Processes

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Introduction

Seasonally frozen soil strongly influences runoff and erosion on large areas of land around the world. In many areas, rain or snowmelt on seasonally frozen soil is the single leading cause of severe runoff and erosion events. As soils freeze, ice blocks the soil pores, greatly diminishing the permeability of the soil. This is aggravated by the tendency of water to migrate to the freezing front, causing elevated ice content and frost heave.

Soil freezing and thawing also play a role in a variety of other environmental processes. Frost heave poses significant problems for structures, roads, and plant roots. Soil freezing and thawing can create stress fractures and alter soil physical properties, including pore continuity and aggregate stability; these alterations can influence soil hydraulic properties and erodibility long after the soil is thawed. Water migration associated with soil freezing can strongly influence solute movement. Artificial freezing of the soil has been used to create a barrier in order to isolate contaminants within the soil.

Freezing and thawing of the soil are controlled by the complex interactions of heat and water transfer at the soil surface governed by meteorological and environmental conditions at the soil–atmosphere interface. Different types of frost may form, depending on soil moisture content, rate of freezing, ground cover, and soil physical characteristics. Soil permeability, erodibility, and frost heave depend largely on the type of frost formed.

Heat and Water Relations During Freezing and Thawing

Due to negative water potentials, soil water exists in equilibrium with ice at temperatures below the normal freezing point of bulk water and over the entire range of soil-freezing temperatures normally encountered. When ice is present in the soil, the soil matric potential is strongly influenced by the temperature. As temperature at the freezing front decreases, more and more water freezes, water potential becomes more negative, and liquid water content continues to drop, creating a gradient in water potential and liquid water content. This drop in liquid water content at the freezing front has a similar effect to drying of the soil, and water migrates from moist regions to the freezing front. This often results in elevated ice content, ice lenses, and frost heave.

When ice is present, soil water potential is a function of temperature. This relation is expressed by the Clausius–Clapeyron equation as:

$$\phi = \pi + \psi_{\mathrm{m}} = L_{\mathrm{f}}\left(\frac{T - T_{\mathrm{frz}}}{T}\right) \qquad [1]$$

where ϕ is total water potential, π is soil water osmotic potential, ψ_{m} is soil matric potential, T is absolute temperature, T_{frz} is the freezing point of bulk water (typically 0°C or 273.16 K), and L_{f} is energy, termed the latent heat of fusion, required to freeze water. Thus, when ice is present in the soil, heat and water flux through the soil are tightly coupled, i.e., the matric potential and therefore liquid water content are defined by the temperature and osmotic potential. The relation between matric potential and liquid water content defined by the moisture-release curve is typically assumed valid for frozen conditions.

Darcy's equation can be used to describe steady-state, one-dimensional water flux through the soil:

$$q_{\mathrm{l}} = -K\left(\frac{\partial(\psi_m + \psi_{\mathrm{g}})}{\partial z}\right) \qquad [2]$$

where K is the unsaturated conductivity, ψ_{m} is soil water matric potential, ψ_{g} is gravitational potential, z is depth within the soil, and $\partial(\psi_{\mathrm{m}} + \psi_{\mathrm{g}})/\partial_z$ is the gradient in soil water potential. The transient mass balance equation for water, including the effects of freezing and thawing within the soil, can be written as:

$$\frac{\partial\theta_{\mathrm{l}}}{\partial t} + \frac{\rho_{\mathrm{i}}}{\rho_{\mathrm{l}}}\frac{\partial\theta_{\mathrm{i}}}{\partial t} = -\frac{\partial q_{\mathrm{l}}}{\partial z} + U \qquad [3]$$

Terms on the left-hand side of the equation represent: the time rate of change of liquid water content, and the time rate of change of ice content. Terms on the right-hand side are: the gradient in water flux (i.e., the net flux of water into a layer of soil), and a source/sink term for water. Here, θ_{i} and θ_{l} are volumetric ice and water content, and ρ_{i} and ρ_{l} are the density of ice and water. Eqn [3] states that the net liquid water flux into a soil layer must equal the combined change in ice and water content of the soil. When the net flux is equal to zero, any change in liquid water content must be offset by a change in ice content, adjusted for the difference in density. Although this change in density can result in expansion of the soil matrix, it is not the primary cause of frost heave.

