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AD-A219 687

US Army Corps of Engineers
Cold Regions Research & Engineering Laboratory

Proceedings
International Symposium

Frozen Soil Impacts on Agricultural, Range, and Forest Lands

March 21-22, 1990
Spokane, Washington

K.R.Cooley, Editor

CRREL SPECIAL REPORT 90-1

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Sheraton-Spokane Hotel
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K.R,Cooley, Editor
USDA Agricultural Research Service
Northwest Watershed Research Center
Boise, Idaho

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Special Report 90-1 • March 1990

U.S. Army Cold Regions Research and Engineering Laboratory
72 Lyme Road, Hanover, New Hampshire 03755-1290

PREFACE

This symposium grew out of more than ten years of intensive investigations concerning frozen soil impacts in the northwestern United States and other northern regions. The work was coordinated through the STEEP (Solutions to Environmental and Economic Problems) organization, a research consortium consisting of: Oregon State University, University of Idaho, Washington State University, and the USDA Agricultural Research Service. Many other institutions and agencies participated in, and contributed to, this focused research. While much remains to be done, significant progress has been made that should be documented. These symposium proceedings make much of this documentation available in one volume.

The organizing committee chairmen were:

Symposium	Dr. Keith E. Saxton, USDA, ARS, Pullman, Washington
Vice Symposium	Mr. Gary E. Formanek, USDA, SCS, Portland, Oregon
Program	Dr. Myron Molnau, University of Idaho, Moscow, Idaho
Publication	Dr. Keith R. Cooley, USDA, ARS, Boise, Idaho
Local Arrangements	Dr. Don K. McCool, USDA, ARS, Pullman, Washington

We gratefully acknowledge the financial support provided by the USDA Soil Conservation Service West National Technical Center, Portland, Oregon, and the USA Corps of Engineers Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire (DA Project 4A161102AT24, Research in Snow, Ice and Frozen Ground). We thank the many individuals who also contributed to the success of the symposium and proceedings. We especially thank Shari L. Hennefer of the USDA, ARS, Northwest Watershed Research Center, Boise, Idaho, for assistance in proofreading/editing the individual papers, and Stephen L. Bowen of CRREL's Information Management Division for his assistance in producing this publication.

The papers published in these proceedings are in the order shown in the Symposium program. The authors submitted camera-ready copy to facilitate the publication of the proceedings prior to the conference.

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FROZEN SOIL IMPACTS ON AGRICULTURAL, RANGE, AND FOREST LANDS--AN INTRODUCTION

by

Keith E. Saxton, Gary E. Formanek and Myron Molnau[1]

INTRODUCTION

Soil freezing and thawing is a natural phenomena of any landscape in which the climatic conditions and physical setting are such that heat transfer results in a soil water phase change. Around the world, soil freezing takes on many patterns of cycles, depths, and severity ranging from permafrost to never occurring.

Frost in the soil is extremely important to all the sciences and activities which are represented on most landscapes such as the ecology, hydrology, agriculture, and climatology. But it is where man's activities rely on land production that we begin a special interest because now soil freezing may deny desirable results or significantly modify expected benefits. Plants may be torn from their roots, infiltration prevented, soil structure severely weakened, erosion magnified, and roads heaved.

Soil freezing impacts are particularly important on agricultural, range and forest lands where mankind gleans valuable food and fiber while desiring to preserve the sustainability of the land. It is on those lands were the climate is sufficiently cold to cause intermittent to permanent soil freezing but yet mild enough for plant production that soil freezing must be understood and the management options understood. Such lands represent a large percent of the worlds land mass found generally in a band from mid-latitude in both northern and southern hemispheres to often quite near the poles.

Large areas of cereal production, grasslands, and forests are included. Some 45% of the land of the United States is impacted by freeze-thaw (Formanek et al., 1990). This includes 1.2 million square kilometers of cropland, 1.3 million square kilometers of forestland, and 1.8 million square kilometers of grazing lands.

Measurements and documentation of soil freezing have been a part of natural sciences for a long time. The basic physics and interactions relative to agricultural production have been well documented for many circumstances around the globe, for example the observations of Post and Dreiblelbis (1942). Yet, detailed understandings, explanations and predictions are wanting today when we attempt to describe such experiences as the amount of infiltration and runoff from a melting snowpack or cold rain on a frozen or thawing soil, the effect of tree removal on two areas with different slopes, or fertilizer translocation after several freeze-thaw cycles.

Of particular interest is the management of agricultural soils where tillage is an option for manipulating the structure and porosity for liquids and air. Empirical evidence often has been obtained but the physical description and predictability are still quite lacking. Thus we continue to attempt to develop more detailed theories and experimentation with which to build our knowledge base.

The papers found in this publication present a broad summary of the current understanding about frozen soils and their impact on the production activities found on these lands. It is intended that the compiled information of these individual papers will provide a sound base from which to develop the current status of frozen soil impacts and launch additional research toward improved understanding and applications.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

[1]Research Agricultural Engineer, USDA Agricultural Research Service, Pullman, WA; Agricultural Engineer, USDA Soil Conservation Service, Portland, Oregon; and Prof. of Agricultural Engineering, Univ. of Idaho, Moscow, ID, respectively.

OCCURRENCES

Soil freezing is widely variable across regions, land forms, and management areas. At mid-latitudes, the shallowest soil layers freeze and thaw in an intermittent pattern throughout the winter. Here surface residues, soil moisture and snow cover can significantly impact the field-to-field variations of frost occurrences in both time and depth.

In more northerly climates, a single, deep freeze occurs most winters regardless of the field or forest treatment. Slight shifts in time and depth of freezing may occur, but usually this is not significant enough to attempt major management strategies. However, climatic variances such as radiation and snow depths may vary soil freezing patterns in these regions.

Still further north in harsher climates, permafrost is the rule and all but limited forest or rangeland production ceases and soil freezing is accepted as an unmanageable natural occurrence. Interesting exceptions occur in these northern regions such as those areas of Alaska where land naturally in forest and permafrost has been cleared for agriculture and now has an unfrozen period each year in the upper several meters which allows cereal production (Restad and McNicholas, 1983).

While much of the variation of soil freezing in all regions occurs as the result of the climatic input variability, man's activities can often impact the soil-atmosphere exchanges sufficiently to alter the occurrences and impacts. Vegetation or residue removal or configuration can dramatically change the radiative and aerodynamic heat exchange. Soil tillage will alter the soil's water and thermal characteristics (Kenny and Saxton, 1988). These shifts need to be understood to improve management methods.

MEASUREMENTS AND RESEARCH

Recent advances in electronics, sensors, data acquisition, and computers have enhanced the opportunity to more fully document and understand the frozen soil phenomena. The capability to simultaneously monitor the energy exchange between the atmosphere and a soil mass has greatly enhanced the understanding of the complete system. Multiple natural and created interface surfaces allow the separation and definition of influences by the major layers of the atmosphere, plant material, and soil mass (Campbell, 1977; Flerchinger and Saxton, 1989b).

Computer models have probably advanced the understanding and predictability of frozen soils more than any other single capability. It is now possible to represent a rather complete mathematical description of the energy and mass transfer physics within the landscapes such that quite realistic and accurate predictions of soil freezing depths and durations can be made (Flerchinger and Saxton, 1989a). These relatively complex recent developments demonstrate why earlier, more simplified approaches often provided only estimates for some circumstances and time periods (Cary et al., 1978).

Predictions with detailed models on todays computers can provide much needed insights about soil freezing impacts and management opportunities. While these soil freezing estimates are now acceptably accurate for many applications, the required input data are still often not readily available for the regions and locations of interest. Hard decisions will need to be made concerning the quantity and distribution of input data to be acquired for many frozen soil applications such as large scale flood predictions.

IMPACTS OF SOIL FREEZING

Soil freezing and thawing impact productive lands in many aspects which may be either positive or negative for the production of food and fiber. Soil morphology reflects the natural occurrence of freezing over the eons during which it developed; the associated plant community is that adapted to the expected freezing conditions; and the mean hydrologic occurrences reflect the impacts of freezing. Thawing of permafrost allows soil surfaces to dry which results in a wind erosion hazard; logging methods on fragile sites can be highly effected by soil freezing; rangeland trampling and winter access for livestock feeding can be impacted.

Land application of agricultural wastes must consider limited application periods to avoid impacting runoff quality. Methods such as residue and tillage management may provide longer application periods and better trafficability. Agricultural waste applications are becoming an even more important issue with the recent emphasis on low input, sustainable agriculture and organic matter management.

Soil erosion impacts on crop production is a serious problem and runoff on frozen soils is clearly associated with years of high soil loss (McCool and Molnau, 1974). Soil frost may greatly reduce the infiltration capacity of the soil through changes in hydraulic characteristics of the soil (Kane, 1980).

It is primarily when management by mankind is imposed on the landscape that we become most concerned with more fully understanding and manipulating the soil freezing of the land. The introduction of new plant species often finds them adversely impacted by soil freezing and heaving. Baring large acreages through cultivation, severe grazing and unwise logging creates more opportunity for shallow freezing, freeze-thaw weakening of the soil structure, and increased resource degradation through runoff and erosion. Downstream flooding can dramatically increase. Chemical applications and redistributions can become increasingly complicated and hazardous with the advent of soil freezing.

Defining soil freezing impacts on the soil system and its functions as related to plant-water-environmental relationships is of high importance in modern agricultural science of colder climates. Not only the occurrence of soil freezing must be predictable, but we must define the changes and impacts which result from that freezing as they relate to other natural process and production practices. With improved knowledge of both occurrence and impact of soil freezing, appropriate management choices can be made for resource utilization and conservation on our farms, rangelands, and forests.

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FROZEN SOIL, RUNOFF AND SOIL EROSION RESEARCH IN NORTHEASTERN OREGON

by

John F. Zuzel and J. L. Pikul, Jr.[1]

INTRODUCTION

Soil erosion is excessive on much of the nonirrigated cropland used to produce small grain east of the Cascade Mountains in the Pacific Northwest and in the northern parts of the Intermountain West. Average annual soil losses range from 5 to 50 t/ha but losses in excess of 200 t/ha have been reported (McCool et al., 1976). There is a strong association of frozen soils with increased runoff, flood producing runoff, soil erosion, and sediment production (Tigerman and Rosa, 1951; Storey, 1955; Johnson and McArthur, 1973). Rain on snow can contribute significantly to flood damage but may also contribute significantly to annual water yield in some low elevation basins. Severe winter flooding occurs frequently and has produced some of the highest peak discharges recorded at some runoff stations (Johnson and McArthur, 1973). These runoff events usually occur from December through March while the soil is frozen and play a major role in erosion and sediment production (Yen and Molnau, 1982; Zuzel et al., 1982).

The climate of these areas is characterized by cold, humid winters and dry, hot summers. Winter rainfall intensities are relatively low, usually less than 4 mm/hr (Brown et al., 1983). The winter climate is also characterized by a shallow, transient snowpack. Frozen soil is also transient and may freeze and thaw several times each winter (Zuzel et al., 1986).

Because of the adverse impacts of soil frost on soil erosion, sedimentation, and water conservation, the prediction and eventual modification of frost formation and thawing has been identified as a major research need by the Soil Conservation Service. In response to these concerns, a research program addressing soil frost was begun at the Columbia Plateau Conservation Research Center beginning in 1979. The objectives of this research were to: 1. predict depth and duration of soil freezing and thawing; 2. quantify the effects of tillage and residue management on depth and duration of soil frost; 3. predict the effects of both extended freezing periods and diurnal freezing and thawing on development of low permeable soil crusts; 4. relate the importance of soil frost to runoff, erosion, and sediment production for field sized areas; and 5. validate soil moisture conservation responses to repeated diurnal freezing and thawing. The purpose of this paper is to briefly review the progress of this research program.

SOIL FROST PREDICTION

Extended freezing

We simulated the presence or absence of soil frost for a thirty year period for sites in northcentral Oregon using a physically based frost model described by Cary (1982). Model inputs are daily maximum and minimum air temperature, snow depth, and solar radiation. The model predicts net daily heat flux across the soil surface. Cumulative daily soil heat flux is then used to predict the presence or absence of soil frost. Negative values of soil heat indicate the presence of soil frost while positive values indicate an unfrozen soil. The model was modified to predict solar radiation from potential radiation, slope, aspect, and daily temperature range. The model was validated using data from field research sites near Moro, Oregon. Measurements of plot runoff, erosion, weather variables, soil temperatures and soil frost were collected from a field planted to winter wheat each year for a 5 yr period. The model correctly predicted the presence or absence of soil frost 80 percent of the time for the 5 yr calibration period (Table 1 and Figure 1). Once the model was calibrated, 30 yr of soil frost were simulated for the Moro weather station. The results showed that the soil was frozen from 6 to 116 days per year with an average of 57 days per year. The number of freeze-thaw cycles varied from 1 to 7 cycles per year. The simulation and analysis has been accomplished for other sites in northcentral Oregon.

Presented at the Frozen Soil Symposium, March 21-22, 1990, Spokane, Washington.

[1]Hydrologist and soil scientist, USDA-ARS, Columbia Plateau Conservation Research Center, P.O. Box 370, Pendleton, Oregon. Joint contribution of OSU and USDA-ARS. OSU Tech. Paper No. 8954.

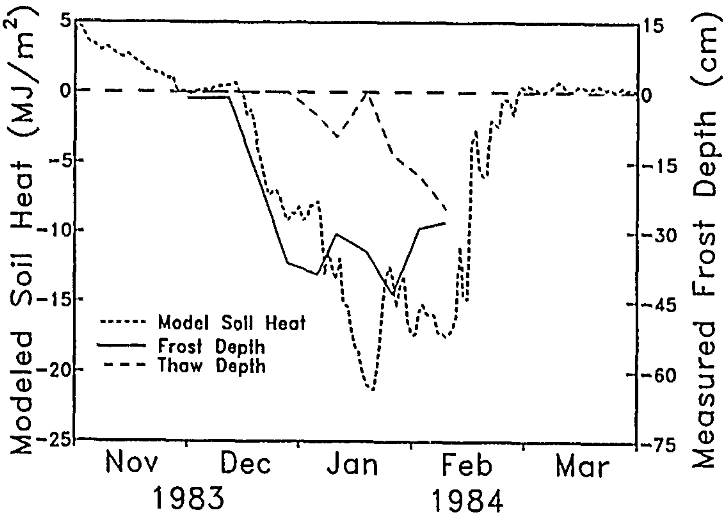

Figure 1. Soil frost simulation model output and measured frost for Moro, Oregon. Modeled cumulative soil heat values < 0 predict the presence of soil frost.

Table 1. Characteristics of sites near Moro, Oregon used to verify the frost prediction model.

Site and winter number	Elevation (m)	Slope (percent)	Aspect (deg)	Prediction (percent)	Distance from Moro weather station (km)
1	561	14	S40W	83	0
2	567	11	S61W	82	0.5
3	539	24	N65E	91	2
4	539	25	N65E	63	2
5	539	24	N65E	80	2

Prediction column is the percent of days between 1 November and 31 March in which frost occurrence was correctly predicted.

Diurnal freezing and thawing

Repeated freezing and thawing may accelerate surface crusting, resulting in decreased water infiltration and erosion resistance. In addition, diurnal freezing and thawing is a wasteful natural mechanism in terms of conserving soil moisture because freezing induces water flow to the soil surface resulting in high evaporative water loss during subsequent thawing periods. A finite difference numerical model was developed and used to simulate soil temperature, depth of freezing, and water movement. A 7 hr freezing period followed by a 12 hr thawing period was simulated. The results of the simulation were compared with measured depth of soil freezing, water content, and temperature. Simulated ice and water content of the top 1 cm of soil compared favorably with the measured value. Simulated maximum depth of freezing also closely approximated the measured value. Details of the model and results are given by Pikul and Zuzel (1990).

TILLAGE AND RESIDUE IMPACTS

Tillage and crop residue can both significantly affect frost formation and thawing in agricultural soils (Benoit and Mostaghimi, 1984). Soil frost was measured at a site near Pendleton, Oregon under several tillage and residue treatments over a 4 yr period. Residue and tillage treatments included standing stubble, fall plow, fall chisel, straw mulch, bare surface, and winter wheat. Frost depths during freezing weather were determined by hand sampling and duplicate frost tubes (Ricard et al., 1976) in each treatment. In addition to meteorological variables, frost depth, snow depth, and soil temperatures, we measured soil heat flux at the 1-cm depth during selected time periods. We used duplicate sensors, each of which averaged the signal from 2 heat flux plates. The maximum frost penetration occurred in the bare surface treatments for all 4 yr, while the minimum occurred in the standing stubble (Pikul et al., 1986). Frost depth and snow cover for the 1983-1984 winter are shown in Figure 2. The persistence of soil frost

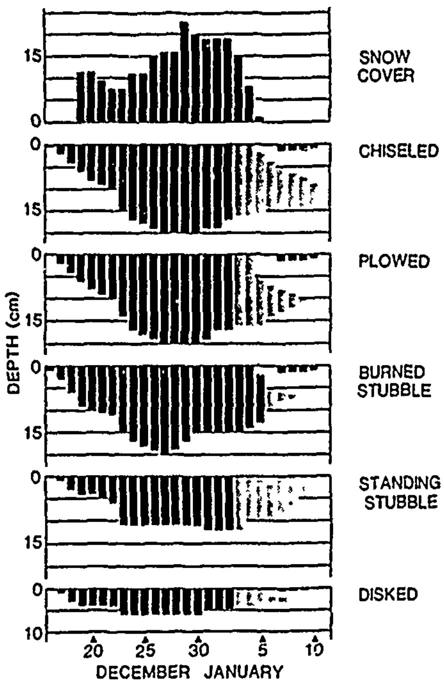

Figure 2. Frost depth and snow cover for 5 tillage treatments during the 1983-1984 winter. Lighter colored bars indicate a porous concrete frost.

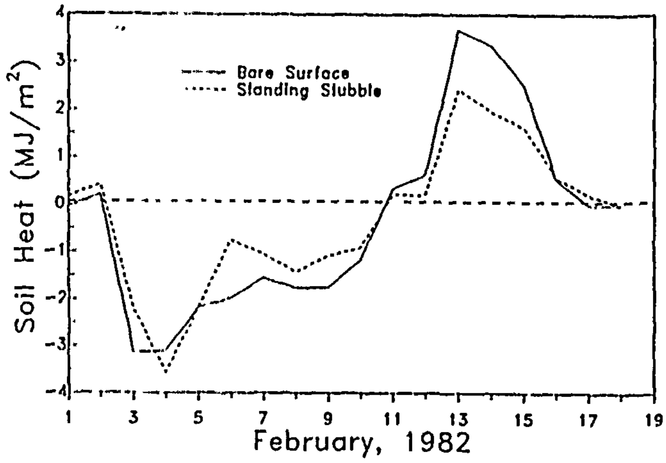

Figure 3. Soil heat flux for a thawing period of clear skies, high radiation inputs and warm air temperatures.

during thawing weather is linked to both surface residue cover and weather conditions during the thaw. Average heat flux in the bare surface treatment was 40% greater than in the standing stubble treatment in 1982 when clear skies and warm air temperatures prevailed (Figure 3). Because of greater heat input in the bare surface treatment, both the standing stubble and bare surface treatments thawed the same day, even through the bare surface treatment was frozen to a greater depth. During 1980 the weather was characterized by lower air temperatures and cloudy conditions. Soil heat flux in the bare surface treatment was only 20% greater than the standing stubble (Figure 4). Under these conditions, the stubble treatment thawed 4 days earlier than the bare surface treatment.

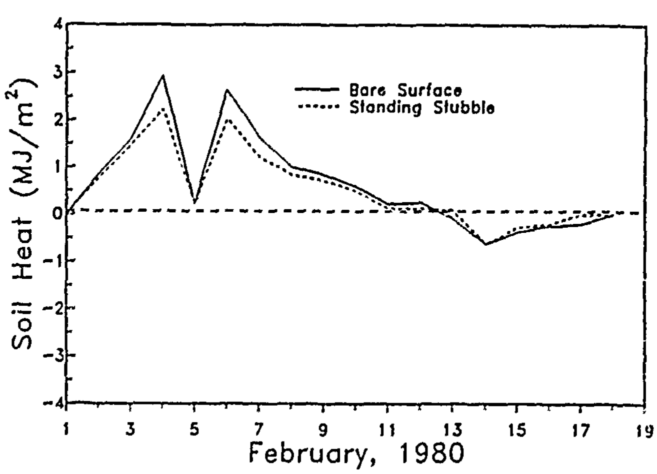

Figure 4. Soil heat flux for a thawing period of cloudy skies, low radiation input and cool temperatures.

RUNOFF AND EROSION

Beginning in 1979 and continuing for 5 yr, we obtained soil erosion and runoff measurements from widely scattered sites in the dryland small grain producing area of northeastern Oregon. The purpose of this research was to evaluate the climatic, physiographic, cultural, and soils factors responsible for producing runoff and erosion. Each site was equipped with holding tanks to collect all runoff and sediment from duplicate bordered plots. Each site was equipped to provide continuous recordings of meteorological variables and soil temperatures as well as plot runoff. The data in Table 2 illustrate the importance of frozen soil in the runoff and erosion process as well as the influence of snowmelt and rain on snow. The data also show the extreme site to site variability, even though all the sites are located in the same general physiographic area. Sources of this variability are slope, aspect, soil type, and depth and duration of the snowpack (Zuzel et al., 1982).

Table 2. Soil and climatic conditions responsible for November-March measurable runoff and erosion from experimental plots in northeastern Oregon, 1979-1984.

Condition	Number of events in five years			
	Site A	Site B	Site C	Site D
Frozen soil, snowmelt	8	2	2	1
Frozen soil, rain and snowmelt	4	4	2	2
Frozen soil, rain	1	1	3	0
Thawed soil, snowmelt	2	1	0	0
Thawed soil, rain and snowmelt	2	1	1	0
Thawed soil, rain	11	5	3	0

Locations of sites were: A, Wasco county; B, Sherman county; C, Gilliam county; and D, Morrow county.

The importance of snowmelt was evaluated from the site data by partitioning the observed snowmelt hydrographs into the individual melt components using an energy balance model (Zuzel et al., 1983). This procedure identified two distinct types of snowmelt. The first type of snowmelt event is typified by warm air temperatures, a largely negative energy balance, and low turbulent transfer rates as the result of low windspeeds. The snowmelt produced by these conditions is refrozen in the pack, and evaporation is dominant in ablation. Since no runoff is produced, this scenario is unimportant in runoff and erosion, but is important from the standpoint of water conservation because of the potentially large evaporative losses. The second type of event is characterized by rain on snow, high windspeeds providing an effective turbulent transfer mechanism, and high dewpoint temperatures resulting in condensation at the snow surface. These conditions accelerate the snowmelt rate beyond that expected from radiation, sensible heat, and advected heat. These conditions are associated with the rapid intrusion of warm, moist Pacific air masses and if the soil is frozen can be a major factor in accelerated runoff and erosion. Because these soils usually freeze at a high moisture level, the infiltration capacity is severely curtailed or nonexistent and nearly all precipitation and snowmelt runs off.

A method to quantify daily rainfall amounts associated with different antecedent conditions was developed and the relationships between daily rainfall, frozen soil, snow on the ground, and runoff were investigated (Zuzel, 1986). The procedure incorporates the soil frost model described under the extended freezing section with an algorithm for extracting rain and rain-on-snow events in the presence or absence of frozen soil from 30 yr weather records. A flowchart of the algorithm is shown in Figure 5. Outputs from this procedure are probability distributions of rain and rain on snow in the presence or absence of soil frost. An example of these distributions is shown in Figure 6. The events produced by the algorithm were correlated with crest gage data for a site in northcentral Oregon. This analysis provided an estimate of the precipitation amount necessary to produce runoff. The crest gage had been in operation 20 yr and of the 20 peak discharges recorded, five years had no flow, one peak occurred in June and 14 occurred during the November-March period. The analysis showed that eight of these 14 events were the result of rain on snow while the remaining six were the result of rain with no snow cover. Moreover, the soil was frozen during seven of the eight rain-on-snow events and one of the rainfall events. Regression analysis of the available variables indicated that daily precipitation amount and snow depth jointly accounted for 73% of the variance in runoff peaks.

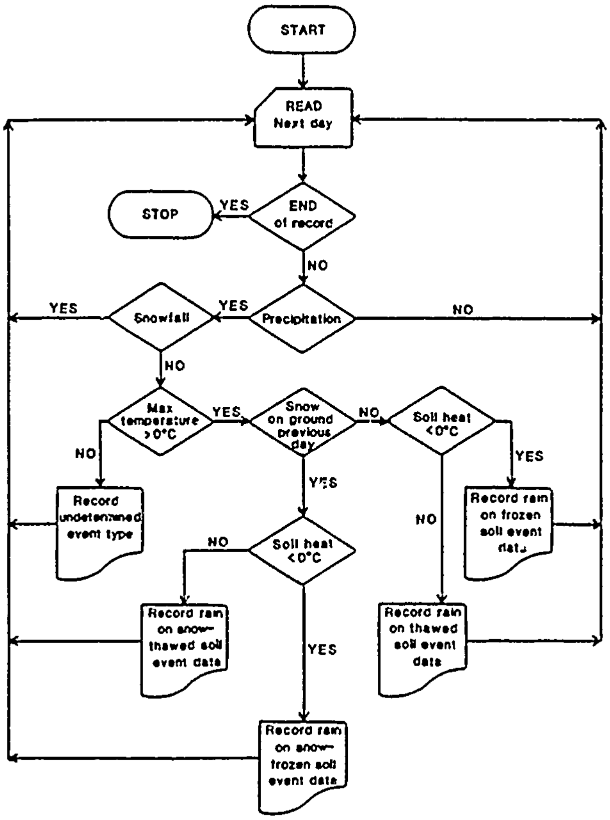

Figure 5. Flowchart of the algorithm used to classify type of precipitation event from daily weather records and cumulative soil heat.

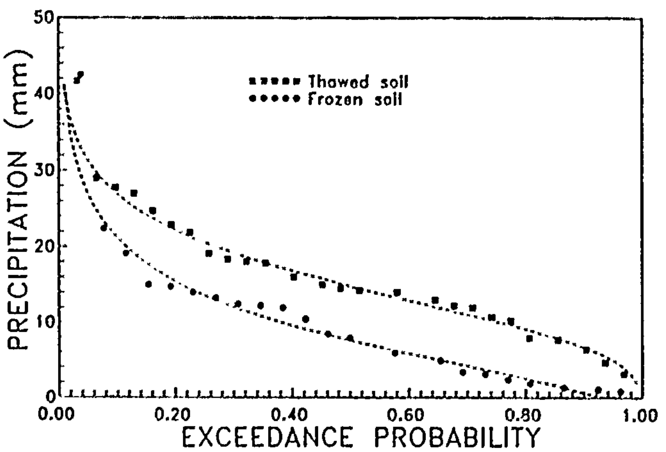

Figure 6. Cumulative frequency and fitted theoretical distributions of the annual series for rain on thawed soils and rain on frozen soils in the absence of snow cover.

INFILTRATION

We conducted a study designed to determine the effects of three tillage treatments on water infiltration. The study was done near Pendleton, Oregon on a Walla Walla silt loam soil (Typic Haploxeroll, coarse-silty, mixed, mesic). We measured infiltration rates using a Palouse rainfall simulator (Bubenzer et al., 1985) before, during, and after a period of frozen soil on three tillage treatments. Infiltration rates were compared for standing stubble, chiseled stubble, and winter wheat (Triticum aestivum, L.) after summer fallow. The tests were conducted in the fall after seeding the winter wheat and chiseling the standing stubble, in the winter when the soil was frozen, and in the spring when the soil was thawed. The experimental area was in a summer fallow-winter wheat rotation. The tillage sequence of this cropping system was spring moldboard plow, three rod weedings during the summer, and seeding in late October. Soil bulk density measurements in the standing stubble indicated a maximum bulk density at 17 cm, about the depth of plowing. A chiseling depth of 25 cm was chosen in order to fracture this high density layer and create zones of high water flow through the compacted surface layers of soil. The results of the three infiltration tests are show in Table 3. For the fall and spring runs, the chiseled stubble had the highest infiltration rate, followed by the standing stubble and the winter wheat. The final infiltration rates for the three treatments in fall and spring did not differ appreciably. Under frozen soil conditions, some infiltration occurred in both the standing stubble and chiseled stubble, but no infiltration occurred on the winter wheat treatment (Figure 7). The results also showed that the beneficial effects of chiseling on water infiltration were not diminished over the winter recharge period. Details of this study are given by Zuzel and Pikul (1987).

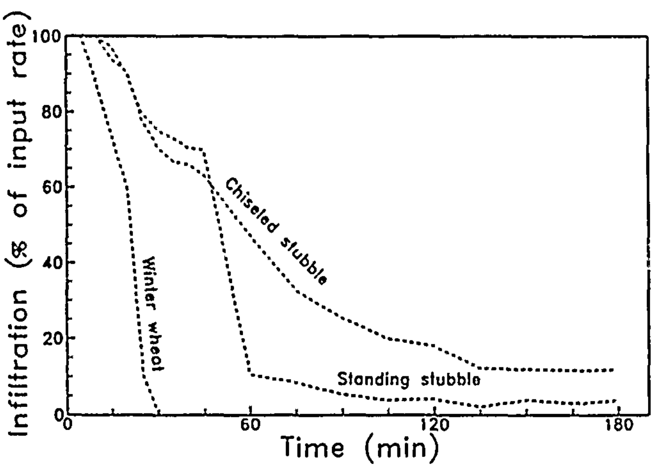

Figure 7. Measured infiltration for the three tillage treatments in January, 1985.

Table 3. Final infiltration rates in mm/h by treatment and season.

	------------- Season -------------		
Treatment	Fall	Winter	Spring
Winter wheat	3	0	3
Chiseled stubble	11	6	16
Standing stubble	7	1	8

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NATURE OF THE CRYIC THERMAL REGIME OF AGRICULTURAL SOILS IN THE YUKON TERRITORY, CANADA

by

C.A.S. Smith[1]

INTRODUCTION

The Yukon Territory is located in northwestern Canada between 60 and 69 degrees north. It is bounded by the state of Alaska to the west and the province of British Columbia to the south (Figure 1). The Territory occupies the northern extension of the western cordilleran, being an area of north-south trending mountain ranges with broad plateau areas in between.

Much of the southern and central portions lies in a rainshadow created by the coastal St. Elias range. Whitehorse receives annual precipitation of 261 mm and has a mean annual temperature of -1.2 degrees C (Table 1). Dawson receives 306 mm of precipitation and has a mean annual temperature of -5.1 degrees C (Atmospheric Environment Service, 1982). Dawson has a pronounced continental climate; winters at Dawson are significantly colder than in Whitehorse but summers are warmer. Agricultural conditions are more favourable in the Dawson area with over 1000 growing degree days (GDD >5°C) as compared with approximately 900 GDD in the Whitehorse area (Wahl et al., 1987). Air temperatures, therefore, limit agricultural activities to the production of forage crops and cold-hardy vegetables.

Table 1. Climatic conditions of the two study areas.

Location	Elev. (m)	Latitude	Longitude	Mean Temp.(°C) Annual	Mean Temp.(°C) Summer	Precipitation (mm)	GGD>5°C	FFP (days)
Dawson	320	64°03'N	139°26'W	-5.1	14.2	306	1015	91
Whitehorse	703	60°43'N	135°04'W	-1.2	12.1	261	890	82

Both Dawson and Whitehorse lie within the discontinuous permafrost zone (Brown, 1978), but there is no permafrost within the control sections of the soils monitored. The soils are weakly weathered, lack organic matter in their surface horizons and have alkaline to neutral reactions (Tarnocai, 1987). In the Whitehorse area, cultivation is most common on glaciolacustrine and fluvial sediments in the Takhini and Yukon valleys. In the Dawson area, cultivation is most common on the alluvial soils of the Klondike Valley.

The objective of this paper is to present data on the thermal regime of soils used for agricultural purposes in northern Canada. Data recorded at two sites over a period five years are analyzed for soil climate properties. Data from a third site recorded on a daily basis over a nine month period are examined to describe seasonal effects.

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Table 2. Comparison of soil climate parameters at 50 cm depth of Yukon sites with two sites from Alaska and southern Canada.

Location	MAST	MAAT[1]	MSST	Degree Days >5°C		Soil Thermal Regime	
				Soil	Air	U.S.	Canada
Dawson, YT	0.9	-5.1	7.4	269	1015	Cryic	Very Cold
Whitehorse, YT	-0.1	-1.2	4.1	13	890	Pergelic	Very Cold/ Extremely Cold
Fairbanks, AK	2.1	-2.7	9.5	--	--	Cryic	Cold
Palmer, AK	3.0	1.3	11.2	--	--	Cryic	Cold
Swift Current, SK	6.1	3.3	16.4	1576	1675	Frigid	Cool
Ottawa, ON	8.8	5.9	17.5	1841	2072	Mesic	Mild

[1]MAAT - mean annual air temperature

Table 3. Soil climate parameters based on best-fit curves for three depths.

		0°C			5°C			15°C		
Location	SEE[1]	Thaw	Freeze	Days >0°	Spring	Fall	Days >5°	Spring	Fall	Days >15°
Dawson										
10 cm	2.94	Apr.29	Oct.7	161	May 19	Sept.18	122	July 7	July 30	23
20 cm	2.37	May 4	Oct.13	163	May 25	Sept.20	118	--	--	0
50 cm	1.47	May 12	Nov.10	182	June 13	Sept.25	104	--	--	0
Whitehorse										
10 cm	2.06	Apr.26	Oct.15	172	May 26	Sept.17	114	--	--	0
20 cm	1.59	May 3	Oct.22	172	June 10	Sept.18	100	--	--	0
50 cm	1.12	May 23	Nov.4	165	July 23	Sept.2	41	--	--	0

[1]SEE - standard error of estimate, a measure of variation of observed temperature about the estimated regression curve. Higher values indicate greater range in possible dates when specified temperatures are reached.

METHODS

All atmospheric climate data presented in this paper are from the 30 year normals for Dawson and Whitehorse as reported by Atmospheric Environment Service (1982). All soils are classified and soil properties described according to Agriculture Canada Expert Committee on Soil Survey (1987) for Canadian taxonomies and Soil Survey Staff (1975) for U.S. equivalents.

Soil temperatures in two pedons representative of Yukon agricultural soils were monitored between September 1984 and July 1989. At Dawson, the soil belongs to the Klondike soil association (Rostad *et al.*, 1977), and is formed on fluvial parent material of silt loam overlying gravelly sandy loam. The soil is well-drained and classified as an Orthic Eutric Brunisol (Typic Cryorthent). This soil was initially underlain by permafrost prior to its clearing and cultivation 30 years ago. A soil belonging to the Shaneinbaw association (Day, 1962) was monitored in similar fashion near Whitehorse. The soil is imperfectly drained, formed on silty clay loam glaciolacustrine parent material and is classified as Humic Gleysol (Typic Cryaquept). At both sites the perennial forage crop bromegrass (*Bromus inermis* Leyss) is grown.

Soils were instrumented with YSI 44004 thermistors (calibrated to ±0.1°C) at depths of 2.5, 5, 10, 20, 50 and 100 cm. Measurements were taken with a 3.5 digit multimeter on a monthly basis during the winter and somewhat more frequently during the summer months. Readings were usually obtained during the late morning. These data were analyzed according to the method of Hayhoe *et al.* (1987) and Mills *et al.* (1977). The normal annual soil temperature cycle for the depths 10 through 50 cm were fit as a function of time to a Fourier series curve with two harmonics (Hayhoe *et al.*, 1987; Eqn. 2). Values were calculated for mean annual soil temperature (MAST); mean summer soil temperature (MSST), i.e. mean of June, July and August readings; and days and degree days above 0, 5 and 15 degrees C.

A third site was also instrumented near Whitehorse, belonging to the Lewes soil association. The soil is formed on fine sandy loam glaciolacustrine parent material, is well drained and classified as an Orthic Eutric Brunisol (Typic Cryocrept). Thermistors were installed at depths of 10, 20, 40, 60, 80, 100, 120 and 140 cm. Unshielded thermistors were also located 1.5 m above ground to record air temperature and at 5 cm above ground to record snow pack temperature during the winter months. Readings taken each day at midnight were recorded on a Campbell Scientific CR-10 datalogger. The period of monitoring ran from November 1, 1988 to July 30, 1989. These data were analyzed for seasonal trends in cooling and thawing.

RESULTS AND DISCUSSION

Soil climate monitoring

Annual soil temperature curves for the 50 cm depth are given for the Klondike and Shaneinbaw soil in Figure 2. Curves are based on mean monthly values which peak at both sites in August and reach a minimum in March. The continental climate at Dawson is reflected in cooler winter soil temperatures and warmer summer temperatures. However, the MAST of these two sites are very low for agricultural soils (Table 2). The Shaneinbaw soil with a MAST of <0 degrees C would have a pergelic thermal regime. In contrast, soils near Ottawa, Ontario, and Swift Current Saskatchewan, have MAST between 6 and 9°C. Further comparison is noted in the growing degree days >5°C (GDD>5). Although there are

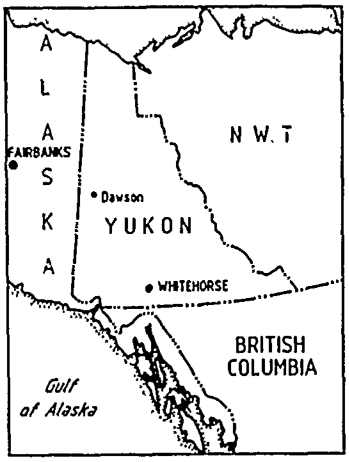

Fig.1 Location of study sites in
Yukon Territory.

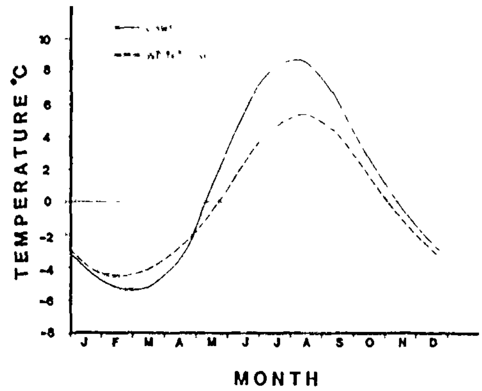

Fig.2 Annual soil temperature curves for
the 50 cm depth.

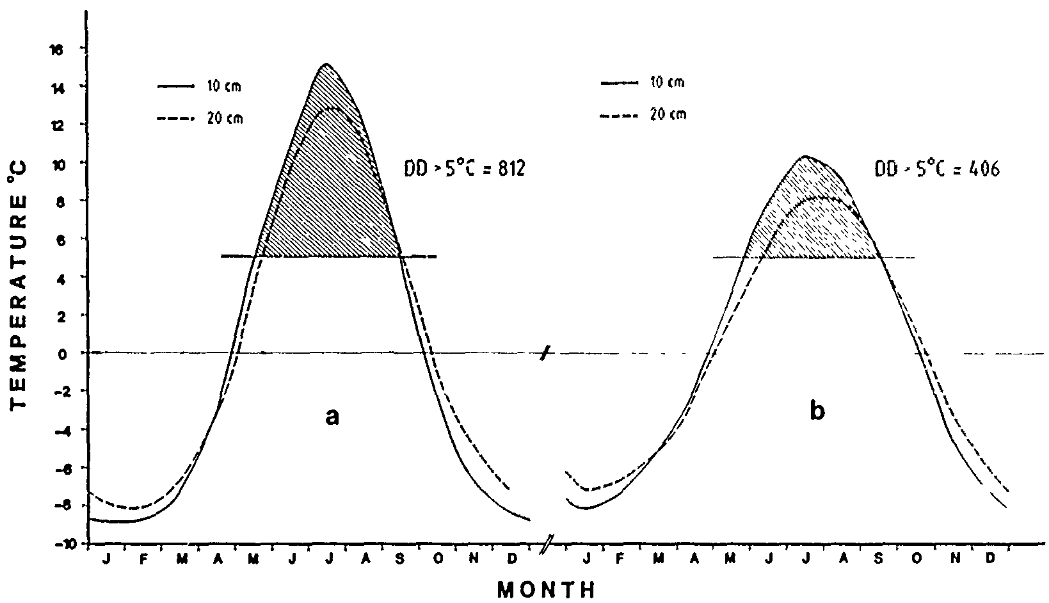

Fig.3 Soil temperature curves for the 10 and 20 cm depths for
a) Klondike soil at Dawson and b) Sheneinbaw soil at Whitehorse

1015 GDD>5 at Dawson this is recorded as only 269 GDD>5 in the soil at 50 cm depth. Swift Current with 1675 GDD>5 has over 1500 GDD>5 at this depth and is presumably a reflection of the much colder subsoil temperatures at the Yukon sites.

In comparison with Alaska agricultural soil temperatures the MAST for the Dawson and Whitehorse sites are only slightly cooler than those at Fairbanks and Palmer. Ping (1987) reports values of 2.1 and 3.0 respectively at 50 cm depth (Table 2).

The nature of soil temperatures during the growing season is shown for the two sites in Figure 3. The temperatures in the plow layer (10 and 20 cm) at the Dawson site reach a maximum of 15.5°C (based on mean monthly values) although individual values as high as 22°C have been recorded occasionally at this depth. Soil temperature in the plow layer is less at the Whitehorse site with maxima of only 10°C. However, individual readings as high as 20°C have been recorded at this site. While the length of time that the plow layer (10 cm depth) remains thawed is greater in the Shaneinbaw soils than in the Klondike soil (172 versus 161 days), the number of days when the soil temperature is above 5°C as well as the GDD>5 are greater in the Klondike soil (Table 2 and 3). The 5°C soil temperature value is known to be important for germination and biological growth. Hence the effective growing season within the Klondike soil is greater as shown by the hatched areas in Figures 3a,b. The higher moisture regime of the Shaneinbaw soil contributes to its generally cooler thermal regime.

Soils seldom reach temperatures over 15°C for any length of time (Table 3). The soils remain frozen for more than half the year and do not begin to thaw until the end of April. Surface layers return to the frozen state by mid October. The effect of cold deep ground temperatures is evident in the Shainebaw soil from Whitehorse. Cooling occurs from below at the 50 cm depth in the fall. An inverse cooling trend is observed in the fall dates at which soil temperatures drop below 5°C.

Both these soils at the field sites are irrigated and, with ample fertilization, produce up to 6,000 kg/ha/yr of grass hay. These annual curves may represent the minimum soil temperatures which can support commercial field crops. Agriculture has not yeat proven to be viable in locations at higher elevation or further north than the sites reported here.

Seasonal monitoring

The monitoring of soil temperatures on a daily basis was undertaken at a second Whitehorse site on the Lewes soil association. This soil has formed on similar parent material about 10 km east of the Shaneinbaw site. The well-drained soil remained fallow during the period of monitoring.

Figure 4 presents a summary of the thermal regime. Comparison of the geotherms with the annual curves for the Shaineinbaw soil indicate that the winter temperatures recorded were near the monthly mean values plotted in Figures 2 and 3 but that the summer temperatures were higher. These higher summer temperatures are a result of the bare dry soil surface and an unusually warm July in 1989, approximately 2°C above normal.

The annual curves indicate that cooling begins in late August. The freezing front had progressed to a depth of 40 cm when monitoring began on November 1. The thermal gradient is small during the early winter with only 3 to 4°C difference in temperature between the 10 and 140 cm depths. Cooling progresses at all depths until mid February when the minimum temperatures were recorded at 10 cm. Minimum temperatures were recorded at 140 cm in April and May when the freezing front progressed to depths greater than the deepest sensor.

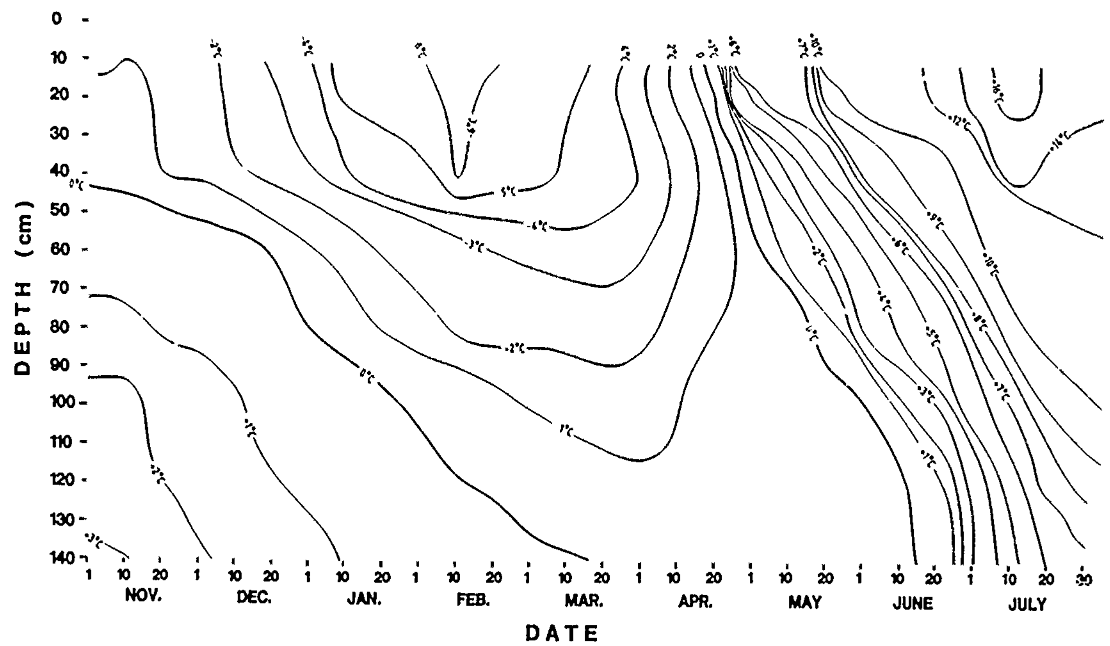

Fig.4 Geotherms showing seasonal temperature distribution in the Lewes sandy loam soil at Whitehorse.

Warming begins rapidly towards the end of April once the winter snow pack has melted and almost isothermal conditions are reached in mid May. The difference in the nature of the cooling and warming rates can be largely explained by the presence of a large snow pack which insulates the soil from the effects of cold winter air temperatures. There is no summer insulating factor and temperatures over 16°C were observed at the time of midnight recording in July.

In order to examine more closely the cooling phenomena during the winter one extreme "cooling event" was analyzed to calculate maximum cooling rate, temperature lag, and the thermal diffusivity within the surface meter of soil.

The effect on soil temperatures of an intense cold spell is presented in Figure 5. The air temperature dropped from -2°C on January 28 to a minimum of -33 on February 2 and warmed back up to -2.5 on February 9th, 1989. Temperature curves for the snow pack and four soil depths are plotted and the slopes of the cooling portion of the curves highlighted. Note that the slope angles (cooling rates) are greatest near the surface and decrease with depth. The lag effect is shown by the dates at which the coldest temperatures are recorded at each depth. Minimum temperatures were recorded on February 3 in the snow pack, February 7 at 10 and 20 cm and on February 9 at 40 cm. No distinct minimum beyond the seasonal cooling was noted at 80 cm or deeper.

The rates that were observed during this cooling event are taken as maximum rates for the season and are plotted on Figure 6. The snow pack cooled at a rate of almost 1.4°C/day over a period of four days. Within the soil, rates of between 0.6 and <0.2°C/day were observed. These values are larger by an order of magnitude than the

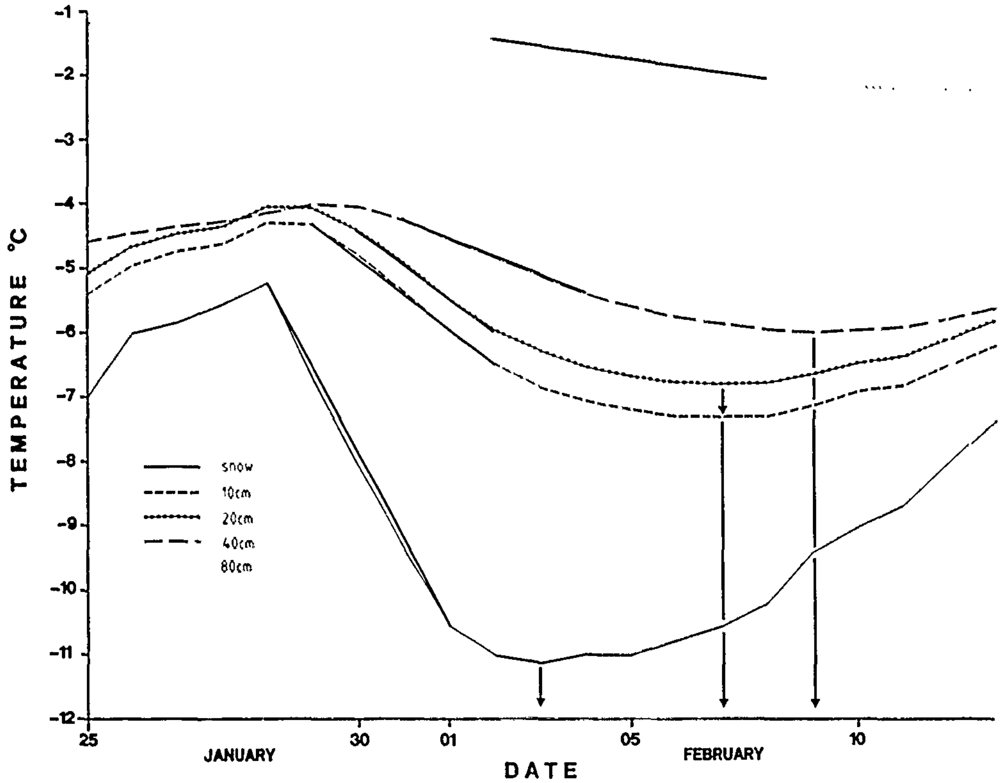

Fig.5 Curves showing temperature drop at 5 depths in the Lewes soil during severe cooling event, January-February 1989. Slopes of cooling trends are indicated in bold and values are represented in Figure 6 as maximum cooling rates.

values taken as seasonal cooling rates. These were calculated as the total temperature drop between November 1 and the date at which the annual minimum temperature was achieved at each depth. A seasonal value of 0.06°C/day was calculated at 10 cm depth and 0.025°C/day at 100 cm and deeper.

Warming trends in the soil between April 20 and July 15 indicated that surface soils (10 cm) warm at a rate of approximately 0.20°C/day and soils at the base of the rooting zone (50 cm) warm at the rate of approximately 0.12°C/day. Hence seasonal warming rates are larger, by an order of magnitude, than seasonal cooling rates.

Thermal diffusivity was calculated using the equation $K=z^2/2wt^2$ as described in Burn and Smith (1988). Thermal diffusivity, K, is the ratio of thermal conductivity to volumetric heat capacity and thus is an index of the change in temperature with time and depth in the soil (Armson, 1977). It is calculated as a function of the thickness, z, of the lag interval (m), the angular velocity of the wave, w,(s^{-1}) and the lag time, t, (s). Using the depth interval of 0.30 m between the readings at 10 and 40 cm, and w equal to 6.05×10^{-6} based on a wavelength of the cold period of 12 days (the period between A and A' on Figure 7), the thermal diffusivity was calculated as 2.4×10^{-7} m^2s^{-1}. This is less than the value of 2.5×10^{-7} m^2s^{-1} reported for frozen peat in the central Yukon (Burn and Smith,

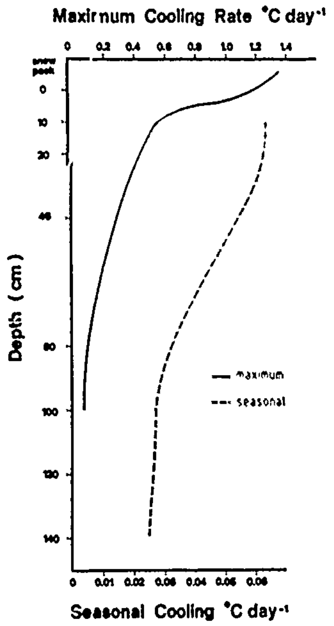

Fig.6 Cooling rates through the Lewes soil profile for the period November 1988 to March 1989.

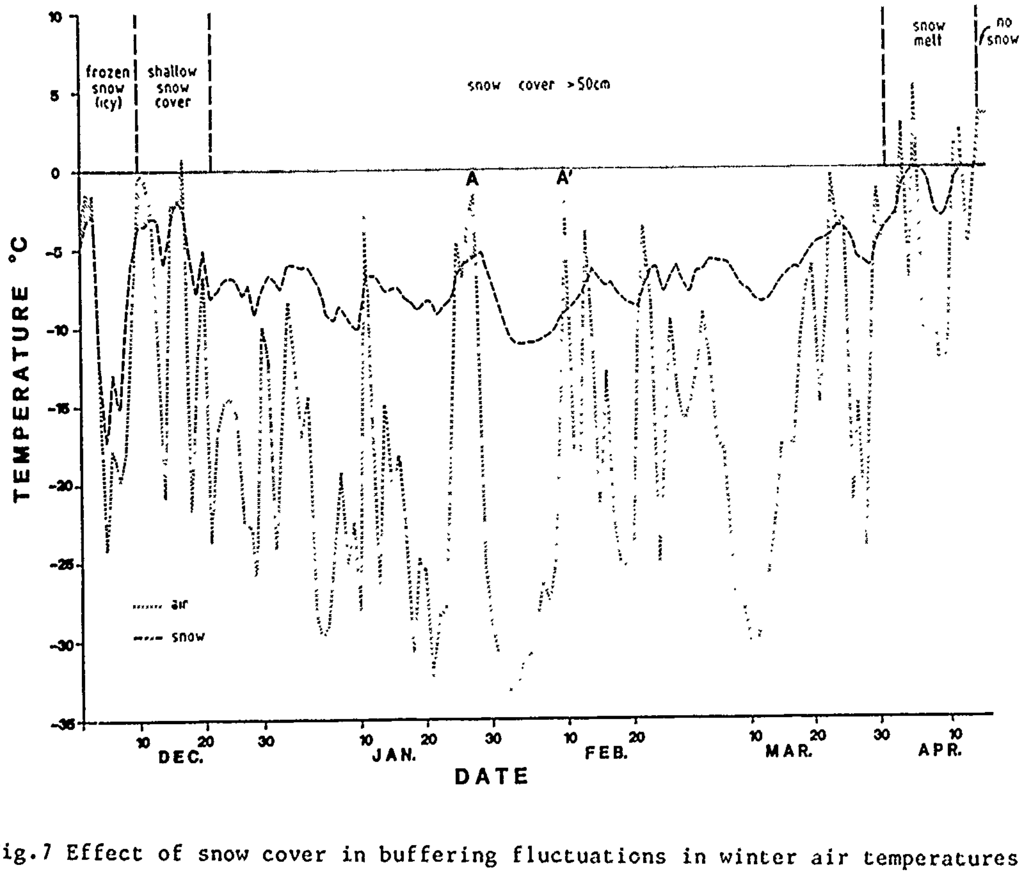

Fig.7 Effect of snow cover in buffering fluctuations in winter air temperatures.

1988) but equivalent to the value reported by Nakshabandi and Kohnke (1965) for unfrozen dry loam. Thermal diffusivity has been shown to increase with decreasing temperature from -1 to -50°C (Farouki, 1981). The dynamic nature of the property is attributed mainly to the considerable changes in the unfrozen water content through this temperature range.

The temperature buffering effect of snow cover and the condition of the snow pack is illustrated in Figure 7. Whitehorse winter temperatures are highly variable. The standard deviation of 6°C for January mean temperatures for south central Yukon are the largest in Canada (Wahl et al., 1987). Midnight air temperatures varied through the period of monitoring from +1°C to a low of -34°C. The equivalent snow pack temperatures during that same period only varied between -3 and -10°C. Most of the winter the snow pack temperature is greater than the air temperature. However, when air temperature rises quickly during incursions of maritime air into the Whitehorse area, the snow pack may temporarily remain colder than the air temperature. These intervals occurred about a dozen times during the winter.

Maximum insulating effect is felt when the snow pack is thick (>50 cm) and of low density. A thaw in November left the early season snow pack in a dense icy condition which provided little insulation. In this condition the snow pack temperature reached almost as low as the air temperature. As the snow melted in April, the snow pack temperature approached 0°C and upon melting the surface soil was heated and cooled directly according to air temperature (Figure 7). Ping (1987) concluded that observed warmer winter soil temperatures in Fairbanks relative to Palmer, Alaska were the result of the ephemeral nature of the snow pack at the climatically warmer Palmer site. Similarly at Whitehorse and Dawson, colder soil temperatures could be expected in winters with less snow.

CONCLUSIONS

1. Soils with very cold (cryic) and extremely cold (pergelic) soil thermal regimes still have value as media for agricultural production. A MAST of -1.0 and MSST of 4°C at 50 cm depth after cultivation may represent the minimum temperature necessary to sustain annual commercial crop production.

2. The soils studied have MAST and MSST values up to 8°C colder than soils from the traditional areas of agriculture in Canada but are roughly similar to agricultural soils in the central interior of Alaska.

3. The general thermal properties and thermal responses of these northern soils are no different than soils of more temperate environments. Colder ground temperatures below the soil control section seem to inhibit subsoil warming.

4. Slight differences in the continentality of the climate at Dawson and Whitehorse are reflected in the soil climates at each location. The soils in the Whitehorse area are subject to long slow periods of cooling (0.02-0.06°C/day) lasting eight months of the year and short, intense periods of warming (0.1 to 0.2°C/day) lasting three to four months of the year.

5. Snow has an extremely important effect on MAST and on the cooling, thawing and minimum temperatures experienced in cleared soil in any given year.

ACKNOWLEDGEMENTS

The author wishes to thank H.N. Hayhoe, Agriculture Canada, Ottawa, for undertaking the analyses of monthly temperature data and C.R. Burn, Department of Geography, University of British Columbia, for his critical review of this paper.

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SOIL FREEZING IN A SUBARCTIC DECIDUOUS FOREST

by

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INTRODUCTION

The soil hydrologic and thermal regimes are dynamic systems which change not only in response to the meteorologic control at the surface but also change due to the coupling that exists between them. The hydrologic and thermal regimes interact to such an extent during phase change that one can not be understood without considering the other. This interaction is manifested in the rate and direction of soil moisture movement, rates of temperature change and rates of soil freezing.

We studied the process of freezing in the field in order to compare techniques of measuring and calculating depths and rates of freezing. Our intensive analysis of a site-specific data set allowed comparison of some commonly accepted methods and some potentially useful new approaches. The changes in the thermal and hydrologic variables which accompanied soil freezing were monitored during the 1987-1988 winter.

EXPERIMENTAL PROCEDURE

The selected site was a subarctic deciduous forest on a north-east facing slope near Fairbanks, Alaska. This site is well drained in a non-permafrost area with a 10% slope. The soils are Fairbanks silt loam and consist of 5 cm of highly organic soil overlying 10 m of very uniform loess. To enhance coupled heat transport, the soils were irrigated for several weeks prior to freezing in mid-October to increase soil moisture to 40% by volume in the top 1 m. These soils have relatively high percolation rates and by November the moisture had drained to about 30%. Primary vegetation is birch (Betula papyrifera), with a typical understory of horsetails (Equisetum spp), bluejoint grass (Calamagrostis cavadensis) and scattered alder (Almus crispa).

The site was instrumented with soil thermistors in small increments near the surface, at 10 cm above the surface, 0, 10, 20, 30, 40, 60, 80, 100, 125, 150, and 240 cm below the surface and in greater increments as the depth increased, 0.6, 1.6, 2.6, 3.6, 4.6, 5.6, 6.6, 7.6, 9.6, and 11.6 m. The air temperature, relative humidity, net radiation, and soil surface heat flux were also measured. The soil thermistors were sampled every 30 minutes and averaged every hour. The meteorologic sensors were sampled every minute and averaged every hour. All of these measurements were recorded on a Campbell Scientific 21X data logger from October 1987 until April 1988.

The unfrozen soil moisture was measured using a Tektronix time domain reflectometer (TDR) about every 10 days. The TDR measures the dielectric constant of the soil which can be related to the unfrozen soil moisture (Stein and Kane, 1983; Smith and Tice, 1988). Probes, 30 cm in length, were installed horizontally at depths of 5, 10, 15, 20, 25, 30, 35, 40, 50, 60, 70, 80, 120, and 150 cm. Total soil moisture and soil density were measured using a Campbell Pacific Strata Gage. Similar instruments have proven to be very useful in studies of moisture movement in frozen soils (Goit et al., 1978). This is a dual probe system which contains a gamma source and detector for determination of density and a fast neutron source and thermal neutron detector for determination of moisture. Density is measured between the two probes while moisture content is measured in a sphere of influence around one probe. It directly outputs total moisture content and wet and dry density in 5 cm increments from the surface to 60 cm in user specified units. Measurements represent the integration of a 32 second sample period. Access tubes made of 2.5 cm diameter thin-walled aluminum were installed in the field to allow measurement in the same location throughout the winter. The tubes were tightly capped at both ends to

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Presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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prevent moisture from entering and to minimize heat transfer. Total soil moisture was measured with the same frequency as the unfrozen soil moisture.

The Strata gage was recalibrated for moisture measurement in the laboratory using soils collected at the experimental site. The factory calibration for density proved adequate when checked against these soils and pea gravel. The Strata gage was calibrated using large wooden boxes (100 cm x 60 cm x 60 cm) of soil. These boxes were equipped with aluminum access tubes identical to those used in the field and the boxes were surrounded with other containers of soil to emulate field geometry. As suggested by Rawitz et al. (1982), the accuracy of the laboratory calibration was checked in the field by collecting samples in a test site near the experimental site. After measurement with the Strata gage, samples were collected and bulk density and moisture content were determined. This instrument is rated by the factory to operate only at temperatures above 0°C, but in a 17 hour cold test, the instrument was allowed to cool from 22°C to -10°C and reported nearly identical results throughout the day.

MODELING PROCEDURE

The Stefan equation, modified by Kersten (1959) for layered media, was used to calculate frost depth using the Fairbanks airport climatological data. The standard use of these equations, referred to as the St. Paul (Minnesota) equations, employs one set of calculations for the whole freezing season and generally assume no snowpack, i.e., typical of roads and runways. However, we included the snowpack as another layer in the profile and implemented this technique on a daily basis. This allowed a continual updating of the variable snowpack thermal resistance and yielded information on the progression of the frost front. A paper detailing these procedures, which link the St. Paul equations with a soil water balance model, is currently in preparation (Fox, 1989).

Briefly, if the snow-soil profile is broken into a number of compartments and $N_{(n+1)}$ is the number of freezing degree-days available after n compartments are already frozen, then the depth of frost penetration into the (n+1) unfrozen compartment or the daily increment to the previous frost depth is given by:

$$d_{(n+1)}=-K_{(n+1)}\cdot[\sum_{i=1}^{n} R_i]+\{K_{(n+1)}{}^2[\Sigma\ R_i]^2+[48\cdot K_{(n+1)}\cdot N_{(n+1)}/(L\cdot\theta_{(n+1)})]\}^{\frac{1}{2}}$$

where,

K_{n+1} = soil compartment thermal conductivity, cal $cm^{-1}h^{-1}\cdot C^{-1}$

48 = 2 * 24 h/day

L = latent heat of fusion, cal/g

θ = volumetric moisture content, g/cm^3

N_{n+1} = number of degree-days necessary to freeze n+1 compartment

$= [L\cdot\theta_{n+1}\cdot x_{n+1}/24]\cdot[\sum_{i=1}^{n}(R_i)+R_{n+1}/2]$

x_{n+1} = thickness of the n+1 soil compartment, cm.

$R_{n+1} = x_{n+1}/K_{n+1}$ = thermal resistance of n+1 compartment, h cm^2°C cal^{-1}

$\Sigma(R_i)$ = cumulative thermal resistance of the profile above the n+1 compartment, h cm^2°C cal^{-1}

The total frost depth is the sum of the completely frozen compartments plus the frost penetration into the partially frozen (n+1) compartment or

$$D_{fr} = \sum_{i=1}^{n} (x_i) + d_{(n+1)}$$

where,

D_{fr} = Depth of frost from surface, cm.

Soil thermal conductivity was calculated using Johansen's (1975) method (as cited by Farouki, 1981). Soil latent heat of fusion is the latent heat for water multiplied by the volumetric water content of the soil. The thermal conductivity of snow was estimated from Jumukis (1977) while the thermal resistance of the snowpack was simply the snowpack thickness divided by the snow thermal conductivity.

The second method of simulating the frost depth is the TDHC model. TDHC stands for Two Dimensional Heat Conduction and was developed by Goering and Zarling (1985). It is a nonsteady state finite element program written in BASIC and operates on IBM compatible microcomputers. The program calculates phase change and accounts for latent heat through the Dirac delta function. Frozen and thawed thermal properties are incorporated into linear triangular elements. Calculations are based upon the two dimensional heat conduction equation:

$$\delta/\delta x\ [K\ (\delta T/\delta x)] + \delta/\delta y\ [K\ (\delta T/\delta y)] = C\ (\delta T/\delta t)$$

Where;

x = coordinate in horizontal direction, m
y = coordinate in the vertical direction, m
T = temperature, °C
C = heat capacity, $J/m^3 \cdot °C$
K = thermal conductivity, W/m°C
t = time, s

This equation of transient heat flow is solved using the Galerkin weighted residual process. TDHC has been used by the authors with good results for calculating the depth of thaw in continuous permafrost (Kane et al., 1989). It also performed well in comparison to another model and to an exact analytical solution (Zarling et al., 1989).

TDHC can be driven by many different forms of data input including harmonic or fixed surface temperatures, constant heat flux, convection at the surface and/or an external surface temperature or heat flux data file. In this application, the simulation proceeded in one hour time steps using the measured soil surface temperature to drive the model. A constant temperature specified the lower boundary condition at 11 m. Convection at or below the surface was not considered. The soil thermal properties were determined from empirical relationships of the soil material properties as described by Zarling et al. (1989). Thermal properties, such as frozen and thawed thermal conductivity, frozen and thawed heat capacity and latent heat can be input for each element and thus varied for the entire profile, however only two soil zones were used with uniform properties in each zone. The grid extended from the surface to 11 m consisting of 63 nodes and 80 triangular elements with smaller elements near the surface and increasing with depth.

The use of the TDHC finite element model driven by on-site measured soil surface temperatures, and the St. Paul equations driven by routinely collected National Weather Service local climatological data, provides an opportunity to compare the effectiveness and utility of the two models in relation to data requirements and availability.

RESULTS AND DISCUSSION

The 1987-1988 winter was especially mild by Fairbanks standards. Air temperature only dropped below -30°C twice (Figure 1). Maximum snow depth was 38 cm and was less than the average of 51 cm. Due to relatively high amounts of soil moisture and mild air temperatures, the maximum depth of freeze was probably less than normal for this locale. The net radiation graphically describes a subarctic winter (Figure 2). Net radiation can be quite high in the autumn and early spring, however as our daylength decreases, the intensity of solar radiation also decreases and by late December, the net radiation is near zero. The surface heat flux remained negative for most of the winter (Figure 3). This actually simplified analysis of soil freezing. In subarctic and arctic climates, sufficient warming to cause soil thawing usually happened only once a year, in the spring. Consequently, during the winter, there was only one freezing plane and the freezing front was usually continually advancing, though at a variable rate.

Although the air temperature dropped to -30°C, the soil, insulated by a snowcover and warmed by phase change, never dropped below -7°C (Figure 4). The data in Figure 4 confirm the validity of the assumption of constant temperature at 11 m. Measurements were recorded until the soil temperature began to increase in April. The depth of freeze at a particular depth was determined from the thermistors as the time when the temperature dropped below 0°C.

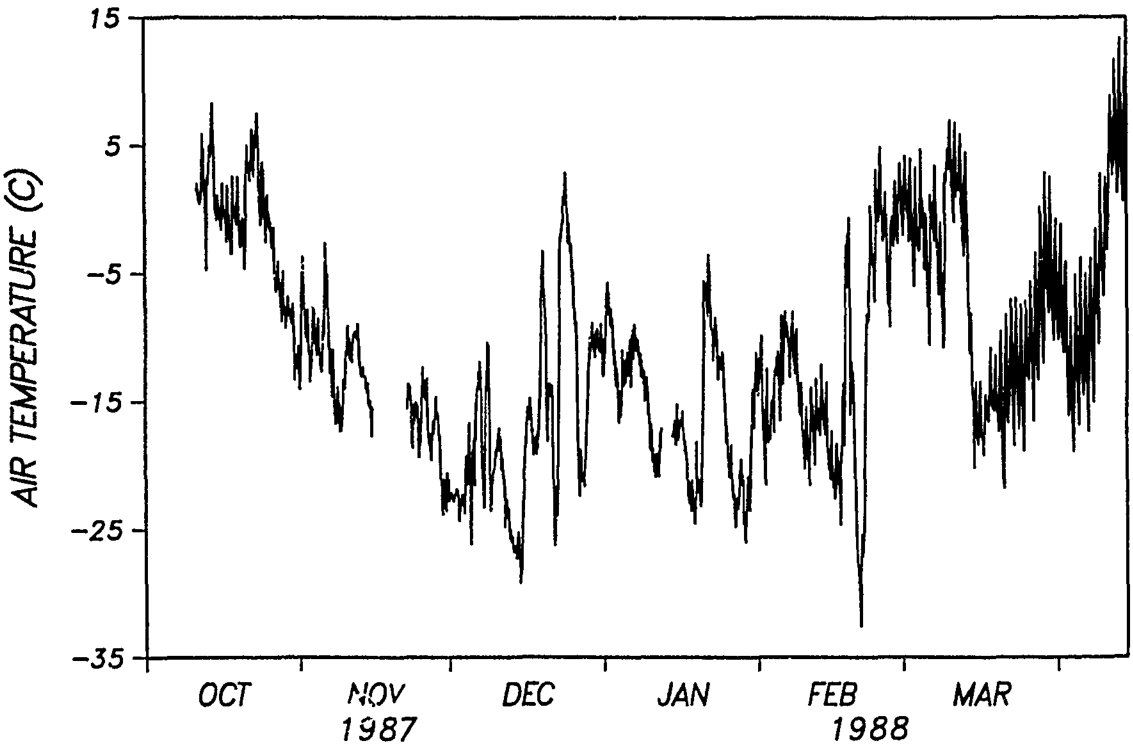

Figure 1. Air temperatures during the 1987-1988 winter at the research site.

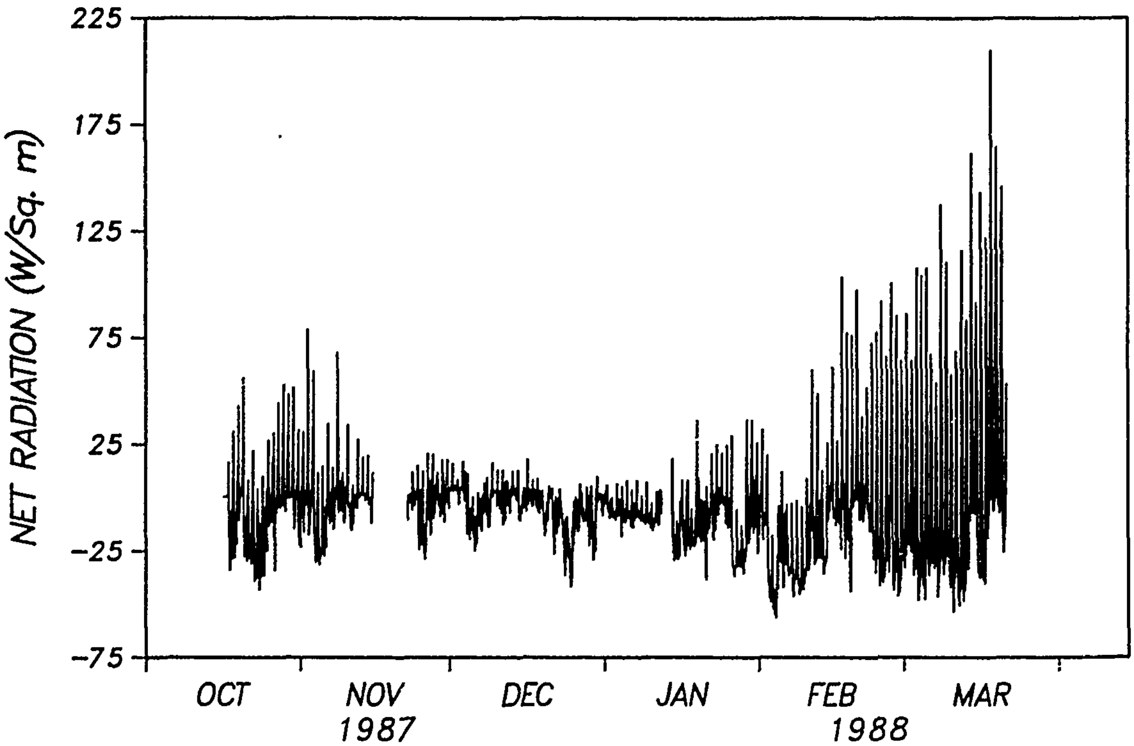

Figure 2. Net radiation measured during the 1987-1988 winter.

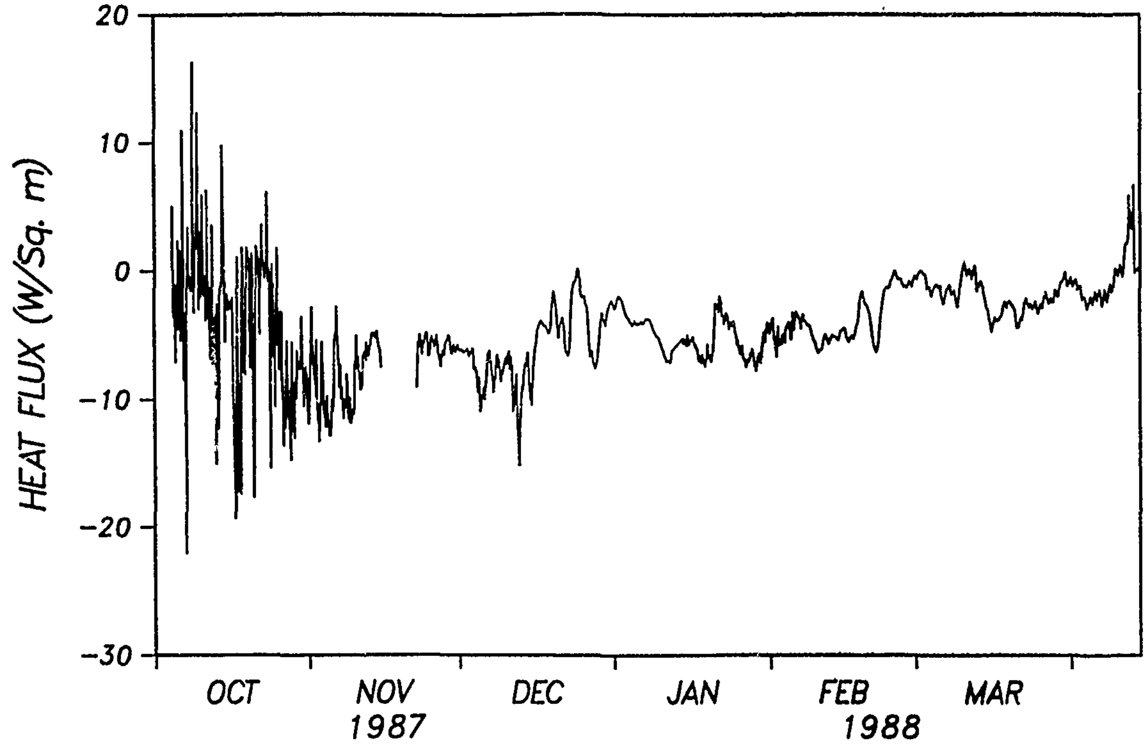

Figure 3. Surface heat flux measured during the 1987-1988 winter.

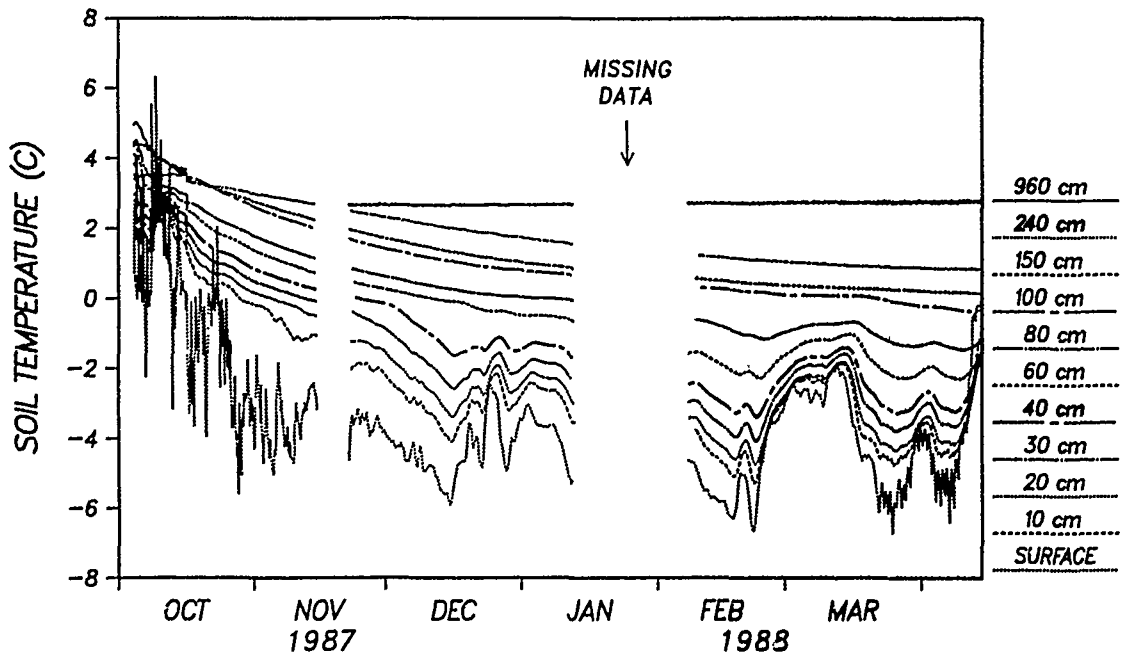

Figure 4. Soil temperatures measured at several depths during the 1987-1988 winter.

The frost depth is quite apparent in the plot of decreasing unfrozen moisture content with time (Figure 5). As the freeze front passed, the unfrozen moisture content dropped from 30 to 5%. The unfrozen soil moisture represents an excellent indicator of the phase of the soil and has been used for locating the frozen-unfrozen interface (Baker et al., 1982). Even in very cold soils, some tightly held water will remain unfrozen. A frozen silt will commonly contain 5 to 10% unfrozen water content (Stein and Kane, 1983). At the point when the freezing front advances past the TDR probes, the unfrozen moisture content will drop substantially. Although we did not monitor them continuously, the TDR method of monitoring phase change offers several advantages over thermistors. In very wet soils, the temperature may remain near 0°C for several weeks as the proportion of ice increases and liquid water decreases. In these cases, freezing of the soil profile becomes more of a slow process than an event, and determination of the depth of freeze involves more of a range than a level.

The measurements of total moisture using the Strata gage yielded some useful information. Measurement accuracy lapsed near the surface where the soil density was less than the recommended minimum for operation of this instrument. After initially draining to about 30%, the total moisture content of the soil did not change as drastically as compared to the unfrozen soil moisture. However there is a distinct increase in moisture content in the 30 to 60 cm layer over the course of the winter (Figure 6). Although this change over the season may not be hydrologically important, the effect on the thermal regime is significant. The migration of water vapor toward colder temperatures and liquid water toward the freezing front is well documented and has been observed in numerous studies (Goit et al., 1978; Williams and Wood, 1985). The process of moisture migration will influence the freezing temperatures by adding sensible and latent heat and slowing the advance of the freezing front. An excellent discussion of a field study of coupled heat-moisture transport was published by Sheppard et al. (1981).

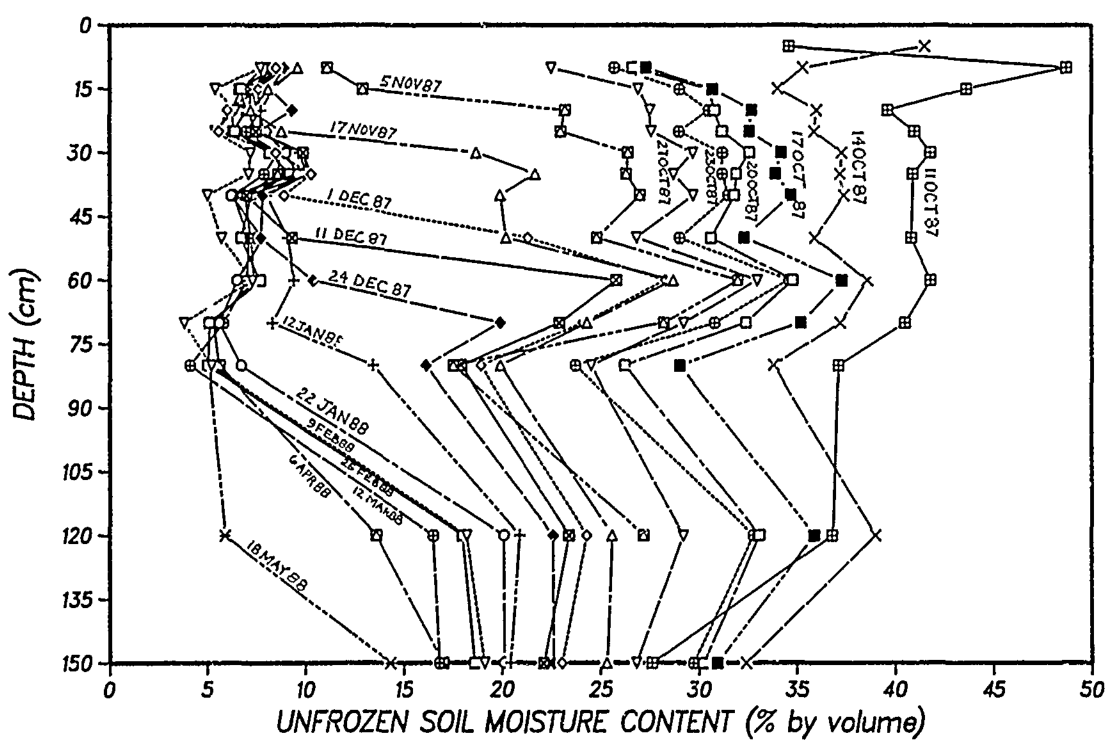

Figure 5. Unfrozen soil moisture content measured at several depths during the 1987-1988 winter.

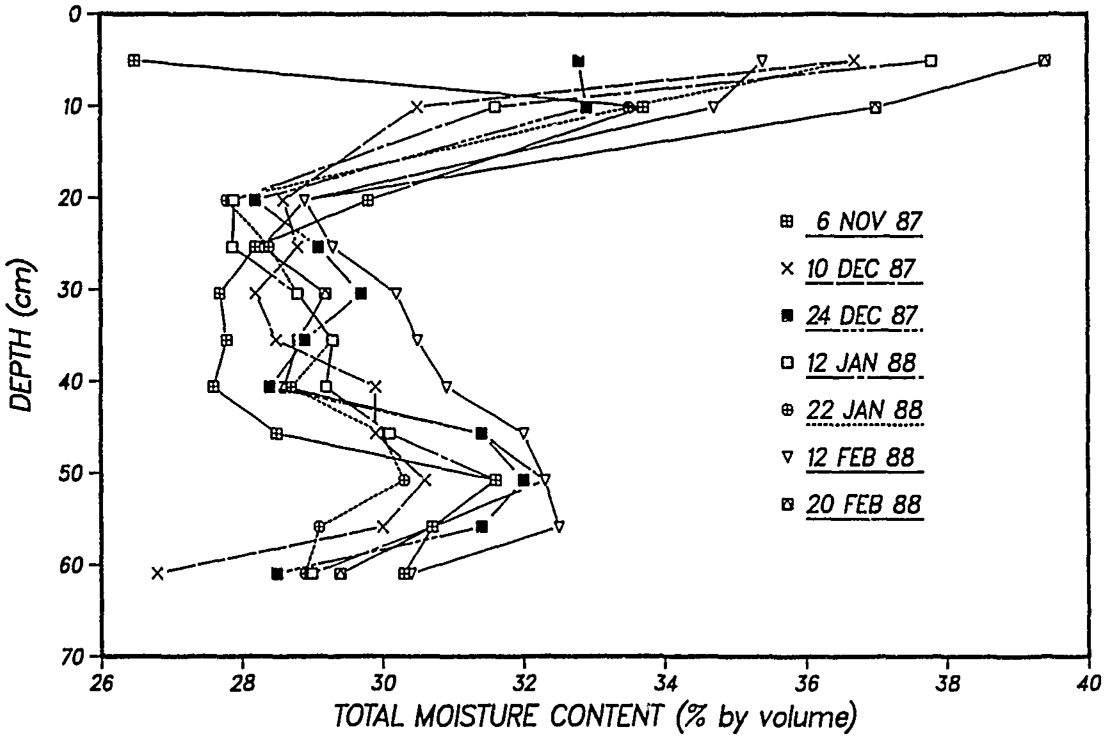

Figure 6. Total moisture content at several depths in the soil profile during the 1987-1988 winter.

TDHC functions well to model the thermal regime of the entire soil profile. Partial results from a selected depth are displayed in Figure 7. The model is driven by measured surface soil temperatures for which there are two periods of missing data during the winter. Data to run the model during these periods were created using a linear regression of the air temperature from the periods just prior and just after the missing data. The predicted temperatures suffer the greatest errors after these periods of missing data and thus this error does not necessarily reflect the accuracy of the model.

The freezing depth as predicted by TDHC is plotted in Figure 8 along with the freezing depth as predicted by the Stefan-St. Paul equations and the observed freezing depths as measured by thermistors and TDR. A regression analysis of the thermistor data shows the characteristic fit to the square root of time normally expected in freezing soils (Carslaw and Jaeger, 1959). The curve representing the thermistor data shows a rapid increase in the rate of freeze in late March. This corresponds to much colder temperatures in the air and throughout the soil profile. Although the TDR curve does not reflect this rapid increase in freezing, this is largely due to the fact that no TDR data were collected between 6 April and 18 May. The maximum depth of freeze normally occurs in late April in this region.

The Stefan-St. Paul formulation was initially oversensitive to changing air temperatures, but settled to a slight under-prediction during the beginning of the winter. This underestimate continued through January followed by over-estimation during February and March. The final estimate of the maximum depth of freeze was essentially the same as that of the thermistors.

There were many possible differences between the parameters and variables of the study site and those used to drive the Stefan-St. Paul equations. The most likely source of error was the use of only three soil zones to represent the top 126 cm of soil and the assumption of uniform moisture distribution within each zone. Figure 6 shows a reduction in soil moisture below the 50 cm depth during the December-January period. This is the

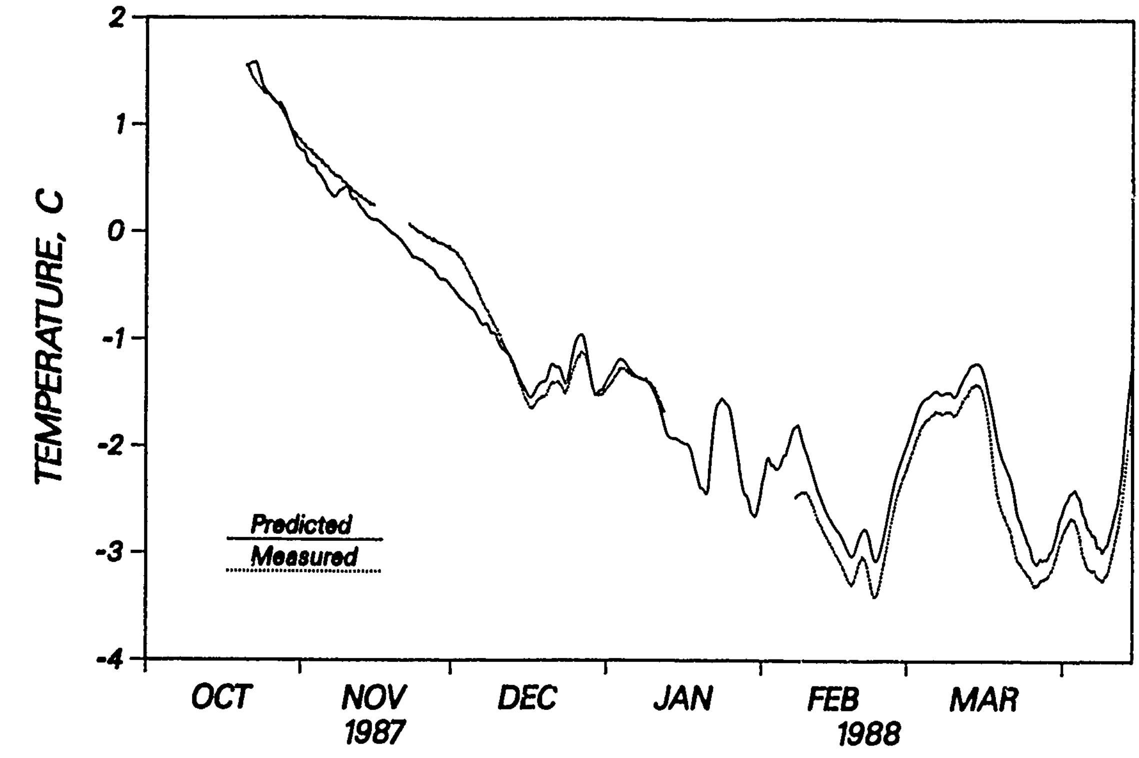

Figure 7. Observed and simulated soil temperatures at 40 cm depth during the 1987-1988 winter as predicted by TDHC.

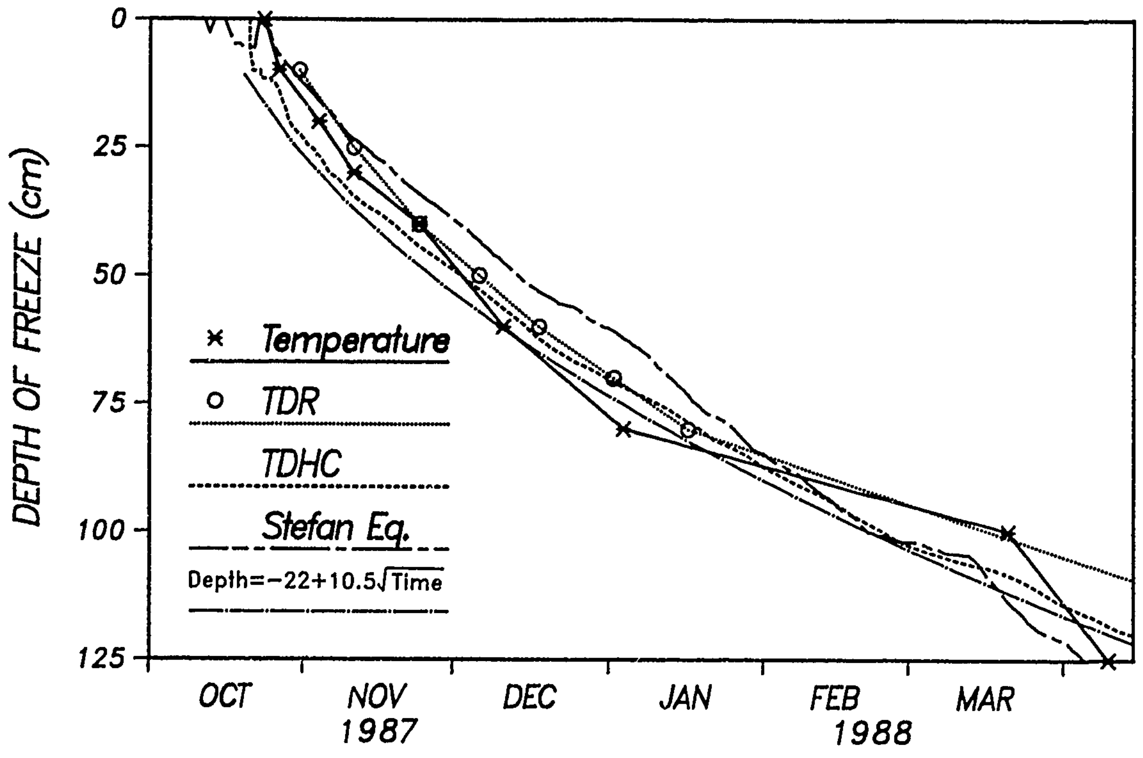

Figure 8. Depth of freeze measured by thermistors and TDR and predicted by TDHC, Stefan equation and regression analysis.

period of maximum error for both the Stefan-St. Paul model and the TDHC model. Apparently, both models responded to the increased surface temperatures during this period by slowing the rate of freeze. In contrast, the freezing front accelerated in the real soil profile because it had reached a relatively dry soil zone. It is interesting to note that the Stefan-St. Paul model does simulate the rapid frost front advance in March while TDHC does not.

TDHC predicts freezing sooner than either thermistor or TDR measurement during the first two months of freezing. This is probably for two reasons. At the surface, there was some evaporative and convective heat loss in October which may have depressed readings on the surface thermistor unrepresentatively. This would tend to make the simulation cool faster than reality. There was also convective heat transfer within the soil profile as the moisture migrated to the freezing front. This would tend to slow the advance of the freezing front as compared to predictions based upon pure conduction theory. The calculated frost depth was quite close to the measured depth in December and January but again over-estimated during February and March. The calculated maximum depth of freeze was 5 cm less than the measured frost depth.

In the final analysis, the best method of predicting frost depth depends upon what information is available. Both techniques are easy to use and easily accessible. The Stefan-St. Paul equations are driven by more readily available data, but may not simulate the rate of freezing as well as the maximum frost depth. TDHC will present more information regarding the entire thermal regime, but in this case required on-site soil surface temperatures. With regard to measuring frost depth, TDR requires special equipment and is not readily amenable to continuous measurement, but does give much more information regarding the status of the soil profile. Thermistors are straightforward and generally available, but only provide information on the soil thermal regime.

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TILLAGE AND CROP RESIDUE EFFECTS ON SOIL FROST DEPTH

by

D.H. Rickerl and J.D. Smolik[1]

INTRODUCTION

Soil frost depth is greatly influenced by snow cover. In agricultural soils snow catch is influenced by crop residues and fall tillage practices. The Northern Great Plains Region frequently experiences minimal snowfall with open periods during the winter months. The objectives of this study were to determine the effects of crop production practices on snow catch, soil frost depth, and soil thawing.

PROCEDURES

A farming systems study initiated in 1985 at the South Dakota State University, Northeast Research Station near Watertown, S.D. was chosen as the research site. The three established farming systems were Alternate (A), Conventional (C), and Ridge-till (R). The crop rotation treatments in system A were oat/alfalfa, alfalfa, soybean, and corn. In systems C and R the crop sequence was corn, soybean, and spring wheat. Corn in the Alternate system was disked in the fall and alfalfa plots were chisel plowed. In the Conventional system, corn was disked in the fall of 1987 and 1988 but not in 1986. The spring wheat in C was turned under with a moldboard plow. Corn ridges (15-20 cm) were established at second cultivation in 1987 and 1988, and in the wheat stubble the fall of 1986. Soybeans were planted on the ridges from the previous corn crop, but not ridged at cultivation. This resulted in a level seedbed for spring wheat rotated after corn in the Ridge System. Depth of tillage was approximately 8-10, 18-25, and 20-28 cm for disk, chisel plow, and moldboard plow, respectively. Crop rotations, primary tillage, and treatment numbers (TRT) are summarized in Table 1.

Table 1. Crop rotation and primary tillage in each system during 1986, 1987 and 1988.

System	Crop Rotation	Treatment Number	Year 1986	1987	1988
			- Primary Tillage -		
Alternate	Oat/alfalfa	1	none	none	none
	Alfalfa	2	chisel	chisel	chisel
	Soybean	3	none	none	none
	Corn	4	disk	disk	disk
Conventional	Corn	5	none	disk	disk
	Soybean	6	none	none	none
	Spring Wheat	7	moldboard	moldboard	moldboard
Ridge	Corn	8	ridge cultivation	ridge cultivation	ridge cultivation
	Soybean	9	ridge plant	ridge plant	ridge plant
	Spring Wheat	10	fall ridge	chisel	chisel

Soil frost tubes (Rickard and Brown, 1972) were installed in the fall of 1986, 1987, and 1988. A Giddings probe was used to remove a soil core from each plot to a depth of approximately 100 cm. Placement below this depth was inhibited by a gravel layer in the glacial till. The soil is a Brookings silty clay loam classified as a Pachic Udic Haploboroll. Frost tubes were placed between the crop rows. They were

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read at approximately 7 day intervals until spring wheat planting in April. Snow depth was also recorded as an average of 3 random measurements per plot. In the fall of 1987 and 1988, residue was estimated using a point method sited by Sloneker and Moldenhauer (1977) calculated as a percentage of ground cover.

The data were analyzed with a General Linear Model program for PC-SAS (1985). Fisher's protected Least Significant Difference was used to separate means.

RESULTS

Frost tubes were installed in early January in 1987 and removed on April 4th. There was no snow accumulated during these months. Depth of soil frost was greatest during mid-February (Table 2). In the Alternate System, oat/alfalfa (TRT 1) producing soil had significantly less frost depth than other treatments. This effect was consistent throughout the winter. Conventional and Ridge systems were similar to each other in frost patterns. Corn producing soils froze less deeply than soybean soils and the pattern for frost in spring wheat soils lie between the two. The lack of significant differences between corn and soybean soils in the Alternate system was probably due to the corn being disked. C and R systems chopped the corn residue, but it was not incorporated into the soil. Differences in frost depths within crops, due to systems, were significant.

Table 2. Soil frost depths in the winter of 1987 as influenced by treatment.

	Date									
TRT	1-23	1-30	2-5	2-13	2-20	2-25	3-13	3-20	3-26	4-4
						- cm -				
1	45	54	55	55	55	53	50	48	48	43
2	56	68	69	69	66	64	59	62	51	47
3	60	74	74	76	76	72	66	65	67	55
4	55	69	74	72	70	69	65	64	60	53
5	54	65	67	68	67	66	60	61	66	54
6	62	75	77	79	77	76	67	60	66	65
7	59	71	74	74	71	70	61	64	60	52
8	54	65	67	65	65	64	58	54	56	61
9	60	76	76	77	74	72	68	65	71	52
10	57	67	68	70	69	68	65	62	60	56
$LSD_{.05}$	8.5	8.5	8.7	9.2	8.2	7.7	8.5	10.4	10.8	9.2

Thaw depths (from the surface downward) were highly variable and significant differences were not found (Table 3).

