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Frost Depth

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Freezing and thawing of soil is a common occurrence throughout the world. Indeed, approximately 50% of the Earth landmass is frozen at some time during the annual cycle, with 20% of the land underlain by permafrost (Sharratt et al., 1997). Seasonal freezing of soils with sparse vegetation and snow cover can occur to depths of 3.5 m (Kennedy & Sharratt, 1997; Shul'gin, 1965) while seasonal frost has been found to penetrate to depths of >6 m below paved runways (Carlson & Kersten, 1953). The extensiveness of soil freezing and the impact of freezing and thawing on the physical, chemical, and biological properties of soils demand a thorough assessment as to the timing and depth to which freezing occurs in soil.

Soil freezing and thawing often result in damage to buildings and roads. Cracks in building foundations and roads caused by heaving or plastic deformation is of economic concern in cold regions. This damage can be minimized, however, by considering in the design process the depth to which soils freeze. Runoff can be enhanced as a result of frozen layers within the soil profile impeding drainage. Indeed, knowledge of frost penetration into soils has recently improved the prediction of runoff and stream flows in the northern USA (Cherkauer & Lettenmaier, 1999). Soil freezing and thawing also can affect soil stability and therefore erosion. Erosion is of particular concern when snowmelt or rain events occur as thawing proceeds from the soil surface.

Depth and duration of soil freezing can influence the viability and behavior of biological organisms as well as the chemical composition and distribution in the soil profile. For example, frost heave and the consequential exposure of viable plant tissues to subfreezing temperatures often result in death of plants such as alfalfa (*Medicago sativa* L.). In addition, survival of some soil-dwelling organisms depends on their ability to migrate to subsoil regions that remain unfrozen (Gerard, 1967). Freezing of pore water causes a decline in soil water potential and thereby magnifies the water potential gradient across the freezing front. Water migrates toward the region of low water potential, carrying with it solutes. Freezing also can cause formation of vertical cracks in the soil, thereby creating

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preferential pathways through which chemicals can readily move. Many more engineering and environmental processes are influenced by soil freezing, all of which occur in response to the change from water to ice within the soil pore.

Visual manifestations of the impact of soil freezing are all too familiar; one of the most recent in the northern USA and southern Canada being the 1997 flood of the Red River of the North. This flood was caused in part by rapid snowmelt while the soil was frozen (Bell & Halpert, 1998). Despite the importance of frozen soil, relatively little is known about the depth and duration of soil freezing around the world. This disparity in information is probably due to a lack of concern for frozen soil phenomena until the advent of the automobile in the early 1900s. Indeed, comprehensive field and laboratory studies of frozen soil phenomena did not begin until the 1940s with the rapid construction of roads and manufacturing of vehicles (Johnson, 1952). In addition, few automated methods have been developed to assess frost depth.

Manual exploration was and continues to be a laborious method in assessing frost depth. Air temperature observations were employed as early as 1890 (Stefan, 1890) in empirically determining the depth of soil freezing. This empirical relationship, called the Stefan equation, has been widely employed during the last century. Instrumentation to assess soil temperature has been used to detect soil freezing and thawing. Callendar and McCleod (1896) made some of the first temperature observations as soils undergo seasonal freezing. Only in the last five decades has interest in the depth of soil freezing resulted in further developments in instrumentation to accurately assess the depth of soil freezing and thawing. This chapter outlines some of the theoretical principles involved as soils freeze and some of the methods used to assess the depth of frozen soil.

PRINCIPLES

Water and air exist in a dynamic state of equilibrium in unfrozen soil pores. This dynamic equilibrium is influenced by the chemical, biological, and physical forces that affect the energy status of soil water. Chemical forces arise due to solutes lowering the free energy status of pure water. Biological forces such as those associated with root water extraction can deplete water reserves and thereby decrease the free energy of pure water. Physical forces such as adsorption also affect the free energy status of pure water.

The energy status of soil water is also influenced by temperature. At temperatures less than 0°C, solutes are excluded during ice formation and therefore become more concentrated in the unfrozen soil water. In field soils, however, the low concentration of solutes in soil water has little effect on the freezing point of water. As ice forms in soil pores, the decline in liquid water content is increasingly affected by the physical forces of capillarity and adsorption. Capillarity reduces the free energy of water as a result of the greater forces that are required to freeze water in progressively smaller pore spaces. Adsorption also reduces the free energy of water as a result of greater forces that must be overcome to freeze the fewer layers of adsorbed water on the soil particle surface. Capillarity appears to be the predominant force affecting the amount of unfrozen water in a soil pore

at temperatures between 0 and –1.5°C. Below –1.5°C, adsorption forces largely govern the unfrozen water content (Bouyoucos, 1917; Williams & Smith, 1991).

Soil water freezes progressively over a range of temperature below 0°C. This range in temperature, or freezing point depression, is caused primarily by capillary and adsorption forces influencing the free energy status of soil water. The freezing point depression of soil was first observed by Bouyoucos and McCool (1915), but Schofield (1935) later identified the relationship between matric potential (measure of the free energy of water) and temperature in frozen soils.

Based upon thermodynamics, the change in Gibbs free energy (ΔG) of water and ice must be equal at the freezing point:

$$\Delta G_{\mathrm{i}} = \Delta G_{\mathrm{w}} \qquad [1]$$

where the subscript *i* refers to ice and *w* refers to water. The change in Gibbs free energy can be further specified by considering the conservation of energy, mass, and entropy in a freezing soil. In so doing, Eq. [1] becomes:

$$V_{\mathrm{i}}dP_{\mathrm{i}} - S_{\mathrm{i}}dT = V_{\mathrm{w}}dP_{\mathrm{w}} - S_{\mathrm{w}}dT \qquad [2]$$

where V is the specific volume, dP is the change in pressure with respect to atmospheric, S is the specific entropy, and dT is the change in temperature with respect to the freezing point. The difference in entropy between ice and water is equivalent to the heat released during the phase change (L) divided by the temperature at which the phase change occurs (T_{o}). Assuming that ice is at atmospheric pressure, Eq. [2] can be written in a generalized form of the Clausius-Clapeyron equation:

$$dP_{\mathrm{w}} = -(L/(V_{\mathrm{w}}T_{\mathrm{o}}))*dT \qquad [3]$$

which expresses the relationship between matric potential (dP_{w} in MPa) and temperature in a freezing soil. In Eq. [3], L equals 333.5 J g^{-1}, V_{w} is the specific volume of water (1×10^{-6} m^3 g^{-1}), and T_{o} equals 273.15°K. This equation assumes that solutes have no effect on the free energy of soil water and has been found to correspond well with experimental data (Williams & Smith, 1991). This expression indicates that the matric potential declines by about 1.2 MPa $°C^{-1}$ below 0°C.