Neglecting vapor flux terms, the equation used for describing one-dimensional heat flow within a partially frozen soil is:

$$C_s \frac{\partial T}{\partial t} - \rho_i L_f \frac{\partial \theta_i}{\partial t} = \frac{\partial}{\partial z}\left[k_s \frac{\partial T}{\partial z}\right] + S \qquad [4]$$

The terms in eqn [4] represent: energy stored in the soil due to a temperature increase; latent heat required to freeze water; net thermal conduction into a layer; and a source term for heat added to the soil. Here, C_s is the volumetric heat capacity of the soil, $\partial T/\partial t$ is the time rate of change of temperature, L_f is the latent heat of fusion, and k_s is soil thermal conductivity. This equation states that the net heat transfer into a volume of soil by thermal conduction and source terms is offset by a change in temperature and a change in ice content of the soil.

Water Migration and Frost Heave

From eqn [1], as the temperature drops below freezing and ice begins to form, the water potential becomes more negative. This creates a gradient in water potential and causes moisture movement toward the freezing zone. Water movement to the freezing zone is described by eqn [2]. If water movement to the freezing zone is sufficient, ice lenses occur, causing the soil matrix to expand. In the case of vertical frost penetration from the soil surface, ice lenses can cause the soil to heave upwards if the pressure associated with freezing exceeds overburden pressures.

The extent of water migration and ice accumulation in the freezing zone is controlled primarily by the rate of freezing front advance in relation to the unsaturated hydraulic conductivity. When the soil is frozen rapidly, there is little opportunity for water to migrate to the freezing front and soil water is essentially frozen in place. Similarly, if the unsaturated conductivity is low, water migration to the freezing front will be slow. Very dry and/or coarse-textured soils have relatively low unsaturated conductivities and exhibit much less frost-related water movement and frost heave than moist, fine-textured soils. Excessive water migration and frost heave are experienced most often when a very moist, fine-textured soil is frozen relatively slowly.

Freezing Dynamics

Soil temperature and water dynamics during soil freezing are shown for a silt loam soil in **Figure 1**. Continuous measurement of near-surface soil ice content is problematic; however, liquid water content during soil freezing can be measured quite accurately with time domain reflectometry (TDR). Therefore, for illustrative purposes, total water (liquid plus ice) and liquid water content plotted in **Figure 1** were simulated by the simultaneous heat and water

