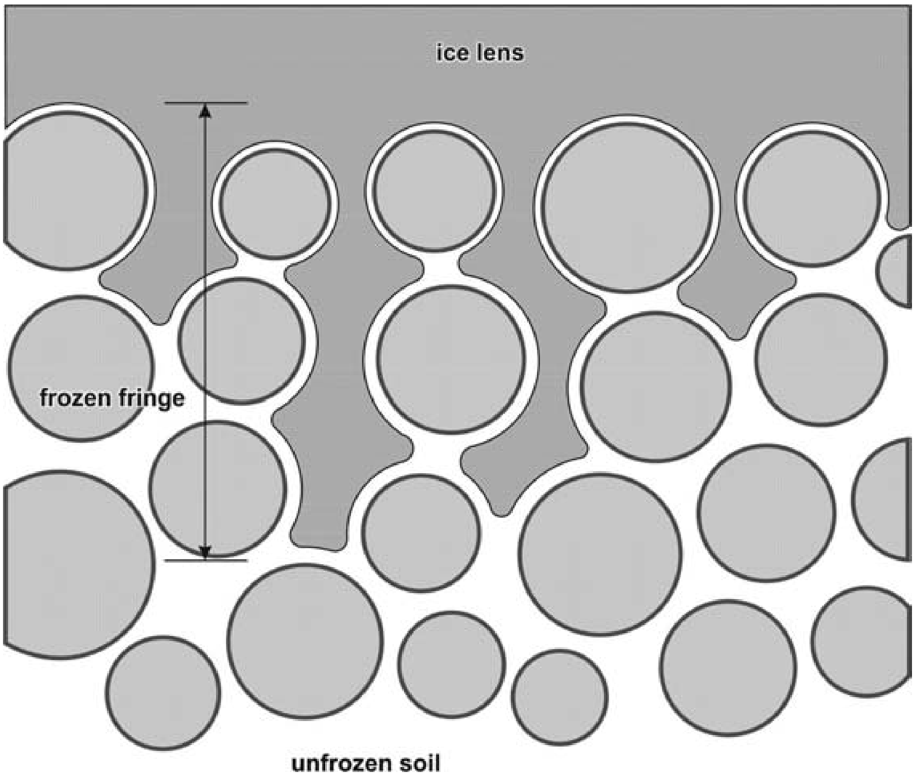

Fig. 3. Schematic diagram of the frozen fringe underlying an ice lens within a freezing soil.

is extracted if it is not supercooled. Ice formation in pores that are not connected to an ice-containing pore, through an ice-containing pore neck, will require some supercooling below the depression of the freezing-point of the water in that pore for an ice nucleus to be initiated. Once ice has penetrated through a specific pore or has been initiated in a pore, continued heat extraction causes ice to propagate through the adjacent pore space until pore necks are reached that have a lower freezing-point than the temperature at that location. The result is that ice 'fingers' through the larger pore spaces, with horizontal linkages to adjacent fingers through some pore necks, and by-passes small pores in which ice would be unstable at the temperature present. This fingering process creates an imperfectly interconnected network of ice-filled pores and leaves a network of ice-free pores. The ice-free pores will become ice-filled when their temperature decreases appropriately. This zone that contains the ice-filled pore network constitutes the 'frozen fringe' and extends a short distance (up to 14 mm, but generally much less (Akagawa 1988)) downward to the freezing front.

Water flow through the frozen fringe

The frozen fringe may be visualized as a zone in which the natural discontinuities of pore sizes in soils maintain a mix of ice-containing pore sequences extending downward from an ice lens and ice-free pore sequences extending upward from the freezing front. For any horizontal cross-section in this zone, the proportion of ice-containing pores is normally the greatest near the ice lens, where the temperature is the lowest, and decreases with increasing distance from the ice lens. The three-dimensional surface connecting the points of farthest downward advance of ice fingers in pore sequences below an ice lens can be a very irregular surface. Where ice-free pore sequences remain in this zone, they serve as preferred conduits (relative to the ice-containing pores) for water movement through the fringe to feed continuing growth of ice lenses. The frozen fringe is not perceived to be as ice dominated as the 'frozen permeameter' model proposed by Miller and others (1975).

The strength of the adsorption forces between mineral surfaces and water molecules is little affected by small temperature differences and is almost identical throughout the frozen fringe. Adsorption potential gradients over the frozen fringe will be insignificant. In contrast, the osmotic forces are strongly affected by small temperature differences, and a considerable osmotic potential gradient exists across the frozen fringe (Miller and others 1960; Römkens and Miller 1973; Gilpin 1979).