The ice content of a partially frozen soil can be ascertained by using Eq. [3] and knowing the total water content and the unfrozen water characteristic curve. Initial water content is an important factor determining when soils begin to freeze (Cary et al., 1978). Indeed, ice formation occurs at a lower temperature (larger freezing point depression) in drier than wetter soils. The type of instrumentation used to measure the depth of soil freezing is then important as those instruments that detect the 0°C isotherm may overestimate the depth of soil freezing due to freezing point depression.

Some of the physical differences between ice and water (Table 8–1) have been exploited to detect whether a soil is frozen or not frozen. For example,

Table 8–1. Selected physical properties of water and ice at 0°C.

Property	Water	Ice
Specific heat, J g^{-1} $°K^{-1}$	2.0	4.2
Thermal conductivity, W m^{-1} $°K^{-1}$	2	0.5
Velocity of radio waves, m μs^{-1}	170	0.002
Dielectric constant	80	4

amplification of the dielectric constant due to ice formation in soil pores has been measured using time domain reflectometry to determine the liquid water content in frozen soils (Patterson & Smith, 1980; Spaans & Baker, 1995). In addition, differences in the transmission velocity of radio waves between ice and water have been used to detect the depth of frozen soil (Lawson et al., 1996). Ice content within soil pores also influences the strength of the soil. The strength of a partially frozen soil is dependent on the interaction between the ice, water, air, and soil particle components (Williams & Smith, 1991). Lovell (1957) demonstrated that the bearing strength of partially frozen soil increased exponentially with an incremental increase in ice content.

METHODS OF INSTRUMENTATION OR EXPLORATION

Soil freezing has been temporally and spatially delineated in the laboratory and field by measuring soil temperatures, detecting ice formation in water-filled apparatus positioned in the soil (e.g., frost tubes), detecting changes in the liquid water content of soil pores, and manually exploring the soil profile. Instrumentation to detect changes in the liquid water content of soil pores are perhaps the most numerous. Laboratory techniques to assess liquid water content of partially frozen soil pores include dilatometry, x-ray diffraction, heat capacity, nuclear magnetic resonance, and differential thermal analysis (Kay & Perfect, 1988). In addition, microwave signatures (Wegmuller, 1990) and computer tomography (Yibin, 1993) have been used to assess liquid water content as soils freeze. Electrical resistance tomography (Daily et al., 1992) also may have application to assessing liquid water content in frozen soils, but has yet to be investigated. All of these techniques are ill-suited for determining frost depth under field conditions due to the constraints (i.e., portability, sensitivity, sample size) of the instrumentation. Changes in liquid water content of partially frozen soil pores, however, have been successfully measured in the field using electrical conductivity and electromagnetic radiation sensors. This section highlights those instruments and techniques suitable for determining the depth of frozen soil.

Temperature

Soil temperature has been widely used as an indicator of frozen soil for over a century. Indeed, soil temperature is often used as a standard of comparison in assessing frost depth using other instrumentation or by simulation. Rickard and Brown (1972), for example, used the time varying position of the 0°C tem-

perature in the soil profile (0°C isotherm) as the basis to compare the performance of a frost tube in determining the depth of soil freezing. Baker et al. (1982) also used the 0°C isotherm to examine the performance of time domain reflectometry in locating the frozen–unfrozen interface as soils freeze. Hayhoe et al. (1983) compared frost depth ascertained by the 0°C isotherm with that measured by a frost tube and time domain reflectometry and that simulated at a field site in Ottawa, Canada.

Callendar and McCleod (1896) made some of the earliest observations of seasonal variation in soil temperature in North America. They found seasonal temperatures below freezing (0°C) to depths >1.0 m at Montreal, Quebec, Canada. Bouyoucos (1920) first observed that soils could be cooled to < 0°C without freezing. Beskow (1935) later reported that supercooling was dependent on soil type; sand froze between –0.1 and –0.2°C while clay froze between –0.4 and –1.0°C. Haley and Kaplar (1952) confirmed that soils freeze across a range of temperature from 0 to –1.6°C depending on soil type. These studies illustrate the difficulty in using soil temperature to differentiate between frozen and unfrozen soil due to the inability of the temperature sensor to detect phase changes that occur as soils freeze. Therefore, while soil temperature can be easily measured using a variety of instruments (see Novak, 2005, this volume), soil temperatures provide little information concerning the depth of frost in soils.

Frost Tube

The frost tube is comprised of an outer rigid pipe and an inner flexible tube (Fig. 8–1). The bottom of both the pipe and tube are sealed. The inner tube is filled with water or a low osmotic solution and sealed at the top. A wire or bolt protrudes from the top of the inner tube, which allows the tube to be extracted from the outer casing. The pipe is installed vertically into the soil profile with some length protruding above the soil surface for ease of access during periods of snow cover. An observation of frost depth is made by extracting the inner tube from the pipe and determining the depth of ice in the tube relative to the soil surface. Multiple observations necessitate using controlled traffic patterns to minimize compaction of soil and snow around the tube. The design of the inner tube to ensure an accurate estimation of frost depth has been the subject of much research during the past 30 years.

Frost tubes are inexpensive and easy to fabricate. Ease of use by untrained personnel has prompted worldwide use of tubes as well as refinement for better performance. Harris (1970) compared frost depth using a modified Gandahl tube and a Fernow tube. The modified Gandahl tube consists of a polyethylene tube filled with saturated sand. The medium-sized sand is saturated with a 0.01 percentage of solution of green fluorescein dye. The Fernow tube (Patric & Fridley, 1969) was similar to the Gandahl tube except that the polyethylene tube was filled with a Kool-Aid fruit drink solution and the outer casing was copper instead of plastic. Harris concluded that the modified Gandahl tube performed better during periods of rapid freezing and thawing. Rickard and Brown (1972) compared the performance of the Gandahl tube and the modified Gandahl tube. The Gandahl tube (Gandahl, 1957) consists of inner tube filled with a methylene