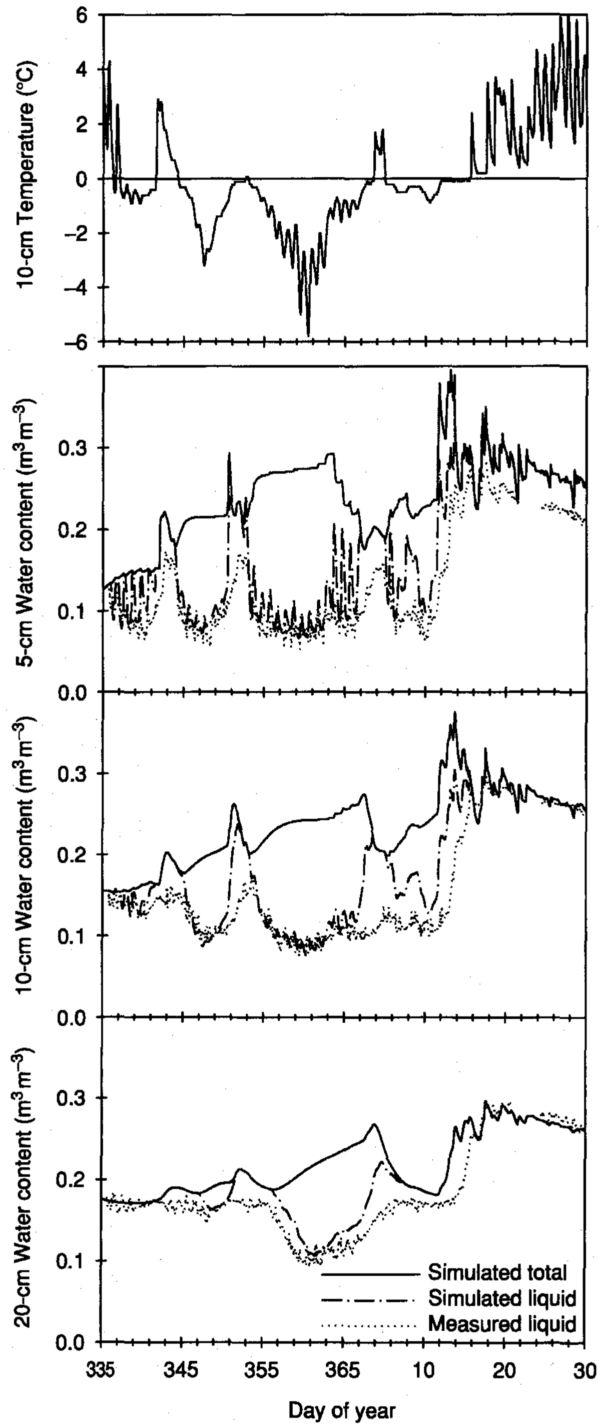

Figure 1 Simulated 10-cm soil temperature, as well as total water content and simulated and measured liquid water content of a silt loam soil for the 5-,10-, and 20-cm depths. (Adapted from Flerchinger GN (2002) Coupled soil heat and water movement. In: *Encyclopedia of Soil Science*. New York: Marcel Dekker, Inc., with permission.)

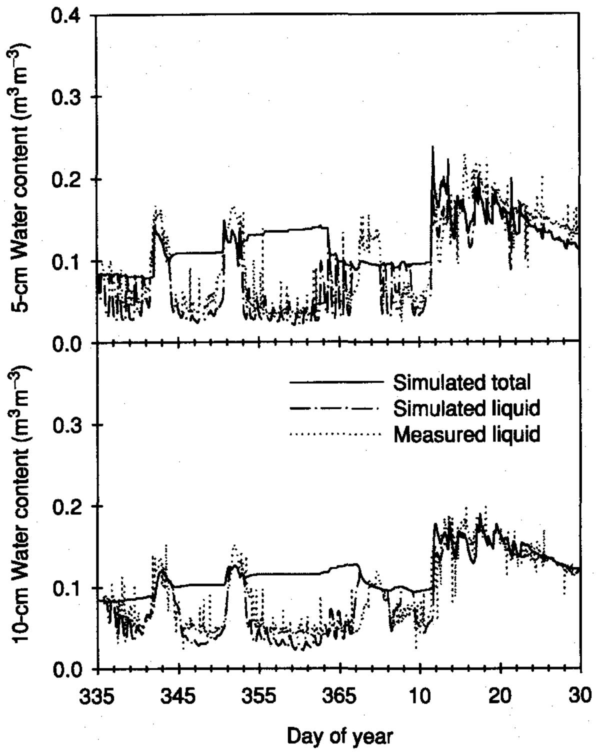

Figure 2 Simulated total water content and simulated and measured liquid water content for a loamy sand soil for the 5- and 10-cm depths. (Adapted from Flerchinger GN (2002) Coupled soil heat and water movement. In: *Encyclopedia of Soil Science*. New York: Marcel Dekker, Inc., with permission.)

(SHAW) model; measurements of liquid water content are plotted for comparison.

The 10-cm soil temperature plotted in **Figure 1** shows several freeze–thaw cycles. Frozen conditions can be observed from the water content plotted for the 5-, 10-, and 20-cm depths by separation of the simulated liquid water content line from the total water content line; the difference between the two lines is ice content. Accumulation of ice and an increase in total water content due to water migration to the freezing front can be observed at all three depths plotted. After 9 mm of rain and snowmelt on days 341 and 342, simulated water content above 20 cm was decreasing on day 343 due to drainage. After initiation of soil freezing on day 344, direction of flow reversed, and simulated water flow above 10 cm was upward toward the freezing front. Total water content of the 5-cm depth began to increase on day 344, while liquid water content continued to decrease. As the frost front advanced and the 10-cm depth began to freeze, water migration into the 5-cm soil layer ceased, and the 10-cm total water content began to increase. Subsequently, the 20-cm depth began to freeze on day 347.

Soil water content is shown in **Figure 2** for a loamy sand soil, which is a much more coarsely textured soil than the silt loam. Soil water dynamics for the loamy sand are considerably less responsive to freeze–thaw processes than the silt loam soil. Due to the low unsaturated conductivity of the loamy sand, there is much less moisture migration to the freezing front than for the silt loam. As a result, increase in total water content is much smaller.