Table 3. Soil surface thaw depths in the spring of 1987, 1988, and 1989 as influenced by treatment.

	1987			1988				1989				
TRT	3-20	3-26	4-4	3-10	3-30	4-5	4-12	3-23	3-30	4-6	4-13	4-19
						- cm -						
1	3	5	7	8	17	29	42	1	19	30	14	48
2	5	14	5	9	17	27	45	3	20	36	40	--
3	5	11	4	11	19	30	52	1	24	35	39	66
4	4	15	5	9	19	34	63	1	23	36	36	71
5	11	14	9	10	20	36	74	1	22	32	27	67
6	7	13	8	10	18	29	48	2	22	31	26	52
7	6	21	7	10	21	30	49	5	22	30	32	44
8	2	7	10	8	16	26	51	0	20	29	30	43
9	6	11	6	11	19	30	51	0	20	32	35	52
10	8	12	10	7	16	21	42	1	21	33	27	47
$LSD_{.05}$	NS	NS	NS	NS	NS	6.8	6.6	2.8	NS	NS	11.6	13.8

Spring temperatures (15 April 1987) at the 15 cm soil depth had two significant extremes (Table 4). The oat/alfalfa producing soils averaged 10°C and the conventional spring wheat soils which had been turned with a moldboard plow averaged 20°C. Other TRT soil temperatures ranged from 15-17°C.

Table 4. Fall residue and spring soil temperature as influenced by treatment.

TRT	Residue		Temperature		
	10-20-87	10-19-88	4-15-87	5-23-88	4-19-89
	- % -			- °C -	
1	99	99	10	18	8
2	31	41	16	20	8
3	66	46	17	20	10
4	61	46	17	20	9
5	66	28	15	19	9
6	78	38	16	20	9
7	26	15	20	19	7
8	66	48	16	19	10
9	69	61	17	20	8
10	91	68	17	19	8
$LSD_{.05}$	12.7	14.1	3.2	NS	1.5

Frost tube readings in the 1987-88 winter began on December 12th and ended on April 12th (Table 5). By the middle of February, the ground was frozen below the depth of the tubes and measurements could not be made. Generally, the oat/alfalfa producing soils froze less deeply than other treatments in the A system. Residues measured after harvest in 1987 were 99% for oat/alfalfa (Table 4) and snow catch was consistently high for this TRT (Table 6). In the conventional system, the disked corn had less residue and less snow than the C soybean and the soil froze deeper than the soybean soil. This was opposite the results in 1987 when the corn was left on the surface and there was no snow cover. Spring wheat in the Ridge system had 91% residue, high snow levels, and less frost than R soybean and corn. In comparing systems, R corn and wheat retained more snow than in the A or C system and frost depth was less. During mid-winter, slightly higher (non-significant) amounts of residue and snow in C soybean resulted in less frost depth for that system than the A soybean.

Table 5. Soil frost depth in the winter of 1987-88 as influenced by treatment.

TRT	Date										
	12-8	12-17	1-8	1-11	1-21	1-26	2-2	2-9	3-30	4-5	4-12
						- cm -					
1	14	20	42	60	70	70	71	82	78	72	16
2*	22	26	52	76	86	--	--	--	87	86	82
3	20	28	57	82	94	98	92	101	98	99	99
4	20	29	60	84	95	97	99	--	98	97	98
5	22	30	60	84	92	93	95	--	93	92	92
6	19	26	51	74	84	86	91	99	95	91	90
7	21	28	53	78	93	95	99	--	99	99	104
8	14	21	49	76	84	79	84	89	89	87	85
9	21	29	56	80	92	94	99	--	98	98	98
10	16	24	45	62	72	73	75	86	86	75	54
$LSD_{.05}$	4.0	3.3	5.2	6.8	8.7	9.2	11.0	NS	NS	9.0	24.8

*Soil frost depth greater than 100 cm is designated by --.

Surface thaw depth in corn soils on the last two dates ranked R < A < C. Soybean soils were not tilled in the fall, and there were no differences among systems. Wheat residues plowed under in the conventional system allowed soils to thaw more rapidly than in the R system. Spring temperatures showed no significant differences.

During February of 1989 soil frost depth in all treatments was greater than the depth of frost tube placement. Only the A alfalfa (TRT 2) soils thawed upward to detectable levels before frost tubes were removed in April. Fall frost readings were less deep in the A oat/alfalfa (TRT 1) than other treatments (Table 7). Treatment 1 also had 99 percent ground cover (Table 4) and generally caught more snow (Table 6). In system R, corn producing soils had less frost than spring wheat or soybean until 12-29 when snowfall eliminated significant differences among R crops.

Table 6. Snow depth in the winters of 1987-88 and 1988-89 as influenced by treatment.

	1987-88					1988-89							
TRT	1-8	1-21	1-26	2-2	2-9	12-29	1-6	1-13	1-20	2-8	2-15	3-8	3-16
						- cm -							
1	5	14	16	10	18	11	8	8	2	5	10	10	6
2	5	9	10	3	10	5	3	3	0	3	4	7	3
3	2	2	4	2	8	9	8	10	3	6	10	11	8
4	4	4	4	1	6	8	6	6	1	3	9	9	4
5	3	4	4	1	5	7	4	5	0	2	7	8	4
6	4	4	6	1	9	5	5	5	0	2	5	10	3
7	5	6	8	4	6	3	2	4	0	2	3	6	0
8	5	12	13	6	9	9	6	3	2	6	8	8	5
9	2	2	3	1	6	9	9	7	3	5	8	10	7
10	10	13	14	8	16	10	9	9	4	7	10	9	6
$LSD_{.05}$	3.7	4.5	5.4	4.7	4.8	2.0	2.7	4.0	1.9	2.7	2.5	2.3	2.5

Table 7. Soil frost depth in the winter of 1988-89 as influenced by treatment.

	Date								
TRT	12-9	12-16	12-23	12-29	1-6	1-13	1-20	1-27	2-8
					- cm -				
1	23	39	48	58	66	75	80	79	90
2*	30	50	59	67	--	--	--	--	--
3	28	47	56	66	79	72	92	92	99
4	28	46	55	68	82	85	99	99	--
5	28	46	54	65	79	84	94	90	--
6	31	48	56	67	81	90	100	100	--
7	28	44	51	61	69	79	--	--	--
8	25	40	48	58	76	90	92	89	96
9	27	44	54	63	73	83	88	88	93
10	29	45	53	61	70	80	88	92	100
$LSD_{.05}$	3.0	4.8	5.0	6.0	6.2	NS	NS	NS	NS

*Soil frost depth greater than 100 cm is designated by --. After 2-8, all treatments were --.

Alternate alfalfa which had been chiseled was the first soil to thaw. Other TRT were grouped as 3, 4, and 5 > 1, 6, 7, 8, 9, 10. Spring temperatures ranged from 7°C (TRT 7) to 10°C (TRT 3 and 8) (Table 4).

CONCLUSIONS

Crop residues act as an insulating material and affect soil frost depth in winters with no snow cover. In this study, soil frost depth was oat/alfalfa < corn < spring wheat < soybean when there was no snow cover. Fall tillage reduced crop residue and its insulating effect. With moderate snow catch (4-8 cm) residues which trapped snow reduced frost depth and slowed spring thaw. When snow catch exceeded moderate levels, differences among crop and tillage systems were not significant.

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Comparison of Numerical Simulations with Experimental Data for a Prototype Artificial Ground Freezing

by
John M. Sullivan, Jr. and Laurie A. Stefanov[1]

ABSTRACT

This project numerically simulates the use of artificial ground freezing as a means of containing and consolidating toxic chemical spills. This innovative treatment technology can cleanse the soil *in situ* through the use of freeze/thaw cycles. Artificial ground freezing has been in practice for over a century in civil engineering applications. Its ability to form impermeable barriers, dewater sludge, and consolidate solids has been demonstrated. Additionally, the environmental dangers of the treatment process are virtually nonexistent even in populated regions. However, predicting or controlling the location of the frozen barrier as a function of time for arbitrary geometries is unsolved. This control is paramount for successful implementation of the process.

The numerical simulations herein are compared to existing, large-scale experimental data for situations involving the solidification of saturated Lebanon silt soil. The successful agreement of the numerical simulations and experimental data supports the use of numerical modeling as a necessary tool for deployment of refrigeration systems for the treatment of toxic spills.

KEY WORDS: Frozen Soil, Hazardous Waste Containment, Artificial Ground Freezing

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Comparison of Numerical Simulations with Experimental Data for a Prototype Artificial Ground Freezing

by

John M. Sullivan, Jr. and Laurie A. Stefanov

INTRODUCTION

Hazardous waste and ground freezing

The containment, consolidation and treatment of toxic chemicals in the soil is an unresolved national tragedy. Recent investigations into new frontiers for hazardous waste management show that current treatment methods are inadequate for ensuring safe immobilization or consolidation of hazardous waste.(Hill, 1985; EPA, 1984; Neely, 1981) A new, innovative application of an established technology is the use of ground freezing as a means of hazardous waste containment.(Sullivan, 1984; Iskandar, 1985 and 1986)

Ground freezing is not a new technology. Its application dates back over a 100 years in the mining industry for shaft sinking.(Sadovsky, 1980) The construction industry uses ground freezing routinely for open excavations, deep unsupported construction trenches, inclined tunnels and subway constructions.(Sadovsky, 1980; Braun, 1982; Dorman, 1971) Some of the advantages of ground freezing over other construction practices are that: a.) ground freezing is not restricted to any specific soil type; b.) ground freezing can handle a large spectrum of boundary conditions and construction site requirements; c.) there are few environmental concerns in the use of ground freezing; and d.) ground freezing is a fast and temporary site application. Essentially, ground freezing is a proven technology for increasing the rigidity of the soil and decreasing the permeability of the soil.

There are three ground freezing scenarios being investigated for the treatment of contaminated soils.(Sullivan, 1984; Iskandar, 1985, 1986 and 1988) In each case consider a hypothetical situation shown in Fig. 1 where a derailed chemical car disperses a toxic substance over an area about the railroad track. (Sullivan, 1984) The first scenario is <u>containment of hazardous waste</u>. This process involves placement of refrigeration rods in the uncontaminated soil about the perimeter of the toxic region. Ideally, the refrigeration rods reach the depth of bedrock. A frozen wall of uncontaminated soil encloses the hazardous waste. Thereafter, clean up crews treat the localized toxic region. The second scenario, <u>immobilization of hazardous waste</u>, involves placement of refrigeration rods about the perimeter of the toxic region as well as within the zone. The method of refrigeration is liquid nitrogen. In this scenario the entire toxic zone is frozen

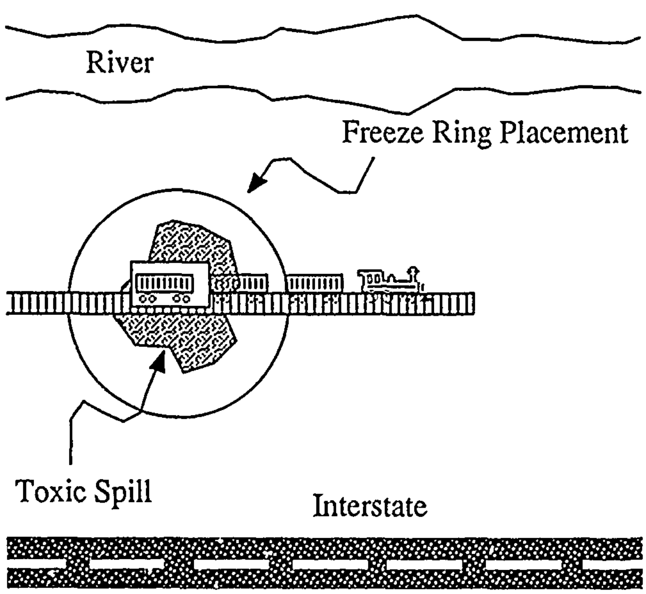


Figure 1 - Hypothetical Train Car Toxic Spill (Taken from Sullivan, 1984)

rapidly. Ice wedges and dendritic fingers form throughout the zone entrapping the solute. The frozen soil is excavated and treated elsewhere. The third scenario is the consolidation of the chemical spill. A toxic zone is initially contained via controlled ground freezing in the uncontaminated perimeter as in the first scenario. The frozen wall or solidification front continues to advance, driving the solute ahead of the front. This consolidates the toxic solute into a region of smaller diameter. The higher the solute concentration of this smaller region the easier the recovery process. In fact, the recovery process may be able to reclaim the hazardous material (as opposed to hazardous waste). In order to carry out these scenarios one must be able to predict and control the solidification front. However, predicting or controlling the location of the frozen barrier as a function of time for arbitrary geometries is unsolved.

This project numerically simulates the use of artificial ground freezing as a means of containing and consolidating hazardous waste. Inherent in the coupled heat and mass transfer system lie nonlinearities associated with the phase transformation and constitutional supercooling. The coupled transport system dealing with solidification of contaminated soils is unamiable to analytical solutions beyond idealized one-dimensional situations. Consequently, it is the third science, i.e. computational science, that holds the greatest promise of illuminating the nonlinear solution to the control of ground freezing. Herein, we delineate the physics and numerical formulation of the coupled heat and mass transport system. Numerical results are presented for the thermal solution of solidification of saturated Lebanon silt soil. These simulations are compared to the experimental data obtained at the U. S. Army Cold Regions Research and Engineering Laboratory (CRREL) in their prototype test facility.(Ayorinde, 1988)

PROBLEM STATEMENT

Consider a solidifying homogeneous soil matrix with an internal moving boundary S that separates the frozen and unfrozen phases, Fig. 2. It is assumed that heat conduction is the only mechanism of thermal transport, i.e. natural convection effects are small for the thermal range of interest; and that the density remains constant during phase transformation. Therefore, the heat transfer process in the frozen and unfrozen phases is governed by

$$c\frac{\partial T}{\partial t} = \nabla \cdot K \nabla T \qquad (1)$$

where c is the volumetric heat capacity; K is the thermal conductivity and T is the temperature expressed as a deviation from the thermodynamic equilibrium temperature for a planar front without solute T_m. Without natural convection or ground water flow in the unfrozen soil the solute must diffuse away from the interface. The governing mass transfer equation is

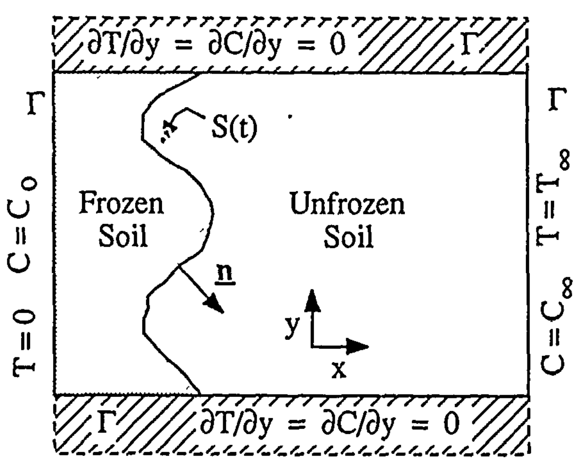


Figure 2 - Two Phase Domain with Interface S

$$\frac{\partial C}{\partial t} = \nabla \cdot D \nabla C \qquad (2)$$

where C is the solute concentration and D is the mass diffusivity in the unfrozen soil. The mass diffusion in the solid is assumed negligible. The usual temperature and concentration boundary conditions apply on the external boundary Γ as displayed in Fig. 2. However, multiple boundary conditions exist on the interface. The solidification temperature is a function of the solute concentration and surface energy effects,

$$T(S(t)) = m_c C - \left|\frac{\gamma T_m}{L}\right| \mathbb{K} \qquad (3)$$

where T(S(t)) is the time dependent temperature on the interface. The first term of (3) accounts for the temperature depression due to the solute where m_c is the slope of the constitutional phase diagram. The geometry effect is a function of the interfacial energy γ, and curvature $\mathbb{K}$ with L/T_m being the entropy of fusion.

A second boundary condition on the interface is required to account for its motion. The equation of motion preserves the balance between the sensible heat transported away from the interface into the frozen and unfrozen regions and the latent heat of fusion L released during solidification at the interface,

$$L\, V \cdot \underline{n} = [\, K_1 \nabla T|_1 - K_2 \nabla T|_2 \,]_{S(t)} \cdot \underline{n} \qquad (4)$$

where subscripts 1 and 2 denote frozen and unfrozen regions, respectively. The unit normal $\underline{n}$ is directed out of region 1. A similar boundary condition on the interface exists to balance the solute displaced ahead of the moving front and its transport away from the front,

$$C_S\,(1-k_c)\, V \cdot \underline{n} = -D\, \nabla C \cdot \underline{n} \qquad (5)$$

where C_S is the solute concentration on the interface S in the unfrozen soil and k_c is the partitioning coefficient, i.e. the ratio of the frozen/unfrozen soil concentration at the interface.

NUMERICAL FORMULATION

The soil solidification problem is solved on deforming finite elements, with element boundaries tracking the phase front. There are therefore three fundamental unknowns at each node; temperature, position and solute concentration. For the temperature, the Galerkin formulation of the heat equation (1) is employed. On a deforming mesh this is (Sullivan, 1987)

$$\begin{aligned} &< c\, \phi_j \phi_i > \frac{dT_j}{dt} + [< -c\, V^c \cdot \nabla \phi_j\, \phi_i > + < K\, \nabla \phi_j\, \nabla \phi_i >]\, T_j = \\ &\int_\Gamma K\, \nabla T \cdot \underline{n}\, \phi_i\, ds + \int_S L\, V \cdot \underline{n}\, \phi_i\, ds \end{aligned} \qquad (6)$$

in which <( )> indicates integration over the domain; $\Sigma\phi_j$ are the finite element basis functions; T_j is the temperature at node j, and (4) is incorporated into the surface integral on S. V^c is the motion of the computational mesh, and the advection term accounts for this motion.

A finite difference approximation is used for dT_j/dt

$$\frac{dT_j}{dt} = \frac{T_j^{k+1} - T_j^k}{t^{k+1} - t^k} \tag{7}$$

where superscripts imply time steps and subscripts indicate node index. The heat equation is evaluated at time $t^{k+\theta}$

$$t^{k+\theta} = \theta\, t^{k+1} + (1-\theta)\, t^k \tag{8a}$$

with

$$T_j^{k+\theta} = \theta\, T_j^{k+1} + (1-\theta)\, T_j^k = T_j^k + \theta\, \Delta t \frac{dT_j}{dt} \tag{8b}$$

and $0 < \theta < 1$. Rearranging (6) to solve for dT_j/dt produces the Galerkin formulation of (1) used in this investigation

$$\begin{gathered}[< c\,\phi_j\, \phi_i > - \Delta t\, \theta < c\, V^{e} \cdot \nabla\phi_j\, \phi_i > + \Delta t\, \theta < K\, \nabla\phi_j\, \nabla\phi_i >] \frac{d\,T_j}{dt} = \\ [< c\, V^{e} \cdot \nabla\phi_j\, \phi_i > - < K\nabla\phi_j\, \nabla\phi_i >]\, T_j^k + \int_\Gamma K\nabla T\cdot\underline{n}\,\phi_i\, ds + \int_S LV\cdot\underline{n}\,\phi_i\, ds \quad ,\end{gathered} \tag{9}$$

which may be solved for the set of dT_j/dt given V^e on the interior, $V{\cdot}\underline{n}$ on S and the thermal boundary conditions.

In exactly the same manner the transport of solute in the unfrozen zone, Eq 2, becomes

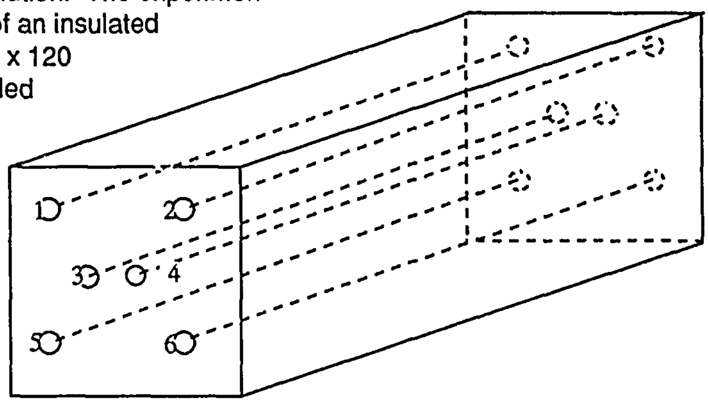
$$\begin{gathered}[< \phi_j\, \phi_i > - \Delta t\, \theta < V^{e} \cdot \nabla\phi_j\, \phi_i > + \Delta t\, \theta < D\, \nabla\phi_j\, \nabla\phi_i >] \frac{d\,C_j}{dt} = \\ [< V^{e}\cdot\nabla\phi_j\, \phi_i > - < D\nabla\phi_j\nabla\phi_i >]\, C_j^k + \int_\Gamma D\nabla C\cdot\underline{n}\,\phi_i\, ds + \int_S C_S(1-k_c)V\cdot\underline{n}\,\phi_i\, ds\end{gathered} \tag{10}$$

where (5) is incorporated into the surface integral on S.

Since the physics of the problem specify only $V{\cdot}\underline{n}$ on the interface, one is free to choose $V\underline{t}$. Our numerical strategy is to maintain uniform node spacing along the phase front by adjusting the tangential velocity of the nodes while satisfying (4).(Sullivan, 1988) The interior node motion V^e of the finite element grids has to maintain a numerically intact and computable topology during the evolution of the physical domain. Additionally, the unfrozen thermal and mass finite element meshes must be coupled at the inte..ace. The finite element meshes are assumed to behave as perfectly elastic materials such that once the motion of the boundaries is specified, the interior motion satisfies the linearized equations of elasticity for plane stress.(Lynch, 1982; Sullivan, 1986) We emphasize that this interior node motion is a numerical issue only. There are no solidification physics involved.

RESULTS

One- and two dimensional simulations of (9) were performed using the experimental data obtained at the U.S. Army CRREL for boundary and initial conditions only. (Ayorinde, 1988) The experimentally measured temperatures of the soil interior were not used by the nurerical simulations rather these data were compared to the simulated results to validate the numerical solution. The experimental apparatus consisted of an insulated concrete tank (180 x 120 x 120 cm), instrumented and filled with saturated Lebanon silt soil. CRREL monitored and recorded 66 thermocouples dispersed throughout the test apparatus as shown in Fig. 3. Table I summarizes the thermal parameters used during the simulation and Fig. 4 displays a typical two dimensional mesh used to simulate the freezing process.



Location of Thermocouple Strings (1 - 6)
Each String has 11 Uniformly Placed Thermocouples

Figure 3 - Schematic View of Thermocouple Placement within Experimental Apparatus (from Ayorinde, 1988)

The duration of CRREL's test was 2200 hours with temperature values recorded approximately every 4 hours.

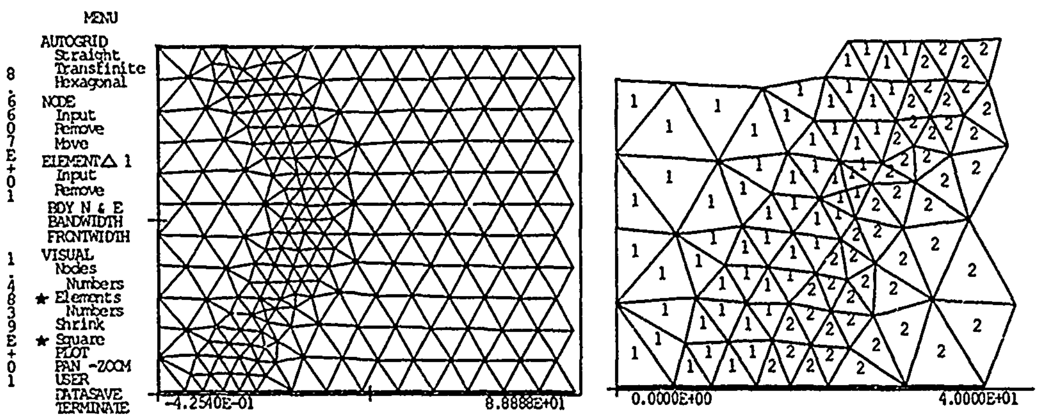


Figure 4 - Typical 2-D Finite Element Mesh Used in Simulation
a.) Complete Y range shown, X range extends to 152.4 cm.
b.) Enlarged lower left section - where 1 = solid and 2 = liquid region

Table I - Thermal Properties Used in the Numerical Simulations of the Freezing of Lebanon Silt Soil.

Thermal Property	Solid Domain	Liquid Domain
Thermal Conductivity Cal/(cm °C hr)	16	14.5
Volumetric Specific Heat Capacity Cal/(cm^3 °C)	0.36	0.478
Latent Heat of Fusion Cal/(cm^3)	17.0	

A series of graphs document the progression of the freezing front and the thermal distribution through the soil system. The initial conditions of the simulation were those of the experimental state throughout the domain. Our data comparison with the experimental is based on the average temperatures of thermocouple strings 3 and 4. Figure 5 displays the numerical temperature profiles and the experimental temperature values at the 11 thermocouple positions along the length of the test facility. Note that the agreement in the temperature profile as well as the location of the solidification front between the numerical simulation and experimental data is excellent throughout the transient and steady state periods. The motion of the phase front was greatest during the initial 400 hours of the test, Fig.5(a-d). Over the next 600 hours the experimental system approached its steady state location with little change in the phase front location or temperature profile, Fig.5(e-f).

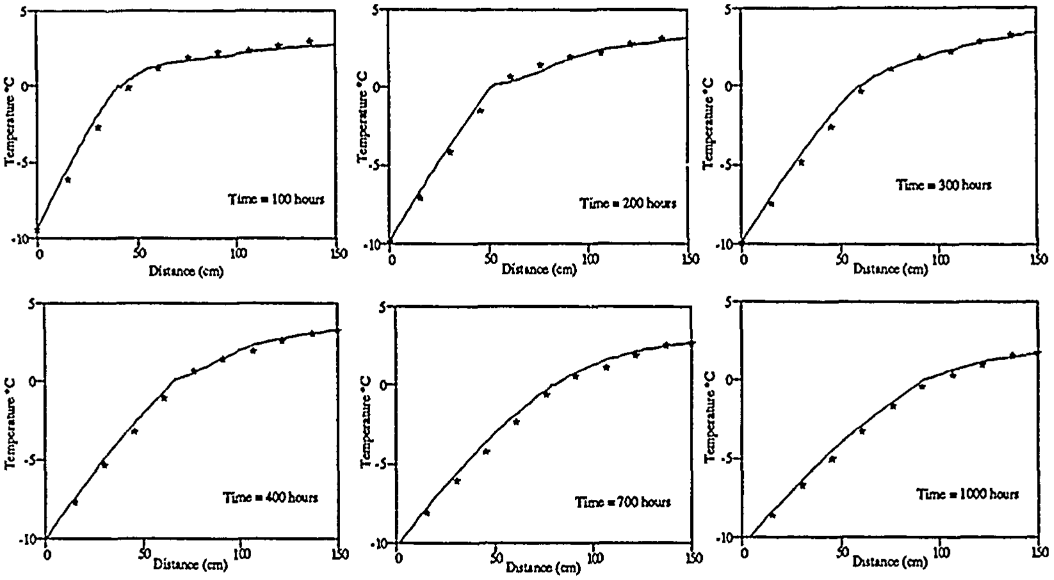


Figure 5 - Numerical (Lines) and Experimental (*) Data During Initial Transient Freezing Through Steady State Modes

The CRREL refrigeration system malfunctioned during the test hours of approximately 1100 hours through 1300 hours. During this time the refrigeration plate lost its coolant and the temperatures oscillated within the test domain as attempts were made to restart the refrigeration system. Fortunately, the automatic data acquisition of the test facility was completely operational during these refrigeration failures. As a consequence of this situation we had an exciting opportunity to simulate the highly transient behavior of the physical system during the malfunction period. Figure 6 documents the experimental temperature oscillations and the results of the robust numerical system. The experimental temperatures rose and fell rapidly causing temperature inflections within the test domain. Our numerical system tracked the experimental data with fidelity. The close agreement of numerical and experimental results establishes the validity of this numerical strategy as a necessary tool for the deployment of refrigeration systems.

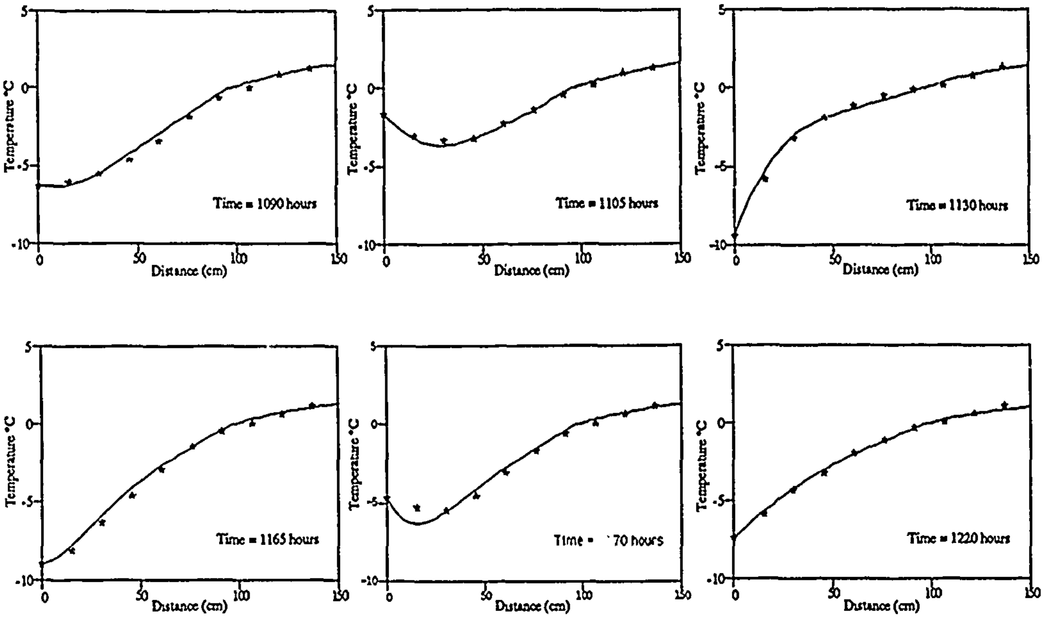


Figure 6 - Numerical (Lines) and Experimental (*) Data During Refrigeration Malfunction Period

CONCLUSIONS

A numerical strategy for the solution of two phase solidification situations with latent heat of fusion released at the interface has been formulated, tested and compared to existing experimental data. The numerical simulations tracked the experimental system throughout all transient and quiescent stages of the test. The close agreement of the numerical work to the observed data support the use of numerical modeling as a necessary tool for deployment of refrigeration systems for the treatment of toxic spills.

ACKNOWLEDGEMENT

This work was supported by the National Science Foundation under Grant CBT-8808477; and by the Research and Development Council, WPI under RDC Award 4-29860.

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EFFECT OF FREEZE-THAW ACTIVITY ON WATER RETENTION, HYDRAULIC CONDUCTIVITY, DENSITY, AND SURFACE STRENGTH OF TWO SOILS FROZEN AT HIGH WATER CONTENT

by

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INTRODUCTION

It has been shown that freezing and thawing affect soil physical properties. However, the type and magnitude of effect reported have not always been consistent. For instance, Baver (1956), Bisel and Nielson (1964 and 1967), Domby and Kohnke (1955), Leo (1963), Slater and Hopp (1949), and Willis (1955) showed that soil frost caused a breakdown of soil physical properties while Gardner (1945), Sillanpaa (1961), and Sillanpaa and Webber (1961) showed that frost action improved a soils physical properties. In addition, Benoit (1973), Chepil (1954), Mostaghimi et al. (1988), and Sillanpaa and Webber (1961) showed that frost action could have a positive or negative affect on soil physical properties depending on the soil involved, its initial water content and degree of aggregation, and rate of freezing. Thus, frost affects soil physical properties as a result of complex actions and interactions of initial soil conditions and freezing temperature.

Most of the reported work has dealt with rates of freezing, degree of aggregation and water content at freezing. Little work has dealt with soil density-frost interactions even though prevailing concepts suggest that frost action will improve the density and water characteristics of compacted soil. Our objective here is to show how cyclic soil freezing and thawing affect soil bulk density, aggregate size, water content, and hydraulic conductivity for various conditions of initial density, aggregate size, and numbers of freeze-thaw cycles.

PROCEDURE

Soil from the Ap horizon of a Barnes loam (Udic Haploboroll) and a Hamerly clay (Aeric calciaquoll, fine-loamy, frigid) was collected near Morris, Minnesota, dry-sieved into three aggregate size groups (0.0 to 0.5, 1-3, and 5-12 mm) and stored in covered, large plastic barrels. Random samples of each aggregate size were packed with a hydraulic press into 76-by 76-mm thick-walled (6 mm) plastic cylinders to target densities of 1.0, 1.2, 1.4 and 1.6 Mg/m^3. Five replications of each aggregate size and bulk density were prepared. All cores were saturated by immersion in water to their upper surface for at least 48 hours and then drained to the atmosphere for 24 hours. Saturation caused soil swelling to occur. After saturation and drainage, all cores were trimmed to remove excess soil and an initial density and a free drainage water content (Kg/Kg) value was calculated for each density-aggregate size group. A duplicate set of five identically treated cores was used for determining initial surface strength values with a fall cone penetrometer. After removal of excess soil, an initial saturated hydraulic conductivity value was determined for each core by a standard vertical flow, constant head procedure. The cores were then inserted into a closed cell foam system that insulated the sides and bottom of the cores from ambient conditions and made it possible to impose freezing treatments from the upper surface that simulated normal vertical freezing and thawing as found under field conditions. Five replications of each aggregate size and density group were then subjected to 1,5 or 10 freeze-thaw cycles. All cores were than resaturated, drained to the atmosphere and final values determined for water content, saturated hydraulic conductivity, bulk density and surface soil strength. Data were converted to express the change that occurred between initial values and values recorded after freeze-thaw action as the ratio of final to initial values (Final÷Initial). Thus, all numbers greater than 1 indicate an increase in value of the property measured while numbers less than 1 indicate a decrease.

Presented at Frozen Soil Symposium, March 21-22, 1990, Spokane, Washington.

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A complete analysis of variance was run on all data collected to evaluate freeze-thaw effects as related to soil type, aggregate size group, initial density and number of freeze-thaw cycles and their interactions. In addition, a stepwise multiple regression analysis was used to develop equations expressing change in bulk density, water content and hydraulic conductivity to freeze-thaw cycles, initial density, and initial aggregate size.

RESULTS

The data collected were expressed as relative change (Final÷Initial) values for bulk density (ΔBD), water content (ΔW), hydraulic conductivity (ΔHC), and surface strength (ΔSS) after 1, 5, and 10 freeze-thaw cycles. Examination of these data showed that the greatest change in relative value occurred after the first freeze cycle with further changes caused by freezing cycles being significant but of smaller value. Relative changes recorded after 10 freeze cycles are the only values shown (Table 1). The zero values shown for changes in hydraulic conductivity at high initial density and 1-3 mm aggregate size are due to a zero initial value (i.e., no water flow detectible after 24 hours). Thus, dividing final values by initial values gives meaningless results in this case.

Table 1. Relative change (Final÷Initial) in bulk density (ΔBD), water content (ΔW), hydraulic conductivity (ΔHC), and surface strength (ΔSS) after 10 freeze-thaw cycles as related to initial soil density and aggregate size.

	BARNES LOAM				HAMERLY CLAY			
AGGREGATE SIZE < .5 mm.								
INITIAL BD	1.04	1.14	1.09	1.12	1.04	1.15	1.19	1.26
	F/I				F/I			
ΔBD	1.00	0.98	1.05	1.03	1.00	1.00	1.01	0.98
ΔW	0.75	0.81	0.90	0.77	0.74	0.84	0.79	0.77
ΔHC	0.62	2.11	3.26	9.79	0.44	0.65	1.98	6.24
ΔFC	5.33	9.17	3.26	6.07	16.36	5.04	5.56	3.67
AGGREGATE SIZE 1-3 mm.								
INITIAL BD	0.97	1.13	1.10	1.50	1.02	1.17	1.24	1.51
	F/I				F/I			
ΔBD	1.13	1.16	1.04	0.86	1.20	1.15	1.00	0.99
ΔW	0.93	0.74	0.84	0.83	0.81	0.63	0.94	0.76
ΔHC	0.16	0.60	24.63	0.00	0.16	0.30	3.07	0.00
ΔFC	0.00	0.00	2.92	0.00	0.00	0.00	0.50	0.00
AGGREGATE SIZE 5-12 mm.								
INITIAL BD	0.89	0.99	1.18	1.20	0.94	1.07	1.29	1.36
	F/I				F/I			
ΔBD	1.00	1.32	1.02	1.03	1.31	1.29	0.98	0.97
ΔW	0.74	0.76	0.81	0.69	0.82	1.30	0.81	0.73
ΔHC	0.12	0.12	4.30	0.55	0.27	0.15	1.50	1.49
ΔFC	4.96	2.66	0.77	2.18	2.54	1.41	0.40	1.81

The data show complex relationships with freezing and thawing causing increased or decreased relative change as related to initial aggregate size and soil density. An analysis of variance of the complete data including numbers of freeze cycles shows that ΔBD is significantly ($p < 0.01$) related to soil type, aggregate size, initial density and numbers of freeze cycles. Interactions between and among these items are also significant. The change in hydraulic conductivity showed similar results except that the probability level for numbers of freeze cycles was 0.04. In contrast, the change in water content was significantly related only to initial aggregate size and numbers of freeze cycles.

Our statistical analysis of surface strength (penetrometer) data is incomplete. However, relative numbers for changes in fall cone penetration show increased penetration after freeze-thaw depending on aggregate size, density, and numbers of freeze cycles.

A stepwise multiple regression analysis was completed for the data for each soil to show the complex, interactive effects of initial aggregate size, initial density, and numbers of freeze cycles on ΔBD, ΔHC, and ΔW. The analysis yielded the equations:

$$\Delta BD1 = .60279*AS + 1.1426*BD - .046434*AS^2 -.32243*BD^2 +.0060098*AS*FC - .35041AS*BD - .0085692*FC*BD$$

$$\Delta W1 = .28964*AS + .65644*BD + .026917*BD^2 - .0054375*AS^3 -.0053023*AS*FC - .20382*AS*BD + .0038805*FC*BD$$

$$\Delta HC1 = -1.5748*AS - .56148*FC + 4.2887*BD + .012972*FC^2 +.090358*AS^3 - 1.1488*BD^3 + .39091*FC*BD$$

$$\Delta BD2 = .17297*AS + 1.703*BD - .73584*BD^2 - .0052373*AS^3 -.000079433*FC^3 = .0063808*AS*FC - .10963*AS*BD$$

$$\Delta W2 = .22899*AS - .060502*FC + 1.4178*BD - .69692*BD^2 -.010914*AS^3 - .10334*AS*BD + .049121*FC*BD$$

$$\Delta HC2 = -51672*FC + .054463*FC^2 + 1.4248*BD^2 - .81796*AS^3 -6.5966BD^3 + .093618*AS*FC + 9.074*AS*BD$$

where ΔBD1, ΔW1, and ΔHC1 and ΔBD2, ΔW2, and ΔHC2 are relative changes (unitless) in bulk density, water content and hydraulic conductivity for Barnes loam and Hamerly clay, respectively. BD (Mg/m^3), AS, and FC represent initial bulk density, aggregate size, and numbers of freeze cycles, respectively. These equations have r^2 values of 0.99, 0.95, 0.48, 0.99, 0.99, and 0.23, respectively, all highly significant ($p < 0.01$). The corresponding standard errors for these equations are 0.049, 0.185, 1.248, 0.066, 0.083, and 7.30, respectively.

The relative change in bulk density, water content, and hydraulic conductivity are all represented by equations that are functions of initial soil bulk density, soil aggregate size group, and numbers of freeze-thaw cycles. By holding any one of the three independent variables constant, a surface can be generated that shows how each dependent variable responds to changes in the remaining two independent variables. For example, at a constant initial bulk density of 1, the density of a Hamerly clay increases for each aggregate size group with each freeze cycle evaluated (Fig. 1a). At an initial density of 1.4, the first freeze cycle caused a density decrease, particularly for aggregate size group 3 (5-12 mm), with subsequent freeze cycles causing increased density (Fig. 1b). Data for a Barnes loam show many of the same general trends except that freeze cycles cause little density change for the first aggregate size group (0-0.5 mm) at an initial density of 1 (Fig. 1c). At an initial density of 1.4, repeated freeze cycles cause increased density after the first freeze for aggregate size group 3 (5-12 mm) but decreased density for aggregate size groups 1 (0-.5 mm) and 2 (1-3 mm).

Response surfaces generated by the multiple regrssion equations show that relative hydraulic conductivity generally increased with repeated soil freezing at initial bulk densities of 1.0 and 1.4, with the amount of increase being greatest for aggregate size group 2 (1-3 mm) for Hamerly clay (Fig. 2a and b). The Barnes soil showed a decreasing relative hydraulic conductivity with repeated freeze cycles for an initial bulk density of 1 (Fig. 2c) with values less than one for aggregate size group 3 (5-12 mm) indicating that final hydraulic conductivity was always less than initial. At an initial density of 1.4 (Fig. 2d) relative hydraulic conductivity increased with freeze cycles indicating improved water flow.

The change in saturated-free drainage water content after freezing Hamerly clay and Barnes loam samples at constant initial bulk densities of 1 or 1.4 (Fig. 3a-d) shows a relative water content less than 1 in all cases indicating reduced final water retention capabilities. However, after the first freeze cycle, increasing numbers of freeze cycles caused a decreased relative water content at an initial bulk density of 1 (Fig. 3a) but an increase at an initial bulk density of 1.4 (Fig. 3b). The least freeze-thaw effect is shown for aggregate size group 2 (1-3 mm). The Barnes soil showed a continually decreasing relative water content for all freeze cycles evaluated for both constant initial bulk

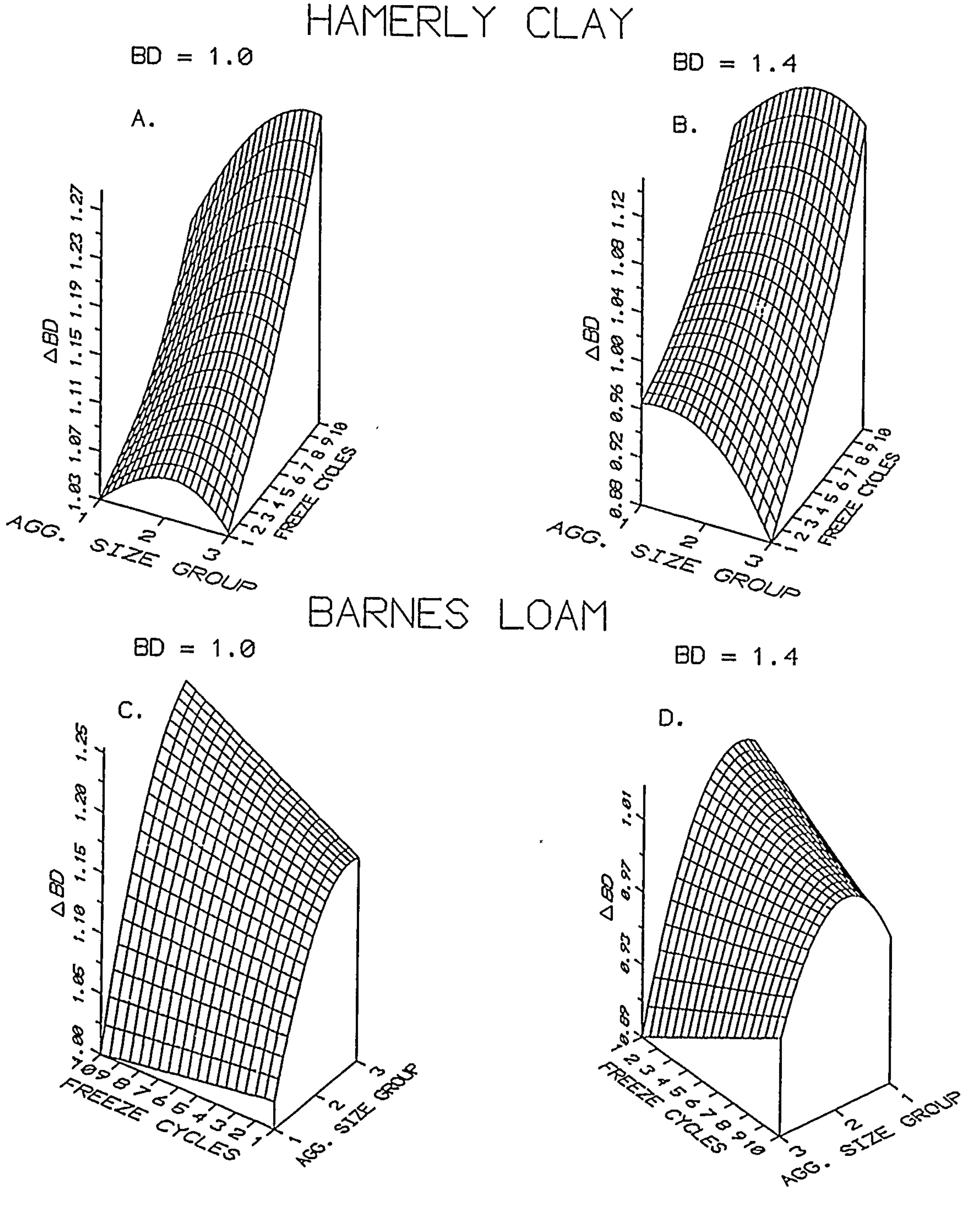

Figure 1: Relative change in soil bulk density after freeze-thaw treatment (ΔBD = Final BD/Initial BD) as related to aggregate size groups 1 (0-0.5 mm), 2 (1-3 mm), and 3 (5-12 mm) and number of freeze-thaw cycles (FC) at constant bulk density (BD) of 1.0 and 1.4 Mg/m^3 for a Hamerly clay and a Barnes loam.

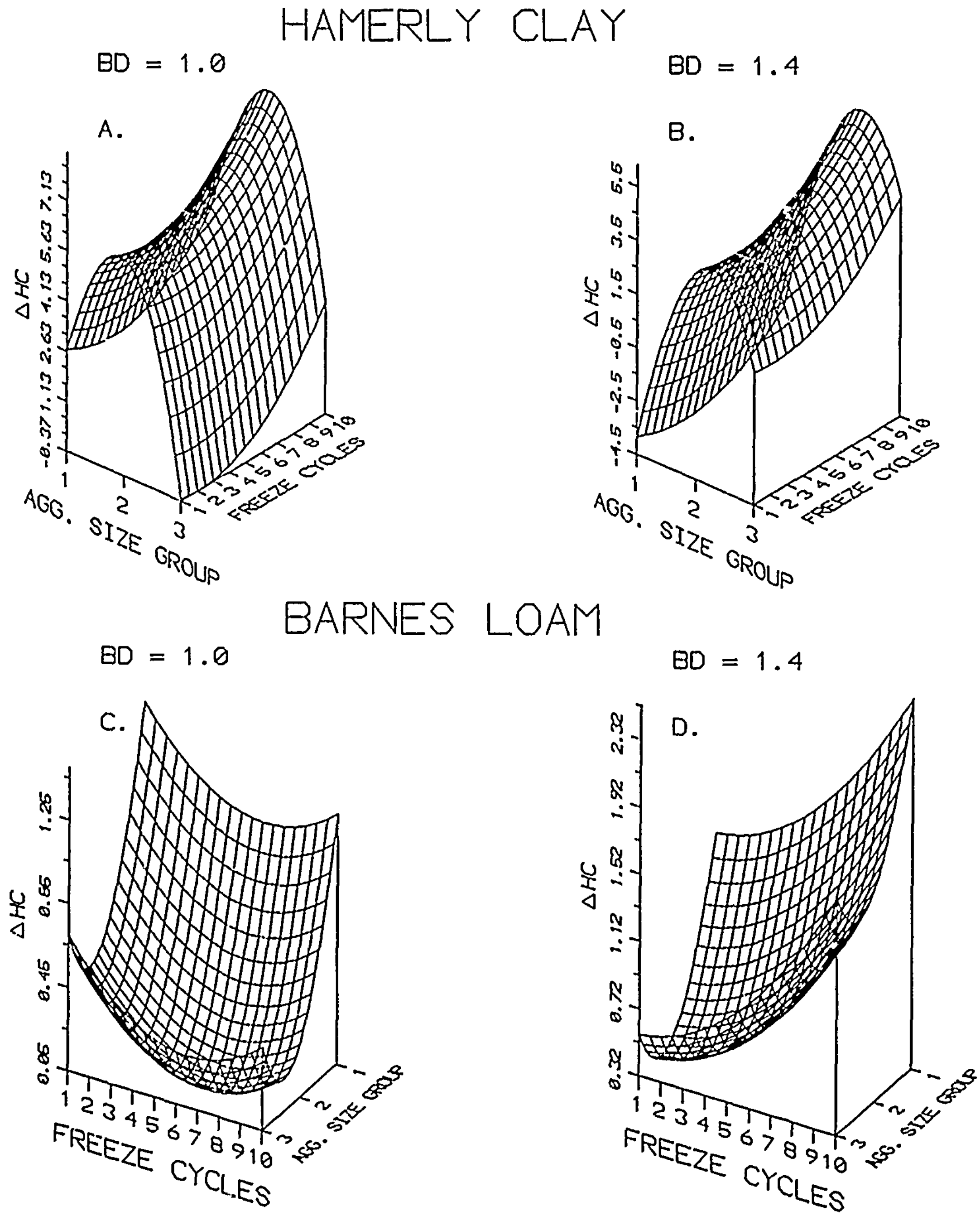

Figure 2: Relative change in soil hydraulic conductivity after freeze-thaw treatment (ΔHC = Final HC/Initial HC) as related to aggregate size groups 1 (0-0.5 mm), 2 (1-3 mm) and 3 (5-12 mm) and numbers of freeze-thaw cycles (FC) at a constant bulk density (BD) of 1.0 and 1.4 Mg/m^3 for a Hamerly clay and a Barnes loam.

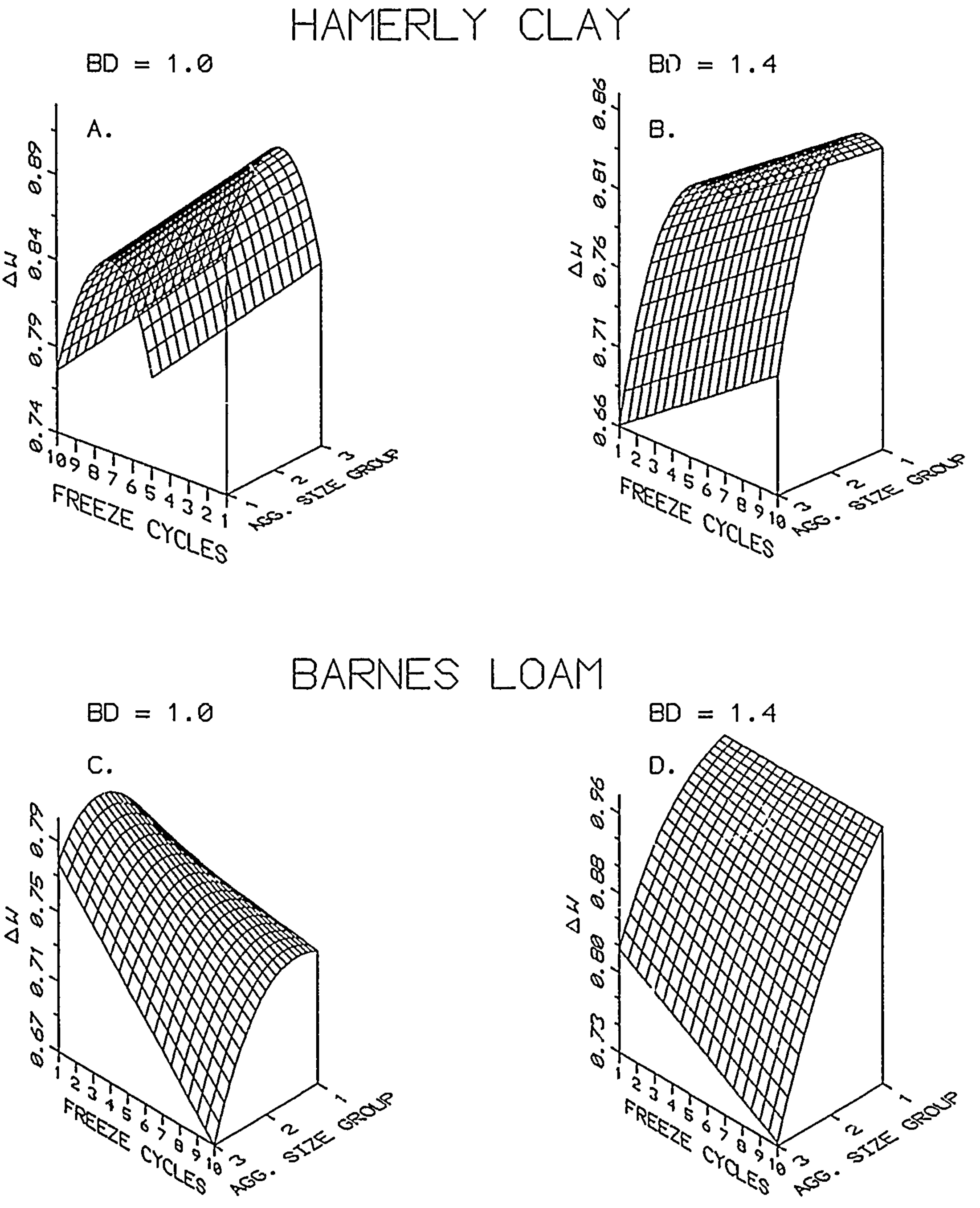

Figure 3: Relative change in saturated-free drainage soil water content after freeze-thaw treatment (ΔW = Final W/Initial W) as related to aggregate size groups 1 (0-0.5 mm), 2 (1-3 mm), and 3 (5-12 mm) and numbers of freeze-thaw cycles (FC) at a constant bulk density of 1.0 and 1.4 Mg/m^3 for a Hamerly clay and Barnes loam.

density values with the greatest total decrease associated with a constant initial bulk density of 1 and aggregate size group 3 (5-12 mm; Fig. 3c).

The relative change in bulk density that occurs after 10 successive freeze cycles as a function of initial bulk density and changing aggregate size groups show that freezing causes an increase in density at low initial densities and a decrease at high initial bulk densities for both Hamerly clay (Fig. 4a) and Barnes loam (Fig. 4b). The greatest density decrease is associated with high initial bulk densities and aggregate size group 3.

For constant aggregate size group 2 (1-3 mm) the relative change in bulk density increases with numbers of freeze cycles for Hamerly clay (Fig. 4c) but not for Barnes loam (Fig. 4d). However, again at low initial bulk density, freezing causes increased final bulk density values while the reverse is true for high initial density values.

SUMMARY

These data show that changes in soil density, saturated water-holding capacity and hydraulic conductivity as a result of freeze-thaw activity is significantly related to soil type, initial aggregate size, initial bulk density, and number of freeze cycles. In general, freezing and thawing effects are complex, with the magnitude of change recorded depending on the particular mix of initial conditions evaluated. For example, changes in density caused by freeze cycles at high initial densities and small initial aggregates may not be the same as for low initial density and large or small initial aggregate sizes. Similar statements can be made for changes in hydraulic conductivity and saturated-free drainage water contents.

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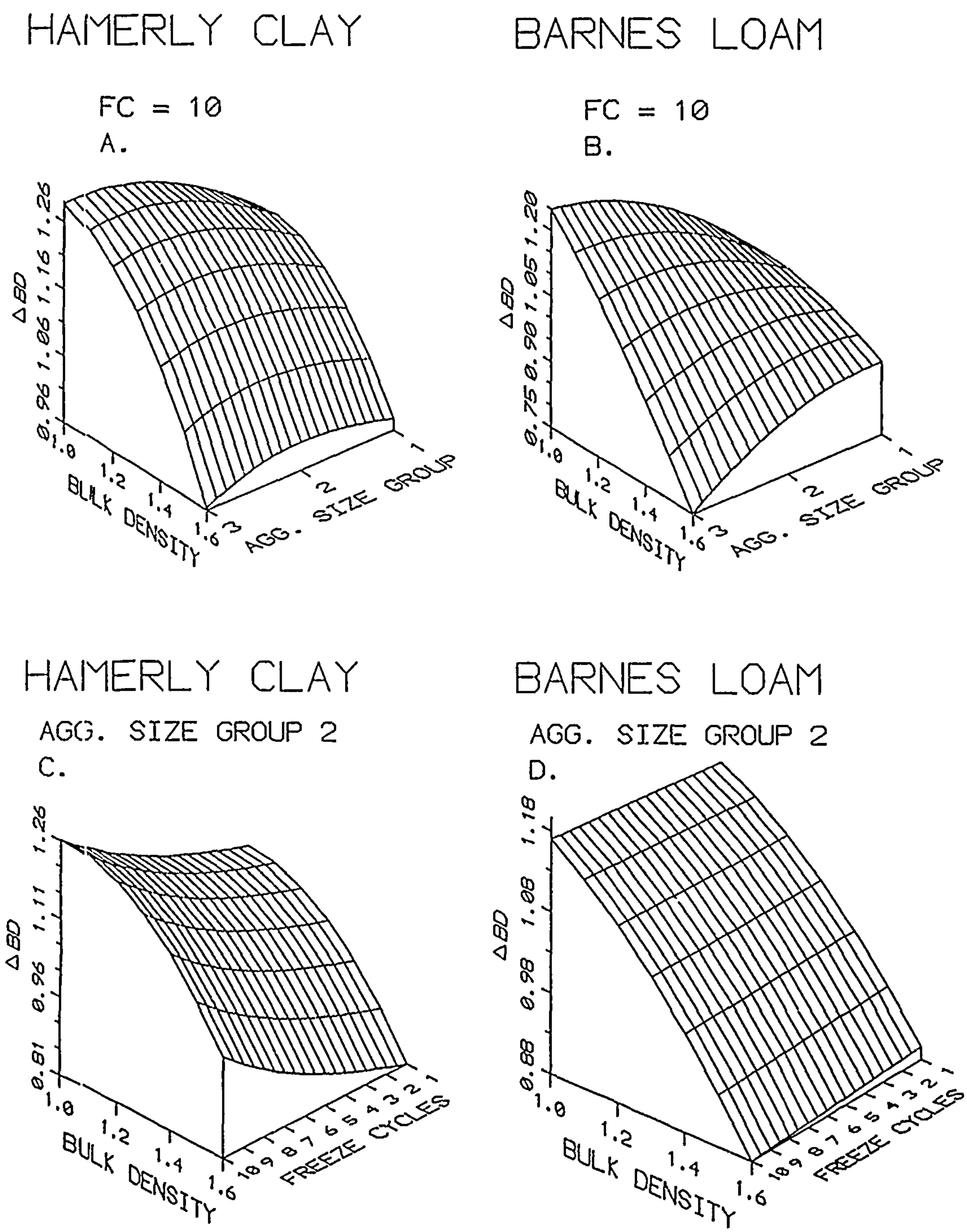

Figure 4: Relative change in bulk density after freeze-thaw treatment (ΔBD = Final BD/Initial BD) after 10 freeze-thaw cycles as related to initial bulk density and aggregate size groups; and for aggregate size group 2 as related to initial bulk density and number of freeze cycles for a Hamerly clay (A and C) and a Barnes loam (B and D).

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PREDICTING UNFROZEN WATER CONTENT BEHAVIOR
USING FREEZING POINT DEPRESSION DATA

P.B. Black and A.R. Tice[1]

INTRODUCTION

It has been recognized since the pioneering work of Taber (1929,1930) and Beskow (1935) that the dynamics of soil freezing is controlled by the continuous liquid layer that separates the soil matrix from the ice in frozen soil. All constitutive relationships describing the transport of mass and energy through frozen soil can be written as explicit functions of unfrozen water content. It thus becomes essential to be able to know the behavior of unfrozen water content if we are to determine the freezing behavior of soil.

Unfortunately, the complex interactions between the soil's physical and chemical environment prohibit, at this time, direct prediction of unfrozen water content from basic physical and chemical properties. Instead, unfrozen water content is empirically determined by nuclear magnetic resonance (NMR), time domain reflectometry (TDR), differential scanning calorimetry (DSC), and dilatometry. There are several detailed reviews of these techniques (Anderson and Morgenstern, 1973; Smith and Tice, 1988), but for the purpose of this paper, the important trait shared by these methods is the requirement of expensive and complicated equipment. We propose another, simpler method, that of freezing-point depression (FPD), to determine the unfrozen water content behavior in frozen soil. While the FPD method is not new (Bouyoucos and McCool, 1916; Schofield, 1935), it is most often used to infer the thermodynamic state of water in dry soils and is only occasionally employed as a method of determining unfrozen water content (Kozlowski, 1989).

This paper presents a framework by which freezing point depression data are interpreted to determine the unfrozen water content behavior of a soil. The transformed data are then fitted to a Brooks and Corey type function and compared to the unfrozen water content behavior determined by separate warming curve data that were measured by NMR.

MODEL

When discussing the state of water in frozen soil-water systems, it is convenient to employ the generalized Clapeyron equation

$$u_w - u_i/\gamma_i = [h/\theta_o]\theta \qquad (1)$$

where the ice and water phases are respectively denoted by the subscripts i and w; u is the pressure, h is the latent heat of fusion, γ is the specific

Presented at the Frozen Soil Symposium, March 21-22, 1990, Spokane, Washington.
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gravity, θ_o is the equilibrium temperature of pure water, and θ is the temperature of the system in question. Now if we define the difference between the ice and water pressures as

$$\phi_{iw} \equiv u_i - u_w \tag{2}$$

the Clapeyron equation can be expressed as

$$\phi_{iw} - [\gamma_{iw} - 1]u_w - [\gamma_i h/\theta_o]\theta \tag{3}$$

Brooks and Corey (1964) found that they could satisfactorily fit ice free water retention curve data to the relation

$$\frac{W-W_d}{W_s-W_d} = \left(\frac{\phi_{wa}}{\phi_b}\right)^{\alpha}; \quad \phi_{wa} < \phi_b \tag{4}$$

where W_d, W_s and W are respectively the water contents at the lower limit of drying, at saturation, and at the matric pressure, ϕ_{wa}, in question. In this case, the subscript wa refers to the difference between water and air pressures, also called the matric pressure, and ϕ_b is the "air-entry" value. The parameter α is obtained from a best fit to the data and has been found to have a large value for nearly uniform grain sizes.

Black and Tice (1989) showed that Eq. 4 can be transformed to represent unfrozen water content and simplified by making the reasonable assumption that W_d is zero. They proposed that unfrozen water content data be represented by

$$\frac{W}{W_s} = \left(\frac{\phi_{iw}}{\phi_b}\right)^{\alpha}; \quad \phi_{iw} > \phi_b \tag{5}$$

where ϕ_b now takes upon the meaning of ice entry pressure. The data collected in this paper are fitted to Eq. 5 using the relationship given by Eq. 3.

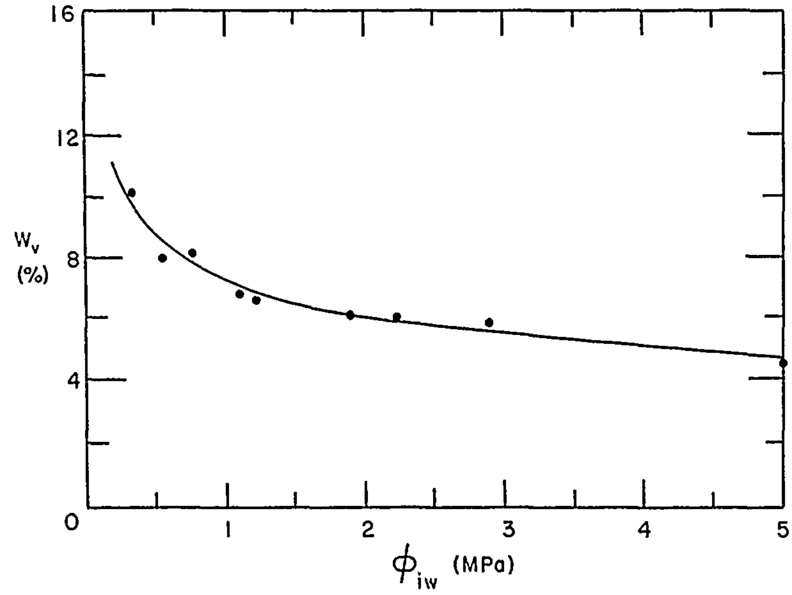
PROCEDURES

We collected unfrozen water content data determined by NMR and FPD for two soils, Chena Hot Springs silt and Tuto clay, whose physical properties are given in Table 1. Unfrozen water content data were determined using NMR in earlier experiments for warming curves by the procedure given by Tice et al.

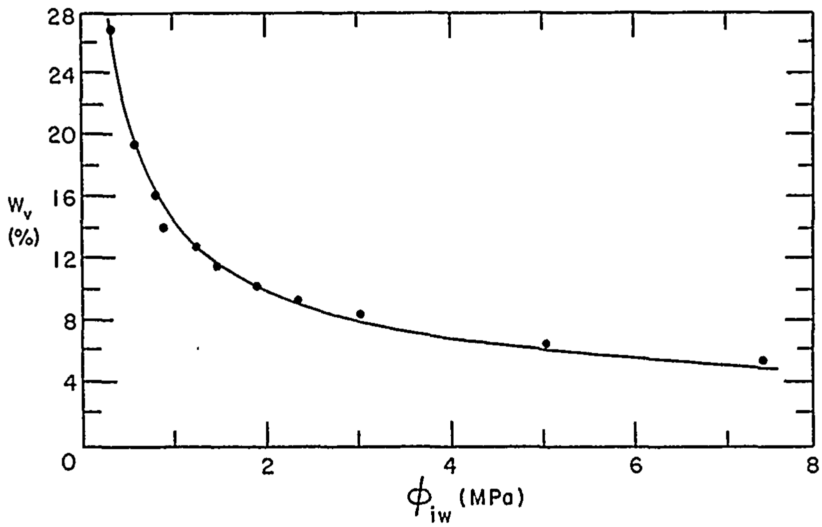
Table 1. Physical properties of Chena Hot Springs silt and Tuto clay.

			Percent passing sieve				
	D60	D10	4.6 (mm)	0.42 (mm)	0.02 (mm)	0.01 (mm)	G (Mg/m^3)
Chena Hot Springs silt[1]	0.027	0.005	100	100	39	20	2.80
Tuto clay							2.54

[1]Personal communication with R. Berg, CRREL.



a. Chena Hot Springs silt.

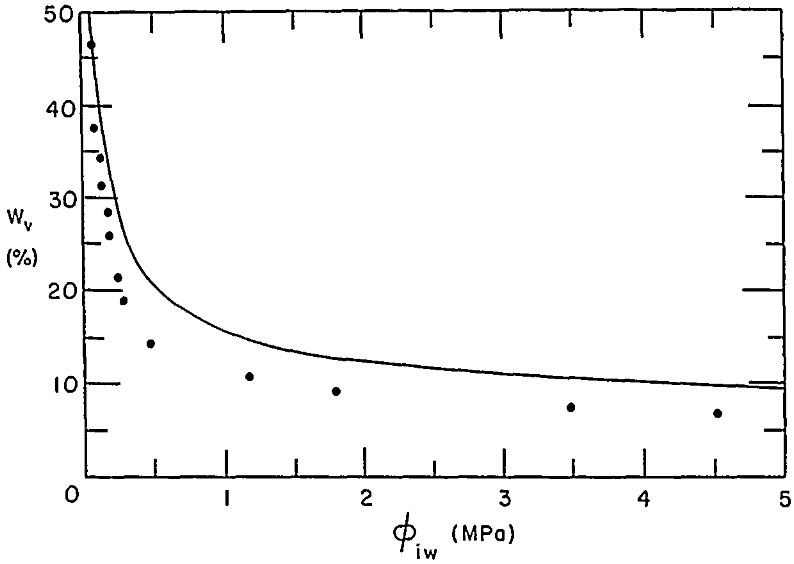


b. Tuto clay.

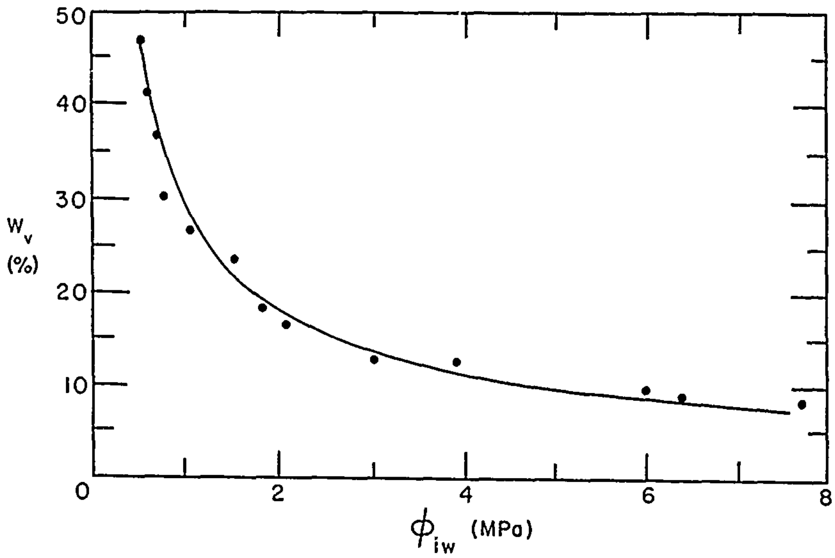
Figure 1. Unfrozen water contents determined by NMR for warming curve data and the best-fit curve to Eq. 5 using the parameters listed in Table 2.

(1978, 1981, 1982) and contained in CRREL's frozen soils database (Tice and Black, in preparation). The results of these measurements are presented in Fig. 1 in which the ϕ-values were obtained by assuming that u_w was zero in Eq. 3. This is a reasonable assumption since the specimens were compacted in tubes exposed to the atmosphere.

Additional 8 cm^3 specimens were prepared in the test tubes for the FPD measurements to different initial water contents ranging from 3 to 68 (cm^3 water/cm^3 total) and individually cooled to -5°C to -20°C. After thermal equilibrium was obtained, the supercooled specimen was struck to initiate ice nucleation and the resulting temperature increase was measured by an embedded calibrated thermistor. The amount of freezing-point depression was determined by calculating the difference between the temperature at the plateau in the re-



a. Chena Hot Springs silt.



b. Tuto clay.

Figure 2. Unfrozen water contents determined by FPD, assuming all soil water freezes, and the best-fit curve to Eq. 5 using the parameters listed in Table 2.

corded temperature rise and 0°C. The specimens were then oven-dried and the actual water contents and densities determined. From these data, the magnitude of freezing-point depression at each water content was obtained. Again Eq. 3 was employed to determine the ϕ-value with uw assumed to be zero and this time θ was set equal to the magnitude of freezing-point depression. The outcome of these transformed measurements are also shown in Fig. 2 in which the water contents are assumed to be equal to the total amount of water contained in each specimen.

Table 2. Statistical results of unfrozen water content measurements determined by NMR and FPD fitted to Eq. 5 for Chena Hot Springs silt and Tuto clay. In each case, W_s was set fixed to the porosity.

	Method	W_s	ϕ_b	α	r^2	error^2
Chena Hot Springs silt	FPD	42.8	0.79	-0.55	0.995	38.86
	NMR	42.8	0.01	-0.26	0.998	0.81
Tuto clay	FPD	51.6	4.42	-0.70	0.995	43.42
	NMR	51.6	0.94	-0.53	0.998	4.21

RESULTS

Table 2 contains the results of fitting the NMR- and FPD-determined unfrozen water contents to Eq. 5 using the Gauss-Newton nonlinear optimization technique (ASYSTANT, 1988). In each case, the high r-squared and low error-squared indicate that Eq. 5 is sufficient to describe these individual data sets. However, these results do not give a measure of how correctly the FPD method measures unfrozen water content. While there may be a suitable statistic that determines how closely these nonlinear data compare, a more physically intuitive approach of comparing these methods is chosen for the purpose of this paper. In this approach, predicted water contents at identical temperatures are simultaneously plotted and trends observed.

The parameters presented in Table 2 are used in Eq. 5 to predict unfrozen water contents by each method for identical temperatures in the range from -0.3°C to -5°C. These predicted water contents are simultaneously plotted in Fig. 3. A perfect match between the two methods would be indicated by the pre-

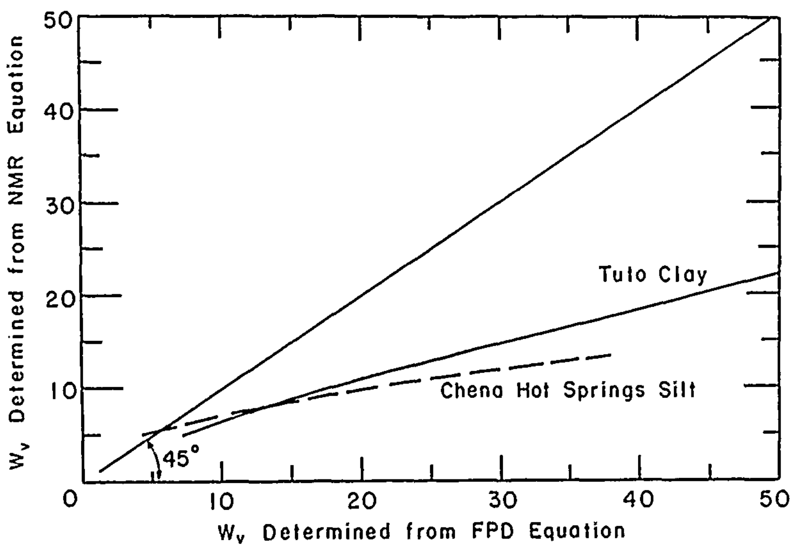


Figure 3. Comparison of unfrozen water contents predicted by FPD and NMR method for the temperature range of -0.3°C to -5.0°C and a 45° line representing identical correspondence.

dicted curves falling on the 45° line. Instead the FPD method tends to overestimate the amount of unfrozen water at a given temperature and skew the prediction curves beneath the 45° line. Further inspection of Fig. 3 reveals that the amount of overestimation is steadily increased as the temperature is increased. It is most unlikely that any statistical method could explain this deviation; hence, our intuitive approach is sufficient to explain the behavior of these methods if we can explain this deviation.

There are two physical factors influencing these data that would result in an apparent overestimation of unfrozen water content by the FPD method. The first physical factor is the bias introduced in the NMR method by using warming curve data. If we assume that the FPD method is more similar to the behavior of a cooling curve, then the hysteresis involved in the warming curve would underestimate the unfrozen water content at a given temperature and give results in the same direction as those in Fig. 3. Past measurements though have found this effect to be slight in the case of an Alaskan silt with characteristics similar to the Chena Hot Springs silt used in this study (Tice et al., 1988). The second physical factor included in these data is the implicit assumption that all available water froze in the FPD method. That is, the amount of water in the soil determined by oven drying at the end of the test was assumed to change completely to ice upon freezing and thus correctly represent the amount of unfrozen water for the measured amount of freezing-point depression. This clearly is not the case since there is always some water that does not freeze, especially for silt and clay; unfrozen water will exist in small pores and the adsorption space. This factor will therefore overestimate unfrozen water content and do so progressively. The warmer the temperature, the more unfrozen water will be remaining in the smallest pores and adsorption space, causing a greater overestimation, just as in the sense of Fig. 3.

CONCLUSIONS

The freezing-point depression (FPD) method of inferring unfrozen water content was compared to the more exact nuclear magnetic resonance (NMR) method and found to overestimate the amount of unfrozen water present in two soils, Chena Hot Springs silt and Tuto clay. It was further observed that the amount of overestimation progressively increased with warmer temperature, resulting in a doubling of estimated unfrozen water content at a temperature of approximately -0.3°C for these two soils. While this aspect might not be suitable for many applications, the ease and low cost of obtaining such data by the FPD method makes it an attractive alternative compared to not having any information on the soil.

ACKNOWLEDGMENTS

We would like to acknowledge the helpful suggestions of T. Pangburn, O. Ayorinde, I. Iskandar and T. Bergstad of CRREL during the preparation of this paper. Funding for this report was provided by DA Project 4A161102AT24, Research in Snow, Ice and Frozen Ground; Task A, Properties of Cold Regions Materials: Work Unit 001, Physical Properties of Frozen Ground.

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EFFECTS OF FREEZING ON AGGREGATE STABILITY OF SOILS DIFFERING IN TEXTURE, MINERALOGY, AND ORGANIC MATTER CONTENT

by

G. A. Lehrsch, R. E. Sojka,
D. L. Carter, and P. M. Jolley[1]

INTRODUCTION

Aggregate stability, a measure of a soil aggregate's resistance to breakdown, influences many soil physical and hydraulic characteristics, such as surface sealing rate, infiltration rate, and hydraulic conductivity. Thus, because aggregate stability is so important, processes that may increase or decrease it should be studied.

Different soils have been observed to respond differently to the freezing process. Hence, it was hypothesized that soils differing in texture, mineralogy, and organic matter content would be affected differently. A laboratory experiment was designed to test this hypothesis.

Recently, numerous studies of aggregate response to freezing have been reported. Aggregate stability has usually been inversely proportional to soil water content at the time of freezing (Bullock et al., 1988; Benoit, 1973; Bryan, 1971; Logsdail and Webber, 1959). Aggregates from poorly aggregated soils, however, are more stable when frozen at intermediate water contents (Mostaghimi et al., 1988; Sillanpaa and Webber, 1961). The number of freeze-thaw cycles to which aggregates have been subjected is important. With increasing freeze-thaw cycles, aggregate stability usually decreases (Logsdail and Webber, 1959; Mostaghimi et al., 1988; Willis, 1955) but may increase for some soils (Mostaghimi et al., 1988; Richardson, 1976).

Aggregates in soil samples constrained from expanding, especially in the horizontal direction, may respond to freezing differently than unconstrained aggregates. The stability of constrained aggregates has decreased more than that of unconstrained aggregates, though affected to an extent by the aggregate's size and water content at freezing (Bullock et al., 1988).

The influence of inorganic bonding agents on aggregate stability has also been widely studied. For example, Al-Ani and Dudas (1988) found that mean weight diameter increased with additions of calcium carbonate from 0 to 4% by weight but thereafter decreased with $CaCO_3$ additions from 4 to 32%. In contrast, Chepil (1954) reported that 3% calcium carbonate had no effect on aggregate stability but that 10% increased it. Others (Kemper and Koch, 1966; Toogood, 1978) found no significant effect due to lime content.

Many of the inconsistencies noted above may be due to different soils being used from experiment to experiment. The current investigation was conducted under a uniform set of conditions with six soils quite different in character, yet having paired soil properties. The experiment was composed of two studies. The objective of Study I was to determine the effects of constrainment, number of freeze-thaw cycles, and water content at freezing on the aggregate stability of six continental U.S. soils differing in texture, mineralogy, and organic matter. In Study II, two potential inorganic bonding agents at two rates were added as additional factors to a subset of the treatments of Study I.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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MATERIALS AND METHODS

The six soils studied (surface horizons only) were obtained from widely separated locations across the United States. The soils were: a Cecil sandy loam (clayey, kaolinitic, thermic Typic Hapludult) from Watkinsville, GA; a Barnes loam (fine loamy, mixed, Udic Haploboroll) from Morris, MN; a Sverdrup sandy loam (sandy, mixed Udic Haploboroll) from Elbow Lake, MN; a Sharpsburg silty clay (fine, montmorillonitic, mesic Typic Argiudoll) from Lincoln, NE; a Portneuf silt loam (coarse silty, mixed, mesic Xerollic Calciorthid) from Kimberly, ID; and a Palouse silt loam (fine silty, mixed, mesic pachic Ultic Haploxeroll) from Pullman, WA. Properties of the six soils (Table 1) were determined (Soil Conservation Service Staff, 1984) by the personnel of the National Soil Survey and Soil Mechanics Laboratories, Lincoln, NE.

Table 1. Soil properties.

Soil type	Particle size distribution: Sand	Silt	Clay	Bulk density	Predominant mineral type	pH	Organic matter content*
	%	%	%	g cm^{-3}		(in $CaCl_2$)	%
Barnes Loam	49	34	17	1.25	2:1	7.1	3.41
Cecil sandy loam	67	16	17	1.69	1:1	4.6	1.24
Palouse silt loam	10	70	20	1.15	2:1	4.5	3.03
Portneuf silt loam	22	66	12	1.24	2:1	7.8	1.24
Sharpsburg silty clay	3	56	41	1.33	2:1	5.4	3.19
Sverdrup sandy loam	76	15	9	1.43	2:1	6.0	2.21

*as estimated from the organic C content using the Van Bemmelen 1.724 factor.

Study I

A randomized complete block design with three replications and a factorial arrangement of treatments was used. Sources of variation were soils, number of freeze-thaw cycles, constrainment, and water content. Freeze-thaw cycles were either 0, 1, 3, or 5 with the 0 level signifying no freezing. The constrainment factor was at one of two levels, either constrained (in brass cylinders 5 cm high with an inside diameter of 2.75 cm) or unconstrained (aggregates placed loosely on Al weighing dishes). The water content factor (qualitatively either low, medium, or high) was quantitatively either 0.05, 0.15, or 0.25 g/g for the coarse-textured Sverdrup and Cecil soil or 0.10, 0.20, or 0.30 g/g for the remaining soils. It should be noted that subsequent figures, for the sake of uniformity, will indicate Sverdrup and Cecil soil samples to have water contents of 0.10, 0.20, or 0.30 g/g. Because some data were missing due to sample mistreatment in the laboratory, statistical analyses were performed using a linear model (SAS Institute, Inc., 1985). Aggregate stability values were reported as least-squares (or marginal) means (Searle et al., 1980). These least-squares means, estimates of the means that would have been obtained had no data been missing, were separated, utilizing a significance probability of 5%, using an option available in SAS (SAS Institute, Inc., 1985).

Samples were prepared by sieving field-moist soil (gravimetric water contents ranged from 7 to 22% and averaged 13%) through a 4-mm sieve. Soil was not permitted to air dry between the time it was sampled in the field and analyzed in the laboratory. To prepare constrained samples, the water content of the sieved soil was first adjusted slowly to the desired level, i.e., either lowered by drying in air or raised by misting in a vaporizer (Kemper and Rosenau, 1986). Moist soil was packed by tapping into each brass cylinder until a dry bulk density of 1.15 g/cm^3 was reached. Each packed cylinder was sealed in a polyethylene bag to both inhibit water loss and prevent water uptake, inserted into a styrofoam tray, and stored at +6°C until the remaining cylinders were packed. The styrofoam, a minimum of 7 cm underneath and 2 cm around each cylinder, served as insulation so that freezing occurred primarily downward from the surface. Unconstrained samples were prepared by sieving the less than 4-mm field-moist soil through a 1-mm sieve and placing the equivalent of 10 g of oven-dry 1- to 4-mm aggregates in an Al weighing dish. Our interest was in the response of the 1- to 4-mm aggregates. When studying unconstrained samples, the less than 1-mm aggregates would by design exert no confining pressure on the desired size fraction of aggregates and thus could be omitted from the unconstrained samples. Each dish was sealed in a polyethylene bag, placed on a plastic tray, and stored at +6°C until the remaining unconstrained samples were prepared.

All prepared samples were subjected to either 0, 1, 3, or 5 freeze-thaw cycles. One cycle was completed when prepared soil samples were frozen at -14°C for 24 hours, and subsequently thawed at +6°C for 48 hours. Samples were convectively frozen without access to water. Freezing-induced vertical expansion of the soil in the cylinders was measured. For all samples, a data logger within each enclosure recorded ambient air temperatures. The 0 cycle samples were not frozen but were stored at +6°C for a minimum of 48 hours. Before the aggregate stability analysis, all samples were brought to room temperature on a lab bench for 2 hours. Aggregate stability was determined using the procedure of Kemper and Rosenau (1986) modified so that field-moist 1- to 4-mm aggregates were vapor-wetted to 0.30 g/g prior to wet sieving.

Study II

The experimental design was the same as for Study I but the sources of variation were different. Factors were soils, number of freeze-thaw cycles, water content, potential bonding agents, and bonding agent addition rates. Freeze-thaw cycles were either 1 or 3. All samples were constrained in brass cylinders. The water content factor was either 0.05 or 0.25 g/g for the coarse-textured Sverdrup and Cecil soil or 0.10 or 0.30 g/g for the remaining soils. Finely ground reagent grade $CaCO_3$ or $CaSO_4$ was added as a bonding agent at a rate of either 0.2 or 1.0% by weight of oven-dry soil. The statistical analysis was similar to that of Study I. Least-squares means were separated utilizing a significance probability of 5%.

Samples were prepared by coating less than 4-mm field-moist soil with ground $CaCO_3$ or $CaSO_4$. The desired mass of $CaCO_3$ or $CaSO_4$ was distributed as evenly as possible over all surfaces of a known mass of sieved soil on plastic sheeting, and thoroughly mixed. Thereafter, in a vaporizer, soil water content was raised to the desired level as described earlier. This moist, coated soil was then packed into a brass cylinder using the procedure of Study I. All packed cylinders were subjected to either 1 or 3 freeze-thaw cycles. After the soil had thawed the last time, it was permitted to warm to room temperature over a 2-hour period, removed from the cylinder, and analyzed for aggregate stability using the procedure of Study I.

RESULTS AND DISCUSSION

Study I

Aggregate stability (averaged over freeze-thaw cycles) is shown as a function of water content in Figure 1 for constrained samples and in Figure 2 for unconstrained samples. Figure 1 reveals that, for each soil, aggregate stability decreased as water contents increased from 10 to 30%. The stability of the Sharpsburg, Palouse, and Barnes soils (all medium-textured or finer with organic matter contents of 3% or more) decreased twice as much from 20 to 30% than from 10 to 20% water content. The stability of the Portneuf silt loam, with nearly 60% less organic matter (1.24%, Table 1), however, dropped just as much from 20 to 30% as it did from 10 to 20%. Thus, elevated organic matter, known to improve the stability of aggregates with diameters over 0.25 mm (Tisdall and Oades, 1982), was also more effective in stabilizing aggregates frozen at lower than at higher water contents. The coarse-textured Cecil exhibited the greatest drop in aggregate stability from 10 to 20%. Large differences in aggregate stability exist from soil to soil even at the same water content. For example, Portneuf and Palouse, two silt loams from the Pacific Northwest, differ in stability by over 25 percentage points at a 10% water content.

Findings for constrained samples (Fig. 1) were similar to those for unconstrained samples (Fig. 2). Some differences were noted, however. In Figure 2, though the difference was not significant, aggregate stability tended to increase slightly from 10 to 20% for the Sharpsburg, Palouse and Barnes soils. Figure 2 also indicates that the Sharpsburg, Palouse, and Barnes, the soils highest in organic matter, had the highest percentages of stable aggregates when frozen at a water content of 10%. These same three soils were still the most stable when frozen at 30%.

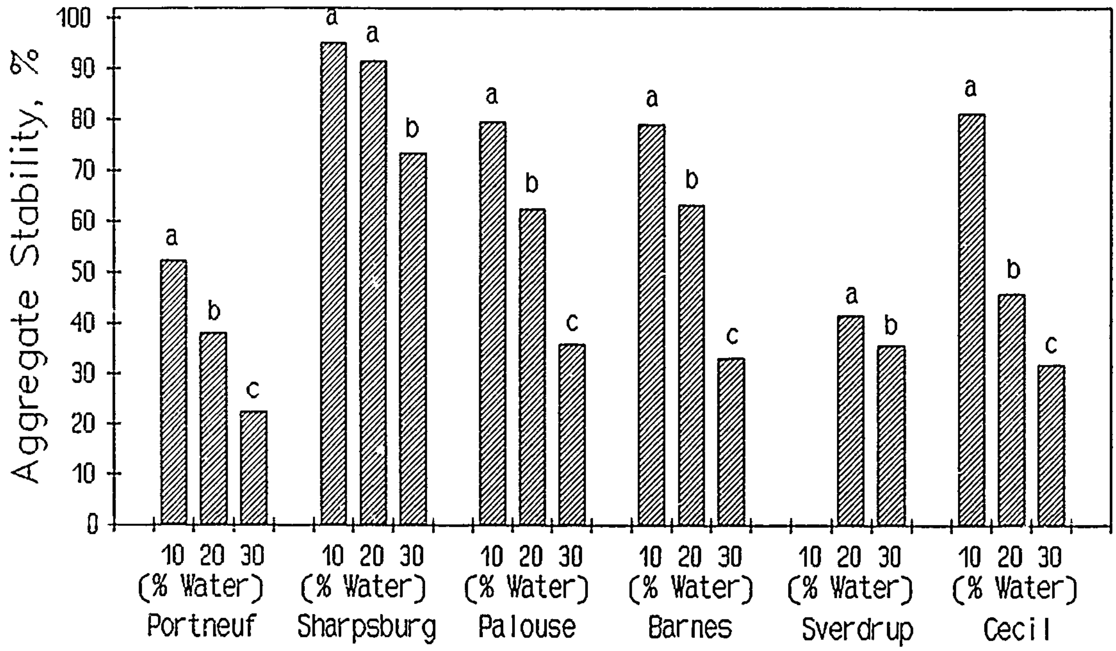

Figure 1. Aggregate stability (averaged over freeze-thaw cycles) as a function of water content for constrained samples of each soil. Within each soil, means without a common letter differ significantly at the 0.05 level.

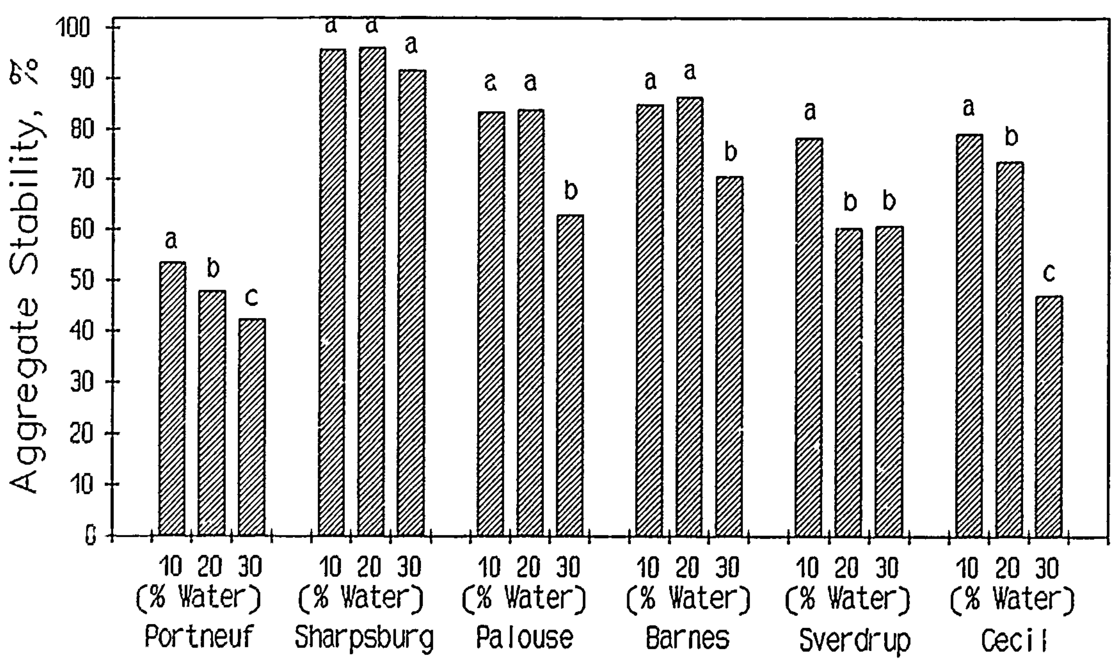

Figure 2. Aggregate stability as a function of water content for unconstrained samples of each soil. Within each soil, means without a common letter differ significantly at the 0.05 level.

In Figure 3, the data (again averaged over freeze-thaw cycles) for a Palouse silt loam illustrate a typical response to constrainment of aggregates while frozen at different water contents. For every soil, constrainment decreased the stability of aggregates frozen at higher (20-30%) water contents. Expansion of ice crystals likely caused planes of weakness in the aggregates whose horizontal displacement was limited by the confining pressure of the brass cylinders. These planes of weakness subsequently manifested their presence during the wet sieving process. Figure 3 also indicates that aggregate stability decreased faster with increasing water content when frozen constrained rather than unconstrained.

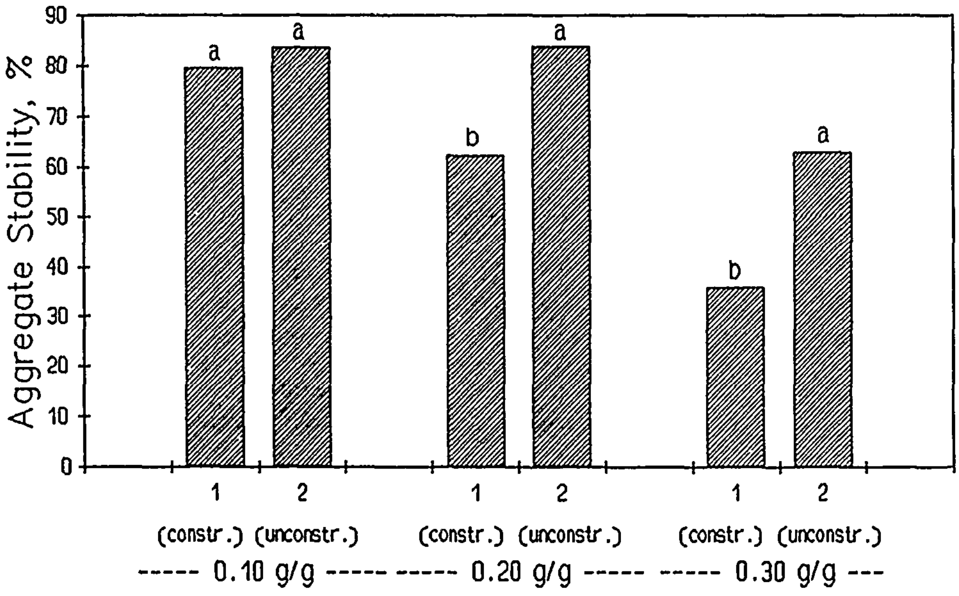

Figure 3. Stability of Palouse aggregates as affected by constrainment at each water content. Within each water content, means without a common letter differ significantly at the 0.05 level.

Table 2 lists aggregate stabilities, averaged over water content, of constrained samples of each soil after each freeze-thaw cycle. The Palouse and Barnes soils increased significantly in stability with increasing number of freeze-thaw cycles. Both are medium-textured soils with relatively high organic matter contents.

Table 2. Effect of freeze-thaw cycles on aggregate stability (constrained samples only).

	Aggregate stability			
	Number of freeze-thaw cycles			
Soil	0	1	3	5
	%			
Barnes	53.0 c*	56.3 bc	59.6 ab	65.4 a
Cecil	49.5 a	55.1 a	55.8 a	51.5 a
Palouse	43.5 c	60.6 b	71.1 a	62.1 b
Portneuf	38.1 a	38.1 a	36.5 a	38.4 a
Sharpsburg	82.8 a	86.0 a	88.7 a	89.1 a
Sverdrup	45.8 a	46.5 a	-	50.6 a

*Means within a row not followed by a common letter differ significantly at the 0.05 level.

When unconstrained, four soils (Fig. 4) increased significantly in aggregate stability with freeze-thaw cycles. In contrast, when constrained, the stability of only two soils increased with freeze-thaw cycles (Table 2). Figure 4 reveals that the coarse-textured Sverdrup and Cecil and, as before, the Palouse and Barnes soils increased in stability. The aggregate stability of Cecil and Palouse increased with each subsequent freeze-thaw cycle, but the response of Sverdrup and Barnes was more variable. The Palouse silt loam was the only soil whose stability increased statistically (and likely practically) from 0 to 1 freeze-thaw cycle. For the remaining three soils, the improvement in stability was minimal, averaging less than nine percentage points over the entire range of cycles studied.

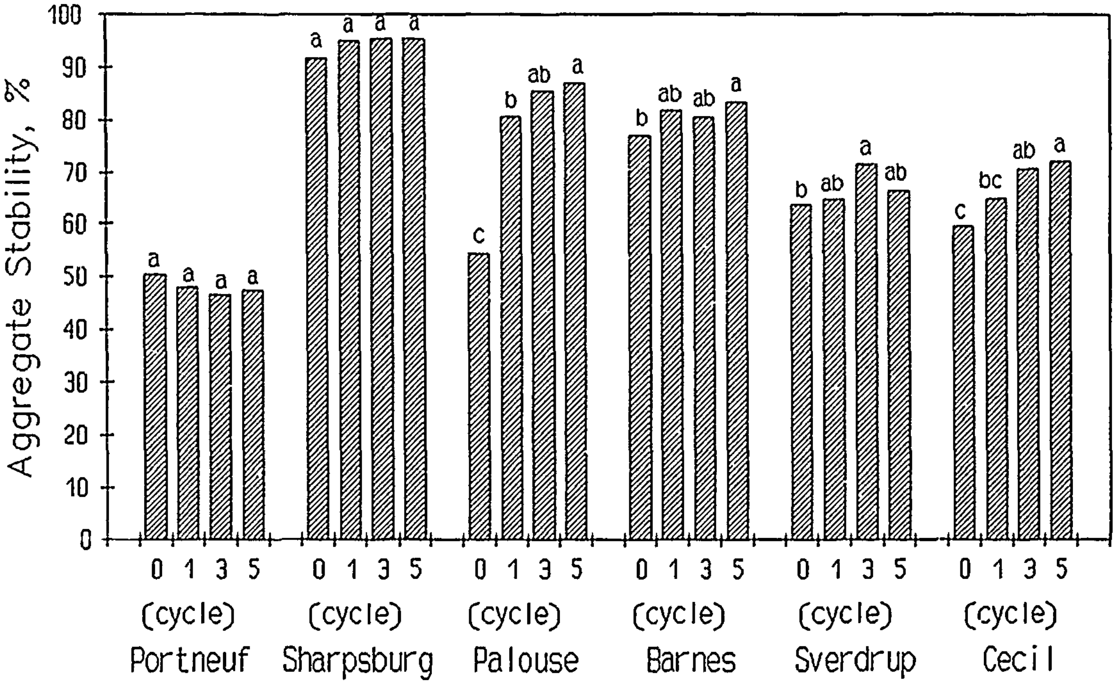

Figure 4. Effects of freeze-thaw cycles on the stability (averaged over water content) of unconstrained aggregates of each soil. Within each soil, means without a common letter differ significantly at the 0.05 level.