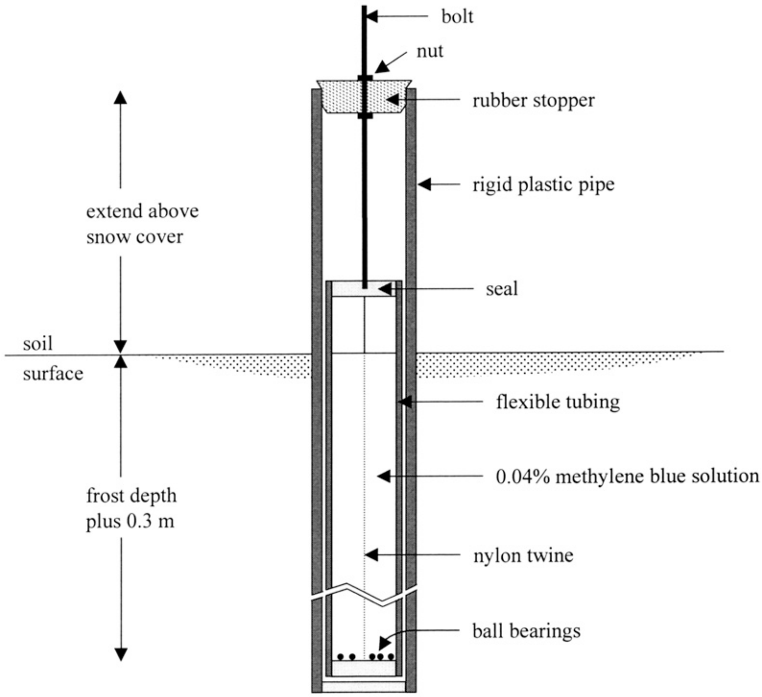

Fig. 8–1. Frost tube.

blue solution (no sand). They found the modified Gandahl tube responded more rapidly to freezing and thawing than the Gandahl tube. In addition, the modified Gandahl tube indicated frost depth within 20 mm of the 0°C isotherm while the Gandahl tube consistently lagged the modified Gandahl tube by as much as 50 mm. Rickard and Brown concluded that the modified Gandahl tube was superior to the Gandahl tube in responding to freezing and thawing (due to the lower mass of water within the tube) and delineating the depth of freezing (the fluorescein dye solution was clearly excluded from ice).

Ricard et al. (1976) later modified the Gandahl tube at the Cold Regions Research and Engineering Laboratory (CRREL). This tube, referred to as the CRREL-Gandahl tube, consisted of an inner polyethylene tube (15.9-mm outside diameter) filled with a 0.05% solution of methylene blue. A nylon string transcended the center of the tube to anchor the ice during periods of thaw. The outer casing was of rigid polyethylene or PVC pipe with a diameter of no more than 6 mm greater than the diameter of the inner tube. They recommended that the gage should be at least 0.30 m longer than expected frost depth to minimize the freezing point depression as dye is excluded from the freezing front. The authors of this chapter have placed ball bearings inside the inner tube for assessing the position of the freezing front. This is accomplished by inverting the tube during a measurement, thus allowing the ball bearings to gravitate through the unfrozen solution to the ice front.

In 1974, the USDA-ARS and the University of Idaho began to search for simple and inexpensive methods to assess frost depth in the Pacific Northwest region of the USA. A modified Gandahl tube, but with a rigid inner tube, worked well over several winters for researchers and technicians acquainted with the idiosyncrasies of the tube. Cooperative observers in the region, however, could not differentiate the color change from yellow-green to gray as the solution froze. The modified Gandahl tubes performed poorly during periods of rapid freezing and on occasion broke (inner tube split) when exposed to very cold temperatures prior to installation. Thus, in 1978, personnel in the Pacific Northwest adopted the CRREL-Gandahl tube. This tube allowed better delineation between frozen and unfrozen regions in the tube, either visually or by squeezing the tube.

Electrical Conductivity

Electrical potential differences arise due to charge separations that occur across ice–water interfaces during freezing of aqueous solutions. Other factors such as temperature gradients across the ice–water interface also may influence potential differences, but those factors are not fully understood (Burn et al., 1998). Potential differences of several hundred mV have been observed during soil freezing in the field (Outcalt et al., 1989). Although Parameswaran and Mackay (1996) suggested that a profile of electrical potential measurements may delineate the depth of frozen soil, little progress has been made in measuring electrical potentials to assess frost depth.

Changes in electrical conductivity or resistance also arise as soils freeze. Electrical resistance is defined as:

$$R = L/(\sigma A) \qquad [4]$$

where L (m) is the distance, A (m^2) is the cross sectional area, and σ is the electrical conductivity (Ω^{-1} m^{-1}) of the material through which charge carriers move. As soils freeze, the mobility of the charge carriers is constrained to the thinner film of liquid water surrounding the soil particles. This decline in mobility results in a decrease in the electrical conductivity (Hoekstra, 1965) and a consequential rise in electrical resistance. The depth of soil freezing has been assessed by measuring changes in electrical conductivity or resistance with resistance probes, soil moisture blocks, and electromagnetic induction sensors.

Resistance Probe

The resistance probe was described by Atkins(1979) and has since been modified by McIntosh and Sharratt(1997). Components of the modified probe (Fig. 8–2) include 20-gauge wire and small diameter (25 mm) PVC pipe. The probe is assembled by wrapping one strand of 20-gauge wire around one-half of the circumference of the pipe at 10 mm increments along the length of the pipe. Leads from the wires run through the center of the pipe, which is filled with insulating foam to minimize heat flow through the pipe. The ends of the probe are sealed to prevent entry of moisture. Probe installation requires drilling or extract-