The ultimate driving force for water flow to and through frozen soil is the temperature gradient. A decrease in the temperature at the base of an ice lens freezes new ice onto the base of the ice lens above the particle, reduces the thickness of the adjacent water film thereby increasing the film's osmotic concentration, which responds by drawing water from water films in adjacent pore spaces to restore its thickness. The temperature gradient also maintains the osmotic (and water) potential gradient downward through the frozen fringe that, in turn, maintains the

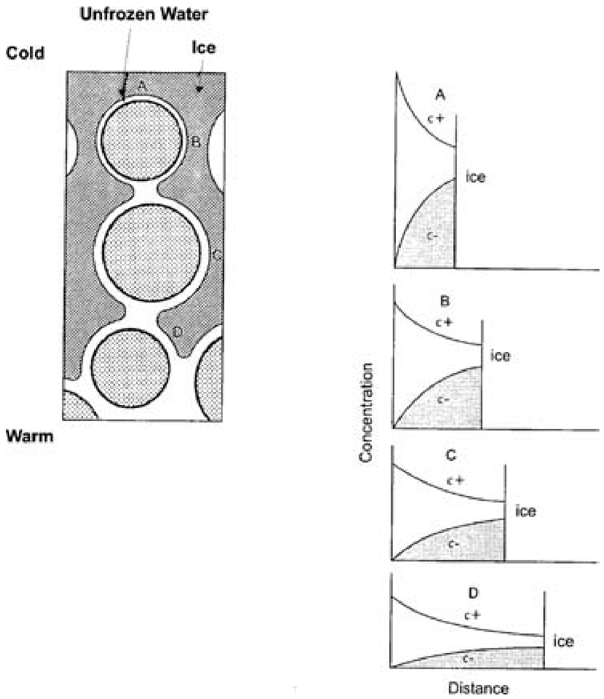

Fig. 4. Left side: Schematic diagram of a portion of the frozen fringe from Figure 3. Right side: Graphs showing the relationship of the thickness of the unfrozen water films, the cation, and anion concentrations in the unfrozen water films, plotted versus distance from the particle surfaces, and the relative solution concentrations at the film/ice interface, for locations A, B, C, and D in the frozen fringe (left side diagram).

required upward water flow through the unfrozen films from the underlying unfrozen soil.

In Figure 4, a microscopic/molecular-level view of the situation at the base of the ice lens and through the frozen fringe is presented. The parts of the ice lens over the pore space can be ignored in considerations of ice-lens growth, because the underlying pores are occupied by ice, and hence there is no ice–water interface at this location onto which ice can form to lift the lens upward. For the ice lens to grow, water must be supplied to the unfrozen films above the particles immediately below the ice lens. If water is not supplied to these unfrozen films, the particles will be incorporated into the ice lens. Such incorporation of soil particles into the base of the ice lens, over its full lower surface, defines the end of growth of that lens as the warmest ice lens, and a new ice lens may be initiated at a lower location. The limiting condition to avoid particle incorporation is that enough water must enter the unfrozen films above the particles to meet the needs of ice lens growth. The reader is cautioned that Figure 4 is a two-dimensional sketch, and that in three-dimensional reality, water flow is entering this zone from all sides of the particle (that is, water supply is easier in 3-D than is implied by a 2-D representation). The driving force for this water transport to the base of the ice lens is the thermally induced, osmotic potential gradient.

In Figure 4, it is shown that the coldest point in the unfrozen film surrounding the particle at the base of the ice lens is at A, the location in the film that is most remote from the unfrozen water source. The unfrozen film is at its thinnest at this location (Fig. 4A) and its interface with the ice is at its most osmotically concentrated, because the osmotic concentration needed to prevent freezing is determined by the temperature. Consequently, the osmotic concentration of excluded salts and counter-ions at the ice–film contact decreases from here towards the slightly warmer portions (for example, location B, Figure 4B) of the particle adjacent to the pores. Water flow from B toward A (and all points in between), in response to the osmotic potential gradient, feeds the ice lens growth. The concentration gradient will not dissipate as long as the temperature gradient remains. The flow of water into the film brings in salts by advective flow and inhibits diffusion to the particle periphery of salt ions excluded from the ice upon freezing. Furthermore, the counter-ions are constrained to remain in the unfrozen films by the need to satisfy the cation exchange sites on the soil particle surfaces. In a similar manner, the unfrozen film thickness between ice fingers and particles increases and the osmotic potential at the film–ice interface decreases downward through the frozen fringe (Fig. 4B, 4C, 4D) because the temperature increases as the distance from the ice lens increases.