Freeze–Thaw Impacts on Infiltration

Rain and/or rapid snowmelt on impermeably frozen soil is the leading cause of severe flooding and erosion in many areas of the world. Soil freezing can dramatically reduce the soil's infiltration capacity. Ice blocks the soil pores, resulting in large runoff events from otherwise mild rainfall or snowmelt events.

The permeability of frozen soil is affected by the occurrence, depth, and ice content of the soil, which is dependent on the interrelated processes of heat and water transfer at the soil surface and within the soil profile. Soil freezing and thawing can also alter soil physical properties or structure that impact infiltration. Changes in aggregate stability or stress fractures caused by freezing affect soil structure and pore continuity and thus affect infiltration even after the soil is thawed.

The type of frost formed influences soil permeability after freezing. Soil frost may be divided into four types: granular, honeycomb, stalactite, and concrete. Granular frost is usually found in woodland soils containing organic matter. It consists of small frost crystals, which aggregate around soil particles, but remain separate from each other. Honeycomb frost is commonly found in highly aggregated organic soils and has a loose porous structure, which resembles a honeycomb. Both granular and honeycomb frost typically have high infiltration rates. Stalactite frost often forms in bare soil, which is saturated at the surface. This type of frost consists of loosely fused columnar ice crystals and absorbs water rapidly because of its open porous structure. Concrete frost usually forms in bare, fine-textured, agricultural soils where upward migration of moisture is significant. It is characterized by a complex formation of many thin ice lenses and leaves the ground very hard, much like concrete. Depending on water content, concrete frost can be almost impermeable.

Frozen soil infiltration rates decrease dramatically with soil water content. Water is held less tightly in the large pores, therefore the largest pores that contain water at the time of freezing are the first to freeze. When these larger pores, which conduct water more readily, are blocked by ice, the tortuosity of flow paths increases and permeability and infiltration are severely reduced. Infiltration into a frozen, relatively dry silt loam can typically decrease from 1.0 cm h^{-1} to 0.01 cm h^{-1} as the water content increases to near saturation.

Ice lenses that form due to water migration from the unfrozen subsoil toward the freezing front have an added effect on infiltration. These lenses, typically formed in wet soils having a high proportion of silt, are often a barrier to infiltration. Melting of ice lenses during infiltration influences the temporal variation in infiltration. If ice lenses melt during an infiltration event, infiltration rate can nearly return to unfrozen infiltration rates.

Tillage and surface characteristics that alter porosity and water and heat transport processes can influence infiltration. Tillage processes that create macropores typically have a positive effect on infiltration. Unless the soil is extremely wet, macropores are not filled with water at the time of freezing and remain open for infiltration. However, tillage is likely to have little effect on infiltration if the freezing front descends below the depth of tillage.

Knowledge of frozen-soil infiltration processes lags considerably behind nonfrozen processes. Accurate quantitative descriptions or algorithms of frozen soil infiltration are lacking, partly due to experimental difficulties in measuring infiltration into frozen soil and characterizing the ice content and structure within the frozen soil. With the exception of expensive laboratory techniques such as nuclear magnetic resonance (NMR), there are no quantitative means of directly measuring ice content and structure in the soil, which is the single most important factor affecting infiltration potential upon freezing. Ice content can be computed as the residual between liquid water content measured by TDR and total water content measured by neutron probe or gravimetric samples; however, the sampling volumes of these techniques are dramatically different, making accurate measurement of ice content difficult.

Our understanding of frozen-soil infiltration processes is further hampered by the fact that ice content, pore blockage, and infiltration rate change as water infiltrates into frozen soil. Introducing water into frozen soil causes freezing of the infiltrating water, thawing of the ice contained within the soil, or both. Thus, there is no steady-state infiltration rate analogous to that in unfrozen soil. An approach to circumvent this problem is to use an alternate fluid that remains viscous at subfreezing temperatures. Fluids such as ethylene glycol and air have been used as test fluids for characterizing infiltration of frozen soils. Measured permeability for these alternate fluids can be related to hydraulic conductivities by accounting for differences in density and viscosity.