Figure 5 reveals that, within each freeze-thaw cycle, aggregate stability, averaged over all six soils and both constrainment levels, decreased with increasing water content. Also, at least for water contents of 20 and 30%, the stability of aggregates increased when subjected to up to three freeze-thaw cycles but then decreased when subjected to five cycles. Thus, for some soils, there seems to be some rather low number of freeze-thaw cycles beyond which aggregates may become less stable with continued freezing and thawing. This suggests that to minimize soil structural damage (and attendant erosion), soil water contents (at least at the surface of the profile) should be as low as possible entering the winter season (Benoit, 1973). Figure 5 also shows that the higher the water content, the greater the increase or decrease caused by freeze-thaw cycles. In other words, freeze-thaw cycles exerted their greatest effect, whether beneficial or detrimental, on aggregate stability at the highest water contents.

Study II

Results from Study II confirmed a number of the findings of Study I. First, aggregate stability decreased when water contents increased from 10 to 30%. Second, the decrease in aggregate stability varied from soil to soil, in general being greatest for coarse-textured soils and least for fine-textured soils. Of particular interest, however, was the influence of $CaCO_3$ and $CaSO_4$ on aggregate stability. Statistical analysis indicated that, at the 5% level, neither of the potential bonding agents affected

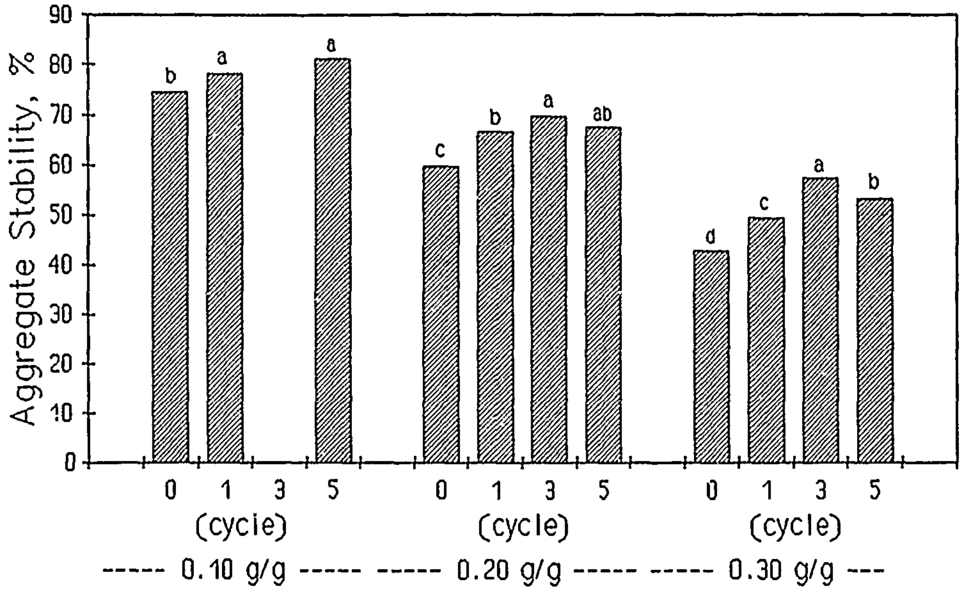

Figure 5. Aggregate stability (averaged over soils and constrainment levels) as a function of freeze-thaw cycles at each water content. Within each water content, means without a common letter differ significantly at the 0.05 level.

aggregate stability. The effect of the bonding agents did, however, approach significance at the 10% level. The trend of the response indicated that aggregate stability was higher with the addition of calcium sulfate than calcium carbonate. Moreover, aggregate stability tended to decrease with increasing rates of the bonding agents. This decrease in stability was greatest when $CaCO_3$ was added to soil samples at a water content of 30%. Statistical analyses of these data are continuing.

Results in light of the objectives

Interactions among the factors examined were the rule rather than the exception in both studies. Constrainment, number of freeze-thaw cycles, and water content at freezing were all found to significantly affect the stability of the six soils, though differently depending upon both the soil in question and the levels of the other factors. In Study I, the effect of constrainment differed depending upon the soil, the water content, and the number of freeze-thaw cycles. When the six soils were taken as a whole, the effect of freeze-thaw cycles on aggregate stability depended upon the water content at freezing. In Study II, the decrease in aggregate stability caused by an increase in water content differed according to the texture of the soil being frozen.

Results in relation to others' findings

The results of these two studies are similar in most but not all respects to the results obtained by other investigators. The decrease in aggregate stability with increasing water content, found by others for loam and silt loam soils (Bullock et al., 1988; Benoit, 1973), was confirmed and found to occur for finer- and coarser-textured soils as well (Fig. 1 and 2). We found that constrained aggregates were less stable than unconstrained aggregates after freezing (Fig. 3) as reported by Bullock et al. (1988). As found by Mostaghimi et al. (1988) for unconstrained samples of a silt loam and Richardson (1976) for a different silt loam, aggregate stability may increase with increasing number of freeze-thaw cycles (see Fig. 4, especially the Palouse silt loam). When, as in this study, soils are not air-dried prior to analysis, this increase in stability with freeze-thaw cycles may be the norm rather than the exception (Figs. 4 and 5). The finding of this study that aggregate stability (as an average response for six soils) increases with up to three freeze-thaw cycles but decreases thereafter (Fig. 5) is supported by the results of Mostaghimi et al. (1988) for a Crofton silt loam (fine silty, mixed, mesic Typic Ustorthent). In contrast, the results presented in Figure 4 showing an increase in stability with freeze-thaw cycles for unconstrained samples of four of six soils are at odds with the results presented by Mostaghimi et al. (1988) showing a decrease in stability with freeze-thaw cycles for two of three soils. Such a discrepancy may well result because their soils were air-dried before their experiment was conducted. Air-drying could have strengthened bonds within the aggregates, thus effectively masking the increase in aggregate stability detected for the first few freeze-thaw cycles (Fig. 5). The finding in Study II that $CaCO_3$ added at either 0.2 or 1.0% by weight did not significantly affect aggregate stability agrees with the results of Chepil (1954), Kemper and Koch (1966), and Toogood (1978) but disagrees with those of Al-Ani and Dudas (1988). There were, however, differences in both soils and procedures between the study of Al-Ani and Dudas (1988) and Study II. The soil samples used by Al-Ani and Dudas (1988) were from the BC horizons of two soils with clay contents of 32 or 42% whereas the soil samples used in Study II were from the surface horizons of soils generally of much lower clay content (Table 1). Procedurally, air-dry soil and 4- to 8-mm aggregates were used by Al-Ani and Dudas (1988) while field-moist soil and 1- to 4-mm aggregates were used in Study II. A number of the earlier studies that found $CaCO_3$ to affect aggregate stability used very high calcium carbonate application rates. The rates considered in Study II were more feasible application rates for agricultural production.

CONCLUSIONS

Constrained aggregates at gravimetric water contents of 20% or more were less stable after freezing than were unconstrained aggregates. For some soils particularly when unconstrained, aggregate stability tended to first increase with increasing number of freeze-thaw cycles and then decrease. The mean response of all six soils showed that the effect of different numbers of freeze-thaw cycles was most pronounced at the highest water content (30%). The one response most consistent throughout the experiment was that aggregate stability decreased with increasing water content at the time of freezing.

ACKNOWLEDGMENT

The authors gratefully acknowledge the contributions of Drs. Russ Bruce, John Gilley, Hans Kok, and Bob Young for initial soil sampling and of Miss Michelle Garrison, Miss Tonya Pearson, Mr. Ron Peckenpaugh, and Miss Deb Watkins for sample preparation and laboratory analyses.

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FREEZE THAW EFFECTS ON SOIL STRENGTH

by
Hans Kok and Donald K. McCool[1]

INTRODUCTION

Erosion occurs on the 4 million hectares of non-irrigated cropland in the Pacific Northwest at rates much higher than tolerable to sustain long-term productive agriculture. For example, ten percent of the Palouse river basin of eastern Washington and northern Idaho has already lost all of its topsoil (USDA, 1978) and current annual erosion rates for cropland average from 5 to 50 ton per hectare (Papendick et al., 1983). Rates as high as 224 ton per hectare have been reported (McCool et al., 1976a). Intensive tillage practices associated with common crop rotations, that frequently contain summer fallow, leave the soils pulverized and very vulnerable to erosion. The long and steep slopes in loessial soils of the region combined with the many freeze-thaw cycles (up to 120 per year) during the winter season result in extremely erodible conditions. Soils accumulate moisture prior to the freezing cycles and often freeze wet with resultant low infiltration capacities (McCool and Molnau, 1974). The majority of the precipitation occurs as low intensity rain or snow during the winter months when soils are bare and unprotected (McCool et al., 1982). Runoff detachment and transport are the dominant erosion processes (McCool et al., 1976a; 1987). More than 50 percent of the soil erosion in the Palouse region is related to rain or snowmelt on frozen or thawing ground, often accelerated by warm, moist Pacific air masses (McCool et al., 1976b; Yoo and Molnau, 1982; Zuzel et al., 1982), yet one of the least understood aspects of the physical erosion process is the effect of freeze and thaw.

This report will briefly summarize research conducted at Pullman, Washington pertaining to the effects of freeze-thaw cycles on soil properties.

FIELD RESEARCH

Field research has been conducted since 1931 at the Palouse Conservation Field Station (PCFS). A series of runoff plots have been installed since 1978 that are subjected to tillages and crop rotations common to the area. Results have been primarily used to adapt the Universal Soil Loss Equation to the Pacific Northwest (McCool and George, 1983, McCool et al., 1976a, McCool et al., 1982). On tillage weakened soils, 60 to 70 percent of the annual runoff and about 55 percent of the yearly soil loss was related to rainfall and snowmelt on frozen ground.

During the 1987-88 and 1988-89 erosion seasons, soil strength was monitored on a Palouse silt loam on a 21 percent slope at the PCFS (Kok, 1989, Kok and McCool, 1989). Several instruments were used; fall-cone, pocket penetrometer and Torvane*, a device with vertical vanes radiating from a central point. Satisfactory results were obtained only with a metal Torvane apparatus. Torvane soil strength varied drastically over the period of field testing, from virtually zero while thawing to 14 kPa under dry soil conditions in early spring.

An example of observations as obtained with the Torvane apparatus for early 1988 is shown in Figure 1, together with soil water content observations for this period. The soil had been tilled with a Lely Roterra (a tiller with vertical teeth rotating around a

* The mention of trade or brand names in this publication does not constitute endorsement or recommendation for use by the authors and is provided for the reader's information only.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990

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vertical shaft) the previous fall and had been in continuous fallow since 1978. Several freeze-thaw cycles occurred before the soil froze for several weeks in early 1988 followed by another series of freeze-thaw cycles until the soil shear strength was 6.4 kPa on day 89 (March 29). Snow and frost occurred and the strength decreased to 5.7 kPa at 1000 PST on day 91 (March 31). By 1600 PST sun and strong wind had dried the surface and the shear strength increased to 11 kPa. Overnight rain increased the soil water content to 25 percent but strength remained constant at 11 kPa on the morning of day 92 (April 1); by 1700 PST it had increased to 13 kPa, due to sun and very strong drying winds. Snow, rain, frost and thaw caused the strength to drop to 4.4 kPa at 1000 PST on day 96 (April 5); by 1600 PST it had increased to 10 kPa, and the next day it was 14 kPa. Rain caused the strength to decrease to 12 kPa overnight. A frost on the night of day 98 (April 7) might have resulted in a lower strength, since the soil water content went up to 36 percent; however by 1100 PST on day 99 (April 8) when the soil was completely thawed and strength measurements were possible, the water content had decreased to 24 percent and the strength was 13 kPa, by 1500 strength was 14 kPa. Further measurements were not possible since the soil was now hard and dry. Insertion of the Torvane caused cracking and crumbling of the soil. Similar trends as described here were observed for late 1988 and early 1989.

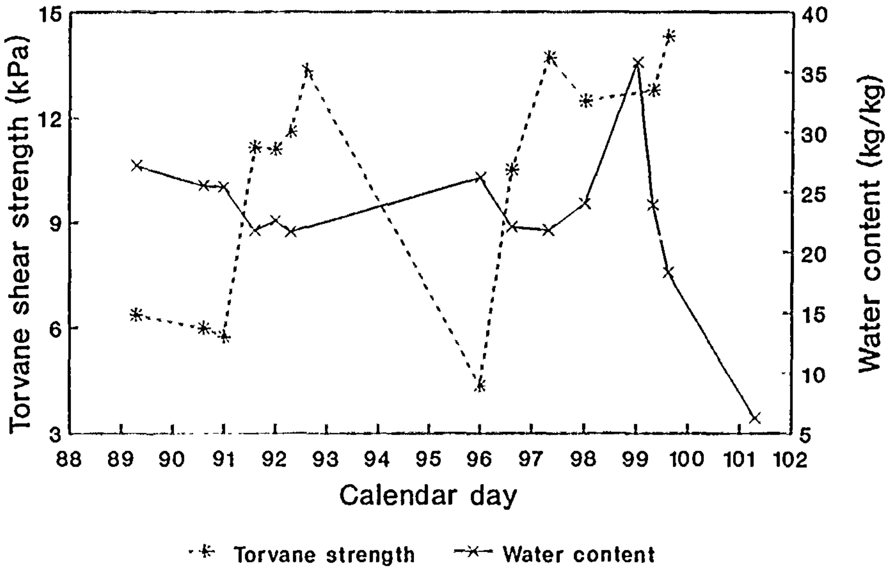

Figure 1. Torvane and water content observations for early 1989

Note the major variations in soil strength and soil water content around freeze-thaw events, shown in Figure 1, especially during diurnal freeze cycles. Similar soil water content fluctuations were observed by Pikul and Allmaras (1984). Soil strength upon thawing would often be 50 percent or less of the pre-freeze strength. Sunny conditions caused rapid thaw of the top several millimeters of the soil profile while the underlaying layers were still frozen. The super-saturated top layer would flow down the hill indicating a near absence of soil strength. Under sunny and windy conditions the soil temperatures could reach up to 8^{o}C while the air temperature remained around 0^{o}C. Under these conditions , the soil would regain its pre-freeze strength within a couple of hours as the soil water content decreased rapidly, indicating an appreciable effect of evaporation. Under overcast conditions it could take several days for the soil to regain its pre-freeze strength.

No clear seasonal trends in soil strength were found. A slight consolidation and strengthening of the soil after fall tillage and a slight strengthening trend in the spring as the soil dried were observed. Soil water content was as high as 58 percent under frozen soil conditions, 44 percent while thawing, and 25 percent during rain or snowmelt. A weak negative linear relationship between soil strength and soil water content of the top ten millimeter of the soil profile was found (Equation 1).

$$\tau = a + b*M_{10} \quad (1)$$

Where τ = Torvane soil shear strength (kPa)
a,b = fitted coefficients
M_{10} = water content of the top 10 millimeter of the soil profile.

Observations were made in four distinct periods that were separated by periods that did not allow for measurements due to frozen conditions in the winter and dry soil conditions in the summer. Coefficients as obtained for the four periods and for the total period of the experiment are shown in Table 1. The wide range in values for the coefficients is currently under study.

Table 1. Regressions between Torvane soil shear strength and soil moisture content.

Period	r^2	Relationship	Number of observations
Late 1987		No data	
Early 1988	0.81	$\tau = 32.16^{**} - 84.07^{**} M_{10}$	8
Late 1988	0.59	$\tau = 22.53^{**} - 60.60^{**} M_{10}$	13
Early 1989	0.59	$\tau = 25.20^{**} - 62.50^{**} M_{10}$	14
Total period	0.51	$\tau = 22.95^{**} - 0.57^{**} M_{10}$	35

** = Significant at the 1 percent level.

LABORATORY RESEARCH

Laboratory experiments on the effect of freeze-thaw cycles on soil strength of a Palouse silt loam have been conducted by means of radiative freezing (Formanek 1983; Formanek et al., 1984). Undisturbed soil samples were frozen with a thermo-electric plate at minus 4°C for 20 to 30 hours to a depth of 50 mm. The experiment was conducted under soil water tensions of 0.49, 1.47 and 4.41 kPa; during fall-cone tests, the soil water tension was set to 0 kPa. Measurements with a fall-cone before freezing and after thawing indicated a 50 percent reduction in average strength over all tensions, from 0.89 to 0.44 kPa, after one freeze-thaw cycle. Subsequent freeze-thaw cycles did not further reduce the average soil strength. The soil was allowed to consolidate overnight at a pre-set tension. Fall-cone strength of the consolidated soil was found to be a function of the logarithm of the soil water tension during consolidation.

$$\tau_s = a + b \ln \psi_m \quad (2)$$

where τ_s = fall-cone shear strength (kPa)
a,b = fitted coefficients for the soil
ψ_m = soil water tension during consolidation (kPa)

Coefficients are shown in Table 2. This relationship shows the extreme weakness of the soil under very wet conditions and explains the high erodibility, and erosion rates, observed in the field.

Table 2. Cohesion of a silt loam at 0 kPa tension after freeze-thaw cycles and consolidation.

freeze-thaw cycle	ψ_m kPa	a	b	r^2	Number of observations
1	0.05-4.41	1.11	0.22	0.64	36
2	0.05-4.41	1.46	0.36	0.85	30
3	0.05-8.82	1.64	0.42	0.82	68

In order to determine the influence of freeze-thaw on rill detachment, a tilting flume apparatus was built (Van Klaveren, 1987, Van Klaveren and McCool, 1987). A large soil sample (2.74m long by 0.45m wide and 0.60m deep) was frozen while maintaining a constant water table at 50, 150 or 450 millimeter (0.5, 1.5, and 4.4 kPa respectively) below the slightly V-shaped soil surface. The soil sample was frozen by circulating a cooled brine solution (minus 12°C to minus 22°C) through a plate suspended over the soil surface. Freezing to a depth of 120 mm was achieved in 7 to 14 days. The soil was allowed to thaw in 5 to 14 days. Fall-cone strength measurements were obtained on the thawed, consolidated soil and a relationship similar to the one determined by Formanek (1983) was found (Figure 2). Values for a and b were determined at 1.6 and 1.1 respectively with a coefficient of determination of 0.64.

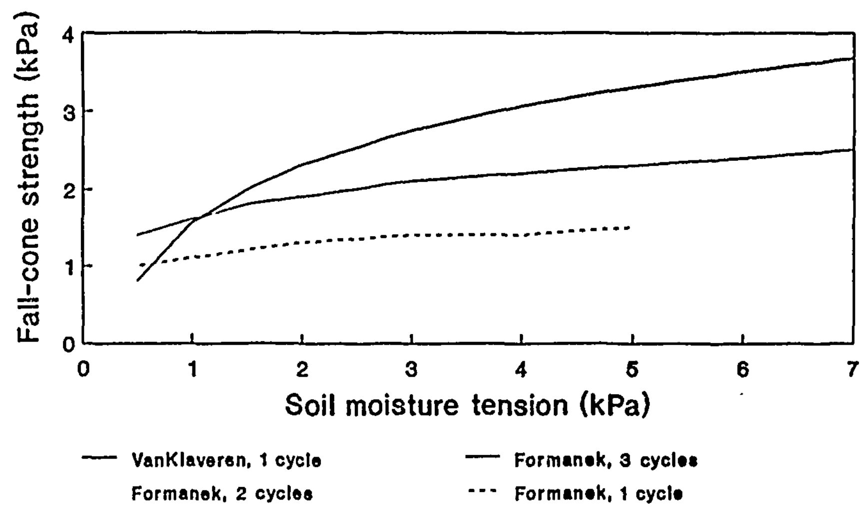

Figure 2. Fall-cone strength after consolidation as a function of soil moisture tension and number of freeze-thaw cycles.

The flume was tilted to a 6.75 percent slope, and surface flow was introduced to the thawed soil surface, causing a rill to develop. Parameters for the relationship between flow shear stress and soil detachment (Equation 3) were determined for the three water table settings. Critical shear strength values were extrapolated from the data (Figure 3).

$$D_{cr} = a\,(\tau-\tau_{cr})^{b} \qquad (3)$$

where D_{cr} = detachment capacity of the rill flow (g m^{-2} min^{-1})
τ = flow shear stress (Pa)
τ_{cr} = critical shear stress at which erosion begins in a channel
a,b = fitted coefficients.

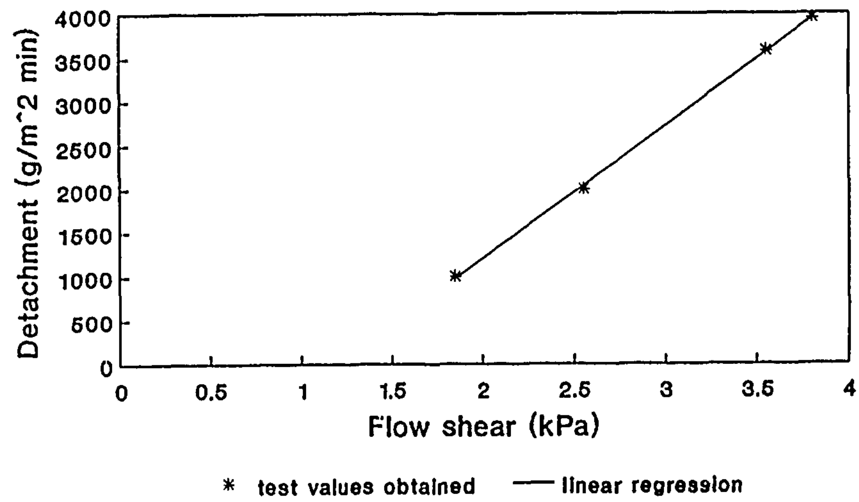

Figure 3. Detachment versus flow shear stress; extrapolation of critical shear.

The coefficient b was found to be near 1, indicating a straight line relationship. This coincides with experiments by Van Liew (1983), who found a value of 0.95, with preformed rills in a Palouse silt loam with slopes of 7 to 24 percent under field conditions. The coefficient a represents the soil erodibility. Erodibility of the soil decreased during initial rill formation, but increased for the 4.4 kPa tension. After about 30 minutes the rill stabilized, but for the 0.5 kPa tension it took an hour for the critical shear strength to stabilize. (Figure 4).

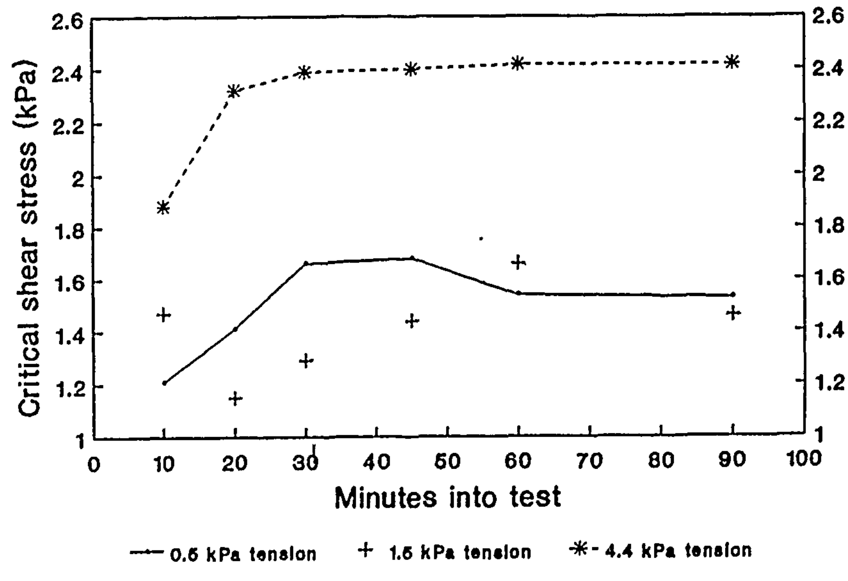

Figure 4. Critical shear stress versus time.

Erodibility followed similar patterns for the 4.4 kPa tension, but showed reversed trends for the 0.5 and 1.5 kPa tensions (Figure 5).

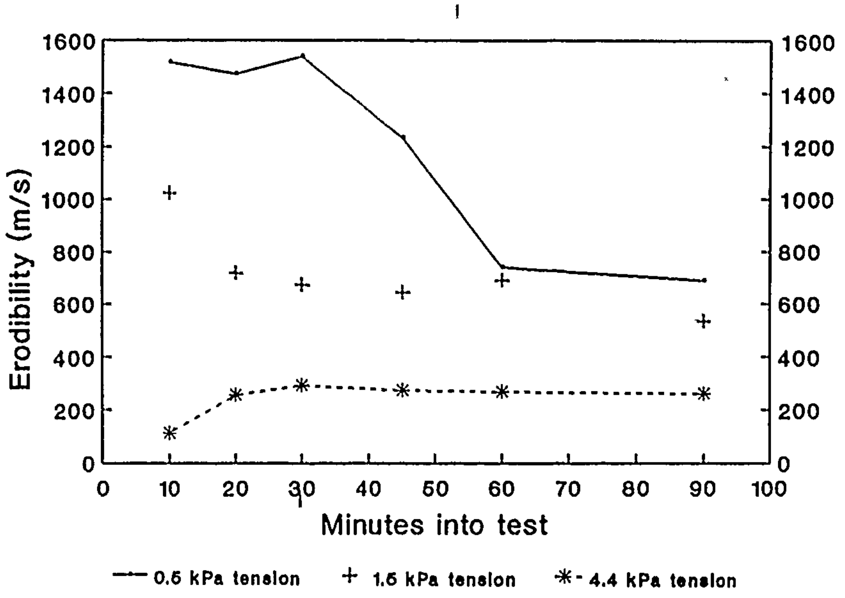

Figure 5. Erodibility versus time.

SUMMARY

Field measurements show extremely low shear strength during soil thaw. Laboratory measurements show a change of strength with applied soil moisture tension after thaw, similar to what actually happens in the field. Erodibility was found to be a function of soil moisture tension. Little change in critical shear stress was found with changing soil moisture tension.

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SOIL FREEZING AND THAWING SIMULATION WITH THE SHAW MODEL

by

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INTRODUCTION

Frozen soil plays a significant role in the hydrology and management of watersheds in northern climates. Soil freezing can inhibit infiltration, increase the potential for flooding and erosion, and delay soil warming and spring planting. Accurately predicting the occurrence and depth of soil freezing and subsequent thawing enhances our ability to evaluate management options for soil and water conservation.

Many methods ranging from simplified, semi-empirical approaches to very detailed, physically-based models of coupled heat and water flow have been developed for predicting frost depth in soils. The Stefan formula and the Berggren equation (Lunardini, 1981) represent some of the earliest models for frost-depth prediction. Benoit and Mostaghimi (1985) developed a simple frost depth model which utilizes mean daily air temperature to calculate heat flux from the snow or soil surface to the frost front. Harlan (1973) was one of the first to formulate a mathematical model for the coupled heat and water transport in partially frozen soils. Harlan's model, like others to follow (Guymon and Luthin, 1974; Jame and Norum, 1980; Pikul et al., 1989; and others), was based on the assumption that the soil characteristic curve and water movement in partially frozen soil are analagous to that in unsaturated soil. Gilpin (1980), and O'Neill and Miller (1985) extended the coupled heat and water flow models to include frost heave. Cary (1987) presented a method for including solute effects on frost heave. While some of these models are very complex, most require surface temperature or surface heat flux for input rather than standard weather data, and none include the effects of snow, residue, and tillage on heat and water flow.

The occurrence, depth and permeability of frozen soil results from the interactions between climate, soil, antecedent water content, slope, cover and management practices. The Simultaneous Heat And Water (SHAW) model (Flerchinger and Saxton, 1989a) uses hourly weather data to simulate the interrelated heat, water, and solute transfer through snow, residue and soil for a wide range of conditions. Because the model uses fundamental equations for heat, water, and solute flux, input parameters are readily definable for different soil, residue, topographic and climatic conditions. This paper illustrates the model's applicability to widely varying site and climatic conditions. Field data and model results are presented for. shallow, transient freeze-thaw cycles which are prevalent in the Pacific Northwest grainlands, elevational effects in the mountainous rangelands in southwestern Idaho; and deep, soil-frost conditions in interior Alaska.

MODEL DESCRIPTION

The physical system described by the SHAW model consists of a one-dimensional vertical profile extending from the snow, residue or soil surface to a specified depth within the soil. The model is sufficiently flexible to represent a broad range of conditions, although transpiring plant canopies are currently not included. The system, illustrated in Figure 1 may or may not have snow, a residue layer or a tillage layer. Hourly predictions include evaporation, soil frost depth, snow depth, runoff and soil profiles of temperature, water, ice and solutes.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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Input to the SHAW model includes initial soil temperature and water content profiles, hourly weather conditions, general site description, and parameters for the snow, residue and soil. General site information includes slope, aspect, latitude, and surface roughness parameters. Parameters necessary for simulating snow include coefficients for the functional relation between snow depth and percent snowcover (Flerchinger and Saxton, 1989a). Residue or litter properties include residue loading, thickness of the residue layer, percent cover and albedo. Input soil parameters for the model are bulk density, saturated conductivity, and coefficients for the matric potential--water content relation, albedo--water content relation, and De Vries' thermal conductivity relation (De Vries, 1963).

Hourly weather conditions above the upper boundary and soil conditions at the lower boundary define the heat and water fluxes to the system. Water and heat flux at the surface boundary include absorbed solar radiation, long-wave radiation exchange and turbulent transfer of heat and vapor. Absorbed solar radiation is dependent on surface albedo, slope, aspect, snow depth, and residue cover, long-wave radiation exchange depends on cloud cover (estimated from solar radiation) and surface temperature, and turbulent transfer is a function of wind speed, surface roughness and atmospheric stability.

A layered system is established through the snow, residue and soil, and each layer is represented by a node (Fig. 1). After computing the flux at the upper boundary, the interrelated heat, liquid water, and vapor fluxes between layers are determined. The heat and water flux equations for the system are solved simultaneously using finite difference equations to balance the net flux into each layer with changing conditions within the layer. The implicit finite-difference equations are solved iteratively using a Newton-Raphson procedure. A more detailed description of the model is given by Flerchinger and Saxton (1989a).

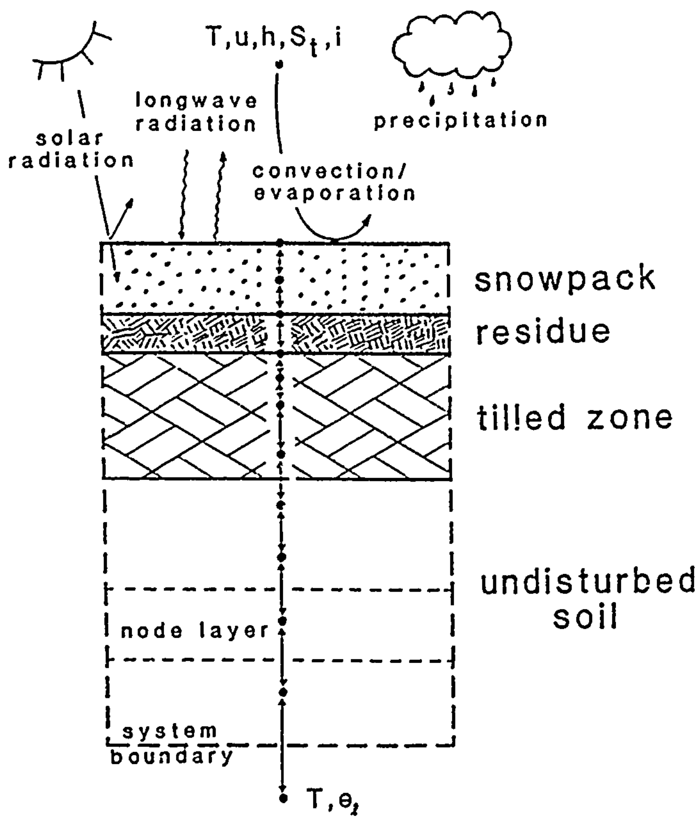

Figure 1. Physical description of system represented by SHAW model (T,u,h,S_t, and i represent air temperature, windspeed, humidity, solar radiation and precipitation, respectively).

Site characteristics

Data were collected from two plots at each of three sites representing widely varying climatic regions. Sites included: the USDA Palouse Conservation Field Station near Pullman, Washington; Reynolds Creek Experimental Watershed located in southwestern Idaho; and Alaska Agriculture Experimental Field Station near Delta Junction in interior Alaska. Specific plot data are presented in Table 1.

Table 1. General site description.

	Pullman, WA		Reynolds Creek, ID		Delta Junction, AK	
Plot	No-till	Convent'l	Quonset	Reynolds Mtn	No-till	Double-disk
Slope (%)	14.6	17.0	0.0	0.0	2.0	2.0
Aspect	S	S	-	-	NW	NW
Residue	wheat	wheat	grass/moss	grass/moss	barley	barley
weight(kg/ha)	10415	0	1250	1250	8500	2000
cover (%)	91	26	40a	40a	90a	40a
layer (cm)	3.0	0.0	0.5	0.5	4.0	0.7
Surface soil						
sand (%)	9	9	52	38	30	30
silt (%)	65	65	39	47	63	63
clay (%)	26	26	9	15	7	7
density(gm/cm^3)	1.36	0.95	1.58	1.39	0.82	0.86

a - Estimated values.

Climate in the Palouse region is characterized by hot dry summers and cool wet winters. The soil typically goes through three to four freeze-thaw cycles annually excluding diurnal cycles (Zuzel, 1986) and seldom freezes to depths greater than 40 cm. Precipitation at the study site averages 540 mm annually, most of which occurs during the winter months. Plots at the Pullman site included a heavy-residue no-till and a light-residue tilled plot, both on a Palouse silt loam having been annually cropped with winter wheat. Residue was applied to the plots in the fall prior to tillage. The tilled plot was plowed and disked to produce a clean-tilled, finely pulverized seedbed, and both plots were planted to winter wheat.

The Reynolds Creek Watershed is a mountainous rangeland site, predominantly vegetated with sagebrush and rangeland grasses. Precipitation, snowcover and frost depth on the watershed are greatly dependent on elevation. The Quonset and Reynolds Mountain sites at elevations of 1193 and 2097 m, respectively, were selected for simulation. Mean annual precipitation at these sites is 284 and 796 mm, respectively. Snowcover is often sufficient at Reynolds Mountain to prohibit soil frost penetration, while the Quonset site often freezes to soil depths greater than 60 cm. Soils at these sites are a gravelly loam at Reynolds Mountain and a sandy loam near the Quonset site. Sagebrush canopy at each site was very sparse and offered negligible resistance to heat transfer.

The Delta Junction site is a forested area cleared for barley production in the interior of Alaska. Mean annual air temperature is -2 °C and mean annual precipitation is 300 mm. Forested areas immediately surrounding the site have permafrost and thaw to a few meters over the summer. The site typically experiences two freeze-thaw cycles annually: one cycle which freezes on the top 2 to 5 cm and the main event having rapid frost penetration which freezes the soil below two meters in depth. Soil profile at the site consists of a 10-cm low-density silt loam soil high in organic matter (4%), underlain by a much denser sandy loam. The plots had been continuously cropped for spring barley since

1980. Post-harvest plot treatments consisted of double-disked and no-till tillage treatments, representing a maximum disturbance, minimum residue condition and a minimum disturbance, maximum residue condition.

Field data

Hourly weather data collected at the Pullman and Reynolds Creek sites included air temperature, humidity, windspeed, precipitation and solar radiation. Soil temperature profiles to a depth of 168 cm on a three-hour basis were taken at the Pullman site, and to a depth of 120 cm on an hourly basis at the Quonset site. Average daily soil temperature profiles to a depth of 90 cm were collected at Delta Junction, and soil temperatures were measured weekly at Reynolds Mountain.

Hourly weather data for Delta Junction were generated from daily weather observations, which included maximum and minimum air temperature, precipitation and solar radiation. Hourly temperatures were extrapolated by fitting a sinusoidal curve through the observed temperatures, assuming that the minimum temperature occurred at 1/2 hour before sunrise, and the maximum temperature occurred midway between noon and sunset. Hourly solar radiation was generated from total daily solar radiation assuming that the atmospheric transmissivity was constant for the day. Hourly values were calculated from (Campbell, 1977)

$$S = (S_t/S^h_o)\, S_o \sin\alpha \qquad (1)$$

where S_t is measured daily solar radiation, S^h_o is total daily solar radiation incident on a horizontal surface at the outer edge of the earth's atmosphere, S_t/S^h_o is the atmospheric transmissivity for the day, S_o is the solar constant (1360 W/m^2) and α is the sun's altitude angle above the horizon for the hour. Wind at the site was assumed at a constant 3m/s. Humidity for each day was assumed constant and equal to the saturated vapor density at the minimum air temperature.

Soil water data were not available for Delta Junction and were estimated at 25 percent by volume near the surface, and 20 percent at depths beyond 15 cm. Soil water content profiles at the other sites were measured weekly. Frost depth was measured using cylindrical gypsum moisture blocks read every three hours for the Pullman plots, and frost tubes filled with a flourescein solution and sand (Harris, 1970) were used at Reynolds Creek. Frost tubes filled with a methylene-blue dye solution were used at Delta Junction, but appeared to unsatisfactorily lag frost penetration when compared with soil temperature data. The 0 °C isotherm was therefore used to indicate frozen soil depth. Gypsum blocks, sand-filled frost tubes, and 0 °C isotherm indicate approximately the same frost depth under moist conditions (Hanson and Flerchinger, 1990).

Values for soil albedo (0.25), residue albedo (0.40) and snow depth-percent cover parameters for the Pullman plots were based on calibration results of Flerchinger and Saxton (1989b). Surface roughness was determined from measured wind profiles and soil thermal and hydraulic properties were measured by Kenny and Saxton (1988).

The model was run without prior calibration for the Reynolds Creek and Delta Junction sites. Soil albedo for these sites was calculated from water content, θ, using the equation

$$A = 0.30e^{-2.4\theta} \qquad (2)$$

based on data presented by Idso et al. (1975). Soil hydraulic properties were measured for the Delta Junction plots and estimated from soil texture (Saxton et al., 1986) for the Reynolds Creek sites. Residue albedo, surface roughness and snow depth--percent cover parameters for the Delta Junction plots were assumed approximately equal to the Pullman plots. Residue albedo for the Reynolds Creek sites was estimated at 0.30, surface roughness was taken as 0.13 of the height of surrounding vegetation (Campbell, 1977), and the minimum depth for 100% snowcover was estimated as half the height of surrounding vegetation. (This was the depth at which surrounding vegetation was estimated to significantly influence snowmelt). Because of the large sand and silt fractions, thermal conductivity of soil minerals for the Reynolds Creek and Delta Junction sites was assumed to be that of quartz (8.8 W/m/C).

MODEL SIMULATIONS

Data from the winter of 1986-87 were used at all sites for simulation with the SHAW model. Soil freezing was simulated from November 4 to February 23 (day 308 to 54) for the transient freeze-thaw cycles on the Pullman plots, from September 7 to May 30 (day 250 to 150) for the Reynolds Creek sites, and from October 15 to November 17 (day 288 to 321) for the rapid frost penetration on the Delta Junction plots. Daily average air temperature for these sites is plotted in Figure 2.

Simulated and measured hourly temperatures were compared using a coefficient of efficiency, E, calculated as (Kitanidis and Bras, 1980)

$$E = 1 - \frac{\Sigma(T(i) - \hat{T}(i))^2}{\Sigma(T(i) - \overline{T})^2} \qquad (3)$$

where $\hat{T}(i)$ are simulated temperatures, $T(i)$ are corresponding measured temperatures, $\overline{T}$ is the average measured temperature for the simulation period, and E (≤ 1) is the fraction of variation in measured values explained by the model. Values of E less than zero indicate that the average measured value for the period is a better estimate than simulated values. The coefficient of efficiency for simulated soil temperature of all plots is given in Table 2.

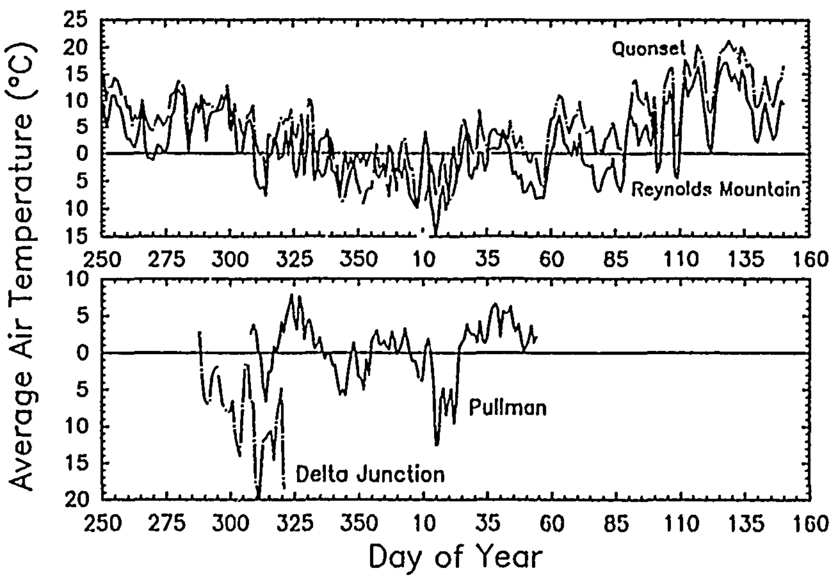

Figure 2. Average daily air temperature for simulation periods.

Table 2. Coefficient of efficiency for simulated soil temperatures.

					Pullman, WA						
Depth (cm)	0	8	15	25	38	53	69	84	107	138	168
No-till[1]	0.70	0.78	0.84	--	0.87	--	0.89	0.85	0.91	0.96	1.00
Conventional[1]	0.70	0.79	0.80	0.83	0.84	0.76	0.74	0.78	0.88	0.96	1.00
					Reynolds Creek, ID						
Depth (cm)	2.5	5	10	20	30	45	60	75	90	120	
Quonset[2]	0.92	0.92	0.94	0.95	0.95	0.95	0.94	0.98	0.99	1.00	
Reynolds Mtn.[3]	--	--	0.27	--	0.95	--	0.97	--	1.00		
					Delta Junction, AK						
Depth (cm)	0	5	10	15	22	30	45	60	75	90	
No-till[4]	0.56	0.64	0.63	0.67	0.62	0.84	0.89	0.93	0.98	1.00	
Double-disk[4]	0.69	0.80	0.78	0.76	0.76	0.76	0.78	0.81	0.88	1.00	

[1] Soil temperatures measured every three hours.
[2] Soil temperatures measured hourly.
[3] Soil temperatures measured weekly.
[4] Average daily soil temperatures.

Soil temperatures were simulated somewhat better for the Pullman and Reynolds Creek sites than the Delta Junction site where only daily weather data were available. Average absolute error in simulated near-surface temperature ranged from 0.82 °C for the no-till Pullman plot to 1.66 °C for the double-disk Delta Junction plot. Errors in simulated temperature generally decreased with depth. The coefficient of efficiency for simulated surface temperature was 0.70 for both plots at the Pullman site, and generally exceeded 0.80 for most other depths. Efficiency of simulated soil temperature exceeded 0.92 for all depths at the Reynolds Creek Quonset site. The coefficient of efficiency at the Delta Junction site was 0.56 and 0.69 for average daily surface soil temperature of the no-till and double-disked plot, respectively. Efficiency at deeper depths exceeded 0.62 and 0.76, respectively for the two plots. Measured windspeed, soil water content and hourly weather observations would undoubtedly improve model predictions at the Delta Junction site.

Simulated and measured frost depth plotted in Figures 3 and 4 for the Pullman plots illustrate the model's applicability to simulate transient snowcover and freeze-thaw cycles under widely varying tillage and residue treatments. Maximum frost depth indicated by moisture blocks was 3 cm for the residue-covered no-till plot and 10 cm for the bare, tilled plot. Maximum simulated frost depths were 5 and 12 cm, respectively.

Maximum frost depth and the rate of freezing and thawing were simulated quite well at the Reynolds Creek Quonset and Reynolds Mountain sites as illustrated in Figures 5 and 6. Snow depth at Reynolds Mountain was simulated reasonably well except where the model underestimated the large snow accumulation between days 65 to 75 due to problems in distinguishing rain from snow at near-freezing temperatures and drifting of snow. Precipitation for the simulation period totaled 127 mm at the Quonset site and 455 mm at Reynolds Mountain. As a result, the Quonset site had little or no snowcover and the soil froze much deeper than at Reynolds Mountain. Maximum measured frost depth was 58 cm at the Quonset and 31 cm at Reynolds Mountain. Simulated frost depths of 58 and 34 cm, respectively, compare very well with measured values.

Rapid frost penetration at the Delta Junction site was simulated well using extrapolated hourly data from daily weather observations. Simulated frost depth and depth of the 0 °C isotherm determined from measured soil temperatures are plotted in Figure 7 for the no-till and double-disked plots. Night-time air temperatures of -15 °C and colder caused frost to penetrate the double-disk plot more than 90 cm in less than 25 days. Results from this site illustrate that reasonably good predictions can be obtained using daily weather observations.

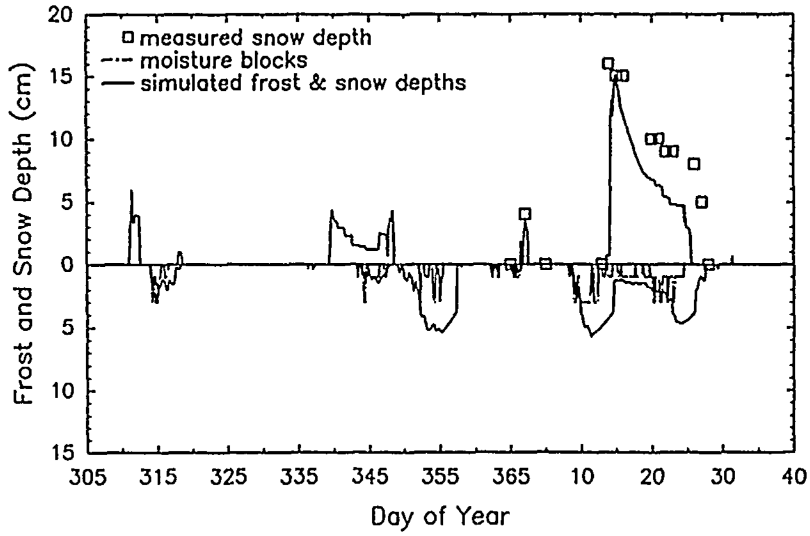

Figure 3. Snow and frost depth for a heavy residue, no-till plot at Pullman, Washington, USA.

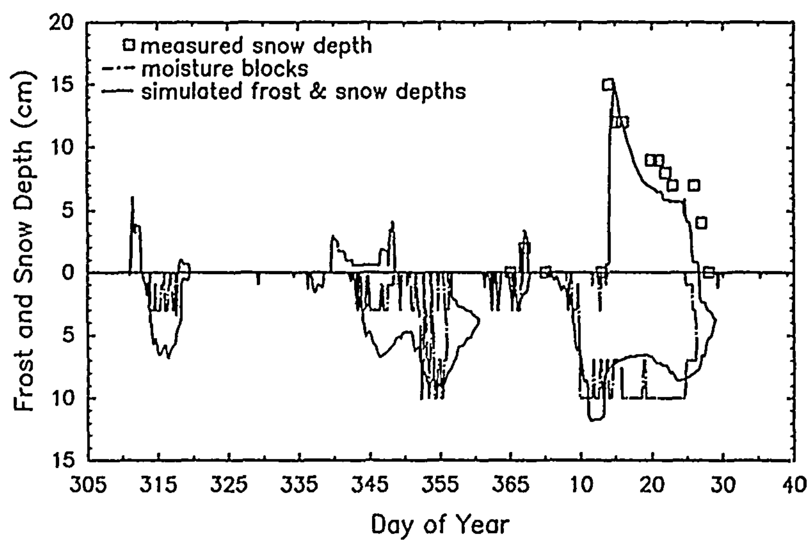

Figure 4. Snow and frost depth for a bare-surface, tilled plot at Pullman, Washington, USA.

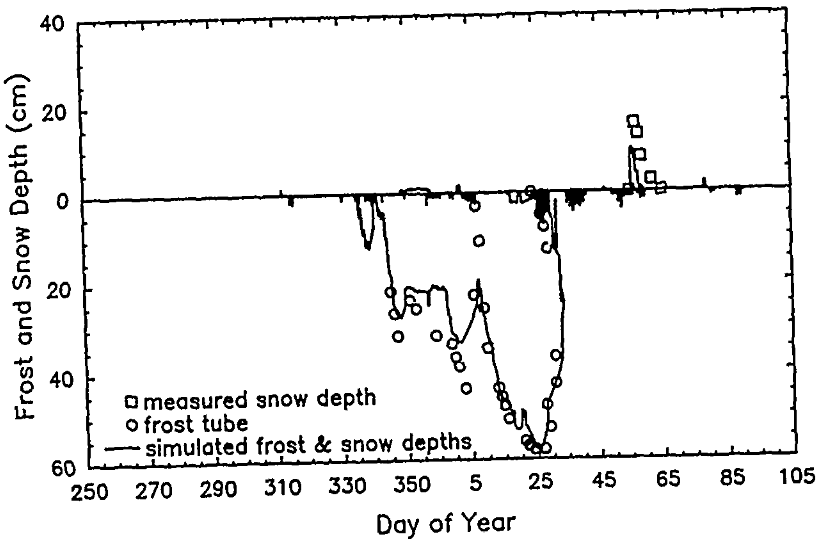

Figure 5. Snow and frost depth for the Quonset site on the Reynolds Creek Watershed, Idaho, USA.

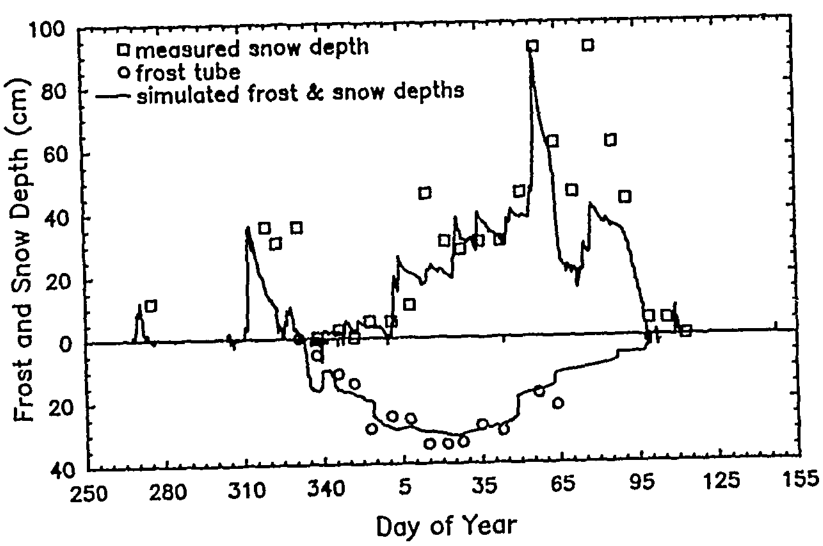

Figure 6. Snow and frost depth for Reynolds Mountain on Reynolds Creek Watershed, Idaho, USA.

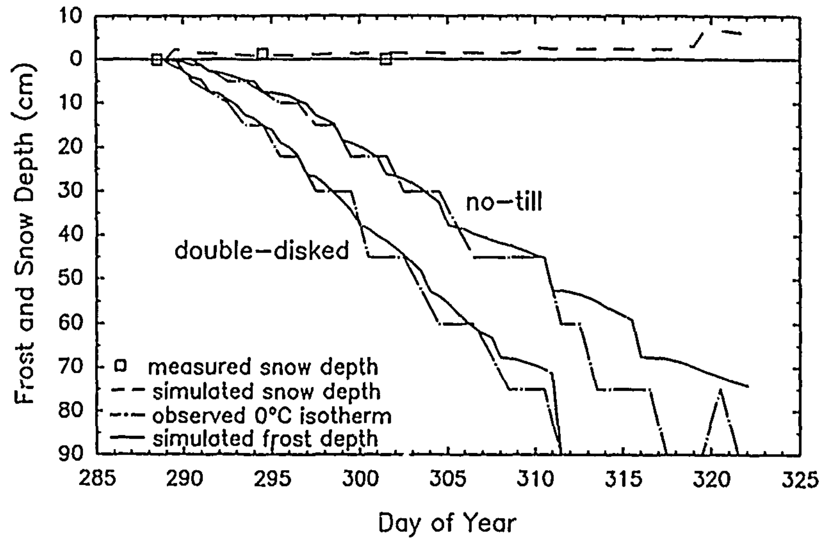

Figure 7. Snow and frost depth for double disked and no-till plot at Delta Junction, Alaska, USA.

SUMMARY AND CONCLUSIONS

The Simultaneous Heat and Water (SHAW) model is a physically based model of the interrelated heat, water and solute transfer through snow, residue and soil and may be easily parameterized for diverse site conditions. The SHAW model was applied to two diverse plot conditions at each of three sites representing widely varying climatic conditions. Effects of tillage and residue management were simulated for shallow, transient freeze-thaw cycles of Pullman, Washington and for deep, rapid frost penetration near Delta Junction, Alaska. Effects of elevation and snowcover in mountainous rangeland were simulated on the Reynolds Creek Experimental Watershed in southwestern Idaho. Tilled plots with light residue cover froze faster and deeper than heavy-residue no-till plots at the Pullman and Delta Junction sites. High-elevation sites with adequate snowcover on the Reynolds Creek Watershed had relatively little frost compared to low-elevation sites.

Results obtained from the three sites illustate the versatility of the SHAW model to simulate heat and water transfer for widely varying tillage, residue, topographic and climatic conditions. Maximum frost depth was simulated within 3 cm for all plots at the Pullman and Reynolds Creek sites. (Maximum frost depth at the Delta Junction site exceeded monitoring depth.) Soil temperature, snow depth and rate of freezing and thawing were successfully simulated for all sites. Application of the SHAW model to the Delta Junction site, where only daily weather observations were available, supports the use of the model even where data is limited.

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SIMULATING THE FREEZING AND THAWING OF ARABLE LAND IN SWEDEN

by

Lars-Christer Lundin[1]

INTRODUCTION

A number of models treating coupled heat and water flows in soils have been presented (*e.g.*, Harlan, 1973; Guymon & Luthin, 1974). The objective with the development of these models was mainly to study frost actions such as heave and ice lens formation. In many cases the applications have been directed towards construction of roads and pipelines, often in permafrost areas. Since there is very little permafrost in Sweden and no pipeline construction in the areas where the permafrost is located, the research on frost processes has here been concentrated on road problems. The theoretical problems in this area were already treated in detail by Beskow (1935) and later research has been very practically oriented.

Frost also influences agriculture in a country with a cold climate. The influence on the paths of the meltwater during spring is very strong. There is also evidence that the hydraulic properties of some soils are influenceded by the frost action throughout spring, several weeks after the thawing of the soil. Such effects are important to consider when studying leakage of nutrients or erosion from arable land because these problems are closely related to the transport of water through or on top of the soil.

The problems cannot be solved in the same manner on arable land as is traditionally done in engineering applications. The soil material present cannot be replaced with less frost susceptible material. This means that the problems often have to be accepted as being properties of the system. Thus, the study of frost action among agronomists tends to be descriptive and focused on measurements more than modeling. Dismissing frost because there is none during the growing season is to simplify the problem to much.

In Sweden, frost modeling started when Jansson & Halldin (1979) developed a coupled heat and water flow model. The first simulated frost depths were presented by Halldin et al. (1979) and Jansson & Halldin (1979). However, these works were not related to arable land. Further developments of the frost description in the model were initiated by Lundin (1984). Simultaneously the model was adapted to agricultural soils (Jansson & Thoms-Hjärpe, 1986). Recently, winter applications of the model to agricultural soils have been presented by Jansson & Gustafson (1987), Johnsson & Lundin, (in press), Lundin (in press) and Thunholm et al. (1989). Engelmark (1986) develops a similar model to study infiltration and runoff but is mainly interested in urban applications.

The objective of this paper is to discuss the problems related to an introduction of frost processes in agricultural hydrology modeling and to point at what can be gained by such an introduction. How the introduction can be done in practice will be demonstrated on a physically based model of the atmosphere-plant-soil-system. The model chosen was the SOIL model (Jansson & Halldin, 1979), being widely used in Sweden (*e.g.*, Grip *et al.*, 1979; Jansson, 1980; 1986; 1987; Jansson & Thoms-Hjärpe, 1986; Dressie, 1987; Gustafson, 1987a; b; Johansson, 1987a; b; Persson & Jansson, 1988) to study the hydrology of both forests and arable land. The aspects of the model that needed to be developed to improve the winter hydrology treatment included the hydraulic conductivity functions, the temporal changes in the soil moisture characteristic curve, the infiltration of melt water through the frozen soil and the discrimination of the runoff into drainage and surface runoff. Results from Johnson & Lundin (in press) and Lundin (in press) will be used as a base for this discussion. Implications to the problems of erosion and nutrient leakage in connection with snow melt and frost will also be discussed.

Presented at the Frozen Soil Symposium, March 21-22, 1990; Spokane, Washington

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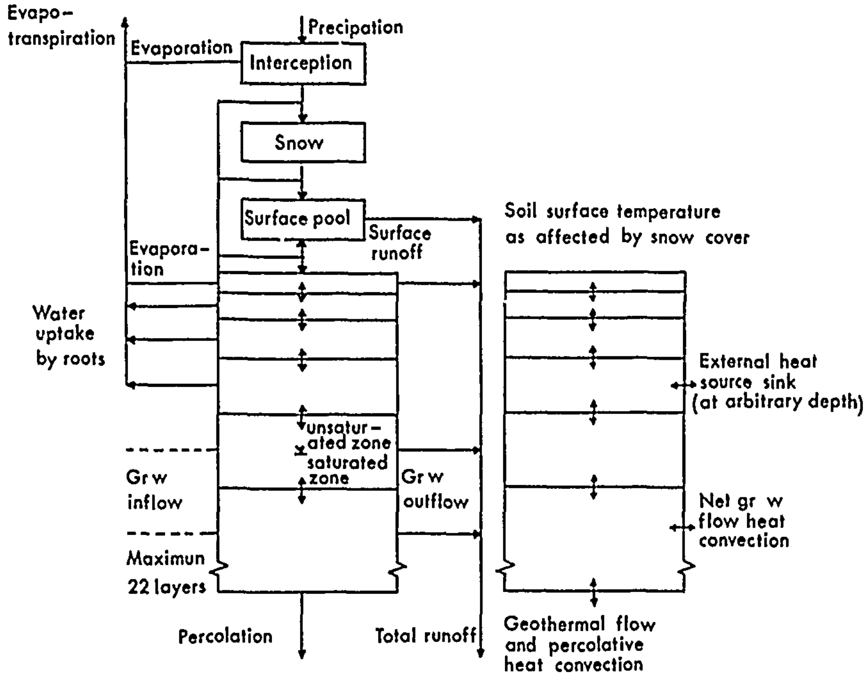

Figure 1. The structure of the SOIL model. (Modified from Jansson & Halldin, 1979 and Jansson & Gustafson, 1987.)

THE SOIL MODEL

The SOIL model (Figure 1) was designed to be operationally simple, *i.e.*, flexible with respect to the handling of different boundary conditions, rather than numerically or physically sophisticated. The numerical integration is based on an explicit finite difference scheme.

A brief description of the model is given below. The technical description by Jansson & Halldin (1980) gives more details. Mass transfer is described by the Darcy equation and the mass balance equation:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta}{\delta z}\left(k_{\mathrm{w}}\left(\frac{\delta p_{\mathrm{w}}}{\delta z} + \frac{\mathrm{g}}{\mathrm{V_w}}\right)\right) + s_{\mathrm{w}}, \qquad (1)$$

where θ is the water content, t is time, z is depth, k_w is the hydraulic conductivity, p_w is the hydraulic potential, g is the gravitational constant, V_w is the specific volume of water, and s_w is a sink term. The hydraulic conductivity, k_w, varies with water content according to the function proposed by Mualem (1976). This function is modified at water contents close to saturation to allow for an increased conductivity due to the influence of macro-pores. This feature is important in cultivated clay soils which commonly show a high conductivity close to saturation.

The heat flow equation, accounting for the latent heat of phase change, is formulated with the heat content, E, as the state variable:

$$\frac{\delta E}{\delta t} - \frac{\mathrm{L}}{\mathrm{V_i}}\frac{\delta\theta_{\mathrm{i}}}{\delta t} = \frac{\delta}{\delta z}\left(k_{\mathrm{h}}\frac{\delta T}{\delta z}\right) - \mathrm{C_w}\frac{\delta(Tq)}{\delta z}, \qquad (2)$$

where E is the heat content, L is the latent heat of thawing, V_i is the specific volume of ice, θ_i is the ice content, k_h is the thermal conductivity, C_w is the heat capacity of water, q is the flow of water, and T is the temperature. The terms to the left describe the change of sensible and latent heat. The first term to the right describes the divergence of the heat flow, whereas the second term accounts for convective flow. The thermal conductivity is in mineral soils assessed from Kersten's (1949) equations. Model compartment sizes are normally chosen to increase with depth.

In this study the upper compartments were 5-10 cm thick. State variables and parameters are interpolated linearly between the model compartments.

The two flow equations are coupled and account for the phase change of water during freezing and thawing. This is a central problem in models of frozen soils. An empirical relation between the latent energy content and the temperature is used in the SOIL model. Sensible and latent heat contents from which ice and water contents are calculated, follow uniquely from a known temperature.

Frost heave is treated in a rudimentary way. The soil heaves if the total volume of ice and unfrozen water exceeds the porosity of the soil in a layer.

The SOIL model has a special routine to handle a snow pack as the upper boundary condition. The accumulation and melt functions are driven by precipitation, global radiation, air temperature, and the soil surface heat flow. Liquid water exceeding a given retention threshold infiltrates into the soil. The thermal conductivity of the snow is calculated from the snow density (Corps of Engineers, 1956).

Two alternative formulations are introduced into the model for the hydraulic conductivity at frozen conditions (Lundin, in press). The first aims at improving the numerical performance of the model. The linear interpolation of the hydraulic conductivity at the boundary between a frozen compartment and an unfrozen one is replaced by a procedure in which the hydraulic conductivity is put to the minimum value of the frozen compartment and the linearly interpolated boundary value.

The second approach is an impedence factor similar to the ones of Jame & Norum (1980) and Guymon *et al.* (1983; 1984). The hydraulic conductivity of frozen soil, k_f, is formulated as:

$$k_f = 10^{-E_i \alpha} k_w, \tag{3}$$

where α is the thermal quality, *i.e.*, the mass ratio of frozen water to the total amount of water, and E_i is an impedence parameter.

Measurements and basic parameterizations of the model

Measurements were carried out on a heavy clay profile in Uppsala (Lat. 59° 49') and on two layered loam-sand profiles at Kjettslinge, 40 km north of Uppsala. The measurements are also used by Johnsson & Lundin (in press) and Lundin (1989; in press) in their analyses.

The clay loam soil had a loamy fine sand horizon extending from approximately 20 to 30-70 cm depth underlain by a heavy clay (Steen *et al.*, 1984). Measurements were made on two plots. One plot was cleared from snow once a week and the other was undisturbed. The distance between the profiles was approximately 15 m. Measurements of temperature and total water content were carried out from 1 November 1984 to 6 May 1985. The temperatures in the top layers were measured every three hours, using Pt-100 sensors. Deeper down, temperature readings were taken weekly using thermistors. The measurements of total water content were made with a Wallingford IH II neutron probe (*e.g.*, Morgenschweis & Luft, 1983). The winter of 1984/85 was cold with air temperatures occasionally below -30 °C. A minor frost cycle started in mid November and lasted for two weeks. The major frost cycle started in early December and lasted throughout the winter.

Field measurements of precipitation and soil temperature at the depth of 5 cm were used as driving data together with Penman evaporation, estimated by Alvenäs *et al.* (1986) from climatic data from Kjettslinge, Marsta, 20 km south of Kjettslinge, and occasionally Uppsala. The soil parameters (Figure 2) were taken from Steen *et al.* (1984).

Measurements of profiles of unfrozen water content, using Time Domain Reflectometry (TDR), were conducted in a plot during 1 December 1982 - 5 July 1983. The measurements are reported by Vikström (1984) and Rydén (1986). The winter of 1982/83 was a warm winter for Swedish conditions with an air temperature seldom below -10 °C. Snow started to fall in mid December, but a lasting snow pack was not established until late

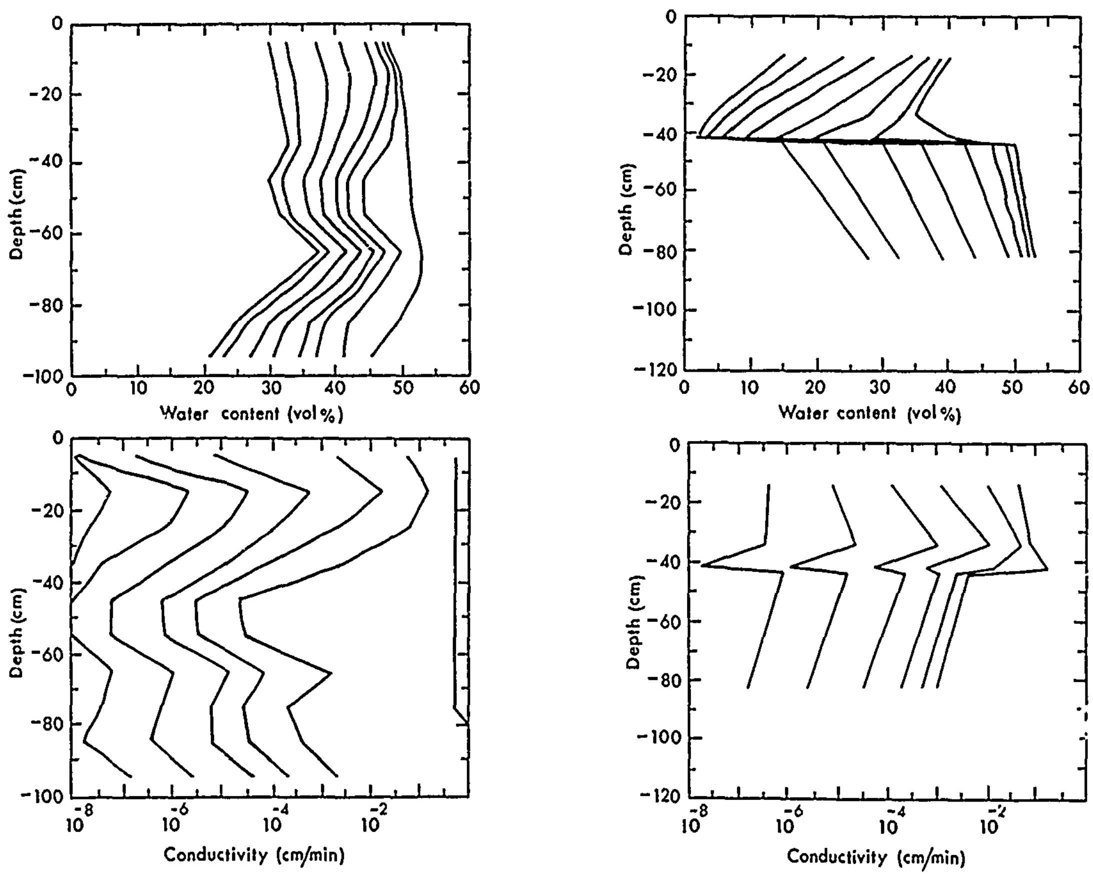

Figure 2. Water content and hydraulic conductivity as a function of depth for the matric potentials 0, 20, 50, 100, 1,000, 5,000, and 15,000 cm (from right to left) for the heavy clay profile, left, and for the loam profile, right. Data from Sandsborg & Wiklert (1976) and Steen *et al.* (1984).

January. The weather was unstable in December and January; a warm winter climate interrupted by occasional cold spells. The maximum snow depth, observed in early March, was less than 20 cm.

Measurements of climatic data from a nearby bioclimatic station were used as driving variables to the model The set of driving data consisted of daily values of air temperature, precipitation, relative humidity, wind speed, net radiation, and global radiation.

Soil moisture retention data from Sandsborg & Wiklert (1976) were fitted to the analytical function given by Brooks & Corey (1964).

In the clay soil, Lundin (in press) assumed that the ice lenses contributed to an increase in the occurrence of pore sizes in the range lower than 0.1 mm. Before thawing, the soil moisture characteristic curve was changed in order to decrease the water content at tensions below pF 1. The new set of soil moisture characteristic curves changed the unsaturated conductivity as well since it is coupled to the matric potential.

Lundin (in press) compares data from the heavy clay profile to

a) simulations using the original conductivity concept with linear interpolation and

b) simulations using the modified interpolation. In this case, a different set of soil moisture characteristic curves was given to the frost affected layers on 1 April, *i.e.*, just before thawing.

The sensitivity to the different conductivity formulations is also tested using data from the layered loam-sand profiles (Lundin, in press). Model experiments are carried out using

a) the original model formulations with a linear interpolation procedure,

b) the alternative interpolation procedure, and

c) the alternative conductivity formulation with values of 4 and 6 for the impedence parameter E_i.

Experiments with the combined use of the alternative interpolation and the impedence parameter were also carried out.

The simulated results were spatially interpolated to be comparable with measured values of temperature and water content. Water contents are expressed as % by volume throughout the paper.

Validation of the model

The loam profile with a sand layer

The measured temperatures for the loam profile with a sand layer were well described by the model (Figure 3).

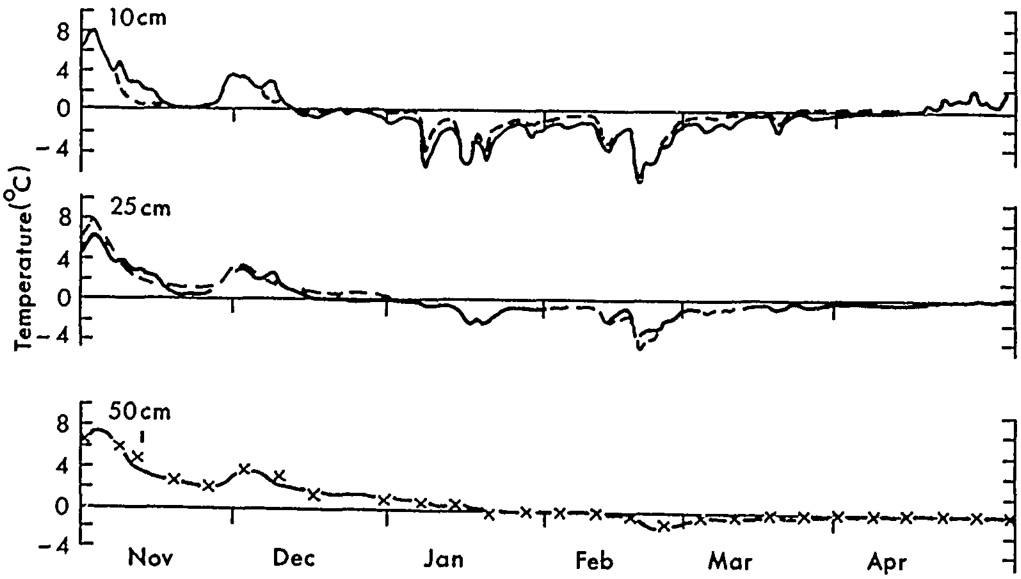

Figure 3. Measured (broken lines and crosses) and simulated (solid line) soil temperature at three depths in the loam profile from 1 November 1984 to 6 May 1985.

The measured total water content (Figure 4) showed an increase in the top layers when freezing started (Lundin, in press). This increase was accompanied by a decrease in deeper layers, indicating a freezing induced redistribution. A marked increase in total water content also occurred in mid March. The simulation did not describe this; an underestimation, observed at all depths, occurred. Ten days after the measured increase, an increase occurred in the simulated total water content at the 10 cm depth.

The sensitivity test made by Lundin (in press) for the snow cleared ground shows that the alternative interpolation procedure gave the best fit at the 10 cm depth (Figure 4). At a depth of 25 cm the alternative

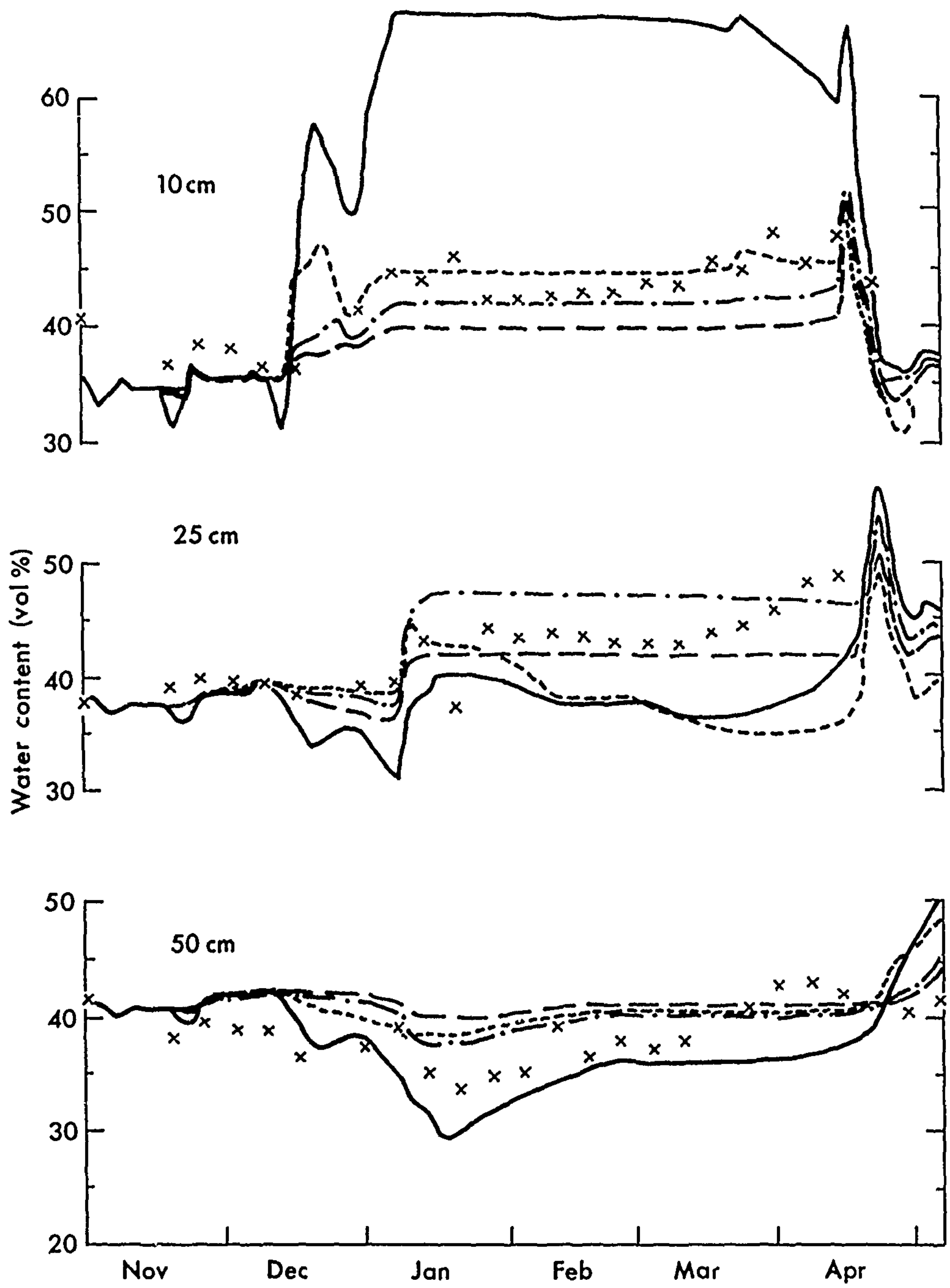

Figure 4. Measured (crosses) and simulated total water content for three depths in the loam profile from 1 November 1984 to 6 May 1985. The different simulations were made with a) the original hydraulic conductivity formulations, *i.e.*, E_i=0 (solid line), b) the modified interpolation of the hydraulic conductivity (dotted line), and the values c) E_i=4 (dot-dashed line), and d) E_i=6 (broken line). (From Lundin, in press.)

interpolation procedure did not give a very good fit; an overestimation of 20 % occurred at this depth in January. The decreasing water contents, seen at the 25 cm depth, created by this interpolation procedure were not found in the measurements. Percolation was taking place below the frozen zone in the model when the alternative interpolation procedure was used. This feature was seen to a small extent in the original approach but not in the measurements.

Using the original description of the hydraulic conductivity of frozen soil resulted in an overestimation of 10 % at the 10 cm depth and 15 % at the 25 cm depth was seen in January. An underestimation occurred at the 50 cm depth.

The greatest sensitivity was seen at a depth of 25 cm. A change in value of the impedence parameter from 4 to 6 (Figure 4) created a 20 % difference at this depth. Maximum differences ranged from 4 to 20 %, dependent on depth. A value of 6 for the impedence parameter was found to be the optimal if the linear interpolation procedure was used. Using the alternative interpolation procedure allowed the use of an impedence parameter of 2.

Johnson & Lundin (in press) could use the same parameters for the snow covered plot when they studied runoff. During the snowmelt period (March-April) almost all simulated total runoff from the field consisted of surface runoff (Figure 5). The limited infiltration capacity was caused by frost in the soil. The simulated tile drainage did not start until the beginning of May, *i.e.*, after thawing of the frost and after the snowmelt period. However, measurements showed that the drainage flow started 2-4 weeks earlier, in early April.

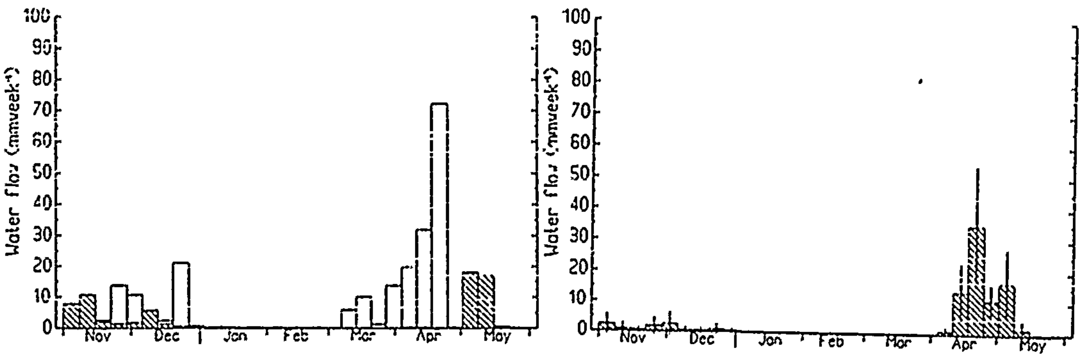

Figure 5. Simulated total runoff and tile drainage (left) and observed tile drainage (right) from the loam soil field at Kjettslinge. The mean of four observations, with the standard deviation as thin bars, is shown in the right subpicture. (From Johnsson & Lundin, in press.)

The heavy clay profile

The dynamics of the unfrozen water content (Figure 6) was depicted quite well during winter in the simulation of the heavy clay profile made by Lundin (in press). When the alternative interpolation procedure was applied no impedence correction was necessary. However, after thawing, the simulated and measured water contents did not correspond in the simulation with static soil properties. The discrepancy was especially marked in the upper layers. The water content was overestimated by 15 % in late May at the 5 cm depth. The discrepancy was noticeable to a depth of more than 50 cm and decreased successively from May, at the top layers, to finally disappear in early July. The correspondence between measured and simulated water content was good when the soil physical properties were modified upon thawing of the soil.

DISCUSSION

Freezing induced redistribution of water

The modeling of frost processes in the field is associated with a number of difficulties. One is the description of the upper boundary, including processes such as snow accumulation and melting, infiltration into the thawing soil, and the assessment of the surface temperature of the ground. Others are the coupling of water and heat flow

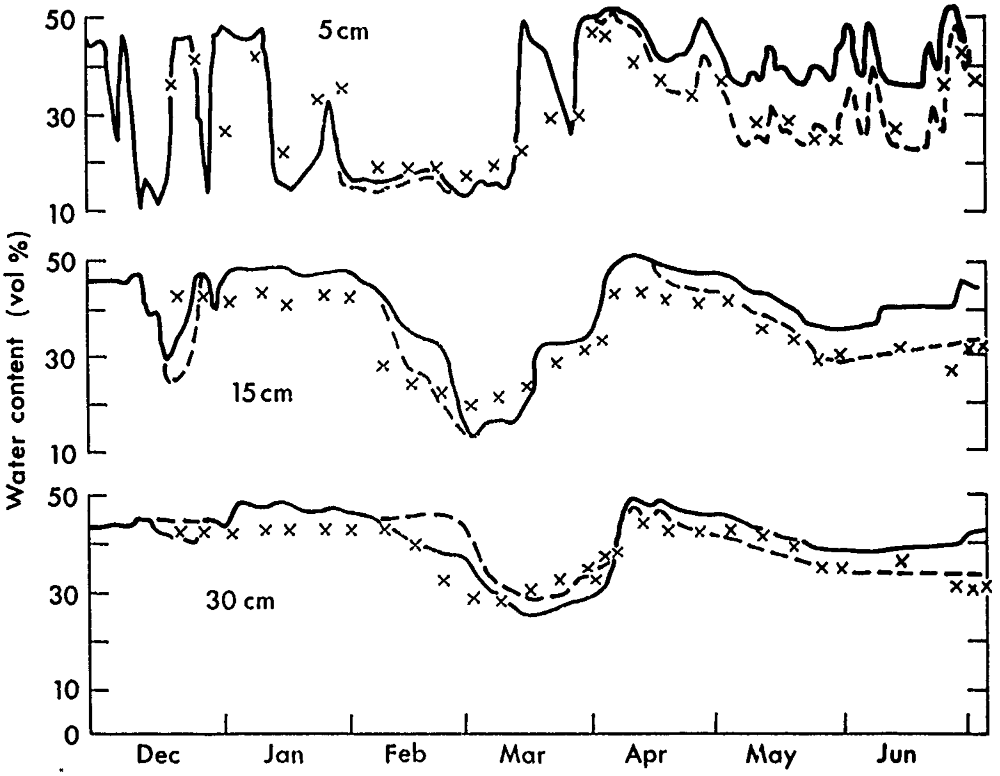

Figure 6. Measured (crosses) and simulated (solid and broken lines) unfrozen water contents for three depths in the heavy clay soil from 1 December to 5 July 1983. The soild line represents a simulation with the unaltered model. The broken line represents a simulation with the modified interpolation procedure and, from 1 April, the modified soil moisture characteristic curves. (From Lundin, in press.)

through the freezing point depression function, the formulation of a hydraulicconductivity function for frozen soil, the effects of frost on soil structure, and the spatial and temporal variability of parameters and variables.

Freezing of fine-grained soils is accompanied by the development of ice lenses. This is not explicitly considered by the model. Instead, a reduction of the hydraulic conductivity of the frozen soil is done by the introduction of an impedence parameter. This procedure is applied in the model studies by Taylor & Luthin (1976), Jame & Norum (1980) and Guymon *et al.* (1983) as well. No measurements of the impedence parameter exist; it is a calibrated parameter. Thus, the impedence parameter will include all numerical and other errors present in the simulation, making it difficult to compare the impedence parameter between models and perhaps even between different fields. Lundin (in press) reviews parameter values for various models and various applications and gives a range from 2.2 to 10.

From the sensitivity tests done by Lundin (in press) and Johnsson & Lundin (in press) it seems that the impedence parameter is not dependent on the thermal regime. They could use the same value on the undisturbed plot as on the snow cleared one. This means that a single value for the parameter can be used for a series of years. When the alternative interpolation procedure was used in the SOIL model, the impedence parameter could be decreased form 6 to 2 for the loam soil.

In the SOIL model it is assumed that the soil-moisture characteristic curve is valid for frozen soils as well. However, for a coarse, non-colloidal soil a correction for the difference in surface tension between the ice-water interface and the air-water interface has to be applied. The correction factor was reported to be 2.2 by Koopmans & Miller (1966), implying that, at a given water content, the pressure difference in a frozen soil is about half the difference found in an unfrozen soil. In soils containing a mixture of non-colloidal and colloidal particles the factor would be somewhere between unity and 2.2, possibly varying with water content.

For the SOIL model, these findings means that water contents obtained from the freezing point depression curve could give a matric potential that is more than twice its actual value. The error will be smaller in the gradient of matric potential than in the potential itself since the error is systematical. The error will affect the hydraulic conductivity estimations as well. The resulting flows will not be affected, however, because the errors are corrected by the impedence parameter.

It could be argued that the surface tension correction factor is to be preferred from a theoretical point of view because it is has a more physical base and because the possibilities to measure the fraction of non-colloidal particles in a soil may be greater than the possibilities to measure the ice lens impedence. However, no attempt to measure the surface tension correction factor in a field soil sample is known to the author. In practice, the problem of estimating the two parameters are probably comparable.

Frost induced structural changes

The difference in water content between simulation and measurements seen in the upper layers during spring at the heavy clay site at Uppsala is related, by Lundin (in press), to changes in the soil structure, influencing the soil moisture characteristics. He intuitively altered the soil moisture characteristic curve, increasing the pore size distribution index by a factor of five, decreasing the air entry value to half, and decreasing the wilting point by 5 %. A less intuitively model should establish the time dependent relation between the soil moisture characteristic curve and the thermal quality of the soil, so that changes are related to the degree of freezing. Different behavior should be expected for different soil types. Changes in porosity because of frost heave may also occur. As the soil thaws it consolidates and the retention characteristics gradually return to their original values, making the process fully reversible in seasonally frozen soils. Series of soil moisture characteristic curves from winter and spring are needed to certify the effects of frost on the spring drainage properties of the soil.

The spring drainage properties are important not only because they influence the drying up of the soil but also because they affect the runoff volumes during and after snow melt. Improved drainage during spring leads to runoff volumes in excess of the water contained in the snow pack and can also lead to increased leakage of nutrients.

Drainage - surface runoff

Jansson & Gustafson (1987) tried to simulate the discharge from drainage pipes on an agricultural field in Röbäcksdalen in northern Sweden. They observed a consistent delay in simulated discharge in connection with snow melt for the studied period (1977-1981). In order to solve the observed problem they added a surface pool to the model, containing melt water that was not able to infiltrate immediately. Water in the surface pool was directed either as surface runoff or infiltrated at a later time step. The model development contributed to a more correct partitioning between surface runoff and tile drainage but did not improve the timing of the simulated discharge in the drainage pipes.

Thunholm *et al.* (1989) performed infiltration experiments on a thawing heavy clay soil and tried to model their results. Problems occurred in that it was not possible to simulate infiltration unless the soil was fully thawed, although infiltration capacities were considerable in the experiment.

Johnsson & Lundin (in press) experienced a similar problem in their loam soil simulations at Kjettslinge. The measured pipe discharge started two weeks before the soil profile was thawed, while the simulated pipe discharge commenced in connection with the complete thawing of the soil.

Several possible causes to the discrepancies have been suggested, *i.e.*, macropore flow, spatial variability of the field, frost influence of the water flow in the model not accounted for, or an overestimation of the soil water deficit at the beginning of infiltration. Surely, the first two of these propositions have influence on some fields. However, a more plausible general explanation is presented by Thunholm *et al.* (1989) and further developed by Johnsson

& Lundin (in press). They lean towards the findings that the spring infiltration capacity is related to the ice content of the soil, reported by, *e.g.*, Kane (1980) and Granger *et al.* (1984).

Most soils are unsaturated when they freeze in field conditions. The result is that the finest pores are filled with unfrozen water, the intermediate pores are filled with ice, and the large pores are filled with air. If no thawing occurs during winter this situation will be conserved until spring. Snow melt water can then infiltrate in the large air-filled pores and the infiltration capacity is not related to water content in the way assumed in the SOIL model. Instead it is related to the volume of voids.

A very interesting aspect of the system is its irreversibility with respect to infiltration capacity, *i.e*, the infiltration capacity can only decrease during winter. For each melt event followed by refreezing, such as the diurnal cycle of freezing and thawing, a portion of the infiltrating water is frozen, jamming the large pores. The infiltration of water contributes to the decrease of the infiltration capacity. To restore the fall infiltration capacity a complete thawing of the ground or infiltration of water with a temperature considerably above 0 °C is needed.

Johnson & Lundin (in press) presents an algorithm and states that the infiltration process as described could be simulated with the SOIL model provided a discrimination between bound water in the fine pores and free, infiltrating, water is done. However, they do not present any results using the new procedure. Further work is needed to this end.

Erosion and nutrient leakage

A correct description of the pathways of the melt water is imperative when studying problems related to erosion and leakage of nutrients. The transport processes are quite different depending on whether the melt water contributes to surface runoff or to infiltration and pipe discharge.

Surface runoff is a major cause of erosion and leakage of phosphorus form arable land (*e.g.*, Alström & Bergman, 1988). There is currently a limited knowledge of the extent of the erosion problem in Sweden. However, recent investigations indicate that serious problems exist, at least locally. Alström & Bergman (1988) report phosphorous losses of up to 3.63 kg/ha from October1986 to April 1987 on a field in southern Sweden. The nitrogen loss caused by surface runoff for the same period was 11 kg/ha.

Often the nitrogen losses are related to nitrate dissolved in the soil water. This nitrate is located in the unfrozen water in the fine pores during winter. Rather high concentrations develop here. If infiltrating water can pass through the air-filled large pores, as described earlier, the exchange between the nitrate-rich, bound, water and the infiltrating water will be very limited. On the other hand, if no infiltration is possible before the soil is completely thawed, the exchange with the nitrate-rich water will be well developed. In addition to this, the water in the fine pores will participate in the percolation as soon as the soil is thawed.

Thus, a cold, uninterrupted, winter leads to a fairly rapid infiltration of melt water through large pores and very limited surface runoff. No problems of erosion or leakage are to be expected. A warm winter, interrupted by melt events, leads to clogging of the large pores with ice, limited infiltration capacity, and surface runoff. Severe problems of erosion and leakage could be expected.

It must be concluded that the ability to model erosion and leakage of nutrients during spring is intimately coupled to the frost treatment in the model used.

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MODELING THE EFFECTS OF SOIL FROST AND SNOWMELT ON RUNOFF AND EROSION

by

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INTRODUCTION

Successful watershed planning efforts require accurate assessment and predictions of the movement of sediment and chemicals throughout a catchment. This can be accomplished with sediment yield and/or water quality models. However, providing accurate estimates of the average annual runoff and erosion occurring on a watershed as a result of a representative series of weather events requires consideration of the effects of winter processes on erosion and runoff in addition to the effects of land management during the growing season. This paper describes a procedure to model overwinter processes on a watershed so that snowmelt runoff and changes in soil erodibility brought about by freezing and thawing can be predicted.

BACKGROUND

Freezing modifies the physical characteristics of a soil, changing its ability to transmit or retain water (Benoit and Bornstein, 1970; Benoit and Mostaghimi, 1985; Campbell et al., 1970; Loch and Kay, 1978), its structural stability (Benoit, 1973; Mostaghimi et al., 1988), and its erodibility (Bisal and Nielsen, 1967). The development of soil frost is the result of complex interactions of several primary factors, including soil characteristics, type of tillage and residue management, surface roughness, type of vegetative cover, duration and extent of freezing temperatures, and the extent and timing of snow cover. The freezing process itself modifies those soil physical properties that, along with temperature, determine the depth and duration of soil frost. The magnitude of soil changes that take place as a result of soil freezing depends on temperature, soil water content at freezing, initial size of soil aggregates, initial soil density, and the number of freeze-thaw cycles that take place over winter. As a result, tillage-residue management combined with over winter frost action determines a soil's erodibility during winter thaw periods and from spring snowmelt to planting (Benoit et al., 1986).

The snowmelt-frozen soil procedure described here is composed of three separate components which interact with each other on a daily basis. These components deal with soil frost, snowmelt, and snowdrift. The frost component estimates the extent of frost development and thawing over the winter period as well as changes in soil water content and infiltration capacity. The snowmelt component estimates the amount of snowmelt occurring and how much snowmelt water is available for runoff in the spring. The snowdrift component estimates the depth, density, and distribution of snow cover over a watershed. Prediction of the effects of soil frost and snowmelt on runoff and soil erosion at any point within a watershed is achieved by dividing the watershed into sections and using a continuous simulation modeling process in which the three components of the snowmelt-frozen soil process interact on a daily basis.

SOIL FROST

The soil frost component is based on simple heat flow theory. It assumes that heat flow in a frozen or unfrozen soil or soil-snow system is unidirectional and that the average 24-hour temperature of the system surface-air interface is approximated by average daily air temperature. The routine predicts daily frost and thaw development for various combinations of snow, residue and tilled and/or untilled soil at any point and is driven only by daily inputs of maximum and minimum air temperature and snow depth. Snow and soil thermal conductivity and water flow components are considered as constants. The routine

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Presented at Frozen Soil Symposium, March 21-22, 1990, Spokane, Washington.

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yields values for daily frost depth, thaw depth, number of freeze-thaw cycles, water accumulated in frozen soil, and infiltration capacity of the tilled layer or top 20 cm of soil if the soil is untilled.

The soil frost routine operates by a daily bookkeeping process that compares calories of heat lost or gained at the soil surface to heat flow from deeper unfrozen soil layers. Net calories of heat lost or gained are converted to centimeters of frozen or thawed soil. Unidirectional heat flow through the frozen soil or soil-residue-snow system is calculated from the relation:

$$Q_{srf} = K_{srf} T_{srf} / Z_{srf} \qquad [1]$$

where Q_{srf} is the heat flux through the snow-residue-frozen soil system (W/m²), K_{srf} is the average thermal conductivity through the combined snow-residue-frozen soil depth thickness (W/m-°C), T_{srf} is the temperature difference the snow-residue-frozen soil thickness (°C), and Z_{srf} is the depth or thickness of the combined snow-residue-frozen soil layer (m).

Thus, heat flow through the snow-residue-frozen soil layer is the product of an average thermal conductivity for the layer and an average temperature gradient, with the gradient being the difference between average daily air temperature and the zero degree isotherm at the bottom of the frozen soil.

The basic assumption is made that the average temperature of the soil (snow)-air interface over a 24-hour period is equal to the average air temperature for the same period. The validity of this assumption varies with location as a function of items such as emissivity, radiation, cloud cover, and wind. For this reason, the average daily surface temperature that drives the frost routine is computed by a surface energy balance routine that modifies average daily air temperature by a local accounting of wind speed, solar radiation, cloud cover, and atmospheric emissivity (Flerchinger, 1987).

The average thermal conductivity for a layered system can be shown to equal the harmonic mean for the layers in the system and is given by:

$$K_{srf} = Z_{srf} / \sum_{i=1}^{N} (Z_i / K_i) \qquad [2]$$

where Z_i is the thickness of each layer (m), K_i is the thermal conductivity of each layer (W/m-°C), N is the number of layers, and

$$Z_{srf} = \sum_{i=1}^{N} Z_i.$$

The soil frost routine is designed to handle a system with up to four layers--snow, residue, tilled soil, and untilled soil. In this case the average thermal conductivity equation becomes:

$$\begin{aligned} K_{srf} = {} & (K_{snow} * K_{res} * K_{ftill} * K_{futil}) * (S_{nowd} + R_{esd} + T_{illd} + U_{tilld}) / \\ & (K_{snow} * K_{res} * K_{ftill} * U_{tilld} + K_{snow} * K_{res} * K_{futil} * T_{illd} + \\ & K_{snow} * K_{ftill} * K_{futil} * R_{esd} + K_{res} * K_{ftill} * K_{futil} * S_{nowd}) \end{aligned} \qquad [3]$$

where K_{snow} is the thermal conductivity of snow (W/m-°C), K_{res} is the thermal conductivity of residue (W/m-°C), K_{ftill} is the thermal conductivity of frozen tilled soil (W/m-°C), K_{futil} is the thermal conductivity of frozen untilled soil (W/m-°C), S_{nowd} is the snow depth (m), R_{esd} is the residue thickness (m), T_{illd} is the frozen tilled soil depth (m), and U_{tilld} is the frozen untilled soil depth (m). With this approach, if any or all of the snow, residue or tilled depths are zero, the thermal conductivity reduces to the harmonic mean of the remaining layers.

Over any 24-hour period, Q_{srf} must be balanced by heat flow (Q_{uf}) from the unfrozen soil below the frozen layer. The frost routine defines Q_{uf} as the sum of heat transferred by the thermal conductivity properties of the soil matrix, the latent heat of fusion in freezing transferred water, and losses in heat content of the soil. That is:

$$Q_{uf} = K_{uf}(T_{uf}/Z_{uf}) + LK_w(P_{wp}/Z_{uf}) + C_{uf}dT_{uf}Z_c \qquad [4]$$

where Q_{uf} is the heat flow from unfrozen soil (W/m^2), K_{uf} is the thermal conductivity of unfrozen soil (W/m-°C), T_{uf} is the change in temperature from 0 degree isotherm to depth of stable temperature (°C), Z_{uf} is the depth of unfrozen soil to point of stable temperature (m), L is the latent heat of fusion (W-s/m^3), K_w is the unsaturated hydraulic conductivity of soil (m/s), P_{wp} is the change in total water potential (m), C_{uf} is the heat capacity of the unfrozen soil (W/m^3-°C), dT_{uf} is the change in temperature of a unit volume of soil in unit time (°C), and Z_c is the depth of unfrozen soil that supplies heat as a result of changes in soil temperature (m) (assume a constant value of 1 m).

In this equation, the soil temperature and water potential gradients are those that exist just below the 0 degree isotherm. As a practical convenience, the model assumes that heat flow through soil thermal conductivity and soil water movement are separate and discrete units of heat transfer.