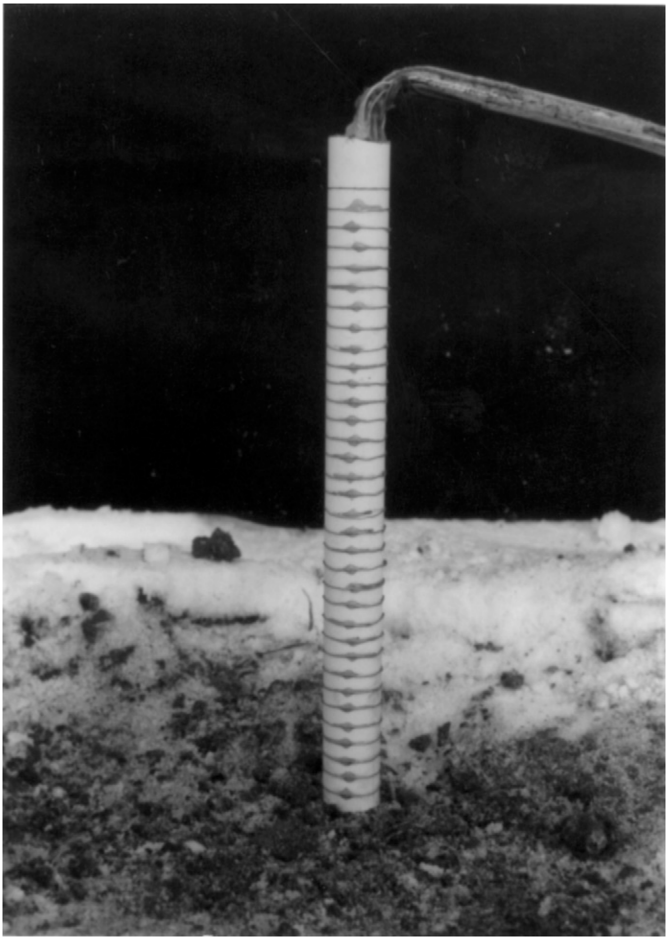

Fig. 8–2. Electrical resistance probe being installed at a field site. Wires are located at 10 mm intervals along the PVC pipe with the top strand of wire buried 10 mm below the soil surface. Leads from the wires protrude from the top of the probe.

ing a core to create a vertical channel in the soil, inserting the probe into the channel, and tamping soil around the wires (electrodes) to ensure good contact with the soil. Electrical resistance is measured between each of the nearest pair of electrodes. A data logger can be used to measure the electrical resistance using an AC bridge circuit.

Frost depth ascertained using the resistance probe has compared favorably with temperature sensors (Atkins, 1979; Hayhoe & Balchin, 1986) and frost tubes (McIntosh & Sharratt, 1997). Atkins (1979) found that resistance probes are superior to temperature sensors in determining frost depth in soils with significant freezing point depressions (saline soils). One disadvantage of the resistance probe is the difficulty in assessing the onset of freezing in dry soil due to the high resistance of dry soils (Brach et al., 1985).

Soil Moisture Block

Soil moisture blocks comprise a pair of electrodes encased in a porous medium such as gypsum, fiberglass, or nylon. The resistance of the electrodes can be monitored with an AC bridge circuit. Conductivity between the pair of electrodes approaches zero when moisture freezes within the soil moisture blocks. A soil moisture block allows the state of soil water to be assessed at a discrete position in the soil, thus multiple blocks must be placed at various depths within the soil profile to accurately assess the depth of soil freezing or thaw.

Garstka (1944) used gypsum blocks to assess frozen soil. He found that blocks were superior to temperature sensors for detecting frozen soil because the formation of pore ice did not occur universally in soils at a temperature of 0°C. Sartz (1967) also found that gypsum blocks were more accurate than thermistors in determining depth of frozen soil. McCool and Molnau (1984) compared gypsum blocks and frost tubes in assessing the depth of soil frost for various soil management practices. They found similarity in frost depth using both sensors during periods of active freezing, but that gypsum blocks responded more rapidly than frost tubes during periods of thaw. Hanson and Flerchinger (1990) illustrated that gypsum blocks do not necessarily agree with other techniques that assess frost depth based upon the 0°C isotherm. They found gypsum blocks more accurately portrayed the delay in freezing or advance in thawing caused by freezing point depressions in soils.

Electromagnetic Induction

Small currents can be generated in the soil using electromagnetic induction. This technique consists of applying a time-varying magnetic field to the soil using an electromagnetic induction device. This primary magnetic field generates a secondary magnetic field in the soil, both of which can be sensed using the device. The ratio of the primary to secondary magnetic fields is proportional to the soil electrical conductivity.

An electromagnetic induction device (Fig. 8–3) consists of a transmitter, receiver, and power supply. The device is placed on or just above the soil surface while making a measurement and can be transported across a field by an operator or vehicle. These devices require calibration or multiple soundings to ascertain changes in soil conductivity associated with soil freezing. The principle advantages of using electromagnetic induction lies in the speed and accuracy of taking non-invasive spatial measurements of soil conductivity. In addition, soil conductivity is ascertained for a large volume of soil. Disadvantages of this technique include the difficulty associated with magnetically-inducing sufficient current in the soil at low water content. Electromagnetic induction has been used with success in delineating frozen soil (Osterkamp et al., 1980).

Electromagnetic Radiation

The propagation velocity of an electromagnetic wave through the soil is dependent on the electrical and magnetic properties of the soil. Electrical properties are altered as soils undergo freezing. Indeed, the dielectric constant of a soil varies according to the proportion of pore space filled with air, water, and ice.

Fig. 8–3. Electromagnetic induction device. The Geonics EM38-DD is 1-m long with digital meters located on the top and side of the device for horizontal and vertical dipole measurements. Photograph courtesy of Geonics Limited, Mississauga, Ontario, Canada.

The individual dielectric constants of air, water, and ice are approximately 1, 80, and 4, respectively (Table 8–1). Consequently, the dielectric constant of a field soil will decline as pore water freezes. Differences in the propagation velocity of electromagnetic waves between frozen and unfrozen soil has been ascertained using ground penetrating radar and time domain reflectrometry.

Ground Penetrating Radar

Ground penetrating radar is a geophysical method for imaging the subsurface. This method involves transmitting electromagnetic pulses of radio waves into the soil through a transducer or antenna. The transmitted energy is reflected from various layers in the soil. Another antenna receives the reflected wave pulses, with the strength of the reflected pulse determined by the contrast in the dielectric constant between the frozen and unfrozen layer. Data analysis of the reflected pulse requires detailed knowledge of the subsurface structure. For example, the depth of frost penetration can be interpreted from the time delay in the reflected pulse at the frozen-unfrozen soil interface only by knowing the dielectric constant of the frozen soil layer.

A radar system includes a control unit, assorted antennas, cables, and a power supply (Fig. 8–4). Ground penetrating radar profiles are made by towing antennas along the ground behind an operator or vehicle. The principle advantages of using ground penetrating radar lies in the speed and accuracy of taking noninvasive, spatial measurements. A disadvantage of this technique is the knowledge required of the subsurface electrical properties to interpret the reflected pulses.