The freezing of water onto the ice lenses and the replacement of the freezing water by water flow from the underlying frozen fringe sets up a water potential gradient along which water flows into and through the frozen fringe. This flow, in turn, induces a more-negative matric potential in the underlying unfrozen soil. Some consolidation of the unfrozen soil below the freezing front may occur in low-hydraulic-conductivity cohesive soils in response to the matric potential decrease during freezing of the soil above. Such consolidation would reduce the pore size and the hydraulic conductivity in the affected zone.

As suggested earlier, water flow will be easier through unfrozen pore sequences than through ice-containing pore sequences. In those portions of the frozen fringe where ice-containing pores dominate, the local temperature determines the osmotic concentration in the unfrozen water films, which means that the counter-ion plus dissolved-salt concentrations determine the water film thickness. The water entering these films in response to the water potential gradient is passing through to the base of the ice lens where it freezes. The local thickness or the local osmotic concentration of the unfrozen films in the frozen fringe changes only if the local temperature changes. Salt exclusion from the water that enters the fringe from the unfrozen soil does not occur exclusively at the base of the ice lens. Two factors lead to some salt exclusion at the freezing front: 1) the replacement of the ends of the ice fingers as the pore ice is ‘dragged up’ (as outlined below) through the pore space by the growing ice lens; and 2) the water passing through a restricted volume (the unfrozen film) is influenced by the negative surface charge of the particle, which leads to anion exclusion (along with their accompanying cations). Regardless, the maximum solution concentrations and the most-negative osmotic potentials are in the unfrozen films at the base of the ice lens where the temperature is coldest.

In summary, ice formation in the pores of the frozen fringe provides, at most, only a minor amount of uplift that is mainly associated with the volume expansion upon freezing of water. While an ice lens is growing and has a frozen fringe, the formation of pore ice that occurs is on the tips of the ice fingers at the freezing front. If this formation of pore-ice at the freezing front was the source of the uplift force, it would also push up the soil particles in the frozen fringe, thereby leading to their incorporation into the ice lens, with the consequence that growth of that ice lens would cease. Because ice formation above the particles at the base of the ice lens and osmotically induced re-supply of water to these films lift the lens, the upward moving ice lens will place the attached ice fingers into tension and act to drag up the pore ice of the frozen fringe. In order for the pore ice not to drag the soil particles with it, there must be pressure and osmotic melting of ice at locations on the film–ice interface that lie below a soil particle (such as the surface below C, Fig. 4), with the melted water being transferred to points above the particle where it freezes onto the ice that overlies the particle (such as the surface above C, Fig. 4) in order to maintain the appropriate film thickness for the local temperature. This process is analogous to regelation flow (Miller and others 1975; Miller 1978). With upward dragging of the pore ice of the frozen fringe, a small amount of new ice will form on the tips of the ice fingers at the freezing front to replace that pulled up.

Eventually, water (with its sensible heat and heat release upon freezing) may not be supplied to the base of the ice lens sufficiently rapidly to satisfy the heat extraction rate and the soil below the lens will cool sufficiently that, in turn, a new ice lens is initiated at a lower position. The described sequence of events continues and a series of ice lenses, separated by zones of soil dominated by ice-containing pores, but without segregated ice, forms.

Initiation of new ice lens

The conditions under which growth of a new ice lens is initiated below an existing ice lens have yet to be adequately resolved. This is dominantly a mechanical problem in which the strength of the soil along the plane in which the ice lens is initiated must be overcome. Miller (1978), Taylor and Luthin (1978), Gilpin (1979), Konrad and Morgenstern (1980), and Wood (1985), among others, addressed this issue. While the mechanical requirements for initiation of an ice lens are not addressed in this paper, it is useful to explore whether chemical factors, as discussed here, provide useful insights.

The conditions for continued growth of an ice lens are that the heat delivered to the base of the ice lens by conduction, sensible heat transport, and latent heat release must satisfy the heat removal requirements for maintenance of a constant temperature at the ice lens base. If heat transfer associated with water flow to the lens is inadequate to offset heat extraction, the temperature at the base of the ice lens will decrease, and the temperature decrease will propagate downwards through the frozen fringe. When the temperature decrease reaches the freezing front, the front will advance through accessible pores into unfrozen soil. Within the frozen fringe, the unfrozen film thickness in ice-containing pores will decrease and ice will propagate into accessible pore spaces that were previously ice-free. The net result is a substantial decrease in the ability of the frozen fringe to transmit water to the ice lens and a greater proportion of the latent heat removal will be associated with advance of the freezing front.