The routine operates on each section of the watershed by iteratively balancing over each 24-hour period the heat lost through the snow-residue-frozen soil zone with heat flow through the unfrozen soil to the freezing front. Iteration is on an hourly basis for each 24-hour period. During the balancing process, it is assumed that heat lost through the frozen zone is first balanced by heat flow in the unfrozen soil as a result of the soils temperature gradient and thermal conductivity. Additional heat loss is balanced by the heat of fusion released by freezing water that is held in place or migrates to the freezing front. Further heat loss is balanced by changes in soil heat content of the unfrozen soil, the magnitude of which is computed by difference.

SNOWMELT

The snowmelt routine is based on a modification of a generalized basin snowmelt equation for melt in open areas developed by the U. S. Army Corp of Engineers (1956, 1960). This equation was modified by Hendrick, et al. (1971) to adapt it for use with readily available meteorological and environmental data. We further modified the Hendrick equation to make it compatible for use with the other two components of the winter processes modeling procedure.

The equation used in the snowmelt component in its modified form is:

$$M = [0.0606R(1-F)-0.84(1-N)(1-F)+0.0268v_2(1-0.8F)(0.18T_x+1.404T_d) + (T_x+T_m)(0.0225+0.248P)][0.0245] \qquad [5]$$

where M is snowmelt (m), R is estimated radiation on a sloping surface (MJ/m^2), F is forest cover (dec %), T_x is daily maximum temperature (°C), N is estimated cloud cover (dec %), v_2 is mean daily wind speed measured at a height of 2 m (m/s), T_d is mean daily dewpoint temperature (°C), T_m is daily minimum temperature (°C), and P is mean daily precipitation (m).

Since some snowmelt can occur in direct solar radiation to about 3°C below freezing (Hendrick et al., 1971), the first term in equation [5] is multiplied by the quantity $(0.36T_x + 1)$ whenever $-3°C \le T_x < 0°C$. The values for T_x, T_m, v_2, T_d, and P can be obtained either from weather records or from a climate generator subroutine. The amount of cloud cover, N, is estimated from the relationship:

$$N = (1-R_m/R_c)/0.7 \qquad [6]$$

where R_m is the mean measured daily solar radiation (MJ/m^2) and R_c is the potential clear sky radiation on a horizontal surface (MJ/m^2).

This equation is based on the fact that clouds reflect approximately 70% of solar radiation and transmit only 30% to the earth's surface (Sutton, 1953). Both R and R_c can be calculated in a separate subroutine based on slope inclination (I) in radians, slope facing direction (A) in degrees from north, calendar day (J), measured radiation (R_m) in MJ/m^2, a solar constant (S_c) (equal to 0.081 MJ/m^2), and latitude (L_{at}) in degrees (Swift and Luxmoore, 1973). This subroutine takes into account cloud cover effects and atmospheric transmissivity. Slope inclination, I, is calculated as:

$$I = \tan^{-1} (S/100) \qquad [7]$$

where S is land slope (%).

To run the snowmelt routine, the average values of F, A, and S are input for any section, L is input for the entire watershed, S_c is a constant value and J is generated within the program.

Equation [5] deals with four major energy components of the snowmelt process--temperature, radiation, vapor transfer, and precipitation. When calculating snowmelt, the following assumptions are made: any precipitation that occurs on a day when maximum daily temperature is <0°C is assumed to be snowfall, no snowmelt will occur if maximum daily temperature is <-3°C, the snowpack will not melt until snowpack density is ≥50 kg/m^3, the temperature of the surface soil-snow interface = 0°C during the melt period, and cloud base temperature approximately equals surface air temperature. The albedo of melting snow is approximately 0.5 (Sutton, 1953), and maximum daily temperature is approximately 2.2 times the mean daily temperature (Hendrick et al., 1971). Using equation [5], if the calculated value of snowmelt, M, is less than 0, then M = 0. If it is greater than the existing snow depth, D, from the preceding day, then M = D.

SNOWDRIFT

The snowdrift routine determines the distribution of snow over the watershed by estimating the depth of snow on the ground at the end of a day in any section, depending on the weather that day and the topography. Calculations are based on several initial assumptions:

- -the density of fresh new-fallen snow ≈ 100 kg/m^3,
- -the density of a ripe snowpack must be ≥ 350 kg/m^3 before it begins to melt,
- -the threshold wind velocity for moving falling snow ≈ 0.89 m/s measured at a height of 2 m,
- -the surface roughness of a uniform snow pack ≈ 0.0002 m, and
- -the snow storage capacity of a tilled layer = the random roughness.

The amount of snow trapped and stored in a section by standing vegetation is the storage capacity, S_t, and is a function of the height and the projected stem area, or basal density, of the vegetation, the surface roughness and the amount of standing biomass. S_t is calculated as:

$$S_t = \epsilon H(d_b)R_t/R_o + z_o \qquad [8]$$

where S_t is the storage capacity of snow (m), ϵ is a trapping efficiency (%), H is the height of standing vegetation (m), d_b is the basal density of standing vegetation (m/m), R_t is the standing residue mass after tillage (kg/ha), R_o is the standing residue mass before tillage (kg/ha), and z_o is the surface random roughness (m). The trapping efficiency, ϵ, reflects the effect of vegetative height and is calculated by:

$$\epsilon = (e^{-0.1H}) - 0.1. \qquad [9]$$

The basal density of the standing vegetation, d_b, is a function of the mean stem diameter and the plant population and is calculated by:

$$d_b = d_s \, p_o^{0.5} /25 \qquad [10]$$

where d_s is the mean stem diameter of standing vegetation (m) and p_o is the plant population (plants/ha).

User inputs to the subroutine consist of the slope facing direction (A) in degrees from north, the land slope (S) in percent, and the length and width of the section in meters. The surface roughness (z_o) in m is obtained either from roughness measurements, values in the literature for given tillage conditions, or from a soil tillage model. The snow depth (D) in m is obtained from the snowmelt component, and precipitation (P) in m, mean minimum daily temperature (T_{min}) in °C, mean daily wind speed (v_h) in m/s, and mean daily wind direction (W) in degrees from north are all obtained from weather records or a climate generator subroutine. The height of standing residue (H) in m, mean stem diameter (d_s) in m, plant population (P_o) in plants/ha, and standing residue mass before and after tillage (R_o and R_t) in kg/ha are all obtained from either measurements or a plant growth subroutine.

The snowdrift component works in two parts, first calculating the amount of scouring or drifting of falling snow occurring on a section of the watershed during the day, including any snow drifting into the section from an upwind section, and then calculating the amount of drifting or scouring of the existing snowpack. If the drift rate (D_r) calculated for an upwind section is negative, indicating that snow in that section is drifting out of the section (scouring), then the amount of snow scoured from that upwind section is added to the snow available for movement in the next downslope section. For falling snow, a threshold velocity of 0.89 m/s at a height of 2 m is assumed for the incipient blowing of snow.

In order to route the blowing snow across a section, certain assumptions must be made. An upwind section at the top of a hillslope must accumulate snow unless the wind is blowing in a direction directly perpendicular to the direction in which the slope faces. Conversely, a downwind slope section at the top of a hillslope must scour unless the wind is blowing directly perpendicular to the direction in which the slope faces. If the wind blows perpendicular to a slope of a section, then no scouring or drifting occurs and the net change in snow accumulation in that section due to wind is zero.

The friction velocity at the snow surface is calculated using a commonly used mathematical representation of the wind profile (Schlichting, 1979):

$$v_h = (v_*/k) \ln (h/z_o) \qquad [11]$$

where v_h is wind velocity measured at height h (m/sec), v_* is friction velocity at the snow surface (m/sec), k is von Karman's constant (assumed to be 0.4), h is height above the surface (m), and z_o is surface roughness (m).

After v_* is determined, if the value of v_* is <0.087 m/sec (the friction velocity corresponding to a wind velocity of .89 m/s measured at a height of 2 m), then no movement of falling snow will occur and the new snow depth will be equal to the snow depth from the preceding day plus the depth of new snowfall. If $v_* \geq 0.087$ m/s, falling snow will begin to drift and the transport capacity of the wind is then calculated from an equation developed by Bagnold (1941) and modified by Iversen et al. (1975):

$$q_{sf} = c(d_a/g)(v_f/v_{thf})(v_{*2})(v_*-v_{thf}) \qquad [12]$$

where q_{sf} is the transport rate of snow (kg/m-s), c is a proportionality constant (= 100), d_a is density of air (kg/m³) (≈ 1), g is acceleration of gravity (m/s²) (= 9.82), v_f is the settling velocity of a snow particle (m/s) (for falling snow assume 0.35 m/s for a 0.150-mm snow particle falling in still air) (Schmidt, 1982), v_{thf} is the threshold velocity for incipient motion of falling snow (m/s) (assume 0.039 m/s) and v_* is the friction velocity at the snow surface (m/s).

The drift rate of falling snow over a section is then determined from:

$$D_f = 86.4 q_{sf}/d_f L_p \qquad [13]$$

where D_f is the drift rate of falling snow (m/day), d_f is density of falling snow (kg/m³) (≈100 kg/m³), and L_p is distance across a section parallel to wind direction (m).

While threshold velocity for incipient movement of snow varies with the nature of the snow surface (Radok, 1977), for a uniform surface of freshly fallen snow, the threshold friction velocity for movement of snow from the snowpack is approximately 0.25 m/s (Tabler and Schmidt, 1986), which is equivalent to a wind velocity of about 5.76 m/s at a height of 2 m. However, the threshold friction velocity for movement of snow from a snowpack is a function of snowpack density and, thus, increases with time since deposition (Schmidt, 1980). The threshold velocity for movement of snow from a snowpack, v_{thg}, in m/s can be estimated from:

$$v_{thg} = -0.023(z_o/3.2\text{x}10^{-7})^{0.5}/\{(1\text{-}\sin[\tan^{-1}\{S/100\}])(\ln 0.01d_g)\} \qquad [14]$$

where z_o is surface roughness (m), S is slope (%), and d_g is snowpack density (kg/cm³). If the calculated value of v_*-v_{thg}, then snow on the ground will begin to move. The transport capacity of wind for moving ground snow is then calculated in a fashion similar to that for calculating the transport capacity of wind for moving falling snow, as:

$$q_{sg} = c(d_a/g)(v_g/v_{thg})(v_{*2})(v_*\text{-}v_{thg}) \qquad [15]$$

where q_{sg} is transport rate of ground snow kg/m-s, v_g is the settling velocity of a ground snow particle (m/s) (for ground snow assume 0.75 m/s for a 0.220-mm ice sphere falling in still air) (Schmidt, 1982), and v_{thg} is the threshold velocity for incipient motion of ground snow (m/s). The drift rate of ground snow for the section is then calculated from:

$$D_g = 86.4q_{sg}/d_gL_p \qquad [16]$$

where D_g is the drift rate of ground snow (m/day), and p_g is the density of the snowpack (kg/m³).

The density of a snowpack on the ground is a function of several factors, including time and temperature. Daily changes in snowpack density are calculated on the basis of the initial depth of the snowpack and how much snowmelt occurs each day. In the absence of snowmelt, changes in snowpack density are estimated daily from the relationship:

$$d_g = 0.522\text{-}(20.5/D)(1\text{-}e^{-0.0148D}) \qquad [17]$$

where D is existing snow depth (m). This relationship is based on 14 years of premelt snowdrift data (Tabler, 1985).

If snowmelt occurs while snowpack density is less than 350 kg/m³, the depth of the snowpack is reduced by the amount of melt but the amount of melt water is added to the remaining snowpack, thus, increasing its density. Once snowpack density equals or exceeds 350 kg/m³, any additional melt water will either infiltrate the ground or run off.

If D_g exceeds the existing snow depth, D, in that section, then D_g is set equal to D. The total drift rate, D_r, is the sum of the drift rates for falling snow and ground snow, or:

$$D_r = D_f + D_g. \qquad [18]$$

After the total drift rate in a section is calculated, the wind direction and topography of that section will determine whether snow is drifting into the section or out of it. Maximum scouring will occur if wind direction is the same as the direction which the slope is facing and maximum drifting will occur if the slope facing direction and wind direction are exactly opposite each other. As previously stated, if the wind is blowing perpendicular to the slope, net scouring or drifting will be zero. Thus, to determine the net movement of blowing snow into or out of an area, the total drift rate, D_r, must be multiplied by a drift factor, s_c, to reflect either scouring or drifting:

$$s_c = 0.0111|A\text{-}W|\text{-}1 \quad [19]$$

where A is the slope azimuth (degrees from north), and W is the wind direction (degrees from north).

As the degree of slope inclination increases, the efficiency of the drifting process tends to decrease. This is accounted for by multiplying the net movement of blowing snow by an efficiency factor, i, based on land slope:

$$i = 1 - \sin[\tan^{-1}(S/100)] \quad [20]$$

If, due to wind angle and slope azimuth, the net movement of snow is positive, i.e., drifting into a section rather than out of it, the drifting snow will be distributed along successive downwind sections according to an exponential decay function of the form:

$$D_p = 1\text{-}[e^{-Lr}/(1+10L_r)] \quad [21]$$

where D_p is the total percentage of available drifting snow falling on an upslope section (%) and L_r is the ratio of the length of the upslope section to the total slope length in a downwind direction.

Not all blowing snow will be deposited, since some of it will evaporate. Net sublimation or evaporation losses can be an important consideration in climatic hydrological balances (Branton et al., 1972). The amount of evaporation is a function of air temperature, relative humidity, solar radiation, and particle diameter (Sturges and Tabler, 1981). An estimate of the amount of evaporation occurring can be obtained by considering the distance snow is being blown along a slope and assuming that under average conditions, complete evaporation would occur after being blown a distance of about 3050 m (Tabler, 1975). Then:

$$D_r = D_r(e^{-0.00066Lp}) \quad [22]$$

where L_p is the distance across a section parallel to the wind direction (m). Evaporation losses are only calculated for those sections in which drifting is occurring. Evaporation losses of snow from areas that are scouring would be accounted for in their downwind areas and are neglected here.

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LIST OF SYMBOLS

Symbol	Definition	Units
A	slope azimuth	° from north
C	a proportionality constant	
c_{uf}	heat capacity of the unfrozen soil	W/m-°C
D	existing snow depth	m
D_f	drift rate of falling snow	m/day
D_g	drift rate of ground snow	m/day
D_p	total percentage of available drifting snow falling on an upslope area	%
D_r	total drift rate	m/day
d_a	density of air	kh/m^3
d_b	basal density of standing residue	m/m
d_f	density of falling snow	kg/m^3
d_g	density of ground snow	kg/m^3
d_s	mean stem diameter of standing vegetation	m
F	forest cover	%
g	acceleration of gravity	m/s^2
H	height of standing vegetation	m
h	height above surface	m
i	efficiency factor based on slope inclination	%
K_{ftill}	thermal conductivity of frozen tilled soil	W/m-°C
K_{futil}	thermal conductivity of frozen untilled soil	W/m-°C
K_i	thermal conductivity of any layer, i	W/m-°C
K_{res}	thermal conductivity of residue	W/m-°C
K_{snow}	thermal conductivity of snow	W/m-°C
K_{srf}	average thermal conductivity of the snow-residue-frozen soil system	W/m-°C
K_{uf}	thermal conductivity of unfrozen (a) tilled soil, (b) untilled soil	W/m-°C
K_w	unsaturated hydraulic conductivity of (a) tilled soil, (b) untilled soil	m/s
k	Von Karman's constant (0.4)	
L	latent heat of fusion	W-s/m^3
L_{at}	latitude	
L_l	distance across a slope section parallel to wind direction	m
L_p	distance across a slope section perpendicular to wind direction	m
L_r	ratio of length of the upslope section to the total slope length	
M	snowmelt	m
N	estimated cloud cover	dec %
N	number of layers	
P	mean daily precipitation	m
P_o	plant population	plants/ha
P_{wp}	change in total water potential	m
Q_{srf}	heat flux through the snow-residue-frozen soil system	W/m^2
Q_{uf}	heat flow from unfrozen soil	W/m^2
q_{sf}	transport rate of falling snow	kg/m-s
q_{sg}	transport rate of ground snow	kg/m-s
R	estimated radiation on a sloping surface	MJ/m^2
R_c	potential clear sky radiation on a horizontal surface	MJ/m^2
R_{esd}	residue thickness	m
R_m	mean measured daily solar radiation	MJ/m^2
R_o	standing residue mass before tillage	kg/ha
R_t	standing residue mass after tillage	kg/ha
S	land slope	%
S_{snowd}	snow depth	m
S_t	storage capacity for snow	m
s_c	scour or drift factor	%
T_d	mean daily dewpoint temperature	°C
T_x	daily maximum temperature	°C
T_{illd}	tilled soil depth	m
T_{srf}	temperature gradient across the snow-residue-frozen soil thickness	°C
T_{uf}	change in soil temperature from °C isotherm to depth of stable temperature	°C
dT_{uf}	change in temperature of unit volume of soil in unit time	°C
U_{tilld}	untilled soil depth	m
v_f	settling velocity of a falling snow particle	m/s
v_g	settling velocity of a ground snow particle	m/s
v_m	wind velocity measured at height h	m/s
v_{thf}	threshold velocity for incipient motion of falling snow	m/s
v_{thg}	threshold velocity for incipient motion of ground snow	m/s
v_*	friction velocity at the snow surface	m/s
v_2	mean daily wind speed measured at a height of 2 m	m/s
W	wind direction	° from north
$z_.$	surface random roughness	m
Z_c	depth of unfrozen soil that supplies heat as a result of changes in soil temperature	m
dZ_i	thickness of any layer, i, in the snow-residue-frozen soil system	m
Z_{rf}	total thickness of the snow-residue-frozen soil system ($\Sigma\, dZ_i$)	m
dZ_{uf}	depth of unfrozen soil to point of stable temperature	m
ϵ	trapping efficiency	%

CONSERVATION APPLICATIONS IMPACTED BY SOIL FREEZE-THAW

By

Gary E. Formanek, Gary B. Muckel, and W. R. Evans[1]

INTRODUCTION

Many impacts are considered by land owners and users of lands where soil freeze-thaw occurs. For example, the rancher drains the stockwater pipeline on the high elevation range before winter. The logger plans for harvest considering loads on access roads when frost boils could occur. The farmer selects crops for planting considering the frost free period. The decisions reflect the knowledge of the impacts of freezing and thawing soils on their activities.

Local Soil and Water Conservation Districts, with assistance from the Soil Conservation Service (SCS), have some basic knowledge of the local conditions. Generally the available information includes soils with potential for frost action, climate and hydrology data, local resource management systems, and other similar data.

The SCS, as part of the United States Department of Agriculture (USDA), is directed to: (1) give top priority to reduction of erosion on agricultural land with emphasis on protection of highly erodible croplands and preservation of wetlands; and (2) improve and protect the quality of surface water and subsurface water by undertaking efforts to avoid harmful contamination from nonpoint sources and thereby maintain the quantity of water available for beneficial uses. The USDA continues to provide assistance in reducing upstream flood damages through its Small Watershed Program with many of the new projects in the past five years being primarily for land-treatment projects (USDA, 1989). Erosion, surface, and subsurface water, and damages from runoff are national priorities with erosion on cropland having the greatest emphasis by the USDA.

LOCATION OF FREEZE-THAW IMPACTED AGRICULTURAL LANDS

About 4.2 million square kilometers (1.2 million square miles) of agricultural lands, which is nearly half of the lands of the United States of America (USA), are impacted by freezing and thawing. In areas such as the discontinuous permafrost of Alaska the impact can be severe, for example, land clearing that allows decline of permafrost and soil subsidence. Frost heave, where it occurs, can raise young plants out of the soil, break root systems and cause dehydration and crop failure. In areas along the 40th parallel the impact of freeze-thaw may be infrequent.

Figure 1 shows the areas identified as impacted. In Alaska, the boundary includes farmlands, timber producing areas, and grazing land for cattle and sheep. The tundra range of Alaska is not included (although important to reindeer herds and the local economy). Along or south of the 40th parallel the boundary follows a 139 degree celsius freezing index adjusted by the authors to a land resource region boundary. The boundary dips south to include the cooler, higher elevations of the mountainous areas of the west. Included are the forest lands, croplands, and rangelands of the west, the wheatlands of the Great Plains, the specialty crops of the Great Lake states and the northwest, as well as, the forage, dairy, livestock, and feed grains regions of eastern and central USA.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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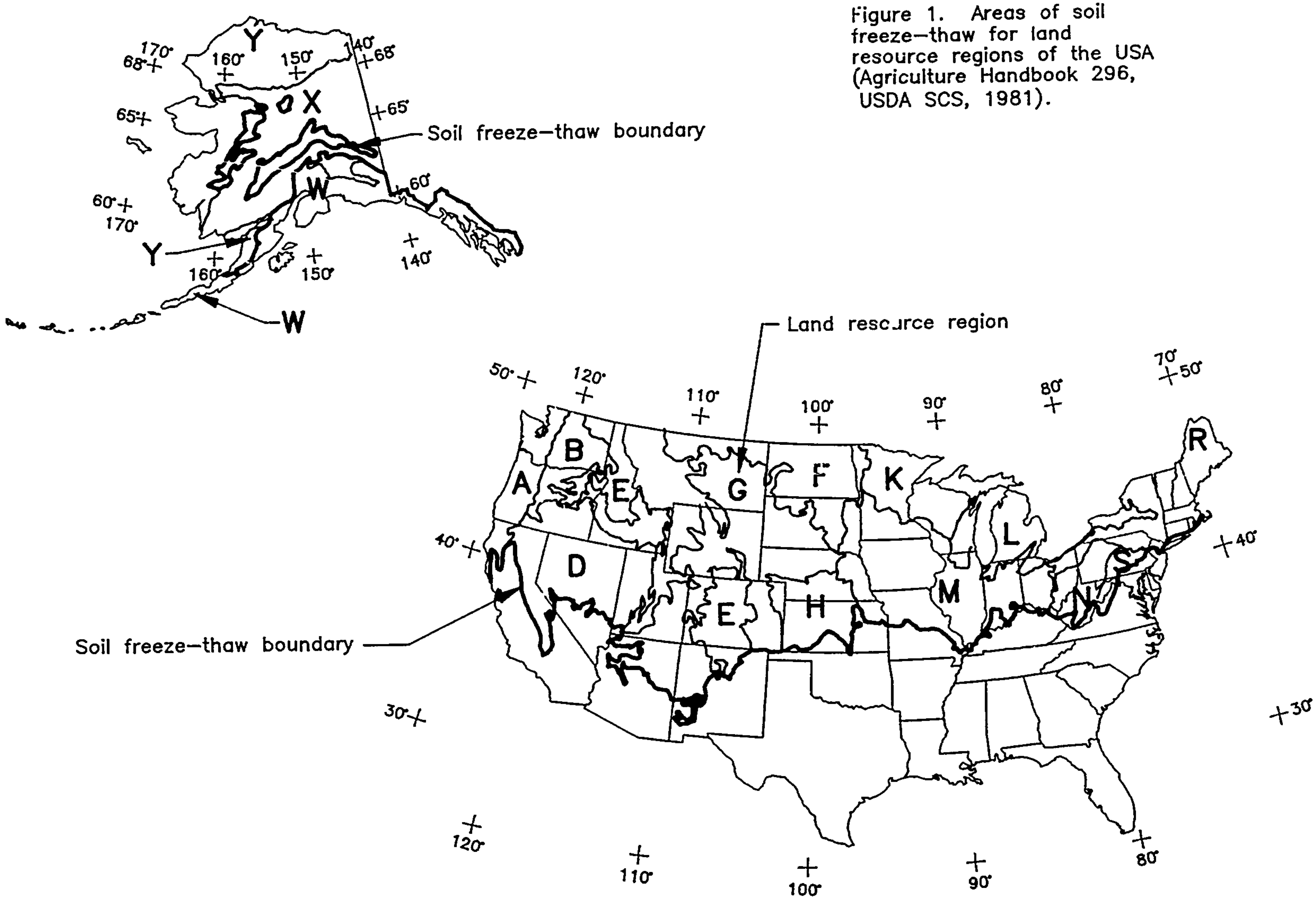

Figure 1. Areas of soil freeze–thaw for land resource regions of the USA (Agriculture Handbook 296, USDA SCS, 1981).

The land uses shown in Table 1 were interpreted from the land resource regions found in the Agriculture Handbook 296 (USDA SCS,1981).Rangeland,pasture land and grazed woodland are combined into the one land use of grazing land. The resulting areas of cropland, forest land, and grazing land impacted by soil freeze-thaw are shown.

Table 1. Approximate USA area of freeze-thaw impacted agricultural land

Land resource region *	Cropland sq.km.	Forest land sq.km.	Grazing land sq.km.
A	16720	158440	5060
B	60030	38510	113710
D	32010	144140	73890
E	24410	290820	158930
F	220280	7710	116990
G	77590	2450	376180
H	132270	--	72070
K	58210	171400	32720
L	108060	41560	18800
M	435880	78270	102850
R	39270	231830	11130
X,W(Alaska)	120**	105470	18000
	1204850	1270600	1765420

* USDA SCS, 1981.
**Alaska Agricultural Statistics, 1988.

SOIL SURVEY

Much of the SCS knowledge of soil freeze-thaw is contained in the National Soils Handbook (USDA SCS, 1983) and the individual soil surveys published by the National Cooperative Soil Survey. In these documents potential frost action is defined to be the likelihood of upward or lateral movement of soil by the formation of segregated ice and the subsequent loss of soil strength upon thawing. The soil moisture state and depth of frost is not available in a national data base. In regions where frost action is a potential problem soil scientists rate the effect on soil into three classes:

(a) Low -- soils are rarely susceptible to the formation of ice lens,
(b) Moderate -- soils are susceptible to the formation of ice lens, resulting in frost heave and subsequent loss of soil strength,
(c) High -- soils are highly susceptible to the formation of ice lens, resulting in frost heave and subsequent loss of soil strength.

These classes consider soil moisture regime, soil temperature class, and family particle size class. The estimates are based on bare soil that is not covered with insulating vegetation or snow. The natural soil moisture regime is assumed, although the ratings can be related to man-made modifications of drainage or irrigation on an onsite basis. Frost action estimates are made for the whole soil to the depth of frost penetration, to bedrock, or a depth of two meters (6.6 feet), whichever is shallower. The resulting ratings are generally displayed in a table of soil features for each soil name and map symbol of the soil survey.

RUNOFF AND EROSION

For design of individual conservation practice applications there is no national SCS procedure to directly estimate peak runoff and volume from frozen or thawing soils. Only unfrozen soils are represented by the hydrologic soil groups and runoff curve numbers used for direct estimates of runoff from ungaged watersheds (USDA-SCS, 1972). The antecedent moisture limits for "dormant season" apply when the soils are not frozen and there is no snow on the ground.

Site specific design procedures, within the SCS, generally include an adjustment to increase runoff by adding about four points to the runoff curve number. This adjustment is considered adequate for volume estimates only. The frequency of water available for runoff from snowmelt and precipitation is available for selected agricultural areas in the northwest U.S.A. (USDC NOAA, 1983).

In the western Oregon part of Land Resource Region A, Figure 1, subsurface drainage is used to reduce winter runoff and erosion from soils saturated above a restricted layer (USDA SCS 1987). Sheet and rill erosion on non-irrigated cropland in Land Resource Region B, Figure 1, are dominated by runoff occurring on thawing soil. Practices such as removal of soil moisture by annual cropping, and residue management to slow heat transfer at the soil surface are effective in reducing erosion. In the northwest USA, the SCS uses a regional relationship (McCool and George, 1983) to relate erosivity to mean annual precipitation, and erosion to slope length and steepness.

QUESTIONS FOR CONSERVATION APPLICATION

The general effects of freeze-thaw are known for many applications. The impact on soil erosion, for example, was stated by Bennett fifty years ago; "With the freezing of water in the surface soil frost heaving occurs and serves to loosen the soil so that the particles and aggregates can be more easily carried away." (Bennett, 1939). Yet questions remain such as rates of sheet (or inter-rill) and rill erosion and amounts of residue or surface roughness or other practices needed for erosion control. National programs such as the Food Security Act of 1985 (P.L.-99-198) need quantities as well as comparisons to be most effective. Quantification of effect is available for few areas and for few applications. The effect of the freeze-thaw cycle on infiltration, runoff rates and runoff volumes continues to be needed for flood protection. More information is needed for the effect on water quantity and quality as well. Additional soil water by infiltration could destabilize steep, sensitive slopes or increase frost susceptibility or affect water quality, particularly where soils are shallow or near water tables. The evaluation of systems adds complexity not yet considered. The landowner and landuser will continue with their questions until satisfactory alternatives are available.

SUMMARY

Land owners and users on about half the USA agricultural land, which has frozen soil with freeze-thaw cycles, need more information to practice conservation on their lands. Although the effects of freeze-thaw are qualitatively known for soil erosion, more quantitative information is required to plan for new programs. Effects on peak runoff, surface or subsurface water quality and quantity, are not well established and require qualitative and quantitative information to plan conservation applications.

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HEAT AND WATER FLUX IN A DIURNALLY FREEZING AND THAWING SOIL

by

J. L. Pikul, Jr. and J. F. Zuzel[1]

INTRODUCTION

Nighttime soil freezing and daytime thawing cycles are numerous in the inland Pacific Northwest. Frequent wetting and drying, that is associated with freezing and thawing, may accelerate breakdown of soil aggregates (Hinman and Bisal, 1968). Surface crusting results in decreased water infiltration (Moore, 1981). Laboratory studies also reveal a decrease in erosion resistance in repeatedly frozen and thawed silt loam soil (Formanek et al., 1984). Management practices that leave crop residue on the surface have been shown to decrease the penetration of soil frost (Pikul et al., 1986). These same management practices should also lessen problems with reduced infiltration and soil erosion.

Soil freezing induces water migration from unfrozen subsoil to the freezing front. As the soil surface freezes, a fraction of the water in the soil pores freezes. Ice particles remain separated from the soil by a thin water film; the thickness of the unfrozen water film depends upon temperature, pore size distribution, solutes in the pore water, as well as freezing and thawing history (Anderson and Tice, 1972). Freezing of water effectively dries the soil in the region of ice formation thereby decreasing the matric potential. Water flows towards the region of low matric potential.

The objective of this paper is to: 1) summarize field research, conducted at Pendleton, OR, on diurnal soil freezing and thawing and 2) illustrate further work on simulating heat and water flux during diurnal freezing and thawing.

PROCEDURES

Two field treatments were located 15 km northeast of Pendleton, OR. Annual precipitation is 400 mm and occurs mostly as rain from October through June. Soil at the site is a Walla Walla silt loam (member of the coarse, silty, mixed mesic family of Typic Haploxerolls) with no appreciable slope. Winter wheat (*Triticum aestivum* L.) was harvested in August 1981, leaving a straw residue of 2.2 Mg/ha. Two soil surface covers for different heat transfer characteristics were located adjacent to each other in the field. The stubble was burned on a plot 20-m by 40-m to free it of all harvest residue (B). In a second unburned 20-m by 40-m plot the chaff was spread to produce a uniform distribution of residue (C).

Hydraulic potential of unfrozen soil, defined as the sum of matric and gravitational potential referenced to the soil surface, was measured using porous cup tensiometers. Duplicate ceramic cups, 1 cm diameter by 3 cm long, were placed at 2, 5, 9, 17, 25, 35, 45, and 60 cm depths on each treatment. Additional experimental details are given by Pikul and Allmaras (1985).

Soil temperature, on each treatment, was measured using duplicate temperature probes which electronically averaged the signal from three thermocouples. Soil temperature was recorded at 0.5 hr intervals. Depths of thermocouples were 1-cm increments from 1 to 5 cm; 2-cm increments from 7 to 25 cm; 30, 35, 40, 50, and 60 cm. Soil temperature near the surface was measured with thermocouples placed at the 0.1-cm depth.

Soil water content was measured at about 1-hr intervals. Gravimetric samples were a composite of three subsamples taken at 1-cm depth increments in the surface- to 11.0-cm soil layer. This sampling method has been described by Pikul et al. (1979).

Depth of frozen soil was determined at the time of gravimetric water sampling. The term *frozen soil* indicates a discernible mass of soil. The line of demarcation between unfrozen soil and frozen soil was determined by the resistance of the soil to cutting.

Presented at the Frozen Soil Symposium, March 21-22, 1990, Spokane, Washington

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Additional experimental details on the measurement of soil bulk density, saturated hydraulic conductivity, and the soil water characteristic curve are given by Pikul and Allmaras (1986). Unsaturated hydraulic conductivity was estimated using the Marshall pore-interaction model (Green and Corey, 1971). Isothermal water diffusivity (Fig. 1), as a function of water content was obtained as the ratio of hydraulic conductivity to specific water capacity. Calculation methods for soil thermal conductivity and soil heat capacity are given by Pikul et al. (1989).

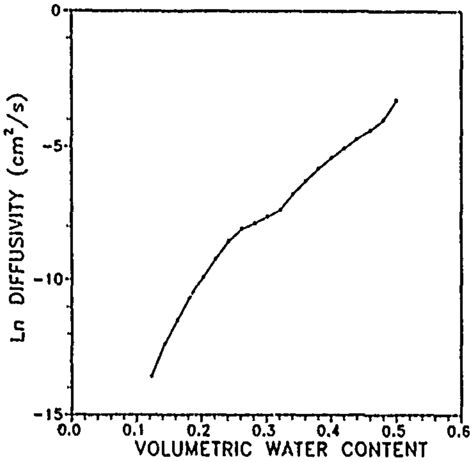

Figure 1. Isothermal soil water diffusivity as a function of volumetric water content.

MODEL

Finite difference methods were used to solve the partial differential equations for heat and water flow. An experimental relationship, given by Pikul et al.(1989), between unfrozen water content and temperature was used to estimate the change in volumetric ice content with respect to time. The change in volumetric ice content was used to couple the heat and water flow equations by providing an estimate of heat and water sinks or sources. Soil heaving was not considered and as such accumulated ice and water are restricted to be less than the pore volume.

The partial differential equation for heat flow where there is phase change between liquid and ice is written as

$$c\rho \frac{\partial T}{\partial t} - L_f \rho_f \frac{\partial \theta_f}{\partial t} = \frac{\partial}{\partial x}\left(\lambda \frac{\partial T}{\partial x}\right) \qquad [1]$$

(i) (ii) (iii)

where x is the position coordinate (cm), λ is apparent soil thermal conductivity (W/cm °C), T is temperature (°C), $c\rho$ is volumetric specific heat (J/cm³ °C), t is time (s), L_f is latent heat of fusion (J/g), ρ_f is density of ice (g/cm³), and θ_f is volumetric ice fraction (cm³/cm³).

Term (i) in Eq. [1] represents the rate change in energy of the soil layer due to a change in temperature; (ii) is a sink or source that is based on the rate change of ice content within the freezing soil; and (iii) is soil heat flux by conduction.

Solution to Eq. [1] was obtained numerically using a Crank-Nicholson finite difference scheme. Known temperatures at 0.1-cm and 17.0-cm depths provided boundary conditions for the heat flow solution. Finite difference equations for Eq. [1] are given by Pikul et al. (1989).

The partial differential equation for water flow where there is phase change between liquid and ice is written as

$$\frac{\partial \theta_w}{\partial t} + \frac{\rho_f}{\rho_1}\frac{\partial \theta_f}{\partial t} = \frac{\partial}{\partial x}\left[D(\theta_w)\frac{\partial \theta_w}{\partial x}\right] + \frac{\partial K(\theta_w)}{\partial x} \qquad [2]$$

(i) (ii) (iii) (iv)

where $D(\theta_w)$ is isothermal liquid diffusivity (cm^2/s), θ_w is volumetric water content (cm^3/cm^3), ρ_l is density of water (g/cm^3), and $K(\theta_w)$ is the hydraulic conductivity (cm/s).

Term (i) in Eq. [2] is the rate change of soil volumetric water content; (ii) is the water sink or source term, whereby water is added or removed from storage based on the rate change of ice content in the freezing soil; and (iii) is the flux of water into and out of the soil layer due to the water content gradient. In respect to term (iii) the water content gradient implicitly represents the matric potential gradient, which is the true driving force for water flow. Term (iv) is the flux of water due to gravity.

Solution to Eq. [2] was obtained numerically using a Crank-Nicholson finite difference scheme. Boundary conditions for the water flow solution were zero flux at the 0.1-cm and 17.0-cm depths. Finite difference equations for Eq. [2] are given by Ungs et al. (1985).

Predicted values were compared to measured values using the standard deviation of the differences calculated as $(\Sigma(\text{Predicted} - \text{Measured})^2/n - 1)^{1/2}$.

RESULTS

Field measurements

Surface cooling during radiation frosts of 20 to 25 March froze the soil each night on the B treatment. On the C treatment there was no measurable frozen soil, however a discontinuous ice lattice about 2 mm thick formed on the soil surface between the rows. During one freezing cycle on 22 March the soil froze to a depth of 1.5 cm on the B treatment; water content of the 0- to 1-cm layer increased from 0.31 at 1800 hours to 0.49 cm^3/cm^3 at 0600 hours. On the C treatment water content increased from 0.33 to 0.39 cm^3/cm^3 over the same time interval.

On the C treatment, in contrast to the B treatment, wetter soil and slow rate of cooling may have favored ice crystal growth near the surface. During these nighttime freezing cycles soil temperature at 1 mm remained above freezing on the C treatment. The numerical model described here uses measured soil temperature at 1 mm as an upper boundary condition, consequently this method was not applicable on the C treatment where freezing occurred above the depth of the first temperature sensor.

Tensiometers in the unfrozen soil provided a means to monitor profile water redistribution during freezing and thawing cycles (Figs. 2 and 3). A typical freezing,

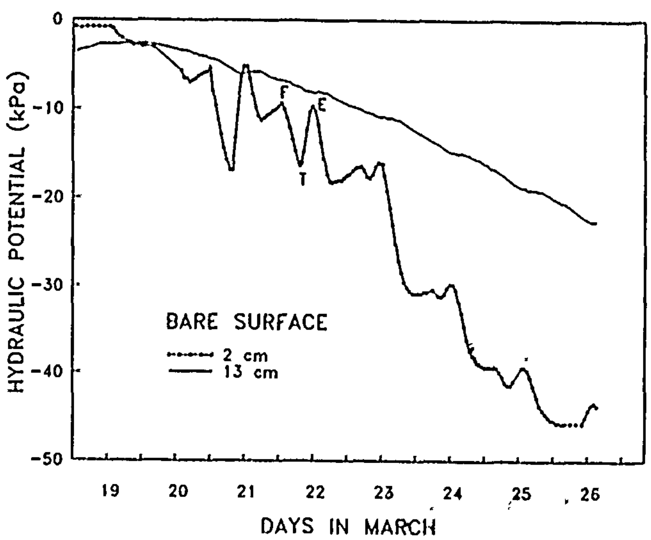

Figure 2. Hydraulic potential at 2 and 13 cm for the bare surface treatment. Calendar days are labeled at 1200 hours. The letters F, T, and E indicate the onset of surface freezing, thawing and evaporation, respectively.

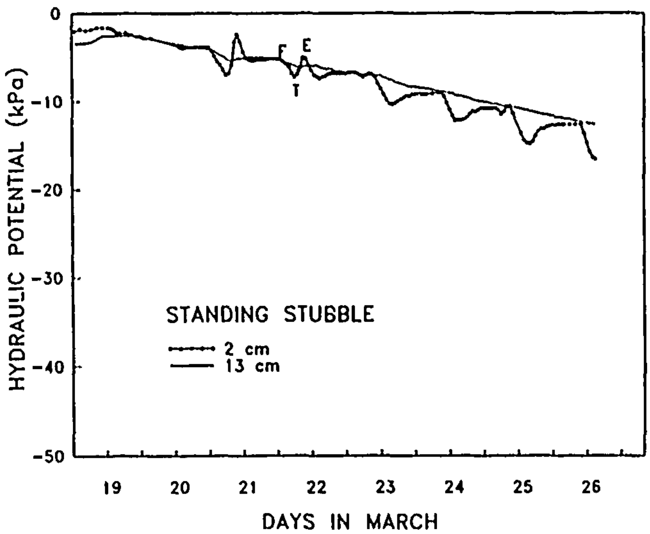

Figure 3. Hydraulic potential at 2 and 13 cm for the residue covered treatment. calendar days are labeled at 1200 hours. The letters F, T, and E indicate the onset of surface freezing, thawing, and evaporation.

thawing, and evaporation cycle, for 22 March, is marked, on Figures 2 and 3, by the letters F, T and E, respectively. For example, during one freezing cycle on 22 March when the soil froze to a depth of 1.5 cm on the B treatment, hydraulic potential at 2 cm decreased from -9.4 at 0100 hours to -16.0 kPa at 0700 hours (Fig. 2). This decrease in potential is a consequence of soil drying as water moved from the 2 cm depth and towards the surface. Hydraulic potential at 2 cm on the C treatment decreased from -5.2 to -6.9 kPa over this same time (Fig. 3). Hydraulic potential at 5 and 9 cm (not shown) on the B treatment also exhibited diurnal changes. At 13 cm there was little diurnal fluctuation in hydraulic potential (Fig. 2).

Soil water evaporation was greater from the B treatment than from the C treatment. During the 24-hr period starting at 1800 on 22 March, the B treatment lost 2.4 mm of water from the top 11 cm as compared to 0.3 mm from the same depth on the C treatment. Evaporative water loss in the 0- to 11-cm profile was determined from changes in measured water content. Hydraulic potentials at 13 cm (Figs. 2 and 3) and below (not shown) did not indicate water drainage.

Diurnal fluctuations in hydraulic potential at 2 cm, due to freezing, diminished as the surface on the B treatment dried (Fig. 2). The rate of water movement to the freezing front is determined by both the hydraulic potential gradient and soil hydraulic conductivity. For this soil, Allmaras et al. (1977) have measured a hundred-fold decrease in hydraulic conductivity for a decrease in water content from 0.34 to 0.24 cm^3/cm^3. Water content, at 2 cm on the B treatment, decreased from 0.34 to 0.22 cm^3/cm^3 during 21 to 23 March.

Simulation Results

Soil temperature and water content was simulated for 21 and 22 March, when the soil froze to a depth of 1.5 cm on the bare surface treatment. A comparison of measured and predicted soil temperature at 1-cm is shown in Figure 4. Correlation of measured soil temperature and predicted soil temperature was 0.99. Standard deviation of the differences between predicted and measured soil temperature was 1.04°C.

Predicted and measured ice and water content of the 0- to 1-cm soil layer are shown in Figure 5 for 21 and 22 March for the B treatment. Ice and water content tend to be overpredicted at the onset of soil freezing. Simulated ice and water content at the time of maximum frost penetration was 0.48, which compares to a measured water content of 0.49 cm^3/cm^3.

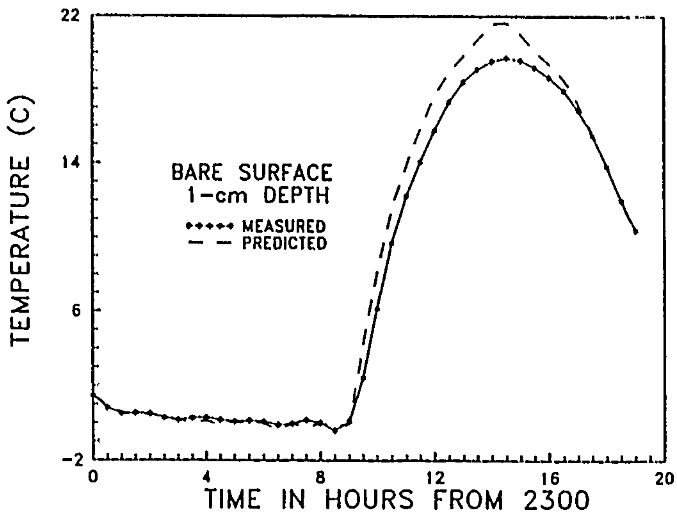

Figure 4. Comparison of predicted and measured soil temperature at 1.0 cm for the bare surface treatment during 21 and 22 March.

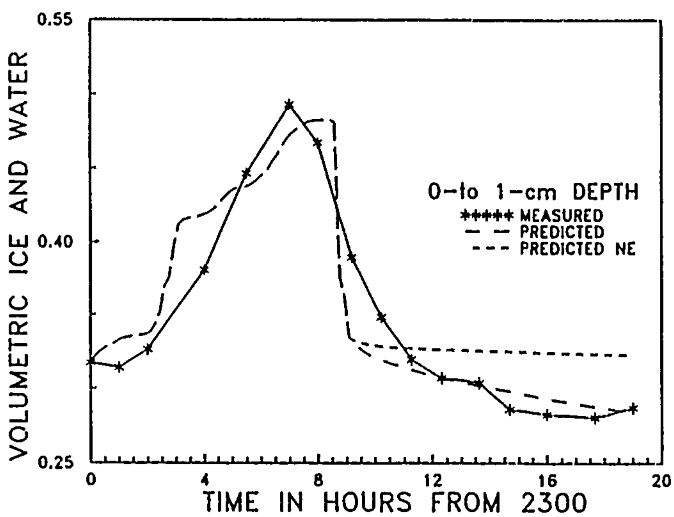

Figure 5. Comparison of predicted and measured ice and water content of the 0- to 1-cm soil layer for the bare surface treatment during 21 and 22 March. Predicted NE labels simulation results obtained with no evaporative water loss during non-freezing periods.

Previously, Pikul et al. (1989) neglected soil water evaporation during non-freezing periods. A comparison of simulation results including and excluding water loss to evaporation during non-freezing periods is shown in Figure 5. Soil water evaporation was simulated by specifying a water sink at 0.1 cm. The strength of the evaporative sink was determined from the measured change in water content of the 0- to 11-cm profile. Initial and final water content of the 0- to 17-cm profile for the evaporation simulation was 5.53 and 5.31 cm, respectively, corresponding to a water loss of 2.2 mm.

Additional simulations were run including the flux of water due to gravity, which is term (iv) in Eq. [2]. Previously, Pikul et al. (1989) used the simplifying assumption that gravitational fluxes of water are small compared to isothermal liquid flux. For this 19-hour simulation there was little difference in the results obtained by including or excluding term (iv). Results reported in this paper include the flux of water due to gravity.

Predicted depth of frozen soil was estimated from simulated soil temperature (Fig. 6). The soil was considered frozen when temperature at a node was $\leq 0.0°C$. Freezing point depression of the soil water solution was estimated to be -0.016°C. Maximum predicted depth of soil freezing was 1.4 cm, by comparison measured depth of frost was 1.5 cm.

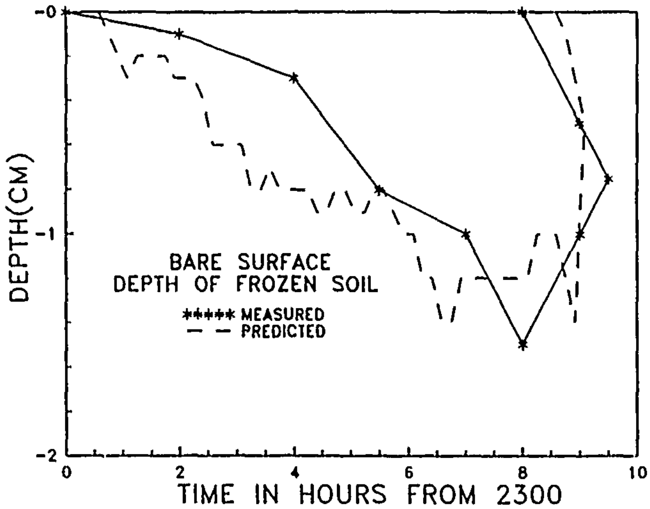

Figure 6. Comparison of predicted and measured depth of freezing for the bare surface treatment during 21 and 22 March.

CONCLUSIONS

Our field measurements show that surface cover reduces the incidence of soil freezing and that the degree of soil wetness, at the time of freezing, influences the amount of water that migrates to the freezing front. On the bare surface treatment water moved readily from unfrozen subsoil to the freezing front when the upper 2 cm of soil was wet ($0.34\ cm^3/cm^3$). Under these conditions water redistribution, due to freezing, resulted in near saturated conditions at the surface and consequently high evaporative water loss during the subsequent daytime thaw. During one 24-hr freezing and thawing cycle the bare surface treatment lost 2.4 mm of water from the upper 11 cm of soil. In contrast the residue covered treatment lost 0.3 mm of water. High water loss from the bare surface treatment is a consequence of the wet surface condition, resulting from soil freezing, and negligible surface cover to impede water evaporation. These observations underscore the importance of residue cover to reduce water evaporation and lessen the incidence and degree of soil freezing. The frequency of soil freezing may also be important from the standpoint of soil structure degradation overwinter, because wetting and drying cycles have been shown to accelerate the breakdown of soil aggregates and formation of surface crusts.

Soil temperature and water content during freezing and thawing were simulated. Finite difference methods were used to solve partial differential equations of heat and water flow. An experimental relationship between unfrozen water content and subzero temperature was used to estimate the change in volumetric ice content with respect to time. The change in volumetric ice content couples the heat and water flow equations by providing an estimate of heat and water sinks or sources. Measured temperatures at 0.1 cm and 17-cm were used to drive the simulation model. Boundary conditions for the water flow solution were specified as zero flux at 0.1-cm and 17-cm. Soil water diffusivity, as a function of water content, was determined experimentally on unfrozen soil. The functional relationship between soil water diffusivity and liquid water content was assumed to be the same for frozen and unfrozen soil. Soil heat capacity and thermal conductivity were calculated using mechanistic models that include ice in the system.

Results from the simulation compare favorably with field measured soil temperature, freezing depth and water distribution. The good agreement between measured and predicted temperature during freezing and thawing also suggests the sinks and sources of heat are being correctly estimated. This study supports the usefulness of this modeling approach for diurnal simulations of soil heat and water flux near the surface during freezing and thawing.

ACKNOWLEDGEMENTS

The authois wish to thank Allen Tice, located at the U. S. Army Corps of Engineers Cold Regions Research and Engineering Laboratory, Hanover, NH, for determining unfrozen water content of the test soil.

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INFLUENCE OF MANAGEMENT PRACTICES ON SNOWMELT RUNOFF

by

D.S. Chanasyk and C.P. Woytowich[1]

INTRODUCTION

Soil water deficiency limits agricultural production on both farmland and rangeland and is the major constraint to increased production on over two-thirds of the agricultural lands on the Canadian Prairies (Oosterveld and Nicholaichuk, 1983). Snow comprises 30 to 40% of the total annual precipitation on the Canadian Prairies and represents a significant potential source of manageable water. The quantity of potential soil water available from snow depends primarily on the depth of the snowcover and weather conditions during snowmelt. The amount of snowmelt water infiltrating and thus the amount of snowmelt runoff is dependent on soil properties affecting infiltration, slope degree, aspect, and surface roughness.

Willis et al. (1969) in a North Dakota study determined that quantity of snowmelt runoff increased with increased stubble height for the same soil water status, but soil water status had more influence on the quantity of runoff than did stubble height. They found that snowmelt generally began first on plots with tall stubble and duration of snowmelt increased with decreasing stubble height. The average duration of melt was six days. The proportion of snowmelt water that runs off is often high. Willis and Haas (1969) found if snow was held where it fell and if the soil was dry in the fall before freezing, spring runoff amounted to approximately 50% of the snowpack; if the soil was wet before freezing, approximately 80% ran off. Bauder et al. (1975) reported 80% or more of the winter precipitation in eastern North Dakota became runoff. Chanasyk and Woytowich (1986) studying four cropping practices in the Peace River region of Alberta found between 35 and 71% of the snowpack ran off.

A better understanding of the effects different cropping practices have on amount of runoff is essential before soil water enhancement through runoff reduction is possible. Most studies on the Canadian Prairies have concentrated on cereal grains and fallow rotations; canola and forages are also common on the Prairies but relatively little is known about their effect on snowmelt runoff. Farmers on the prairies generally cultivate their stubbled land in the fall after harvest to reduce their springtime workload. This practice reduces stubble cover but increases surface roughness.

The objective of the study reported here was to assess the effects of such management practices and crop cover type on runoff during the springmelt period. The study was initiated in the Peace River region of Alberta which is considered a high water erosion risk area. Major portions of the Peace River region utilized for agriculture have long (often up to 1.0 km in length) but gentle (between 1 and 5%) slopes, which combined with a fine-textured soil, pose a potentially high water erosion hazard.

RESEARCH METHODOLOGY

The study site was located near La Glace, Alberta, 50 km northwest of Grande Prairie and approximately 500 km northwest of Edmonton. Slopes located were 5% slope with a westerly aspect. Seventeen runoff plots, each 5 m wide and 75 m long, were established. Fifteen cm boards were used as borders and small H flumes were positioned at the bottom end. All plots were delineated by a board border. Treatments were: fallow, fescue, barley, and canola, replicated four times. Two barley, two canola and the fallow plots were cultivated upslope after harvest.

The experimental design was a paired randomized block to allow adjacent cultivated and non-cultivated plots, minimizing treatment differences due to soil variability. Paired plots were immediately adjacent to each other; these pairs were separated by a 1.0 m buffer zone. In the spring, each plot was cultivated upslope, fertilized, and seeded.

Presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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Swathing, tillage of selected plots, and preparations for winter were generally completed by October. All harvested straw was removed from the plots. Soil texture, water retention properties, and saturated hydraulic conductivity of the uppermost 30 cm were determined. Long term average meteorological data (1951-80) for the Grande Prairie airport were obtained from the Atmospheric Environment Service.

1985 and 1986 springmelt periods were monitored. Antecedent soil water conditions were measured during the preceding fall using a neutron probe at three locations within each plot to a depth of 1.0 m. A meteorological station was centrally installed at the site at the start of springmelt to measure air temperature, windspeed, radiation, relative humidity, and precipitation. Snow water equivalent was measured prior to melt using a Montrose snow sampling tube. Soil temperature was measured using thermistors. Three thermistor nests were installed at one third, one half, and two thirds of the way up each plot. Measurements at each nest were made at the surface and at depths of 10 and 20 cm daily during springmelt between 0900 and 1030 h local time. Runoff flow rates were determined at the time of sampling by measuring time required to fill a container of known volume. Depending on the rate of discharge, one of three different sized containers (284, 560, or 4500 ml) was used. Specific values were summed to obtain daily and event totals for each plot.

RESULTS AND DISCUSSION

Soil properties

The surface horizon (0-15 cm) was 20% sand, 41% silt, and 39% clay and was classified as clay loam according to the Canadian System of Soil Classification. Bulk density was 1.14 Mg m^{-3} and saturated hydraulic conductivity was 8.4×10^{-5} cm s^{-1}. Field capacity (-33 kPa) was 0.305 g g^{-1} and wilting point (-1500 kPa) was 0.157 g g^{-1}. The underlying horizon (15-30 cm) was 15% sand, 28% silt, and 57% clay and was classified as clay. Bulk density was 1.35 Mg m^{-3} and saturated hydraulic conductivity was 1.2×10^{-5} cm s^{-1}.

Meteorological parameters

The study area has a mean annual precipitation of 453 mm, approximately 40% of which occurs as snow. The months of June, July, and August have approximately equal amounts of precipitation: 70.0, 65.1, and 60.5 mm, respectively. Precipitation is equally distributed over the entire winter period. Snow may fall as early as September and as late as December, with October being average. Melt usually occurs in April, but earlier melts do occur. Mid-winter melts may also occur, usually due to a warming trend affiliated with a chinook. Rain storms during the winter are not common. July is the warmest month with a mean monthly temperature of 15.0 °C and January the coldest with a mean monthly temperature of -17.7 °C. Average frost-free period is between 105 and 118 days.

The two study years were characterized by high precipitation in September: 92.3 mm (247% of the long term average) in 1984 and 80.5 mm (215% of the long term average) in 1985, likely resulting in high soil water contents at the time of fall freezing. However, precipitation for the winter period (November - March inclusive) was lower than the long term average in both years: 90.9 mm in 1984-85 and 98.9 mm in 1985-86, compared to the long term average of 138.2 mm for the same period. Both years were characterized by significantly warmer than normal mean monthly temperatures in January (9.4 °C higher in 1985 and 9.0 °C higher in 1986. This may have resulted in some intermittent snowpack-air interface melting and mid-winter snow pack ripening.

Antecedent soil water

Fall water contents were more variable among treatments in fall 1984 than in fall 1985, ranging from 23.7 to 34.1% (volume) in 1984 but only from 33.3 to 36.2% in 1985. Soil porosity was approximately 50% so degree of saturation was approximately 47 - 68 % in 1984 and 67 - 72% in 1985.

The highest and second highest water contents were in the fallow in 1984 and 1985, respectively. Water content in the fescue was the same in the two years (33.8 %), ranking second in 1984 but last in 1986. In 1984 fall water content was higher in canola than in barley, but this trend was not evident in 1985. Water contents were not significantly different among treatment pairs (non-cultivated versus cultivated) in either year.

Snow water equivalent

The lowest spring snow water equivalent in both study years was in the fallow, but no other treatment trends were evident. There were greater snow water equivalents in 1986 than in 1985 for all treatments.

Snow water equivalent (m^3)

	Fallow	Fescue	Canola Stubble	Canola Cultivated	Barley Stubble	Barley Cultivated
1985	20.5	25.3	24.0	25.9	28.4	29.0
1986	27.6	35.3	31.1	27.7	34.9	41.2

There is no indication that stubble cultivation reduced the snow water equivalent of the plots. Of the eight possible cases to test this hypothesis, the standing stubble had a greater snow water equivalent in five. In almost all cases the margin of difference was very small.

Some of the differences among plots within each treatment were possibly due to location. None of the seventeen plots had equivalent wind exposure, fetch, or aspect. For example, plots 3 and 4 (fescue) had extensive drifts forming over them during the winter of 1985-86 due to vigorous vegetative growth in 1985. These drifts extended partially on either side of the plots; fallow on one side and barley on the other.

Soil temperatures

During the 1985 melt period, surface soil temperatures rose above 0 °C on March 23, with four plot exceptions. The 10 and 20 cm soil temperatures did not increase above 0 °C for any plot at any time during the melt period. In 1986, surface temperatures on many plots rose above 0 °C on April 2 or 3 with all plots above freezing by April 4. As in 1985, soil temperatures at 10 and 20 cm did not rise above freezing during the melt period.

Snowmelt

The snowmelt period was defined as beginning with the first appearance of runoff from any plot through to the termination of runoff from the last plot. The 1985 melt period began on March 15 and ended on April 1, although melt was not continuous. Maximum daily temperature was generally above 0 °C for the entire period except for March 28. Minimum temperature exceeded 0 °C only twice: March 18 and April 1. On March 23 a noon snow storm interrupted melt and there was no runoff on March 28 (maximum air temperature -0.8 °C). There were four events of measurable precipitation (snow) during the melt period (March 20, 23, 28 and 29), yielding a total of 4 mm, which was added to the snow water equivalent for each plot.

In 1986 the melt period was more prolonged, beginning on March 4 and ending on April 4. Again melt was not continuous; stopping on March 5 and resuming on March 26. Air temperatures dropped sharply on the 5th with maximum air temperature reaching only -10 °C on the 7th. Maximum air temperatures rose above 0 °C on the 10th remaining there until the 23rd. Minimum air temperatures remained well below 0 °C from March 5 until March 19. Five precipitation events during the melt period (March 8,18, 19, 24 and April 4) yielded 13 mm of snow.

Runoff

In 1985 the fescue plots had the greatest mean total runoff volume followed by barley (standing stubble), canola (standing stubble), barley (fall cultivated), canola (fall cultivated) and fallow in descending order. In 1986 the order of decreasing total runoff volume was barley (fall cultivated), fescue, barley (stubble standing), canola (stubble standing), canola (fall cultivated), and fallow. Runoff trends in 1986 were not as clear as those in 1985 due to snowdrifting effects from adjacent plots. Increased soil water status as a result of substantial precipitation in September and October 1985 and a higher snow water equivalent resulted in greater total runoff volumes from all treatments in 1986 compared to 1985.

Total Runoff Volume (m^3) for individual plots in spring 1985 and spring 1986

	1985			1986			
Fescue	25.6	24.9	22.5	46.6	30.6	26.0	27.0
Fallow	13.6	11.5	13.7	23.5	33.2	18.7	17.5
Barley Stubble	19.1	26.5		31.6	28.5		
Barley Cultivated	18.3	24.9		43.5	25.4		
Canola Stubble	20.9	23.8		27.4	25.4		
Canola Cultivated	25.1	16.5		28.2	22.7		

The data for the fesuce and fallow plots can be used to assess the variability of springmelt runoff for a given treatment.

Statistics for springmelt runoff (m^3) for fescue and fallow for the two study melts

	1985			1986		
	Mean	SD	CV (%)	Mean	SD	CV (%)
Fescue	24.3	1.6	6.6	32.6	9.6	29.4
Fallow	12.9	1.2	8.6	23.2	7.1	30.6

The reproducibility of the data in 1985 was very good (coefficient of variation less than 10%) but poor in 1986 (CV of approximately 30%). The cvs were quite similar for the two treatments, as was the trend in CV, despite the fact that the two treatments had dramatically different covers.

Runoff coefficients

Runoff coefficients (ratio of runoff to snow water equivalent) were high both study years, ranging from 57-94% in 1985 and 72-95% in 1986. Fescue had very high coefficients, averaging 90% for the two years.

Runoff Coefficients (%) for spring 1985 and spring 1986

	1985			1986			
Fescue	84	82	94	95	86	95	91
Fallow	69	57	60	86	93	80	72
Barley Stubble	80	81		86	87		
Barley Cultivated	69	79		93	72		
Canola Stubble	87	99		94	77		
Canola Cultivated	83	77		90	95		

The average overall runoff coefficient in 1985 was 79% (CV 12%) and 87% (CV 8%) in 1986. The low runoff coefficients for fallow in 1985 may be attributable to the rather late cultivation of these plots the preceding fall (October 10), as compared to September 12 in 1985. Cultivating barley reduced the runoff coefficient by approximately 5% in each of the two study years. The trend for canola was not clear; cultivation reduced the runoff coefficient by 13% in 1985 but raised it by 8% in 1986.

Prediction of total snowmelt runoff

The effects of management practice, fall soil water and spring snow water equivalent on plot runoff were investigated using Pearson correlation matrices and stepwise linear regression. Significant parameters were then used to test the general linear hypothesis. Tillage was non-parametrically coded: 1=fall tilled (including fallow); 2=stubble standing; and 3=fescue. Results of the multivariate general linear hypothesis revealed that total runoff was dependent only on snow water equivalent and fall tillage. The equation derived for predicting snowmelt runoff was:

$$\text{Runoff} = 0.883 * \text{snow water equivalent} + 2.123 * \text{tillage parameter} - 4.578$$

with runoff and snow water equivalent expressed in m^3.

CONCLUSIONS

Total runoff was greatest from the fescue plots; least from the fallow. Total runoff was significantly correlated to snow water equivalent and fall cultivation. Greater snow water equivalents increased runoff while fall cultivation decreased it. For all treatments, between 80 and 90% of the snow melt water ran off. The high runoff coefficients are likely due to the inability of the snow melt waters to infiltrate because of frozen soil and sloped land.

The study has shown that runoff can be reduced by cultivation. Thus farmers who cultivate fields in the fall may increase their soil water content, although the increase would not likely be large.

The long term representativeness of the results is difficult to assess. The above normal precipitation in both Septembers of the study would likely mean that subsequent springmelt runoff would be above normal. However, this is compensated by the below normal over-winter precipitation that occurred. The above normal January temperatures the two study years would have meant an earlier ripening of the snowpack, but not necessarily an earlier melt. Both springmelts were prolonged, suggesting the runoff values may be conservative. Thus, the runoff values measured were likely below average for the region.

ACKNOWLEDGEMENT

The authors gratefully acknowledge funding received from Farming for the Future.

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ENVIRONMENTAL CONDITIONS AND PROCESSES ASSOCIATED WITH RUNOFF FROM FROZEN SOILS AT REYNOLDS CREEK WATERSHED

by

Mark S. Seyfried, Bradford P. Wilcox and Keith R. Cooley[1]

INTRODUCTION

Most severe flooding and erosion events in the Pacific Northwest occur in the winter and are the result of rain and snowmelt on frozen soil (Johnson and McArthur, 1973). Frozen soil runoff is also important in other parts of the United States and Canada (e.g., Dunne and Black, 1971; Granger et al., 1984). The high runoff is generally attributed to the limited infiltration capacity of concretely frozen soil (Zuzel et al., 1982).

Accurate description of runoff from frozen soils with simulation models has proven to be difficult (Wilcox et al., 1989). This is partly due to the complex nature of the relationship between soil frost and soil infiltration capacity. Freezing soil may result in either increased or decreased infiltration capacity (Trimble et al., 1958; Haupt, 1967). Cooley and Robertson (1983) found that infiltration into frozen soil ranged from 0 to 100% of the total rain plus snow water storage at a rangeland site. In addition, macroporosity, which is difficult to predict, affects infiltration into frozen soil (Harris, 1972).

Climate and vegetation also affect runoff from frozen soil. These effects are clearly illustrated in mountainous terrain where large climatic differences due to elevation changes can be observed over short distances. In northern California (Taylor, 1969) and Oregon (Hale, 1950) impermeable, concrete frost has been observed to be more prevalent at lower elevations where sagebrush dominates than at higher, forested sites. This corresponds to what Cooley and Robertson (1983) refer to as the transient snow zone. They consider that zone to be most susceptible to frozen soil runoff.

Soil frost data have been collected at the Reynolds Creek Experimental Watershed in southwest Idaho for about 20 years (Hanson et al., 1988). The watershed is about 240 km^2 in extent and ranges in elevation from 1097 to 2195 m. Annual precipitation increases with elevation from 237 to 1587 mm. Several smaller subwatersheds have been instrumented which represent different climatic and vegetational conditions.

Although the entire watershed is subject to frozen soil conditions, frost is deepest and soil temperatures coldest at lower elevations. On the other hand, runoff is greatest from higher elevations. This suggests that environmental conditions at Reynolds Creek bracket a zone most conducive to frozen soil runoff generation. In order to better understand processes controlling runoff from frozen soil, our objectives were to: (1) locate the zone of maximum frozen soil runoff at Reynolds Creek using soil frost and runoff data; and (2) determine what conditions within that zone cause it to be so conducive to generation of frozen soil runoff.

STUDY AREA AND METHODS

Site description

Hydrographs from six subwatersheds, Flats (FL), Nancy's Gulch (NG), Lower Sheep Creek (LS), Reynolds Mountain (RM), Tollgate (TG), and Salmon Creek (SC), all located within the Reynolds Creek watershed, were used to analyze runoff. Characteristics of those subwatersheds are shown in Table 1. Runoff from all the subwatersheds is ephemeral and event-based except RM, which is perennial and shows a strong spring snowmelt peak. About 80% of the annual precipitation at the higher elevations occurs from November through April and is mostly snow. This compares to about 60% of annual precipitation in the same period

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Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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at lower elevations of which about 25% is snow (Hanson and Morris, 1982; Hanson et al., 1988). FL and NG are at low elevations, RM is at high elevation and LS is intermediate between the two. TG and SC range from high to low elevations.

Table 1. Subwatershed characteristics.

	Elev	Area	MAT	MAP	Vegetation
	(m)	(ha)	(°C)	(cm)	
Flats	1183-1192	0.9	6.8*	25.5	Wyoming Big Sagebrush
NG	1414-1426	1.3	6.8*	31.7	Wyoming Big Sagebrush
LS	1588-1658	13	6.0#	37.9	Low Sagebrush
RM	2018-2143	40	3.4#	113.8	Mountain Big Sagebrush
SC	1387-1829	129	Vary	Vary	Vary
TG	1402-2225	5444	Vary	Vary	Vary

* From Quonset, 1964-1987
\# 1967-1987
MAT- Mean annual temperature
MAP- Mean annual precipitation

Data collection

There was a period of 13 years, from water years 1972 to 1984, in which all six subwatersheds were continuously monitored for runoff and soil frost data were collected. Runoff was reported on a per unit area basis. Precipitation, air temperature, and frost depth were collected at sites immediately adjacent to the NG, LS, and RM subwatersheds. FL data were taken from the quonset weather station located at approximately the same elevation about 3 km away. Precipitation was measured with a dual gauge system (Hanson et al., 1980). Daily maximum and minimum air temperatures were measured by a hygrothermograph at the quonset, LS, and RM weather stations.

From 1972-1976 depth of soil frost up to 10 cm was measured. After that frost tubes, similar to those described by Harris (1970), and frost blocks (Burgess and Hanson, 1979), located immediately adjacent to the site in all cases, were used. Calibrated neutron probe readings were made at LS from 1976-1984 at depths of 15, 30, 60, and 90 cm. All measurements were made at approximately biweekly intervals.

Data interpretation

Runoff data from the FL, NG, LS and RM subwatersheds were used to establish the effects of a climatic (elevational) gradient on frozen soil runoff while SC and TG were used to establish the extent of the elevational band of frozen soil runoff events within the Reynolds Creek watershed. Frozen soil runoff was characterized in terms of quantity (average annual runoff) and frequency (number in 13 years) of events.

Runoff events were considered to be frozen soil related if the measurement made prior to the event indicated frozen soil conditions. Air temperature was used to estimate if soil frost was present where unusually long extrapolations from measurement to event were required. We refer to runoff events as frozen soil related, as opposed to frozen soil caused, because, although we have good data concerning soil conditions when the events commence, we have very little information on the progress of conditions during the events. This was due to the biweekly sampling regime. This is important because measurements taken subsequent to events frequently indicated unfrozen conditions.

Determination of when specific runoff events commenced and ended was generally quite straightforward except at RM where most runoff occurs as a single, large, fluctuating spring snowmelt event. This was not considered to be frozen soil related because the soil was generally not frozen at the outset and because virtually no surface runoff is observed during these events. That is, the normal snowmelt from RM generates runoff by subsurface flow. The events considered to be frozen soil related at RM were distinct, winter runoff events.

The frequency (number during the 13 year measurement period) of "significant" frozen soil related events at different subwatersheds was also determined. "Significant" runoff events were defined as having a peak daily runoff greater than 1.0 mm or greater than 6 mm total runoff. Some distinction of this kind was necessary due to the many small, indistinct events at all the subwatersheds that contributed little to the total runoff. As defined, significant runoff events accounted for 71 to 90% of the total frozen soil related runoff except at RM where frozen soil related runoff was relatively minor (Table 2).

Table 2. Runoff Summary (Water years 1972-1984)

Site	Average Annual Runoff	FSR* Runoff Fraction	Number of Significant FSR* Runoff Events	Fraction of FSR* Runoff from Significant Events
	(mm)			
Flats	1.6	0.91	3	0.90
NG	1.8	0.94	2	0.71
LS	8.3	0.83	16	0.86
RM	600.9	0.04	5	0.22

* Frozen Soil Related

Air temperatures were used to distinguish between rain and snow with air temperatures above 0°C assumed to be rain, those below assumed to be snow. In some cases there was a mixture of rain and snow using this approach. In these cases the events were classified as rainfall events.

RESULTS AND DISCUSSION

Amount and frequency of frozen soil related runoff

Average annual runoff from FL, NG, LS, and RM increased with elevation, as expected (Table 2). The portion of that runoff from frozen soil related events shows the opposite trend, with runoff from the three lower elevation subwatersheds almost entirely frozen soil related while that from RM was relatively insignificant.

The domination of frozen soil related runoff at FL, NG, and LS results from a combination of soil and climatic factors. The climatic factors include the precipitation distribution and intensity of most storms in the region. As mentioned earlier, about 60% of the precipitation falls from November through April at the lower elevations, which brackets the time that the soil is frozen. Most of these storms are regional frontal systems characterized by fairly low (<5.0 mm/hr) precipitation intensities (Hanson and Morris, 1982). This compares with measured soil infiltration capacities of greater than 60 mm/hr at LS (Johnson et al., 1984), and about 40 mm/hr at NG (unpublished data), which is very similar to FL (Johnson et al., 1984). Since overland flow is the chief pathway for runoff from these watersheds (there is no base flow), either a reduction in soil infiltration capacity and/or an increase in precipitation intensity are required to generate runoff. Frozen soil is known to be virtually impermeable under the proper water content and temperature conditions.

Conditions at RM are quite different. Almost all runoff there is of subsurface origin and overland flow is rarely observed. This, plus the relatively thick snowpack that tends to insulate the soil and reduce soil frost, results in a large amount of runoff with relatively little effect of frozen soil.

There was substantially more frozen soil related runoff from significant events and more significant runoff events at LS than the other subwatersheds during the monitoring period (Table 2). All significant runoff events at FL, NG, and LS were frozen soil related while RM had significant runoff events each year which resulted from subsurface flow of snowmelt and were not frozen soil related.

Distribution of frozen soil related runoff

The data in Table 2 indicate that conditions at LS are more conducive to the generation of frozen soil related runoff than conditions at the other subwatersheds. If LS is representative of a climatic zone that is active throughout the larger watershed, then events at LS should be common to other subwatersheds with overlapping elevation (and therefore climate) ranges. Two such subwatersheds are Salmon Creek (SC), and Tollgate (TG). The LS subwatershed contributes runoff to neither.

Figures 1 and 2 illustrate how runoff patterns from LS, SC, and TG are related. In 1975 (Fig. 1), there were three distinct frozen soil related events at LS on days 24, 44, and 57, all of which were significant. In each case, LS runoff was of shorter duration and had greater peak flow than either TG or SC, but the timing of the events was very similar. There were three smaller peaks after day 61 in which runoff from TG and SC greatly exceeded that from LS. Those events occurred after the soil thawed at LS and were primarily the result of snowmelt from higher elevations in SC and TG.

The 1984 runoff patterns (Fig. 2) also show distinct frozen soil related events. The first two events, occurring near days 3 and 23, were not significant at LS while the third, near day 43, and fourth, near day 67, were. As in 1975, and all other years observed, there was considerable snowmelt runoff from TG and SC after the soil thawed and the snow had melted at LS. 1984 differed from 1975 in that peak flow from both TG and SC was greater from the frozen soil related events than LS. Runoff from both 1975 and 1984 illustrate how closely events at LS matched those at TG and SC. Over 13 years of comparison, there was runoff at SC and TG corresponding to all 16 significant events at LS. At TG, 11 of those events were significant while 12 were significant at SC.

Although we have no direct soil frost measurements representative of TG or SC, it seems reasonable to assume that runoff events from those subwatersheds that correspond to frozen soil related events at LS are also frozen soil related. The strong correspondence between events at LS and the other two subwatersheds is consistent with the existence of an event-specific elevational (or climatic) band of frozen soil related runoff throughout the Reynolds Creek watershed. The alternation of maximum peak runoff between LS, SC, and TG suggests that the location of the band varies with event although other factors, such as the precipitation distribution, may also account for the observed alternation.

Comparison of runoff from FL, NG, and RM, which represent different elevational zones, further shows how the band of frozen soil related runoff varies with event. Figure 3 (1976) illustrates some "typical" runoff patterns at FL, NG, LS and RM. The peaks for NG and FL occurred early in the year and were relatively small. All runoff from FL ended by day 10 and that from NG was complete by day 18. Three distinct peaks (two significant events) at LS were observed between days 55 and 75 with smaller events following. Runoff at RM was unaffected by the events at lower elevations and did not rise above the base level until shortly after day 90.

The small runoff events at FL and NG (Fig. 3) near day 10 were the result of rain on frozen soil. That same storm produced snow at LS and RM and thus no runoff. When rain fell on days 56, 57, and 59 snow covered frozen soil at LS, while the snow had melted and the soil thawed at FL and NG. Thus there was runoff from LS but not at FL or NG. Subsequent runoff events at LS were due to rainfall "sandwiched" between snowstorms. The smaller peaks that followed were due to a combination of rainfall and snowmelt without rain. The runoff producing storms at LS either produced snow at RM or the rain produced was absorbed by the snow pack and no runoff resulted. Runoff from RM came later when the general snowmelt occurred. In this year the band of frozen soil related runoff moved from the lower elevations early in the year up to LS later. This is consistent with long-term maximum average monthly runoff trends. The month of maximum runoff for the four subwatersheds is: January for FL and NG, March for LS, and May for RM (Wilcox et al., 1989).

The data presented above indicate that frozen soil related runoff occurs in event-specific bands. The extent of the bands and the intensity of runoff within them determines the severity of individual runoff events downstream. LS is apparently representative of an elevational zone between about 1450 and 1750 m in which the event bands are most often located (Table 2).

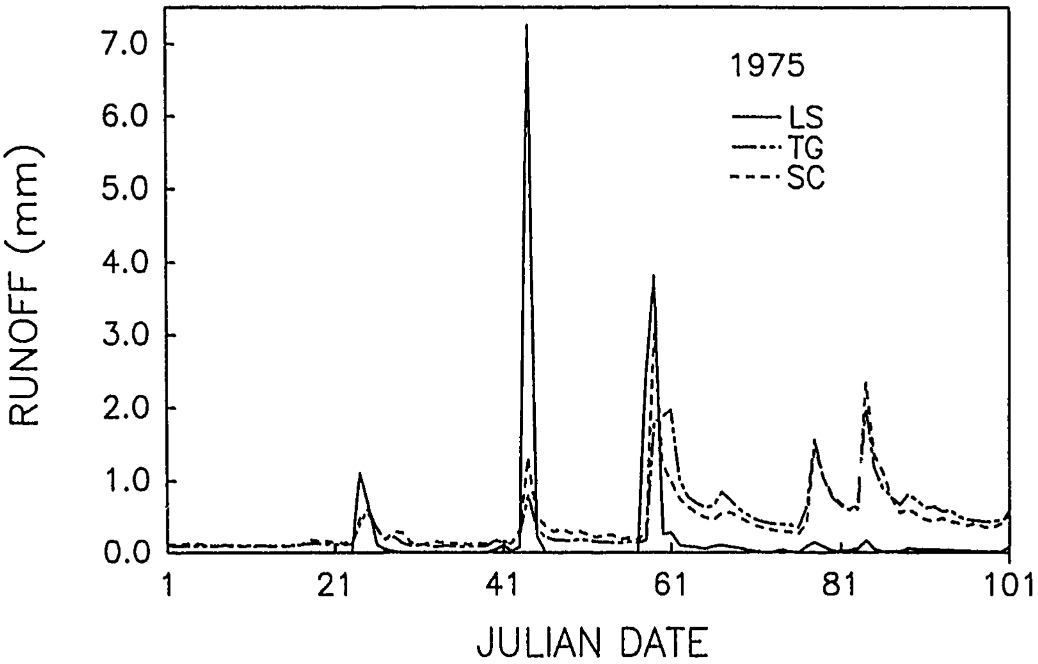

Figure 1. Daily runoff from the Lower Sheep Creek, Tollgate, and Salmon Creek subwatersheds during the first 101 days of 1975.

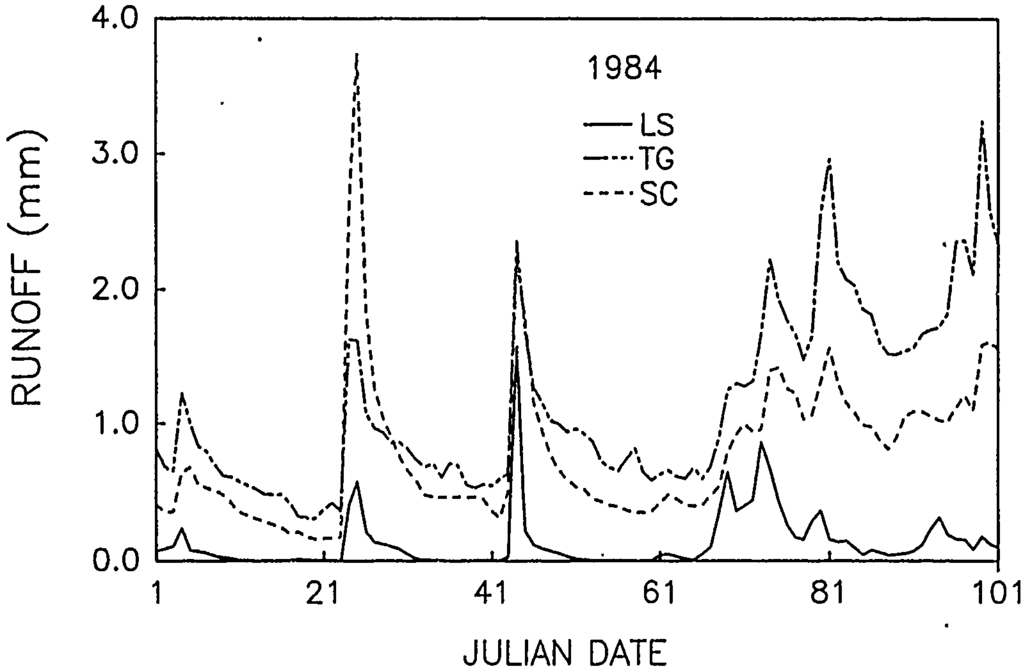

Figure 2. Daily runoff from the Lower Sheep Creek, Tollgate, and Salmon Creek subwatersheds during the first 101 days of 1984.

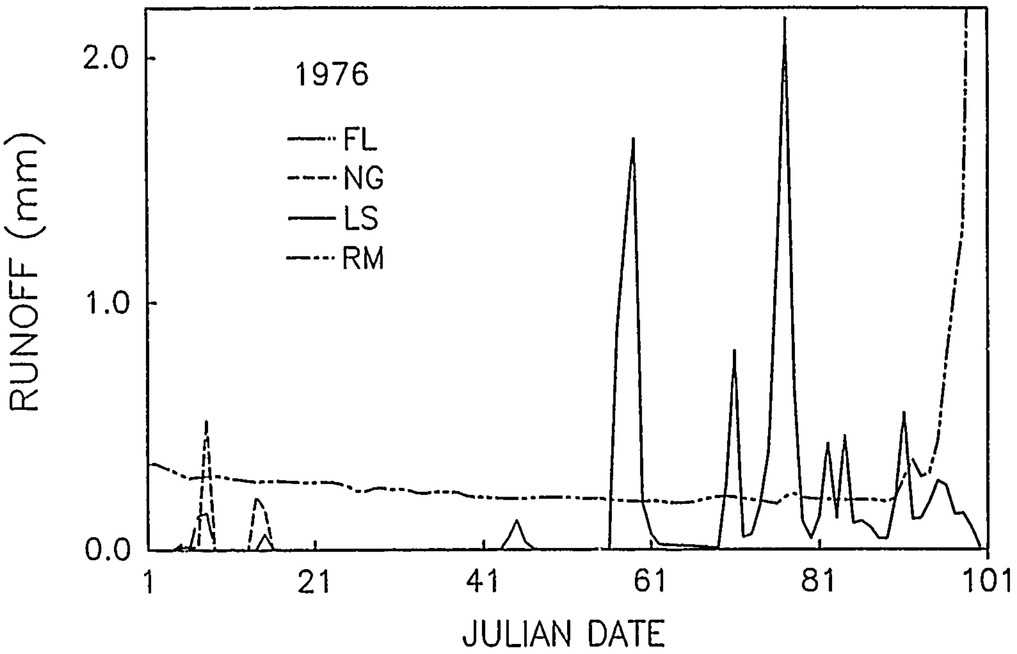

Figure 3. Daily runoff from the Flats, Nancy's Gulch, Lower Sheep Creek, and Reynolds Mountain subwatersheds during the first 101 days of 1976.

Precipitation/runoff relationships at LS

Since LS represents the zone of maximum frozen soil related runoff, we investigated runoff characteristics in relation to precipitation at LS to establish conditions most conducive to the generation of frozen soil related runoff. There was very poor linear correlation (r=0.427) between total annual precipitation and runoff at LS (Fig. 4). This has also been observed at the FL and NG subwatersheds (Wilcox et al., 1989). The scatter of data in Figure 4 reflects the fact that while about 84% of runoff was frozen soil related, only about 23% of the precipitation occurred during frozen soil conditions. Thus, the majority of precipitation is ineffective at producing runoff.

A large improvement in linear correlation (r=0.813) resulted when yearly frozen soil related runoff and frozen soil precipitation were regressed (Fig. 5). Further improvement was expected in correlation between frozen soil related rainfall (precipitation minus snowfall) and runoff, but the linear correlation was actually reduced (r=0.627). This indicates that snow is an important component of frozen soil related runoff. In fact, snow was on the ground at the commencement of all 16 significant runoff events at LS. There was also rainfall on 15 of those events. Thus, it was the combination of rainfall on snow and frozen soil that resulted in frozen soil related runoff events.

The effects of snow largely explain why LS represents the maximum zone of frozen soil related runoff. At higher elevations, where snowfall is much greater, snow cover tends to insulate the soil and reduce soil freezing. In addition, the snow cover buffers the system from occasional winter rainfall and warming trends by absorbing rain and melt water within the snow pack. The process of ripening the snow pack further assures that the soil is thawed during snowmelt. By the time snow has melted, the likelihood of freezing conditions is greatly reduced. These effects of snow explain why frozen soil related runoff is relatively insignificant at RM and other similar watersheds (Troendle, 1985).

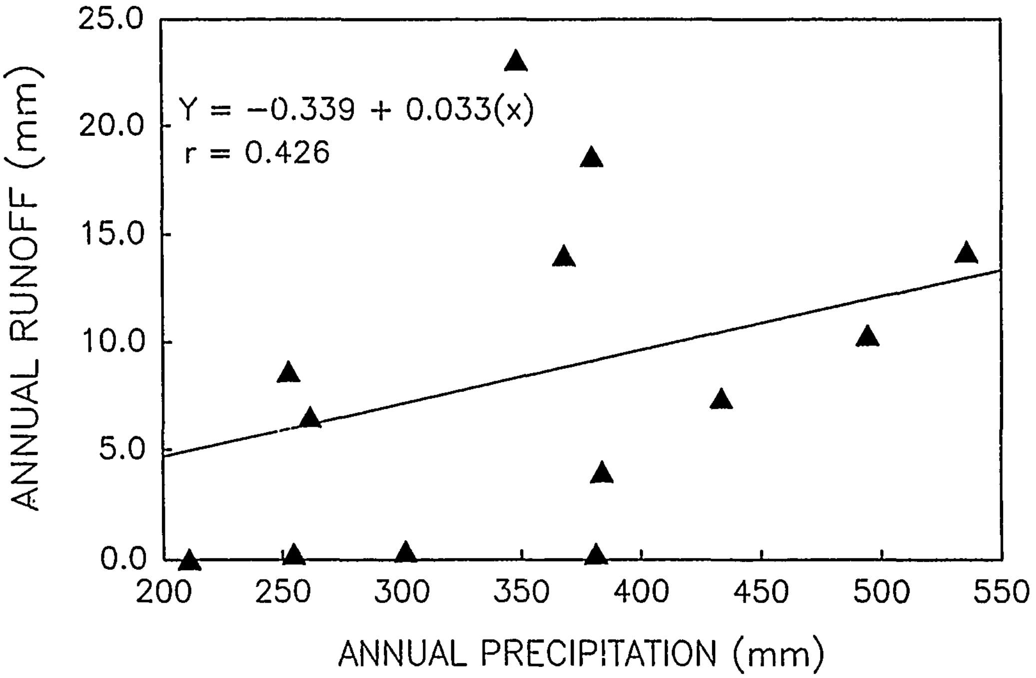

Figure 4. Relationship between annual runoff and precipitation at Lower Sheep Creek for water years 1972-1984.

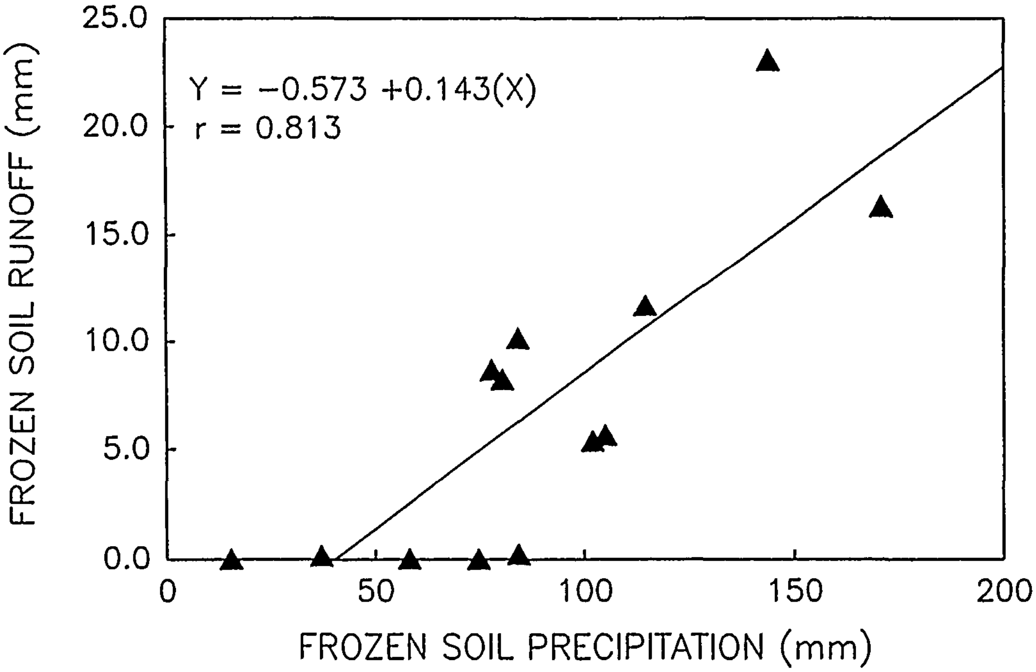

Figure 5. Relationship between frozen soil related precipitation and frozen soil related runoff at Lower Sheep Creek for water years 1972-1984.

A shallow (10 to 35 cm) snow cover, on the other hand, can facilitate frozen soil related runoff in conjunction with rain. Snowmelt alone produces little runoff either because it is of insufficient quantity or low intensity and allows for soil thawing. Similarly, rainfall alone appears to produce little runoff, presumably for the same reasons. In combination, however, the effective rainfall rate and quantity is increased due to the added melt water. A shallow snow cover does not provide much insulating effect so that soil freezing proceeds freely, and the buffering effects of snow are greatly reduced.

The dominant vegetation in the subwatershed is low sagebrush, which rarely exceeds 35 cm in height. Strong winds are common throughout the winter and the only snow retained in the subwatershed is found in topographic lows or is "caught" by the sagebrush. Thus the vegetation and climate combine to make LS most conducive to frozen soil related runoff.

Although the conditions conducive to generating runoff occur at LS most often, they can extend throughout Reynolds Creek. For example, during two large runoff events in 1972, most of the Reynolds Creek watershed produced runoff simultaneously, and on one occasion, in 1982, all six of the subwatersheds contributed significant runoff from the same rainfall event. On these occasions the threat to flooding and erosion is greatest.

Infiltration during frozen soil related events

Examination of Figure 5 indicates that only a small portion (10-15%) of the precipitation landing on frozen soil appears at the weir as runoff at LS. This could be partly due to infiltration, transmission losses, sublimation and the effect of wind redistribution of snow. Although the dual gauge system is quite accurate for rain and snowfall measurement, the actual water input to the watershed from snow is strongly affected by wind redistribution during and subsequent to snowfall. This results in areas of blow-on (drifts) and blow-off. We have assumed that blow-on and blow-off are balanced within the watershed, which appears to be reasonable but is not confirmed by measurement. Other unmeasured losses include channel transmission losses and sublimation. Even if these losses resulted in a 50% reduction in water available for runoff, the majority of precipitation on frozen soil would still not be accounted for.

The remaining possible fate for the water is infiltration into the soil. The average soil-water content of the two access tubes at LS is shown in Figure 6 for 1976. There was a distinct increase in soil-water content in the two upper depths, while the two lower depths changed little between days 34 and 49. The reading on day 49 immediately followed a small runoff event (Fig. 3). This was followed by a substantial soil-water content increase at all depths between days 49 and 77. The reading on day 77 followed the highest peak runoff for the year at LS. In all 8 of the significant runoff events at LS during the neutron probe monitoring period (1976-1984) there was an associated increase in soil-water content. In this case (day 77) it extended to a depth of at least 90 cm. Three explanations for these observation are that:

1. There is infiltration through the frozen soil. Soil frost does not necessarily render the soil impermeable and can, in some cases, actually increase infiltration capacity (Haupt, 1967; Hinman and Bisal, 1973). Intermediate infiltration capacities are also possible, which could allow for large amounts of infiltration;

2. Spatial variability results in areas of high infiltration capacity. It has been shown that the unfrozen infiltration capacity varies considerably in sagebrush rangelands (Johnson and Gordon, 1988). This may also affect frozen soil infiltration capacity. In addition, microclimatic, vegetational, and aspect interactions may result in highly variable frost conditions; and

3. The infiltration front may follow the thaw front. Observations of soil frost indicate that frozen soil related storm events often result in soil thawing. Each of these explanations is plausible and carries different implications in terms of modeling or predicting soil behavior.

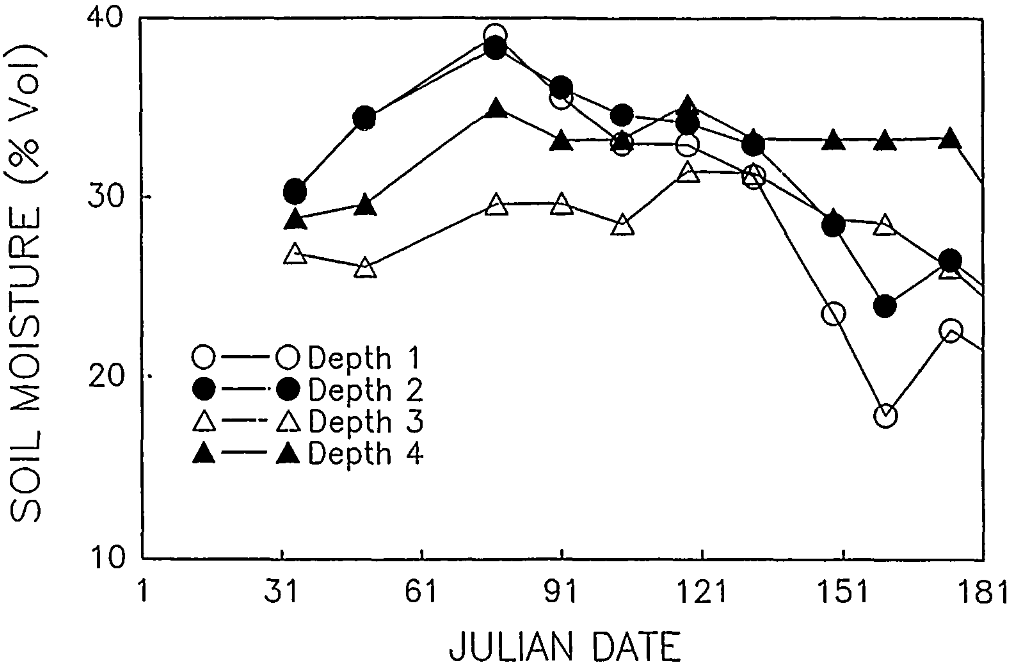

Figure 6. Average volumetric soil moisture content measured at two locations on the Lower Sheep Creek subwatershed during 1976. Depths 1, 2, 3, and 4 are 15, 30, 60, and 90 cm, respectively.

SUMMARY AND CONCLUSIONS

Most of the runoff generated in the Reynolds Creek Watershed and similar areas results from snowmelt that enters streams by subsurface flow. The area contributing to this flow is relatively small, however, being restricted to elevations above 1750 m. Runoff from the far more aerially extensive lower elevations is mostly via surface flow related to frozen soil. Severe flooding and erosion events can result when these areas contribute runoff because they are so aerially extensive. The severity of these events depends, in part, on the extent of the area contributing to runoff and the intensity of runoff from those areas.

The following conditions are associated with most runoff from the lower elevations: frozen soil when the event begins, shallow (<35cm) snow cover; and rainfall. These conditions, and the runoff associated with them, are generally found in event-specific elevational bands within the watershed. These fall most often in a zone between 1450 and 1750 m at Reynolds Creek.

It appears that most of the water potentially available for runoff during frozen soil related events infiltrates into the soil. Thus the intensity of runoff seems to be controlled by infiltration processes. There are several possible explanations for this which require further investigation.

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THE EFFECT OF FROZEN SOIL ON EROSION - A MODEL APPROACH.

by

Peter F. Botterweg[1]

INTRODUCTION

At the end of the seventies the Norwegian authorities recognized that agriculture contributes considerable to the total environmental pollution. Losses of soil by erosion and losses of nutrients, especially phosphorus, by both erosion and surface runoff are the main causes of eutrophic fresh waters in agricultural areas (Min. of Agriculture, 1984). Besides the non-point source pollution, point pollution occures by leakage of silage effluent and leakage from manure storages. In 1987 an algae bloom sprang up in the coast waters between Denmark, Sweden and Norway. It is assumed that nutrient leakage, especially nitrogen from agriculture, was one of the causes for this algae bloom.

In 1984 the authorities initiated an extensive research program to find measures to reduce non-point source pollution and point pollution from agricultural areas. The research program finished in 1989 (Rognerud, 1989). One project in the program was the development of a mathematical model that (1) predict non-point source pollution and (2) evaluate the effect of changes in agricultural management on pollution. The model should be a tool for local and regional agricultural and environmental service authorities to select optimal management. The model project is not finished yet and in this paper provisional results are presented.