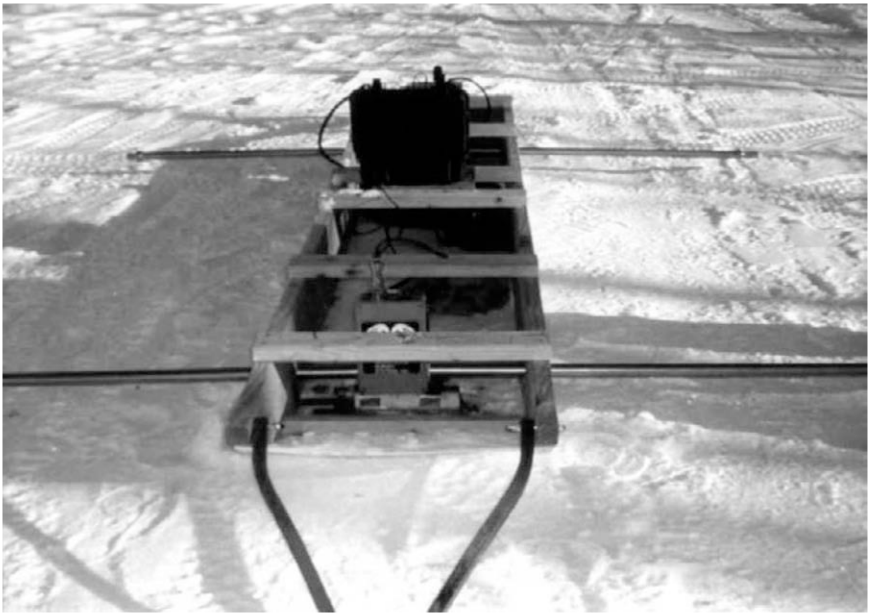

Fig. 8–4. Ground penetrating radar unit. The sled is equipped with a control unit (on seat at middle of sled), transmitter and receiver antennas (front and rear of sled), and a power supply (below control unit). Photograph courtesy of Larry Hinzman, University of Alaska, Fairbanks.

Time Domain Reflectometry

A time domain reflectometer (TDR) consists of a timing control unit, pulse generator, and a receiver. The generator produces an electromagnetic pulse that is transmitted through the receiver and into a transmission line. For application to soil science, a pair of transmission lines is embedded in the soil and acts as a wave guide for the pulse. The propagation velocity of the pulse along the wave guide is inversely proportional to the dielectric constant of the soil (Topp et al., 1988). Thus, as the dielectric constant decreases due to the conversion of water to ice, an increase occurs in the velocity of the pulse through the soil. In practice, the velocity of the pulse is determined from the length of the transmission line in the soil and the measurement of the pulse travel time in the soil. This information is then used to calculate the dielectric constant and therefore liquid water content of the soil (Topp et al., 1988).

A TDR system commonly used in measuring soil water content is shown in Fig. 8–5. A data logger and wave guides are used along with a TDR; the data logger aids in processing and storing information. The wave guides can be inserted horizontally or vertically into the soil profile. Depth of soil freezing can be determined at discrete depths by installing wave guides horizontally into the soil. Frost depth can be continuously monitored by installing wave guides vertically into the soil; vertical orientation, however, requires an assessment of the dielectric constant within the frozen or unfrozen layer of soil (Baker et al., 1982). Vertical installation may also result in heat and water transfer along the length of the wave guide. Care should be taken when inserting the wave guides into the soil to ensure good contact with the soil and proper distance between wave guides.

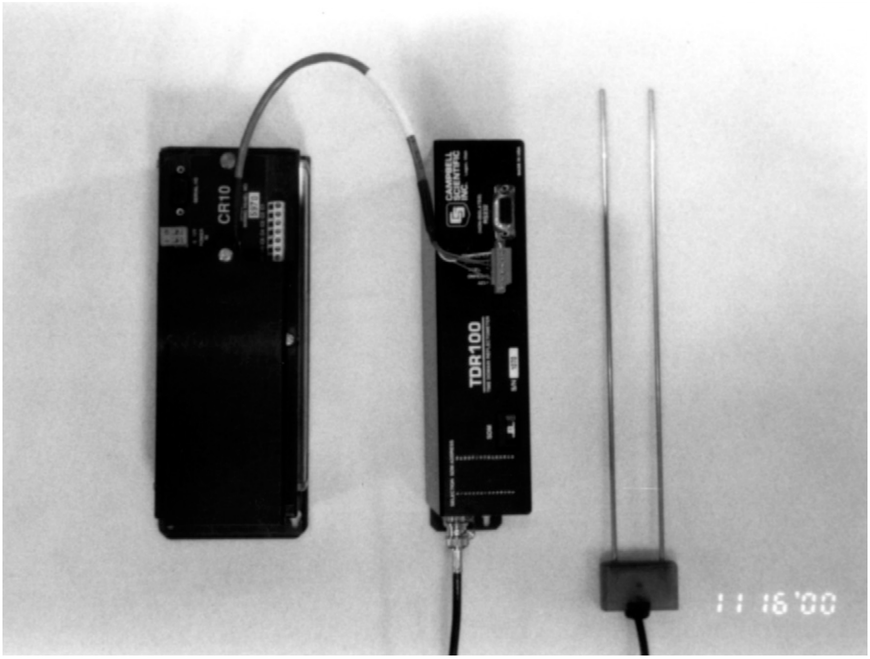

Fig. 8–5. Time domain reflectometer unit. Components of the unit from top to bottom include a data logger, time domain reflectometer, and 30-cm long waveguide.

Frost penetration ascertained by a TDR has agreed well with techniques that assess penetration based upon temperature measurements (Baker et al., 1982) and frost tubes (Hayhoe et al., 1983). As the soil profile becomes isothermal near 0°C, however, a TDR is more accurate than temperature measurements and frost tubes in determining thaw or frost depth (Hayhoe et al., 1983).

Manual Exploration

Manual exploration is a direct method to which all other methods can be compared in assessing frost depth. The laborious and time-consuming task of manually exploring the soil, however, makes such measurements infrequent. Manual exploration is commonly used to assess soil profile descriptions in areas of permafrost. We review in this section those methods that have been used to manually explore frozen soils.