Because of the essentially constant temperature conditions within the frozen fringe of a 'well-fed' ice lens, a new ice lens will not be initiated before the currently growing lens has inadequate water supply and the temperatures in the frozen fringe decrease. Formation of a new ice lens within those portions of the frozen fringe and underlying unfrozen soil that have experienced significant consolidation, as a consequence of induced decreases in the matrix potential, appears to be improbable because of the smaller pores and the increased strength of consolidated soil, both of which would inhibit initiation of the ice lens. Ice-lens initiation should 'leap-frog' beyond such consolidated zones to less-dense underlying soil.

This leap-frogging should occur particularly in clays in which, under rapid freezing conditions, a thin consolidated zone would form quickly under a growing ice lens and would have the effect of 'starving' the lens. The consequence would be formation of a series of thin, closely spaced ice lenses. At lower heat-extraction rates, which require a lower rate of water supply, the consolidated zones could become thicker, and ice lenses would be thicker and more widely spaced; this dependence of lens spacing on heat-extraction rate has been observed in bench-scale experiments (Penner and Walton 1979) and a large-scale field experiment (Smith 1992).

This analysis of the formation of new ice lens in clays has not invoked a chemical explanation. While not specifically addressing mechanical issues, the following consideration on the dynamics of soil freezing as developed in this paper may be a factor in the initiation of a new ice lens. Namely, the upward tension on the ice fingers in the frozen fringe may lift some soil particles and open up horizontal space into which pore ice can develop a continuous sheet of ice. Where this lateral growth is sufficiently extensive, it may serve as the initiation location of a new ice lens. Natural inhomogeneity in the soil pore-size distribution could also be a significant factor in the initiation location of a new ice lens. Most probably the two factors work together, with the lateral extension occurring in zones of lower density (larger pores) and weaker soil.

Continued growth of ice lenses within the frozen soil

Increase in thickness of ice lenses in the frozen soil above the warmest ice lens has been observed (Goto and Takahashi 1982; Smith and Patterson 1989; Smith 1992). These lenses are connected to the freezing front by the zones of frozen soil, with pore ice and unfrozen films, that occur between successive lenses (and by the frozen fringe of the warmest ice lens). Water-flow from warm ice lenses to cooler ice lenses can occur, because a temperature gradient and its induced osmotic-potential gradient are present in the frozen soil between lenses and because unfrozen films of water surround all the soil particles. These films, which are thinner than in the frozen fringe of the warmest ice lens, provide the path for water flow (with an appropriate amount of regelation flow). The colder lens will grow (albeit slowly) at the expense of the upper surface of the next warmer ice lens. The base of this warmer lens will grow at the expense of the upper surface of the next warmer lens. With the unfrozen films becoming thinner as the soil becomes colder, this process decreases in importance the colder the location at which it is occurring. Over long periods this process could still increase the separation of points in already-frozen ground (Smith and Patterson 1989).

Frost heave

The essential conditions for frost heave are sub-freezing temperatures, the presence of water, and a frost-heave-susceptible soil. The sub-freezing temperatures normally arise by natural cooling of the soil during winter, and may be a result of human-emplaced freezing sources, such as buried gas storage and transmission structures or thermo-syphons. The presence of water is required for formation of the ice. Rates of soil freezing and the nature of the frozen soil resulting are influenced, among other factors, by the rate of heat extraction, which in turn is controlled by the surface boundary temperature, the thermal gradient, and the heat and moisture transmission properties of the soil system.

The major factor accounting for frost heave is the formation of ice lenses within the soil. These lenses represent water that has been drawn into the freezing portion of the soil from unfrozen soil, such that if the frozen soil is thawed it has a water content much higher than in its original state of saturation. The portion of the frozen soil between lenses remains saturated, although some consolidation of this zone may have occurred before it was overtaken by the advancing ice front. Because of the volume increase of water when it is frozen (Buchan 1996), the ice lenses occupy 9% more space than did the water that migrated to form them. The extent to which formation of ice in the pores of the frozen fringe may increase the volume of the fringe is unknown (and undoubtedly inconsistent), but in many cases the volume increase associated with the phase transformation in the pores of the fringe is expected to be at least partially accommodated by movement of water to the growing ice lens — in which case the water equivalent within the frozen fringe would be less than before the ice formed within it.