MODEL SELECTION

It has not been the intention to develop a complete new model because of lack of resources. In Norway up to 40 percent of yearly runoff may be related to a snowmelt period of about 2 weeks (this study). The ice content and soil structure will influence both the amount of surface runoff during snowmelt and soil erodibility. These factors depend on the snow cover and the frost regime during the foregoing winter (Lundin, 1989). This will demand simulation of snow dynamics based on at least daily values for precipitation and temperature. In the Norwegian grain districts the soil doesn not have a plant cover from late autumn till after snowmelt in April. During this period the soil is greatly exposed for erosion by autumn rainstorms and snow melt. A list of desired model qualities has been formulated and existing models are selected from the literature (Decoursey,1985; El Kadi and van der Meijde, 1983; Haan et al., 1982; Rose, 1985) for testing (Table 1). SWAM, ARS Small Watershed Model (DeCoursey, 1982) has not been tested up until now because an improved version of the model, OPUS, was expected in 1988 (DeCoursey, pers. comm.). CREAMS (Knisel, 1980) and derived models are probably the most frequently used models for estimating non-point source pollution. For the years 1979-1982 the CREAMS model and the GAMES model (Cook et al.,1985; Rosseau et al.,1987) have been applied to the Årungen watershed situated in south-eastern Norway. The main lake in the watershed is heavily eutrophieted and it is well studied (Grøterud et al.,1981). Simulation results given by Sannes and Seip(1984) and Seip(1984) were correct for some years but wrong for other years when compared with the available measured values. Seip(1984) mentions that the measured data may not have been as accurate as is necessary to precisely evaluate model performance. With the models it was not possible to correctly simulate selected episodes, as for exemple snowmelt. Changes made in the snowmelt routine of CREAMS gave better results (Kalgraf,1986). Nevertheless, as long as monthly mean temperatures are used as input it can not be expected that the model will precisely simulate snow dynamics. Although CREAMS does not simulate soil temperature it does includes one of the best routines for erosion available.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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TABLE 1. Characteristics of hydrology/erosion models selected from literature for priliminary testing under Norwegian climatic conditions.

desired model characteristics	MODEL			
	CREAMS	SWAM	SOIL	GAMES
simulate: hydrology	yes +	yes +	yes ++	yes-
erosion	yes ++	yes +	no	yes +
soil temperature	no	no	yes +	no
snow dynamics	no	no	yes +	no
Physically based	+/-	+/-	++	+/-
Input easy to quantify	+	+/-	+	+
agricultural management input	++	+	no	+
heterogeneous areas, more then one field	yes -	+	no	yes +
easy to apply	+	+/-	+/-	+
runs on Personal Computor	yes	yes	yes	yes
documentation available	yes +	yes -	yes -	yes +

The advantage of the GAMES model is the possibility to route runoff from one field to an other. The erosion routine in GAMES is the USLE and therefore comparable with CREAMS. A limited test of the model did not give satisfactory results, but this may be caused by uncertainties in the input variables.

The SOIL model (Jansson, 1988; Jansson and Haldin, 1980) developed at the Swedish University of Agricultural Sciences is a hydrology and energy model with routines for snow dynamics based on daily temperature and precipitation as input. The model simulates the soil water balance quite well (Jansson and Gustafson, 1987; Gustafson, 1987); however, rutines for soil freezing and thawing and its effect on infiltration can be improved (Lundin, 1989). The existance of macro pores and by-pass flow are other details in the model that can be improved by methods developed by Jarvis and Leeds-Harrison(1986) or Bronswijk(1988).

Based on the preliminary test of CREAMS and the qualities of SOIL it was decided to build a new model where the hydrology part of CREAMS is replaced by the SOIL model. At a later stage it will be tried to incorporate the routing routines from GAMES to simulated larger heterogeneous areas.

MODEL DESCRIPTION

To understand the SOIL/CREAMS model a short description of the two contributing models and the way they are connected is given.

SOIL was originally designed to predict annual soil climate for biological purposes within the Swedish Coniferous Forest Project. The model is physically based and general enough to predict annual water and heat dynamics, including frost, for a variety of layered, unsaturated soils and vegetation covers. Soil boundery conditions are supplied by submodels of snow dynamics, precipitation interception, evapotranspiration, root water uptake and net horizontal groundwater flow. Given a measured soil surface temperature, the model can simulate soil temperature and heat flow variations within the day. For long-term simulations driving variables include daily means or sums of air temperature, precipitation, relative humidity and windspeed. Evapotranspiration is predicted from input of either net or global radiation, or cloudiness, or duration of bright sunshine. Soil properties are determined by independent methods but most surface characteristics must be determined by optimization. The most difficult and important parameter in the hydrology model surely is saturated conductivity. Because of variation in time and space it is not posible to get a precise estimate for it for a field. In this study the input parameters KSATC (saturated conductivity) KSATCT (saturated conductivity including the effect of macro pores) and groundwater flow are used for calibration. Mass and heat balanse schemes for the model are illustrated in Figure 1.

The two partial differential equations for heat and water flow are solved with an explicit forward differencing method (Euler integration). This requires the soil profile to be approximated with a discrete number of internally homogenous layers. Slowly changing state variables are updated at larger time-steps, and integration time-step varies dynamically as a function of conditions (including frost occurance) during simulation to minimize execution times. Water flow rates into the two top soil layers are used as tests. The version used in this study included neither improvements for hydrolic conductivity at frozen conditions as given by Lundin(1989) nor improvements for bypass flow.

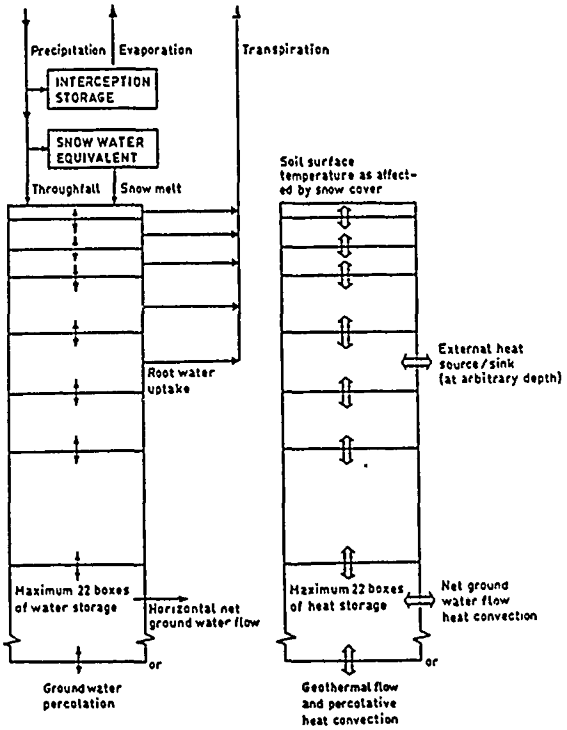

FIGURE 1. Mass and heat balance schemes of the SOIL model (from Jansson et al.,1980).

CREAMS is a field scale model for Chemicals, Runoff, and Erosion from Agricultural Management Systems. The version of the model used in this study consists of three seperate modules which treat respectively hydrology, erosion and loss of nutrients (P and N) and pesticides. A newer version of the model, GLEAMS, is now available where the three separate modules are intergrated and routines for nutrient leakage to the groundwater have been improved (Knisel et al.,1987). In this study, the hydrology module has been replaced by the SOIL-model described above. The erosion component of the model is a modification of the USLE(Foster, 1981), an emperical formula derived from plot data that employs physically related factors. Total soil loss is calculated as the maximum sediment yield from the erosion proces or from maximum sediment transport capacity of the overland flow in case erosion exceeds transport capacity. The calculations are made seperately for five soil particle size classes. Losses of chemicals are calculated as a sum of losses by runoff, sediment transport, and leaching. A sensitivity test of the CREAMS' erosion part by Lane and Ferreira(1980) showed that changes in the USLE factors for soil erodibility (K), soil loss ratio(C) and contouring factor(P) have a linear effect on erosion while the effect of Manning's n is larger than the relative changes in the parameter.

Programs have been developed to combine the two models with differences in variable definitions, input/output formats and units. Necessary input variables for simulation of erosion as excess rainfall rate and erosion index for a given storm have been calculated following formules given by Foster et al.(1981). A flow chart of the model is given in Figure 2. The model is run on an IBM-PC in batch mode.

FIELD DESCRIPTION

For the national research program, experimental watersheds have been selected for studies of agricultural practices, and measurements of surface runoff, erosion and nutrient losses (Øygarden, 1989). At each watershed homogeneous fields ranging 0.5-3.0 ha were selected for detailed studies. In this study data are used from the watershed situated 40 km north of Oslo. Marine clay and silt are the dominating soil types in the area. In the last two decades most of the original ravine landscape has been destroyed by leveling. This has reduced hill slopes and increased slope lenghts. In the same period a development from mixed small-scale agriculture to predominally large-scale grain produc-

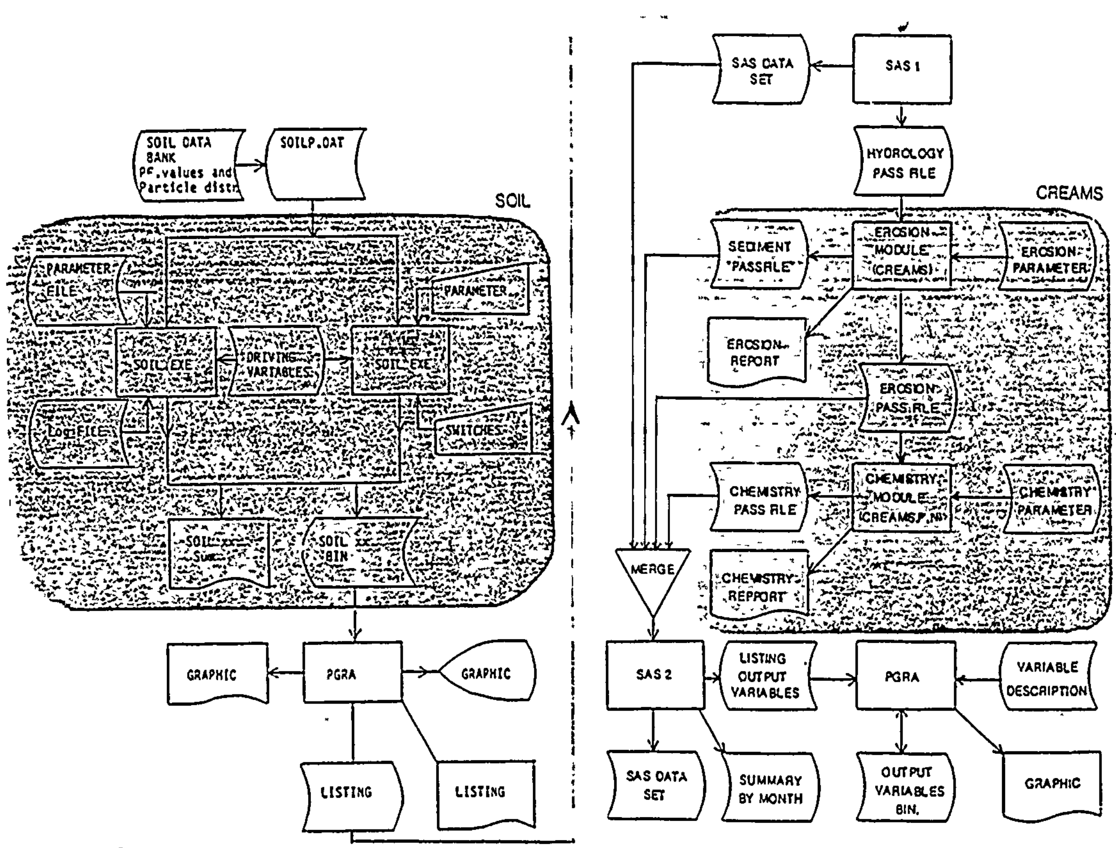

FIGURE 2. Flow chart of the SOIL-CREAMS model as it runs on IBM-PC in batch mode.

TABLE 2. Sumary of climatic data for the test fields for the years 1987, 1988 and the 30-year normal.

	Temperature °C			Precipitation (mm)		
	1987	1988	1931-60	1987	1988	1931-60
January	-14.2	-0.7	-6.9	23	135	58
February	-7.8	-3.1	-6.3	41	115	41
March	-6.6	-3.4	-2.3	91	123	30
April	4.2	1.8	3.2	27	53	50
May	8.2	10.7	9.4	49	31	53
June	11.3	17.6	13.6	176	60	79
July	15.8	15.5	16.0	46	141	92
August	12.1	13.9	14.6	82	150	95
September	8.6	11.1	10.0	133	131	87
October	6.3	3.6	4.5	213	79	85
November	-1.3	-3.8	-0.6	116	31	82
December	-5.6	-4.1	-3.9	23	56	73

tion has taken place. These changes have contributed to an increase of environmental pollution, especially as a result of erosion. A summary of climatic characteristics of the area are given in Table 2.

Daily values for surface runoff (m^3), concentration of phosphorus, nitrogen and pottassium (μg/l) and soil loss (mg/l) are calculated from logged data both for the whole watershed and for the small fields within the watershed. A discussion of measurements methods is outside the scope of this paper and is discussed elsewhere (Øygarden et al., in prep). In this study data are used from 2 fields in the watershed for the years 1987 and 1988. A list of the field characteristics is given in Table 3.

TABLE 3. Characteristics of field A and B used for testing the model.

Field	area (ha)	soil type	slope (%)	slope lenght (m)	drain distance (m)
A[1]	0.86	clay/clay loam silt	4-8	155	4
B	0.44	silt loam	6-16	115	7

1)leveled.

MODEL APPLICATION

In the first test of the model described here it was decided to run the hydrology and erosion module separately. The module for chemicals has not been tested yet. The model has been calibrated against measured data from 1987 and validated against those from 1988. A summary of the relevant observed data is given in Figure 3. The relation between surface runoff and soil loss is not constant due to soil cultivation during a year (Figure 4). Differences between the years can be explained by changes in agricultural management and the difference in the distribution of yearly precipitation. In the spring of 1988 in field B, a grass vegetation was established where runoff concentrates. Differences between the fields are caused by differences in soil type and agricultural practices. It may also be result of field size and the difficulty to estimate exactly the area that contributes to the surface runoff sampled. The observed data are discussed in more detail by Øygarden (1989).

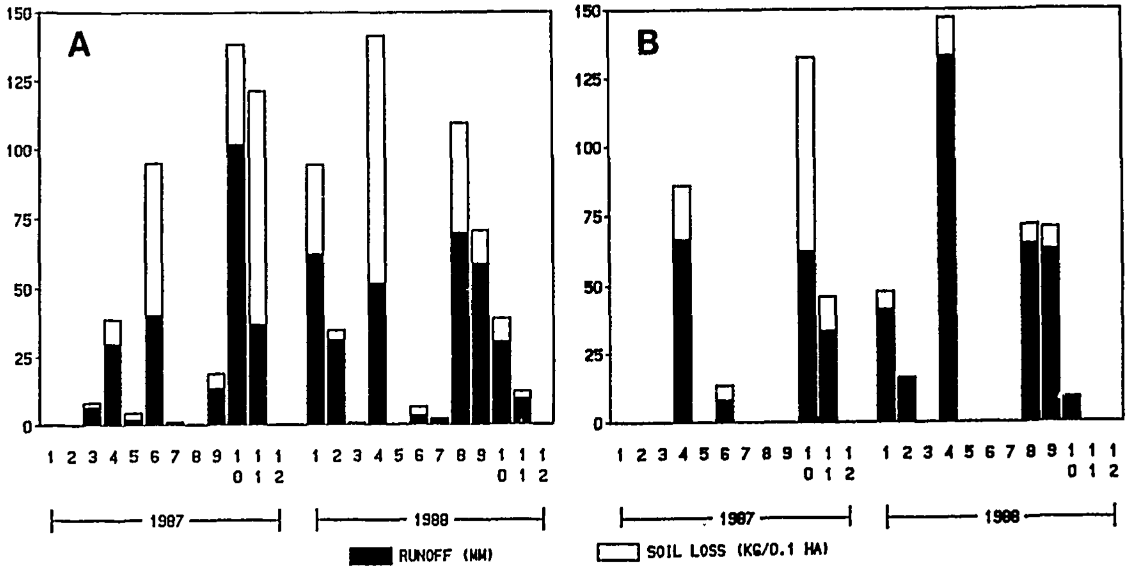

FIGURE 3. Monthly sums of observed surface runoff (mm) and soil loss (kg/0.1 ha) for field A and B, 1987 and 1988.

Hydrology module (SOIL)

For 1987 runc"' could be simulated satisfactory for field A and B (Fig. 5). It was necessary to increase KSATCT during the snowmelt period together with a high value for groundwater flow. This can be explained by accepting a high bypass flow during snowmelt. The figures for temperature (Table 2) show that the winter of 1987 had a continued period of frost existed from January til snowmelt, while in 1988 the winter was mild in the beginning. For this mild period with alternating frost and thaw the model underestimates runoff, resulting in a delay til snowmelt in April. This shows that an overestimation of snow accumulations has taken place. The runoff at the end of August is underestimated too. This is because in 1987 there wasn't runoff in August and the model isn't calibrated for that period of the year. From the results sofar it is concluded that replacing the original hydrology module of CREAMS by SOIL gives improved estimates for runoff.

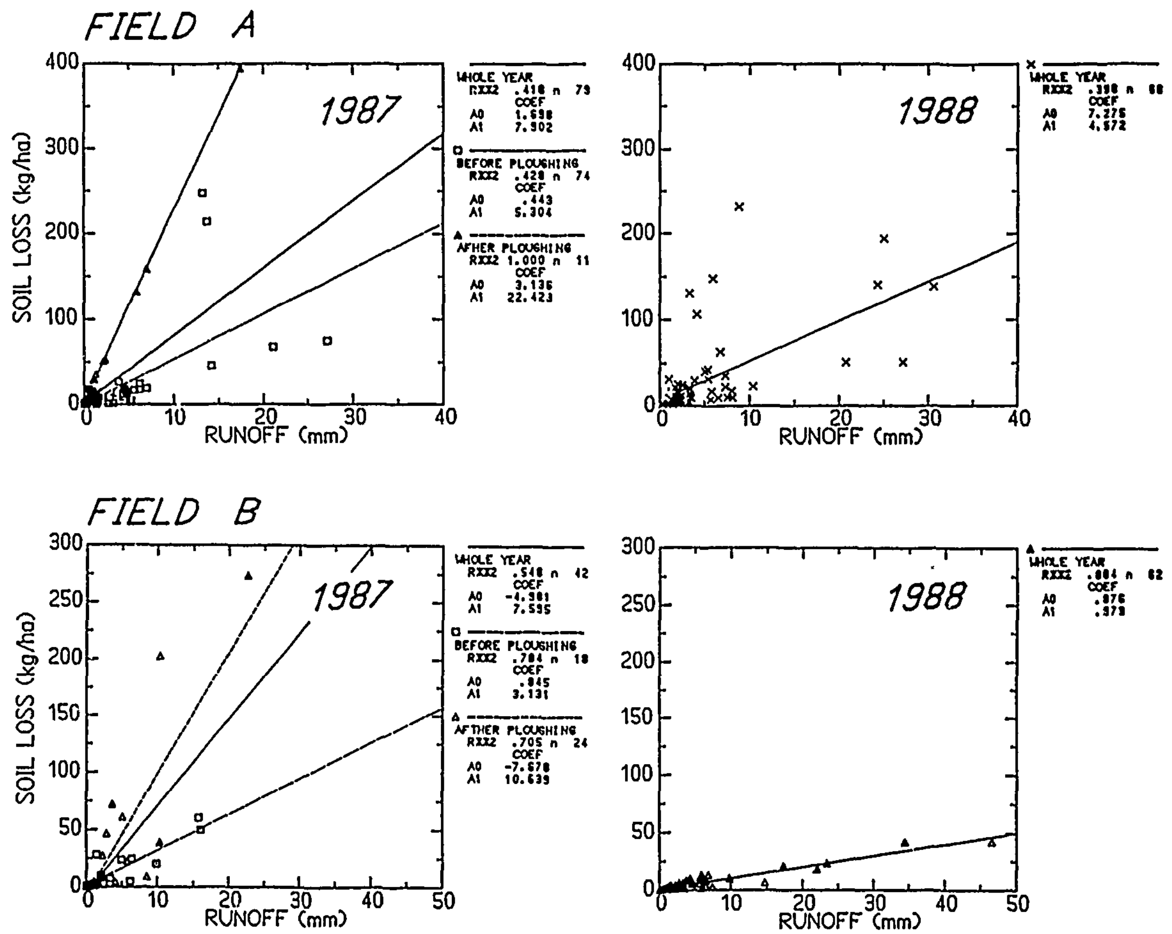

FIGURE 4. Correlation between observed runoff and soil loss for field A and B, 1987 and 1988.

<u>Erosion module</u>

A first run of the erosion module gave a sediment yield about 10 times higher than observed. The parameter values used were based on the information given in the CREAMS manual(US.Dep. of Agr.,1984). In Figure 4 it is shown that the relation between runoff and soil loss depends on the period of the year. Calibration is then only possible with the time dependend parameters C (soil loss ratio for overland flow), P (contouring factor) or N (mannings' n for overland flow). As stated by Lane and Ferreira(1980) the effect of C and P on soil loss has a linear form. To reduce soil loss by a factor of 10 demands very low values for C (0.5-5%), when the lowest value for P is 0.5. Such low values for C are no longer related to the cultural practices on the fields. It was therefore decided to use N as a calibration parameter for the erosion module. The value of N was changed depending on the date of soil cultivation and the presence of snow cover. Ploughing normally takes place in autumn just before soil frost is established. The seed bed is prepared shortly after the snow has disappeared. Simulation results are presented in Figure 6. Because of a minor time delay in the simulated output, the regression analysises are based on 5-day sums of soil loss. Figure 7 shows that after snowmelt, N can be reduced by about 50% for field A and B (1987). During the rest of the year best results are achieved by changing N according to date of harvesting and ploughing. The difference between field A and B is caused by an accepted higher soil erodibility for the silt loam soil type in field B. It was not possible to simulate soil loss in 1988 satisfactory without changing N. In 1987 erosion took place during snow melt and in June and October. In 1988 erosion occurs mainly in January and in August (Fig 6). These periods where not calibrated with the erosion events that took place in 1987.

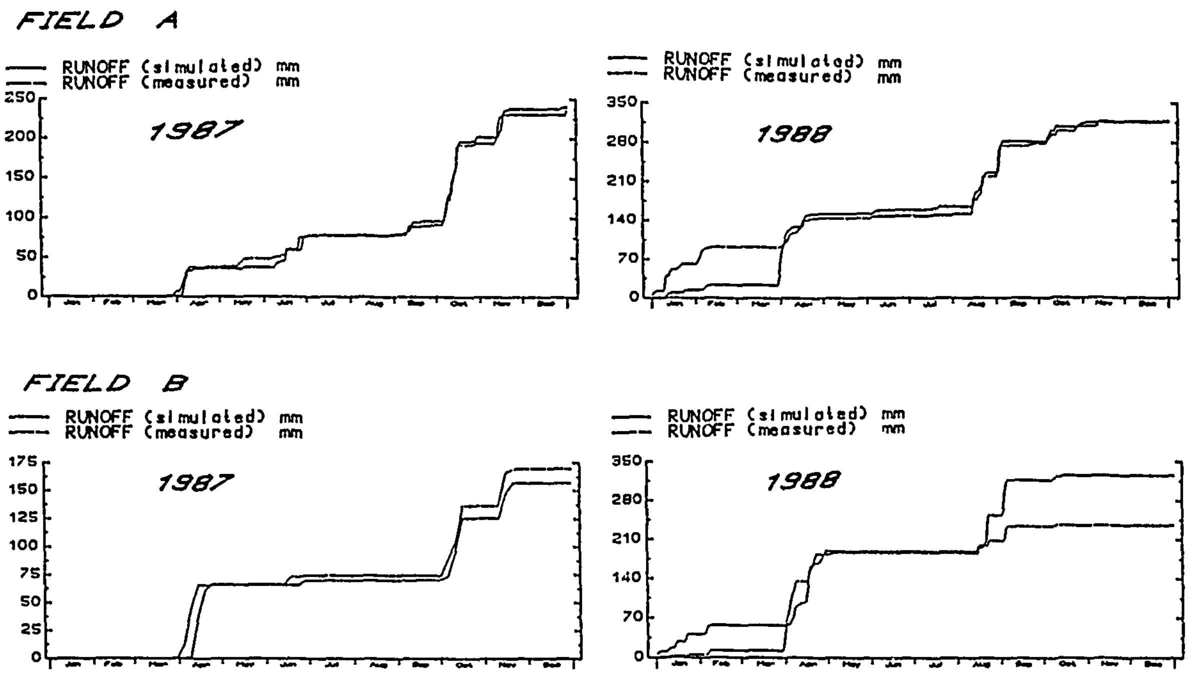

FIGURE 5. Observed and simulated accumulated runoff (mm/day) for field A and B, 1987 and 1988. Correlation and regression coefficients between measured and observed values:
Field A:1987:R^2=0.50, a=0.67, n=364 ; 1988:R^2=0.47, a=0.67, n=364
Field B:1987:R^2=0.47, a=0.63, n=74 ; 1988:R^2=0.18, a=0.36, n=74

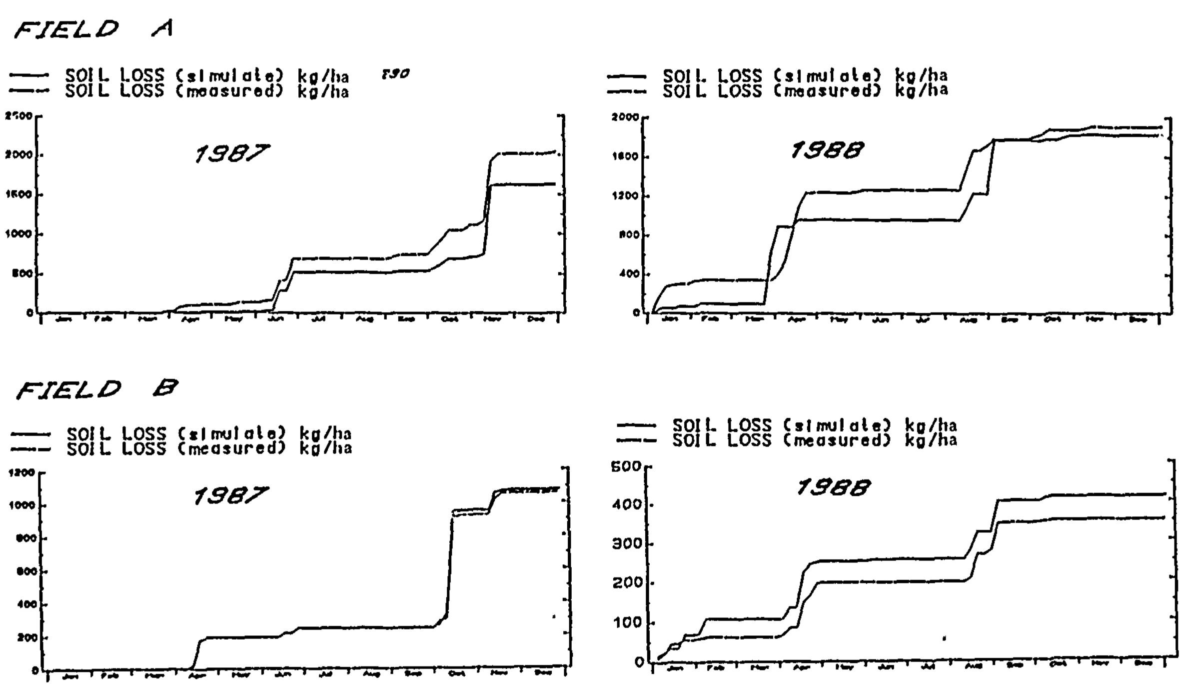

FIGURE 6. Observed and simulated accumulated soil loss (kg/ha) for field A and B, 1987 and 1988. Correlation and regression coefficients between measured and observed values(based on 5-days sum):
Field A: 1987:R2=0.97, a=1.08, n=73; 1988:R^2=0.04, a= 0.30, n=73.
Field B: 1987:R2=1.00, a=0.98, n=73; 1988:R^2=0.78, a= 1.06, n=73.

DISCUSSION

The model approach used in this study gives reasonable results for surface runoff and soil loss. The SOIL model gives better estimates of surface runoff than the hydrology modul of CREAMS does because of better routines for snow dynamics. A satisfactory simulation of runoff will result in good estimates of soil loss. The data presented show that the time of the year precipitation falls strongly determines the amount of runoff. The relation between runoff and soil loss has been shown to depend on soil erodibility at the time of runoff. This soil erodibility is the result of foregoing precipitation, temperature and soil cultivation. The effect of the temperature regime on ice in the soil is described by Lundin (1990). The interaction between all these factors of which precipitation and temperature are stochastic, makes the relation between runoff and erosion complicated. It is shown that a by-pass flow has been important during the snowmelt in 1987. In the model, by-pass flow is simulated by a conductivity parameter that includes the contribution of macropores. This parameter has been used in calibration. Saturated unfrozen soil on top of frozen soil was not observed during snowmelt in 1987 og 1988. In such case a macropore system would not exist and the result is an increase in runoff.

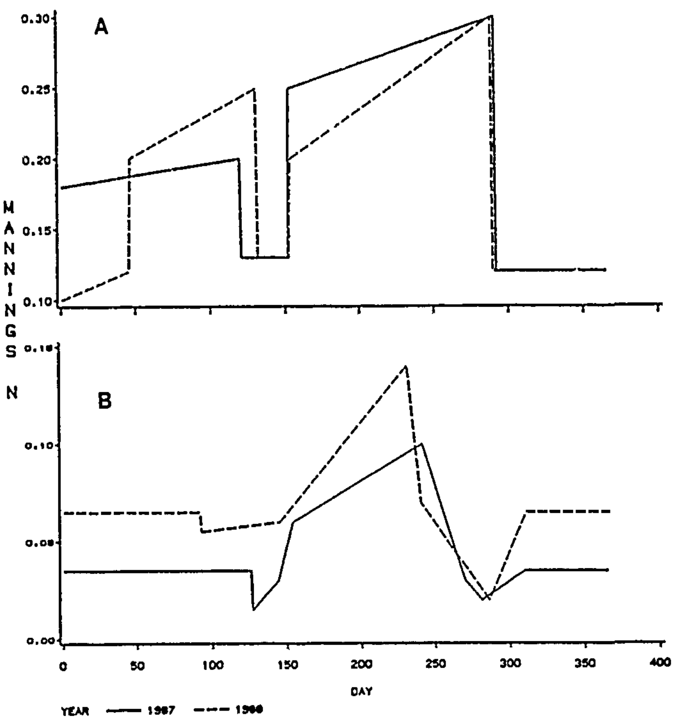

FIGURE 7. Values of calibration parameter N (mannings'n) in relation to snowmelt and soil cultivation.

The effect of frozen soil on erosion during snowmelt could in this study be simulated by a higher value of N (Mannings'n for overland flow). Physically this means that the running water meets a higher surface resistance. Possible explainations for that are 1: the micro relief of frozen soil is not so easy leveled out by the water or 2: the snow cover gives resistance to the running water. As mentioned, during the winters of 1987 and 1988 no saturated unfrozen soil was observed on top of an impermeable frozen soil. Such a situation would certainly have resulted in a higher soil loss per mm runoff, because of a higher erodibility. Therefore, the way the effect of frozen soil has been treated in this study may give incorrect results in other situations. The way Mannigs N has been used for model calibration doesn't have any relation with the original physical meaning of N.

To be able to simulate a situation as described above, it is necessary to have a combined temperature-hydrology model with a short time step as SOIL has. Depth of soil frost on a given day will be determined by the daily combination of air temperature, precipitation and snow cover during the foregoing days. To estimate actual soil erodibility, the simulated daily temperature and water content in the top soil layers should be input variables for the erosion model. In this way it may be possible to calculate a soil erodibility factor based on soil type (particle size distribution), soil temperature and soil water content. Such a soil erodibility factor is not known from the literature. The presented values of observed soil loss (Fig 3 and 4) show that erosion depends on an combination of circumstances during one precipitation event. To evaluate measures to reduce erosion, soil loss has to be estimated as a one event phenomenon, not as a process based on yearly or monthly averages of input variables.

The model desired by the authorities should be a tool for local and regional services to advise individual farmers and regional authorities what to do to reduce soil and nutrient losses from agricultural fields. On that scale a model with a daily output is necessary. Possibly, their further development of the model presented can be that tool.

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EFFECT OF FREEZE-THAW CYCLES ON THE PERMEABILITY AND MACROSTRUCTURE OF SOILS

E. Chamberlain,* I. Iskandar* and S.E. Hunsicker†

INTRODUCTION

Hazardous waste treatment and disposal is one of the major environmental concerns. In the United States alone, about 50 million tons of hazardous waste is produced each year. Clay liners and clay caps are commonly recommended and used for containing and covering hazardous and toxic waste as well as solid municipal waste. The purpose of the liners is to impede the flow of contaminants to ground water and to sorb the chemicals, thus protecting the ground water from contamination. The purpose of the caps is to prevent water infiltration into the contaminated soil and the release of toxic gases.

Chamberlain and Blouin (1978) and Chamberlain and Gow (1979) found that the permeability and structure of fine-grained dredged material and soil were greatly changed by freezing and thawing. They reported reductions in void ratio and large increases in vertical permeability. They attributed these changes to particle reorientation and reduction of particle spacing due to thinning of water films. They concluded that the permeability increased because of the formation of polygonal shrinkage cracks and/or the reduction of the volume of fines and the associated bound water in the void spaces separating the coarse fraction. The structural changes were attributed to the development of high soil moisture suctions in the freezing zone. This causes an increase in the effective stress in the region immediately below the freezing front and the formation of segregated ice lenses. In clay soils, this results in a shrinkage of the soil structure, the formation of cracks and a change in the pore size characteristics. All engineering properties of soils containing significant amounts of clay particles are affected by this process(Chamberlain 1989)

The objective of this study was to investigate the effect of freeze-thaw cycling on the permeability and structure of compacted clay soils used as caps or barriers for containing hazardous waste materials.

MATERIALS AND METHODS

Apparatus

All tests were conducted in the consolidometer described in detail elsewhere (Chamberlain and Gow 1979). A schematic is shown in Figure 1. The consolidometer consists of a Teflon-lined Plexiglas cylinder with outer and inner diameters of 15.24 cm and 6.35 cm, respectively. The base and the piston are made from stainless steel, each containing a glycol cooling chamber, a thermoelectric cooling plate and a porous drainage plate. The piston, which is guided by linear bearings, has a loading plate mounted on top on which lead weights are placed for simulating surcharge and for conducting consolidation tests. The drainage ports in the base and the piston lead to a valving system (Fig. 2). This system allows a constant head water supply to be connected to the test sample during the consolidation and freezing and thawing phases, and for falling head and constant head devices to be connected to the sample during the permeability tests. A rotary pump is linked to the valving system to pump the water manually to either the constant or variable head devices. A de-aerator is directly connected to the system to provide de-aired-distilled water.

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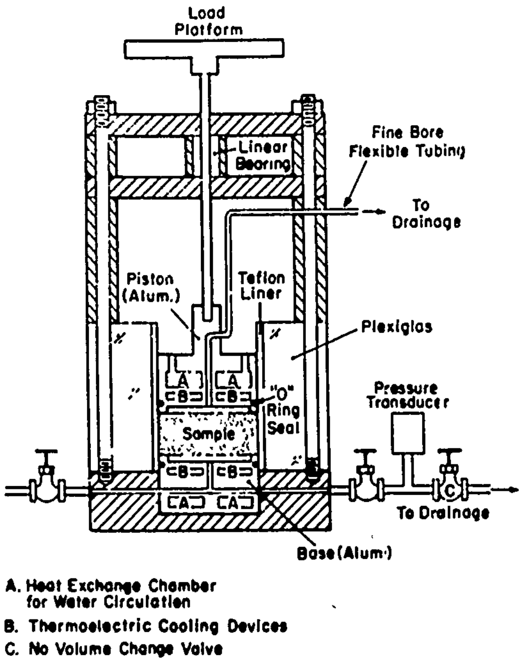


Figure 1. Schematic of the freeze-thaw permeameter.

During freezing, ethylene glycol from a refrigerated bath is circulated into the chamber in the base to the chamber in the piston and back to the cooling bath. This allows the control of the soil temperature to a fixed ambient temperature (usually near 0°C). The thermoelectric freezing plates in the base and piston allow precise control of the temperature during freezing. Two thermistors, one mounted on each of the porous plates at the top and bottom sample contact surfaces, are used to measure the boundary temperatures. A computer-controlled data logger reads the temperatures. Adjustments are made automatically through a computer-controlled power supply connected to the thermoelectric plates to obtain the desired boundary temperatures. This system allows precise maintenance of the temperature gradient and the rate of freezing.

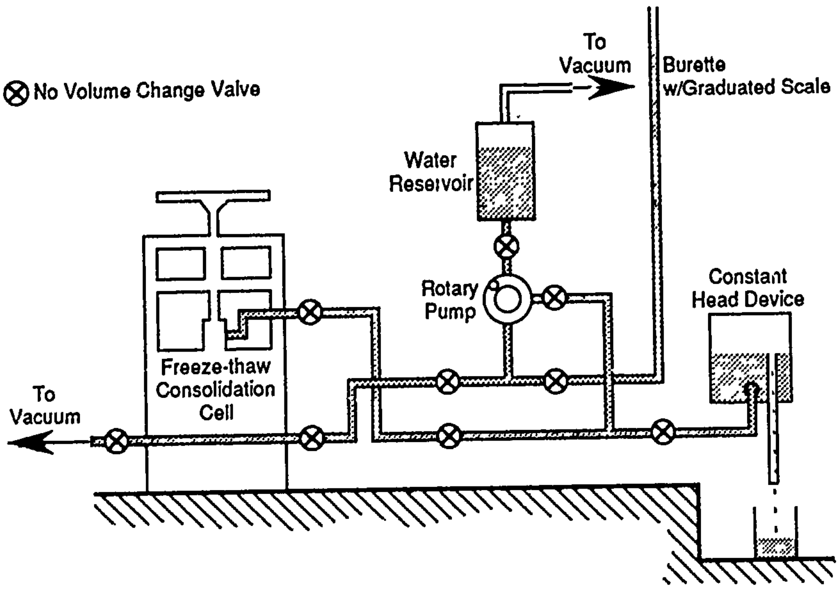


Figure 2. Schematic of the permeability system.

Sample Properties and Characteristics

One of the soil samples used in this stud' is Fort Edwards silt-loam-clay. It has been recently used as a barrier cover for a hazardous waste site near Fort Edwards, New York. The other four soils were being used by the U.S. Department of Energy (DOE) Uranium Mill Tailings Remedial Action (UMTRA) Projects Office as caps for processed uranium mill tailings in the western United States. The purpose of these barriers is to control radon gas release. The soil texture of the four UMTRA samples ranges from loam to clay loam.

Particle size distributions for the five soils studied were determined using the Department of the Army, Office of the Chief of Engineers, Engineering Manual (EM-1110-2-1906). Atterberg limits (liquid limit, plastic limit and shrinkage limit) were also determined using the same standards. Figure 3 shows the gradation curves and Table 1 summarizes the index properties of the five soils. Both engineering and agricultural soil classifications are given.

Sample Preparation

Both compacted and consolidated samples were prepared with the Fort Edwards soil. Only compacted samples were prepared for the UMTRA soils. The UMTRA soils were provided by the DOE ready for compaction. They were compacted in the consolidometer using a Harvard miniature compactor with a 12.7-mm-diameter tamping foot and a 40-lb (178 N) rated spring. The nominal length of each sample was 25 mm and the diameter 62.5 mm. The samples were compacted according to DOE specifications to the design water contents and densities. Generally, 40 to 50 tamps for each of two layers were employed. Details for the compaction of the UMTRA soils are given in Chamberlain (1989).

The Fort Edwards samples were compacted with a much higher compactive effort to reach in-place conditions. The soil at a natural water content of 13% was passed through a sieve with 5-mm openings to remove any larger particles and to help break down the aggregates. The water content was then adjusted to 25% and the mixture was allowed to stabilize for 24 hours. The samples were compacted in the consolidometer in two layers with a special hammer with a 35-mm-diameter foot. Collar spacers on the hammer restricted compaction to the specified density. Details of the compaction of the Fort Edwards soil are given by Hunsicker (1987).

The consolidated Fort Edwards samples were prepared from a degassed slurry with an initial water content about twice the liquid limit. Loading was applied in doubling increments to the desired stress much like in a standard consolidation test. Details are given by Hunsicker (1987).

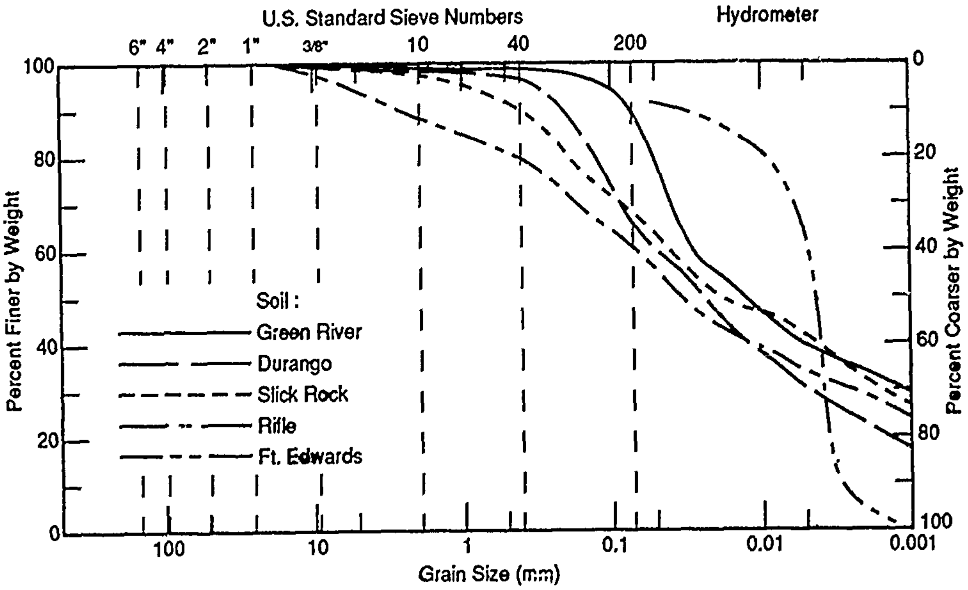


Figure 3. Particle size distribution curves for the test soils.

Permeability

Falling head permeability tests were conducted before freezing and after selected cycles of freezing and thawing. The permeability tests were conducted by allowing de-gassed water to flow from the graduated burette (Fig. 2) through the sample base and out from the top piston into the constant head apparatus. The procedure generally follows that outlined in the EM 1110-2-1906 (1970). The initial upper head in the burette was about 60 cm above the lower constant head. The lower head at the outflow was about 1 cm above the top of the soil sample. Because the compacted samples were initially unsaturated, de-aired water was allowed to flow through them until equal rates of inflow and outflow were achieved. The permeability readings for all samples were made only when this condition was satisfied. The permeability observations after freeze-thaw cycling were made after 400 minutes of flow or after 6 ml of water passed through the sample, whichever occurred first. All of the permeability tests were conducted at room temperature (24°C) and calculations were adjusted to 20°C.

Freeze-Thaw Cycling

The samples were frozen from the bottom up with free access to water at the top. A surcharge of 14 kPa was placed on the compacted samples to simulate 0.6 m of overburden. The consolidated samples were frozen in the normally consolidated state with the applied load in place. A frost penetration rate of 8 cm/day and a temperature gradient of 0.022°C/mm were normally used. The boundary temperatures were programmed to maintain this rate and gradient for 10 hours. During thawing, the bottom temperature was held at the freezing temperature of the soil water and the top temperature at 0.6°C higher for two hours. Thus, two freeze-thaw cycles were imposed per day. The freeze-thaw cycling was then repeated to the prescribed number of repetitions. A total of 15 cycles were imposed on the UMTRA samples. The compacted Fort Edwards samples were normally frozen and thawed six times and the consolidated samples three times. Before conducting a permeability test, the temperature of the sample was raised to room temperature. After the last permeability test was run, each sample was frozen once more to prepare it for the final moisture content and density determinations and for the thin section analysis.

Thin Section Sample Preparation

The frozen sample was removed from the cell and cut vertically in half with a band saw in a coldroom. A vertical slice and two horizontal slices were then made, each approximately 2.5 mm thick. The slices were frozen to glass plates using supercooled water. The exposed surfaces were scraped until a few millimeters thickness of the soil sample remained. The thin sections were then placed on a light table and photographed with both back and front lighting.

RESULTS AND DISCUSSION

The particle size distribution curves (Fig. 3) and index properties (Table 1) show that the Fort Edwards soil is a silty clay with a Unified Soil Classification (USC) of CH. Engineers call this a highly plastic clay; soil scientists refer to it as a silt-clay-loam. The other four UMTRA soils are in a different class of soils with USC values of CL. These are called clays of low plasticity by engineers and are referred to as silt or clay-loams by soil scientists. The Atterberg limits (Table 1) show the significant differences in consistency between the Fort Edwards and the UMTRA soils.

Figure 4 summarizes the relationships observed between the number of freeze-thaw cycles and the permeability of the compacted Fort Edwards soil. We observed a wide range of permeabilities after freezing and thawing with no particular pattern of increasing or decreasing values. A null-hypothesis statistical analysis based on the Student-t test showed that permeability was <u>not</u> related to the number of freeze-thaw cycles. (1987) We did, however, observe increases in permeability after freezing and thawing of consolidated Fort Edwards soil (Fig. 5). In this case we saw increases in permeability greater than one order of magnitude when freeze-thaw cycling was initiated at an initial water content (or void ratio) just below the liquid limit

Table 1. Index properties of soils investigated.

Soil	Atterberg limits: Liquid limit (%)	Plastic limit (%)	Shrinkage limit (%)	Organic matter content (%)	Specific gravity	Optimum water content (%)	Optimum dry density (g/cm³)	U.S. Dept. Agri. class	Unified soil class
Fort Edwards	65	28	18	–	2.80	25.0	1.605	silt-loam	CH
Green River	24	14	12.6	0.48	2.69	13.0	1.896	silt-clay-loam	CL
Durango	38	14	11.2	< 0.01	2.70	15.9	1.793	loam	CL
Slick Rock	30	15	10.0	0.37	2.76	14.8	1.832	clay-loam	CL
Rifle	36	14	9.5	0.19	2.71	14.0	1.817	clay-loam	CL

(or its equivalent void ratio). Much smaller increases were observed as the initial water content before freezing approached the plastic limit. Straight lines drawn through the data in the log permeability vs void ratio plane (Fig. 5) appear to intercept at a void ratio just below the void ratio equivalent to the plastic limit.

Figure 6 summarizes the relationship between the number of freeze-thaw cycles and the permeability of the compacted soils from the UMTRA project. The test results show that, in contrast to observations for the compacted Fort Edwards soil, the permeabilities of these soils substantially increased during freeze-thaw cycling. The initial permeabilities of the four soils were all about 1×10^{-8}cm-sec^{-1} or less. The permeability of the Rifle soil increased to about 4×10^{-6}cm-sec^{-1}, a nearly 300-fold increase, after nine freeze-thaw cycles. The permeabilities of the other three soils increased by factors ranging from 35 to 50. In all cases the permeability increases were larger during the first few cycles, most of the changes occurring within nine freeze-thaw cycles.

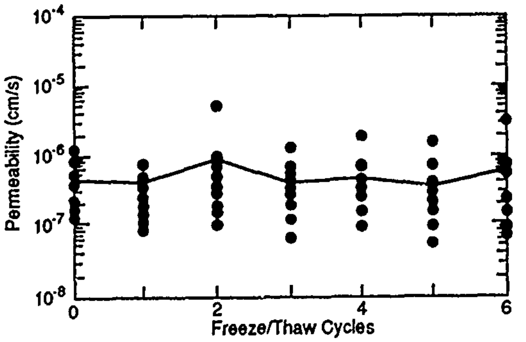


Figure 4. Permeability change with freeze-thaw cycling for compacted Fort Edwards soil (all tests).

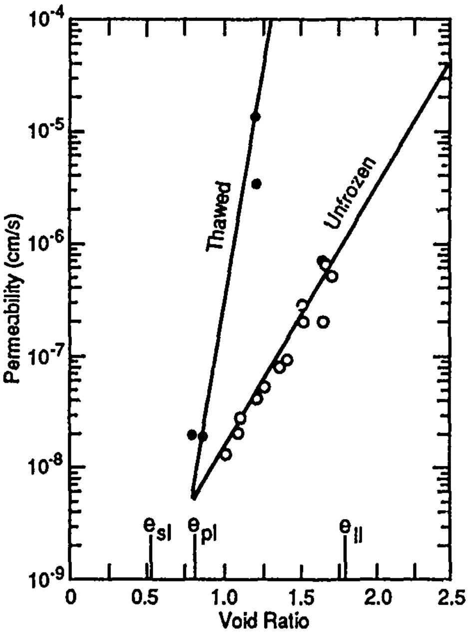


Figure 5. Summary of freeze-thaw permeability tests on cconsolidated Fort Edwards soil (subscripts sl, pl and ll refer to the shrinkage, plastic and liquid limits,respectively).

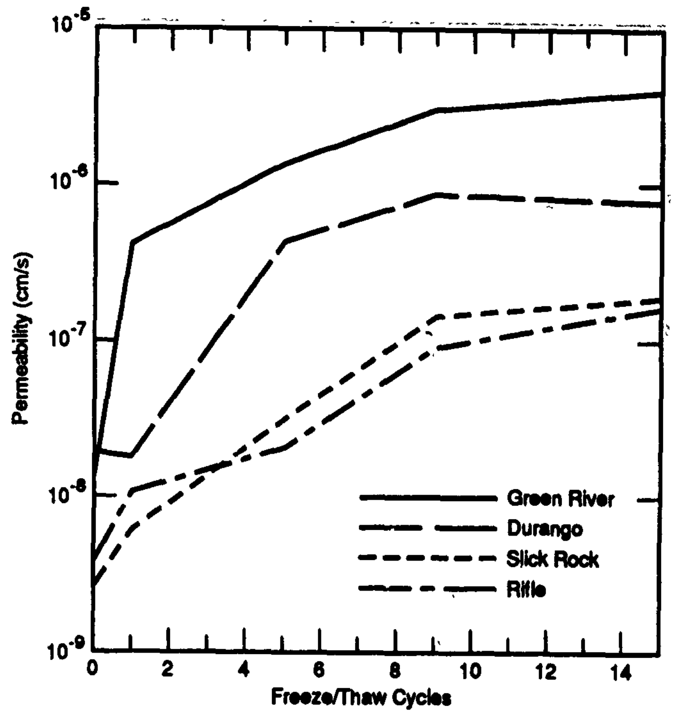
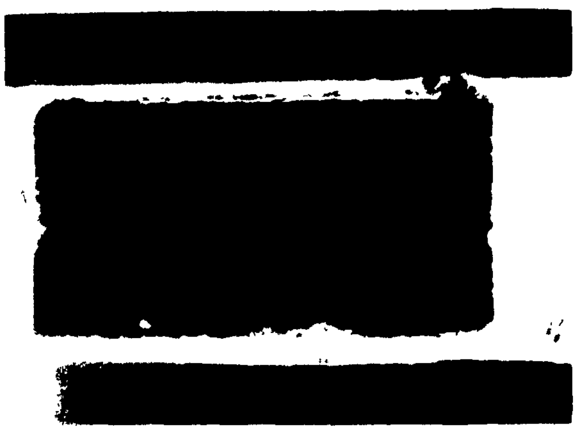


Figure 6. Summary of freeze-thaw permeability test results for the four compacted UMTRA soils.

The changes in permeability result from structural changes in the soil and from increases in the degree of saturation. Photographs of thin sections of a consolidated Fort Edwards sample (Fig. 7) show the soil-ice structure that developed during freezing. Both horizontal and vertical ice features are visible. These structures are also visible for the Green River and Durango soils, while only the horizontal ice features are visible for the Slick Rock and Rifle soils. There are also structural changes that occur on a microscopic scale. Scanning electron

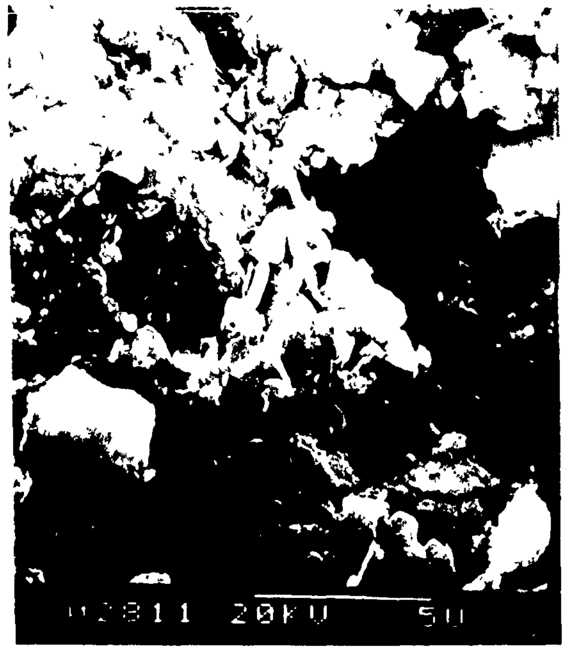
a) Horizontal plane.



b) Vertical plane.

Figure 7. Thin section of a consolidated UMTRA Durango sample after freeze-thaw Cycling.

a) Before freezing.



b) After freeze-thaw cycling.

Figure 8. Horizontal views of a consolidated Fort Edwards sample at 6000X magnification.

microscopic images (Fig. 8) of samples of Fort Edwards soil before and after freezing show major changes in the soil microstructure. Large voids appear where there was previously a relative homogeneous structure. The macroscopic cracks and microscopic voids created by freezing provide paths of reduced flow resistance and thus increase the permeability.

The permeability also increases because of increases in the degree of saturation. The initial degrees of saturation for the compacted test samples ranged from 65 to 93% (Table 2). Saturation procedures with de-gassed water increased the degree of saturation to above 90% (Table 3). Full saturation by back-pressuring was not attempted because of concerns for structural changes and time constraints. However, after a succession of freeze-thaw cycles and permeability tests the degree of saturations approached 96 to 98%. Figure 6 shows that the saturated permeabilities (obtained from UMTRA) are above the initial partially saturated values for the four UMTRA soils, but well below the permeabilities achieved after freeze-thaw cycling. This, however, does not fully explain why the permeabilities of the compacted UMTRA soils increased after freeze-thaw cycling and the permeability of the compacted Fort Edwards clay did not.

Table 2. Sample Properties after Preparation.

Soil	*Compaction Method*	*Water Content (%) (W_c)*	*Dry Density (g/cm³) (D_d)*	*Void Ratio (ρ_c)*	*Degree of Saturation (%) (S_c)*
Fort Edwards	Space hammer	~25	~1.60	~0.75	~93
Fort Edwards	Consolidation	~130	~0.60	~3.64	~100
Green River	Harvard minature	12.6	1.900	0.416	81.8
Durango	Harvard minature	15.5	1.828	0.477	87.7
Slick Rock	Harvard minature	14.1	1.826	0.512	76.2
Rifle	Harvard minature	14.0	1.708	0.587	65.3

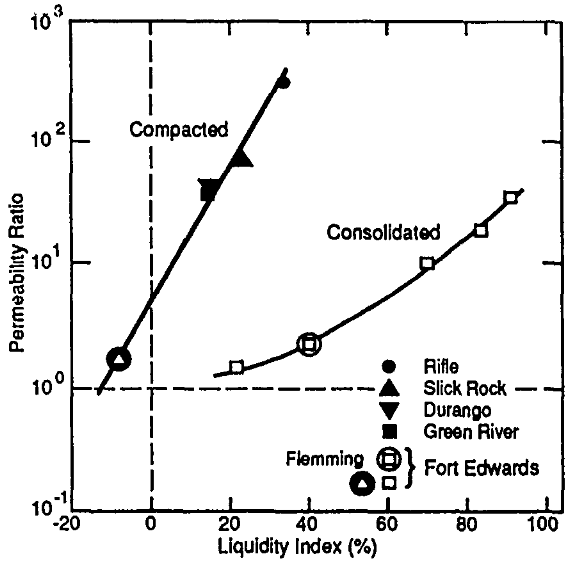


Figure 9. Effect of freeze-thaw cycling on the permeability of compacted and consolidated soils in the permeability vs void ratio plane.

A better understanding of the changes caused by freeze-thaw cycling can be made by examining Figure 9, where the permeability is plotted vs void ratio for both the compacted and consolidated Fort Edwards soil. Test sample S5 was consolidated from a slurry in load-doubling increments to a void ratio of about 1.0 and frozen and thawed three times. Permeability tests were made after every load increment and freeze-thaw cycle. Test sample S6 was consolidated to a void ratio of only 1.7 and also subjected to three freeze-thaw cycles. After the third thaw, this sample was again consolidated and permeability tests continued. The result is the polygon in Figure 9 formed by the void ratio-permeability paths. The path taken during the consolidation of sample S5 represents the permeability for the unfrozen saturated condition. The path observed during consolidation after thawing of sample S6 is for the thaw consolidated condition. The paths taken during the freeze-thaw cycling of both samples show how the transition is made from one state to the other.

The results for the compacted Fort Edwards soil plot to the left of the thaw consolidated curve (Fig. 9). The permeability measurements made before freezing have a range of about one order of magnitude. After five freeze-thaw cycles, all the void ratios increased. The permeabilities both increased and decreased in an unpredictable manner. However, the locus of points for the compacted-thawed condition appears to approach the thaw-consolidation path for this soil. These results begin to make sense if we consider that the compacted-thawed material may have a similar structure to that of the thaw-consolidated material. We started the slurry consolidated tests with a highly flocculated and disaggregated material. Individual soil particles (or very small aggregates of particles) were in suspension. Freezing caused the formation of larger aggregates with a more dispersed structure. The discontinuities formed at aggregate boundaries during freezing formed paths of reduced flow resistance. Thus, the permeability increased, even though the void ratio decreased (density increased). In contrast, the compacted samples started with a dense aggregated and dispersed structure. Freeze-thaw cycling caused the void ratio to increase slightly, probably because of the expansion of water to ice. The permeability did not follow a regular pattern of increasing or decreasing after freeze-thaw cycling. This may have been caused by non-uniform compaction. It also may have been the result of the extreme sensitivity of permeability to void ratio for the thawed condition (Fig. 9). A change of void ratio of only 10% in this region results in a 1000% increase in permeability. Furthermore, differences from sample to sample in the percent of saturation may have also contributed to the problem. The permeability of soil is very sensitive to the degree of saturation.

Table 3. Test results.

Soil	Before freezing: Water content (w_o) (%)	Saturated water content (w_s) (%)	Liquidity index (LI) (%)	Degree of saturation (S_o) (%)	Void ratio (ρ_o)	Permeability (K_o) ($\times 10^{-7}$ cm/sec)	After freezing and thawing: Water content ($w_{f/t}$) (%)	Degree of saturation ($S_{f/t}$) (%)	Void ratio ($\rho_{f/t}$)	Permeability ($K_{f/t}$) ($\times 10^{-7}$ cm/sec)	Permeability ratio[2] (K_r)
Ft. Edwards (compacted)	> 25.0[3]	26.5	8.1	> 93.3[3]	0.75	2.10 to 13.0	~26 to ~36[3]	> 93.3[3]	0.78 to 1.09	0.62 to 104	1
Ft. Edwards (consolidated)	36.1	36.1	21.8	100	1.01	0.128	28.2	100	0.79	0.190	1.48
	53.9	53.9	70.1	100	1.51	3.27	43.6	100	1.22	33.1	10.12
	59.3	59.3	84.6	100	1.66	7.08	43.6	100	1.22	131.0	18.50
	61.8	61.8	91.3	100	1.73	5.12	36.8	100	1.03	182.0	35.54
Ft. Edwards (consolidated after Fleming)	39.6	39.6	40.4	100	10.10	0.194	32.3	100	0.90	0.445	2.29
Green River	> 12.6[3]	15.4	14.7	> 81.8[3]	0.416	0.038	15.60	97.5	0.427	1.58	41.4
Durango	> 15.5[3]	17.7	15.0	> 87.7[3]	0.477	0.180	19.7	96.9	0.562	7.60	42.2
Slick Rock	> 14.1[3]	18.5	23.5	> 76.2[3]	0.511	0.026	19.5	96.6	0.511	1.86	72.4
Rifle	> 14.0[3]	21.4	33.9	> 65.3[3]	0.581	0.131	19.6	96.4	0.552	39.79	303.3

1—$LI = [(w_s - w_{pe})/(w_{e1} - w_{pe})]\ 100$.
2—$K_r = K_{f/t}/K_o$.
3—Exact values not determined, but $S_c \le S_o \le S_{f/t}$ with $S_o \ge S_{f/t}$.

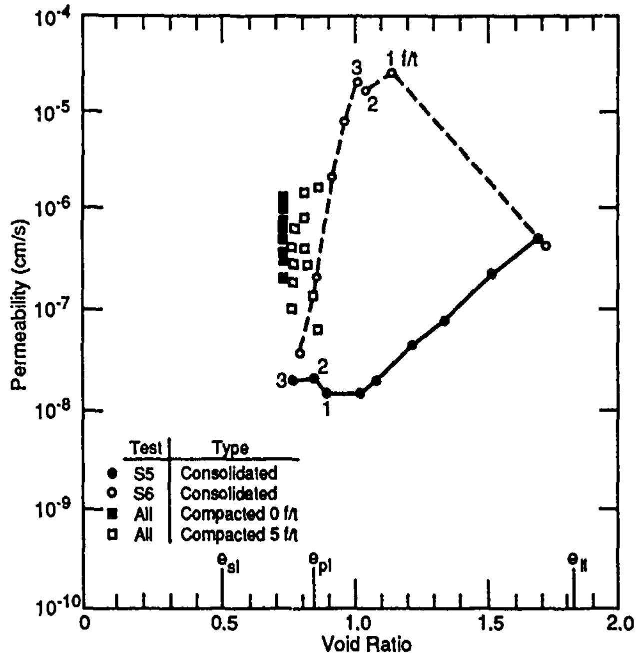


Figure 10. Effect of the liquidity index on the permeability ratio.

There is another important difference between the Fort Edwards samples and the UMTRA samples that may further explain why the permeability of the former soil did not change significantly. The initial saturated water content w_s for the compacted Fort Edwards clay samples (Table 3) was always below the plastic limit w_p (Table 1), whereas w_s for all the UMTRA soils was well above w_p. In past research (1988), we identified the plastic limit as the saturated water content state below which little change in engineering behavior is caused by freezing and thawing. We found that when thaw settlement was plotted vs the ratio W_r of w_s to w_p no settlement is predicted if W_r is equal to one, i.e., $w_s = w_p$.

In the case for these freeze-thaw permeability tests we decided to plot the permeability ratio K_r (ratio of the permeability after freeze-thaw cycling to the permeability before freezing) vs the liquidity index LI. This index conveniently ranges between 0 and 100% for water contents ranging between w_p and the liquid limit w_l. This is the probable normal consistency range for compacted and consolidated fine-grained soils. The data are given in Table 3 and the results plotted in Figure 10. Data from Fleming (1987) for consolidated Fort Edwards soil are also included. There are two distinct curves, one for the compacted state and the other for the consolidated condition. The curve is linear for the compacted soils. The LI has a value of about -12% for no permeability change (K_r = 1).

For the consolidated Fort Edwards soil, the curve is curvilinear in the K_r vs LI plane. The LI for no permeability change after freeze-thaw cycling is not so well defined in this case. However, it appears that LI needs to be at or below 0% to achieve the K_r= 1 condition.

The results of these tests can be used to establish preliminary guidelines for specifying soils and compaction conditions for frost-resistant soil caps and liners. It is clear that very high compaction efforts are required to reach the condition where the saturated water content will be below the plastic limit. Even if this is accomplished, the void ratio and the permeability of the soil will probably increase a small amount with successive freeze-thaw cycling, particularly if the overburden is shallow. If the soil is highly compacted and in a near normally consolidated condition (requires large overburden stresses) before freezing, then little change in void ratio or permeability will occur. However, a thick overburden will generally prevent frost penetration into the protective soil barrier and the frost effects problem will be trivial.

IMPACT ON DESIGN AND CONSTRUCTION ACTIVITIES

The results of this study, although preliminary, indicate that maintaining the initial permeability in clay barriers in seasonally frozen soils may be difficult. Protective clay barriers must be compacted to densities at which their saturated water contents are at or below their plastic limits to minimize damage due to frost action. The alternatives may be to bury the clay below the frost depth, to modify the freeze-thaw behavior of clay with additives or to use a combination of a clay barrier and a synthetic impermeable membrane. Burying a clay barrier, however, can be a very expensive procedure. Certain soil stabilizers may reduce the permeability of clay soils, but their freeze-thaw durability is not known. Moreover, stabilizers may also add significant construction costs. Similarly, the freeze-thaw durability and cost of synthetic membranes may also be problems.

RECOMMENDATIONS FOR FURTHER STUDIES

The data obtained from this study indicate that more research needs to be done before clay liners can be recommended as moisture barriers in seasonally frozen regions.The studies needed include the following topics:

1. Nondestructive monitoring of moisture and macro-structure changes after freeze-thaw cycles.
2. Effect of freezing rates and conditions on soil permeability and structure.
3. Effect of freeze-thaw on unsaturated soil permeability and macro structures.
4. Modifying the response to freeze-thaw cycling with soil stabilizers.
5. Healing and recovery of permeability with time.
6. Effect of soil water solution chemistry on the permeability and macrostructure.
7. Correlation of laboratory results with larger scale prototype systems.

Once the effects of freezing and thawing on the permeability of soil barriers are well understood, we need to develop design criteria based on soil index properties such as the liquidity index. Freeze-thaw permeability tests may provide more precise design data; however, these tests are difficult and time consuming to conduct.

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INFILTRATION INTO A SEASONALLY FROZEN CLAY SOIL

by

Bo Thunholm and Lars-Christer Lundin[1]

INTRODUCTION

Infiltration into the soil is of importance for many physical, chemical and biological processes. A low infiltration capacity may lead to surface runoff and loss of plant nutrients, such as phosphorus. Persistent high soil water content may lead to deterioration of soil structure as well as denitrification.

The infiltration capacity is reduced by high ice content in the soil (Willis et al., 1961; Kane, 1980; Granger et al. 1984). The ice content is increased by high water content prior to freezing or by infiltration and freezing of snow melt water during winter.

This paper was based on data presented by Thunholm et al. (1989). The objective of the present study was to further interpret the effect of irrigation before soil freezing and during winter respectively on the infiltration rate during the thawing period. Measurements of total volumetric water content, soil temperature and infiltration rate in 3 small field plots were analysed by a numerical model (Jansson & Halldin, 1979). During the thawing period the differences in infiltration rate between the plots as well as temporal variations were studied.

FIELD MEASUREMENTS

The site (59°49'N, 17°39'E) was located in an agricultural field. The soil type is very similar to the heavy clay soils described by Sandsborg and Wiklert (1976). The clay content is around 50 % down to 1 m depth and the macro structure is well developed.

On three plots, denoted as plot 1, 2 and 3, measurements of total water content and soil temperature down to 1 m depth were made during two winter periods (Nov. 1985 - April 1987). Each plot covered 1 m^2 and lateral water flow was prevented by a rubber sheet that was dug down, extending from 10 cm above the soil surface down to 1 m depth. In order to prevent precipitation the plots were covered with aluminum plates, thus only artificially supplied water was considered. Soil temperature down to 50 cm depth was measured every 3 h with Pt-500 probes. At deeper levels the temperature was measured with thermistors. The total water content was measured every 5-10 days with a neutron probe.

In April 1986, after thawing, equal amounts of water (170 mm) were poured out on the three plots and the time for the water to infiltrate was measured. This study was made to estimate the differences in infiltration rate without any previous preparations of the plots. In order to create differences in water content prior to freezing, plot 2 and 3 were irrigated with 100 and 200 mm respectively in fall 1986. In order to increase differences in ice content, further 14 and 21 mm was added to plot 2 and 3 respectively prior to thawing 1987. Thereafter, during the thawing period, the measurements of infiltration rate were made. A total amount of 120 mm was added to each of the three plots on 14 occasions.

MODEL DESCRIPTION

The one-dimensional SOIL model used (Jansson & Halldin, 1979, Jansson & Halldin, 1980) simulates the coupled heat and water flow in a soil profile. The flow equations of the model are

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Presented at the Frozen Soil Symposium, March 21-22, 1990; Spokane, Washington.

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$$\frac{\delta(CT)}{\delta t} - L_f \rho_i \frac{\delta\theta_i}{\delta t} = \frac{\delta}{\delta z}\left(k_h \frac{\delta T}{\delta z}\right) - C_w \frac{\delta(Tq_w)}{\delta z} \qquad (1)$$

$$\frac{\delta\theta}{\delta t} = \frac{\delta q_w}{\delta z} + s_w \qquad (2)$$

Where C = soil heat capacity, T = temperature, t = time, L_f = latent heat of freezing, ρ_i = ice density, θ_i = volumetric ice content, k_h = thermal conductivity, z = depth, C_w = water heat capacity, q_w = water flow, θ = volumetric water content, s_w = water sink/source. Equation (1) describes the heat flow and Equation (2) describes the water flow. The equations are solved with an explicit finite difference method.

The calculation of unfrozen water content is made in a similar way during drying-wetting and freezing-thawing. A freezing-point depression curve, based on the soil-moisture characteristic curve, is used to estimate the unfrozen water content of the partially frozen soil. The water-retention function is estimated by a modified version of the Brooks & Corey (1964) equation. For water contents lower than 4 % below porosity, the hydraulic conductivity is estimated from the retention properties and an estimated saturated conductivity (Mualem, 1976). Close to saturation a logarithmic dependence of the conductivity on the water content is used (Jansson & Thoms-Järpe, 1986). To prevent overestimation of the conductivity at the freezing front an interpolation procedure suggested by Lundin (1989) is used. Thermal conductivity is estimated from Kersten's (1949) equation. Contrary to the measurements, surface runoff is accounted for by a procedure described by Jansson & Gustafson (1987).

In the present study, daily mean values of the measured soil temperature at 5 cm depth were used as the temperature at the upper boundary (i.e. the temperature at the soil surface). Since the plots were shielded from precipitation only the artificially supplied water was accounted for. Parameters for the water retention properties and the unsaturated conductivity were estimated from Sandsborg & Wiklert (1976).

RESULTS

The measured and the simulated total water content increased during the early part of the winter (Fig. 1). The simulated water content increased rapidly, contrary to the notable decrease in measured water content during thawing. Irrigating plot 2 and 3 in the fall of 1986 did not increase the total water content during winter although the pre-thawing irrigation in March 1987 seemed to have increased the measured water content in the upper layers (Fig. 2).

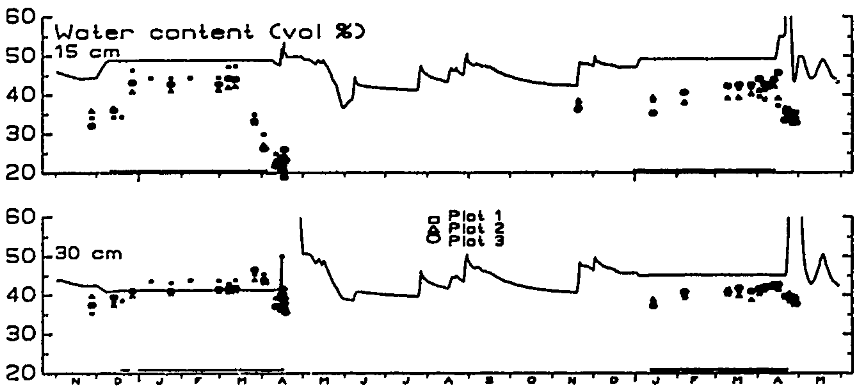

Figure 1. Simulated (solid line, plot 2) and measured total water content from Nov. 1985 to May 1987 at 15 and 30 cm depth. Time periods with soil temperatures below 0°C at the corresponding depths are marked with bars.

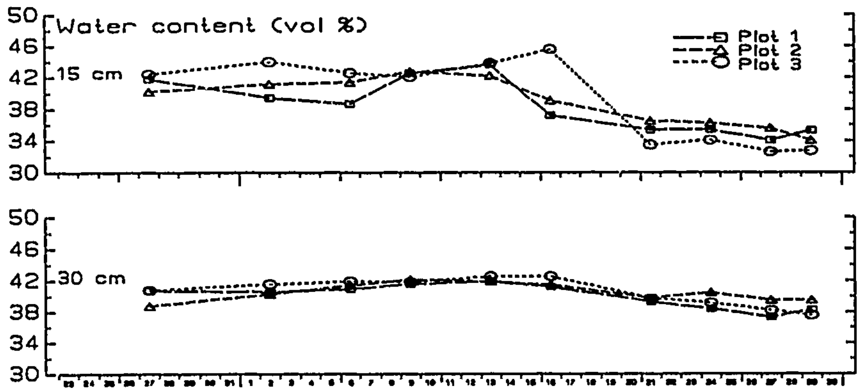

Figure 2. Measured total water content from 23 March to 30 April 1987 at 15 and 30 cm depth.

The measured infiltration rate during April 1987 was related to the soil temperature variation around the freezing point, although the different amounts in pre-thawing irrigation influenced the rate (Fig. 3). The simulated infiltration rate did not agree at all. Most of the added water was calculated as surface runoff and the simulated infiltration started 11 days after the infiltration actually began (Fig. 4). The delayed infiltration seemed to be related to a time lag in simulated soil temperature in the upper layers.

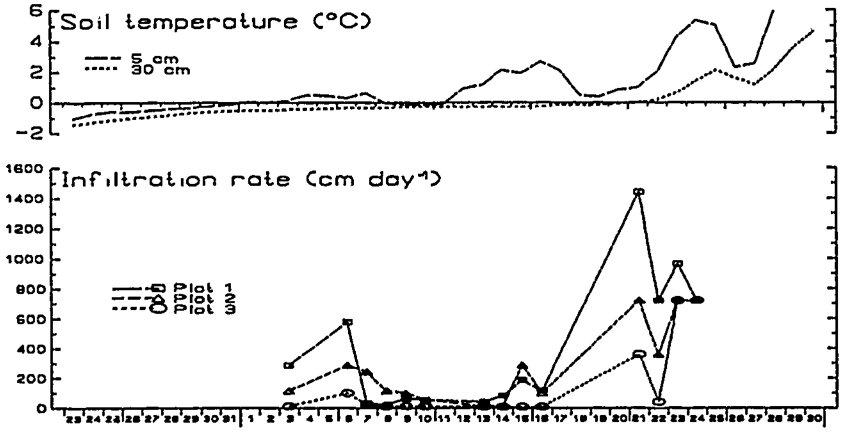

Figure 3. Measured soil temperature and infiltration rate from 23 March to 30 April 1987.

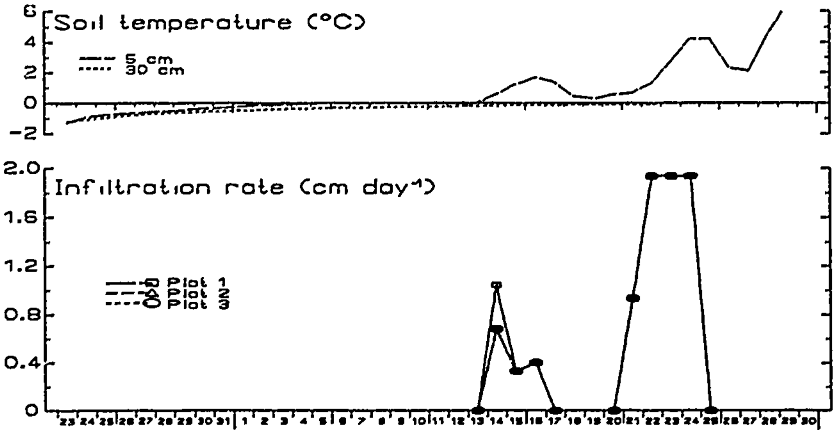

Figure 4. Predicted soil temperature and infiltration rate from 23 March to 30 April 1987.

DISCUSSION

The overestimation of the simulated water content (Fig. 1) during the thawing period indicates an influence of the freezing on the water retention properties. The formation of ice lenses in the macropores extracts water from the aggregates, thus the micro-pore system within the aggregates would be affected as a result of the extremely high tensions. Due to the slow swelling processes of clays this effect should be observeable after the thawing period. Since aggregates and cracks are not accounted for in the model, the calculated water content after the thawing period was almost the same as prior to freezing.

The high measured infiltration rates could be explained by a water flow in the macropore system. When the soil is frozen the ice lenses should only partially occupy the macropore system (Kenny & Saxton, 1986), thus the frozen layer may still have a relatively high conductivity. The relationship between temperatures close to freezing point and measured infiltration rates during April 1987 (Fig. 3) emphasizes the effect of the ice content on the hydraulic conductivity. The temperature variation indicates repeated thawing and freezing which may cause percolated water to freeze and fill the macropores with ice, thereby decreasing the infiltration rate. The inverse relationship between measured infiltration rate and the amount of added water prior to the thawing period of 1987 was very likely a result of freezing of the added water.

The discrepancies between simulated and measured infiltration rates were partly due to an overestimated ice content, associated by a time lag in simulated soil temperature. The delayed increase in soil temperature was probably caused by using the 5 cm soil temperature as the upper thermal boundary condition, resulting in a time lag around a week during freezing and thawing periods. As mentioned before, the inability of the present model to describe the aggregates and cracks should also contribute to discrepancies.

Another reason for different infiltration rates between the plots could be the influence of the freezing on the soil structure. The effect of freezing on aggregates varies between soil types but in general the destructive effect increases with increasing water content of the soil (Slater & Hopp, 1949; Logsdail & Webber, 1959; Benoit, 1973; Hinman & Bisal, 1973). Thus the inverse relationship between the pre-thawing irrigation and infiltration rate might as well be related to differences in structural condition as to the ice content. Destruction of the structure of a freezing soil at high water contents can also explain a commonly observed low infiltration capacity after thawing.

CONCLUSIONS

The results support the general consensus of an inverse relationship between ice content of the soil and infiltration rate. A high ice content can be caused by repeated thawing and freezing during the winter. A possible additional reason for a low infiltration capacity, though not verified, is a freezing-induced destruction of soil structure, especially at high water contents. If the soil structure is well developed a relatively high infiltration rate may be sustained troughout the winter. When modeling water flow in soils with high clay contents the soil structure should be taken into account. Further development should include the influence of freezing on the water retention properties and aggregates at different water contents.

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SNOWMELT INFILTRATION INTO COMPLETELY-FROZEN, SUBSOILED SOILS

by

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INTRODUCTION

Throughout the most part of the semi-arid prairie region of central Canada annual snowfall, which ranges in depth between 90 and 130 mm water equivalent, comprises on average about 30% of the annual precipitation and, excluding large-scale water development projects, constitutes the most amenable source of fresh water supply available for management. Meltwater from the shallow snowcovers serves many beneficial purposes as wildlife habitat, as a local supply for domestic and livestock use, and for recharging groundwater supplies, soil water reserves and lakes.

Agriculture is the primary industry of the region with cereal grains, pulse and oil seeds the principal crops. The importance of water to the production of cereal grains under dryland farming is well-recognized. For example, the Saskatchewan Council on Soils and Agronomy (1982) recommends that root-zone water reserves of 125 mm and 100 mm are needed at the time of seeding to achieve average yields in the Brown and Dark Brown soil zones, respectively. This recommendation is based on the expectation of normal rainfall in the growing season. Staple and Lehane (1952, 1954) and Staple et al. (1960) reported that each 25-mm addition of soil water to a moisture reserve in the range between 262 mm and near 412 mm may increase the yield of spring wheat by 230 - 400 kg/ha. Corroborating data are given by de Jong and Rennie (1969) who reported increases in the range of 200 - 275 kg/ha for each additional 25 mm of soil water above the long-term normal growing season precipitation for the relatively-humid, east-central part of Saskatchewan.

In recent years on the Prairies there has been renewed interest on the potential of "managing" snow to increase soil water reserves. Because of the type of crops and the nature of the farming operations, stubble management practices employing non-competitive vegetative barriers ("tall" stubble, "alternate-height" stubble, "trap" strips) to trap snow are commonly-used. Vegetative barriers also reduce the distance of snow transport, thereby reducing the amount of snow water lost by sublimation during blowing snow events. This loss may not be insignificant. A recent study by Gray et al. (1989) using the Prairie Blowing Snow Model developed by Pomeroy (1988) and five years (1970-1976) of meteorological data recorded at stations throughout the Prairie Provinces shows average annual sublimation losses from 1000-m fetches, expressed as a percentage of annual snowfall, ranging between 21 and 55% on stubble and 34 and 55 % on fallow in areas where snowcovers are continuous.

Evidence showing that snow management practices can be successfully applied in a prairie environment to increase snowcover depth is indisputable (for example see Nicholaichuk et al., 1986). However, an increase in depth of snowcover does not assure a proportional increase in meltwater infiltration. The problem arises because the period of snowcover ablation is short and infiltration takes place into "completely-frozen soils": herein defined as those soils whose frozen profile does not thaw to appreciable depth during snowcover ablation. Infiltration into frozen soils involves the complex phenomenon of coupled heat and mass transfer through porous media and the process is affected by many factors: the hydrophysical and thermal properties of the soil; the soil moisture and temperature regimes; the rate of release of water from the snowcover and the energy content of the infiltrating water. In the absence of major structural deformations e.g. cracks or other macropores, the most important hydrophysical property of a frozen soil which governs its ability to absorb and transmit meltwater is its moisture (ice) content.

Presented at Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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Granger et al. (1984) and Gray et al. (1986a) suggest that for practical purposes "completely-frozen" soils may be grouped into three broad classes in respect to their meltwater infiltration potential.

1. Restricted - infiltration is impeded by an ice lens on the surface or at shallow depth. The amount of meltwater infiltrating the soil is negligible and most of the snow water goes to direct runoff and evaporation.

2. Limited - infiltration is governed primarily by the snowcover water equivalent and the frozen water content of the 0-300 mm soil layer. The average depth infiltrating meltwater penetrates "Limited" soils is about 260 mm and the limit of saturation of the wetted depth (L) can be approximated as: $L = 0.6 + 0.4\theta_p$, in which L and θ_p, the average ice/water content of the wetted depth, are expressed as a degree of saturation in mm^3/mm^3. These findings demonstrate the capacity of "Limited" soils to absorb meltwater. For example, a soil with a porosity of 50% at an initial moisture content of 20% by volume (dry to medium wetness) would likely infiltrate on average less than about 47 mm of meltwater ([0.5(0.6 + 0.4*0.4) - 0.2] * 260 = 46.8), or the water equivalent of a "normal" snowcover with a depth in the range of 15-20 cm.

3. Unlimited - soils containing a high percentage of large, air-filled macropores which are capable of infiltrating most or all of the meltwater.

The material above suggests that snow management practices for augmenting soil water reserves from meltwater infiltration in completely-frozen soils are likely to prove most successful in those soils which naturally crack on drying. Non-cracking, medium and fine-textured soils must have their structure significantly altered to increase their macropore content in order to have the same potential. This paper reports on studies conducted in Saskatchewan on meltwater infiltration into naturally-cracked soils and the effects of subsoiling of non-cracking soils on infiltration enhancement and the yield of spring wheat.

FIELD EXPERIMENTS AND PROCEDURES

During the past nine years, snow and soil moisture monitoring programs have been conducted on cracked and subsoiled soils within the Brown and Dark Brown soil zones of the Province of Saskatchewan. The majority of the cracked sites are located in heavy, lacustrine clays (primarily Sceptre clay, >55% clay) on flat to gently undulating land under cultivation of cereal grains by dryland farming. The subsoiled sites, located near the village of Kerrobert, SK, are in glacial till whose principal texture is clay loam, although heavy clays dominate the lower slope positions. The general topography varies from flat to moderately rolling. On the rolling land the variations in soil texture and topographic position cause large gradients in yield over a field.

In the fall of 1983 a small area (0.5-ha) was ripped with a single - shank Killefer plow to a depth of 600 mm on a spacing of 1.85 m, the wheel spacing of the power unit. This treatment was repeated on different plots of land in the fall of 1984. In 1985 and 1986, other small areas were subsoiled with an Ebson plow (depth = 400 mm, spacing = 0.75 m) and a Kello-bilt plow (depth = 600 mm, spacings = 0.70 and 1.4 m). In 1987, two, 3.75 ha plots were ripped with a "Hubee" subsoiler to a depth of 500 mm, one with a spacing of 0.7 m, the other with a spacing of 0.9 m. Where possible, "tall stubble", "trap strip" or "leave strip" stubble-management practices were used for snow trapping. Smaller plots on the subsoiled areas were ringed with snow fences to ensure large accumulations of snow.

Changes in soil moisture were monitored with a two-probe gamma density meter before, during and immediately following snowcover ablation. This system provides non-destructive sampling of a cube of soil approximately 50 mm wide, 250 mm long and 20 mm thick. With the method, 50-mm diameter plastic access tubes spaced 305 mm apart are installed vertically into the soil. A radioactive source is inserted in one tube and detector at the same depth in the other and a measurement is taken of the number of photons striking the detector in one minute. The density of the soil is calculated knowing the intensity of the source and the attenuation coefficients of gamma radiation for soil and water. By assuming the mass of soil between the tubes on successive measurement dates remains constant i.e., no major structural changes occur in the soil profile, changes in attenuation of the gamma radiation are attributed to changes in the mass of water contained in the volume. (Note: It is pertinent to mention that openings in frozen soils remain relatively stable until water is imbibed by the surrounding soil. causing it to thaw and to swell). The equivalent moisture change is calculated from the readings assuming a density of water equal to 1000 kg/m^3. Measurements are taken at 20-mm increments of depth to 400 mm and at 40-mm increments between 400 mm and the bottom of the access tubes. The equipment was extensively tested and calibrated to operate reliably to temperatures of -20^{o}C. Measurements were taken with a "crack" or "rip" centred between the access tubes. with the access tubes aligned perpendicular to a "crack" or "rip" on centres of 150 mm and 450 mm from the opening. and in tubes placed in uncracked or undisturbed soil.

The depth of snowcover was measured routinely and the snowcover water equivalent obtained either from nuclear or gravimetric density measurements. Some locations were equipped with a Nipher precipitation gauge and these measurements were used to update the snowcover data. Several sites were also equipped to monitor the ambient air temperature and soil temperatures at different depths.

Yield samples were taken from the different treatments each year. The size of samples varied from 1 - m^2 to field size. Small plots were harvested by hand. the larger units by swathing or straight combining.

RESULTS AND DISCUSSION

Infiltration into undisturbed (uncracked). cracked and subsoiled soils

Figures 1 and 2 show changes in soil-moisture distribution patterns in undisturbed, cracked and subsoiled soils due to meltwater infiltration. In Figs. 1b and 2b the changes are calculated from measurements taken with the crack and "rip" located between the source and detector of the twin probe system: Figs. 1c and 2c are changes measured perpendicular to the fractures at a distance between 150 and 450 mm. The subsoiling treatment was a "Killefer" plow and the measurements were made on a plot which was located in an "upper" slope position and protected against flow along the lines (rips) by plastic liners inserted to a depth of about 1 m. The data in Figs. 1 and 2 substantiate comments made in the INTRODUCTION. namely:

1. Undisturbed soil: The amount of infiltration into undisturbed soil is limited. Both sites (Figs. 1a and 2a) show approximately 27 mm of meltwater infiltrating the undisturbed soils to depths of 375 and 185 mm. The snowcover water equivalents at the sites were 78 mm and 194 mm, respectively.

2. Cracks and Rips: Infiltration into "cracks" and "rips" is substantially greater than to undisturbed soil. The increases in soil moisture in the fractures were 94 mm and 173 mm respectively. Note the amount of infiltration to the "crack" (94 mm) is substantially greater than the snowcover water equivalent of 78 mm. This is attributed to interflow in inter-connecting cracks and surface runoff from the undisturbed soil between the fractures entering the opening directly. The subsoiled soil. which was "ripped" to a depth of 600 mm. shows large amounts of soil water recharge to a depth of 1 m (Fig. 2b).

3. Adjacent to Cracks and Rips: The amount of infiltration into the block of soil immediately adjacent to a crack or rip is less than that into the aperture, but greater than the amount into undisturbed soil. Only 36 mm of water infiltrated the soil adjacent to the crack (Fig. 1c). a small increase over the amount observed at the "uncracked" site. However. a significant increase in moisture (>78 mm), which extends below the depth of measurement. occurred adjacent to the "rip". This is attributed to an increase in hydraulic conductivity caused by rupture and fracture of the soil during installation of the lines and to the rip acting as a source of water for lateral movement. The vertically-elongated pattern is the result of increasing hydraulic head and soil temperature with depth. At the beginning of snowmelt the frost depth was between 1200 and 1300 mm.

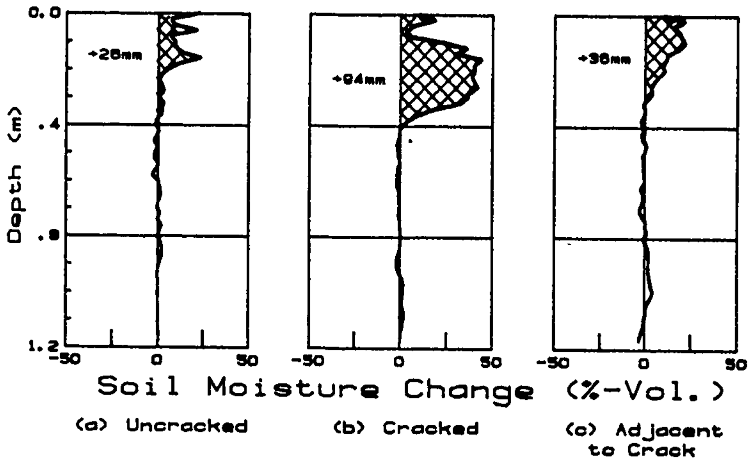

Figure 1. Soil moisture changes due to snowmelt infiltration in completely-frozen, uncracked and cracked heavy clay.

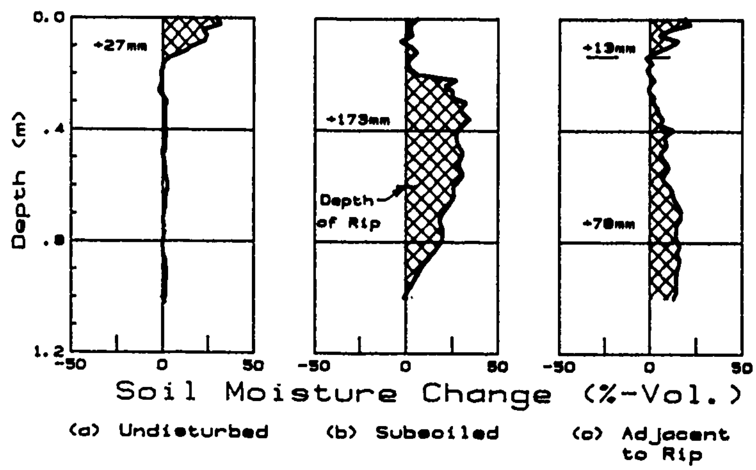

Figure 2. Soil moisture changes due to snowmelt infiltration in completely-frozen, undisturbed and subsoiled clay loam. Rip was installed to a depth of 600 mm by a Killefer plow.

Table 1 summarizes the findings of the soil moisture monitoring program as average amounts of snowmelt infiltration into undisturbed (uncracked) soils, cracks, and rips for different ranges of snowcover water equivalent. These data show infiltration increasing with snowcover water equivalent in both cracks and rips with a trend for rips to infiltrate more water than "cracks". However, the ratio of the amount of infiltration into an opening to the amount infiltrating undisturbed soil varies only slightly with depth of snow water. On cracked soils the ratio-values vary between 3.4 and 4.3 with a mean of 3.8 and on "ripped" soil between 5.4 and 6.5 with a mean of 5.8.

Depth of rips

Analyses of soil moisture profiles following meltwater infiltration into undisturbed, cracked and subsoiled "completely-frozen" soils show a strong relationship between infiltration (INF) and the depth of penetration of the wetting front (d) (Gray et al., 1986b). The curve enveloping these variables can be approximated as INF = 0.2d, in which INF and d are in mm. However, unless a snowcover is unusually deep (≈ >60 cm; SWE > 150 mm), d is less than 1 m (average = 630 mm). Based on this finding and a review of profiles of soil moisture changes monitored at numerous subsoiled sites and assuming a root zone of 1 m, it is suggested that lines be installed at depths in the range of 400 - 500 mm for efficient moisture conservation.

Table 1. Average amounts of snowmelt infiltration into uncracked/undisturbed soils. cracks and rips for different ranges of snowcover water equivalent (SWE).

SWE mm	INFILTRATION* Uncracked mm	Crack mm	Ratio	INFILTRATION** Undisturbed mm	Rips mm	Ratio
<30	12.4 (5)***	49.7 (2)	4.0	10.2 (2)	57.1 (3)	5.6
30-50	18.6 (7)	70.7 (2)	3.8	13.7 (3)	89.4 (4)	6.5
50-70	23.7 (11)	84.1 (9)	3.5	17.8 ****	99.7 (7)	5.6
70-100	28.0 (23)	95.7 (5)	3.4	21.5 (8)	117.1 (5)	5.4
100-150	30.5 (17)	116.6 (9)	3.8	23.0 ****	134.4 (9)	5.8
>150	34.5 (9)	147.0 (4)	4.3	27.6 (4)	154.9 (12)	5.6
		Mean	3.8		Mean	5.8

* Values for uncracked and cracked soils from all sites within Brown and Dark Brown soil zones of the Province in which soil cracking was observed. The major textural groups are heavy clay and clay loam.

** Values for subsoiled and undisturbed soil are for Kerrobert where the principal soil is a glacial clay loam. The subsoiling treatments included: (a) Killefer plow -depth = 600 mm. spacing = 1.9 m: (b) Ebson plow - depth = 400 mm. spacing = 0.75 m: (c) Kellobilt plow - depth = 600 mm, spacings = 0.70 and 1.4 m; (d) Hubee subsoiler -depth = 500 mm. spacings = 0.70 and 0.90 m.

*** Numbers in parentheses refer to the number of samples

**** Estimated values.

Areal infiltration and line spacing

Figure 3 shows group mean values of soil water changes due to snowmelt infiltration in rips. cracks. adjacent to rips and undisturbed stubble plotted against snowcover water equivalent. Expressing these relationships in equational form gives:

Cracks
$$INFC = 8.48SWE^{0.553} . \quad [1]$$

Rips
$$INFR = 16.86SWE^{0.432} . \quad [2]$$

Adjacent to Rips

For SWE < 80 mm $INFA = 1.65INF = 8.25(1 - \theta_p)SWE^{0.584}$ and [3a]

For SWE > 100 mm $INFA = -27.5 + 0.89SWE.$ [3b]

Undisturbed Soils
$$INF = 5.0(1 - \theta_p)SWE^{0.584}. \quad [4]$$

where: INFC = infiltration into a crack (mm). INFR = infiltration into a rip (mm). INFA = infiltration into the column of soil lying 150 to 450 mm adjacent to a rip (mm). INF = infiltration into undisturbed soil (mm). SWE = snowcover water equivalent (mm). and θ_p = average premelt soil water (ice) content of the 0-300 mm soil layer expressed as a degree of saturation (mm^3/mm^3).

Equations 2. 3 and 4 are used to develop a simple, conceptual model describing the average depth of snowmelt infiltration over a subsoiled field (INF(areal)) as a function of SWE and line spacing (s). Consider a soil which has been subsoiled with minimum disturbance of structure to the narrow block of soil adjacent to a rip. The system can be represented by three individual soil blocks (rip, adjacent to rip. and undisturbed (see insert in Fig. 4), each having its own infiltration characteristics. Taking the widths of the "rip" and "adjacent" block equal to the width of measurement of soil moisture change (≈ 300 mm), the water balance equation can be written as:

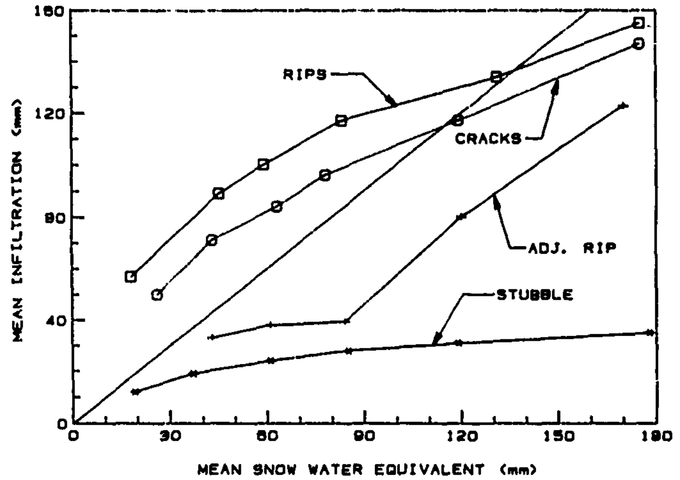

Figure 3. Group mean values of snowmelt infiltration into undisturbed stubble, adjacent to rips, cracks and rips as a function of mean snowcover water equivalent (mm).

$$INFR(areal) * s = INFR * 0.3 + INFA * 0.6 + INF * (s - 0.9). \qquad [5]$$

in which the spacing of the rips is in metres. Substituting the expressions for INFR, INFA and INF given by Eqs. 1 - 4 into Eq. 5 and solving gives:

For <u>s ≥ 0.9 and SWE ≤ 70 mm</u>

$$INF(areal) = \{5.06SWE^{0.432} + (s - 0.09)[5(1 - \theta_p)SWE^{0.584}]\}/s \qquad [6a]$$

and for <u>s ≥ 0.9 and SWE ≥ 100 mm</u>

$$INF(areal) = \{5.06SWE^{0.432} - 16.5 + 0.534SWE + (s - 0.9)[5(1 - \theta_p)SWE^{0.584}]\}/s \qquad [6b]$$

For snowcovers with 70 mm ≤ SWE ≤ 100 mm, the lesser amount of INF(areal) given by Eqs. 6a and 6b is used.

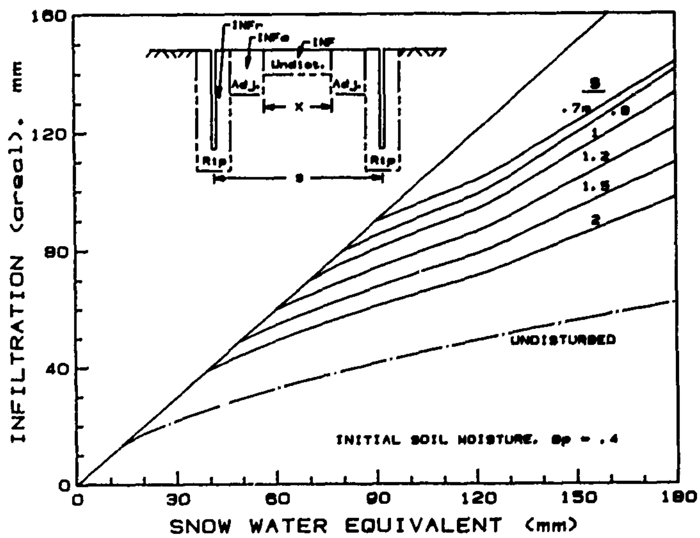

Figure 4. Relationship between "areal" infiltration, line spacing and snowcover water equivalent for subsoiled soils modelled by Eqs. 6a and 6b with $\theta_p = 0.4$.