Probes and Tubes

Modified Hoffer Probe. The modified Hoffer probe is a hand held device (Fig. 8–6) capable of cutting through a frozen soil layer. The probe has been described by Zoltai (1978) and includes a coring bit, extension rods, and handle. The bit and extension rods are made from seamless steel tubing. The bit is serrated and hardened. The probe is operated by driving the bit into frozen soil. The

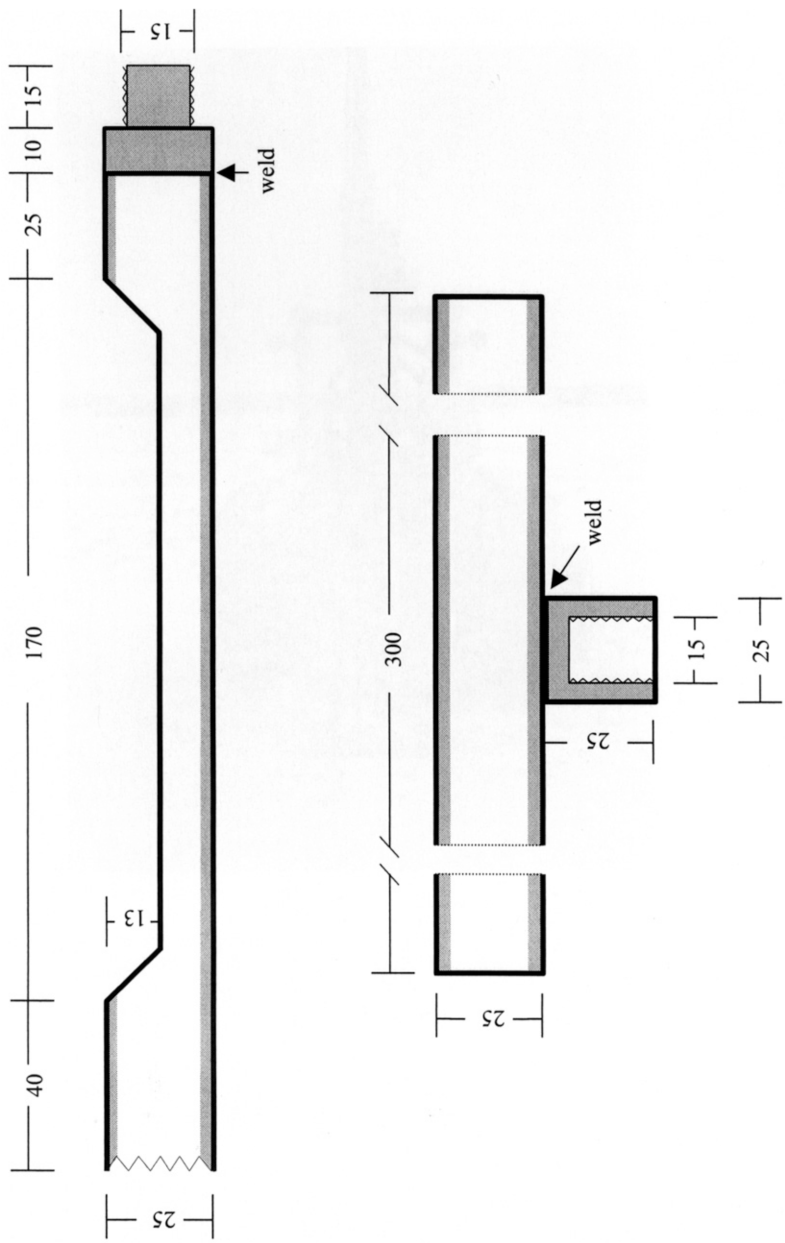

Fig. 8–6. The modified Hoffer probe. The diagram depicts the coring bit with serrated edge and T-handle (dimensions in mm).

probe is rotated as the bit contacts the soil, thus causing the bit to cut through the frozen soil. The soil core sample taken with the probe can be examined for the presence or absence of ice crystals. The probe is light weight and most effective at exploring soils with low ice content. The probe is capable of penetrating several meters of frozen soil.

Soil Sampling Tubes. Soil sampling tubes have been used to ascertain frost depth. Soil tubes are driven into frozen soil and used to extract a soil core sample. The sample is examined for ice crystals; lack of ice crystals at some distance along the core sample is considered the depth of frost penetration. Frost penetration also can be ascertained by monitoring the resistance of the tube to penetration in frozen soil. A large decrease in the resistance of the tube to penetration of the soil occurs at the frozen–unfrozen interface. Atkinson and Bay (1940) used a hammer to drive a soil sampling tube into frozen soil. The authors have driven a sampling tube into frozen soil using a Giddings machine (Fig. 8–7). This machine has been used to extract frozen soil samples to a depth of 1.0 m.

Fig. 8–7. Soil sampling tube (1-m long) driven into frozen soil using a Giddings machine (Giddings Machine Company, Fort Collins, CO).

The depth and time required to penetrate a frozen soil layer depends largely on soil type and ice content of the frozen layer. Any device used to drive a soil sampling tube into frozen soil exerts considerable force on the sampling tube, at times causing deformation or breakage of the tube.

Power Hammers

The Cobra hammer (Fig. 8–8) can be used for excavating frozen soil. The hammer is powered by a two-stroke gasoline engine and weighs about 20 kg. Tools used with the hammer include drills, spades, and chisels. The hammer has both rotary and hammering motions. This hammer can be transported to field sites via a pack frame.

The PICO hammer also is powered by a gasoline engine and has both rotary and hammering motions (Fig. 8–9). The hammer can be used with various attachments similar to those used with the Cobra hammer. Tarnocai (1993) built a

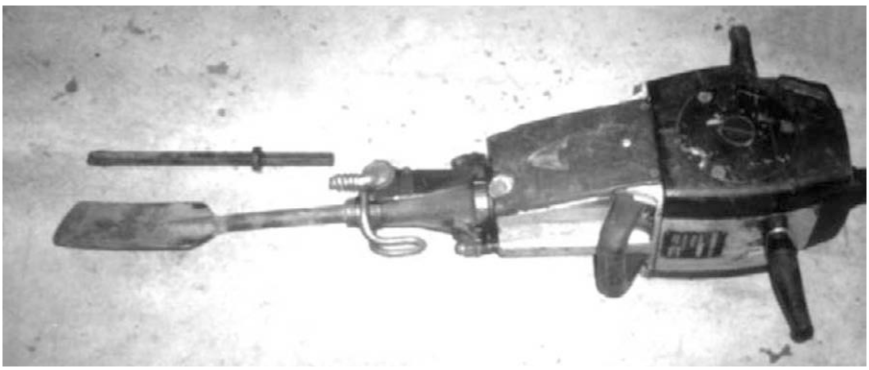

Fig. 8–8. Cobra hammer used to excavate frozen soil. The hammer is shown with the spade and chisel attachments. Photograph courtesy of Larry Hinzman, University of Alaska, Fairbanks.