The heaving of the soil surface may be categorized as derived from three components, all of which commonly occur simultaneously:

- Type 1 heave: increase of volume of the soil system due to the expansion of water when it freezes in situ.
- Type 2 heave: increase of volume of the soil system related to inflow and freezing of water from the unfrozen soil through the frozen fringe to the warmest ice lens.
- Type 3 heave: increase of volume of the soil system related to inflow and freezing of water from movement of water from warmer ice lenses to colder ice lenses within the frozen ground, if the flow is converging (Konrad 1994: Fig. 6). Type 3 heave, if flow is not convergent, should only increase separation of points on opposite sides of the growing ice lens.

Lifting of the soil surface requires the development of a lifting force, commonly expressed as the ‘frost heaving pressure,’ capable of lifting the overlying soil, which dominantly originates from the osmotic gradient that re-supplies water to the unfrozen films at the base of the ice lens.

Susceptibility to frost heave

A frost-heave-susceptible soil is one in which the pore space is such that ice segregation to form ice lenses can occur (Chamberlain 1981). Not all soils are susceptible to frost heaving. Clean gravels and pure sands normally do not experience the process, but small amounts of silt and clay are enough to make them frost susceptible; medium-textured soils are extremely susceptible and can heave large amounts; and fine-textured clay soils are very susceptible, but the amount of heave is usually limited. Susceptibility to frost heaving is related to the soil’s ability to support growth of ice lenses, as determined by the ability of ice to propagate into pore space, the ability to supply water to the portion of the ice lens situated above soil particles, and the ability to transmit water through the unfrozen soil to the freezing front.

Coarse-grained soils, such as gravels and clean sands, are not frost-heave susceptible because their large pores exhibit only a very small depression of the freezing-point, and because of the very long distances (relative to unfrozen-film thickness) that water must travel through the unfrozen films above the particles to feed ice-lens growth. Even Type 1 heave is unimportant (except in a completely closed system where water cannot escape) because it is easier for water to flow away from the freezing front to accommodate the volume change than to lift the soil above.

In medium-textured soils, such as silts, the conditions discussed above are satisfied. Substantial depression of the freezing-point inhibits ice growth into the soil pores and limits free propagation of ice fingering into the soil pore space; the unfrozen films at the base of the ice lens are able to transmit water around to the top of soil particles to support ice-lens growth; and the frozen fringe and unfrozen soil are capable of supplying adequate water for ice-lens growth. Considerable frost heaving can result.

A lesser amount of frost heave in clays is a result of the lesser ability of the clay to supply water through the frozen fringe and unfrozen clay. The very low hydraulic conductivity in the clay also allows the development of substantial matrix potential gradients within the unfrozen soil. These gradients cause consolidation of the clay immediately adjacent to the freezing front, decreasing both the pore size and its already low hydraulic conductivity and increasing its depression of the freezing-point for ice entry into pores. These combined effects inhibit ice propagation through the pores. As soon as water flow to the ice lens no longer satisfies the heat removal requirements, the underlying soil cools until ice propagates through some pores or ice formation is initiated in unfrozen soil and creates the conditions for formation of ice lenses at a new location. The ice lenses formed in a clay soil are usually thinner and more closely spaced than those in a medium-textured soil (Konrad 1994) and add up to a lesser total thickness. Frost heave can be substantial in clays, but, for similar thermal conditions, is normally of lesser magnitude than in silts and other medium-textured soils.

Small amounts of silt and clay in sandy soils greatly increase their frost susceptibility. Concentrations of fine particles at pore constrictions can set up freezing-point depression and water supply conditions that favour growth of ice lenses. The fine particles can be accumulated at appropriate sites during freezing of dirty sand by their exclusion from ice at the freezing front as the ice propagates through the pores, or by having been filtered from water flowing through the sand. Chemical factors play only an indirect role in considerations of frost susceptibility.

Osmotic conditions and frost-heave pressures

The argument has been made that the force that lifts an ice lens and the overlying soil is temperature-induced, osmotic-gradient-driven re-supply of water to the ice–unfrozen film interface at the base of the warmest ice lens as water is removed from the film by freezing onto the base of the ice lens. In turn, the osmotic gradient between this unfrozen film and the freezing front is the major component of the temperature-gradient-induced, water-potential gradient that generates water flow through the frozen fringe to the warmest ice lens.