Equations 6a and 6b are plotted for a "premelt" soil moisture, θ_p = 0.4 (dry to medium wetness) and a range of SWE-values in Fig. 4; also plotted is INF(areal) for rips spaced at 0.7 m and for undisturbed soil. To demonstrate the use of the curves assume average annual snowfall water equivalents of 90 and 130 mm (range in snowfall over a large part of Saskatchewan) and snow management trapping efficiencies of 60 and 75% of annual snowfall. Using these values the limits in available SWE would be 54 to 98 mm and the spacing of lines required to infiltrate all the snow water by Fig. 4 would be 1.3 and 0.7 m, respectively. Similar calculations with θ_p = 0.6 (wet soil) gives spacings of 0.8 and <0.7 m. The results of the calculations suggest that throughout the semi-arid Prairies lines spaced between 0.7 m and 1.3 m, depending on soil moisture conditions, will likely absorb the water from snowcovers produced by normal snowfall. A spacing of 0.7 m is considered very conservative. "Field observations have shown that most commercial subsoilers, when operated on spacings of 0.7 m or less, rupture and fracture the mass of soil between the rips to such an extent that, at least in the first year, it will have the capacity to infiltrate most of the snow water which is likely to be accumulated by a stubble management practice, independent of the soil moisture content". It can be expected that the infiltration characteristics of subsoiled soils will decrease naturally in time due to settlement and packing. However, field observations suggest that the beneficial effects of subsoiling on meltwater enhancement may be "longer lasting" in fields ripped at wide spacings (to some maximum) due to the natural cracks which develop between the lines in response to the strong gradients in soil moisture established during crop growth. There is an upper limit to the spacing of lines however. Visual observations of crop growth on lines spaced at 1.4 m or wider show nonuniform stand, the plants directly over the lines are taller and mature more slowly.

In summarizing the material above, it is suggested that subsoiling most glacial till soils of the Province to a depth between 400 500 mm on a line spacing between 0.9 and 1.3 m will allow efficient meltwater augmentation under normal snowfall and soil moisture conditions. For slightly solenetzic soils the spacings should be decreased slightly.

Areal infiltration of cracked soils

Measurements of the physical dimensions of cracks in heavy lacustrine clays showed a mean length:area ratio of 1.75 m/m^2 (Gray and Granger, 1986). Assuming the width of the soil column sampled by a twin probe equal to 300 mm, a point measurement of infiltration into a crack is representative of a surface area of 0.53 m^2/m^2. Combining the snowmelt infiltration equations for a crack (Eq. 1) and an undisturbed soil (Eq. 4) and weighting each according to their respective surface area gives an expression for areal infiltration into a 'cracked soil" as:

$$INFC = 4.53SWE^{0.553} + 2.33[(1 - \theta_p)SWE^{0.584}] \qquad [7]$$

The application of Eq. 7 will change as the length:area ratio for cracks varies from the assumed value of 1.75 m/m^2.

Effects on yield of spring wheat

Throughout the study all subsoiled plots were in continuous stubble and no supplemental fertilizer was applied. Yield samples collected in the 4-yr period showed that each additional 25 mm of water to a maximum of 160 mm produced an average annual increase in the yield of spring wheat of 185 kg/ha. This rate (7.3 kg/ha-mm) is consistent with rates reported by Staple and Lehane (1954) and de Jong and Rennie (1969).

Table 2 lists average yield data from subsoiling treatments without snow management. Because these data are largely based on samples from 1-m^2 plots, which for the subsoiling treatments were centred over a rip, they should be treated as indices of potential increases that may be expected in fields which have been subsoiled at narrow spacings. The data show a decrease in treatment effect on yield with time. In the first two years following subsoiling the decrease is small: in the first year there is an average difference of 435 kg/ha compared to the yield from undisturbed stubble (34% increase), while in the second year a difference of 377 kg/ha (30% increase) is observed. In the fourth year the difference has reduced to 20 kg/ha (only 1% increase). The small difference in the fourth year is likely due in part to the low amount of precipitation (snow and rain) received in 1987, the only year with fourth year data. Over the duration of the trials annual precipitation has been much lower than normal (eg. mean May-August rainfall ~200 mm).

Table 2. Effect of subsoiling on the yield of spring wheat <u>without</u> snow management.

Landuse and Age of Rips	SWE* mm	Rain** mm	Yield (kg/ha) Difference***	Ratio****	Range in Ratio
Fallow*****		110	610	1.44	1.05 - 1.71
Stubble Rips					
1st year	52	110	435	1.34	1.18 - 1.60
2nd year	50	122	377	1.30	1.18 - 1.52
3rd year	42	122	132	1.11	1.06 - 1.15
4th year	26	100	20	1.01	1.01

* Mean snowcover water equivalent.
** Average growing season rainfall.
*** Difference in yield from that on undisturbed sites of continuous stubble .
**** Ratio of mean yield on treatment to that on stubble.
***** 4-yr average yield from fallow without snow management was 1896 kg/ha.

Table 3 summarizes the effects on the yield of spring wheat of "combined" snow management and subsoiling practices relative to the yield from undisturbed plots of continuous stubble. As with Table 2 these data should be used as indices of potential increases because of the sampling procedures and because the depths of snowcover water are considerably higher than would naturally cover a field. During the four - year period the average snowcover water equivalent on undisturbed stubble was 39 mm. The data in Table 3 exhibit a trend similar to that reported above, namely a decrease in the average yield with age of treatment. If the tests for the fourth year (1987), the low snow year are excluded, the decrease is independent of available snow water. Regardless, the average increase of 1033 kg/ha on the snow-managed, subsoiled plots is higher than the increase of 610 kg/ha measured on fallow.

Table 3. Effect of snow management and subsoiling on the yield of spring wheat.

Age of Treatment	SWE* mm	Rain** mm	Yield (kg/ha) Difference***	Ratio****	Range in Ratio
1st Year	152	110	1435	2.06	1.52 - 2.06
2nd Year	145	106	1335	1.71	1.40 - 1.86
3rd Year	180	122	855	1.60	1.33 - 1.69
4th Year	64	100	510	1.28	1.28

* Mean snowcover water equivalent on plots.
** Average growing season rainfall.
*** Difference in yield from undisturbed sites of continuous stubble.
**** Ratio of mean yield on treatment to that on stubble.

The average, 4-yr yield increases in spring wheat over that monitored on undisturbed, continuous stubble of: (a) 1033 kg/ha on subsoiled, snow-managed plots, (b) 523 kg/ha on snow-managed, undisturbed plots, and (c) 240 kg/ha on subsoiled plots <u>without</u> snow management clearly demonstrate the value of "combining" subsoiling and snow management practices. It is reasonable to assume that larger increases would have been achieved with the addition of supplemental fertilizer.

SUMMARY

The results of a field monitoring program conducted within the Brown and Dark Brown soil zones in the Province of Saskatchewan on the meltwater infiltration characteristics of 'completely-frozen' uncracked, undisturbed, cracked and subsoiled soils are presented. A completely-frozen soil is one whose frozen profile does not thaw to appreciable depth during snowcover ablation. The importance of the macropore content of these soils to the infiltration process is emphasized.

Meltwater infiltration into soil cracks and "rips" averaged 3.8 and 5.8 times the amounts to the same soils in uncracked and undisturbed condition, independent of available snow water. Based on the field observations a practical model is developed which gives the average depth of infiltration over a field subsoiled to a depth between 400 and 500 mm as a function of line spacing, snowcover water equivalent and soil moisture (ice) content. Using this model with long-term snowfall statistics and reported catch efficiencies of stubble management practices, it is suggested that subsoiling on spacings between 0.9 and 1.3 m should enhance the infiltration properties of frozen soils within the semi-arid Prairie region so they are capable of absorbing most of the snow water accumulated by stubble management practices in a normal snow year.

The annual yields of spring wheat over a 4-yr period on subsoiled plots with snow management averaged 1033 kg/ha higher than the yields from plots of continuous, undisturbed stubble. This difference is in contrast to an average increase on subsoiled plots without snow management of only 240 kg/ha and emphasizes the need to "combine" the practices for maximum return. An average yield increase in the range of 7.3 kg/ha for each additional mm of available water found in the study corresponds closely with the values reported by soil scientists. The difference between the annual yields from subsoiled areas and plots of undisturbed stubble decreases substantially following the second year of treatment. This trend coupled with field observations of changes in the infiltration properties of rips with time suggests a useful life for the treatment effects of the order of 4 to 5 years.

ACKNOWLEDGEMENTS

The writers wish to acknowledge the technical support provided the project by Mr. Dell Bayne and the financial support of the Strategic Grants Program, Food and Agriculture, Natural Sciences and Engineering Research Council of Canada and the Agriculture Development Fund, Province of Saskatchewan under the project "Snow Management and Meltwater Enhancement", ERDA Project 3.2.3

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CROP MANAGEMENT EFFECTS ON RUNOFF AND SOIL LOSS FROM THAWING SOIL

by
Donald K. McCool[1]

INTRODUCTION

Frozen and thawing soil has a major impact on runoff and soil loss in the northern tier of states in the US. The effects are perhaps most pronounced in the dry-farmed cropland of the Pacific Northwest, where the soil freezes and thaws several times each winter. Agriculture Handbook 537 (Wischmeier and Smith, 1978) indicates as much as 90 percent of the soil loss in that area is caused by surface thaws and snowmelt. Researchers in north-central Oregon, in a short-term study, found about 86 percent of the soil erosion on winter wheat was caused by snowmelt, rainfall, or frost melt on thawing soil (Zuzel et al., 1982).

The influence of crop treatment on frozen soil runoff and erosion is a subject for frequent debate among action agency personnel, producers, and researchers. The statement is sometimes heard that conservation tillage has little or no influence on runoff and erosion from frozen soil.

The purpose of this paper is to examine some of these issues, using data collected on runoff plots at the Palouse Conservation Field Station near Pullman, WA and field observations made in Whitman County, Washington by V. G. Kaiser, deceased, Agronomist with the SCS, from 1940 through 1979.

EXPERIMENTAL SETUP

Runoff plots at the Palouse Conservation Field Station were established prior to the winter of 1972-73 and have been in operation since. The first treatments involved only conventionally-tilled winter wheat after small grain (WW/SG-T) and winter wheat after summer fallow (WW/SF-T). The project was expanded in 1977-78 to include various treatments including no-till seeding of winter wheat after small grain (WW/SG-N), conventional and no-till seeding of winter wheat after spring peas (WW/P-T and WW/P-N), and various rough primary tillages (RTS). The plot area is on a south-facing slope with a 19 to 26 percent steepness. The soil is a Palouse silt loam. Most plots are 3.66 m wide by 22.1 m long, although a plot length study is included. Runoff and sediment are collected in tanks which are sampled and emptied on an event or daily basis. Instrumentation on the plots includes frost tubes to determine frost depth, neutron access tubes to allow collection of soil moisture data, and weighing-type recording rain gages. Air temperature data are available from max.-min. thermometers in a standard shelter. The data presented are from the winters of 1978-79 through 1987-88.

The Kaiser soil loss data were collected in the late winter or spring from sample winter wheat fields throughout Whitman County. The data were largely observational in nature although rill cross sections were obtained as a check on the method. The same fields were observed year after year or, when in a crop rotation involving summer fallow or spring crop, every other year.

DATA ANALYSIS

The runoff and soil loss events on the plots were separated into frozen or unfrozen based on weather records and observations at the time the collection tanks were emptied. Frozen ground events include both those when the soil was frozen at the surface and those when it was thawed at the surface but a frost layer was present. Unfrozen events are those when frost was not present and had not been present immediately prior to the event.

Paper presented at the Frozen Soil Symposium, Spokane, WA, March 21-22, 1990.

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The Kaiser data for soil loss from individual years were separated by checking weather records for periods of freezing followed by rainfall during the warming period four days following. Examination of weather records indicate that frost is frequently gone within four days after the start of a warming period (McCool and Molnau, 1974).

RESULTS AND DISCUSSION

The winter of 1978-79 was unusually cold. Snow and cold temperatures occurred in early December. The snow melted and compacted in late December. January was quite cold and more snow fell on the deeply frozen ground. The snow was of insufficient depth and the air temperature too cold for the soil to thaw. In early February, warming and rain occurred. The snow melted and runoff ensued. Little or no erosion occurred until the ground was bare and had thawed 25 to 75 mm. Then three days of rain occurred. At the end of this period the soil was still frozen at a depth of approximately 150 mm. Most of the snow and rain had run off, and the plots had suffered various amounts of soil loss depending upon crop treatment. On all treatments but WW/SF-T, nearly all of the runoff and soil loss for the winter occurred in the one extended frozen ground event. Runoff ranged from 83 to 132 mm on the different treatments during the event. Soil loss ranged from 1.17 to 22.9 t ha^{-1} (Table 1, Figure 1). Residue and roughness were enough to protect the soil even from large quantities of runoff.

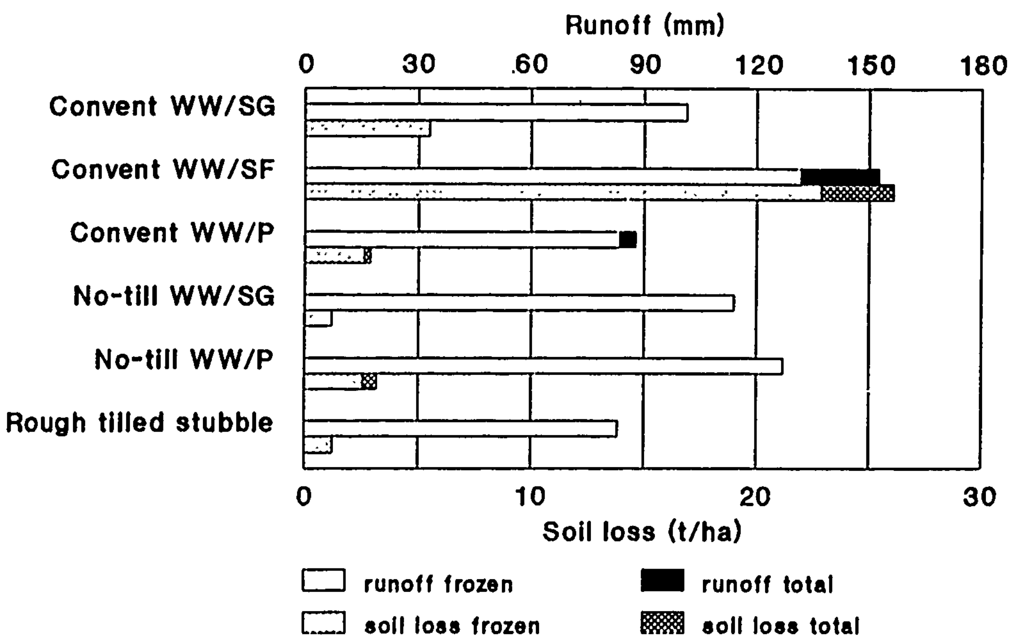

Figure 1. Runoff and soil loss for 1978-79 from plots at Palouse Conservation Field Station

The WW/SF-T plots suffered runoff and erosion from other events in addition to the frozen ground event. An additional 20 mm of runoff and 3.21 t ha^{-1} of soil loss occurred in non-frozen ground events.

The following two years of 1979/80 and 1980/81 were fairly open winters. Yet, the influence of frozen ground on runoff and erosion was evident (Table 1). In 1979/80, soil loss on WW/SF-T was 7.62 t ha^{-1}, of which 5.02 was from frozen ground events. Runoff was 21 mm, of which 16 mm occurred during the frozen ground events. Other treatments had lesser amounts of runoff and soil loss. Runoff occurred from other than the frozen ground events, but erosion occurred almost exclusively from the frozen ground events.

In 1980/81 a large quantity of runoff occurred only from the WW/SF-T (Table 1, Figure 2). About half of the 72 mm of runoff occurred from frozen ground events. Only 5.11 t ha^{-1} of the 18.23 t ha^{-1} of erosion occurred from frozen ground events, however. Of the other treatments, only WW/P-T had any quantity of runoff and erosion. Frozen soil was involved in slightly more than one-third of the runoff, but almost unmeasurable soil loss.

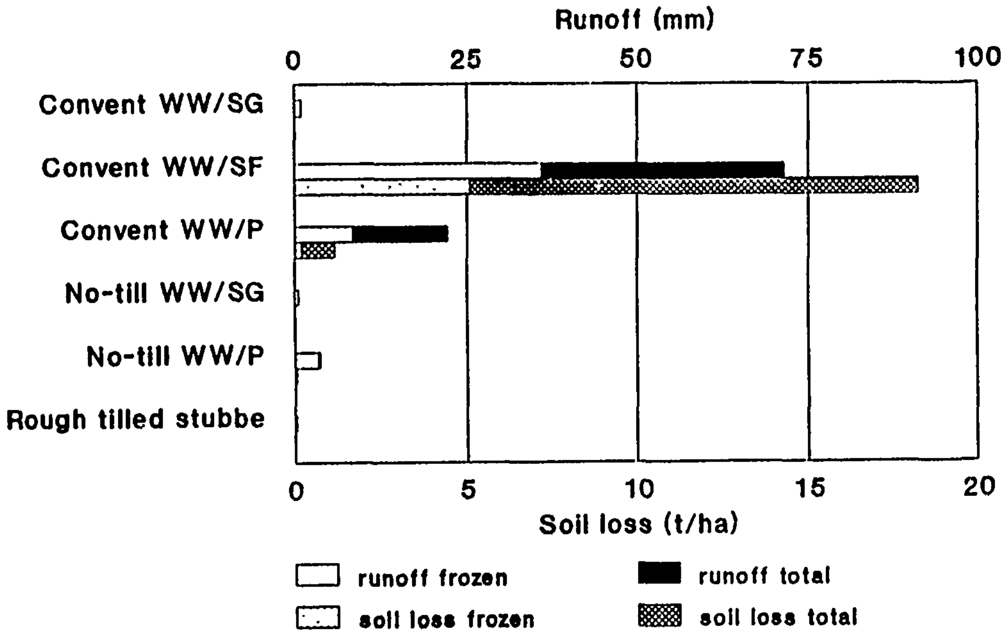

Figure 2. Runoff and soil loss for 1980-81 from plots at Palouse Conservation Field Station

Data from 1981/82 (Table 1) are quite similar to those from 1980/81. Appreciable runoff and soil loss occurred only from WW/SF-T. About half of the runoff came from frozen ground events but only 7.06 t ha^{-1} of the 28.02 t ha^{-1} total soil loss. The lack of soil loss during frozen soil events in 1980/81 and 1981/82 indicates that the soil was frozen at the surface while much of the runoff was occurring.

Frozen ground effects were varied in 1982-83. WW/SF-T and RTS both produced runoff and soil loss during events when the soil was not frozen. The runoff and soil loss quantities were both low from the RTS plots.

All treatments, including rough tilled stubble (RTS), produced runoff in 1983/84 (Table 1). On all plots but WW/SF-T, almost all runoff and soil loss came from frozen soil events.

Runoff and soil loss were quite small in 1984/85 (Table 1). Even on the WW/SF-T, runoff and soil loss came only from frozen ground events.

The remaining three years of the data show no unexpected trends, except that in 1987-88, a larger than normal amount of erosion occurred on WW/SF-T when the soil was not frozen. Again, this may indicate the soil was frozen at the surface when much of the runoff was occurring

Combining the ten years of data provides an interesting picture. For many of the crop managements, frozen soil runoff events account for the bulk of the runoff and erosion. These include all of the less erosion-prone treatments. For WW/SF-T, however, frozen soil events account for only 40 of the 60 mm of average runoff (about 67 percent), and 7.13 of the 15.40 t ha^{-1} of soil loss (about 46 percent).

These short term plot data can be compared with the long term erosion records collected by V. G. Kaiser from 1940 through 1979. These data are estimates of the total soil loss in Whitman County, Washington. The data were developed by expansion from a vast number of sample fields. Out of the 40 years of data, temperature, precipitation and stream flow records indicate a major frozen ground runoff event in about 9 of the 40 years. These may be identified, essentially, as the years with greater then 13.5 million t of soil loss (Figure 3). These 9 years (summation of 13.5-18.0 and above 18.0) account for 41 percent of the erosion in the entire 40-year period. Other years undoubtedly included frozen ground events of lesser magnitude.

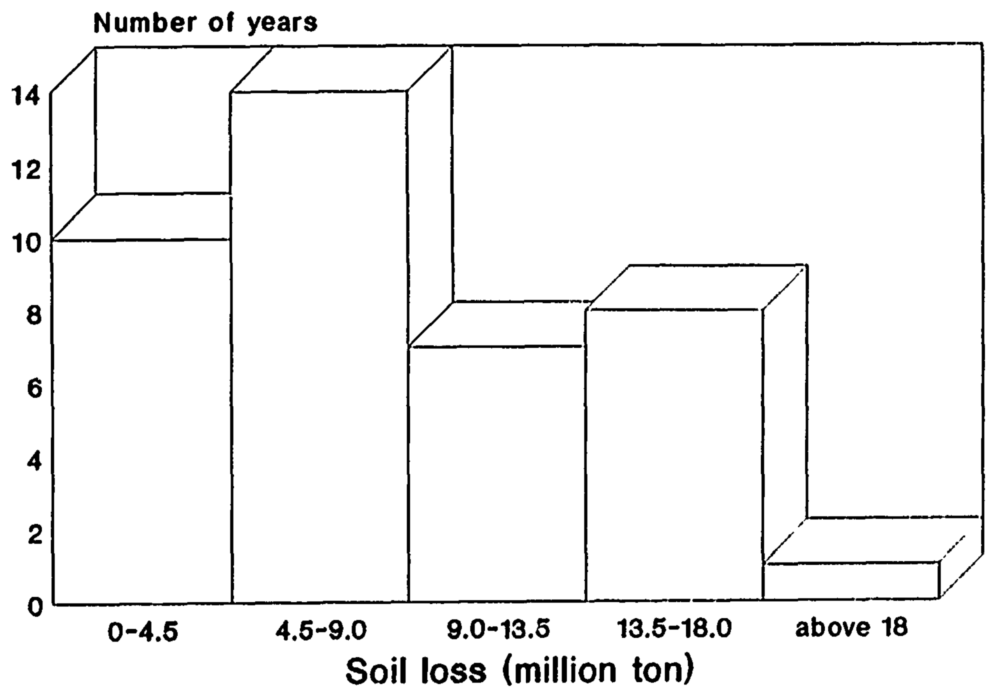

Figure 3. Estimated soil loss in Whitman County, Washington, 1939 through 1979.

CONCLUSIONS

From the plot data and examination of the Kaiser data, the following may be concluded:

1. Frozen and thawing soil runoff events account for a significant portion of the erosion in the Palouse region of the Pacific Northwest. The percentage, however, is dependent on crop treatment. A figure of 90 percent for the loss from frozen soil events appears too large for the crop managements in general use in the area.

2. Crop management can have a major impact on both runoff and erosion for events when there is no frozen soil or if the soil is only frozen to a shallow depth and thaws rapidly.

3. Crop management can have a major impact on erosion for events when the soil is deeply and impermeably frozen.

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Table 1. Runoff and Soil Loss Data, Palouse Conservation Field Station

	RUNOFF		SOIL LOSS	
YEAR	Frozen (mm)	Total (mm)	Frozen (t/ha)	Total (t/ha)
WW/SG-T				
78/79	101.3	101.3	5.51	5.51
79/80	4.8	7.1	0.34	0.34
80/81	1.0	1.0	0	0
81/82	1.5	2.3	0	0
82/83	15.7	16.5	0.69	0.76
83/84	30.0	30.0	0.31	0.31
84/85	0	0	0	0
85/86	24.9	25.1	1.34	1.34
86/87	0.5	0.8	0.02	0.02
87/88	16.0	16.5	0.04	0.20
AVG	19.6	20.1	0.82	0.85
WW/SF-T				
78/79	131.8	152.4	22.86	26.07
79/80	15.5	21.3	5.02	7.62
80/81	36.1	71.6	5.11	18.23
81/82	38.1	75.4	7.06	28.02
82/83	30.0	93.5	11.05	34.39
83/84	69.8	90.4	13.34	18.54
84/85	9.9	10.2	1.10	1.12
85/86	20.8	27.9	0.81	2.04
86/87	18.0	18.5	4.21	4.26
87/88	29.7	39.6	0.76	13.70
AVG	40.0	60.1	7.13	15.40

YEAR	RUNOFF Frozen (mm)	RUNOFF Total (mm)	SOIL LOSS Frozen (t/ha)	SOIL LOSS Total (t/ha)
WW/P-T				
78/79	83.6	87.9	2.60	2.87
79/80	0.8	1.0	0.02	0.02
80/81	8.4	22.1	0.02	0.94
81/82	7.6	29.0	0.18	1.57
82/83	32.0	32.8	2.11	2.13
83/84	42.9	44.2	3.74	3.79
84/85	1.5	1.5	0	0
85/86	51.8	55.9	8.47	9.01
86/87	8.4	8.6	0.72	0.72
87/88	31.2	33.3	0.43	1.12
AVG	26.8	31.6	1.83	2.22
WW/SG-N				
78/79	114.0	114.0	1.17	1.17
79/80	5.6	6.1	0.09	0.16
80/81	0.5	0.5	0	0
81/82	10.4	15.7	0	0.02
82/83	0.5	0.5	0	0
83/84	8.6	8.6	0.04	0.04
84/85	0.2	0.2	0	0
85/86	4.6	4.8	0	0
86/87	0.8	0.8	0.02	0.02
87/88	2.0	2.3	0	0.02
AVG	14.7	15.4	0.13	0.14
WW/P-N				
78/79	127.0	128.5	2.56	2.62
79/80	9.1	12.4	1.57	1.64
80/81	3.3	3.6	0.04	0.04
81/82	13.7	19.3	0.13	0.36
82/83	34.0	40.9	0.74	1.08
83/84	51.8	54.4	1.97	2.22
84/85	17.3	17.3	0.09	0.09
85/86	56.4	65.3	1.61	1.88
86/87	3.8	4.1	0.20	0.20
87/88	6.1	6.6	0.02	0.09
AVG	32.2	35.2	0.89	1.02
RTS				
78/79	83.0	83.0	1.23	1.23
79/80	0	0.5	0.02	0.07
80/81	0.2	0.2	0	0
81/82	0.5	0.5	0	0
82/83	0.8	5.1	0	1.57
83/84	8.9	8.9	0.13	0.13
84/85	0.2	0.2	0	0
85/86	4.3	4.6	0.04	0.04
86/87	0	0	0	0
87/88	0.5	0.5	0	0
AVG	9.8	10.4	0.14	0.30

Table 1 (continued). Runoff and Soil Loss Data, Palouse Conservation Field Station

EFFECT OF FREEZING ON MASS AND HEAT TRANSFER IN POROUS MEDIA

by

Neil N. Eldin, Leonard R. Massie, and Nagwa Shafik Aggour[2]

INTRODUCTION

Frost action in soils causes: 1) An upward movement (heaving) of the ground surface during the cold season due to the freezing of soil moisture and the formation of growing ice lenses, and 2) A subsequent loss in the supporting capacity of the soil during the warming season due to the over saturation caused by water trapped above the frozen underlayer. These problems often cause structural damages varying from simple cracks to complete collapse of the involved structures.

Artificial salting of soils as a frost modifier has been studied by many researchers (Yong et al, 1980, Banin et al, 1974, Lamb et al, 1971, Kronik et al, 1969). In this technique, the ion concentration of the pore solution is raised by applying water soluble salts to the ground. The absorbed cations are replaced by those of the introduced salt ion resulting in changes in particle orientation, hydro-physical properties, and physiochemical properties of the soil. Many investigators (Sheeran and Yong, 1975, Brandt, 1972, Korniket et al, 1969, Lamb, 1956) believe that salt treatment is one of the most efficient and least expensive methods to alleviate the detrimental effects of frost heaving in soils, but much still remains to be learned.

This paper presents the results of an experiment conducted to measure the effect of increasing the ion concentration in the pore water on the freezing behavior of a silt soil. The study investigated the effect of sodium chloride (NaCl) on moisture transfer, the temperature profile, and frost penetration in the tested soil. NaCl was chosen because it is highly soluble in cold water, safe to use, inexpensive, commercially available, and used by most authorities for deicing of highways, runways, and railroads. Also, calcium chloride ($CaCl_2$) and ferric chloride ($FeCl_3$) were included in this study to determine the significance of the cation's valence on the frost heaving reduction. Freezing tests were conducted on 5x15 cm specimens under temperature gradients of $1^oC/cm$.

EXPERIMENTATION

Open-system unidirectional freezing tests were conducted to establish the frost heave reduction efficiency of NaCl as a function of the concentration level when applied at different soil dry densities. Test specimens of 5 cm inside diameter and 15 cm height were prepared at initial soil dry densities ranging from a relatively loose 1.5 gm/cm^3 to a dense 2.0 gm/cm^3 condition. The mixing water for the specimens contained salt concentration ranging from 0.01N to 1.0N. During the freezing test, specimens were subjected to an air temperature of -10^oC while their bottom temperature was maintained at $+4^oC$. Overall surface heave, the final moisture profile and the temperature profile were measured. The final position of the frost front was determined visually.

Presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990

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Test specimen

The tested soil consisted primarily of quartz and feldspar, with traces of illite and chlorite. The grain size distribution of the soil showed 78% by weight finer than 0.2 mm in equivalent particle diameter, 52% finer than 0.6 mm, 37% finer than 0.02 mm, 25% finer than 0.006, and 17% finer than 0.002 mm. The soil had a liquid limit of 20% and a plastic limit of 15%.

Multi-ring plexiglass test cells (Figure 1) were used to minimize friction between the soil and the test cell during the freezing process. This design was chosen to avoid side friction between the soil and a rigid test cell which may alter the freezing point of the pore water, the magnitude of frost heaving, and the depth of frost penetration as demonstrated by Hammamji (1969).

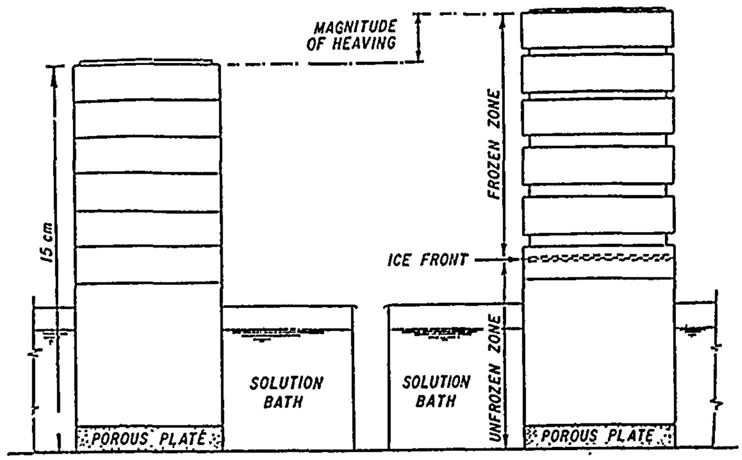

Figure 1. Multi-Ring Test Cell

Test equipment

A cold room having inside dimensions of 2.5 x 2.5 x 2.0 m was used to control the freezing temperatures and house the test box. The cold room was equipped to maintain constant air temperature of -10°C and to avoid temperature zoning. A test box with inside dimensions of 100 x 75 x 25 cm was built to accommodate up to 20 test specimens. An electrical heating pad controlled by a thermoregulator was used to maintain the temperature of the solution bath for each specimen at +4°C during the freezing test. A schematic view of the test setup is given in Figure 2.

Test measurement

A system of dial gauges, reading to 0.025 mm and affixed to the test box, was used to measure the uniaxial heaving of the test specimens. A thin metal plate was placed on the top of the specimen to provide a smooth solid surface for the dial gauges and to reduce evaporation of soil moisture. Copper-Constantan thermocouples were used with a potentiometer to determine the sample temperature profiles during the test.

DISCUSSION OF RESULTS

The effect of artificial salting on the freezing of soil moisture and on the ability of soil to transport moisture during freezing is discussed.

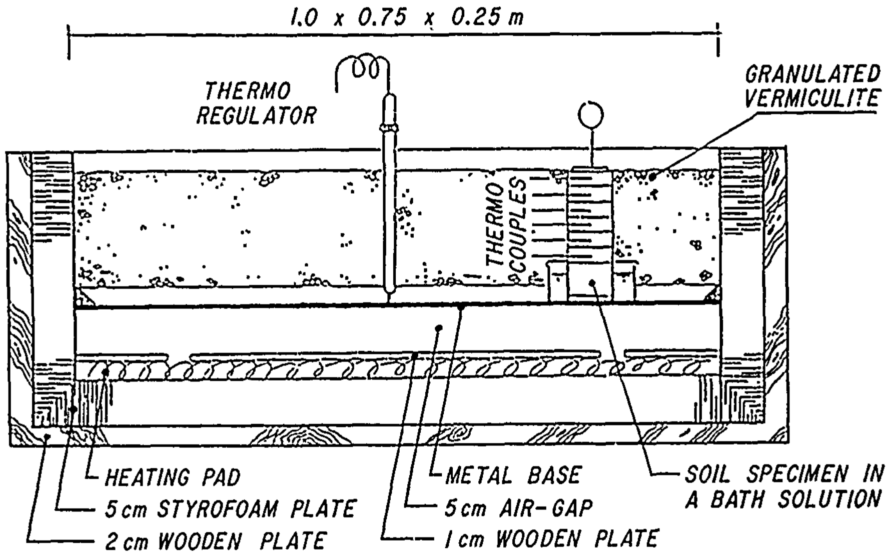

Figure 2. Schematic for the Test Box

Effect of artificial salting on moisture freezing

The presence of soluble salts in pore fluid lowers its freezing point. The magnitude of the freezing point depression depends on the concentration and valence of the salt. Thus, by adding a salt to the soil moisture a lower temperature has to be reached before a change in phase (water to ice) occurs. An ice-free soil could be maintained throughout the cold season, if the temperature does not fall below the freezing point of the pore fluid. Even if the temperature falls below the freezing point of the pore fluid, the artificial salting acts as a frost retarder. More heat, as compared to an untreated soil, has to be extracted from the soil before a frost front can advance in the soil. As water turns to ice, the freezing temperature of the pore fluid is lowered continuously because of the increasing concentration of salt in the remaining unfrozen moisture. We would expect treated soils to have less frost penetration than untreated soils. Although there is disagreement on the correlation between the frost penetration and the magnitude of frost heaving (Penner, 1963, Beskow, 1935, Taber, 1929) frost penetration in itself is of vital importance. Treated soils freeze after and thaw before untreated soils. The effect of artificial salting is equivalent to exposing the soil to a shorter freezing season which optimally can be of zero duration.

Supporting data was obtained experimentally in the present study by determining the depth of frost penetration at the end of the freezing tests. A regression analysis was performed on the data using two models: a linear and a second degree polynomial. The second order model provided a higher correlation in all cases. The results are shown in Table 1 and Figure 3.

Table 1 - Regression Analysis Using Second Degree Polynomial

Concentration	Equation of Best Fit	R^2
0.00N	$Y = -49.3 + 32.2\,x - 5.1\,x^2$	0.880
0.10N	$Y = -75.4 + 65.4\,x - 15.0\,x^2$	0.887
0.25N	$Y = -83.5 + 80.2\,x - 19.8\,x^2$	0.957
0.50N	$Y = -76.2 + 74.8\,x - 158.9\,x^2$	0.933

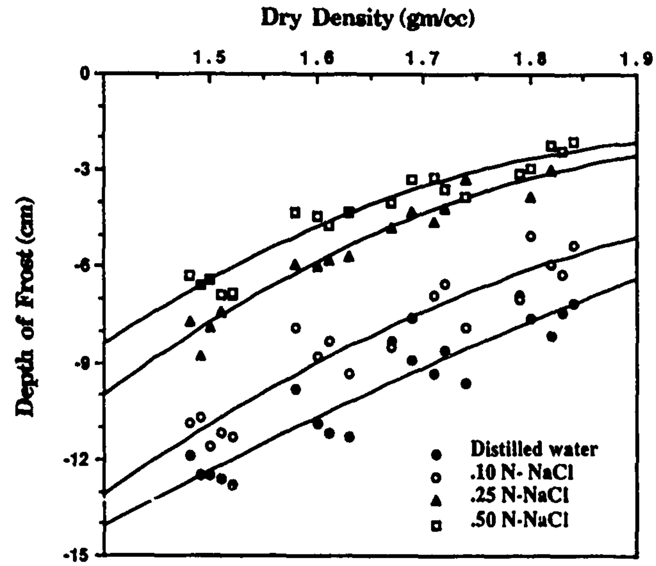

Figure 3. Depth of Frost Penetration

The data suggests that a non-linear relationship exists between frost penetration and the level of treatment. In other words, equal increases in the salt concentration does not yield equal reduction in frost heaving. Figure 4 also shows a slight tendency to less reduction in the magnitude of frost heaving at higher soil dry densities. At such densities, there seems to be more profound effect of variables other than salt concentration. Temperature profiles were also determined before test termination. Figure 4 shows the temperature profiles for specimens treated with different levels of salt concentration. Frost penetration in the specimens tested at higher salt concentration stopped at elevations of lower temperature than those treated at a lower level of salt concentration. Specimens treated at higher salt concentration also kept warmer temperature profiles. This is because specimens with higher concentrations have higher unfrozen water contents in the frozen zone (Chung 1978). Since both the specific volume and the thermal conductivity of a soil matrix increase with the increase in its water content (Crawford 1951), specimens with higher salt concentration have warmer temperature profiles.

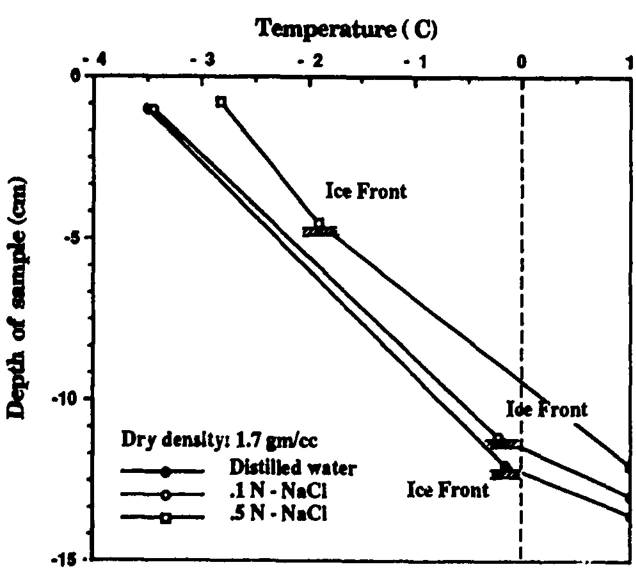

Figure 4. Temperature Profiles

Effect of salt treatment on moisture flow

An obvious method to combat the detrimental effects of frost heaving is to stop moisture in the soil mass from migrating upwards to an active frost front or a growing ice lens. This can be achieved by replacing the natural soil with clean gravel or crushed stones, which is quite an expensive solution to soil freezing problems. A more economical way of accomplishing this objective is by the use of chemical additives that reduce the ability of soil to transport moisture. Many investigators believe that artificial salting is a much less expensive way of accomplishing this objective (Sheeran et al, Yong 1975, Lamb 1971, Tsytovich et al, 1960). Salt addition to the pore fluid affects the soil's ability to transport moisture in its liquid and vapor forms through several mechanisms. As hypothesized by Taber and confirmed by the experimental work of other researchers, the most significant method of water transport upwards to an active frost front is through the unfrozen double-layer along the soil particle surfaces (Yanagisawa et al 1985, Miller, 1975, Jumikis, 1966, Penner, 1956, Taber, 1929). It is now well known that frozen and unfrozen water exists in fine-grained soils at temperatures well below the freezing point of the pure pore solution (Fukuda et al, 1985).

Addition of salt to the pore fluid reduces the thickness of the double-layer and, hence, reduces the thickness of the water film in contact with the frost front or a growing ice lens. The relationship between the thickness of the double-layer and salt concentration is given by the Goug-Chapman equation:

$$n_+ = n_o[\coth 0.16z\ (c_o)^{1/2}\ x]^2$$

where: n_+ = number of cations per unit volume at any distance x from the particle surface; n_o = number of cations per unit in the pore water away from the influence of the particle surface; z = valence of cations; c_o = concentration of cations in mols/liter away from the influence of the particle surface; x = distance from surface in Angstrom. The extent of the double-layer can be taken as the distance to which n_+ is still appreciably larger than n_o. The lower the concentration and the valence, the larger is the ratio n_+/n_o. Increasing either the concentration or the valence in the pore water would reduce the thickness of the double-layer.

Solutions of higher concentrations have lower vapor pressures. This is of great importance because the vapor pressure of ice is lower than that of pure water. Moisture will flow to the frost front and to ice lenses because of this pressure deficiency. Moisture flow will continue until the vapor pressure of the pore solution is reduced to that of ice under the effects of matrix, osmotic, or other forces acting within the soil mass. If the freezing process is started with a pore fluid which already has low vapor pressure (due to the presence of salt), the pressure equilibrium between ice and pore fluid is expedited. In other words, the amount of water that can be transported to the ice lenses due to the pressure difference is less and consequently the magnitude of surface heaving is reduced. Brine exclusion causes even further local increase in the salt concentration at the frost front which may accelerate the attainment of that equilibrium.

There are two secondary mechanisms by which artificial salting may reduce frost heaving. These are the effect of salt on the viscosity of the pore solution and on the capillary rise. The effect of reducing the thickness of the double-layer and the fact that solutions of higher concentrations have higher viscosities, may result in further reduction of moisture transport, and hence reduction in frost heaving. The effect of viscosity can be better appreciated if we remember that viscosity increases exponentially with decreasing temperature, and the temperatures considered in this study are in the subfreezing range.

Adding salts to the pore fluid also reduces the capillary rise in the soil mass. Figure 5 shows the relation between salt concentration and capillary rise for the tested soil. The reduction of soil capillary rise may cause partial desaturation in the active layer, thus reducing the soil permeability and the ability for moisture to flow.

Direct measurements of moisture plus ice content profiles for soil specimens treated at different levels of salt concentrations were made in the study. The data verifies the above conclusion and suggests that specimens containing higher concentration of NaCl experiences less moisture transfer (Figure 6). For instance, specimens treated at concentration of 0.5N and 0.1N experienced 13% and 86% moisture transfer, respectively, compared to the untreated specimens. Figure 7 shows the relationship between the magnitude and rate of frost heaving and the salt concentration. The magnitude and rate of frost heaving is inversely proportional to the salt concentration. A reduction of approximately 50% is achieved for the 0.5N level of treatment compared to the untreated specimens.

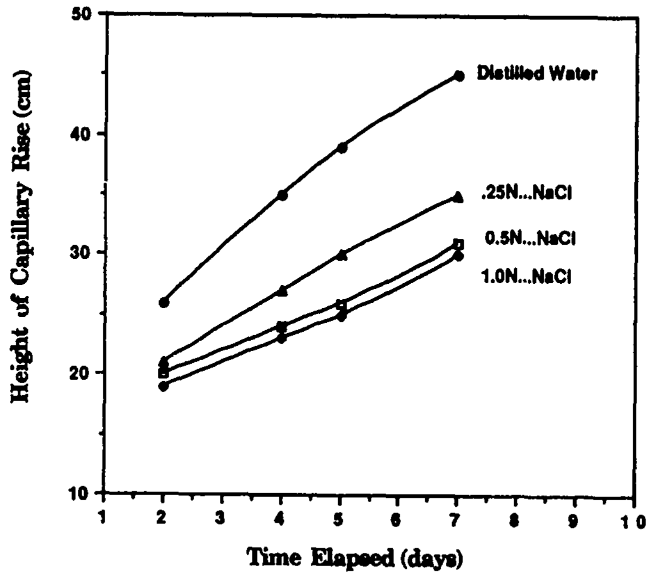

Figure 5. Capillary Rise

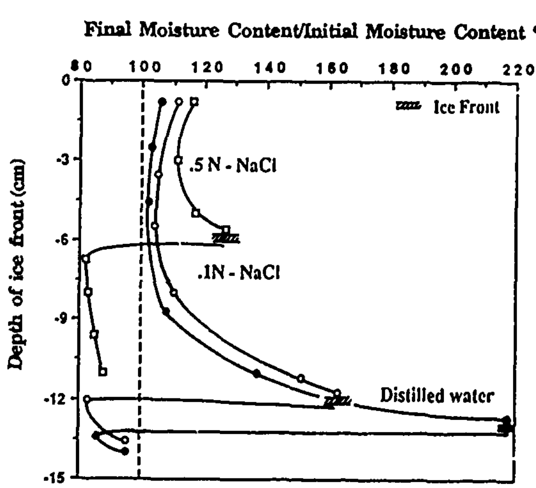

Figure 6. Moisture Transfer

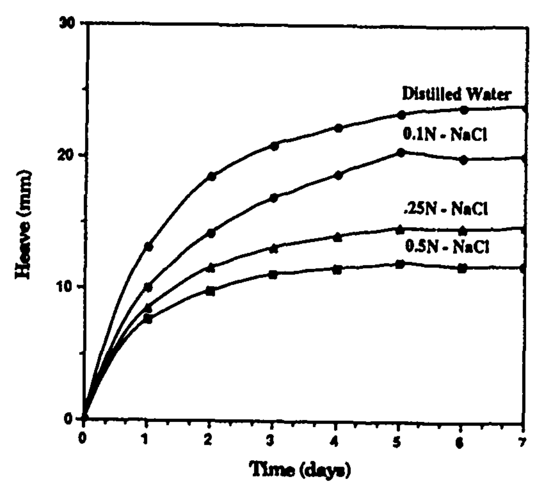

Figure 7. Frost Heaving

Interesting observation

The freezing tests were conducted on washed soils (free of natural salts) to determine the optimum level of salt concentration. The results show that every increase in salt concentration was accompanied by a reduction in the magnitude of frost heaving up to a concentration of 0.5N above which little or no effect on the magnitude of frost heaving was observed. However, an exception to this trend did occur. Specimens treated at a very low salt concentration (0.01N) heaved more than untreated samples. Figure 8 shows the results obtained using concentrations ranging from 0.0N to 1.0N.

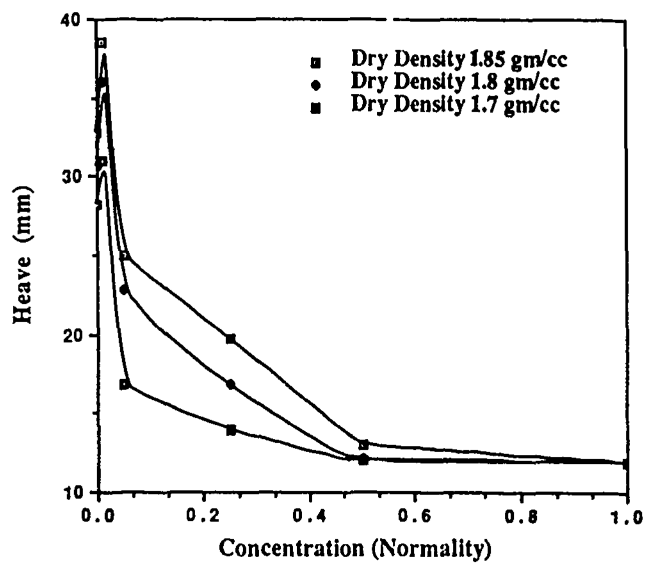

Figure 8. Frost Heave at Low Concentrations

A possible explanation for this exception is that the reduction of the double-layer thickness, due to the presence of salt, transferred some of the bounded water into free water in the pores. This free water was then readily frozen without much influence from the surface forces. It should also be noted that at such very low concentrations the counter effect of lowering the freezing point of the pore water is not significant.

Effect of salt valence on frost heaving

Specimens treated with $CaCl_2$ and $FeCl_3$ were tested under the same freezing conditions and compared to specimens treated with NaCl to determirc the significance of cation valence on the effectiveness of the treatment. Figure 9 shows the relative efficiency of each salt in reducing frost heaving. Figure 10 compares the moisture profiles and the frost penetration of specimens treated with the three salts. There seems to be a correlation between cation valence and both the depth of frost penetration and the amount of moisture transferred to the frozen layer. The Ca^{++} resulted in 17% more moisture transfer than Na^{+}, while Fe^{+++} resulted in 56% more moisture transfer. However, more work is certainly needed to draw a firm conclusion.

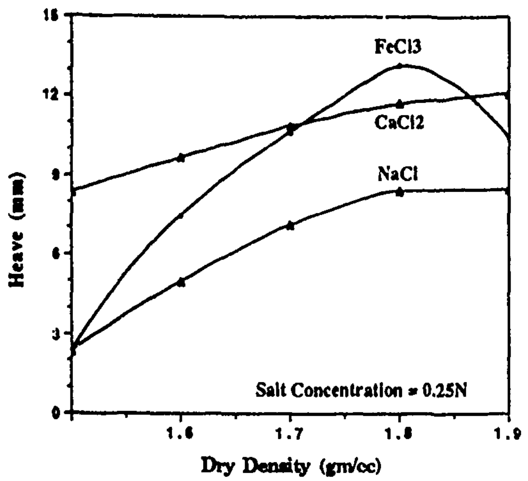

Figure 9. Heaving & Salt Valence

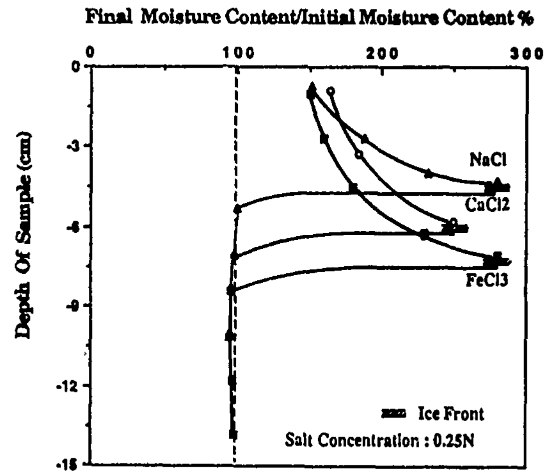

Figure 10. Mass Transfer & Salt Valence

CONCLUSIONS

The following conclusions can be drawn:

1. Artificial salting provided protection against the detrimental effects of frost heaving for the soil tested. Increasing the ion concentration in the pore fluid reduced the rate and magnitude of heave, limited frost penetration in the soil, reduced moisture transfer to the frozen layer, and kept the soil at a warmer temperature profile.
2. The protection provided was directly proportional to the concentration of the salt in the pore fluid. There is an optimum salt treatment for each soil which can be determined by laboratory freezing tests. For the soil tested and for the freezing conditions used that level was 0.5N.
3. A very low level of salt treatment may cause more frost heaving than in an untreated soil.
4. A correlation between the effectiveness of the treatment and the valence of the cation used in the treatment exists, but more research is needed.

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APPLICATION OF TIME DOMAIN REFLECTOMETRY TO MEASURE SOLUTE REDISTRIBUTION DURING SOIL FREEZING[1]

by W.K.P. van Loon, E. Perfect, P.H. Groenevelt and B.D. Kay[2]

INTRODUCTION

The influence of solutes on ground freezing is threefold. Firstly, the presence of solutes determines, in part, the freezing point depression (and thus the unfrozen water content) [Banin and Anderson, 1974; Yong et al., 1979]. Secondly, solutes oppose frost heave, since the osmotic pressure reduces the maximum heaving pressure [Cary, 1987; Chamberlain, 1983]. Thirdly, solutes are themselves redistributed during the freezing process. This is due to the exclusion of solutes by ice formation [Kadlec et al., 1988; Kay and Groenevelt, 1983; Mahar et al., 1983]. Solutes are also transported into the frozen fringe when water is drawn upwards to form ice lenses [Gray and Granger, 1986]. The net result is a concentration of solutes in the vicinity of the freezing front.

The redistribution of solutes during soil freezing is of interest to agronomists and environmental engineers seeking to minimize contamination of ground water supplies during winter. The phenomenon is also important for construction projects in areas of saline and subsea permafrost. It forms part of the matrix of coupled heat and mass transfer processes in frozen porous media [Perfect et al., 1989].

Different techniques have been used to monitor solute redistribution as a result of soil freezing. Starting with a known solute concentration profile, the most direct way is destructive sampling at the end of a freezing experiment [e.g. Cary and Mayland, 1972]. A major disadvantage with this method is that it is only possible to sample once per volume element, so changes over time are impossible to determine. Gray and Granger [1986] solved this problem in the field by sampling different volumes at different times and assuming that all samples had the same history. In the laboratory, Cary et al. [1979] extracted small amounts of pore water over time from different locations within a freezing column. The amount of dissolved salts was determined by measuring the electrical conductivity of these samples. Mahar et al. [1983] also used extracted soil water to monitor changes in salinity due to freezing.

In order to obtain a better understanding of the mass transfer processes, non-destructive in situ measurements are desireable. An applicable method is the measurement of bulk electrical conductivity (BEC). Baker and Osterkamp [1988] and Hayhoe and Balchin [1988] have used the in situ BEC as a measure of the solute concentration during soil freezing. However, the relationship between BEC and solute concentration is quite complex. The influence of liquid (unfrozen) water content is non-linear and the surface conductivity of soil particles must be taken into account [van Loon et al., 1989b].

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1 Presented at: Frozen Soil Symposium, March 21-22 1990, Spokane WA, USA

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Recently, time domain refectometry (TDR) has been used to measure BEC as well as the water content of unfrozen soils [Dalton et al., 1984]. It may also be possible to measure the BEC under frozen conditions using TDR [van Loon et al., 1989a]. Measurement of the unfrozen (liquid) water content of frozen soils using TDR is well established [Patterson and Smith, 1985]. Thus, TDR is a promising method for the simultaneous measurement of unfrozen water content and solute concentration (as indicated by BEC) in frozen soils.

In this research, unfrozen water content and BEC profiles were measured during freezing using the TDR technique. The major objective was to identify which parameters (solute concentration, unfrozen water content and temperature) the measured BEC is most sensitive to under frozen conditions. The possibility of using TDR to monitor solute redistribution during transient soil freezing will be examined.

MATERIALS AND METHODS

Uniaxial freezing tests were performed using twelve 270 mm diameter columns with a height of 180 mm. The soil used was Guelph silt loam, classified as an Aquic Eutrachrept. This material is described in van Loon et al., [1989a]. To obtain a more homogeneous medium the field soil was air dried and sieved. Different sieve sizes were used (Table 1). The columns were packed to specific bulk densities, depending upon the degree of aggregation (Table 1). The influence of aggregate size and bulk denisty on the freezing process will be reported in a future publication. Parallel TDR probes and thermocouples were installed horizontally into the dry soil at different heights (Fig. 1). These were allowed to move (in response to frost heave) by cutting a narrow slit in the columns at a height > 70 mm. All heights were measured relative to the base of the columns and are reported as positive upwards.

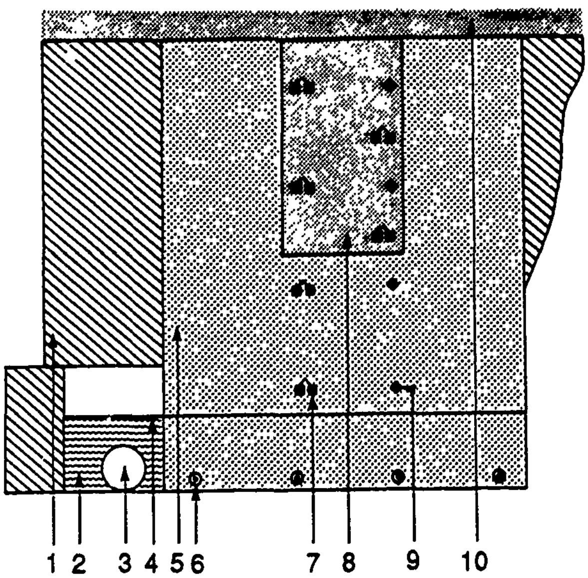

Figure 1 Cross section of column used for uniaxial freezing test. 1 - insulation, 2 - $CaCl_2$ solution reservoir, 3 - tubular copper heat exchanger (T=4°C), 4 - constant water table, 5 - soil column, 6 - water inlet with porous screen, 7 - parallel TDR probes, 8 - slit with bare soil, 9 - thermocouple and 10 - soil surface contacting air (T<0°C)

Table 1 Details of column packing and $CaCl_2$ solutions added.

Column no.	Sieve size [mm]	Bulk density [Mg/m^3]	Solution concentration [g/l]
1,2,3	10	1.15	1
4,5	5	1.31	1
6	crushed	1.39	1
7,8,9	5	1.31	2
10,11,12	5	1.31	4

All of the columns were placed into containers with different well defined $CaCl_2$ solutions (Table 1). Holes (with porous screens) at the bottom of each column allowed the solutions to wet the soil by capillarity. The fluid level was kept constant by adding solution of the same concentration to the reservoirs. The columns with their solute reservoirs were then set up in an environmental chamber, which could be maintained below freezing. The reservoirs were kept unfrozen by a heat exchanger, which was connected to a thermostatic bath (Fig. 1). Both columns and reservoirs were carefully insulated with styrofoam, so that the 0°C isotherm could only penetrate from the (upper) soil surface. Temperatures in the reservoirs and in the air (near the soil surface) were also measured with thermocouples. All temperatures were recorded automatically every six hours using a data logger. The TDR readings were taken once a day by hand.

The TDR method is based upon the reflection of an electromagnetic pulse within transmission lines, which are installed in the medium. By this means the (effective) dielectric constant (ϵ) of the medium can be determined. This value depends on the volume fractions of the components in the medium. Topp et al. [1980] obtained an empirical relation to convert the effective dielectric constant into the volumetric water content, θ_w:

$$\theta_w = -0.053 + 0.0292\epsilon - 5.5x10^{-4}\epsilon^2 - 4.3x10^{-6}\epsilon^3 \qquad (1)$$

Similarly, Smith and Tice [1988] developed an empirical relation between Ú and the unfrozen water content (θ_u) at subzero temperatures:

$$\theta_u = -0.146 + 0.0387\epsilon - 8.5x10^{-4}\epsilon^2 + 9.9x10^{-6}e^3 \qquad (2)$$

The bulk electrical conductivity of the medium can be obtained from the amplitude of the reflected pulse. However, this reflection is influenced by the measuring system, as well as by the medium itself. Van Loon et al. [1989a] have developed a method to correct for influences of the measuring system. In this method, a reflection measurement in soil (r) is compared with a reference measurement in air (r_0). The bulk electrical conductivity (BEC) is then given by the following equation:

$$BEC = \{\sqrt{\epsilon}/(N_0L)\}\ln(r_0/r) \qquad (3)$$

Where N_0=120π Ω and L is the length of the probes.

RESULTS AND DISCUSSION

Regression analyses were performed on the 915 data points obtained from the twelve columns (exceptions are always indicated). As a first approximation, the measured BEC was assumed to be a function of the solute concentration on a bulk basis. In this model, however, no interaction with the soil particles is included, so it can only be applied close to saturation. For $\theta_w \geq 40\%$ we tested the following model:

$$BEC = A + B.C.\theta_w \tag{4}$$

where A and B are regression coefficients and C is the solute concentration of the pore water. This simple model was highly significant: $R^2=0.88$. From the regression it follows that A=61 mS/m and B=80 mS m^2/kg.

The measured BEC in column number 12 was examined, as an example, to show the relationship between BEC and temperature and unfrozen water content.

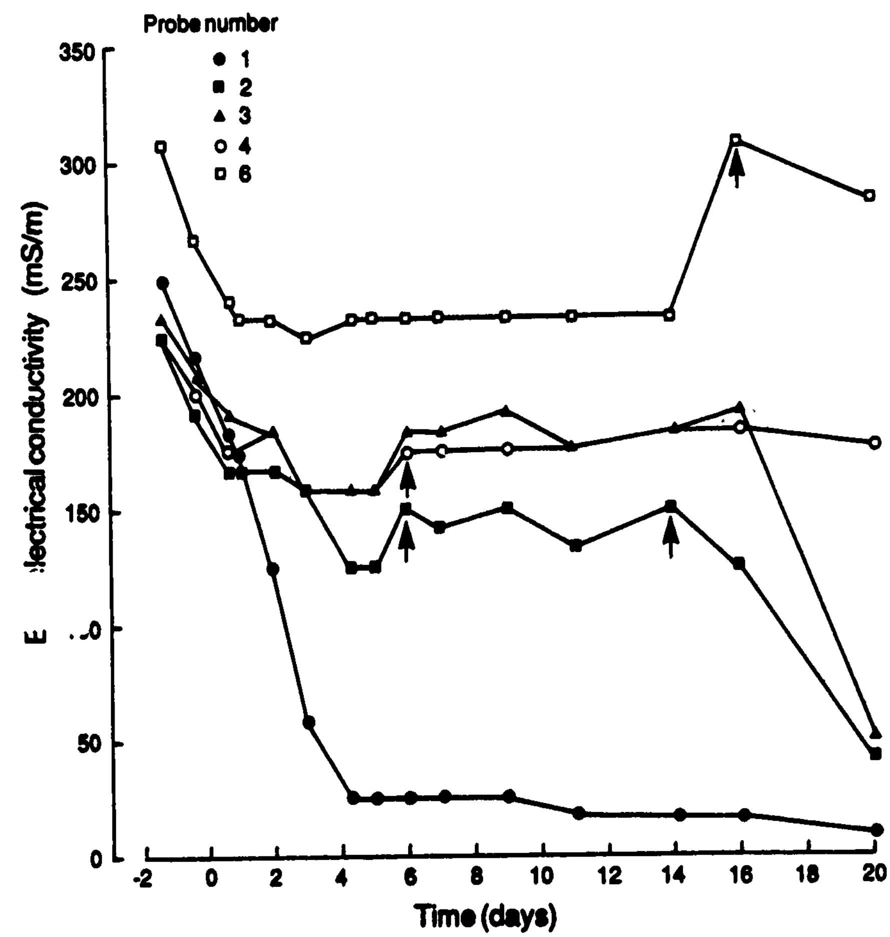

Figure 2 Measured bulk electrical conductivity as a function of time in column number 12 (with a solute concentration 4 g/l). Initial positions of TDR measuring probes: 1=160 mm, 2=140 mm, 3=120 mm, 4=100 mm and 6=40 mm. Arrows indicate significant increases in measured BEC.

Influence of temperature

Changes in BEC over time for five different TDR probes in column number 12 are shown in Fig. 2. Before time t=0 the climate room was cooled from 12°C to 4°C. At t=0 the temperature was dropped to -5°C. The first thing that happened was a sharp decrease in BEC for all five measuring probes (Fig. 2). At the same time the temperature in the soil went from 12°C to 1.5°C (averaged over column). The water content profile, which was also measured using TDR, did not change by more than 2% during this period (t=-1.2 to t=0.6 day). Thus, the measured BEC is temperature dependent. The next step is to include this temperature effect into the model for BEC, where the regression is again limited to water contents close to saturation (i.e. $\theta_w \geq 40\%$):

$$BEC = A + B.C.\theta_w + D.T \tag{5}$$

The R^2 for this model was 0.92 and the regression coefficients were: A=60 mS/m, B=79 mS m^2/kg and D=3.4 mS/°C/m. The inclusion of temperature was statistically highly significant. The physical meaning of a positive D coefficient is increased mobility of ions with higher temperatures. This effect is probably related to the viscosity of water. The electrical conductivity of a solution is inversely proportional to the viscosity of water, and the viscosity of water is directly temperature dependent [Weast, 1978]. Thus, if we express the measured BEC relative to a standard BEC at room temperature (i.e. 20°C), the following expression is obtained:

$$BEC(T) = BEC(20)\{viscosity(20)/viscosity(T)\} \tag{6}$$

In the temperature range -8°C < T < +30°C, the ratio of viscosities, V(T), can be modeled with a second order power expansion on T:

$$V(T) = viscosity(20)/viscosity(T) = A + B.T + D.T^2 \tag{7}$$

where A=0.5584, B=0.0193 $°C^{-1}$ and D=0.00014 $°C^{-2}$ with T in °C. The agreement between eqn. (7) and Weast [1978] is excellent: $R^2 > 0.99995$ with a standard error of 0.001.

The effect of viscosity can now be incorporated into a new regression model for bulk electrical conductivity:

$$BEC = A + B.C.\theta_w.V(T) \tag{8}$$

The regression coefficients for this model were A=61 mS/m and B=130 mS m^2/kg and the coefficient of determination was the same as that for the model which was extended to include the influence of temperature (i.e. R^2=0.92). However, eqn. (8) has two important advantages over eqn. (5): firstly, it uses fewer degrees of freedom, and secondly, it has a better physical background.

Influence of unfrozen water content

Another feature of Fig. 2 is the continued reduction in BEC at z=160mm (probe number 6) for t > 0.6 day. Between t=1 and t=4.2 day the measured BEC fell from 171 to 26 mS/m. During this time interval the temperature remained constant (within the measuring accuracy), while the liquid water content decreased from 0.41 to 0.12. This was probably due to ice nucleation and freezing of the pore water. It is unlikely, however, that the decrease in liquid water content (by a factor of 3.4) would cause a such large reduction in bulk electrical conductivity (a factor 6.6). In fact, the opposite might be true if solutes were transported into the freezing zone by thermally-induced water migration. Therefore, the reduction in BEC (if present at all) should be less than or equal to the reduction in liquid water content.

At low water contents the BEC is a non-linear function of the liquid water content [e.g. Hayhoe and Balchin, 1988]. Van Loon et al. [1989b] developed the following theoretical relation between BEC, solute concentration and liquid water content at room temperature:

$$BEC(20) = (1+\theta_w)\sigma_s + \theta_w^2(2-\theta_w)\sigma_w/(2-\theta_{sat})\theta_{sat} \tag{9}$$

where θ_{sat} is the water content at saturation, σ_s is the electrical conductivity of the soil particles and σ_w is the electrical conductivity of the pore solution. The relation between the latter property and solute concentration is simply: σ_w = QxC, with Q the specific electrical conductivity at 20°C.

Equations (6), (7) and (9) can now be combined to obtain a quasi-theoretical model for the bulk electrical conductivity at different temperatures, water contents and solute concentrations:

$$BEC(T,\theta_w,C) = \{(1+\theta_w)\sigma_s + \theta_w^2(2-\theta_w)\sigma_w/(2-\theta_{sat})\theta_{sat}\}V(T) \tag{10}$$

To generate an equivalent regression model the following assumptions were made:
(1) σ_s is unknown and can be found by regression (i.e. $A=\sigma_s$),
(2) the specific electrical conductivity for $CaCl_2$ is Q=161 mS m^2/kg [Weast, 1978],
(3) θ_{sat} is unknown, so a second regression parameter is needed (i.e. $B = 1/(2-\theta_{sat})\theta_{sat}$),
(4) eqn. (7) can be used for the viscosity effect,
(5) no intercept is needed in the regression, since it is included in the A coefficient.

The resultant regression model for BEC can then be written as:

$$BEC(T,\theta_w,C) = (1+\theta_w).V(T).A + \theta_w^2(2-\theta_w).Q.C.V(T).B \tag{11}$$

The results of the regression analysis are summarized in Table 2. It can be seen that the combined model for BEC(T,θ_w,C) was highly significant.

Table 2 Regression coefficients for the model: BEC(T,θ_w,C) = $(1+\theta_w)$.V(T).A + $\theta_w^2(2-\theta_w)$.Q.C.V(T).B (eqn. 11).

Data range	Number [-]	A [mS/m]	B [-]	R^2 [-]	SD* [mS/m]
ALL	915	37	1.24	0.985	13
T≤0°C	403	24	1.60	0.974	10
T≥0°C	614	42	1.17	0.990	13

* SD = standard deviation.

It is to be expected that the coefficients will be different for the frozen and unfrozen cases: in the frozen case $\theta_w = \theta_u$, the unfrozen water content, and in the unfrozen case θ_w equals the total water content. However, the influence of ice versus air on the measured BEC is surprisingly small. If all the data points are taken together, the model explains 98.5% of the variation in BEC. If only the data points for frozen conditions are considered, the R^2 decreases slightly. Conversely, if only the data points for unfrozen conditions are taken, the R^2 increases slightly (Table 2).

The particle conductivities (σ_s) given in Table 2 agree well (within the accuracy of the regression model) with the value for the same material at 20°C (σ_s=26 mS/m) estimated by Van Loon et al., [1989b]. The agreement is closest for the frozen material (σ_s=24 mS/m). In all probability, σ_s was determined less accurately in the unfrozen case, because the water contents were always rather high: θ_w=0.40 ± 0.10.

Since A was overestimated in the unfrozen case, B was probably underestimated. The B coefficient for the entire data set looks more reasonable. A value of 1.24 for B would give a saturated water content of 0.56, which is compatible with the lowest bulk density value in Table 1. The B coefficient for the frozen soil is larger than one would expect from the bulk density values given in Table 1. The higher B coefficient might indicate an increase in the solute concentration (C) due to ice formation; solutes are excluded from the ice and confined to the unfrozen water films under frozen conditions.

Hayhoe and Balchin [1988] gathered a similar data set using an AC conductivity signal conditioner module to measure the electrical conductance. This parameter can be transformed into BEC as follows: BEC=conductance/d, with d the spacing between the probes (i.e. d=50 mm). From Figures 1, 2 and 3 in Hayhoe and Balchin [1988] we obtained data sets of 33 points for sand, 26 points for loam and 19 points for clay. When these data sets were analyzed with our BEC model (eqn. 10), we again found very high coefficients of determination: R^2=0.96, 0.97 and 0.95 for the sand, loam and clay respectively. However, the model produced negative estimates of σ_s in each case, which is physically impossible. Inaccuracies in determining σ_s may be related to the restricted range of water contents available for analysis.

Redistribution of solutes

Apart from the initial decrease in BEC, Fig. 2 also shows some significant (i.e. > 13 mS/m) transitory increases in measured BEC. Temperature and unfrozen water content always decreased over time, so these peaks cannot be explained with the above model. Excluding measurement errors, an increase in BEC could only be caused by solute redistribution effects. Altogether 29 increases (some determined with more than one probe) were identified from the 12 columns over the 13 measuring dates. From these, two variables were evaluated: the amplitude of the increase (dBEC) and the mean height at which the increase occurred (Z). Parameters that might be important in the redistribution process include position of the 0°C isotherm (Z_0), initial solute concentration (C) and time (t).

Using these parameters the following regression model was developed for the increase in bulk electrical conductivity (dBEC):

$$dBEC = A + B.Z_0 + D.C + E.t \tag{12}$$

The regression coefficients were B=0.2 S/m^2 and D=4 mS m^2/kg; A and E were not significantly different from zero. The coefficient of determination was low (R^2=0.18) but statistically significant at $p < 0.15$. Because of the poor correlation, further conclusions are hard to make.

A similar model was developed for the location (Z) of the measured increases in BEC:

$$Z = A + B.Z_0 + D.C + E.t \tag{13}$$

where the regression coefficients are A=9 cm, B=0.5 and E=-6 mm/day, with D not significantly different from zero. This model appears to describe the measurements much better: R^2=0.77. We can use eqn. (13) to calculate the redistribution of solutes during soil freezing; predicted values for t=0 days, Z_0=18 cm and t=16 days, Z_0=8 cm are Z=18 and Z=3 cm, respectively. Thus, the solute enriched peak appears to have moved faster than and ahead of the 0°C isotherm.

Kadlec [88] suggests that 'the mechanism of solute transfer in the liquid phase during freezing in solutions include molecular diffusion and the brine sinking phenomenon'. The driving force for the latter process is the density difference between the higher concentration solution at the freezing front and the bulk solution. These remarks are for sea water. However, similar phenomena have been reported in soils, namely 'fingering' of pollutants. In all these processes the solute is allowed to move faster than the advancing freezing front. Thus 'brine sinking' or 'salt fingering' might explain our observations.

The velocity of the solute peak (E=-6 mm/day) is in good agreement with analyses performed by Kay and Groenevelt [1983]. These authors measured the redistribution of nitrate in the same Guelph silt loam soil under field conditions; the nitrate peak moved down 15 cm during the first 26 days of freezing. Their hypothesis that solute redistribution occurs on a local scale is also applicable to this research. The increases we measured were very small compared to the total BEC (see Fig. 2). Thus, only a small fraction of the solutes could have been transported over macroscopic distances.

CONCLUSIONS

Time domain refectometry can be used to measure the unfrozen water content and bulk electrical conductivity simultaneously under frozen conditions. The in situ bulk electrical conductivity depends on the solute concentration, unfrozen water content and temperature. The temperature effect can be related to the mobility of ions, through the viscosity of water. The influence of unfrozen water content on BEC is non-linear. Redistribution of solutes during soil freezing can also monitored using the BEC measurement. After temperature and water content corrections are applied, it is possible to identify increases in BEC associated with solute enrichment. These solute peaks appear to move faster than and ahead of the 0°C isotherm.

ACKNOWLEDGEMENTS

The authors wish to express their gratitude to I.A. van Haneghem, R.G. Kachanoski and I.J. van Weesenbeeck for scientific discussions and to A.M.J. van Hoof for participating in the measurements.

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MODELING OF SOLUTE REJECTION IN FREEZING SOILS

by

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INTRODUCTION.

The need to understand freezing processes in saline soil solutions has prompted several laboratory experiments and field tests. Kay and Groenevelt (1983) noticed solute redistribution during natural downward freezing in their field tests. However, upward unidirectional freezing in laboratory sand column tests did not indicate any solute exclusion from the frozen zone. This phenomenon was also observed in laboratory tests by Baker and Osterkamp (1988). Wilson and Vinson (1983) conducted laboratory tests on downward freezing of saline solutions in a granular soil. The degree of solute redistribution was noticed to increase at colder surface temperatures with decreasing soil solution salinity. Baker and Osterkamp (1988) and Kaldec (1984) further observed increased salt redistribution with decreasing freezing rates. Diffusion seemed to be a secondary process in solute movement. Mahar et al. (1982) report a few of their results obtained in unidirectional freezing tests of fresh water and saline water solutions. Mahar et al. (1983) attempted a numerical model to explain the freezing process they observed. Their model performed well for the distilled pore water case. However, the saline pore water comparisons with test data were not very favorable. The discrepancy seems to stem from modeling phase change at a discrete temperature, while for saline pore water solutions, freezing occurs over a range of temperatures and depths due to changing solute concentrations as a result of salt exclusion from the ice phase. Other mathematical modeling attempts of solute redistribution include the study by Osterkamp (1987) which considers a single heat balance equation to obtain an analytical solution for the temperature distribution. Cary and Mayland (1972) perform experiments to identify the significant transport parameters. The Soret effect (i.e., transport of solutes due to temperature gradients) was found to be inconclusive. Cary (1987) proposes a model to estimate the effects of solutes on heaving. Fuchs et al. (1978) note the importance of the osmotic potential, but neglect it in their effort to analyze the heat flow in freezing soils. Kadlec et al. (1988) developed a model to describe the solute segregation at the freezing front in a peatland. A critical review of studies on the processes occuring within freezing soils is presented by Kay and Perfect (1988) in a very comprehensive article.

Freezing and thawing of saline pore water solutions is a complex phenomenon. The freezing process excludes the impurities from the ice phase. As a result, salt concentrations in the unfrozen pore-water increase as freezing progresses (Hallet, 1978). This elevated salt concentration further lowers the freezing point of the liquid water. The influence of salt concentrations on unfrozen water content has been studied by Banin and Anderson (1974), and Yong et al. (1979). With ice pressures being limited by the total overburden, the capillary pressure effects between the liquid water and the ice create water pressures which are driving forces for saline pore-water flow. Increased solution concentrations due to salt rejection upon freezing, further cause osmotic potentials creating additional driving forces. Diffusion and dispersion also contribute to salt migration. Ice lenses are formed if a sufficient amount of mobile water exists to feed them. In this study we will assume a rigid soil column. Energy transport is governed by conduction, convection, and phase change. Phase change also indirectly

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Presented at the Symposium on Frozen Soil Impacts on Agricultural Lands, Spokane, Washington, March 21-22, 1990.

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Table 1. Model Equations

Conservation of mass equation of unfrozen water phase

$$\frac{\partial}{\partial t}(nS_w) + \frac{\rho_i}{\rho_w}\frac{\partial}{\partial t}(nS_i) - \frac{\partial}{\partial z}\left[\frac{K}{\rho_w g}\frac{\partial}{\partial z}(p_w - \rho_w gz)\right] = 0 \qquad (1)$$

Conservation of mass equation of dissolved salt in the water phase

$$\frac{\partial}{\partial t}(nCS_w) - \frac{\partial}{\partial z}\left[D\frac{\partial C}{\partial z}\right] - \frac{\partial}{\partial z}\left[\frac{CK}{\rho_w g}\frac{\partial}{\partial z}(p_w - \rho_w gz)\right] = 0 \qquad (2)$$

Conservation of energy equation

$$\frac{\partial}{\partial t}((\rho C)_m T) - L\rho_i\frac{\partial}{\partial t}(nS_i) - \frac{\partial}{\partial z}\left[\frac{C_w TK}{g}\frac{\partial}{\partial z}(p_w - \rho_w gz)\right]$$

$$- \frac{\partial}{\partial z}\left[\lambda_m\frac{\partial T}{\partial z}\right] = 0 \qquad (3)$$

Definitions of the heat capacity and the effective thermal conductivity of the soil.

$$(\rho C)_m = \rho_w C_w n S_w + \rho_i C_i n S_i + (1-n)\rho_s C_s ,$$
$$\lambda_m = nS_w\lambda_w + nS_i\lambda_i + (1-n)\lambda_s \qquad (4)$$

Clapeyron equation

$$\frac{p_i}{\rho_i} - \frac{p_w}{\rho_w} + \frac{RTC}{M\rho_w} + \frac{L}{T_o}[T - T_o] = 0 \qquad (5)$$

Expressions for the characteristic retention curve, the definition of capillary pressure, and a constraint

$$S_w = f(p_c) \qquad p_c = p_i - p_w \qquad S_w + S_i = 1 \qquad (6)$$

affects energy transport by changing the thermal properties (i.e., conductivity and heat capacity) of the medium.

Mathematical model

In this study, we present a mathematical model to model the salt rejection process in freezing saline soils.

Quantification of processes occuring during the freezing and thawing of air free soils will be presented by employing the conservation of mass equations for the solute and the unfrozen water and the conservation of energy equation. In Table 1, equations (1), (2), (3), (5), (6a), (6b), and (6c) are a set of seven equations in the seven unknowns S_w (degree of unfrozen water content), S_i (degree of ice content), p_c (capillary pressure), p_i (ice pressure), p_w(pressure in unfrozen water phase), C (salt concentration) and T (temperature), and can be solved by using appropriate initial and boundary conditions.

Table 2. Model Parameters

$n=$	0.256
$R=$	8.3 $J/mol./K$
$T_0=$	273 K
$M=$	0.05844 $kg.\ mol.$ [for Na Cl]
$\rho_w=$	1000 kg/m^3
$\rho_i=$	920 kg/m^3
$\rho_s=$	2500 kg/m^3
$g=$	9.81 m/s^2
$C_w=$	4180 $J/kg/K$
$C_i=$	2056 $J/kg/K$
$C_s=$	940 $J/kg/K$
$L=$	335,000 J/kg
$\lambda_w=$	0.59356 $J/m/s/K$
$\lambda_i=$	2,253 $J/m/s/K$
$\lambda_s=$	14.5 $J/m/s/K$
$D=$	$[7.0\ 10^{-9}\ nS_w]\ m^2/s$ with tortuosity accounted for.
$K=$	$3.8\ 10^{-5}\ S_e^2\ m/s$ where S_e = Effective Saturation.
$S_e=$	$\dfrac{1}{1+[3.9246\ 10^{-7}p_c]^6}$
$S_w=$	$0.1 + 0.9\ S_e$

The resulting equations are solved by using a Newton-Raphson linearization, and implicit iterative procedure. The details of the numerical solution are given in Panday and Corapcioglu (1990).

The test case simulated is of a 30 part-per-thousand saline solution in a 25 inches long sand column with properties given in Table 2. Figures 1 and 2 show temperature and concentration profiles over the freezing period. The temperature profile decreases the curvature near the base after around 71 hours indicating that the sample froze to the base as observed in the experiment. The curves given in Figure 2 indicate concentration peaks occurring in the column. This observation initiated a numerical experiment wherein the diffusion coefficient was lowered by two orders of magnitude and surface freezing rates were increased. At sufficiently high freezing rates and/or low diffusion coefficients, it was noticed that such concentration 'bumps' are trapped at a fixed location surrounded by higher ice saturations on either side. Such phenomenon have been observed at a field scale level (Kay and Perfect, 1988), and are attributed to the lower freezing point of the concentrated solution, since the concentration increase due to freezing is not dissipated off into the soil. The temperatures however, are low enough to cause more freezing in the less concentrated solution further below.

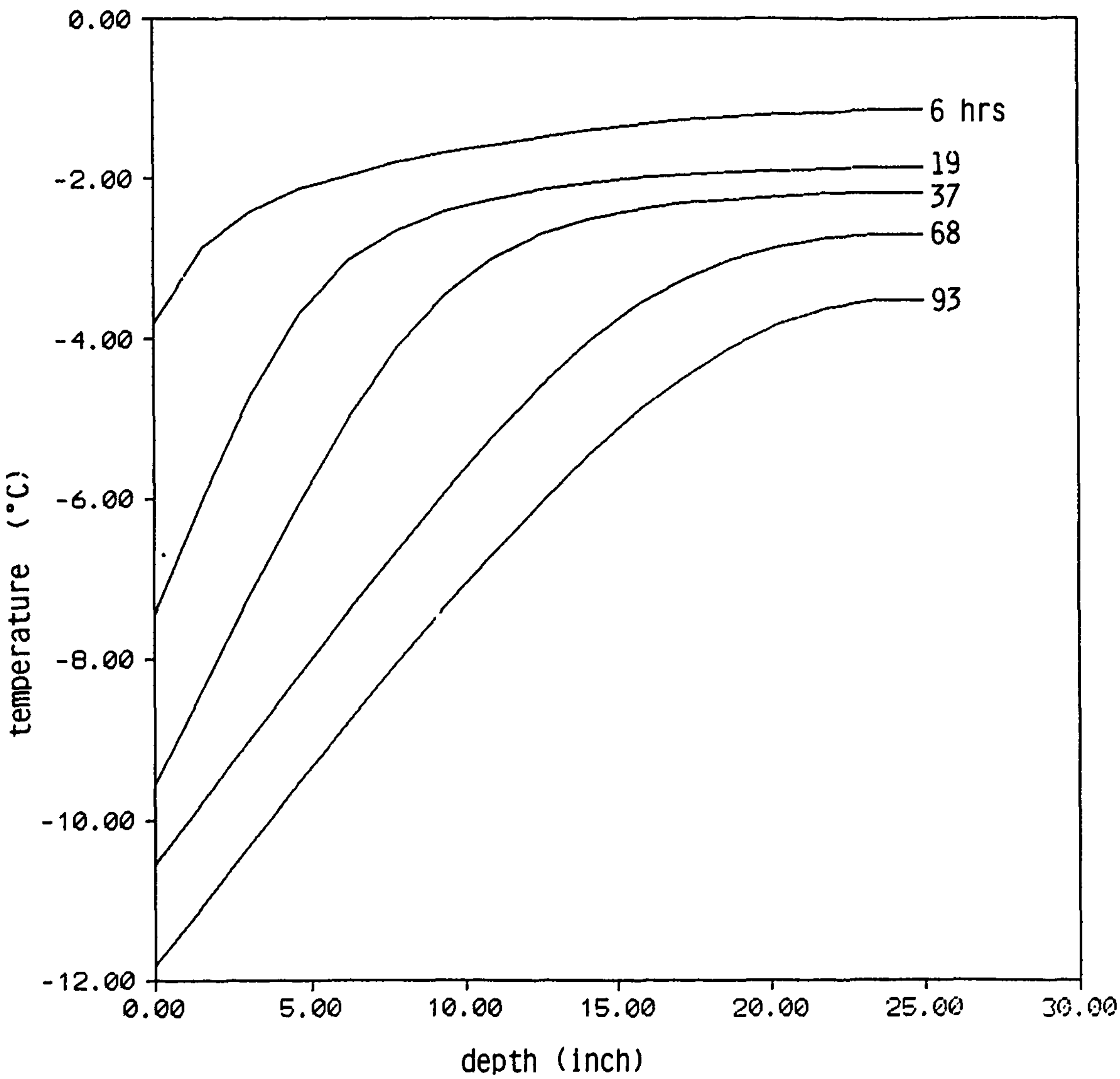

Figure 1. Variation of temperature along the soil column at different times. Initial salinity is 30 ppt.

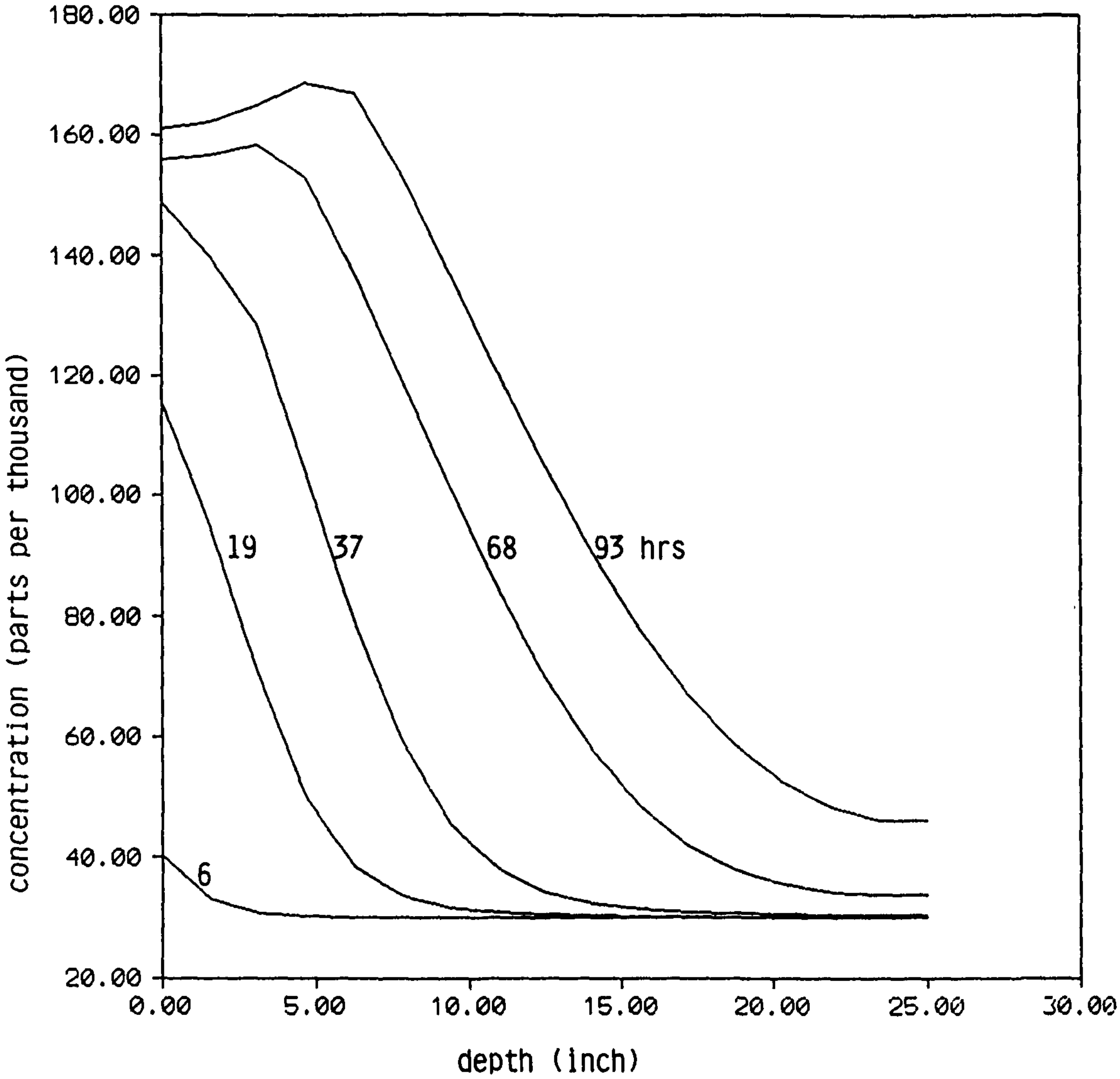

Figure 2. Variation of concentration along the soil column at different times. Initial salinity is 30 ppt.

List of Notations

C	concentration of solute
C_i	heat capacity of ice
C_s	heat capacity of soil grains
C_w	heat capacity of water
D	hydrodynamic dispersion coefficient
g	gravitational acceleration
K	hydraulic conductivity
λ_i	thermal conductivity of ice
λ_s	thermal conductivity of soil grains
λ_w	thermal conductivity of water
L	latent heat of fusion for water
M	molecular weight of the solute
n	porosity
ρ_i	ice density
ρ_s	density of the soil solids
ρ_w	water density
p_c	capillary pressure
p_i	pressure in the ice phase
p_w	pressure in the unfrozen water phase
R	universal gas constant
S_i	ice saturation
S_w	degree of unfrozen water saturation
t	time
T	temperature
T_0	freezing point of the pure solvent
z	vertical coordinate (positive downward)

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FATE AND TRANSPORT OF CONTAMINANTS IN FROZEN SOILS

O. A. Ayorinde* and L. B. Perry*

INTRODUCTION

The transport of organic compounds through soils (especially frozen soils) is complex and is governed by many physical, chemical and biological processes which are not yet fully understood. Prior to the current CRREL research program on the use of artificial freezing for immobilization and decontamination in soils, few, if any, research studies had been conducted on the fate and transport of explosive-based contaminants in frozen soils.

The manufacture, use, and disposal of organic-based explosives at several military bases and U. S. Army ammunition plants have resulted in the contamination of soils and sediments with explosive residues. Some of the explosive compounds include 2,6-DNT, M-NT and O-NT. One of the environmental problems and challenges facing most industrialized countries, including the U.S., is to develop a cost-effective and safe method of controlling, treating and disposing of explosive hazardous/toxic wastes. In particular, the transport of the explosive contaminants in soils and the potential contamination of ground and surface water by these explosive residues is a major concern to the Army. Iskandar and Houthoofd (1985) and Iskandar and Jenkins (1985) reviewed and summarized available techniques for remedial action at uncontrolled hazardous waste sites, and concluded that current techniques are not adequate and cost-effective for protecting ground and surface water.

In order to fill this gap, there is an urgent need for fundamental research on other possible innovative and cost-effective techniques to treat and control hazardous/toxic wastes. Iskandar et al. (1986) and Ayorinde et al. (1988) have evaluated laboratory methods for the potential use of artificial freezing to decontaminate soils contaminated with volatile organic compounds (VOCs) and their data indicated moderate success. Perry and Ayorinde (1988) and Taylor (1988) have experimentally demonstrated that different types of explosive residues can be effectively moved in water by freezing.

The objective of this investigation was to evaluate the effect of freezing on the fate and transport of 2,6-DNT, O-NT and M-NT explosive residues in soils. This paper describes (a) the development of experimental methods for obtaining reliable data that can be used to model freezing-induced transport of contaminants in soils and (b) the analytical approach used to interpret these data.

MATHEMATICAL FORMULATION

Most theoretical and experimental studies on solute or contaminant transport have focused on transport processes in repacked laboratory soil columns assumed to be relatively uniform. Transport in such nonfreezing soils during a steady state flow is generally predicted fairly well with an advection- or convection-dispersion solute transport equation (Freeze and Cherry, 1979):

$$R(\partial c/\partial t) = D(\partial^2 c/\partial z^2) - v(\partial c/\partial z) \qquad (1)$$

where

c = dissolved contaminant concentration (kg/m^3)
R = retardation factor

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D = dispersion coefficient (m^2/sec)
v = average pore water velocity (m/sec)
z = distance (m)
t = time (sec).

It should be noted that Eq. (1) was derived based on the assumption of a homogeneous porous medium under saturated conditions for which

$$\partial\Theta/\partial z = 0 \tag{2}$$

where

Θ = volumetric pore water content (m^3/m^3).

To better understand the transport of solutes or contaminants in nonfreezing soils, Eq. (1) is commonly used to analyze the breakthrough curve (BTC) data obtained in a soil column experiment. However, due to the assumptions of soil or porous medium homogeneity and full saturation, Van Genuchten and Wierenga (1977) and Yu et al. (1984) have shown that, in most cases, the experimental BTC tends to drop more slowly than Eq. (1) predicts. Nonetheless, BTC data can provide an insight into the fate of the contaminants as they move through soils. Hence, BTC data were generated for 2,6-DNT, O-NT and M-NT explosives during the soil column tests in which freezing was used to move these contaminants in Lebanon silt. In this paper, no attempt was made to use Eq. (1) to predict and model the experimental BTC data.

EXPERIMENTAL APPROACH AND PROCEDURES

Sample Preparation

A detailed description of the test apparatus and the procedures for sample preparation and preparation for contaminant spike solution is given by Ayorinde et al. (1989). Four sample soil columns were prepared using Lebanon silt and a 7-cm-OD by 0.325-cm-thick by 12.5-cm-high Lucite cell into which the Lebanon silt was carefully packed. A spike solution of the explosive contaminants was made by adding 2,6-DNT, O-NT and M-NT explosive residues to 200 mL of Milli-Q water, and the resulting solution was poured into a 2-L carboy in which the solution was mechanically mixed for about 24 hours. Before backsaturating the soil columns with the spike solution, the contaminant concentration was measured for each of the explosive contaminants. The expected and measured concentrations for each of the explosive residues were 20 mg/L and 10 mg/L, respectively. The difference was probably due to biodegradation that took place before the backsaturation was performed.

One column was used as a control (column E) while the other three columns (columns F, G and H) were used as replicates for the explosive-contaminated soil. Each Lucite column was packed with oven-dried Lebanon silt to an average density of 1.67 g/cm^3 and a porosity of about 36%. The packing was done in 2-cm layer lifts to achieve relatively uniform density throughout the column. Once the test cell was full, an end cap was placed with the filter on top of the soil, and the cell was sealed. Both the bottom inlet and the top outlet were then attached with three-way control valves fitted with Teflon tubing. A typical Lucite test cell used for the experiment is shown in Figure 1 with a soil sample height of 12 cm.

Backsaturation Method and Breakthrough Curve Procedure

To obtain a reasonably uniform distribution of the contaminant concentration through each soil column prior to freezing, each column was backsaturated slowly with 10 pore volumes of spike solution. An increment of 0.5 pore volume was used until a total of 10 pore volumes was reached. The backsaturation apparatus used is shown in Figure 2. The burette with the spike solution was connected to the Lucite

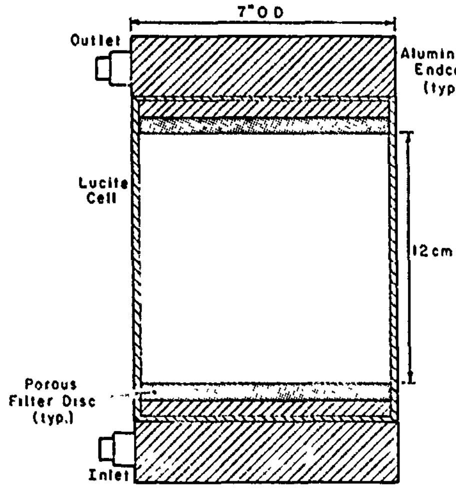

Figure 1. Typical Lucite soil test cell

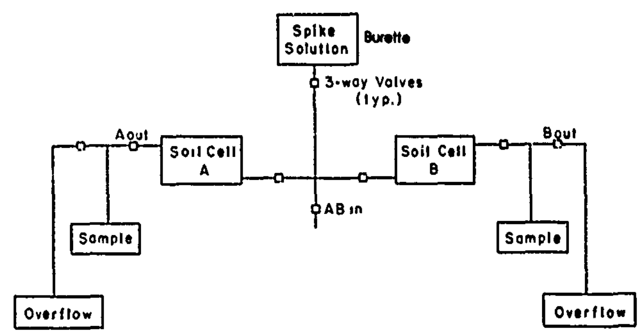

Figure 2. Backsaturation apparatus.

soil columns using the Teflon tubing to form a closed loop as shown in Figure 2. The burette was placed high above the soil columns to provide enough head to move the spike solution very slowly through the soil from the bottom up. Two soil columns were simultaneously backsaturated at each time. During the backsaturation of a new set of soil columns, previously saturated columns were kept inside the environmental chamber. The incremental volumes of the effluent solution collected from each column outlet at known periods of time were carefully measured.

For each incremental pore volume, the inlet solution at the bottom of each soil column and the outlet solution at the top of the column were continuously monitored and collected for chemical analysis. HPLC was used to measure the concentration levels of the explosive contaminants in the collected solution based upon the procedure developed for determining nitroaromatics and nitramines (Jenkins et al., 1988). As shown schematically in Figure 2, the inlet solution from the burette is identified as AB_{in}, while the effluent solution coming out of samples A and B are identified as A_{out} and B_{out}, respectively. The concentrations of A_{out} and B_{out} were averaged and used to generate the BTC for each contaminant.

After the flow of 10 pore volumes of the spike solution, the soil was assumed to be fully saturated, and both the inlet and outlet valves were closed. The soil sample columns were then placed in the environmental chamber for subsequent one-dimensional freezing from the bottom up.

Sample Freezing and Sampling Methods

Ayorinde et al. (1989) provides a detailed description of the sample freezing and sampling methods used in this experiment. The environmental chamber that housed the four explosives-contaminated saturated soil columns was maintained at +1.0°C throughout the experiment. Prior to freezing, all the soil columns were equilibrated at +1.0°C for 48 hours. Soil columns F, G, and H were frozen bottom up by means of a large cooling plate inside the environmental chamber, while the control column E was kept unfrozen throughout the test. Using another large cooling plate placed on top of the four soil columns, the top end boundary temperatures of the columns were maintained at +1.0°C.

The freezing of the soil columns was gradual and was from bottom up in order to eliminate the possible effect of gravity on the mobility of contaminants. Using the bottom plate cooling rate of about 0.4°C per day to freeze the soil columns, an ap-

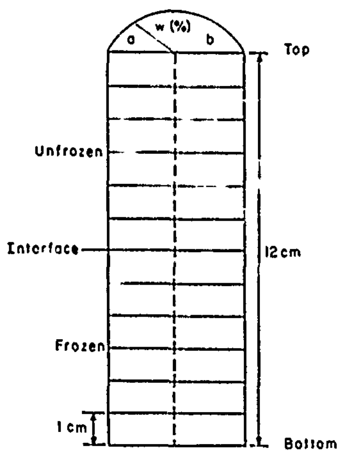

Figure 3. Sampling layout of half soil column after flash-frozen in liquid nitrogen.