Various approaches exist for estimating infiltration of frozen soils. Depending on the level of sophistication, adjustments for frozen conditions may be based on: simply whether the soil temperature is below freezing; the amount of ice present in the soil; or the available porosity remaining in the frozen soil. Very simple approaches use essentially a simple on/off switch for accounting for frozen-soil effects, in which the curve number or hydraulic conductivity is set to an arbitrary value to cause reduced infiltration when the soil is frozen. Slightly more sophisticated methods use an adjustment factor to hydraulic conductivity based on antecedent water content or ice content of the soil. Many detailed approaches for estimating infiltration in frozen soils assume the hydraulic conductivity and water retention characteristics are the same for frozen and unfrozen soils. Thus, hydraulic conductivity for infiltration is based on the unsaturated hydraulic conductivity computed from the available porosity (total porosity less volumetric ice content).

Freeze–Thaw Impacts on Soil Erodibility

As soil freezes, water migration to the freezing front can cause ice lenses to form. When the ice lenses melt, the soil often cannot reabsorb all of the excess water, particularly if an impermeably frozen layer still exists below the thawed layer. This supersaturated state results in soil that is extremely weak and susceptible to erosion. However, after drainage and consolidation, soil strength returns. Thus, partially thawed and thawed but unconsolidated soil is highly susceptible to erosion.

Shear strength of a soil is indicative of its resistance to erosion. Specifically, it is defined as the resistance to deformation by the action of tangential (shear) stress. Soil shear strength is made up of cohesion between particles and resistance of particles sliding over each other due to friction or interlocking. Cohesion is composed of true cohesion and apparent cohesion. True cohesion is a function of soil mineralogy and results from chemical bonds between particles. Apparent cohesion, however, is determined by water tension within the soil and is strongly influenced by water content. As the soil thaws at a high water content, soil strength due to apparent cohesion is nil.

Shear stress, τ, caused by water flowing over the soil surface, is defined as:

$$\tau = \gamma RS \qquad [5]$$

where R is the hydraulic radius of the flow, γ is density of water, and S is the slope of the channel of surface. The minimum amount of shear stress required to initiate particle movement is termed the critical shear stress, τ_{cr}, which is a measure of the soil's shear strength. Soil detachment is computed as:

$$D = \alpha(\tau - \tau_{cr})^{\beta} \quad [6]$$

where α is referred to as soil erodibility and β is a fitted exponent. Studies on thawing soils have shown that moisture content has some effect on τ_{cr}, but the major effect is on α, the soil erodibility. The erodibility of a soil thawed at 5-cm water tension can be approx. 14 times that of the same soil thawed at 45 cm of tension. However, the thawed soil can regain its strength upon drying after tensile forces related to water tension are restored. This strength can be restored in a matter of hours if an impermeable soil layer thaws and allows drainage to occur or if strong evaporative conditions exist. Thus, shear strength can fluctuate dramatically, and timing of rainfall or snowmelt during the first hours of thawing can make a big difference to the erosion that occurs.

Freeze–Thaw Impacts on Aggregate Stability

Aggregate stability, a measure of an aggregate's resistance to breakdown when subjected to external forces, is an important soil property, because soil susceptibility to water and wind erosion increases as aggregate stability decreases, in general. Moreover, many soil physical and hydraulic properties, such as surface-sealing rate, infiltration rate, and hydraulic conductivity, are influenced by aggregate stability. In addition, on medium-textured soils with unstable surface aggregates, crusts can form that hinder or, in some cases, prevent the emergence of seedlings of sown crops.

Mode of Action

As the soil temperature drops below freezing, ice crystals form in the soil matrix, forming first in the soil pores. Once a crystal forms, water flows to the crystal due to a potential gradient, enlarging it, which exerts pressure on nearby aggregates. If those aggregates are constrained and cannot move away from the expanding ice crystal, the pressure exerted upon them can fracture the aggregates directly or develop planes of weakness that can, upon subsequent wet-sieving, cause the aggregates to fracture.

Factors Affecting the Impact

Antecedent water content is the greatest single factor that determines how an aggregate responds to freezing. Aggregate stability decreases, often linearly, with increasing water content at freezing. This relationship appears to hold for many soils ranging in texture from sandy loams to silty clays. As soil water contents increase, more water is available to form ice crystals or ice lenses. Moreover, wetter soil has more water-filled pore space and thicker water films surrounding soil particles, increasing the area through which unsaturated water flow can occur. Also, in wetter soil, tortuosity is less, thereby shortening the water's flow path and speeding its movement to the ice crystal, increasing the latter's rate of expansion. To preserve aggregation and reduce erosion in temperate regions, it is often recommended that producers minimize autumn soil water contents near the soil surface, whenever possible.