This mechanism is a more satisfying explanation for the origin of heaving pressures than the capillary theory in which it was posited that frost-heaving pressures originated with the pressure difference across the curved ice-water interface above the pores at the base of the ice lens. While a pressure difference that is related to the curvature of the ice in the pore does exist across this interface, with the water pressure being lower than the ice pressure (Gold 1957; Miller 1963; Williams 1967), it is pushing downward against water, not against the skeleton of the soil. The measurement of frost-heaving pressures in excess of what is possible based on these considerations of pore size (Loch and Miller 1975; Takashi and others 1981) has long ago disproved the capillary theory argument. The existence of a frozen fringe below an ice lens also negates the pore-size argument for the upper magnitude of heaving pressures. The proposed osmotic argument is affected by neither of these problems; it both pushes against the soil skeleton, and it provides the water potential difference for water flow through the frozen fringe to ice lenses, and through already frozen soil to colder ice lenses. The pore-size-related temperature depression is nonetheless important in that it delays the progress of the freezing front into the soil below an ice lens and limits the thickness that the frozen fringe can attain.

According to the osmotic model, the maximum heaving pressure will be determined by the difference between the water potential in the unfrozen film at the base of the growing ice lens, and the water potential at the water source in the unfrozen soil. The osmotic potential in the unfrozen film is, as argued, determined by the temperature at that location, and the total potential at the location will be slightly lower (as affected by overburden pressure — usually a relatively minor factor). The water potential at the water source in the underlying soil (also affected by osmotic, matric, and gravitational potentials) serves as the reference potential.

In assessments of frost heave, the temperature, potential, and pressure differences are usually calculated between the base of the warmest ice lens and the freezing front. While the temperature at the base of the warmest ice lens is probably the main factor governing frost-heaving pressures in short-term freezing situations, the temperatures at the bases of colder lenses are important in the Type 3 heave that takes place in long-term freezing conditions, such as frost heave around a buried pipeline operated for many years at sub-freezing temperatures. This latter statement is consistent with the observation by Takashi and others (1981) that the maximum observed heaving pressure in long-term experiments depended on the temperature at the cold end of the specimen, where the ice lens formed in their experiments. Horiguchi (1987) calculated that the osmotic pressure in the unfrozen film between the ice lens and the soil particles at this location is equal to the observed maximum heaving pressure. The upper limit on heaving pressure observed by Takashi and others (1981), at temperatures below –25°C, probably represents the limit of film thickness capable of transmitting water to supply further growth.

It should be noted that hydrodynamic models that do not rely on knowledge of the origin of the driving forces are favoured currently for the predictive modelling of soil frost heave in North America (for example, Konrad and Morgenstern 1980, 1981; Nixon 1991; Guymon and others 1993). These empirical models, based on considerations of heat and water flow, implicitly acknowledge the presence of appropriate driving forces for water flow to the location of the growth of ice lenses, but express them in terms of the presence of 'free energy' differences. They do not require knowledge of the source(s) of the driving forces, other than that they are related to the existence of a temperature gradient. The argument that the pressure difference driving water flow through the water films at the base of the ice lens and through the freezing fringe is the osmotic pressure gradient provides the basis for modelling flow within these zones. Horiguchi (1987) has presented the elements of an osmotic model for frost heaving that is based on the principles elaborated in this paper. However, his model did not include a criterion for formation of new ice lenses.

Summary and conclusions

This analysis of the role of chemical factors in soil freezing and frost heave is derived from accepted principles in chemistry and soil science. In particular, it links together knowledge of osmotic solution properties, the charge properties of soil particle surfaces, and the distribution in the adjacent solution of the adsorbed cations that balance this charge. Osmotic re-supply of the unfrozen films between an ice lens base and the underlying soil particles has been identified as the source of the force that lifts the overlying material to generate frost heave, with the osmotic pressure being controlled by the local temperature.

The temperature-induced osmotic potential gradients in the unfrozen films between ice and soil particles are also the major driving force for water flow through the frozen fringe to the ice lens. Because the osmotic potential of brine solutions (which the unfrozen films are) increases dramatically as they are concentrated by ice formation in response to temperature decrease, the osmotic potential gradients associated with small temperature differences are very large compared to normal pressure gradients in saturated soil. While the empirical models of frost heave based on considerations of heat and water flow have proven to yield satisfactory predictions for most considerations, we argue that the more complete understanding of the role of chemical factors in freezing and frozen soil that this paper presents removes some mystery from what occurs during soil freezing. The fuller understanding of the process of soil freezing provided, should prove useful to analysis, among other things, of the influence of contaminants on the properties of frozen soils.

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