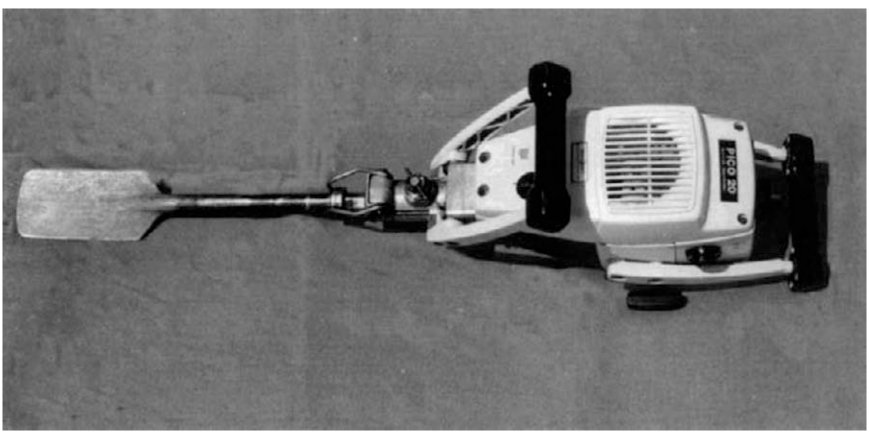

Fig. 8–9. PICO hammer with the spade attachment. Photograph courtesy of Charles Tarnocai, Agriculture and Agri-Food Canada, Ottawa, Ontario.

coring attachment for the PICO hammer that allows extracting a soil core sample from frozen soil. The hammer has the advantage of being lightweight (10 kg) and also can be carried on a pack frame.

CRREL Auger

The CRREL auger was developed by the U.S. Army Cold Regions Research and Engineering Laboratory for sampling snow, ice, and mineral soils. The auger consists of a steel core barrel with an inside diameter of 76 mm and a length of 920 mm (Fig. 8–10). The outside of the barrel has a welded, double helix flight configuration with a 30° slope. The bottom of the barrel has a stainless steel cutting shoe with two cutting bits. These bits can be sharpened or replaced as necessary. A removable head fastens to the top of the barrel and can be connected to a rotary motor. The authors have used both a portable gasoline-

Fig. 8–10. The CRREL auger for coring in frozen soil. The auger consists of a steel core barrel (barrel depicted in this photograph is 920-mm long) and a stainless steel cutting shoe with two cutting bits.

powered motor and a Giddings machine (Fig. 8–11) to extract frozen core samples using the CRREL auger. The rotary action of the auger allows the bits to cut through the frozen soil. The core sample is removed through the top of the barrel and examined for ice crystals along the length of the sample. The presence or absence of ice crystals signifies whether the sample is frozen or unfrozen. The auger provides an undisturbed soil core and can penetrate to several meters with the aid of extension rods. The auger requires constant rotation during drilling to avoid in-situ freezing of the barrel.

Penetrometers

Lake States Penetrometer. Stoeckeler and Thames (1957) developed the Lake States penetrometer as a method to assess the depth of frozen soil. The penetrometer consists of a 13-mm diameter, 1.0-m long solid steel rod and a 25-mm

Fig. 8–11. CRREl auger attached to a Giddings machine.

diameter, 0.4-m long steel pipe. The rod has 25-mm incremental markings along its length. The top and bottom of the pipe are fitted with a 50- and 10-mm reducing coupler, respectively. The 50-mm coupler and upper portion of the pipe are filled with lead. The solid rod slides inside the 10-mm coupler and pipe. The pipe is used as a hammer having a 0.3-m stroke. The total weight of the penetrometer is 7 kg. The steel rod is driven into the soil. After each stroke, a record is made of the depth of penetration of the rod. Depth of frozen soil is determined by examining the depth of penetration after each stroke. A large difference in depth of penetration will occur between regions of frozen and unfrozen soil. Frozen soil can be detected to a depth of 0.6 m with an accuracy of 0.01 m. The Lake States penetrometer was later modified by Sartz (1967) who enlarged and tapered the tip of the rod to reduce the frictional drag on the rod. The penetrometer is portable and accurate, but requires considerable labor in penetrating the frozen soil layer and is ill suited for soils with stones.

Pointed Rod. Depth of soil thaw can be ascertained using a pointed steel rod. Sharratt and Flerchinger (1995) used a 3 mm pointed rod to determine the depth of soil thaw. The rod is inserted into soil until further insertion is prevented by a frozen soil layer. The distance the rod is inserted into the soil can be measured using a ruler. Soil thaw can be determined to a depth of about 0.2 to 0.4 m depending on soil type, water content of the thawed soil layer, ice content of the frozen soil layer, and rod thickness. Error in assessing thaw depth with a rod lies in the ease to which a pointed rod can penetrate a soil having a temperature near 0°C and low ice content.

MODELS

Models have been developed to predict frost depth in soils. These models vary in their degree of complexity in dealing with heat and water transfer processes in soils during freezing and thawing. Simplified models generally require little data to simulate frost depth, but cannot account for the complex physical interactions that govern the freezing and thawing process. The more complex, physically based models require information concerning the weather, soils, vegetation, and topography. In addition, these models attempt to account for the underlying processes that govern heat and water transfer as soils freeze and thaw.

Semi-Empirical

Numerous formulas have been developed for predicting the depth of frost penetration. The simplest and earliest of these is the Stefan formula (Aldrich, 1956):

$$X = \sqrt{(2k_f F/L)} \qquad [5]$$

where X is the depth of frost penetration (m), k_f is the thermal conductivity of frozen soil (W m^{-1} C^{-1}), L is the latent heat of fusion (J m^{-3}), and F is the surface

freezing index (°C s). The F is defined as the cumulative difference between the freezing-point temperature and the soil surface temperature across a given time interval. The Stefan equation neglects the volumetric heat of the frozen and unfrozen soil, and, hence, tends to overestimate frost penetration (Aldrich, 1956).