The rate at which soil freezes determines in large measure the impact of freezing on soil structure, principally by affecting water redistribution within the soil. If the upper, moist horizons of a soil freeze quickly, that is, if the air temperature decreases sharply, water in those horizons is essentially frozen in place. Consequently, ice lenses, even if they form, do not thicken appreciably and do not compress nearby soil. On the other hand, if a relatively wet soil freezes slowly, water moves to an ice lens and freezes there, thickening the lens, and causing structural deterioration and frost heave.

Soil texture and organic matter also influence aggregate response to freezing. Soils containing high proportions of sand are easily weakened or even fractured. On the other hand, soils with a lot of clay are better able to withstand pressures exerted by nearby ice-crystal enlargement, probably because of additional bond strength provided by more or stronger clay bridges that form between silt and/or sand particles within the aggregate. Organic matter, known to increase the stability of unfrozen aggregates with diameters greater than 0.25 mm, also strengthens aggregates that are later subjected to freezing stresses, providing the water content at freezing is not too great. Elasticity provided by organic matter may enable aggregates from medium-textured (or finer) soils frozen at relatively low water contents to withstand ice-lens expansion pressures before fracturing. Available data indicate that organic matter contents of 3% or more are particularly beneficial for soils that are medium-textured or finer.

The number of freeze–thaw cycles that an aggregate experiences also determines how stable the aggregate is after freezing, with responses being somewhat soil-dependent. In the past, aggregate stability was thought to decrease as freeze–thaw cycles increased, beginning with the first freeze–thaw cycle and continuing monotonically thereafter. Most early research subjected initially air-dried aggregates to many freeze–thaw cycles, often five to ten or more. Recent research, however, has demonstrated that aggregates that have not been air-dried between

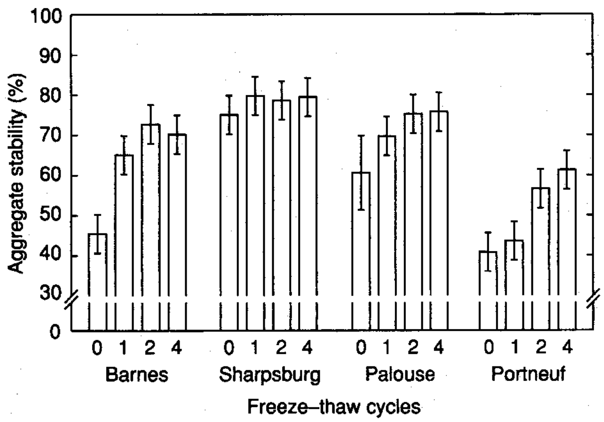

Figure 3 Aggregate stability of four soil types measured after freeze–thaw cycles. (Bars represent 95% confidence intervals.) (Reproduced from Lehrsch GA (1998) Freeze–thaw cycles increase near-surface aggregate stability. *Soil Science* 163: 63–70.)

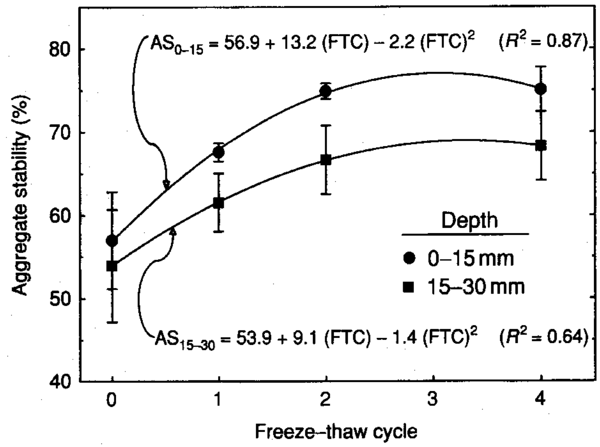

Figure 4 Effects of freeze–thaw cycles on aggregate stability at different soil depths. (Bars represent 95% confidence interval; AS denotes aggregate stability; FTC denotes number of freeze–thaw cycles.) (Reproduced from Lehrsch GA (1998) Freeze–thaw cycles increase near-surface aggregate stability. *Soil Science* 163: 63–70.)

sampling and analysis often increase in stability with the first two or three freeze–thaw cycles (**Figure 3**). The imposition of many freeze–thaw cycles does indeed decrease aggregate stability; what was seldom recognized was that just a few cycles could increase the stability of aggregates from medium- and fine-textured soils. It has been postulated that ice formation in interaggregate pores and the initiation and early enlargement of ice lenses increase particle-to-particle contacts. Migration of water to the enlarging lens then dries the soil matrix surrounding the ice lens, positioning polysaccharides on soil particle surfaces, gathering and arranging clay domains at points of contact between soil particles, and/or precipitating slightly soluble bonding agents such as $CaCO_3$, silica, or iron oxides at contact points. These processes help aggregates reform and increase in strength, after thawing.