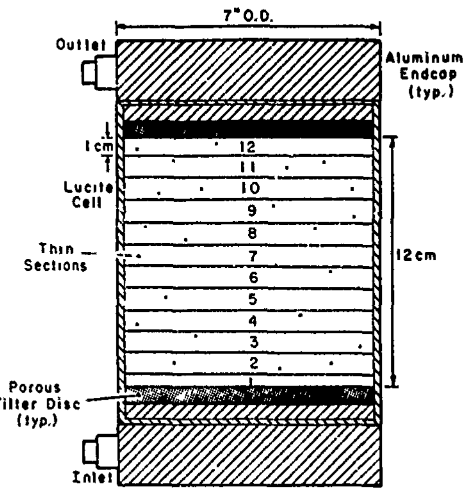

Figure 4. Test cell soil sampling scheme with thin section dimensions.

proximate freeze rate of 0.5 cm/day was achieved. Because of the good temperature control of the environmental chamber, one-dimensional vertical freezing of the soil columns was also achieved. Columns F, G and H were frozen approximately halfway (6 cm) from the bottom.

At the end of the freezing process, all soil columns were taken out of the environmental chamber and flash-frozen in liquid nitrogen. The soil samples were thin-sectioned to provide estimates of the soil solute concentration profile. To do this, frozen soil columns were transferred to a coldroom and cut vertically in half as shown in Figure 3. One half was kept in the freezer, while the other was cut into 1-cm-thick sections on a band saw as depicted in Figure 4.

Two sample duplicates of each thin section were obtained, representing about 2/3 of each thin section made from 1/2 of the original sample. The remaining 1/3 of the thin section was used for estimating the sample moisture content profile. The thin-section sample duplicates were placed in sealed glass scintillation vials to be extracted for chemical analysis. About one milliliter of the extract was then transferred into mini-autosampler vials for direct concentration determination using HPLC. By this procedure, the concentration profile of the contaminants along each soil column was obtained.

ANALYSIS OF TEST DATA

Actual theoretical analysis of the BTC data or the freezing-induced explosive contaminant transport in Lebanon silt was not performed in this paper. Such an analysis for the BTC data would require the use and solution of the advection-dispersion equation given by Eq. (1) with sorption and biodegradation terms. For the freezing-induced transport of the explosive contaminants, Eq. (1) with sorption and biodegradation terms would need to be coupled with the thermodynamic energy equation involving heat transfer with phase change. The resulting set of equations would be very complex and difficult to analyze. However, Eq. (1) can be used to explain the BTC characteristics for each explosive contaminant employed for the test.

Breakthrough Curve Data Analysis

During the transport of contaminant or solute in soils, the solute movement is usually slowed as a result of sorption or ion exchange, precipitation, or other geochemical reactions. The slowing of solute movement in soils is referred to as "retardation." For the BTC analysis, the retardation coefficient (R) in Eq. (1) for each type of the explosive contaminant used can be obtained from the following equation (Freeze and Cherry, 1979):

$$R = 1 + K_p(\rho_d/\Theta) \tag{3}$$

where ρ_d is the dry bulk density of the soil (g/cm^3), Θ is the volumetric water content (cm^3/cm^3), and K_p is the partition coefficient (cm^3/g), which relates mass of contaminant sorbed per unit mass of soil to the concentration of contaminant in solution at equilibrium. The dry bulk density was 1.67 g/cm^3, Θ was obtained by dividing the average soil moisture content by the volume of the Lucite test cell shown in Figure 1. However, K_p (not calculated in this paper) has to be determined separately from typical adsorption tests.

Breakthrough curves are often plotted with pore volumes (PV) of flow on the abscissa and the relative concentration (C/C_0) of the contaminant on the ordinate, as shown in Figure 5. C_0 was obtained by averaging each explosive initial concentration determined from the incremental inflow pore volume. PV represents the cumulative quantity of inflow divided by the volume of the void space (V_v) in the soil. V_v is equal to the soil column average porosity of 0.36 multiplied by the volume of the soil column, using the soil column dimensions of 12 cm high and 7 cm in diameter in Figure 1.

A nonfreezing-induced advective transport of contaminant through soil can be assumed to occur according to Darcy's law:

$$Q = kiAt \tag{4}$$

where Q is the volume of liquid flowing through a soil column with a hydraulic conductivity k and a cross-sectional area A over a period of time t, when the hydraulic gradient that drives the flow through the soil column is i. By carefully monitoring accumulative Q with time, BTCs were generated. For our tests, there was a pressure difference of about a 50-cm head of water across the 12-cm-high soil column. Since 1-cm head of water corresponds approximately to 9.806×10^{-2} kPa pressure, then a 50-cm water head yields a pressure differential of about 5 kPa. Hence, the hydraulic gradient is approximately 0.4 kPa/cm. From Eq. (4), values of the hydraulic conductivity can be estimated, and a plot of the relative concentration (C/C_0) versus hydraulic conductivity (k) can be obtained. This was not developed in this paper. Daniel et al. (1988) used the C/C_0-versus-k plot to estimate the "effective porosity" of a clayey soil. By definition, effective porosity is a measure of the volume of the soil pores that conduct most of the flow divided by the total volume of the soil, since pores of various sizes and degrees of connectivity exist in soils.

Freezing Soil Column Data Analysis

A detailed analytical procedure used to interpret the soil freezing data is given by Ayorinde et al. (1989). As noted earlier, the purpose of the experimental investigation was to test the hypothesis that freezing can move explosive contaminants in soils just as it has been demonstrated for water (Perry and Ayorinde, 1988; Taylor, 1988). For our investigation, two simple analytical approaches were adopted to analyze the measured concentration data.

The first approach compared the average of the concentration profile for the frozen bottom half (0-6 cm) of the freezing-treated columns (columns F, G and H) with the corresponding average concentration for the bottom half (0-6 cm) of the unfrozen control column (column E). It was pointed out earlier that the columns F, G and H, which were subjected to freezing, only froze to about 6 cm from the cold bottom ends of the columns. The bottom-half average calculated concentration was each normalized with respect to the mean concentration value (C_{tmean}) for the whole height of the unfrozen control column E. Normalization was done in an attempt to minimize inherent variability in the measured data arising from unavoidable variations in measurement, sampling, sample preparation and soil heterogeneity. The difference between the frozen column normalized concentration and that of the unfrozen control column was expressed as the percent normalized reduction in concentration as shown in Table 1.

The second analytical method used the Student t statistical significance test to compare (Bauer, 1971; Draper and Smith, 1976) the regression slope of the concentration profile for the unfrozen control column with that of the average concentration profile for the three frozen columns for each analyte. All the data for the three replicate frozen columns were combined. The combined data were least-squares fitted to estimate the frozen slope. Duplicate values at each location were used for the control column.

Table 1: Approximate changes in explosive contaminant concentration in Lebanon silt presumably induced during freezing* (approx. freezing front location = 6 cm from bottom end).

Type of contaminant	Distance from sample Bottom (cm)	Unfrozen control sample conc. (C_t)	Norm. avg. control sample conc. (C_t/C_{tmean})	Avg. frozen sample conc. (Cf)	Norm. avg. frozen sample conc. (C_f/C_{tmean})	Norm.* reduction in conc. ($(C_t-C_f)/C_{tmean}$) (%)
2,6-DNT	Bottom-Half (0-6 cm)	5.86	1.05	4.96	0.89	16.10
O-NT	Bottom-Half (0-6 cm)	3.21	1.07	2.00	0.67	40.36
M-NT	Bottom-Half (0-6 cm)	4.04	1.08	2.65	0.71	37.19

* Calculated concentration changes were obtained by dividing each column into two equal segments, and by averaging concentration profile over the bottom frozen half segment.

Table 2: Comparison between the regression slopes of contaminant concentration profiles for control, and combined frozen silty soil columns using t-test significance analysis

Type of contaminant	Control slope µg/g per cm	Combined frozen slope µg/g per cm	t-test signif. level (%)	Octanol-water partition coefficient (K_{ow})*	General comment
2,6-DNT	-8.92E-02	-2.99E-02	76.0	97.0	NS; < 95%
MNT	-1.07E-01	-6.70E-03	97.5	263.0	S; > 95%
ONT	-7.85E-02	-1.17E-03	96.5	199.5	S; > 95%

S = Statistically significant difference between slopes

NS = No statistically significant difference between slopes

* Values obtained from Hansch and Leo (1979), Leggett (1985) and Jenkins (1986).

For each analyte, a value of t was calculated based on the slope comparison analysis, and the significance level of the difference between the slopes was obtained from the Student t table. A significance level of less than 95% was considered insignificant for any difference between the slopes for the control and frozen sets of data. A probability level of equal to or more than 95% was considered significant. The results of the comparison between the regression slopes for the frozen and control concentration profiles are summarized in Table 2.

RESULTS AND DISCUSSION

The experimental breakthrough curves for 2,6-DNT, O-NT and M-NT are shown in Figures 5, 6 and 7. These curves depict the increasing relative concentration (C/C_0) with increase in pore volume of flow (PV). Combined plots of the BTCs for the three explosive contaminants are also shown in Figures 8 and 9. Figure 8 and Figure 9 indicate that breakthrough occurred earlier for M-NT than for both O-NT and 2,6-DNT. Both 2,6-BNT and O-NT appeared to break through at the same time around two pore volumes, while M-NT broke through at about one-and-half pore volumes. This strongly suggests that M-NT is more mobile in Lebanon silt than either O-NT or 2,6-DNT.

Figure 10 shows a composite of the soil moisture content profiles for unfrozen control sample and the three replicate frozen samples used as triplicates involving the explosive contaminants. The moisture profiles for these four samples appeared to agree very well within the measurement accuracy, in spite of several variabilities inherent in a complex nonhomogeneous system such as soil.

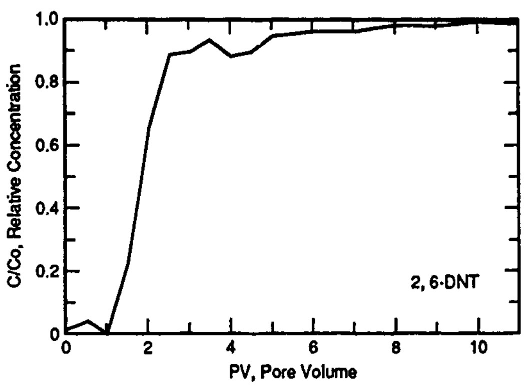

Figure 5. Breakthrough curve for 2,6-DNT.

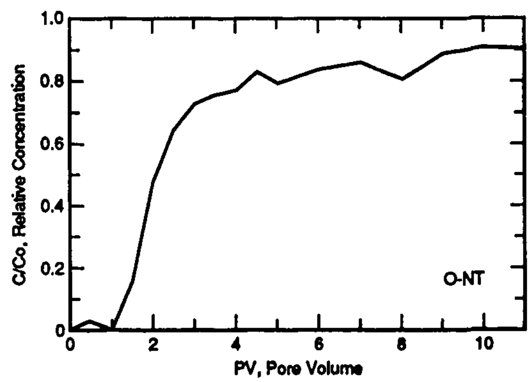

Figure 6. Breakthrough curve for O-NT.

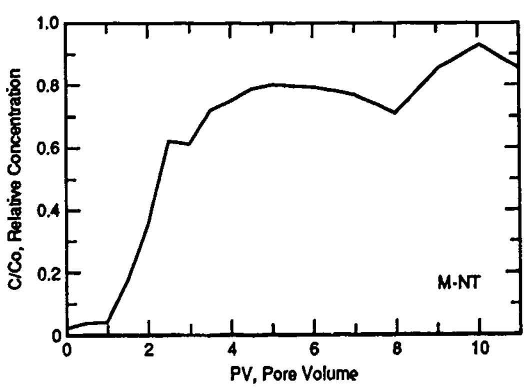

Figure 7. Breakthrough curve for M-NT.

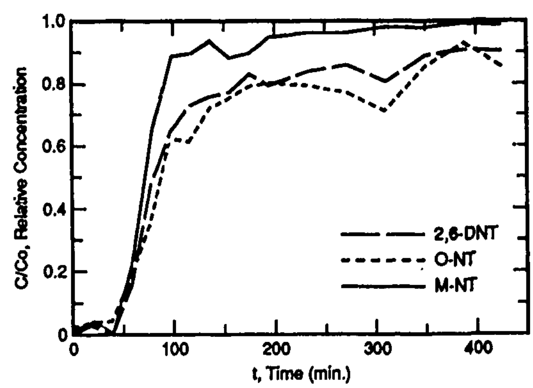

Figure 8. Combined breakthrough curves for 2,6-DNT, O-NT and M-NT C/Co vs.t).

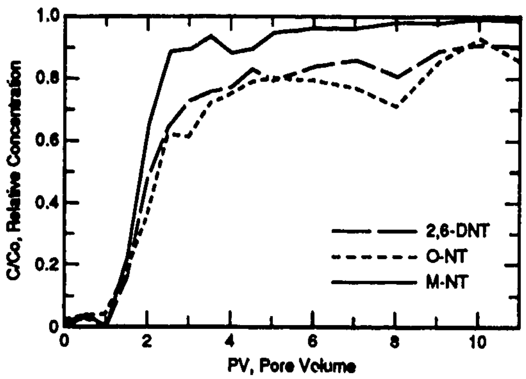

Figure 9. Combined breakthrough curves for 2,6-DNT, O-NT and M-NT (C/co vs.PV).

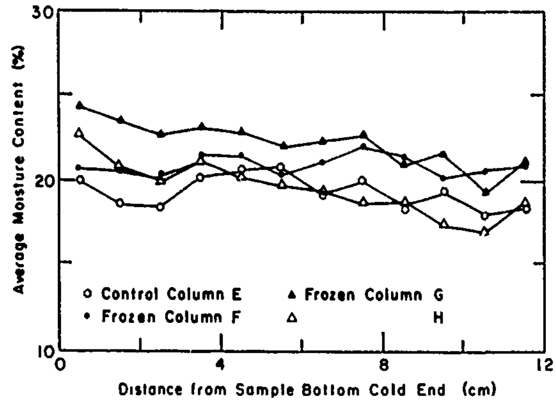

Figure 10. Moisture content profiles of soil columns with explosives.

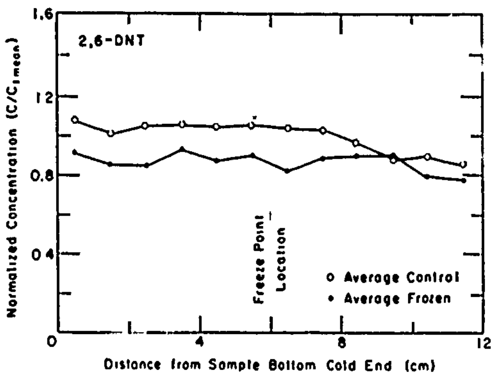

Figure 11. Normalized 2,6-DNT concentration profiles in Lebanon silt.

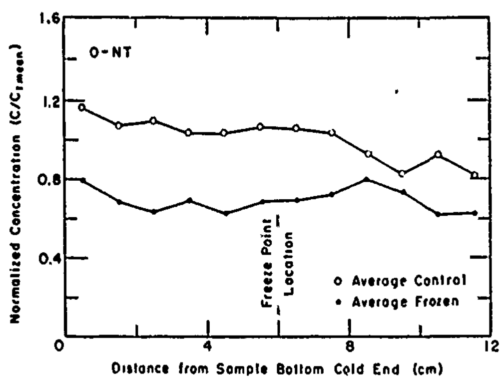

Figure 12. Normalized O-NT concentration profiles in Lebanon silt.

Concentration profiles for 2,6-DNT are depicted in Figure 11. Very little data scatter along the sample height for control and frozen sample profiles was observed for 2,6-DNT. The statistical analysis of the 2,6-DNT yielded low values for standard deviation (STD) and coefficient of variation (CV). Despite the low STD and CV values, only about 20% concentration reduction was calculated for 2,6-DNT, indicating no appreciable movement caused by freezing. In addition, a comparison of the regression slopes between control and frozen samples for 2,6-DNT showed that there was no statistically significant difference between the slopes at the 95% confidence level (Table 2). The average initial concentration for 2,6-DNT was approximately 5.6 µg/g. At such a low concentration, the effect of sorption and biodegradation may be significant enough to adversely impact the freezing-induced mobility of 2,6-DNT.

Figures 12 and 13 compare the concentration profiles between the control and frozen samples for O-NT and M-NT, respectively. Comparing the control and frozen samples, the normalized concentration reductions within the frozen portion were about 40% for O-NT and about 37% for M-NT (Table 2). The lengthwise data scatter for both analytes was minimum, and the STD and CV values appeared to be uniform along the sample height. The average initial concentrations for O-NT and M-NT were 3.0 µg/g and 3.75 µg/g, respectively. Using the Student t-test significance analysis, the regression slopes of the concentration profiles for control and frozen samples were compared. It was found that there was a statistically significant difference

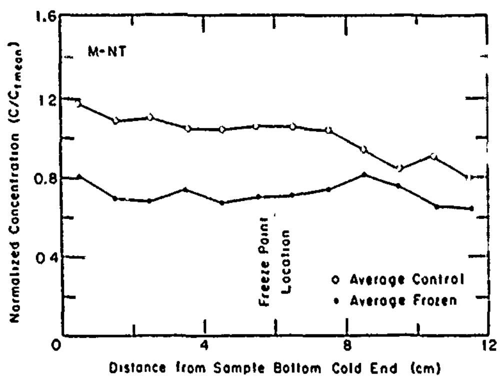

Figure 13. Normalized M-NT concentration profiles in Lebanon silt.

between the slopes at 95% confidence level for both O-NT and M-NT as shown in Table 2. Hence, the resulting data strongly suggest that M-NT and O-NT were moved in Lebanon silt when subjected to one freezing cycle at a freezing rate of 0.5 cm/day.

CONCLUSIONS

From the experimental study, the following conclusions were drawn:

1. For one freeze cycle, about 40% reduction in concentration was obtained for M-NT and O-NT, respectively, with less than 20% reduction for 2,6-DNT.

2. At an average freeze rate of 0.4 cm/day, statistically significant movement was observed for M-NT and O-NT but not for 2,6-DNT.

3. For given freeze rate, freeze-thaw cycles, soil and soil moisture, it was postulated that the ability to move a contaminant by freezing strongly depends on the type, initial concentration level and the soil/chemical interaction of the contaminant. Also the ability to easily detect each contaminant in the soil affects how to assess whether or not freezing moves a given type of contaminant.

ACKNOWLEDGMENTS

This work was financially supported, in part, by the U.S. Environmental Protection Agency (EPA) under the Interagency Agreement Project No. DW96931421-01-2 with the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL), and, in part, by the U.S. Army Corps of Engineers RDTE Project No. 4A161102AT24, Work Unit SS/020, Prediction of Chemical Species Transport in Snow and Frozen Ground. The authors wish to thank T.F. Jenkins, P. Schumacher, D. Pidgeon and P. Miyares for their cooperation and assistance in the chemical and computer data analyses. We also thank E. Wright for editorial review, and Dr. C.M. Reynolds and E. Chamberlain for technical review of the manuscript. The EPA Project Officers for this project were Janet Houthoofd and Doug Keller, Risk Reduction Engineering Laboratory, Cincinnati, Ohio.

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DISCLAIMER

The information contained in this paper represents the authors' opinions and not necessarily those of EPA or CRREL. Hence, no official endorsement should be inferred.

EFFECTS OF FREEZING ON SULFATE SALTS IN NORTH DAKOTA SOILS AND WETLANDS

by

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INTRODUCTION

Cooling and freezing are techniques used in the sulfate salt industry to separate Glauber salts (mirabilite $Na_2SO_4 \cdot 10H_2O$) from Epsom salts (epsomite $MgSO_4 \cdot 7H_2O$) and other brine constituents (Last and Schweyen, 1983). Figure 1 presents the temperature dependent solubility relationships and stability fields of the major Mg and Na sulfate salts. By controlling density, temperature and brine composition a nearly pure sodium sulfate can be harvested. In saline seeps the stability of sulfate evaporite minerals and the relative abundance of their associated ionic constituents in soil extracts is also strongly affected by temperature changes (Keller et al., 1986a, 1986b; Timpson et al., 1986; Timpson and Richardson, 1987). A similar but annual separation of Mg and Na salts should occur in soils and pond water during fall and winter freeze-up if the solution has sufficiently high salt content and contains abundant sulfate.

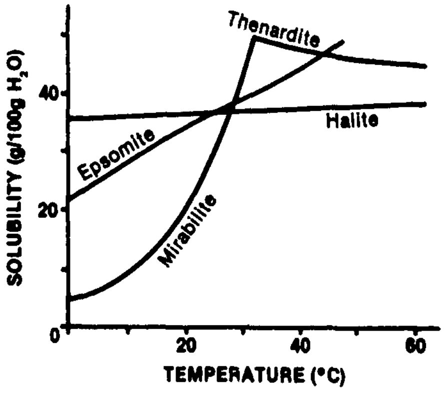

Figure 1. Relationships between solubility and temperature for some common efflorescent salts (after Timpson et al., 1986).

Freezing of saline lakes can also have other unexpected consequences. Sublimation, the transition of a substance directly from the solid state to the vapor state, may explain the observed release of crystalline salts from the ice of a saline pond emplaced in ice-thrust hummocky moraine in east-central North Dakota. During early winter sampling of this pond we noted drifts of salt on the ice surface in emergent vegetation. The dune-like appearance and position of the salt on top of the ice plus the absence of other salt sources suggests that original ice-entrapment of salt occurred. The entrapment was followed by later release through sublimation of the pond ice (Arndt and Richardson, 1985).

This paper investigates the influence of freezing of sulfatic waters on salinization and segregation of Mg and Na in a natural setting associated with brackish to saline ponds in east-central Nelson County, North Dakota.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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PROCEDURES

The identity of evaporite salts collected in winter was determined by x-ray diffraction using the techniques of Keller et al. (1986a). Monthly to bimonthly samples of near-surface groundwaters were collected from hydric soils peripheral to the saline and brackish wetlands using several passive, capsule-type groundwater collectors (Arndt and Richardson, 1988, 1989). The study wetlands are emplaced in calcareous, sulfate-rich till typical of the Northern Plains, the weathering of which results in strongly sulfatic groundwater (Hendry et al., 1986). Electrical conductivity (EC), pH, and soluble ions (CO_3, HCO_3, Cl, SO_4, Ca, Mg, Na, and K) were determined in salt (dissolved in a known quantity of deionized water), ice, and water samples using standard methods (Soil Survey Staff, 1984).

To investigate the magnitude of sublimation as a potential source of crystalline salts over winter, direct measurements of weight loss due to ice sublimation from white plastic pans filled with water and frozen were made during January of 1985 and 1987. Days where snowfall covered the pans and days where the temperature was above freezing were excluded from the results.

RESULTS AND DISCUSSION

Sublimation ranged from 0.06 to 0.69 mm H_2O per day during 12 days in January, 1985. During January, 1987 sublimation ranged from 0 to 0.73 mm H_2O per day during 18 days. The average for all measured days was 0.31 mm per day, a value probably inflated because days with precipitation (snow) or days with periods above freezing were excluded. However, when related to a 20 acre (8 ha) pond similar to the saline pond (20) (Table 1), this represents a loss of over 25,000 $L{\cdot}day^{-1}$ of water. If the concentration of salt in the ice is 10 $g{\cdot}L^{-1}$ (EC approx. 10 $ds{\cdot}m^{-1}$, Table 1), this represents a release of over 250 $kg{\cdot}day^{-1}$ of salt from this wetland. Even if sublimation values are halved, the potential exists for large quantities of slat to be released.

Table 1. Conductivity and relative ionic composition* of unfrozen water, ice, and aeolian salt from wetlands 20 and 18.

Wetland #	Sample	EC dS/m	SO_4	Alk**	Ca	Mg	Na
			----% Anions----		----% Cations----		
20	Fall	18.3	96.3	1.2	4.6	42.5	50.9
	Ice [1]	11.1	95.8	2.6	6.8	40.5	51.4
	Ice [2]	13.7	95.1	2.9	3.5	47.0	46.3
	Ice [3]	8.0	93.5	4.7	4.4	46.3	47.7
	Winter [1]	44.0	94.7	3.5	2.1	54.5	42.3
	Winter [2]	61.9	95.4	2.5	1.5	59.7	37.3
	Winter [3]	40.6	93.2	5.0	2.7	55.9	42.3
	Salt	-	96.5	2.6	8.6	26.1	64.5
18	Fall	2.6	62.4	30.5	19.9	36.1	40.8
	Winter	9.9	84.3	10.3	16.5	51.5	30.2

* mmol (+/-)L basis.
** Alkalinity (Alk) includes HCO_3 and CO_3.

The effect of freezing as a concentration mechanism can be seen in the conductivity and the relative ionic composition of selected water and salt samples from both the brackish wetland (18) and the saline wetland (20) (Table 1). The data from wetland (20) show that the EC of surface ice was two-thirds that of fall water, yet winter water collected under the ice exhibited conductivities 2 to 3 times greater than fall. Similar trends are noted in wetland (18). These data suggest that dissolved solids are exiting from freezing solutions as ice forms and are being concentrated in water underneath the ice. Thompson and Nelson (1956) describe similar conditions.

Differences in the relative amounts of major ions among salt, water, and ice samples indicate that some ions are being concentrated more than others, suggesting a geochemical control on the levels of certain ions with progressive concentration through freezing. In the less saline conditions of wetland (18), SO_4 and Mg are increasing in pondwater at the expense of alkalinity, Ca, and Na. A similar trend is noted in the saline samples of wetland (20), except that differences between SO_4 and alkalinity are minor at these high concentrations. A model developed by Hardie and Eugster (1970) to explain the evolution of closed basin brines by evaporation predicts that with progressive concentration the relative amounts of alkalinity and Ca with decrease; however, in these solutions Na seems to be controlled also. An x-ray analysis of the salt collected on top of the ice indicated a dominance of mirabilite, epsomite, and gypsum ($CaSO_4 \cdot 2H_2O$) with smaller amounts of calcite ($CaCO_3$). Thenardite (Na_2SO_4) was essentially absent. Mirabilite and epsomite are common, extremely soluble Na and Mg sulfate salts and are not normally expected to exist in equilibrium with any but the most saline solutions.

Freezing, however, acts as a concentration mechanism active in saturated soils that removes water from solution and can cause even very soluble minerals to precipitate. At 30 °C, epsomite and mirabilite have similar solubilities (Fig. 1). The solubility of each decreases with temperature. However, at 0 °C mirabilite is one-fourth as soluble as epsomite. In sulfatic solutions where Na and Mg concentrations are high, mirabilite should precipitate before epsomite upon freezing. The salt collected from the surface ice is enriched in sodium, a further indication that mirabilite precipitation occurs prior to epsomite. Under conditions of extreme concentration of sulfatic solutions, mirabilite saturation may act to control the activity of Na in the same way that gypsum saturation controls Ca activity. This would allow the more soluble Mg-sulfates to accumulate in unfrozen water.

Freezing from the surface downward may concentrate dissolved solids at depth in a saturated soil and alter soil-solution composition in much the same way, affecting mineral precipitation and wetland soil morphology. Data from groundwater extractors in the shallow marsh of wetland (20) show an enrichment in magnesium over sodium with depth similar to that seen in the ice and pondwater samples. However, more data from these extractors is needed to show clearly that depth-dependent changes in soil-solution composition occur because of freezing.

CONCLUSIONS

Sulfate minerals frequently contain waters of hydration in their crystal structure. The hydrogen bonds associated with these waters of hydration are exceedingly temperature dependent; therefore, differential solubility of sulfate salts occurs in saline, high sulfate soils. Mirabilite, dominant in salt released from the ice of the saline wetland, in particular changes solubility with temperature. Freezing concentrates the soil solution to levels that may result in mirabilite precipitation. The mineral precipitation sequence that occurs at or near freezing apparently can segregate ions in soil profiles. Because of their greater solubility magnesium minerals do not form until later in the freezing process. Thus Mg may be concentrated over Na in the soil solution or in pond water during freezing.

We hypothesize that in saline soils high in sulfate, top down freezing tends to maximize the sodic influence by removing magnesium; the result is that a matric or sodium affected soil can occur with lower amounts of sodium ions present than in an unfrozen soil or in a soil high in chloride salt where temperature dependent solubilities are not as important.

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AN SCS PERSPECTIVE ON USING RESEARCH MODELS IN PLANNING AND APPLYING CONSERVATION MEASURES

Lee P. Herndon 1/

INTRODUCTION

The 1990s promise to be an exciting time for soil and water conservation professionals. The development of powerful microcomputing capability will make possible utilization of the latest technology to solve problems occurring in agricultural ecosystems in ways heretofore considered impractical. Efficient access to remote data base information, combined with onsite data gathered by conservationists, will fuel the engines of predictive models carried to the field with these microcomputers. Such "real-time" analysis of soil and water conservation problems will provide SCS and others an invaluable tool in providing better service to decisionmakers in protecting and enhancing the Nation's natural resources.

SOIL AND WATER CONSERVATION PRIORITIES

Soil and water conservation priorities have been identified for the next decade that will guide the development of specific technology aimed at solving complex agricultural ecosystem problems. USDA, through its National Program for Soil and Water Conservation, will focus its overall efforts in two general areas:

1. Reducing damage caused by excessive soil erosion on rural lands; and

2. Protecting the quality of surface and ground water against harmful contamination from nonpoint sources and thereby maintaining the quantity of water available for beneficial uses.

In addressing USDA's top priority item, damage from soil erosion on rural lands, offsite (as well as onsite--the primary concern in the past) will be considered. Consequently, activities undertaken by action agencies within USDA will be primarily tuned to developing, transferring, and applying technology that will alleviate such damage.

ROLE OF SCS IN ADDRESSING PRIORITIES

The primary mission of SCS has always been to provide national leadership in the conservation and wise use of soil, water, and related resources through a balanced, cooperative program that protects, restores, and improves those resources. The role of SCS professionals is to cause actions to occur that are compatible with the dominant mission. In accordance with the priorities established by the National Conservation Program, SCS is taking a leadership role in ensuring that conservation provisions of the 1985 Food Security Act (FSA) are carried out.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990
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This Act, through the Conservation Reserve and the Highly Erodible Lands provisions, establishes a commitment to protect the Nation's cropland from excessive damage due to erosion. SCS is now devoting a major part of its funds and resources in efforts to help reduce erosion on highly erodible cropland. For the purposes of the Act, highly erodible cropland (HEL) is land that has an erodibility index of 8 or more based on factors from either the Universal Soil Loss Equation (USLE) or the Wind Erosion Equation (WEQ) (i.e., RKLS/T or CI/T equals or exceeds 8) (USDA, 1978; Woodruff and Siddaway, 1965).

SCS ACTIVITIES PLANNED FOR IMMEDIATE FUTURE

Among other activities associated with these priorities, SCS provides technical assistance in planning and applying conservation systems for this highly erodible cropland. In carrying out their responsibilities, SCS professionals must have up-to-date knowledge of the complex processes of wind and water erosion. Over the next several years, conservation plans incorporating this knowledge must be applied on about 140 million acres of highly erodible cropland.

This symposium concerns frozen soil and its impact on agricultural lands. About 70 percent of the cropland identified as highly erodible land (HEL) is potentially subject to freezing and thawing--roughly that area north of the 38 degree North Latitude line but the high mountain ranges of New Mexico, Arizona, and California. Therefore, knowledge of the complex phenomena associated with this aspect of erosion is also necessary for proper planning and application of all conservation measures in these areas of the country and is directly relevant to complete and comprehensive treatment of lands identified as highly erodible.

USE OF MODELS IN PLANNING AND APPLYING CONSERVATION WORK

The term "model" is often misunderstood, connoting different meanings to soil and water conservation professionals in different agencies. Furthermore, the utilization of modeling to help fashion additional conservation tools to solve soil and water conservation problems is likewise often fraught with difficulties related to such things as misapplication of the model for the purposes intended and/or disagreement over who (in most instances, which agency) should be responsible for ensuring a specific model's proper utilization in developing conservation tools.

Models in the soil and water profession are often constructed to represent, in a relatively simple manner, more complex processes associated with physical phenomena or ideas in hopes that such phenomena or ideas can be better explained, predicted, or manipulated in some way (Rubenstein, 1975; Kaye, 1982). Modeling procedures and techniques are well understood by professional researchers, but not so well understood by conservationists outside of the research profession or others engaged primarily in using technological tools based on models to plan and install conservation measures. Additionally, over the years, confusion has developed regarding the appropriate use of these technological tools, based on models, by personnel in action agencies. We have all heard complaints such as ". . .too data intensive. . ." or ". . .gives answers that do not seem to correspond with field conditions. . .".

It is important that SCS utilize various soil- and water-related models in conservation planning and application. However, to be useful, easily understood procedures based on modeling efforts must be developed for obtaining appropriate answers to problems occurring in agricultural ecosystems. With today's computing equipment, this often means that user-friendly computer programs must be built to support the technology imbedded in models.

Consequently, it is vital that research agencies, academic personnel, and user groups (such as SCS) work closely together to forge proper conservation tools from the raw material of basic research models. Over the last several years, a remarkable and extremely effective partnership between the agricultural scientific and academic communities and user agencies has been in existence. This effort has involved translating the basic technology being developed by the Agricultural Research Service (ARS) for the Water Erosion Prediction Project (WEPP) into useful, user-friendly computer software to be employed by SCS conservationists in planning and applying soil conservation practices. Designed to eventually replace the Universal Soil Loss Equation (USLE), WEPP will be the cornerstone of SCS conservation planning efforts for well into the 21st century. The development of an operational computer program (OCP) for portions of the WEPP model will help ensure timely and successful implementation of this technology at the field level of SCS. A similar effort will be needed to translate technology now being developed by ARS under the Wind Erosion Prediction System (WEPS).

USE OF WEPP AS A "TEMPLATE" FOR SOIL AND WATER MODELS

An important part of WEPP technology involves ascertaining comprehensive knowledge of freeze-thaw relationships. We expect the work of ARS researchers, SCS engineers, and others equally involved in studying these relationships to be effectively utilized in WEPP. But beyond the considerations involved in specific components of WEPP, such as freeze-thaw relationships, it is even more important to use WEPP as a "template" for other soil and water conservation efforts, such as will be required in the water quality arena, to determine the requirements of the user community prior to beginning models efforts.

Publication of the booklet "User Requirements: USDA - Water Erosion Prediction Project (WEPP)" was a precedent-setting event in ascertaining the requirements of the user community for research into new technology (USDA, 1987). Such a study should be made for each and every major soil and water research effort undertaken in the future. Which involved agency initiates such an effort is not as important as ensuring that the work is accomplished.

SCS practioners, ARS researchers, academic personnel, and others should begin now to proactively seek out opportunities to take advantage of the new WEPP technology. With the idea in mind that it is best to conserve soil rather than prevent erosion, a subtle mind set that may one day portend a paradigm shift in attacking soil and water conservation problems, we can begin the process of "inventing" conservation systems that are more holistic in nature. This process will require major new research efforts and the continuing close cooperation of all of us involved in soil and water conservation.

FUTURE RESEARCH NEEDS

The recently published 1989 "Soil and Water Conservation Research and Education Progress and Needs" report identified high priority research needed to address some of the most important soil and water conservation priorities of the 1990s (USDA, 1989). Completion or initiation of research models associated with these high priority needs will require close attention by ARS researchers, academic personnel, and SCS professionals. It is intended that each of the items listed in the report be assigned to someone in SCS so that progress in completing the items can be monitored and so that ancillary problems (and opportunities) associated with the items can be identified at the earliest possible stage. We in SCS foresee an era of extraordinary cooperation between ourselves and the research community in completing high priority and high quality research needed to further the objectives of soil and water conservation in the United States.

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FROZEN SOIL IMPACTS: RESEARCH NEEDS

by

R.I. Papendick and K.E. Saxton

INTRODUCTION

The purpose of frozen soil research on agricultural and forest lands should be to develop new knowledge and technologies that can help provide management strategies for overcoming the adverse effects of soil freezing and thawing on production and environment while at the same time enhancing the beneficial effects. As this symposium has made clear, frozen soil is a natural occurrence on a large percentage of the world's land areas which are under cultivation, or in grasslands and forests. It is these lands that produce food, fiber, and wood products for over a third of the world's population and it is likely that demands on this particular resource base will continue to increase. Some of the world's most productive soils and agroclimatic zones are found where frozen soils occur. It is also where some of the most fragile soils and ecosystems exist.

Because soil freeze/thaw action can dramatically affect many of the objectives of farm, range, and forest management practices, it imperative that frozen soil impacts be given a high level of research attention. Soil freezing is especially an important factor influencing broad scale hydrologic occurrences, and thus, soil water conservation, stream flow, soil water recharge, surface and groundwater storage, soil erosion, and nutrient loss. It can also be an important factor influencing plant survival during the winter. For example, prolonged frost combined with cold winds can desiccate and kill winter wheat and other fall-planted crops. In the Pacific Northwest, soil heaving from freeze action can cause wounds in winter wheat roots which may lead to increased infection by the soil-borne pathogen *Cephalosporium* gramineum and result in serious crop losses.

Clearly a wealth of knowledge exists on soil freeze/thaw processes, predictability of freeze/thaw occurrences, frost measurement techniques, and effects of frost on soil properties. There are also some excellent research simulation models available, such as those of Benoit and Mostaghimi (1985), Cary et al (1978), and Flerchinger and Saxton (1989), which can predict the influence of management practices on the depth of soil freezing, and water and solute movement. However, it is also evident that there are major knowledge gaps which limit application of technologies for management of frozen soil effects on agricultural, range and forest operations. There is also an urgent need for improved transfer of technology that can assist land use planners and land users in coping with frozen soil impacts.

RESEARCH RECOMMENDATIONS FOR THE FUTURE

Priorities for frozen soil research should be a part of or aligned with problem-solving research on existing or emerging major issues in the management or use of agricultural, range and forest lands. For example, major problems confronting farmers in many regions include excessive runoff and soil erosion, low water and nutrient use efficiency, and pollution of surface and groundwater by pesticides and agricultural chemicals. Floods from winter and early spring snow melt can cause severe damage to farmlands, homes, and urban developments, and also create problems in the management of rangelands and forests. Frozen soils can impact all of these problem areas and must be considered in any research plan for potential solutions.

Presented at the Frozen Soil Symposium, March 21-22, 1990; Spokane, Washington.

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Some suggested areas of high priority research are listed as follows:

1. Determine the effect of soil freezing on infiltration rate and runoff of rain and snow melt.

While it is possible to reasonably well predict depth and duration of soil freezing with simulation models it is not yet possible to adequately predict the infiltration and runoff of rain and snow melt on frozen and partially thawed soils. This predictive capability needs to be developed and incorporated with freeze/thaw simulation models to account for the influences of surface conditions, tillage and residue management, slope characteristics, frost type, and antecedent soil moisture on infiltration and runoff from frozen soils. The enhanced model should be applicable to forest and rangelands as well as croplands.

2. Incorporate soil freeze/thaw effects in hydrology and erosion prediction modeling.

Continued effort needs to be given to determining and mathematically describing the effects of freeze/thaw process on runoff and soil erosion by water. Attention should be given to establishing relationships between soil shear strength, and critical hydraulic shear as affected by freeze/thaw cycles for different soil types and runoff conditions. These inputs need to be incorporated into models such as WEPP to improve their performance on frozen soils by accounting for effects of freeze/thaw process on runoff, and rill and interrill soil erodibility values.

3. Determine the effect of soil freezing and thawing on overwinter water and solute movement and soil biological processes in relation to surface and groundwater contamination by agricultural chemicals.

There is considerable evidence that freeze/thaw action can cause significant effects on soil hydrologic processes which in turn influence chemical transport. However, the flux rates and magnitudes of chemical leaching from rain or snow melt during freeze/thaw periods in relation to management practices are poorly understood. Soil freeze/thaw prediction models such as that of Benoit or SHAW should be expanded to include chemical and microbial transformations and movement of nutrients and pesticides for representative wintertime conditions including where there is ephemeral freezing and thawing. The model should serve as the basis for analyzing farm management systems as to potential impacts of the freeze/thaw process on surface and groundwater quality and for chemical management alternatives to improve water quality.

4. Adapt technologies to improve infiltration rate of frozen and partially thawed soil.

Soil management technologies such as slot mulch, the dammer diker, and the paraplow provide potential methods for reducing runoff of rain and snow melt on frozen soils. However, research is needed to determine how these technologies can be adapted into the farm management system for effective results. Consideration must be given to where these specialized tillages fit best in the cropping sequence, frequency of application, range of slopes where they can be used, residue amounts, soil types, and farmer interest.

5. Determine the effect of soil freezing in relieving compaction in shallow and deep compacted layers.

Use of increasingly heavier machinery and on wetter soils has led to formation of compacted layers in many soils extending to depths of 30 cm or more. In some cases the compaction may occur below the level of normal cultivation which makes it expensive and extremely difficult to break up even with specialized equipment. The compaction which may develop can impede water flow, aeration, and root growth compared with soil where compaction has not occurred. Research is needed to determine the extent to which freezing and thawing breaks up compacted soil and at what depths the compaction is relieved, and how management affects this process.

6. Determine the effect of soil freezing on overwinter soil water recharge.

Water movement to the freezing front from shallow water tables and deeper layers can be a significant source of soil water recharge for some soil conditions. Research is needed to determine the extent to which this occurs and how this type of recharge can be enhanced through soil, snow, and residue management. Where this is important this contribution needs to be accounted for in water budget models.

7. Evaluate methods to minimize frost heave damage to crops.

Damage to crops from soil heaving caused by freezing can be extensive for certain soils. The effect of management practices on heave damage is not well understood. Studies are needed to determine how tillage, residue, and snow management can be used to reduce frost heaving where this is a serious problem. Special consideration should be given to soil type, water contents, heat and water flow properties of the soil, and rate of freezing.

8. Determine how residue layers affect the energy balance at the soil surface, and rate and depth of soil freezing.

Surface residues strongly influence the flow of energy across the soil surface. The physics of the radiative and thermal energy balance through the residue layer is not well understood with respect to the quantity, quality, and orentation of the residues. This type of information is needed for improving the predictive capability of soil freeze/thaw models to account for residue management effects on depth, duration, and type of frost.

9. Test the applicability of soil freezing and thawing simulation models for forest conditions.

Results with recent modeling efforts indicate good versatility for predicting soil freezing and thawing for widely varying cropland and rangeland conditions. These models need to be further tested for applicability to forested and cleared forest lands with and without heavy snow cover, and where winter and early spring runoff is important.

10. Develop geographical and topological influences on soil freezing and topographic impacts.

Topography, aspect, and associated snow depths often dramatically influence the depth-duration of soil freezing. Sometimes this benefits the prodiction or resource protection and sometimes it causes drastic effects. These cases need to be studied and understood to provide management alternatives.

11. Develop probability maps of frozen soil occurrences for use in hydrology and erosion models.

Seasonal and annual predictions of hydrologic processes and soil erosion will require as inputs probability values for frozen soil occurrences. Maps for land areas need to be developed from weather records and use of freeze/thaw simulation models that can be used to identify frequency and distribution of frozen soils during the winter season. This information can be used to improve the long term prediction performance of existing erosion and hydrologic models which include subroutines which account for frozen soil effects.

12. Evaluate the potential effect of global warming on the distribution and seasonal patterns of frozen soils.

Models for predicting global warming or the "greenhouse effect" need to include predictions on how climatic change will influence the distribution and seasonal patterns of frozen soils in different regions of the world. Such predictions are needed to evaluate potential effects of increased temperatures on changes in hydrologic phenomena, vegetation, adaptable plant species, and biological production in regions where major changes in frozen soils can be expected to occur.

SUMMARY

Research on frozen soils needs to be a part of ongoing problem-solving research on existing or emerging major issues in the management and use of agricultural and forest lands. The process of soil freezing and thawing impacts a number of important problem areas including watershed runoff, soil water conservation, soil erosion, surface and groundwater contamination from agricultural chemicals, and frost heave damage to plants. Research should give high priority to how management practices can be applied to alleviate the adverse effects of frozen soils on the environmental and productivity goals of crop, range, and forest land use. It should also consider how the beneficial effects of soil freezing can be enhanced.

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RUNOFF AND EROSION DURING SIMULATED RAINFALL ON FROZEN FIELD PLOTS WITH DIFFERENT DEPTHS OF SURFACE THAW AND LEVEL OF ERODIBILITY

by

GEORGE R. BENOIT, ROBERT A. YOUNG, and ALAN WILTS[1]

INTRODUCTION

Freezing and thawing have a marked influence on a soil's aggregate stability, density, hydraulic conductivity, and water content (Baver, 1956; Bisel and Nielson, 1964 and 1967; Gardner, 1945; Domby and Kohnke, 1955; Leo, 1963; Sillanpaa and Webber, 1961; Slater and Hopp, 1949; and Willis, 1955). The influence of frost action on such soil physical properties can be either positive or negative with the magnitude of the effect depending on soil type and initial conditions (Benoit, 1973; Chepil, 1954; Mostaghimi et al., 1988; and Sillanpaa and Webber, 1961). This implies that frost action should have a marked but site-specific effect on soil erodibility when snowmelt or rainfall runoff events occur on frozen or partially thawed soil. Published reports confirm this by showing that from about 10 to nearly 90 percent of annual soil loss can occur during runoff events on frozen or partially thawed soil with the actual amount of erosion that occurs depending on soil type, presence of frost, and the physical conditions of the soil at time of the event (Kirby and Mehuys, 1987; Satterlund and Eschner, 1965; and Spomer and Hjelmfelt, 1983).

The knowledge base available about erosion from frozen soil is broad but general with little information that shows the relation between thaw depth and actual erosion during a runoff event. Our objective is to show, with use of a rainfall simulator, the effect of thaw depth on infiltration, runoff, and sediment transport from three Minnesota soils under two tillage conditions (fall plow and fall chisel) and two levels of residue management (residue and residue removed).

PROCEDURE

Field sites that had been planted to corn were selected on a Hattie clay (Udertic haploboroll) with a 6 percent slope in the fall of 1985, a Barnes loam (Udic haploboroll) with a 7 percent slope in the fall of 1986 and 1987, and a Sverdrup loamy sand (Udic haploboroll) on a 3 percent slope in 1988. Each fall 8 uniform plot areas were selected and split so that all corn residue was removed from half of each plot area. After the residue treatment had been established, half of the plots were moldboard plowed and half chisel plowed. The result was a split plot arrangement with residue level superimposed on main tillage treatment. This approach yielded 16 individual plots, 4 each of all tillage residue combinations. No treatment replication was possible for each run.

After tillage-residue treatments were established, each plot was isolated with metal barriers and a drainage pit was dug at the downslope end. Frost tubes, neutron access tubes, and thermocouples were installed in the treated area but just upslope of the metal barriers. Frost depth, thaw depth, snow depth, soil moisture, and soil temperature were measured weekly throughout the winter and on a more frequent basis during spring thaw. In addition, initial random roughness, aggregate stability, and surface bulk density data were recorded for each treatment. Sill plates were installed at the downslope end of each plot and the 16 plots were divided into 4 groups. Each group had all residue-tillage combinations.

In the spring simulated rainfall was applied to one group of plots at a time at 4 successive stages of thawing. Thaw depths during simulated rain ranged from about 20 mm to 650 mm. A final rain event was run after complete thaw each year. Extra runs in some years covered situations where surface refreezing had occurred, creating a three-layer system of surface frost, thawed soil, and deep frost. All soil measurements were repeated

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Presented at Frozen Soil Symposium, March 21-22, 1990, Spokane, Washington.

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for each simulated rainfall event in addition to measurements of time to first runoff, total runoff, sediment yield, aggregate size distribution, aggregate stability, enrichment ratios, and soil surface strength. In this paper we will examine the overwinter tillage-residue effects of soil physical conditions and concentrate on the relation between depth of thaw and rates of soil or water loss.

RESULTS

Surface density (BD), random roughness (RR), and aggregate stability (MWD) recorded just before each simulated rainfall event was related primarily to soil type (Table 1). The Hattie clay used in spring 1986 had an average surface BD (Uhland core method) of 1.02 while Barnes loam (1987 and 1988) and Sverdrup sandy loam (1989) had average surface BD's of 1.16 and 1.37, respectively. In general, the finer textured soil had the lowest pre-rainfall BD values (Table 1).

Tillage type and residue treatment had secondary and inconsistent effects. Fall plowing as compared to fall chisel decreased surface BD of Hattie clay, slightly increased BD of Barnes loam but had no effect on Sverdrup sandy loam. No apparent significant difference in BD was associated with residue management on any of the soils studied. In addition, no significant relationship was found between surface BD and thaw depth.

Table 1. Soil surface density (BD-Mg/m^3), random roughness (RR-mm), and aggregate mean weight diameter (MWD-mm) by drop tower method recorded just before simulation rainfall events were applied to Hattie clay, Barnes loam, and Sverdrup sandy loam soils. Tillage-residue treatments were fall plow and fall chisel, each with and without residue. Seasonal average maximum frost depth (FD) is also shown for each tillage-residue treatment.

	FALL PLOW				FALL CHISEL			
Residue Level	BD (Mg/m^3)	RR (mm)	MWD (mm)	FD (mm)	BD (Mg/m^3)	RR (mm)	MWD (mm)	FD (mm)
Hattie clay, 1986								
Res	.92	50.9	.945	718	1.09	22.8	.973	735
No Res	1.00	49.4	.916	772	1.09	20.4	.938	710
Barnes loam, 1988								
Res	1.19	26.1	.762	1228	1.11	21.5	.658	1190
No Res	1.18	27.8	.668	1185	1.17	21.8	.726	1220
Sverdrup sandy loam, 1989								
Res	1.36	23.8	.310	1500	1.35	28.0	.266	1500
No Res	1.37	24.2	.287	1500	1.40	33.3	.316	1500
Ave. across soil	1.17	33.7	.648	1151	1.20	24.6	.646	1142

On the fall plowed plots, RR was highest for Hattie clay, with a markedly lower value recorded for Barnes loam and the lowest value for Sverdrup sandy loam. In contrast, fall chisel plots were just the opposite, with lowest RR for Hattie clay and the highest for Sverdrup sandy loam. No residue effect on RR was observed (Table 1).

Aggregate stability, measured by MWD, changed only as a function of soil texture with the highest aggregate stability occurring with Hattie clay and the smallest with Sverdrup sandy loam. No tillage or residue management effect on aggregate stability was observed.

Maximum frost depth varied from year to year as a function of snow depth, winter temperatures, and soil moisture. The greatest depth was observed in 1989 in Sverdrup sandy loam and was associated with a low profile moisture content and little snow cover. The tillage and residue management used in this study never caused differential snow entrapment and showed no effect on frost depth.

Data was also collected from a Barnes soil in the spring of 1987. It was consistent with the data presented in Table 1 and is not shown.

Each rainfall simulation event was applied at 64 mm/hr and followed a two-stage sequence consisting of 60 minutes of rain, 60 minutes of rest, and a final 30 minutes of rain. Table 2 is a sample set of data collected from each rainfall application. In 1988 the time to runoff from a Barnes loam was always greatest for the 60-minute segment of the sequence and varied from 10 to 2 minutes. Time to runoff for the 30-minute segment varied from 1.5 to 0.5 minutes. Over all years, the time to runoff varied from a low of 2 minutes for the 60-minute segment to cases where no runoff occurred for either segment. More importantly, the data sample presented in Table 2 shows a breakdown of the water and soil loss recorded for each runoff segment and their relation to initial surface soil moisture and depth of thaw. For ease of presentation, data for both segments have been combined to give the total runoff and soil loss that occurred for each event each year. These combined data are presented graphically to show runoff and soil loss differences associated with thaw depth as affected by tillage method and residue level (Figs. 1-4).

Table 2. Total soil and water loss from a Barnes loam during 60- or 30-minute simulated rainfall applied at a rate of 64 mm/hr as related to thaw depth and calendar day (CD) of run. Time to beginning and end of runoff and percent soil water in the surface 150 mm of soil are also presented.

		FALL PLOW						FALL CHISEL						
TRT	RUN	RUNOFF BEG (MIN)	RUNOFF END (MIN)	TOTAL WATER (KG)	LOSS SOIL (KG)	SOIL MOIST %	THAW DEPTH mm	RUNOFF BEG (MIN)	RUNOFF END (MIN)	TOTAL WATER (KG)	LOSS SOIL (KG)	SOIL MOIST %	THAW DEPTH mm	CD
RES	60	6.0	60.5	214.52	8.87	23.00	-120	5.5	68.0	517.85	9.79	33.50	-100	70
	30	.7	30.5	168.06	4.84	23.00		1.0	38.0	358.52	6.23	33.50		
NRES	60	3.5	61.0	405.97	16.23	23.20	-110	5.5	62.0	508.91	30.73	27.40	-80	70
	30	.7	30.5	250.96	7.61	23.20		1.0	32.5	335.21	17.05	27.40		
RES	60	3.5	63.5	484.91	33.03	26.70	-50	1.0	65.0	655.10	18.00	33.20	-50	82
	30	.7	31.5	327.33	14.57	26.70		.7	34.5	359.38	7.20	33.20		
NRES	60	2.0	61.5	568.44	44.10	18.50	-50	1.0	61.0	736.21	57.25	25.10	-20	82
	30	.7	31.5	332.34	17.65	18.50		.7	31.5	416.12	23.45	25.10		
RES	60	5.0	60.5	356.85	20.55	22.30	-290	2.0	61.0	385.55	16.44	21.40	-330	95
	30	.5	30.5	240.19	12.77	22.30		1.0	31.0	244.12	7.96	21.40		
NRES	60	4.0	60.5	377.83	28.69	21.10	-310	1.0	61.0	575.06	32.54	22.90	-380	95
	30	.5	30.5	230.31	15.15	21.10		.5	31.0	307.01	10.10	22.90		
RES	60	7.5	60.5	247.93	11.24	18.40	-560	2.0	60.5	85.32	1.60	22.45	-460	98
	30	1.5	31.2	308.57	14.44	18.40		1.5	31.0	251.96	5.70	22.45		
NRES	60	8.0	60.5	396.46	14.28	20.05	-390	3.0	60.5	324.94	7.29	20.05	-410	98
	30	1.5	31.2	423.40	12.83	20.05		1.0	31.0	493.55	9.79	20.05		
RES	60	8.0	60.5	247.30	10.21	17.30	-590	7.0	60.5	128.60	2.90	19.35	-620	104
	30	.5	30.5	189.40	8.12	17.30		1.5	31.0	174.25	3.81	19.35		
NRES	60	10.0	60.5	327.32	20.82	16.85	-560	3.0	60.5	247.56	10.95	19.75	-710	104
	30	.5	31.0	322.40	19.77	16.85		1.0	31.0	228.36	9.19	19.75		
RES	60	5.0	60.5	257.69	14.10	17.35		6.0	61.0	202.28	5.85	23.30		113
	30	.5	30.5	199.40	6.98	17.35		1.0	31.0	247.79	4.61	23.30		
NRES	60	3.5	60.5	468.67	20.35	20.05		3.0	61.0	450.54	9.53	19.90		113
	30	.5	30.5	404.99	11.14	20.05		1.0	31.0	406.43	7.22	19.90		

Water loss from a Hattie clay that was chisel plowed generally increased with increasing thaw depth with the greatest runoff occurring after complete soil thawing (Fig. 1). Only a slight increase in runoff is evident where residue had been removed. Soil loss also increased with increased thaw depth for residue removed plots with the greatest loss occurring after complete thaw. In contrast, residue plots show a decreased soil loss for increasing thaw depth even with increased runoff. Comparable plowed plots, except for one plot with no residue, yielded minimal water and soil loss and in half the cases yielded no loss at all.

Simulated rainfall on Barnes loam (Figs. 2 and 3) resulted in decreased water loss with increased thaw depth for both tillage treatments with and without residue. Runoff from chisel plots tended to exceed runoff from plowed plots. Residue comparisons also suggest increased runoff from residue-removed plots. Soil loss from the Barnes loam followed the same general pattern as shown for runoff, with soil loss values higher for 1988 than for 1987.

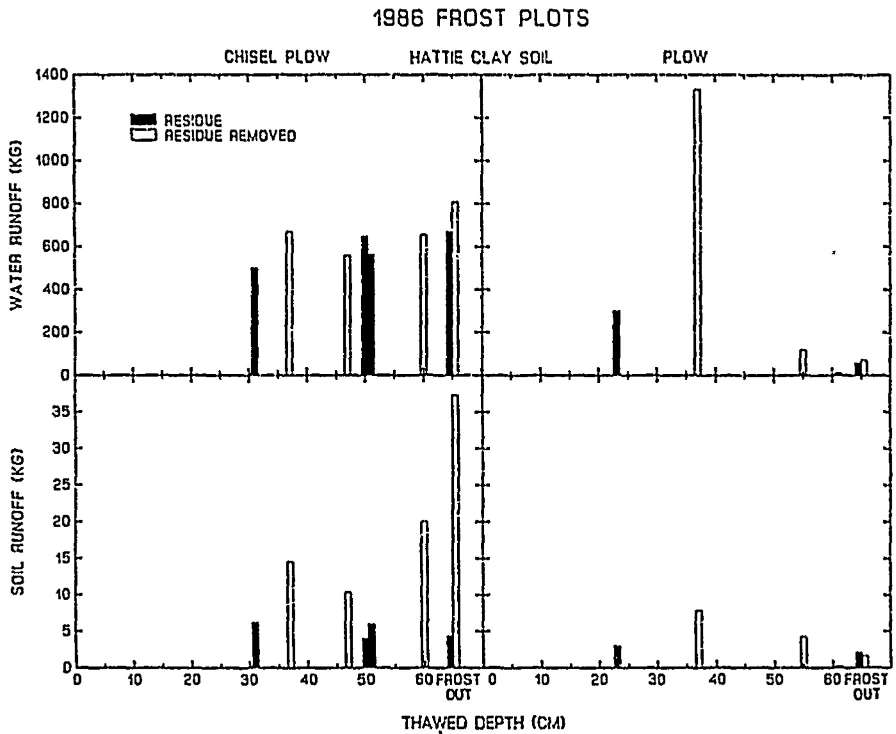

Figure 1. Total water runoff (kg) and soil loss (kg) observed at different depths of soil thawing as a result of 96 mm of simulated rain applied to a Hattie clay under chisel or plow tillage management, each with and without residue.

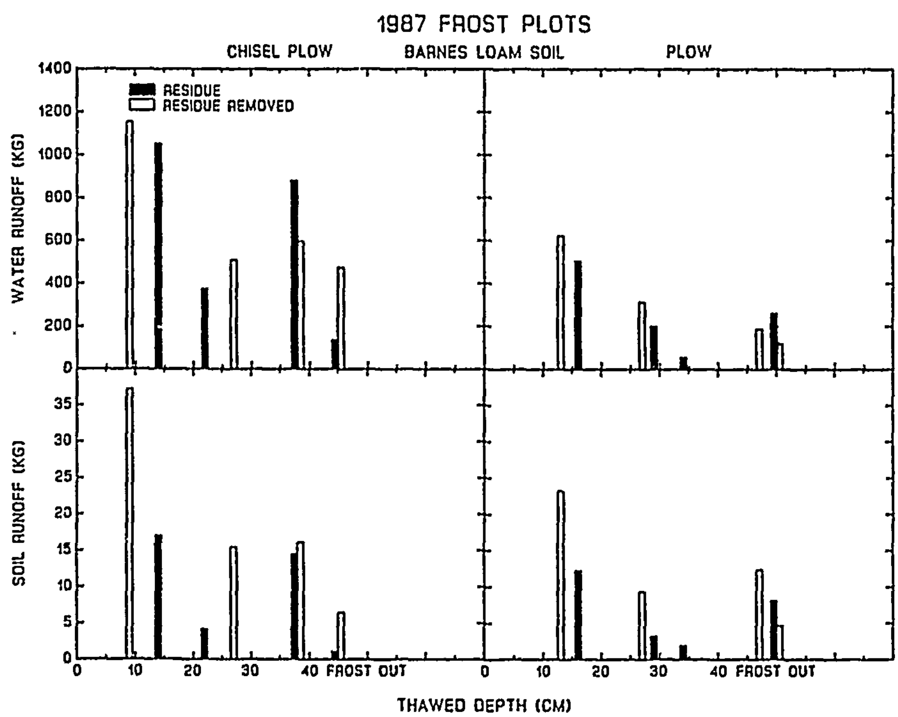

Figure 2. Total water runoff (kg) and soil loss (kg) observed at different depths of soil thawing in 1987 as a result of 96 mm of simulated rain applied to a Barnes loam under chisel or plow tillage management, each with and without residue.

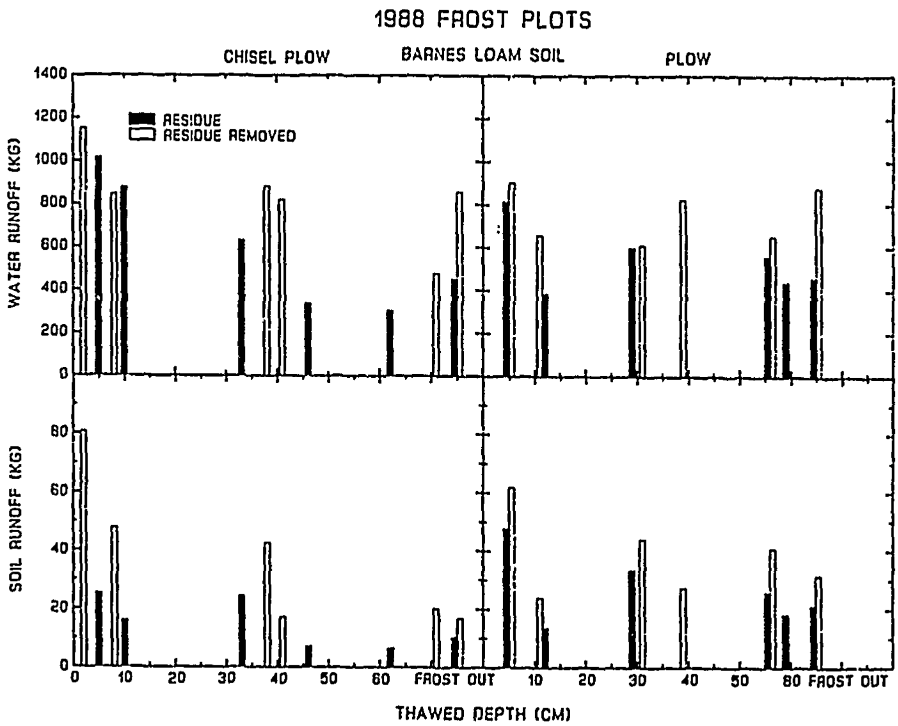

Figure 3. Total water runoff (kg) and soil loss (kg) observed at different depths of soil thawing in 1988 as a result of 96 mm of simulated rain applied to a Barnes loam under chisel or plow tillage management, each with and without residue.

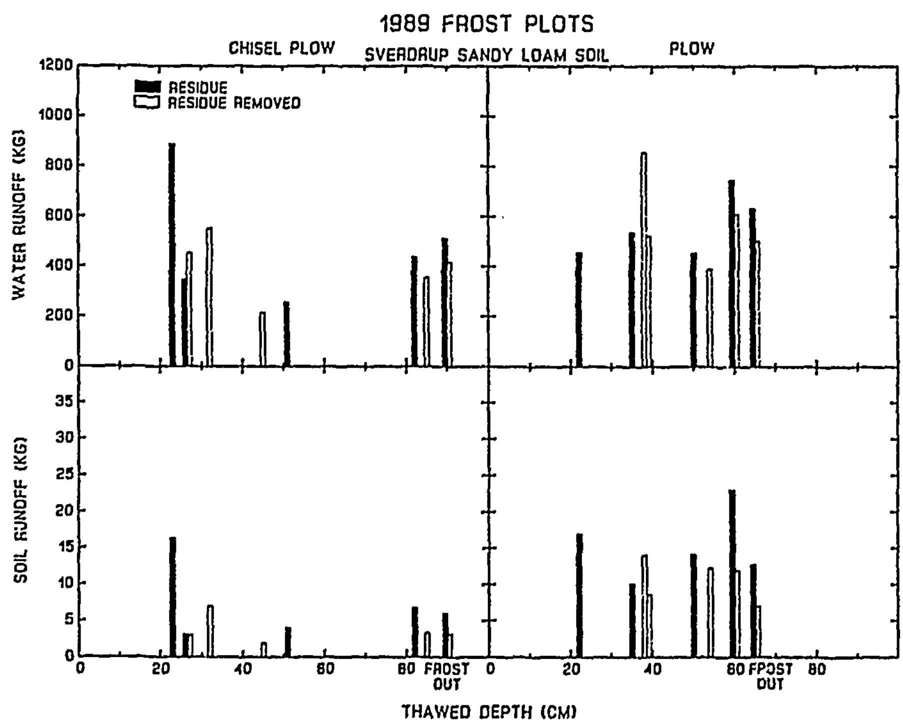

Figure 4. Total runoff (kg) and soil loss (kg) observed at different depths of soil thawing in 1989 as a result of 96 mm of simulated rain applied to a Sverdrup sandy loam under chisel and plow tillage management, each with and without residue.

Despite a higher surface BD, Sverdrup sandy loam yielded lower total runoff and soil loss values than did Barnes loam. In addition, chisel plots usually yielded lower runoff and soil loss rates than did plowed plots. This point is particularly evident for soil loss from both residue and residue-removed plowed plots which show soil loss values roughly twice those for chiseled plots.

The data presented show differences in physical properties related to soil type. These properties are modified by tillage-residue management treatments in complex ways. Trends are apparent that indicate these properties, along with depth to frost (thaw depth), affect the runoff and soil loss observed from partially thawed soil during simulated rainfall events. These trends become more readily apparent if the lumped data for all soils and treatments are analyzed with a stepwise multiple regression technique to determine the relation between (1) the dependent variable total water runoff (TWR) and the independent variables of BD, RR, percent water in the surface 150 mm of soil (W), MWD, and thaw depth (TD) and (2) the dependent variable of total soil loss (TSL) and the above-listed independent variable plus the addition of TWR.

Stepwise multiple regression analysis yielded the two equations:

$$TWR = 4.3435*W+199.41*BD-92.683*RR+16.273*MWD+0.014085*TD \quad [1]$$

$$TSL = .024065*TWR-.011079*W+.19783*BD-.30445*RR+1.2738*MWD+.00169*TD \quad [2]$$

with r^2 values of 0.84 ($P<.01$) and 0.72 ($P<.01$), respectively. Equation 1 indicates that runoff increases with increases in BD, water content (W), and MWD but, since TD is entered as a negative value, decreases with increased TD and also RR. The analysis also showed that the greatest impact on TWR was caused by BD followed by W; the least impact was caused by TD.

Equation 2 indicates that soil loss increases with increased runoff, BD, and MWD but decreases with increased RR, W, and depth to frost. In this case, the greatest impact on soil loss was caused by TWR followed by MWD, RR, and TD; the least impact was caused by W followed by BD.

In summary, regression analyses indicate that 84 percent of the variations observed in TWR under all conditions of soil and tillage-residue management can be explained by W, BD, RR, MWD, and TD. Similarly, 72 percent of the variations observed in TSL can be explained by the same five variables plus TWR.

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SEED ZONE TEMPERATURE AND MOISTURE CONDITIONS IN A PARTIALLY FROZEN SOIL-CROP RESIDUE SYSTEM

by

William R. Bidlake, Gaylon S. Campbell, and Robert I. Papendick[1]

INTRODUCTION

Tillage and residue management options and their effects on soil temperature and moisture may be explored with a mechanistic model for heat and water transport. Several workers have recognized the utility of mechanistic models for predicting heat and water transport in systems of soil and crop residues (Ross et al. 1985, Bristow et al. 1986, Flerchinger 1987). The objectives of this study were: 1) to identify management options that promise to promote rapid crop establishment within the framework of conservation tillage, and 2) to provide a mechanistic model of heat and water transport applicable to partially frozen soil-crop residue systems. The model is adapted from Bristow et al. (1986) to include effects of soil freezing and thawing, and to include effects of varying residue configurations on heat and water transport in fields cropped for small grain production.

PROCEDURES

Heat and water transport model

The model is designed to simulate heat and water transport under a variety of management practices and the configuration of the simulated soil-crop residue system is flexible. In the most complex configuration (Fig. 1), the soil is covered by two distinct layers of crop residues. The lower layer consists of matted straw and chaff and the upper layer is standing straw. Simulation of heat and water transport in the soil-crop residue system is accomplished through numerical solution of the governing

STANDING STRAW

MATTED STRAW

SOIL

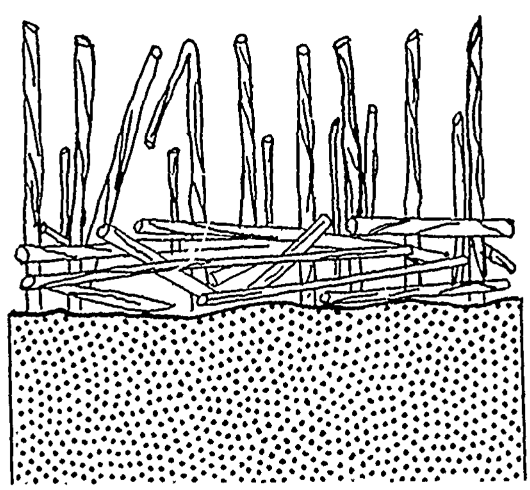

Figure 1. Soil-crop residue system.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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equations for heat and water transport. Solutions are carried out similar to Bristow et al. (1986) in that the soil and crop residues are linked into a single system for heat and water transport.

A one-dimensional heat balance equation for an unsaturated, partially frozen soil may be written (Fuchs et al. 1978)

$$C \frac{\partial T}{\partial t} - \rho_i L_f \frac{\partial \theta_i}{\partial t} = \frac{\partial}{\partial z}(-q_h - L_f q_l - L_v q_v) + R_n \qquad [1]$$

where symbols are defined in the Appendix. Net radiation is zero below the soil surface and may be calculated from analysis of the surface short-wave and long-wave radiation budgets.

A one-dimensional heat balance equation for a crop residue canopy may be written

$$C \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}(-q_h - L_v q_v) + R_n \qquad [2]$$

The short-wave radiation balance of the crop residue canopy is calculated after Norman and Jarvis (1975). The matted straw and standing straw components are each divided into three layers of equal thickness. By assuming that the residue elements do not transmit short-wave radiation, element overlap within layers can be ignored and the residue canopy can be divided into the small number of layers of varying residue area index. Solution of the long-wave radiation balance is accomplished by using the system of crop residue layers described for the short-wave balance.

An equation describing one-dimensional water transport in an unsaturated, partially frozen soil can be written (Fuchs et al. 1978)

$$\rho_l \frac{\partial \theta_l}{\partial t} + \rho_i \frac{\partial \theta_i}{\partial t} = \frac{\partial}{\partial z}(-q_l - q_v) + U_s \qquad [3]$$

water transport in the residue canopy is assumed to take place only in the vapor phase. Vapor density equilibrium is not assumed to exist between residue elements and the air bathing them. An equation for one-dimensional vapor transport in a crop residue canopy may be written (Bristow et al. 1986)

$$\rho_{br} \frac{\partial w_r}{\partial t} + \frac{\partial \rho_v}{\partial t} = \frac{\partial}{\partial z}(-q_v) + U_r \qquad [4]$$

The atmospheric boundary condition consists of radiative and forced convective components, and rainfall. The radiative components are described by atmospheric short-wave direct and diffuse radiation fluxes, and by a downward long-wave flux from the atmosphere. The convective component is described by atmospheric air temperature, vapor density, and wind speed. The rainfall component is described by a flux of water to the soil-residue system. The soil boundary condition is described by soil temperature and soil water matric potential.

Interception and distribution of rainfall within the residue canopy follows directly from Bristow et al. (1986). Throughfall and drip transport in the residue canopy are modeled after penetration of short-wave radiation from an elevation angle of $\pi/2$. All throughfall and drip intercepted by a residue layer are added to the source term of the water balance equation until the residue is nearly saturated. Additional interception is converted to a drip flux arriving at the next layer down. The entire flux of throughfall and drip at the soil surface are added to the source/sink term.

Solutions to the heat and water transport equations are approximated using numerical methods. Each of the transport Eqs. [1-4] is written in difference form. These difference equations are written for each of a vertical series of n nodes spaced through the soil-crop residue system. Heat and water transport are assumed to occur only at nodes. Solution proceeds by discrete time steps. Steady flow of heat and water is assumed to take place between adjacent nodes for the duration of a time step. Atmospheric and soil boundary conditions are applied and the system of equations is solved using the Newton-Raphson iterative technique (Campbell 1985). Coupled heat and water transport may be simulated in a frozen soil by iterating between heat and water solutions over the same time interval. Solutions for soil temperature and soil water matric potential become difficult to obtain as partially frozen soils approach saturation. Further approximations to soil heat and water balance equations may be obtained by solving heat and water balances separately (Bidlake 1988).

Study site, tillage and residue treatments, and field measurements

The study site is in the Tanana River Valley in the Interior Basin of Alaska (63 deg. 58 min. N) and near milepost 1408 of the Alaska Highway. The silt loam soils are in the Volkmar series (Aeric Cryaquept; Schoephorster 1973) and are derived from aeolian deposits of mica schist parent material. Mean annual air temperature is -2 deg. C and mean annual precipitation is approximately 30 cm.

Two tillage and residue treatments were selected and instruments were installed during fall, 1986 and spring, 1987. The crop was spring barley A conventional-till plot was selected as a maximum disturbance and minimum residue condition. The plot was disked in the fall with an offset disk. Crop residues were raked and removed from the plot. The plot was then disk-seeded in the spring. Second, a no-till plot was selected to represent the minimum disturbance and maximum residue condition available to farmers. In this treatment, all crop residues were left in the field and seed was drilled directly into undisturbed soil. Planting was conducted on day of year 131 (May 11) and the target seeding depth was 6 cm for both plots.

Soil temperature, soil water matric water potential, depth to frozen soil, and climatic variables were monitored from day of year 107 to 160 (April 17 to June 9). This time period encompasses the critical stand establishment phase of small grain production in interior Alaska. Soil temperature was measured with insulated copper-constantan thermocouples. Measurements were made with a field portable data logger (21X data logger, and AM-32 multiplexer; Campbell Scientific Inc., Logan, Utah). Soil temperature was measured and recorded hourly at 6, 10, 18, and 30 cm depths at three locations on each plot, and at one location on each plot at 60 and 80 cm depths. Soil water matric potential at 6 cm was measured with the filter paper method on excavated samples. Midday measurements were at 4 to 6 random locations on each plot at 3 to 5 day intervals beginning when the soil had thawed to 6 cm. Depth to frozen soil was measured in mid to late afternoon. A piece of welding rod (dia. 4 mm) was used to probe the depth to frozen soil.

Incoming short-wave radiation was monitored with a pyranometer (Kipp and Zonen, Delft, Holland) and a field portable data logger (CR7; Campbell Scientific Inc., Logan, Utah). Radiation measurements were made every 10 seconds and recorded as hourly averages (Alaska Standard Time).

Other meteorological variables were monitored continuously during the study. A dual function relative humidity/air temperature sensor was suspended 1.5 m above the ground at a border between the two study plots (207 RH Probe; Campbell Scientific Inc., Logan, Utah). The electrical resistance hygrometer employs a sulphonated polystyrene chip (PC-RC11; Phys-Chemical Research Inc, New York, New York). A thermistor in the unit is used to sense air temperature. The sensor was monitored with a data logger (21X) to obtain humidity and air temperature hourly. Wind speed was sensed with a cup anemometer (014A with aluminum cups, switch closure operation; Met One Inc., Grants Pass, Oregon). The anemometer was suspended 1.5 m above the soil surface of the conventional-till plot and was operated with the 21X data logger. Wind speed was integrated over 60 second intervals and recorded as an hourly average. Daily precipitation was measured with a wedge rain gauge.

The model was operated to simulate soil temperature, soil water matric potential, and depth to frozen soil in the conventional-till and no-till plots over the study period. Initial water contents to 30 cm depth were obtained from soil samples extracted from holes drilled in the frozen soil. Initial temperature profiles were obtained from sensors already in place. The lower boundary of the soil-crop residue system was set at 80 cm. Values for many of the important transport parameters were obtained from sampling at the study site (Bidlake 1988).

RESULTS

Soil in the conventional-till plot was warmer than soil in the no-till plot. Measured soil temperature at the 6 cm depth was 1-3 deg. higher in the conventional-till plot that in the no-till plot during most of the study period. Soil temperature as simulated by the model generally compared well with measured temperature. During most of the the study period (day of year 107 to 150), simulated mean daily soil temperature at 6 cm was 1 to 2 deg. higher than measured temperature for both plots (Fig. 2). During the final 10 days of the simulation, simulated temperature closely approximated measured soil temperature. The difference in simulated soil temperature between conventional-till and no-till soils was generally within 1.5 deg. of the measured soil temperature difference.

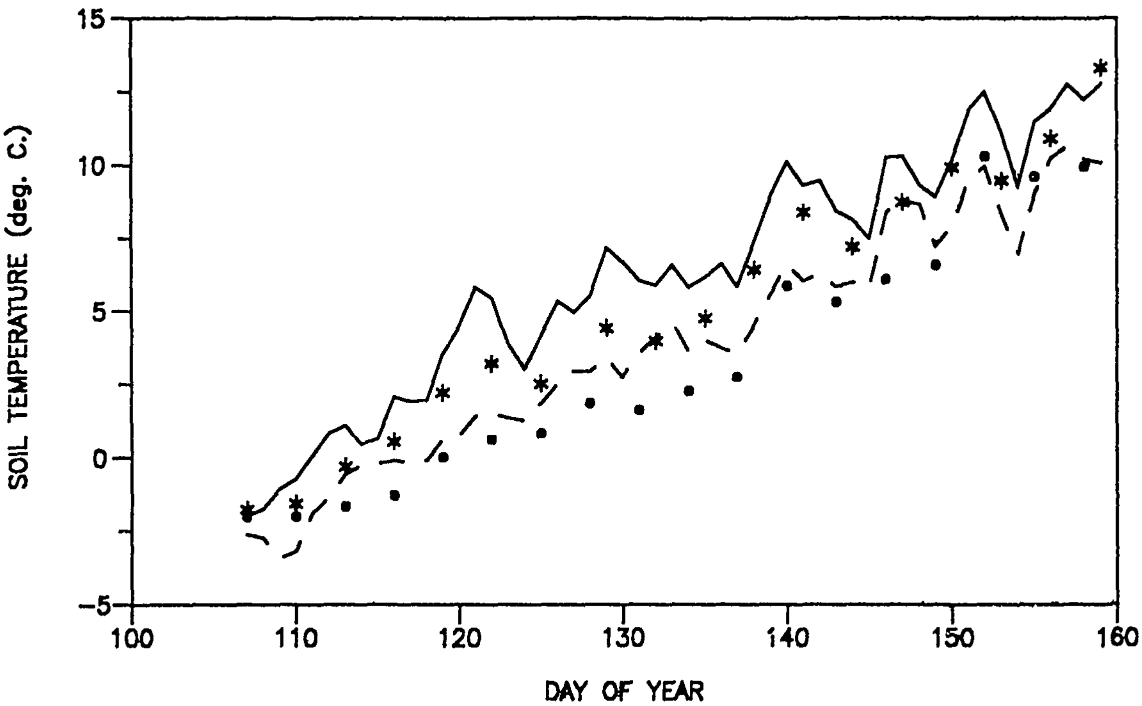

Figure 2. Average daily soil temperature at 6 cm depth in the conventional-till (solid line, * symbol) and no-till (dashed line, • symbol). Lines represent daily average simulated values. Symbols represent averages of measured values with the standard deviation (n= 4).

Soil water matric potential for unfrozen soil at the 6 cm depth was lower in the conventional-till soil than in the no-till soil (Fig 3). Measured soil water matric potential ranged between -15 and -30 J/kg in the conventional-till soil, and it ranged between -8 and -15 J/kg in the no-till soil. Soil disturbance during sampling may have produced some error in measurement of soil water matric potential because of the

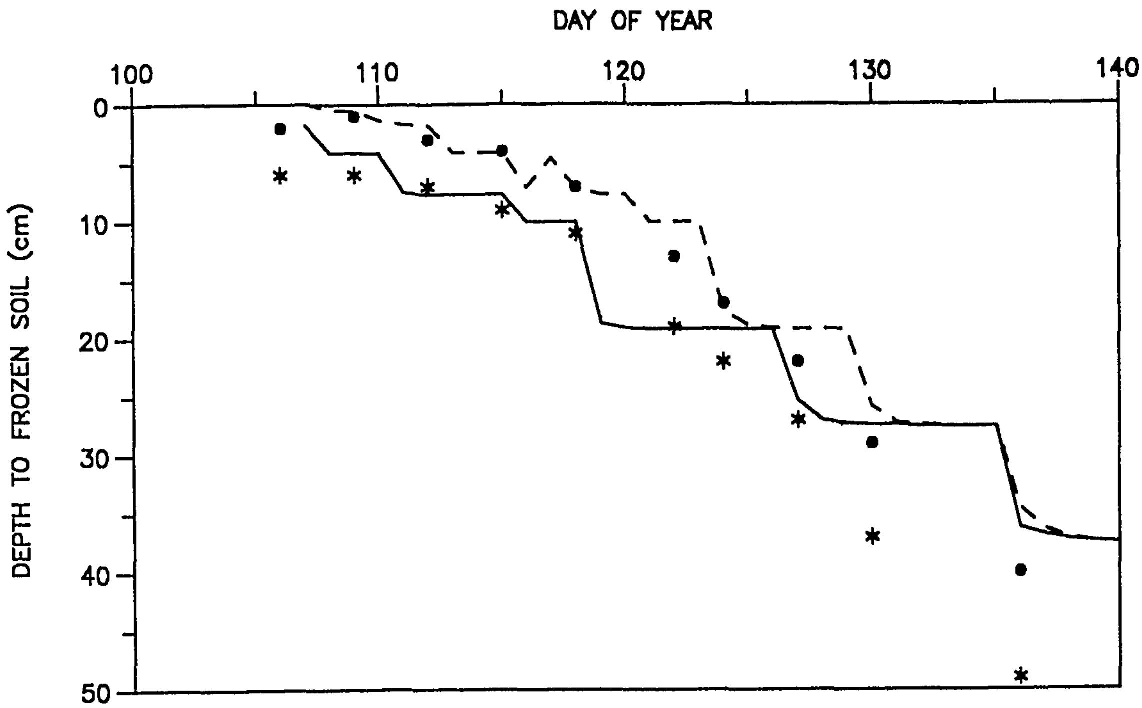

Figure 4. Measured and simulated depth to frozen soil in conventional-till (solid line, * symbol) and no-till (dashed line, • symbol) plots. Measurements were made in the afternoon and simulated values are the maximum daily depth from model simulation.

The oscillating behavior of simulated soil water matric potential in the conventional-till plot is related to thawing of discrete soil layers as simulated by the model (Figs. 2, 3). In the model, ice is immobile, therefore, underlying frozen soil layers are not a source for water to replace that which is ultimately lost from the surface by evaporation. As a result, water from the seed zone flows toward the surface under matric potential gradients, seed zone water is depleted, and soil water matric potential in the seed zone decreases. Once a frozen layer below the seed zone thaws, however, the water is available to flow toward the surface and soil water matric potential in the seed zone increases. The oscillating patterns of soil water matric potential in simulation for the conventional-till plot was due to the intermittent supply of water from soil layers below the 6 cm depth. This result suggests that a thawing soil at depth is a source of water to replace seed zone water that is ultimately lost by evaporation. In the case of the no-till soil, the effect of intermittent water supply from below was not evident because evaporation was suppressed by surface residues to the extent that the initial water at the 6 cm depth was never substantially depleted.

Weather records for interior Alaska indicated that the study period of 1987 was driest such spring period in over 50 years. Despite this fact, simulated and measured moisture conditions in the seed zone of each both plot were likely not inhibiting for germination and emergence. Cool seed zone temperature is likely a limiting factor for germination and emergence of small grains in interior Alaska.

CONCLUSIONS

A mechanistic model for heat and water transport has been presented with which to simulate seed zone temperature and moisture conditions in partially frozen soil-crop residue systems. Simulated and measured results indicated that conventional-till soils are warmer and dryer than no-till soils. Seed zone moisture conditions are likely not dependence of matric potential on soil structure. Simulated soil water matric potential in the conventional-till plot was lower than the measured value until day of year 152 (Fig 3). Simulated soil water matric potential dropped below -45 J/kg but measured values never dropped below -30 J/kg. Simulated soil water matric potential in the conventional-till plot oscillated throughout the study period while the measured value generally changed by less than 6 J/kg between samplings. Simulated and measured soil water matric potential at the 6 cm depth in the no-till plot did not vary by more than 8 J/kg, and simulated matric potential did not exhibit the oscillation typified in simulations for the conventional-till plot.

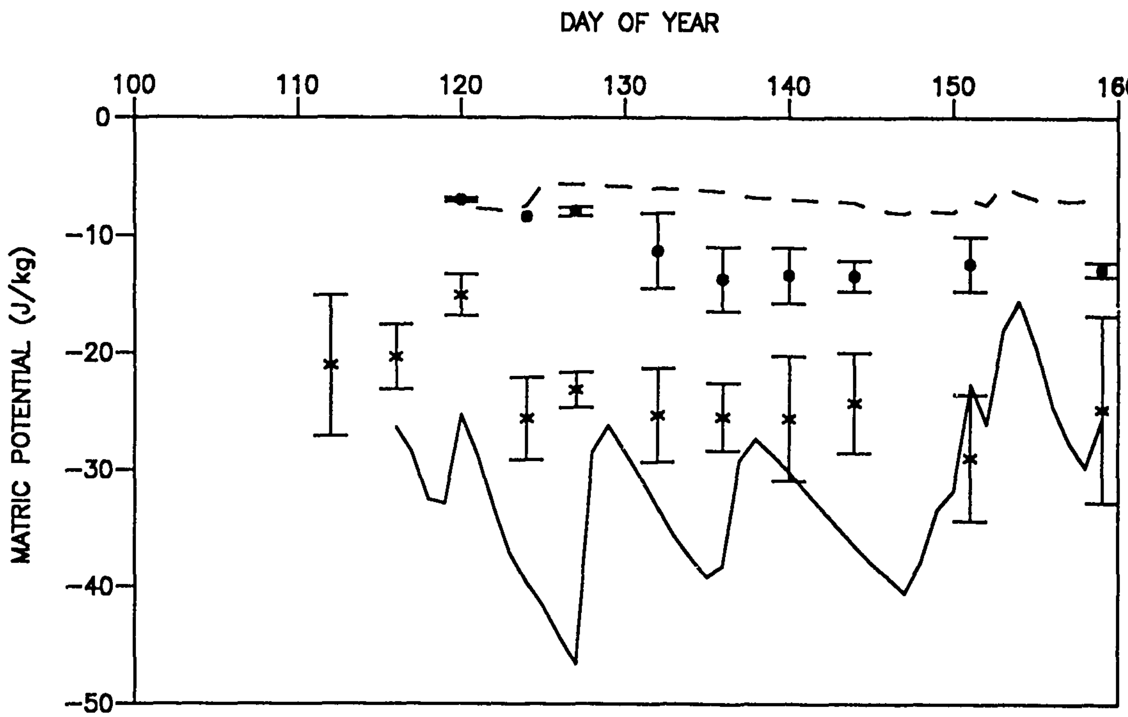

Figure 3. Soil water matric potential at 6 cm depth in the conventional-till (solid line, * symbol) and no-till (dashed line, • symbol). Lines represent daily average simulated values for unfrozen soil. Symbols represent averages of measured values with the standard deviation (n= 4).

The conventional-till soil was unfrozen at a greater depth than the no-till soil during the entire study period. Measurements of afternoon thaw depth indicated that the conventional-till soil thawed to a depth that was 4-9 cm greater than the thaw depth for the no-till soil (Fig. 4). The daily maximum of simulated depth to frozen soil remained within 3 cm of the measured value for both soils until day of year 130. After that time, measured thawing outstripped simulated thawing. Simulated depth to frozen soil was within one standard deviation of the measured depth for all but the final measurement in the conventional-till plot (sd. bars not shown).

Simulated soil thawing proceeded in discrete steps and the step size tended to increase with depth. The reason for this simulation result is that energy liberated in thawing controls the temperature of a discrete layer of soil until it warms to above the freezing point. At that time, the entire layer suddenly thaws out and creates a sharp increase in thaw depth that is equal to the thickness of the layer. Layer thickness was increased progressively with depth for the simulations and this was reflected in the thawing step size.

inhibiting to stand establishment for small grains in interior Alaska. Results from this study suggest that thawing soil at depth may yield water to keep the seed zone moist. Greater moisture availability in no-till systems may be used to ameliorate adverse temperature conditions by planting at shallower depths in order to advance the goals of crop production and soil conservation.

APPENDIX: SYMBOL DEFINITION

C Volumetric heat capacity (J/m³-°C).

T Temperature (°C).

t Time (s).

L_i Heat of fusion (J/kg).

θ_i Volumetric ice content.

z Depth (m).

q_h Sensible heat flux (W/m²).

q_l Liquid water flux (kg/m²-s).

L_v Heat of vaporization (J/kg).

q_v Vapor water flux (kg/m²-s).

R_n Net radiation (W/m³).

$\rho_a c_p$ Volumetric heat capacity of air (J/m³-C).

ρ_l, ρ_i Densities of liquid water and ice (kg/m³).

θ_l Volumetric liquid water content.

U_s Source term for precipitation at the soil surface (kg/m³-s).

ρ_{br} Crop residue bulk density (kg/m³).

w_r Gravimetric water content of residue elements.

ρ_v Vapor density of the bathing air in the residue canopy (kg/m³).

ρ_{vr} Vapor density at the surface of residue elements (kg/m³).

U_r Source/sink term for precipitation, and evaporation and condensation in the in the residue canopy (kg/m³-s).

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TILLAGE AND RESIDUE MANAGEMENT SYSTEMS AFFECT WINTER SOIL TEMPERATURES

By

R.F. Cullum, B.S. Sharratt, C.E. Lewis[1]

INTRODUCTION

The interior of Alaska has a growing season climate favorable for spring cereal crop production. This region, however, is not conductive to the establishment of winter annual crops due to winter-kill from low soil temperatures and snowmold (McBeath, 1985). The average January minimum soil temperature at 20 cm reported by Ping (1987) was -6°C. Average minimum winter air temperatures are below -30°C and indicate the possibility for shallow soil temperatures of less than -20°C, the lower temperature threshold to winter wheat survival (Ulanova, 1975). Winter cereals have been tried in interior Alaska with little success.

The vulnerability of subarctic soils to erosion when cleared of the native vegetation necessitates a soil and water conservation plan to minimize soil loss (Boyer, 1985). Winter soil temperatures are higher when the soil surface is mulched instead of being exposed (Aase and Siddoway, 1980). Soil freezing is generally not as deep (Pikul *et al.*, 1986; Benoit and Mostaghimi, 1985) and the number of soil freezing events are fewer when utilizing no tillage in cereal cropping systems.

The purpose of this study was to characterize the winter climate of soils as influenced by tillage and straw management in interior Alaska. This baseline information may be used in establishing a soil climatology to assess the potential of subarctic soils to diversify crop production.

PROCEDURES

Mile 1408 of the Alaskan highway near Delta Junction, Alaska was chosen as the study site (Figure 1). This area, located within the Tanana River floodplain at 350m above sea level, is characterized by a mean annual temperature of -3°C. Annual precipitation totals 250 to 300 mm. Soils have been formed on stabilized deposits of loess on large glacial outwash plains and are not aggregated when newly cleared. The silt loam soil at the study site (*Typic Cryochrepts*) is typical of the area. The soils are considered moderately to severely susceptible to wind and water erosion. These soils are very low in clay (<10%) and high in organic matter (>6%). They are cold and remain frozen for approximately six months each year. Thus, decomposition of organic matter is slow and tree roots and forest litter remain in the soil long after land clearing (Cochran *et al.*, 1988). The research site was cleared of native spruce vegetation in 1979 and cropped to barley (*Hordeum vulgare* L.) since 1982.

Tillage and straw treatments were initiated in the spring and fall of 1983, respectively, using continuous barley cropping systems (Lewis and Cullum,1985; Siddoway *et al.*, 1984). The experimental design was split plot with tillage as main treatments (23m x 122m) and straw amount as secondary treatments (23m x 41m). The two extreme treatments were evaluated for this study. The maximum tillage-straw removed treatment (maximum-till) consisted of fall and spring disking with threshed and standing straw removed following harvest. The no tillage-straw remaining treatment (no-till) consisted of no tillage with threshed and standing straw remaining after harvest. These treatments were selected to maximize differences in the measured characteristics of the crop and soil.

Presented at the International Frozen Soils Symposium, March 21-22,1990; Spokane, Washington.

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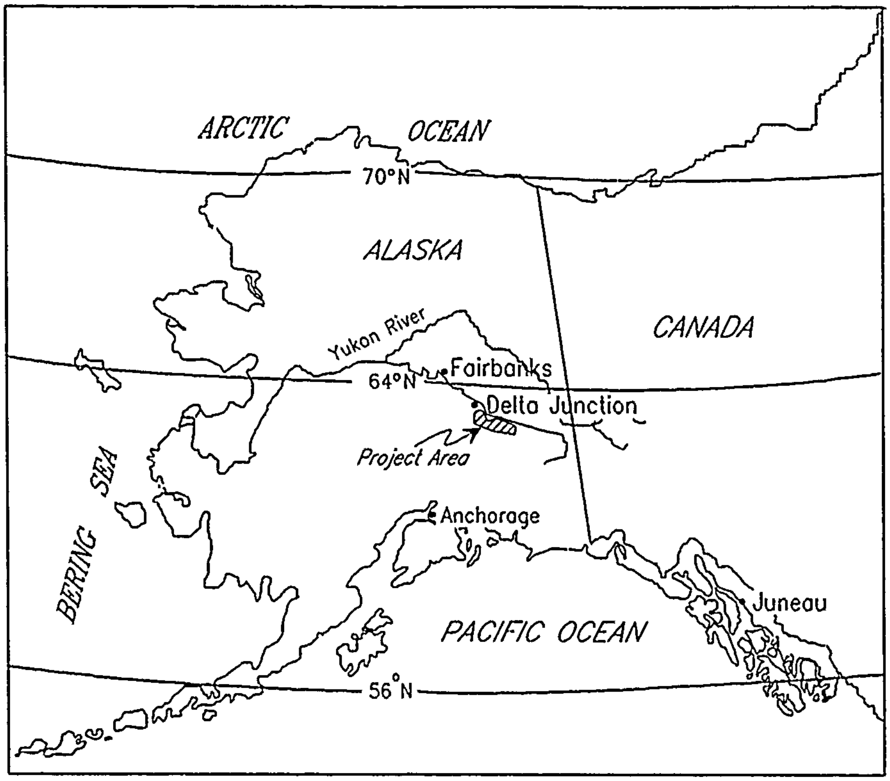

Figure 1. Location of the conservation tillage project.

Thermocouples were used to measure soil temperatures and were installed in each plot in the summer of 1983. They remained buried throughout the duration of the three year study. Thermocouples were placed at depths of 0, 5, 15, 30, 50, 60, 76, 102, and 123 cm. Each thermocouple consisted of three leads wired in parallel to obtain an average temperature for each depth. A micrologger recorded daily averages, minimums, and maximums from 10-minute readings.

Soil frost was determined by use of frost tubes (Richard *et al.*, 1976) that were installed to a one-meter soil depth. Each tube extended two meters above the soil surface and were marked to measure snow depth. These tubes were read once a week beginning the middle of September.

RESULTS

The data in Figure 2 indicates that tillage-residue systems influenced the amount of snow retained on the soil. The no-till system consistently trapped more snow than the maximum-till system (7 cm vs 32 cm on Sept 30 or 2 cm vs. 30 cm on Dec 4). The greater amount of snow on the no-till system was attributed to the snow entrapment by the standing stubble. Stubble also prevents removal of snow during the high wind periods from October through April (Figure 2).

Tillage-residue effects on snow depth influenced soil frost penetration. The maximum-till treatment normally froze first and deeper than the no-till treatment as seen in Figures 2, 3, and 4. The no-till plot trapped and retained more snow than the maximum-till plot. Frost penetration in the no-till treatment was less than observed in the maximum-till treatment. Snow depth and frost penetration data indicate tillage-residue systems influenced snow accumulation and resulted in decreased frost penetration when standing stubble remains over winter. The standing stubble entrapped snow which insulated the soil retarding frost penetration. The no-till treatment retained the snow, thus preventing the extreme -50°C air temperatures from contact with the soil surface. The soil surface of the maximum-till plot were exposed to these sub-zero temperatures and wind speeds greater than 24 kph producing a two to five-cm layer of

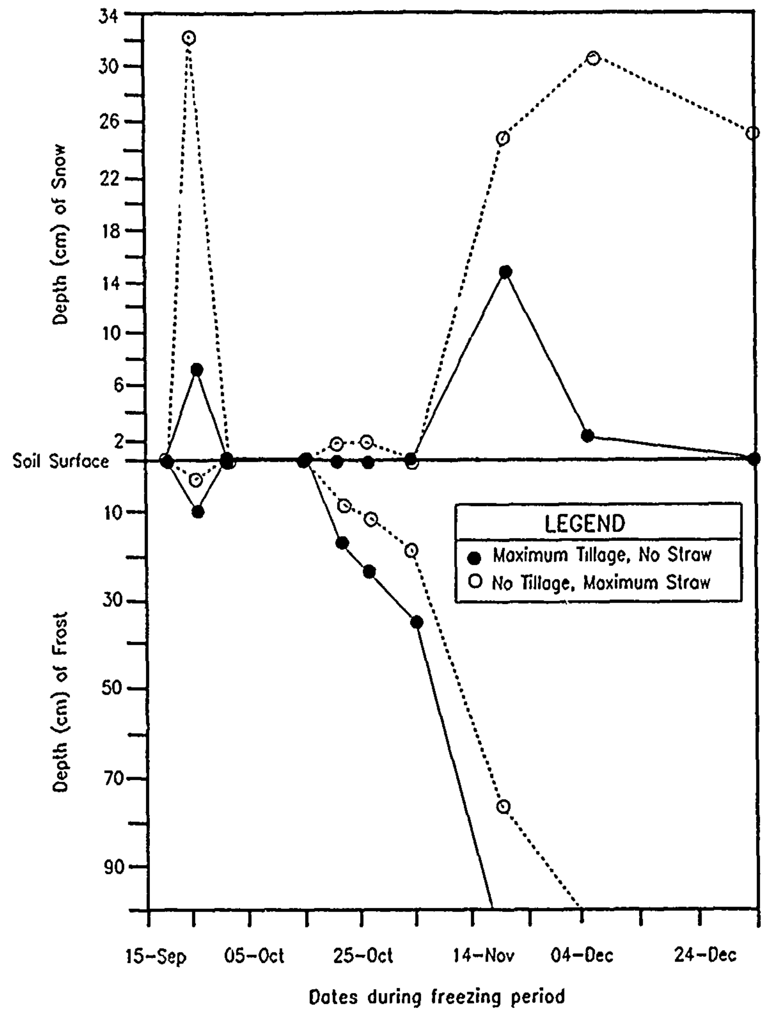

Figure 2. Snow and frost depth during a typical fall in the extreme residue and tillage treatments.

dehydrated soil through sublimation. This process resulted in a layer of soil with water content at the wilting point. This layer, a dust mulch, reduced the thermal conductivity in the maximimum-till treatment and hindered heat transfer during the thaw period (mid-April to early May) (Figure 5). By leaving standing stubble over winter in this area, earlier thaw would be expected.