The modified Berggren formula (Aldrich, 1956) attempts to account for heat storage in a freezing soil by including the volumetric heat capacity in the Stefan formula. The modified Berggren formula may be written in the form:

$$X = \lambda\sqrt{(2k_f F/L)} \qquad [6]$$

where λ is a dimensionless coefficient that corrects for volumetric heat capacity. Since the Stefan formula tends to overestimate freezing depth, the coefficient λ is always less then unity for freezing soil. Equations [5] and [6] require knowledge of the soil surface temperature. Since surface temperatures are seldom measured, other methods are used to estimate surface temperature. For example, Fox (1992) recently incorporated the Stefan formula into a hydrology model to account for the effects of soil freezing and thawing on soil water infiltration and drainage. He recognized the difficulty in empirically deriving the surface temperature from air temperature for agricultural and forest soils and therefore used an energy balance approach to estimate soil surface temperatures.

Heat Balance

Models have been developed that predict freezing and thawing based upon the heat balance of soils. These models rely on semi-empirical relationships between soil heat balance and frost depth and generally predict modes of heat transfer using limited weather data. Two examples of heat balance models are those developed by Cary et al. (1978) and Benoit and Mostaghimi (1985).

Cary et al. (1978) predicted the occurrence of soil freezing by determining the net heat transfer across the surface of a soil whose temperature is near 0°C. Net heat transfer across the soil surface (M) was defined as:

$$M = \Sigma(G_n + G_u) \qquad [7]$$

where M (W m^{-2}) is <0 if the soil is frozen and G (W m^{-2}) is the daily soil heat flux directed either downward across the surface (G_n) or upward from subsoil layers into the zone susceptible to freezing (G_u). The cumulative difference in soil heat flux is determined across a number of consecutive days. They determined G_u as a function of calendar day and defined two approaches for estimating G_n. The first approach determined heat transfer from thermal conductivity and the temperature gradient across a soil surface layer according to Fourier's law. Soil temperatures within the surface layer were determined as an empirical function of air temperature and snow depth. This approach does not require solar radiation data and provided reasonable estimates of frost depth in eastern Washington. The second approach estimated G_n from the energy balance at the soil surface, given by the equation:

$$G_n = R_n - L_v E - H \qquad [8]$$

where R_n is net radiation (W m^{-2}), L_vE is heat associated with the evaporation of water (W m^{-2}), and H is sensible heat exchanged between the surface and atmosphere (W m^{-2}). These parameters were estimated from air temperature, solar radiation, and snow cover. The model of Cary et al. (1978) was developed to assess whether the soil is frozen or unfrozen, not the depth of freezing or thawing.

Benoit and Mostaghimi (1985) developed a model to predict frost depth for agricultural soils subject to various forms of tillage and residue management. The model calculates net heat flow between the frozen and unfrozen regions of the soil profile based upon daily inputs of air temperature and snow depth. Frost depth is determined from discrepancies in heat flow between these two regions. Heat flow (G) within the frozen region of soil was determined according to:

$$G = k_f \Delta T/\Delta z \qquad [9]$$

where ΔT is the temperature (°C) difference between the soil surface and the freezing front and Δz is the distance between the soil surface and freezing front (m). Heat flow from the unfrozen to frozen region of the soil was computed as the sum of heat transferred by thermal conduction within the unfrozen soil, the latent heat of fusion associated with water migration to the freezing front, and changes in heat storage in the unfrozen region. The model assumes discrete modes of heat transfer associated with thermal conduction and water migration, when in fact, these modes of heat transfer are coupled in a freezing soil.

Coupled Heat and Water Transport

Freezing induces water transfer in soils due to a reduction in water potential as water freezes within a soil pore. This reduction in water potential associated with freezing generally results in movement of water from unfrozen regions to the freezing front. Heat movement will also occur in frozen soils due to thermal advection associated with water transfer to a freezing front and as a result of temperature gradients. Therefore, water and heat transfer are coupled processes occurring within a freezing soil. Only recently have physically-based models been developed to account for coupled water and heat movement in agricultural soils subject to various management practices and subfreezing environments.

The Simultaneous Heat and Water (SHAW) model was developed to evaluate the effects of tillage and residue management on soil freezing (Flerchinger & Saxton, 1989). The model uses finite difference methods to solve the heat balance equation:

$$C_s \partial T/\partial t - \rho_i L_f\, \partial \theta_i/\partial t + L_v \partial \rho_v/\partial t = d(k \partial T/\partial z)/\partial z + C_l \partial q_l T/\partial z - L_v \rho_v \partial q_v/\partial z + S \qquad [10]$$

where C_s and C_l are the respective heat capacities of soil and liquid water (J m^{-3} C^{-1}), t is time (s), ρ_i and ρ_v are the respective densities of ice and water vapor (kg m^{-3}), L_f and L_v are the respective latent heats of fusion and vaporization (J kg^{-1}), θ_i is the volumetric ice content (m m^{-1}), k is the thermal conductivity of soil and varies by the state of water in soil (W m^{-1} C^{-1}), q_l and q_v are the respective fluxes of water and water vapor (m s^{-1}), and S is a source or sink term (W m^{-3}).

Hourly inputs of measured or estimated air temperature, wind speed, relative humidity, solar radiation, and precipitation are required to estimate parameters used in the model. Coupled heat and water transfer models, such as the SHAW model, have produced better estimates of frost depth under a range of environmental conditions compared with other less sophisticated models (Kennedy & Sharratt, 1997).

COMMENTS

A variety of methods have been used in determined the depth of frozen soil. These methods range from manual exploration to computer simulation. Manual exploration is laborious, time consuming, and can damage equipment. This method requires the ability to identify the presence of ice crystals within the soil matrix or to detect variations in the resistance of a penetrating rod. Computer simulations range in complexity from simple empirical relationships to those that include physically-based, interactive processes. Simulations with increasing complexity generally require more information concerning the physical environment, but more accurately mimic soil freezing under a range of environmental conditions. Confidence in using a computer simulation to assess frost depth requires familiarity in using the simulation.

Indirect methods for assessing the depth of soil freezing rely on instrumentation to detect phase change in soils. Temperature sensors and frost tubes can be used to assess the position of the 0°C isotherm (or some temperature below 0°C) in soils, but ice formation in soil pores rarely occurs at 0°C. Instrumentation has been developed during the past several decades to detect the presence of ice in soil pores. Ice formation in soil pores can be detected by monitoring changes in the electrical properties of soils. Time domain reflectometry is a common method used to assess changes in the dielectric constant as soils freeze.

Most methods employed today in assessing the depth of soil freezing are constrained by detecting ice formation at discrete positions within the soil profile or by periodic observations. Interpolation between positions in the soil profile or time of observations is then used to identify the temporal nature of the freezing front in the soil. New methods are therefore needed that continuously track the progression of ice formation within a soil profile.

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