It is particularly interesting that aggregate stability often increases with the first few freeze–thaw cycles but then decreases as more and more freeze–thaw cycles accrue (**Figure 4**). Opposing forces may be responsible for such phenomena. A force serving to strengthen aggregates may be the result of the precipitation of slightly soluble bonding agents at points of contact between soil particles during the first few cycles. An opposing force that weakens aggregates may be due to ice-lens formation, compression of nearby aggregates, and development of fracture planes. The strengthening process may well occur, and be dominant, for the first two or three freeze–thaw cycles, until most bonding agents have been precipitated from the soil solution. As freeze–thaw cycles continue to accrue, however, more and more fracture planes may be formed in aggregates near where the ice lenses formed. This persistent weakening process may then begin to play the dominant role, decreasing aggregate stability after two or three cycles, as suggested by the fitted curves in **Figure 4** and commonly reported in the literature.

Aggregates that are constrained from moving about either in a sample or in a soil profile are weakened more by freezing than are unconstrained aggregates. Aggregates below the soil surface are less stable after freezing than aggregates at or near the soil surface, regardless of the number of freeze–thaw cycles to which they are subjected (**Figure 4**). Freezing and thawing processes thus affect aggregate stability and soil structure, in addition to water and heat flow through soil.

Soil Freezing and Thawing Effects on Solute Migration

Solutes directly affect water movement during freezing by altering potential gradients. Solutes in the soil solution add an osmotic component to the total water potential. This additional component affects soil water redistribution within the profile as a soil freezes at a given temperature. Specifically, frost heave is greater, that is, more and thicker ice lenses are formed where the soil solute concentration is low, rather than high.

As previously noted, temperature gradients cause soil water to migrate to the freezing front. Solutes in the soil water are carried to the freezing front via convection. As water freezes at the front, solutes are excluded and accumulate in the unfrozen water contained in thin films around soil particles, within soil pores, or in relatively concentrated brine pockets, which may be entrapped within the ice lens itself. If the concentration of a solute in the unfrozen soil solution exceeds its solubility limit, that solute may be precipitated at particle-to-particle contact points, potentially increasing aggregate stability. Upon thawing, the solute transport to the freezing front via mass flow often results in increased concentrations of soluble constituents in the soil water near where the ice lens had formed.

The concentrated brine can also migrate through the frozen soil under certain climatic conditions. When air temperatures at the soil surface are very low, a temperature gradient induces these trapped brine pockets to migrate downward through the ice and frozen soil, into deeper but warmer portions of the soil profile.

See also: **Energy Balance**; **Polar Soils**; **Swelling and Shrinking**

Further Reading

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FUNGI

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Introduction

Fungi are ubiquitous in soils, but comprise a varying proportion of the microbial biomass in different systems. They tend to dominate in soils of high organic matter and constitute a smaller proportion in intensively managed mineral soils. They are involved in a plethora of functional roles, encompassing biological, chemical, and physical interactions, and are of great ecological and economic significance. The study of fungi is termed mycology (after *mukēs*, Greek for 'fungus').

Classification

The 'true fungi' belong to the kingdom of Fungi and are eukaryotic, exclusively heterotrophic organisms with cell walls that contain chitin or chitosan as a major constituent, and typically have a thread-like hyphal growth-form, although unicellular forms are common in the yeasts. There are four phyla, the Chytridiomycota, the Zygomycota, the Ascomycota, and the Basidiomycota, plus an informal group denoted the mitosporic fungi (formerly the Fungi Imperfecti or Deuteromycota), which lack a sexual phase and are not ascribed to a formal taxonomic position. Only the Chytridiomycetes produce motile zoospores. Representatives from all taxonomic groupings are commonly found in soils, where fungal biodiversity,