The effect of tillage-residue systems on snow accumulation influenced soil temperature patterns (Figures 3 and 4). The no-till treatment showed warmer soil temperatures due to deeper snow accumulation than the maximum-till treatment. During the freeze periods, the no-till treatments consistently maintained the highest average soil temperatures. Benoit *et al.* (1986) presented similar results. The no-till treatment could thus be planted several days (3 to 10) earlier than the other treatment. From mid-May through mid-June (the period of barley canopy development) insignificant temperature differences resulted between the treatments; however, during late June and throughout July the temperatures of the no-till treatment would be greater than maximum-till treatment (Figure 6 and 7). An explanation for this phenomena may be a result of lower moisture content attributed to maximum tillage which would retard heat transfer into the soil than would otherwise be found in no-tillage systems.

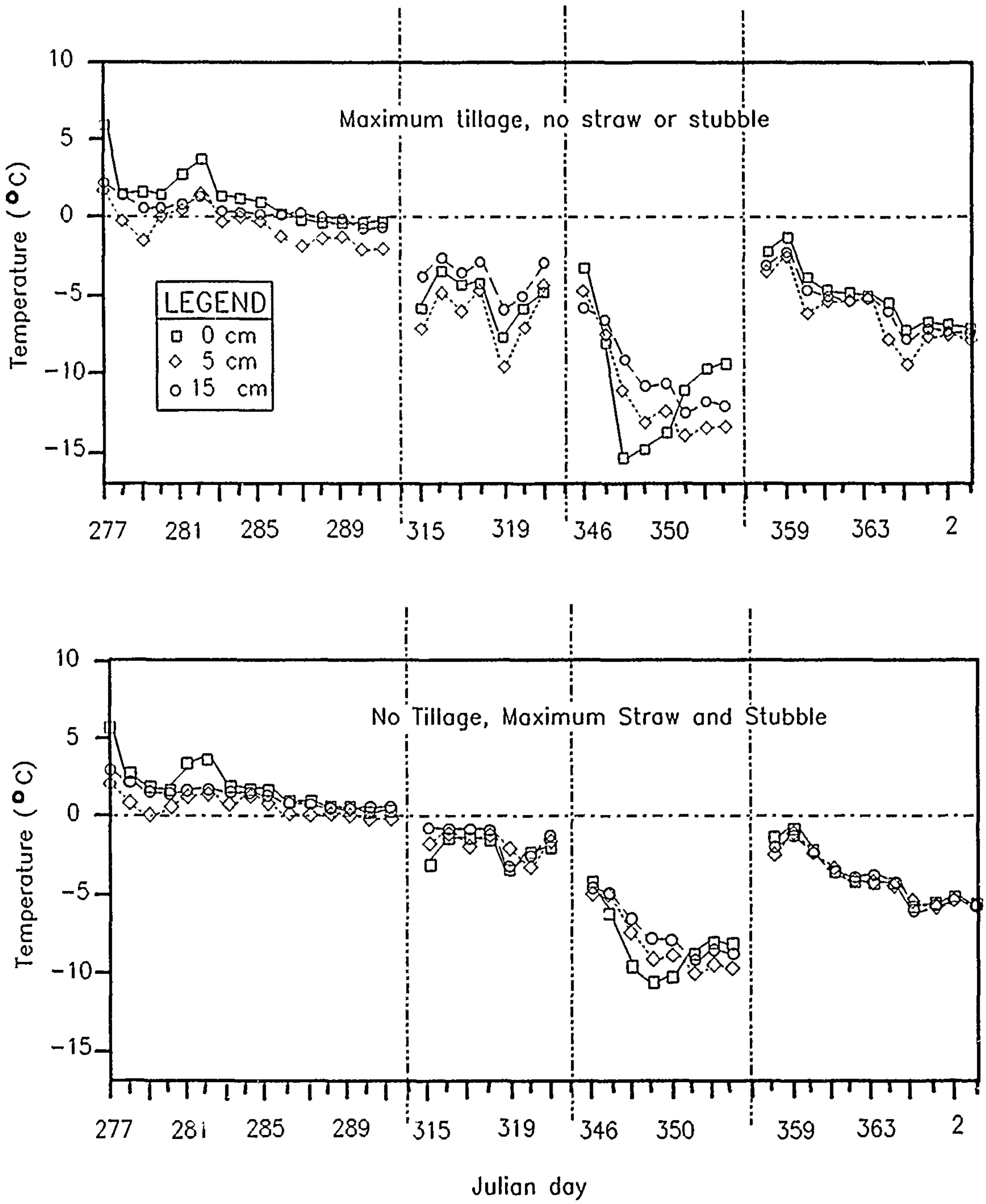

Figure 3. Average daily soil temperatures from one-hour readings during 1985 for plots in a conservation tillage experiment.

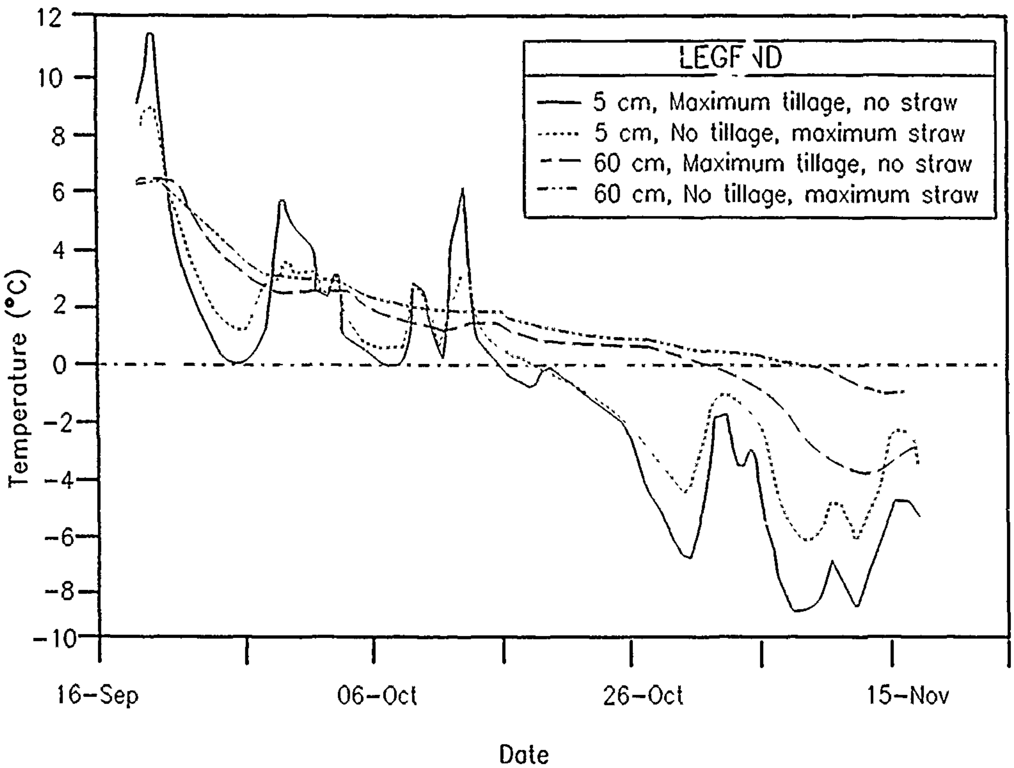

Figure 4. Average daily soil temperatures from one-hour readings during the freeze period of 1986 at soil depths of 5 and 60 cm.

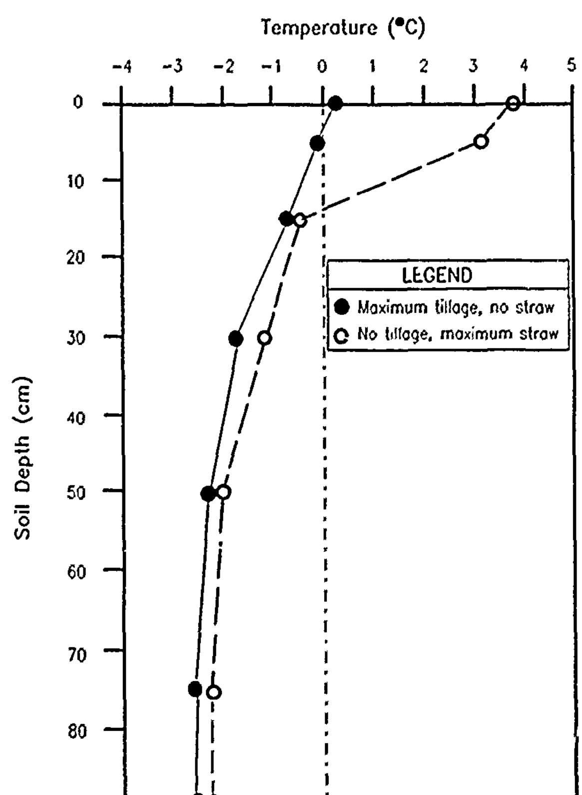

Figure 5. Temperature profiles of barley plots on April 30, 1986.

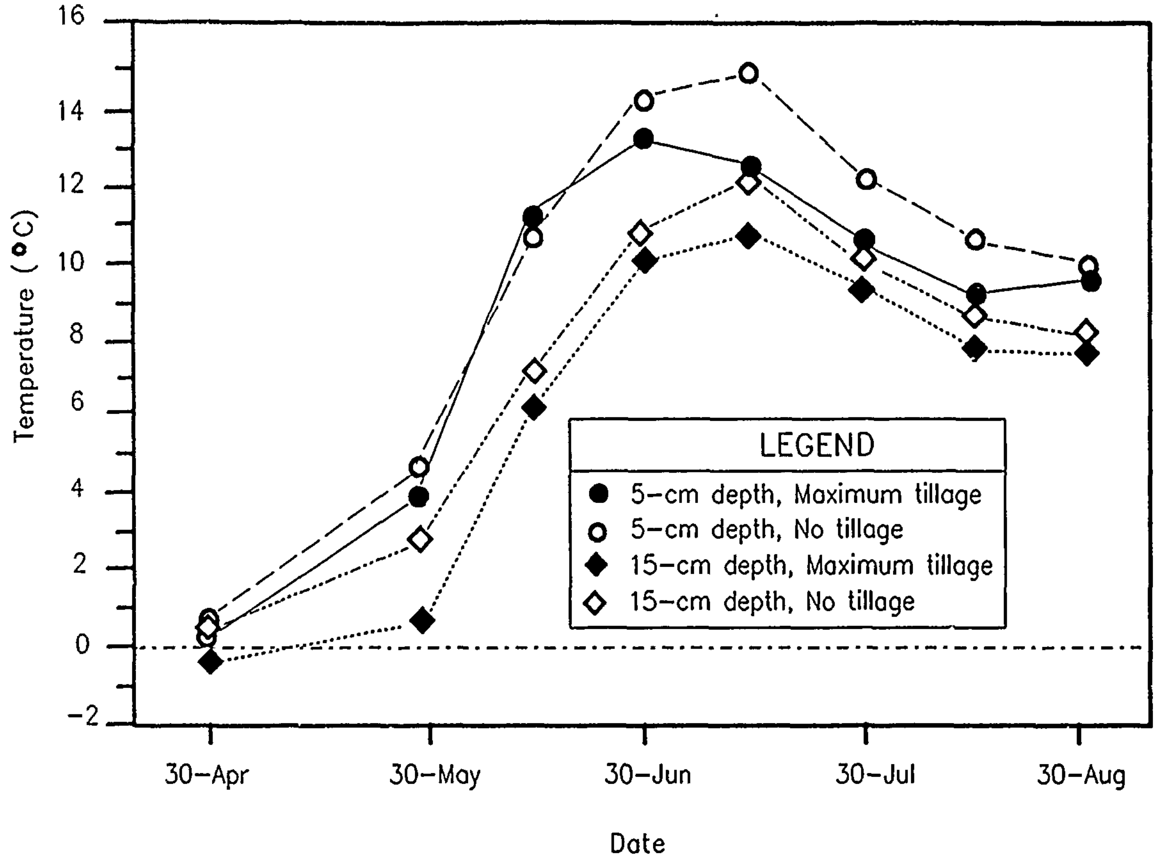

Figure 6. Five-day average soil temperatures at the 5 and 15 cm depth for the 1985 barley growing season under the two extreme tillage-residue treatments. The averages were determined from hourly readings.

Conclusions

A distinct tillage-residue effect on snow accumulation and soil temperatures which affect frost penetration was shown. These effects mean that tillage-residue systems have an influence on conditions found at spring planting (early May). Reduced tillage with standing stubble results in a greater accumulation of snow and maintains the snow cover in high wind areas resulting in reduced frost penetration, early frost disappearance, and warmer soil temperatures. Maximum tillage resulted in less snow accumulation, deeper frost penetration, and colder soils which tended to retain frost later in the spring (mid-May). During winter, this tillage forms a distinct layer, a dust mulch, which retards heat flow retarding early spring thaw. This implies that management of tillage-residue systems might promote earlier planting lengthening the growing period of this subarctic region. The data favor development of tillage-residue systems that minimize tillage and maximize residues left on the field. The residues to be effective should be left in the form of standing stubble.

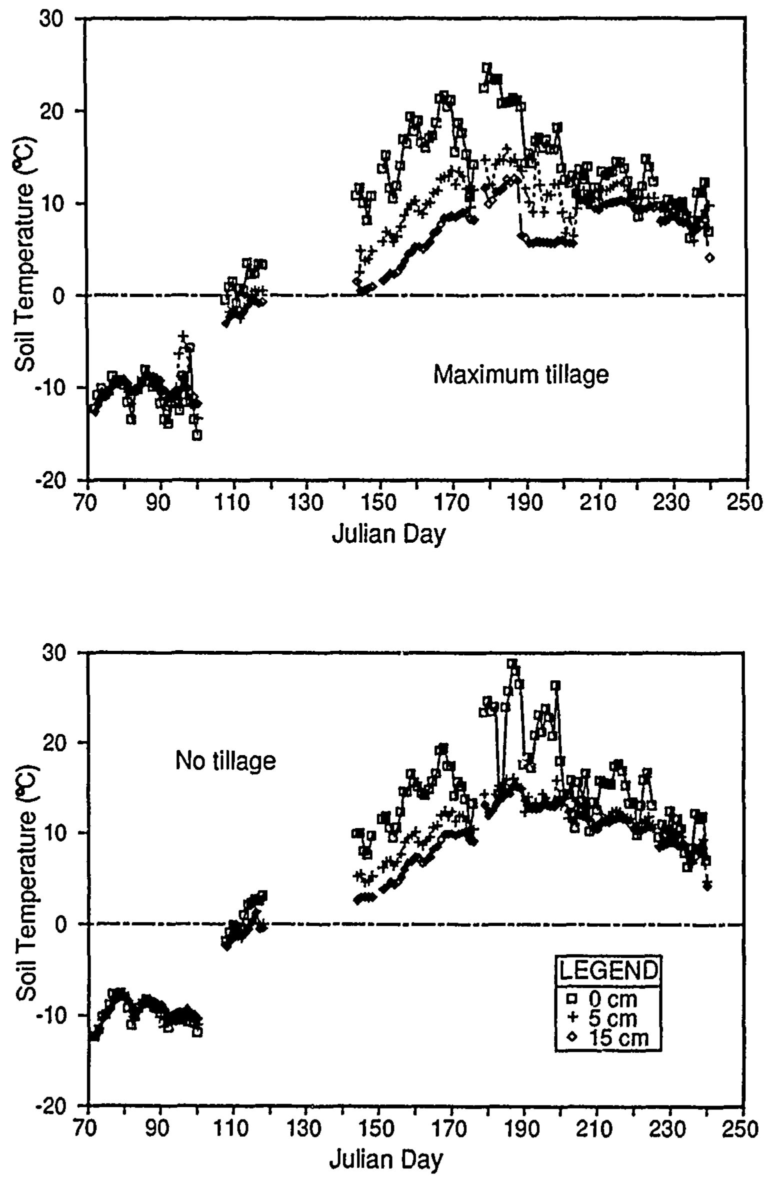

Figure 7. Average daily soil temperatures from one-hour readings during 1986 for plots in a conservation tillage experiment.

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THE IMPACT OF FROZEN SOIL ON PRAIRIE HYDROLOGY

by

R.J. Granger and D.M. Gray[1]

INTRODUCTION

Much of the Canadian Prairies can be classified as a cold climate region, at least during those five months of the year (November to March) when air temperatures, and soil temperatures, are below the freezing point for water. Its hydrology is therefore strongly affected by the seasonal fluctuations in temperatures and the relative lengths of the cold and warm seasons. For example, the extent and continuity of soil freezing, the relative amount of the mean annual precipitation which occurs as snow, and the length of the period of soil thaw after the disappearance of the seasonal snowcover impact directly on local hydrology.

The seasonal fluctuations in temperature have also affected the nature and scope of hydrological investigations in the region. Whereas numerous studies have been conducted during the growing season on evapotranspiration requirements and root-zone soil moisture withdrawal patterns of crops, only a limited number of investigations have studied the soil moisture regime during the winter months. The seasonal difference in research activity is understandable given the inclement conditions under which equipment and personnel must operate during a Prairie winter. As a consequence, there is a general impression that moisture conditions remain relatively static over winter and that the effects of the freeze-thaw sequence on the soil moisture status are unimportant, with the exception of the effects on snowmelt infiltration. It is one purpose of this paper to demonstrate that the moisture regime of a Prairie soil remains active during the winter months, and that significant moisture transfers can occur within the root zone of a crop.

The mean annual precipitation throughout a large part of the grain-producing sector of the Canadian Prairies is in the range of 300-380 mm. Of this total the relative amounts occurring as snow and rain vary widely from year to year, with an average ratio of approximately 30% as snow. Neglecting large-scale water diversions and the development of irrigation schemes, snow represents the major source of manageable fresh water available for agricultural and municipal use. The role of snow as a water resource for agricultural production is still not well understood. In the wake of recent "dry" years, there is increased interest in the potential of managing snow to increase soil water reserves. The basic premise underlying the use of snow management practices for this purpose is that an increase in snow water will result in an increase in snowmelt infiltration: however, the relationship between snow cover and infiltration is not straightforward. The paper examines the potential of the snow resource for augmenting soil moisture reserves in terms of the infiltration characteristics of frozen soils.

The soil moisture status at the end of melt of the seasonal snowcover is not usually the same as that at the time of seeding of annual crops. Losses occur due to evaporation and drainage. The effects of the presence of frozen soil during the period following the disappearance of the seasonal snowcover on the disposition of soil water gained by snowmelt infiltration are discussed.

EXPERIMENTAL STUDIES

The data presented in the paper were obtained from comprehensive field investigations of soil moisture changes and infiltration into frozen soils undertaken by the Division of Hydrology, University of Saskatchewan in the Brown and Dark Brown soil zones of the province of Saskatchewan from 1979-88. The sites included a range of soil textures (fine sand ~80% sand to heavy clay ~63% clay) and a number of land use practices (fallow, stubble and grass) on arable farmland under irrigated and dryland farming practices.

Presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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Soil moisture was monitored with a two probe gamma density meter. This system provides non-destructive sampling of the density of a 20-mm thick soil layer between two access tubes (into which a cesium source and a scintillation detector are lowered) spaced about 300-mm c.c. By assuming the mass of the soil layer remains constant (no structural changes occur) changes in density of the layer can be attributed to changes in the mass of water present in the layer. The equivalent moisture change is calculated assuming a density of water equal to 1000 kg/m^3. Measurements are obtained at 20-mm increments of depth to 400 mm and at 40-mm between 400 mm and 1600 mm. Repeatability tests with the equipment gave a standard error of estimate in moisture content of about ±2.5 mm in a 1-m profile. All systems were tested and calibrated to operate reliably under cold weather conditions (to -20°C).

Profiles were monitored in the fall prior to freeze-up and during winter, snowmelt and post-melt periods, up to the time of seeding of annual crops. An attempt was made to obtain measurements at least once every three weeks during the winter and postmelt periods. During snowcover ablation the frequency of measurements was increased with the final observation taken immediately following the disappearance of the snowcover.

At several locations soil temperature probes with automatic recorders were installed to provide measurements at depths of 25, 50, 100, 200, 400, 800 and 1600 mm. Daily maximum and minimum air temperatures were also recorded. The temperature data were used to help identify layers of soil where overwinter changes occurred, periods of midwinter melt and infiltration, and the times of freeze-up, snowmelt and thaw, and to establish rates and depths of freezing and thawing.

THE WINTER PERIOD

That moisture migrates in response to a temperature gradient during freezing and thawing of soils has been demonstrated by numerous laboratory studies on small specimens of freezing, frozen and thawing soils (Jumikis, 1973; Jame and Norum, 1980; Mageau and Morgenstern, 1980). However, many questions remain unanswered regarding the phenomenon in nature: for example, a clear consensus is lacking on the relative importance of the transfers in the liquid and vapor phases, and it is yet difficult to identify those conditions which favor the migration of moisture in significant amounts during freezing.

Gray et al. (1985a) monitored the migration of moisture in response to soil freezing in-situ. They suggest a soil profile can be divided into two zones in respect to overwinter moisture changes. In the upper zone, extending from the surface to a depth of approximately 300 - 400 mm, overwinter moisture transfer (other than infiltration of meltwater) occurs primarily as a vapor and changes in moisture content are affected by those factors influencing the energy and vapor exchange processes at the air/snow and snow/soil interfaces (ie, large diurnal fluctuations in surface temperature and very strong temperature gradients). In the zone below, the lower zone, moisture changes occur by upward migration or relocation of water in response to freezing and by drainage below the freezing front. In this zone temperature gradients are less pronounced, and, when an "adequate" supply of soil water is available, moisture moves primarily as a liquid.

Figures 1 and 2 show typical plots of soil temperature and changes in soil moisture regimes at selected times during the winter of 1982-83. The Saskatoon site (Fig.1) is a silty clay soil having a relatively high soil moisture content (ranging from 24 to 45% by volume) in the fall and a water table at a depth greater than 4 m. The Outlook site is an irrigated, fine sandy loam having a fall soil moisture near field capacity (~25% by volume) and a water table at 2.55 m on Nov. 17/82. With respect to the overwinter period (before Mar. 28th) the figures show:

1) Moisture losses from the upper zone. That moisture had moved into the overlying snowcover was evinced by the coarse, rounded, large-grained texture of the snow near the soil surface (an indication of strong temperature-gradient metamorphism resulting from the flow of heat and moisture from the soil). Benson and Trabant (1973), Peck (1974) and Santeford (1976, 1978) have also reported the movement of moisture from the soil to the snowcover as a result of thermal gradients. An overwinter loss of moisture from the upper zone is typical of that observed at most sites, with the exception of those situations where significant mid-winter melting had occurred.

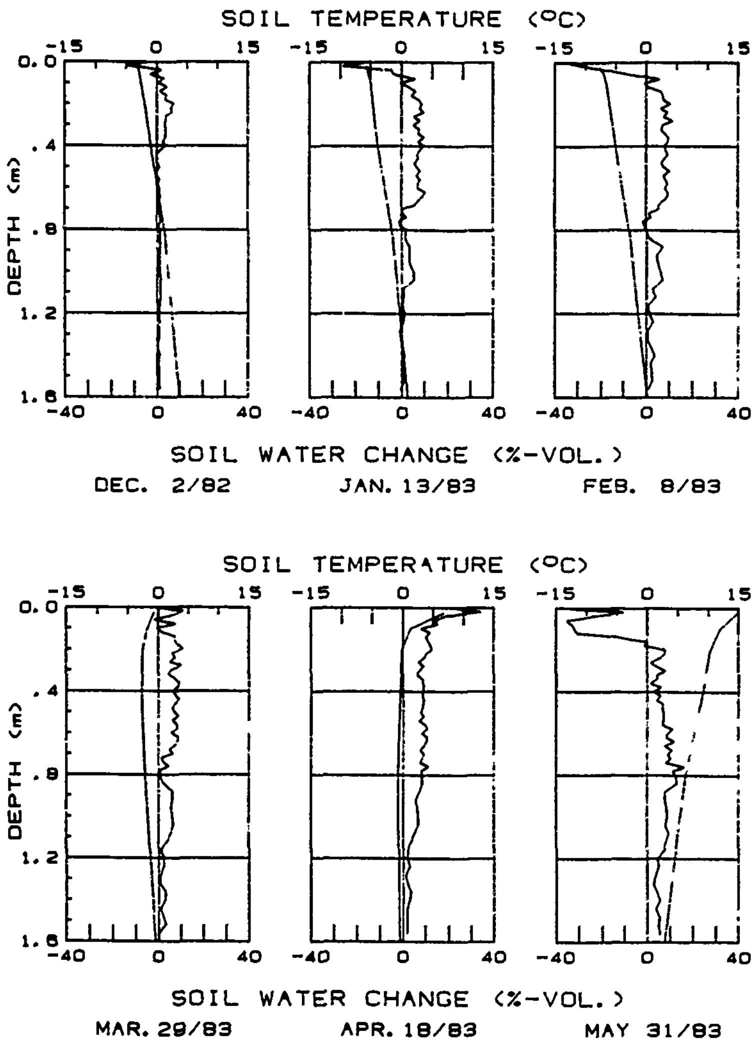

Figure 1. Profiles of soil temperature and changes in soil moisture measured in a silty clay soil at Saskatoon, Sask. Moisture changes are for the period from November 16/82 to the date shown below each frame.

2) Large fluxes of water may migrate upwards into the frozen zone in response to the freezing action. Within the 0 to 1-m depth these increases amounted to ~47.8 mm at the Saskatoon site and ~95.7 mm at Outlook. The lower extremity of the accumulation of water/ice coincides with the freezing front, and extension of the zone-of-accumulation coincides with the downward movement of the 0°C isotherm. At Outlook, a drop in the water table of 400 mm during freezing and its recovery in the spring suggests that the saturated zone is the primary source of the migrating water.

3) No pronounced changes in moisture content in the unfrozen soil. The lack of strong moisture gradients below the freezing front suggests that if moisture is moving as a liquid the transport process bringing water from below is analogous to a coupled hand-to-hand exchange process.

4) The frozen depth in the silty clay soil at Saskatoon is substantially greater than in the sandy soil at Outlook, despite the fact that snowcover conditions were similar and the accumulated degree-days of frost, calculated from soil temperatures at 200 mm, was the same at each location after 80 days of frost. The shallower depth of freezing at Outlook is attributed to the larger amount of water frozen.

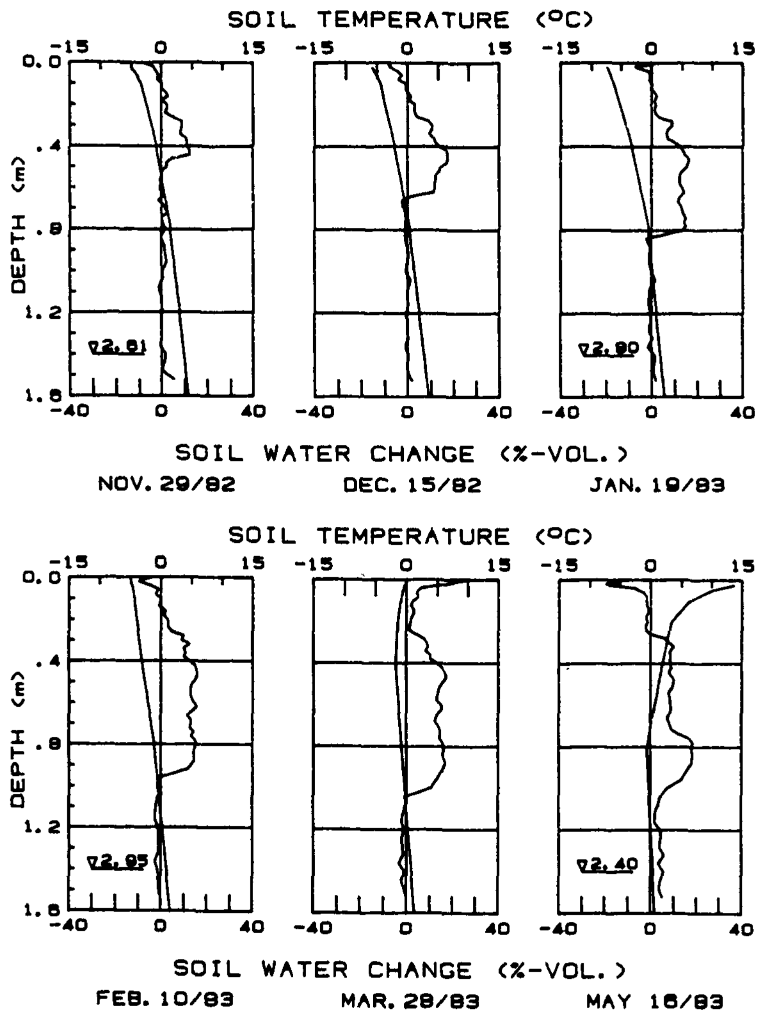

Figure 2. Profiles of soil temperature and changes in soil moisture measured in a fine sandy loam soil at Outlook. Sask. Moisture changes are for the period from November 17/82 to the date shown below each frame. ∇( ) refers to the depth to the water table in metres: depth on November 17/82 was 2.55 m.

These results demonstrate that significant fluxes of water may occur to the freezing front where conditions favor migration. They are in agreement with the findings reported by Ferguson et al. (1964). Willis et al. (1964) and Sheppard et al. (1981) which show larger amounts of moisture migration in a wet soil. and where a water table is near the soil surface. compared to the amounts transferred in dryer soils. Soil moisture observations in soils which were relatively dry at freeze-up showed moisture profiles with alternating layers of gains and losses throughout the frozen depth. indicating localized transfers. but with little change in total moisture content. When a soil layer is sufficiently dry that liquid water does not form continuous films throughout the matrix. it is not possible for water to move completely by capillarity in response to a temperature gradient. The moisture content at which capillary flow predominates depends on such factors as soil texture. porosity. density and layering. and is difficult to predict.

Excluding those sites where mid-winter melting had occurred. the overwinter moisture losses observed in the upper zone ranged to a maximum of 35 mm. Occurrence of mid-winter melting can compensate these losses and result in a net gain of moisture in the period. When all (~150) observed profiles are considered. the overwinter moisture changes in the upper zon- ranged from a loss of 35 mm to a gain of 15 mm. with fallow lands appearing to be the most prone to moisture losses. The fact that fields in this landuse may frequently be snow-free or that the snowcover is shallow results in large fluctuations in surface temperature, and hence larger temperature gradients. conditions favoring the transfer of vapor to the surface.

Overwinter soil moisture increases were observed in the lower zone (300 - 1000 mm depth) at 67% of the dryland sites (85% of fallow sites. and 58% of stubble sites). These moisture gains ranged from 1 to 50 mm. or 7% by volume. The differences in magnitude of the changes between landuse practices can be attributed to differences in snowcover and soil moisture and temperature regimes.

Changes to the moisture regimes of prairie soils during winter impact on the hydrology and agricultural practices of the region. For example:

1) Because the snowmelt infiltration potential of a frozen Prairie soil is inversely related to the soil moisture content of the upper zone at the time of melt (see following section). they directly affect the amount of soil water recharge and snowmelt runoff.

2) The large fluxes of moisture due to freezing in lighter-textured soils with a water table at shallow depth put in question the value of fall irrigation for increasing spring water reserves. The practice may enhance moisture migration. reduce snowmelt infiltration. contribute only small additions of water to the root zone over those that would occur naturally by migration due to freezing. increase runoff and erosion. and retard thawing in the spring.

3) Water that migrates in response to freezing does not drain until the soil profile has thawed in the spring. Therefore established crops. such as biennials or perennials. will benefit most by these additions. particularly in the early part of the growing season.

4) A moisture change calculated from measurements made in the fall and spring may not accurately reflect snowmelt infiltration.

5) They should be included in any model designed to forecast spring soil water reserves which is based on observations of fall soil moisture and snowcover water equivalent.

6) Liquid water which moves upward in a soil may carry dissolved salts. Gray and Granger (1986) showed increases in the amount of exchangeable ions in the 0 to 1-m layer at Outlook of 11.9 and 4.4 t/ha during the 1983-84 and 1984 85 winter seasons respectively. Because of the good drainability of the sandy loam most of these salts were removed by drainage following soil thaw. However it may be up to a month after a snowcover completely ablates before a soil thaws sufficiently to allow free drainage. During this period the surface layer is usually at a high level of saturation (due to snowmelt infiltration) and the hydraulic gradient is directed upwards (due to evaporation at the surface). conditions favor further upward movement of salts to the soil surface.

THE SNOWMELT PERIOD

When dealing with seasonally-frozen soils. the temperature regime (or depth of frost penetration) at the time of snowmelt will affect the infiltration properties of the soil. If the depth of frost is small the energy content of the infiltrating meltwater may be sufficient to raise the temperature of the entire frozen layer to 0°C. thus returning the infiltration characteristics of the soil to those of its unfrozen state. Komorov and Makarova (1973) suggest that a soil frozen to a depth of 150 mm or less absorbs meltwater as does an unfrozen soil. and once a soil is frozen to a depth of 600 mm. freezing to greater depth has no further effect on the infiltration of meltwater. Thus it is appropriate to define a "completely"- frozen soil as one that does not thaw during the active snowmelt period.

Infiltration into a frozen soil is g /erned primarily by the hydrophysical properties of the soil layer near the surface. Granger et al. (1984) showed a direct relationship between effective porosity and the minimum temperature against which infiltrating meltwater advanced. thus confirming the movement of water against low temperatures in soils with large effective porosities. as reported by Steenhuis et al. (1977). Both Komarov and Makarova (1973) and Steenhuis et al. (1977) suggest that in frozen soils the effects of soil temperature at the time of melt on infiltration may be secondary to those of effective porosity. which is inversely related to the ice content. They stress that the entry to and movement of water in frozen soils occurs mainly through large non-capillary pores.

Granger et al. (1984) and Gray et al. (1985b) outline the development and testing of a simple physically-based model describing infiltration into"completely"- frozen soils. They suggest that these soils may be grouped into three broad categories with regard to their infiltration potential:

Restricted - infiltration is impeded by an impermeable layer. such as an ice lens on the soil surface or within the soil near the surface. The amount of meltwater infiltration will be negligible and most of the snowcover water equivalent will go to direct runoff and evaporation.

Limited - infiltration is governed primarily by the snowcover water equivalent and the frozen water content of the soil layer. 0-300 mm, at the time of melt.

Unlimited - most of the snow water infiltrates. A soil in this condition contains a high percentage of large, air-filled. non-capillary pores at the time of melt. Coarse sands, gravels and heavy clay soils which crack naturally upon drying fall into this category.

Figure 3 is a plot of infiltration against snow water equivalent in which the three categories are demonstrated. The 1:1 line represents the "unlimited" situation where all the snow water infiltrates: the data points represent point measurements of infiltration into frozen. cracked soils. Infiltration at a point can exceed measured snow water because of overland flow to and interflow within the soil fractures. The data points on and near the abscissa. representing the "restricted" case, show that even with relatively large amounts of snow water. infiltration is negligible. The family of curves represent different initial, "premelt". moisture conditions for soils in the "limited" category. Granger et al. (1984) found for medium- to fine-textured uncracked, frozen soils in which meltwater is not impeded by ice layers that: (a) the average depth meltwater penetrated a soil was 260 mm (standard deviation = 100 mm). (b) infiltration was relatively independent of soil texture and landuse practice. and (c) the amount of snowmelt infiltration was inversely related to the average moisture content of the 0-300 mm soil layer at the time of melt. Gray et al. (1985b) provided the following relationship between snowmelt infiltration (INF), premelt moisture content (θ_p) and snowcover water equivalent (SWE):

$$INF = 5(1-\theta_p)SWE^{0.584} \qquad [1]$$

in which INF and SWE are in mm and θ_p is the degree of pore saturation, mm^3/mm^3. Equation 1. which was developed from measurements at 130 sites. has a correlation coefficient of 0.85 and a standard error of the estimate of 5.5 mm.

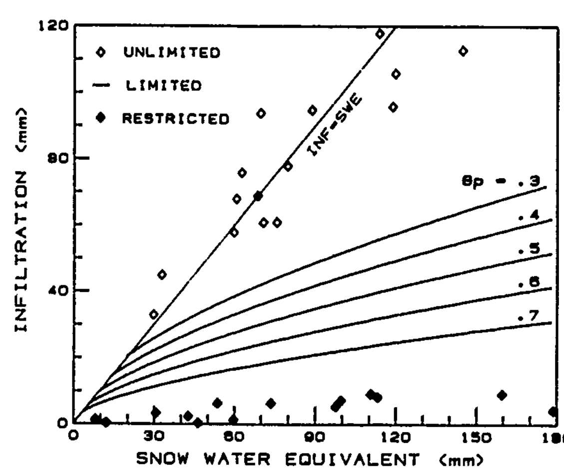

Figure 3. Infiltration plotted against snowcover water equivalent for frozen soils of the Unlimited. Limited and Restricted categories.

Referring to Fig. 3. it can be observed that the rate of increase of infiltration per unit increase in snowcover water equivalent decreases with increasing SWE. In light of this, and the interest in managing the snow water resource, it appears that applying snow trapping techniques on fields with soils in the 'limited" and "restricted" categories can result in situations where the practice is contributing more to runoff (and potential erosion) and evaporation than to recharge. It is evident that those soils which are likely to benefit most in terms of water augmentation from snowmelt, and hence are most amenable to the application of snow management practices, are those in the "unlimited" category. Under natural conditions the arable soils that most-commonly exhibit these properties are those heavy, lacustrine clays which severely crack on drying. The macropore content of soils which do not crack by natural causes can only be increased substantially by some practice such as subsoiling, which fractures soil to depths greater than the normal depth of penetration of infiltrating water. A companion paper in these proceedings (Gray et al. 1990) discusses infiltration into cracked and subsoiled frozen soils.

A number of models have been developed for simulating streamflow from snowmelt. A few examples are: the U.S. National Weather Service River Forecasting System, NWSRFS (Anderson, 1973): the Streamflow Simulation and Reservoir Regulation Model, SSARR (U.S. Army Corps of Engineers, 1972): and the HBV of the Swedish Meteorological and Hydrological Institute (Bergström, 1976). Each model differs from another, either as it calculates hydrological components or simulates the processes of snowcover accumulation and ablation, evaporation, infiltration, changes in soil moisture storage and flood routing, and no one model has been accepted for universal use. The performance of these models applied to natural catchments is directly related to the accuracy with which infiltration is evaluated. In most of the existing operational systems no attempt is made to distinguish differences in the infiltration process in frozen and unfrozen soils. In light of the above discussion this represents a major limitation of these systems for synthesizing snowmelt runoff on watersheds in many parts of central and northern Canada.

Gray et al. (1985b, 1986) demonstrated that incorporating the infiltration model (described above) into the NWSRFS and SSARR models significantly improved the performance of these systems in simulating streamflow from snowmelt on Prairie watersheds. Figure 4 shows observed and simulated hydrographs by NWSRFS for the Creighton Tributary (a small 11.4 km^2 watershed in western Saskatchewan) for 1974. The closer agreement between observed and simulated hydrographs due to the improved estimate of runoff volume is evident; the original model grossly underestimated the runoff volume because it assumes infiltration into an "unfrozen" soil. Tests with revised (containing the frozen soil algorithm) NWSRFS and SSARR models applied to a larger (125 km^2) watershed in south central Saskatchewan (Gray et al., 1986) also showed that simulations of runoff volume are markedly improved when infiltration into frozen soil is accounted for.

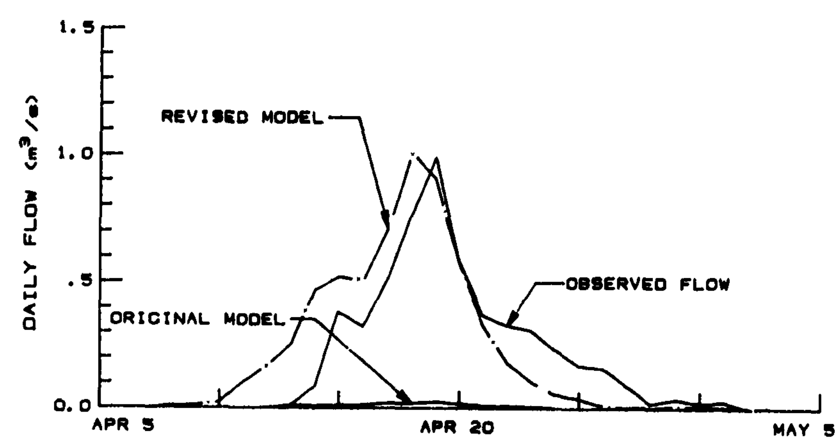

Figure 4. Observed and simulated hydrographs of snowmelt runoff from Creighton Tributary in 1974. Original model:NWSRFS operated with its land phase subroutine unchanged: revised model:NWSRFS operated with the land phase subroutine replaced with the infiltration model.

THE POSTMELT PERIOD

In this paper postmelt is taken as the period extending from the disappearance of the seasonal snowcover up to the time of seeding of annual crops (usually six-nine weeks). During the first two-five weeks of postmelt (depending on the frozen moisture content and the energy supplied) the soil remains partially frozen. The presence of frozen soil affects the disposition of infiltrated meltwater and the moisture which has accumulated in the lower zone. The last clichés of Figs. 1 and 2 show typical changes in the soil moisture regime following melt: a loss of moisture from the upper zone due to evaporation, and drainage of water below the frost line. Note that even in the light-textured soil there is no pronounced change in the moisture profile due to drainage until the soil temperature throughout the frozen zone reaches 0°C. Deep percolation below a root zone of 1 m does not always take place (it was observed in only 23% of the sites): it occurs when the moisture content exceeds the water holding capacity of the soil, and therefore is affected by the drainability of the soil (75% of the sites in fine sandy loam and 15% of those in clay showed deep percolation).

By preventing the vertical movement of the infiltrated meltwater, a frozen soil increases the probability that this water will be lost to evaporation. Measurements of soil moisture changes showed that at all sites (100%) evaporative losses occurred from the upper zone, whereas at only 53% of the sites was the rate of soil thaw sufficiently rapid to allow some of the meltwater to move deeper into the soil profile before it was evaporated. Table 1 shows that evaporative losses during postmelt are significant when compared to gains by meltwater infiltration and precipitation. The data also show that, on average, meltwater infiltration is lost to evaporation in 7 days following the disappearance of the annual snowcover from fallow and in 16 days from stubble.

Table 1. Average snowmelt infiltration and evaporation losses in the postmelt period.

Landuse	No. Obs.	Infiltration mm	Evaporation+	Ratio EVAP/INF	Ratio EVAP/(INF+PPT)	Rate++
Fallow	39	19.9	50.9	2.6	0.9	7
Stubble	77	26.6	46.5	1.7	0.7	16

+ Obtained from the soil water balance: EVAP = ΔMoisture - PPT.

++ The average number of days following the disappearance of seasonal snowcover in which meltwater infiltration is lost to evaporation.

The existence of frozen soil affects the energy available for evaporation. While there is frost in the ground, a greater amount of the incoming energy (solar radiation) is apportioned to the soil heat flux in order to supply the energy required for the solid-to-liquid phase change, thus reducing the amount of energy available for the liquid-to-vapor phase change (evaporation) at the surface. Many schemes for estimating evaporation, although they are based on energy balance considerations, ignore the soil heat flux, and tend to overestimate the evaporation rates during the period of soil thaw. A comparison of the measured evaporation amounts to those estimated by Morton's (1983) evapotranspiration model showed that, whereas the model performed well during the summer months it gave estimates that were generally greater by 22 to 30 mm during the 4-week period following snowcover ablation: an error of approximately 50% (Granger and Gray, 1989).

SUMMARY

This paper uses in-situ measurements of soil moisture changes during winter and spring to demonstrate the impact of frozen soil on the hydrology of the Canadian Prairie region. The discussion focuses on three periods: the winter period (from freeze-up to the onset of snowmelt), the snowmelt period, and the postmelt or thawing period (from the disappearance of the seasonal snowcover up to the seeding of annual cereal grains).

During winter, moisture is usually lost from the upper soil zone (0 to 300-400 mm soil layer); conversely, large fluxes of water may migrate into the adjacent lower frozen zone (which extends to the depth of the frost front) in response to the freezing action. These moisture changes affect the infiltration capacity of a soil and the amount of water contained within the root-zone of a crop. It is also shown that under specific conditions the migrating water may contain soluble salts in amounts sufficient to change the salinity level of the frozen profile. The magnitudes of the observed moisture changes indicate that snowmelt infiltration cannot be reliably estimated from the difference between fall and spring soil moisture measurements and demonstrate the importance of accounting for the migration process in modelling soil moisture reserves.

Snowmelt infiltration into a completely-frozen soil (a soil which does not thaw during the snowmelt period) is governed largely by the frozen water content of the upper soil zone and by the presence of large non-capillary macropores. A simple conceptual model is presented which classifies frozen soils into three categories according to their infiltration potential. The impact of the presence of frozen soil on the use of snow management practices aimed at increasing soil moisture reserves for crop production and on the simulation of streamflow from snowmelt is demonstrated.

The continued presence of frozen soil during the postmelt period affects the disposition of infiltrated meltwater. Since the downward movement of the water is prevented until the soil profile has completely thawed. the water is held near the surface where it is subject to evaporation. Evaporation losses often exceed the gains from snowmelt infiltration. It is during postmelt that conditions are favorable for further migration towards the soil surface of salts which have moved in response to freezing. The frozen soil also represents an effective heat sink. and thus affects the surface energy balance. Evaporation models which assume that soil heat flux is negligible may be subject to significant errors during the period of soil thawing.

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COMPARISON OF THREE METHODS FOR MEASURING DEPTH OF SOIL FREEZING

by

Clayton L. Hanson and Gerald N. Flerchinger[1]

INTRODUCTION

Flooding often occurs during the winter in the northwestern United States (Johnson and McArthur, 1973). These floods generally occur from December through March when warm, moist air masses, accompanied by strong winds, produce rain and/or rapid snowmelt after periods of extreme cold that freezes the soil to considerable depths.

The Reynolds Creek Experimental Watershed located in southwestern Idaho has been operated by the Agricultural Research Service, USDA, since 1960 for the purpose of studying water yield, flood flows, and sedimentation problems from grazing areas of the Pacific Northwest (Robins et al., 1965). Several methods have been used on the watershed to measure the spatial and temporal distribution of frozen soil to enhance our understanding of relationships between frozen soil and the associated runoff and erosion (Hanson et al., 1988). Our experiences with the thermilinear network for measuring soil temperature and gypsum soil-water resistance blocks and a version of the Gandahl gage for measuring depth of frozen soil are reviewed in this paper. Data from the Quonset site (Fig. 1) are used to show how well the depth of frozen soil compares between the three systems.

EXPERIMENTAL SITES AND INSTRUMENTATION

Soil temperature

The 12 locations on the Reynolds Creek Experimental Watershed where depth of frozen soil was measured are shown in Figure 1. Soil temperatures were measured at five of the sites by the thermilinear network (Yellow Springs Instrument Co., Yellow Springs, OH*), which is a resistance network consisting of a thermistor composite and two precision resistors. Soil temperatures have been recorded hourly at the Quonset site since 1985 at depths of 2.5, 5, 10, 15, 20, 30, 40, 60, 70, and 90 cm. Soil temperatures are recorded hourly at two sites and weekly at the other two sites. Measurements are made at depths of 10, 30, 60, 90, 120, 180, and 240 cm at these four sites.

Thermilinear networks were used rather than thermocouples because they can be easily read by a hand-held meter or continuously recorded at remote sites. The only problem we encountered was a breakdown of the material used to pot the sensors. This problem has been solved by protecting the sensors in vinyl.

Gypsum blocks

Gypsum soil-water-resistance blocks (gypsum blocks) have been used in several studies to indicate the presence of frozen soil (Burgess and Hanson, 1979; Sartz, 1967; Wilen et al., 1972). Studies by Colman and Hendrix (1949) and Bethlahmy (1953) showed that there was a change in electrical resistance in blocks due to both soil-water content and temperature. They found that the electrical resistance increased as the soil temperature decreased and that there was a sharp rise in the electrical resistance when blocks froze.

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*Trade names and company names when mentioned are for the convenience of the reader and do not imply preferential endorsement by the USDA-Agricultural Research Service of a particular product or company over others.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

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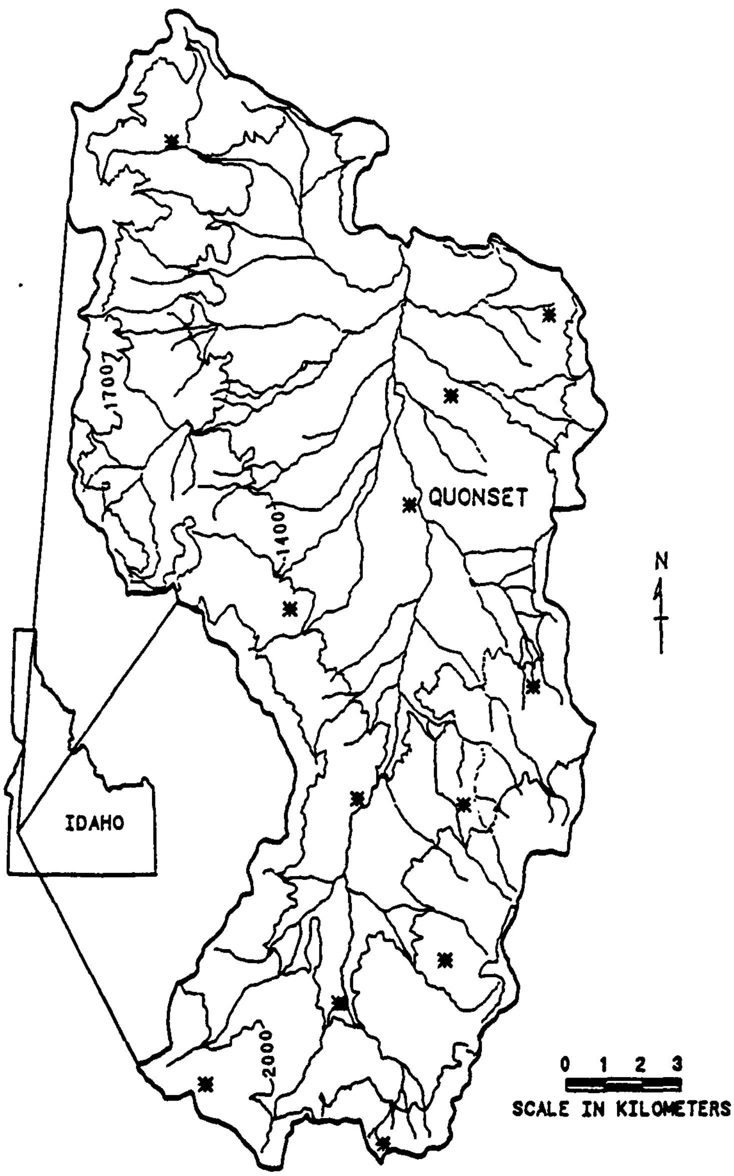

Figure 1. Reynolds Creek Experimental Watershed depth of frozen soil instrumentation network.

Gypsum blocks were installed at 12 sites on the Reynolds Creek Experimental Watershed during the mid 1970's as one of the methods used for monitoring the spatial and temporal distribution of frozen soil (Fig. 1). Two sets of single gypsum blocks were installed at a level site at each location. The gypsum blocks were positioned at depths of 5, 10, 15, 20, and 30 cm. In addition, one set of gypsum blocks at the Quonset site was positioned at 5, 10, 15, 20, 30, 40, 50, and 60 cm depth. Two sets of gypsum blocks were installed at each site, except the Quonset site, to prevent data loss due to gypsum block failure.

During the cold period, frost depth was measured weekly at all sites except the Quonset site where measurements were taken more frequently. When the soil was frozen, frost penetration was assumed to be between two adjacent depths, i.e., frozen at 5 cm but not at 10 cm. When soil frost was indicated at 30 cm, the depth was noted as 30 cm or greater.

The gypsum blocks on the watershed required changing every four years because they dissolved enough during that time period to cause failure. The major problem encountered in using gypsum blocks to measure soil frost on the watershed was the lack of fall precipitation at all of the sites. Because of dry conditions during some years, there was no way to determine if the soil was dry, frozen, or both since gypsum blocks also indicate very high resistance at low soil water contents.

Frost tubes

Frost tubes similar to the modified Gandahl gage (Gandahl, 1957; Rickard and Brown, 1972; Harris, 1970) have been used on the Reynolds Creek Experimental Watershed. The tubes were read manually on a daily basis at the Quonset site and the other tubes were read weekly.

Frost tubes consisted of an outer rigid PVC pipe, 24 mm inside diameter, and an inner plexiglas tube 19 mm inside diameter by 22 mm outside diameter. The outer tubes were installed to a depth which allowed for the inner tube to be 100 cm in the soil. The outer tube extended above the ground surface so they could be found during the winter. The inner tube had a 4.8 mm flexible plastic tube mounted in the center to prevent breaking the inner tube when the solution in the inner tube froze. This happened at a site where 50 to 100 cm of soil frost occurred during some winters without the flexible plastic tube. The flexible plastic tube was held in place by the material used to plug the ends of the inner tube. The inner tube was filled with sterilized, 1 mm blasting sand, and a 0.1 percent fluorescein dye solution. When the fluorescein dye solution froze, it changed from green to a red-brown color. In order to read the depth of freezing, an adhesive, metric tape was placed on the outside of the inner tube.

We currently use sterilized sand to prevent organic growth in the inner tube as this was a problem with the first frost tubes we constructed. As mentioned earlier, we now put a flexible plastic tube in the center of the inner tube to prevent breaking the bottom out of the inner tube.

Annual maintenance includes adding a small amount of sand to almost all frost tubes because of settling. This is accomplished by feeding sand into the top of each inner tube through a small hole that is drilled and threaded for a stopper.

FIELD EVALUATION OF METHODS FOR MEASURING DEPTH OF FROZEN SOIL

Depth of frozen soil measured by the three measuring systems is shown in Figure 2 for the period November 27, 1988 through March 7, 1989. The 0 °C isotherm shown in Figure 2 was obtained from linear interpolation between the depth at a temperature sensor below 0 °C and the next sensor that had a temperature above 0 °C. The gypsum block depths have not been interpolated and thus the actual frost depth was at or below the plotted depths.

During the last three days of November and the first 15 days of December, the gypsum blocks and frost tubes indicated the same depth of frozen soil. The 0 °C isotherm was 5 to 10 cm below the frost line indicated by the gypsum blocks or frost tube but the soil temperatures were all warmer than -0.3 °C. Between December 16 and January 8, the frost depth measured by the frost tube and the 0 °C isotherm were never more than 5 cm different. During this same period, the frost depth indicated by the gypsum blocks lagged the others; however, the frost depth was greater because only the depth of the deepest frozen gypsum block was plotted.

There was a disparity between the gage readings during the period of January 9 through 24. When the soil was at or near freezing after being frozen, the gypsum blocks indicated little or no frozen soil, whereas the frost tube and soil temperature indicated freezing to about 20 cm. However, the soil temperature was generally warmer than a -1 °C below 5 cm for six of these days and field observations indicated that the color in the tube made it difficult to determine if the tube was frozen or not. These remarks about the difficulty of determining color differences in the tube agreed with Harris (1970) who said that "during periods of rapid freezing and thawing the color separation lines in the gages were sometimes diffused and mixed." His gage was the same type as the one used in our study.

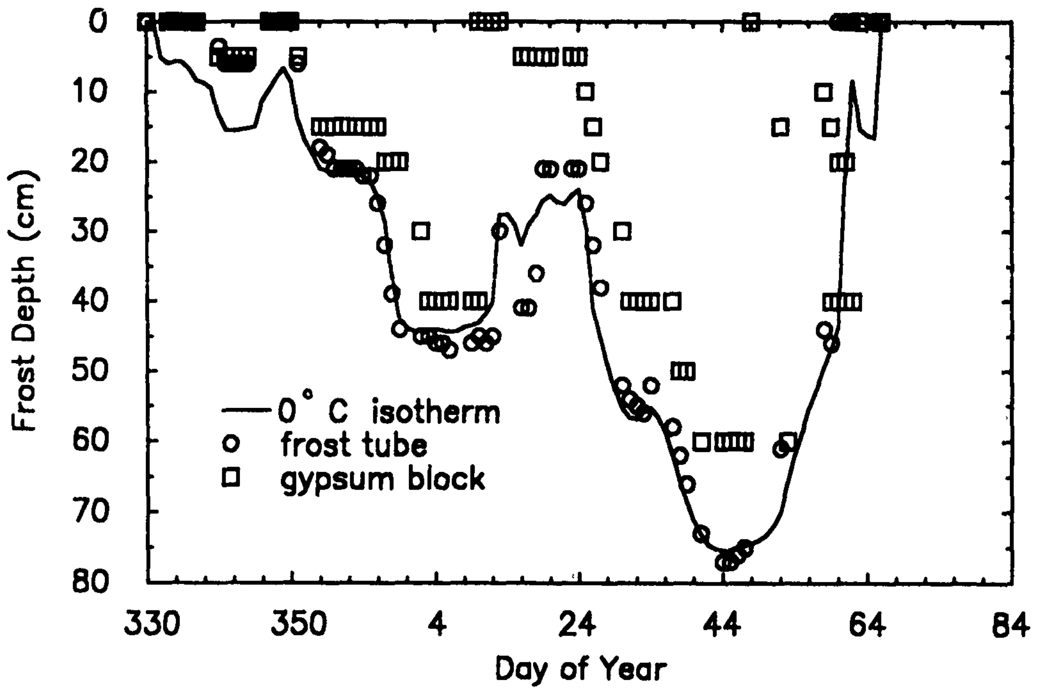

Figure 2. Depth of frozen soil at the Quonset site for the period November 7, 1988 through March 7, 1989.

When the weather cooled at the end of January, all gages indicated that the soil froze rather rapidly. The frost tube and 0 °C isotherm agreed very closely to a maximum freezing depth of about 75 cm. The gypsum blocks lagged the other gages, but only the bottom frozen gypsum block was plotted. During the period of deepest frost penetration, the deepest gypsum block, which was located at 60 cm, froze which suggests that the soil was frozen below 60 cm.

The soil profile started thawing rapidly on February 21 and thawed from the bottom upward from about 70 cm to 40 cm by February 25. From February 27 through March 2, the gypsum blocks indicated that the soil had thawed from the surface and from below leaving a frozen layer from about 15 to 40 cm. The tube indicated frozen soil at about 45 cm on February 28 and all thawed the next day. During this period of rapid thaw, field observations indicated that the colors in the frost tube were not distinct, making it difficult to determine if the frost tube was frozen or not. All of the sensors indicated that the soil profile was not frozen on March 7.

Soil frost monitoring systems do not indicate the type of frost in the soil, i.e., the two extreme types, concrete and granular. Bethlahmy (1953) could not determine soil frost type from laboratory experiments using fiberglass blocks; However, these studies do suggest that the gypsum blocks can be used to infer soil frost type because gypsum blocks indicate soil water status and soil frost types as related to the amount of soil water. Knowing if the soils are wet or dry prior to soil freezing could indicate frost type, which is useful in flood prediction.

SUMMARY

Three measuring systems have been used to measure the depth of frozen soil on the Reynolds Creek Experimental Watershed in southwest Idaho. The systems include a thermilinear network to measure soil temperature, gypsum soil-water-resistance blocks (gypsum blocks), and the modified Gandahl tube (frost tube). The frost tube readings were very well correlated with the 0 °C isotherm for most of the 1988-1989 winter season.

Generally, the frost tube readings were within 5 cm depth of the 0 °C isotherm depth. There were greater differences during a thawing period in mid January and at the end of the season. During these thawing periods, field observers were unable to determine the frost line because the color in the frost tube was not distinct. These results agree with findings by Rickard and Brown (1972) and Harris (1970).

In general, the frost tube has given reliable reconnaissance type data. We have not encountered the problem of the frost tube thawing several days later than the soil. This condition was reported by McCool and Molnau (1984) who did not use sand in the inner tube.

Gypsum block readings followed the same pattern of soil freezing and thawing as indicated by the other two methods. However, they tended to lag when the soil was freezing and indicate that the soil was thawed before the other methods. The gypsum blocks indicated essentially the same maximum depth of frozen soil when all three sensor types indicated frozen soil conditions. Results from this study show that the gypsum blocks are a reliable method of measuring depth of soil freezing and substantiate previous researchers' findings (Burgess and Hanson, 1979; Sartz, 1967; Wilen et al., 1972). Frost tubes are advantageous because they measure all locations from the surface, whereas the resolution provided by the gypsum blocks and thermilinear network are determined by the location of the sensors.

The thermilinear network systems worked well after a method was developed to protect the sensors from water. The gypsum blocks must be replaced approximately every four years because they eventually dissolve. They also do not indicate soil frost conditions when they go into winter dry due to a lack of fall precipitation. The frost tubes need some annual maintenance because of breaks, etc., but this has not required a great amount of money or time.

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REDISTRIBUTION OF SOIL WATER AND SOLUTES IN FINE AND COARSE TEXTURED SOILS AFTER FREEZING

by

L. L. Hofmann, R. E. Knighton, and J. L. Richardson[1]

INTRODUCTION

Winters in the northern Great Plains are characterized by long periods of below freezing temperatures with few if any periods of thaw occurring for two to three months. This is often accompanied by little to no snow cover resulting in soils freezing to depths greater then 1m. Studies indicate this freezing process results in significant increases of water and solute concentrations in the upper portion of the soil profile (Campbell et al., 1970; Gray et al., 1985; Gray and Granger, 1985). The quantity of water translocated towards the surface is affected by soil moisture content before freeze-up, depth of freezing front, and the presence of a shallow water table (Willis et al., 1963; Benz et al., 1968; Benoit et al., 1988).

In saturated soils, water freezing in the pure state concentrates solutes just below the freezing front as it advances into the soil profile (Arndt and Richardson, 1986). In unsaturated soils, upward migration of water and solutes through the unfrozen soil in combination with concentrated solution moving down with the freezing front concentrates substantial quantities of salts near depths of maximum frost penetration (Beke and Palmer, 1989). Movement of soil water is also evident within frozen regions advancing from warmer to cooler areas believed to occur primarily in the vapor phase (Gray et al., 1985; Gray and Granger, 1985). Studies indicate that films of liquid water surrounding soil particles in frozen soil can move along temperature gradients with thickness and rate of movement decreasing as temperatures drop below 0 °C (Hoekstra, 1966). Migration of solutes and particles can also occur by diffusion through these liquid films (Cary and Mayland, 1972; Römkens and Miller, 1973). These phenomenon may result in increases of water and solutes in frozen areas of the soil profile.

This study is designed to quantify the changes in moisture and solute content as the freezing front advances into the soil profile in a sulfatic system containing similar concentrations of Na and Mg salts. The effects of a water table, solute concentration, depth of freezing front, and soil texture on water and solute redistribution within a soil as it freezes are presented.

PROCEDURES

Soil columns 90cm in length of Fargo silty clay (Vertic Haplaquoll) and Hecla fine sandy loam (Aquic Haploboroll) were sampled in 20cm diameter PVC pipe by driving the pipe with a metal cutting bit into the ground with a tractor-mounted post pounder. A pit was dug and the columns were pulled from the pit wall with a PVC base cemented to the bottoms to obtain undisturbed samples. The columns were saturated under a vacuum, then allowed to drain using two solutions: (1) a moderately saline solution of 1-gram equivalent $MgSO_4$ and Na_2SO_4 with 100 mg kg^{-1} KBr, and (2) a control solution of 0.01N $BaCl_2$ plus 5% ethyl alcohol by volume. Thermocouples were placed at 7.5cm depth intervals in the upper 45cm and 15cm intervals in the lower 45cm. The columns were brought to an ambient air temperature of 6 °C and placed in baths maintaining a 2cm hydrostatic head over the column bases. The upper 45cm of the soil columns were positioned within a modified chest freezer surrounded with 7.6cm fiberglass insulation and enclosed in 2.5cm high density foam insulation with only the surface of the columns exposed to freezing temperatures. The columns were then frozen from the top down at -10.0 °C . Soil and air temperatures were logged every 12h with a Wescor TC Thermometer model TH-65. A freezing run consisted of one of three soil columns of a given solution treatment frozen to a depth of 15cm, 30cm, and 45cm. The columns were removed from the freezer when the temperature at the desired depth of freezing reached -0.2 °C and sampled at the depths of thermocouple placement. All samples were analyzed for gravimetric water content, bulk density, EC, pH, alkalinity, soluble salts (Rhoades, 1982), and Br^- (Adriano and Doner, 1982) with sulfate determined by difference based on saturation paste extraction.

Presented at the Frozen Soil Symposium, March 21-22, 1990; Spokane Washington.

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RESULTS

Freezing the soils from the surface down was achieved with mean temperature gradients across the length of the columns at sampling ranging from -7.2 to 4.8°C in the Fargo silty clay and -5.3 to 6.1°C for the Hecla fine sandy loam. The average temperature profiles of the two soils for each freezing treatment are illustrated in Figure 1. Average time to complete each freezing treatment is presented in Table 1. The saline solution (SS) treated columns required 25 to 64% more time then the distilled solution (DS) to freeze to the specified treatment depth.

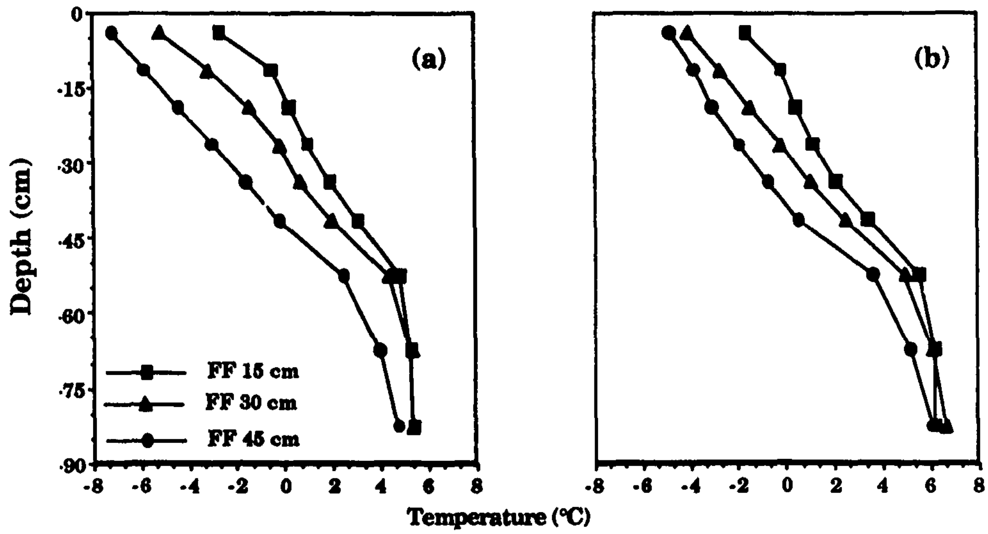

Figure 1. Average temperature profiles of the Fargo (a) and the Hecla (b) soils at three depths of the freezing front (FF).

Table 1. Time of freezing required to obtain depth of freezing treatments in Fargo silty clay and Hecla fine sandy loam soils.

Soil Type	Solution Treatment	Freezing Front	Freezing Time
		- cm -	- d -
Fargo	DS	15	1.8
		30	2.8
		45	19.4
	SS	15	2.3
		30	4.6
		45	24.0
Hecla	DS	15	1.2
		30	6.2
		45	31.0
	SS	15	1.9
		30	10.1
		45	31.0

Figures 2 and 3 illustrate water and EC redistribution as the freezing front advances in the Fargo silty clay soils and Figures 4 and 5 for the Hecla fine sandy loam. Data values are averages of three replications. Exceptions are two replications for the Fargo soil water content values and the 45cm freezing treatment of the Hecla soil. Values for the Hecla soils represent only one replication of the experiment. The control treatment represents average moisture and EC values of columns treated with the corresponding solution and sampled after draining for 14 days. Depletion of H_2O in the unfrozen portion of the columns corresponds to increases near the freezing front in the soils frozen to 30 and 45cm. Average initial Fargo soil water content for the surface depth increment is 0.33 $cm^3 \cdot cm^{-3}$ with losses ranging from 0.01 to 0.04 $cm^3 \cdot cm^{-3}$ as the freezing front advanced from 15cm to 45cm in depth. Average Hecla initial water content for the surface depth increment is 0.21 $cm^3 \cdot cm^{-3}$ with a range of +0.02 to -0.04 $cm^3 \cdot cm^{-3}$ after freezing. The losses of water occurring near the surface are due to sublimation.

Changes in EC values appear to be different between soil types and solution treatments. Average EC values for the DS and the SS solutions are 0.99 and 6.29 $dS \cdot m^{-1}$, respectively. Fargo soils exhibit EC changes inversely proportional to changes in water content, but a direct relationship between water and EC redistribution is evident in the Hecla soils. This may be explained by hydraulic and vapor conductivity properties of the soils. The low hydraulic conductivity of the Fargo silty clay inhibits mass flow of water to a rate below that of sublimated losses. Increases in moisture in the frozen Fargo soil are due predominantly to transport in the vapor phase. The greater hydraulic conductivity in the coarser Hecla fine sandy loam allows enough water transport during freezing to recover sublimated water losses (Koopmans and Miller, 1966).

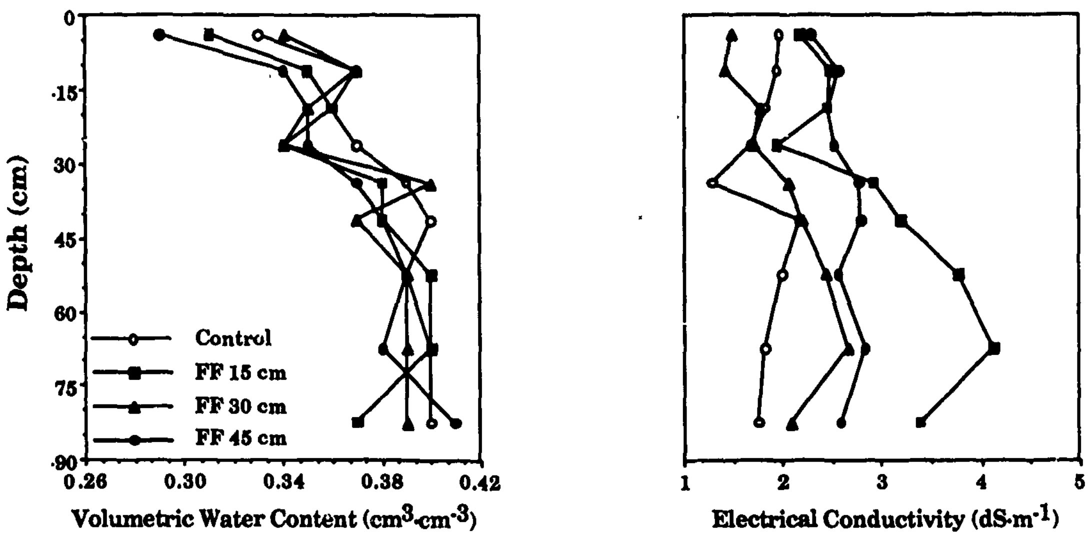

Figure 2. Average volumetric water and electrical conductivity redistribution in DS treated Fargo soil as the freezing front advances.

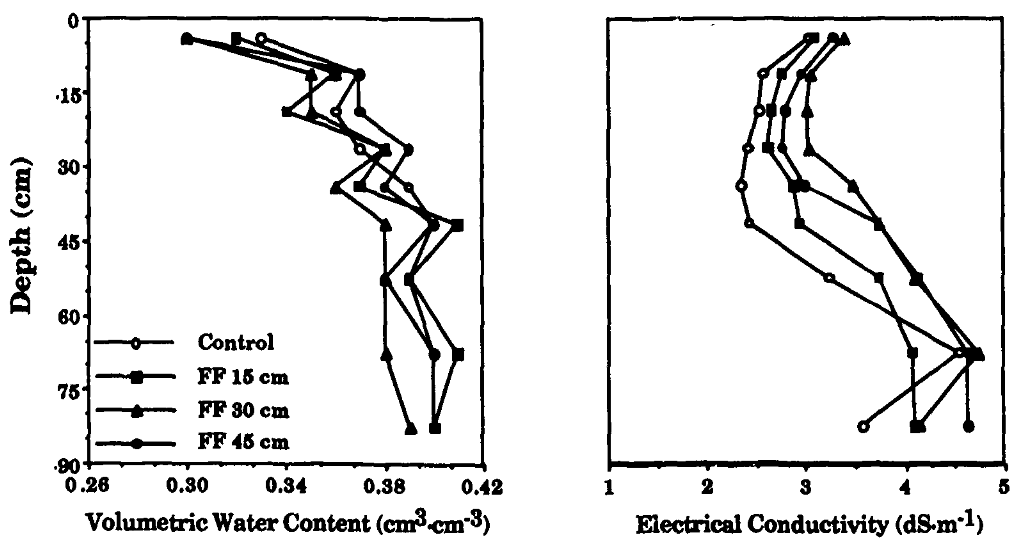

Figure 3. Average volumetric water and electrical conductivity redistribution in SS treated Fargo soil as the freezing front advances.

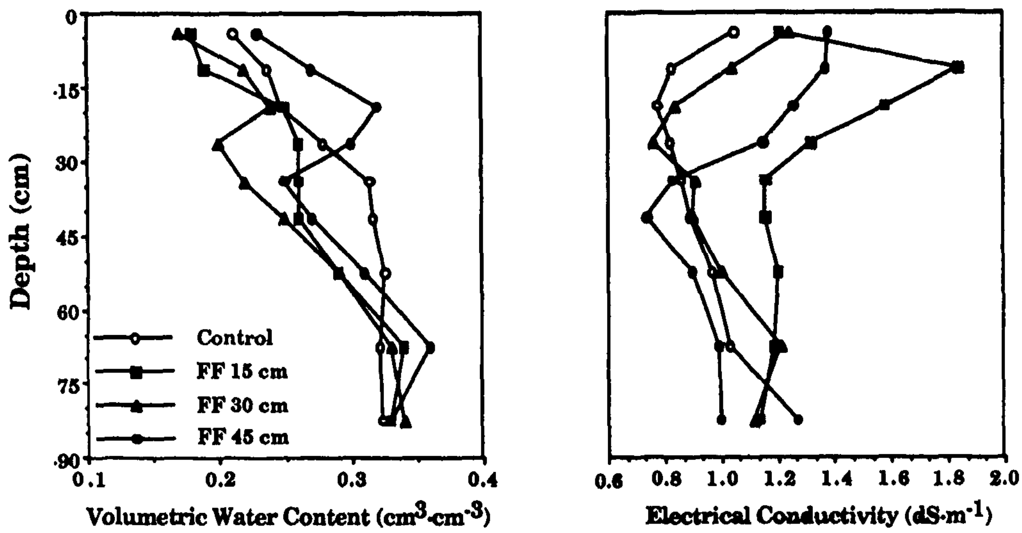

Figure 4. Average volumetric water and electrical conductivity redistribution in DS treated Hecla soil as the freezing front advances.

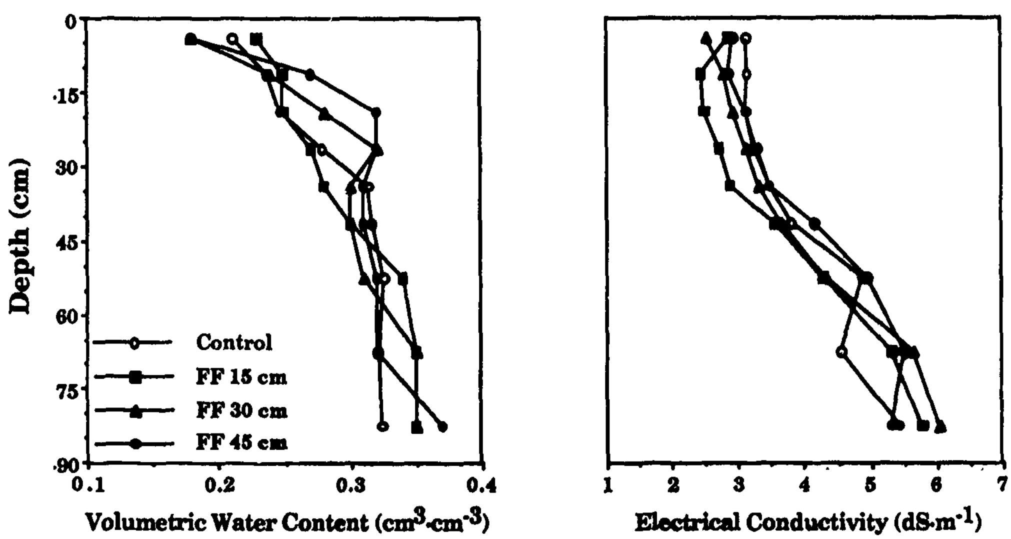

Figure 5. Average volumetric water and electrical conductivity redistribution in SS treated Hecla soil as the freezing front advances.

Trends for DS treated soils indicate an inverse relationship between volumetric water changes and EC changes. This is an indication that water low in solutes is moving up from the watertable with salts dissolving into solution in the unfrozen portion being transported towards the freezing front. Trends of a direct relationship are evident in the SS treated soils but to a lesser extent in the Fargo soils than the Hecla soils. Complex exchange reactions in the Fargo soil due to high exchange capacity could be responsible for attenuating these trends.

Movement of solutes towards the surface in the frozen soil is evident in the 45cm freezing treatment. This increase may be explained by upward thermal transport of thin liquid films surrounding ice and soil particles. This is probable considering it took from 2-4 weeks more freezing time between the 45cm freezing treatment versus the others (Cary and Mayland, 1972; Römkens and Miller, 1973). Redistribution of cations as the freezing front advances into the soil is illustrated in Figures 6 and 7.

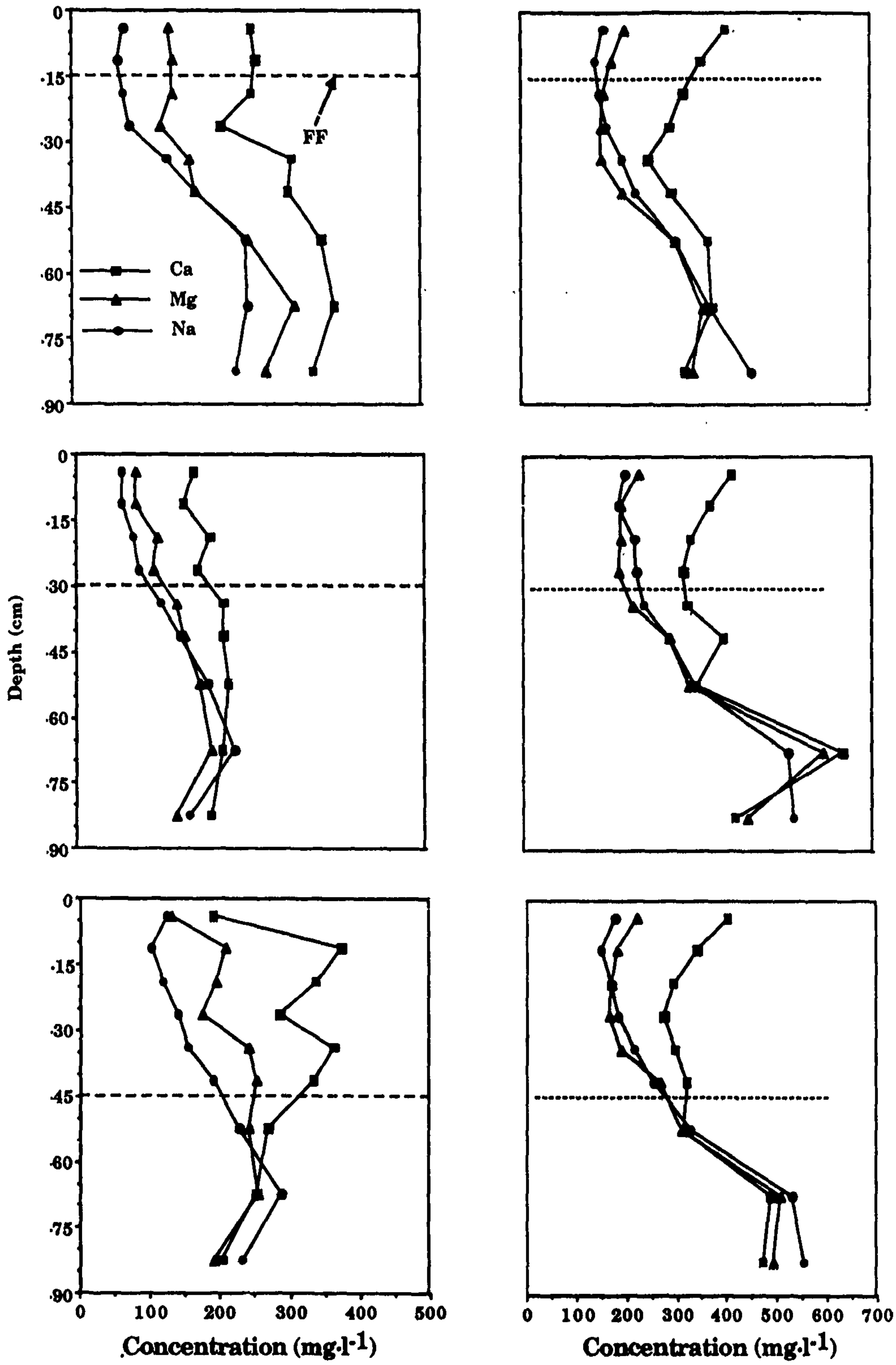

Figure 6. Redistribution of average cation concentrations in the Fargo soil treated with DS (left) and SS (right).

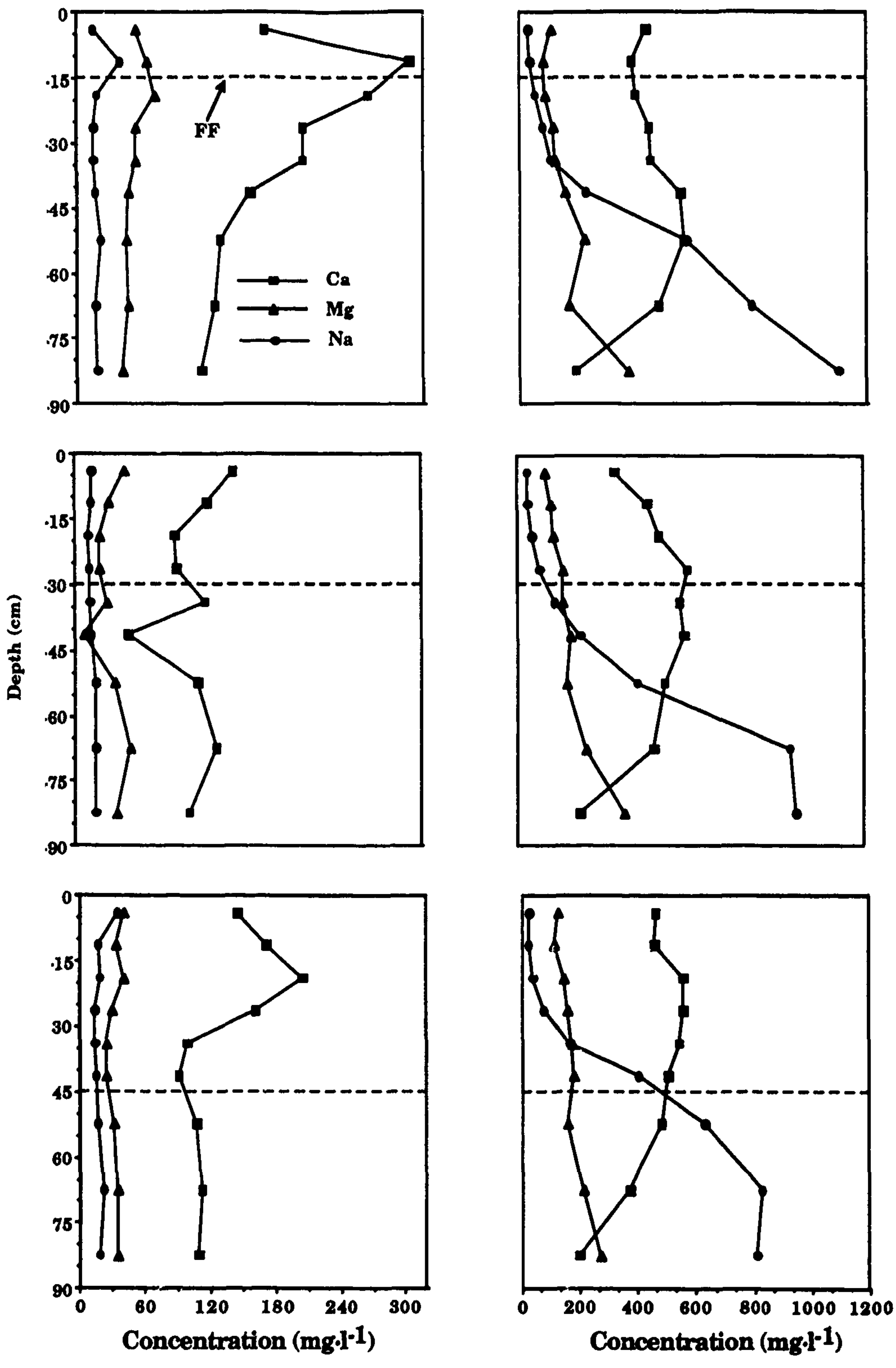

Figure 7. Redistribution of cation concentrations in the Hecla soil treated with DS (left) and SS (right).

The dominant anion in the SS treated soils is SO_4^{-2} with HCO_3^- influencing the chemistry to a large extent in the DS treated soils. Preliminary results indicate movement of salts through the unfrozen portion, concentrating salts near the freezing front. Few trends in ionic redistribution are evident in the DS treated Hecla and SS treated Fargo soils. However, trends in cation redistribution are evident in the DS treated Fargo and the SS treated Hecla soils. The increase in Ca concentrations may be due to exchange reactions and ion pairing as Mg is drawn up from the watertable and unfrozen soil. Correlations conducted between EC values resulted in correlation coefficients between EC and Mg at the three freezing treatments ranging from 0.83 to 0.95 and 0.37 to 0.99 for EC versus Na. This is an indication that the increases in EC are mainly due to movement of Mg and Na. Na appears to concentrate the greatest below the freezing front while Mg is exchanging with Ca associated with the exchange complex. The use of undisturbed soil columns and limited constraints on the number of treatment replications led to the inability to detect any major significant differences between treatments. Trends indicate that appreciable redistribution of soluble and exchangeable ionic species within these soils can occur as water and solutes are drawn from a watertable as these predominantly sulfatic systems freeze. Bromide values at similar treatment levels may enable better interpretation of the mechanisms influencing redistribution of water and solutes.

CONCLUSIONS

Water and solutes move towards a freezing front causing increases in water and solute concentrations within the soil profile drawing from the water table as the freezing front advances. The relationship between water and solute movement is influenced by soil textural and solute concentration differences with movement more restricted in the silty clay soil. Appreciable differences in soluble and exchangeable salts occurs in these systems as water and solutes are drawn from the watertable through the unfrozen soil. Mg and Na appear to be the most mobile cations with Na concentrating below the freezing front while Mg exchanges with Ca. Solute increases in the frozen soil induced by thermal gradients are evident given an adequate time of below freezing temperatures. The use of undisturbed soil columns resulted in the inability to detect significant differences between solution and freezing depth treatments influenced by physical and chemical spatial variability common under naturally occuring field conditions.

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FIELD AND LABORATORY TECHNIQUES FOR FROZEN AND THAWING SOIL

by

D. K. McCool and Hans Kok[1]

INTRODUCTION

The northern tier of states of the US and the Canadian provinces experience frozen soil of varying depth and for periods of varying length every winter. The regions roads and other infrastructure are affected by the occurrence of the frost and its hydrology is influenced by rain or snowmelt on frozen soil which can cause severe floods. In some areas, such as the Palouse of the Pacific Northwest, frozen and thawing soil is a factor in soil erosion because of reduced strength and high erodibility as the soil thaws (Zuzel et al., 1982; Formanek et al., 1984; Van Klaveren, 1987; and Kok, 1989). Research on frozen and thawing soil is necessary to protect infrastructure, evaluate flood hazard to protect lives and property, and to understand and prevent erosion from frozen soil runoff events.

The purpose of this paper is to briefly present and discuss methodology and instrumentation for conducting frozen soil research. References will provide additional information.

FROST DEPTH

The classic method of determining the depth of frost is to dig a pit and observe the occurrence of ice crystals. This requires a great deal of time and effort and destroys the experimental area. However, it is preferred by some investigators, because of the information that can be obtained about frost form. For many situations is necessary that the area not be disturbed or that readings be done more quickly. Thus, indirect methods have been devised to infer the location of the frost line. Garstka (1944), Harrold and Roberts (1960), and Burgess and Hanson (1979) used soil moisture blocks to detect the frost line. Electric conductivity approaches zero when moisture in the blocks freezes. Soil moisture blocks are an accurate means of determining frost depth but provide only point measurements. However, the blocks and instrumentation may be installed at remote sites and visited infrequently. Sartz (1967) reported the results of his test of three indirect indicators of frost depth. These were moisture blocks, thermistors, and a frost meter, a tube containing a series of solution-filled glass bottles; the lowest broken bottle indicated frost depth. He concluded that the moisture blocks were more accurate than thermistors when compared to a penetrometer test.

The low cost, ease of fabrication, and ease of use by untrained personnel have led to numerous experiments to develop and improve tube-type frost gages. Harris (1970) reported on tests of two water-dye filled tubes. These were a gage using a 0.01 percent solution of green fluorescein dye with medium-size clean sand in a polyethylene tube, similar to a gage developed by Gandahl (1957), but as modified at the Cold Region Research and Engineering Laboratory (CRREL), and a similar tube using a Kool-aid* solution. He concluded the modified Gandahl gage performed better during the critical periods of fast freezing and thawing.

Paper presented at the Frozen Soil Symposium, Spokane, Washington, March 21-22, 1990.

* Mentioning of trade or brand names does not constitute endorsement for use by the authors and is provided for the readers' information only.

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Rickard and Brown (1972) compared the performance of two frost tubes. The first was a modified CRREL tube, as reported by Harris (1970), constructed of a 19.0 mm O.D. innertube of polyethylene filled with sandblasting sand and a 0.1 percent fluorescein dye solution. The second tube was filled with a 0.1 percent methylene blue solution (no sand). They found that when the frost was penetrating, the modified CRREL tube responded to frost penetration faster, with the indicated frost depth within 20 mm of the $0^{o}C$ isotherm. The methylene blue tube consistently lagged the modified CRREL tube by as much as 50 mm. During thawing periods, they found the methylene blue tube also lagged. They concluded that the modified CRREL sand-filled tube with green fluorescein was superior to the methylene blue tube because the added water mass of the methylene blue tube slowed the response time and the greater volume expansion of the liquid filled tubes resulted in physical damage. Also, the blue solution tended to give unclear edges.

In 1976, Ricard et al. reported on a field assembled frost gage. This was referred to by one of the co-authors as a CRREL-Gandahl gage. This gage used a polyethylene inner tube of 12.7 mm I.D. and 15.9 mm O.D. and was filled with a 0.05 percent solution of methylene blue. A nylon string up the center of the tube anchored the ice and prevented floating during thaw periods. The outer casing was of rigid polyethylene or PVC pipe. It was specified that the inner diameter of the casing should be no more than 6 mm greater than the outer diameter of the inner tube. They recommended that the gage should be at least 0.30 m longer than expected frost depth to minimize freezing point depression as dye migrates from the freezing front.

Researchers of the USDA-ARS at Pullman, Washington and of the Agricultural Engineering Department of the University of Idaho experimented with versions of the Gandahl and CRREL tubes. A tube filled with fluorescein solution and sand as described by Harris (1970), but with a rigid inner tube, was first used by USDA-ARS during the winter of 1974/75 and by the University of Idaho in the winter of 1975/76. The design seemed to work fairly well for researchers and technicians acquainted with the idiosyncrasies of the gage. However, many cooperative observers could not differentiate the color change from yellow-green to gray as the liquid froze. Also, super-cooling was a problem in that some tubes did not begin to freeze until exposed to very cold temperatures. There was also occasional difficulty with the rigid inner tube freezing and breaking if the gages were exposed to cold temperatures prior to installation. Thus, in the fall of 1978, both USDA-ARS and University of Idaho adopted the CRREL-Gandahl design. The line between frozen and unfrozen material was much easier to locate, and if necessary, the ice blocks could be located by squeezing the flexible tube.

There seemed to be some problems with freezing and thawing of the solution in the tubes lagging behind freezing and thawing of liquid in the soil. Thus, in the fall of 1979, 4 mm diameter glass beads were added to the methylene blue solution in the tube. The intent was to decrease the quantity of liquid in the tubes and hence thermal lag. However, the beads seemed to offer little advantage and were discontinued. The CRREL-Gandahl tube is currently used by these researchers.

SHEAR STRENGTH

While a soil is frozen, its shear strength is determined by the strength of the ice crystals, and the ice-soil bonds. Several tests are available to determine soil strength under laboratory conditions such as the direct shear test and the triaxial shear test. Those tests require undisturbed samples and are not practical for weak agricultural soil conditions, such as tilled or thawing soils. Only in-situ methods to determine soil strength are discussed.

Vane shear device

The shear or Torvane test was described by several researchers (Tuma and Abdel-Hady, 1973, Jumikis, 1962, Caquot and Kerisel, 1967). Vertical vanes radiating from a central point are inserted in the soil and rotated around their common axis. The torque needed to rotate the cylinder of soil between the vanes is related to shear strength by the equation:

$$M_{max} = \tau * \pi (D^2 H/2 + D^3/12) \quad (1)$$

where M_{max} = torsional shear resistance (N m)
τ = shear strength of the soil (N m^{-2})
D = diameter of vane (m)
H = height of vane (m).

Dickenson et al. (1982) found a good relationship between vane shear, bulk density and volumetric water content:

$$\tau = a + b(BD) + c(VW) \quad (2)$$

where τ = vane shear strength (N m^{-2})
BD = bulk density (kg m^{-3})
VW = volumetric water content ($m^3 m^{-3}$).
a,b,c = regression coefficients.

Caquot and Kerisel (1967) proposed rounded edges on the vertical vanes since the shear plane in stiff clays did not follow the cylinder formed by the vanes. Douglas (1986) used a vane apparatus to study the effect of management on shear strength of a clay topsoil. He measured strength 19 times during a season and found a linear relationship between water content and strength, and a similar relationship between bulk density and strength. He stressed that understanding the effects of season on strength would require more frequent measurements and that all pertinent soil physical properties should be closely monitored (such as depth of freezing). Elliot et al. (1988) found the Torvane to produce non-repeatable data on freshly eroded soil during WEPP rainfall simulation tests. Bradford and Grossman (1982) stated that using a vane test on loam and silt loam soils has limited value because inserting the tool into the soil would cause failure and underestimation of the shear strength. Hirschi et al. (1987) determined a weak relationship between Torvane shear strength and soil loss under simulated rainfall. Kok (1989) found a weak relationship between Torvane shear strength and soil water content of the top 10 mm of the soil profile for thawing soil under field conditions. He found the device easy to use for a wide array of soil conditions, but hard to clean between measurements.

Penetrometer

The penetrometer test consists of a cone or foot which is forced into the soil at a constant rate. The force required to obtain a preset depth of penetration is read manually or recorded with a strain gauge. Penetrometers measure a complex of forces due to penetration: cutting and separating of soil, shear, plastic flow, compression and metal-to-soil friction (Hillel, 1980). Penetration resistance is dependent on water content, bulk density, soil type, soil strength, penetration rate, shape of the cone or foot and surface roughness of the cone or foot. Perumpral (1987) found penetrometer resistance to decrease with increasing water content and a decreased effect of density with increased water content. Penetrometers are available in capacities ranging from vehicle mounted for penetration of highly compacted soils such as roads (up to 10,000 kg), to small pocket penetrometers for very weak soil conditions (0.015 kg). Nearing and West (1988) found a pocket penetrometer to detect different consolidation levels in three soils satisfactorily. Kok (1989) used a pocket penetrometer on thawing soil and found fluctuations in soil strength to reach beyond the measuring range of the penetrometer. A series of adapter feet were made to increase the area tested . The device performed well and is very easy to handle and clean, however, no satisfactory relationship between measurements made with the various adapter feet could be established.

Swedish fall-cone apparatus

The Swedish fall-cone apparatus (Hansbo, 1957) is widely used in Europe for measuring the shear strength of clays (Towner, 1973). A cone of known weight and configuration is suspended over the soil surface, just touching it. The cone is released and the depth of penetration is registered. Shear strength is related to depth of penetration by the expression:

$$\tau = K_f * W/H^2 \qquad (3)$$

where τ = shear strength (N m^{-2})
K_f = a factor of proportionality ($9.8*10^3$ m s^{-2})
W = mass of the cone (g)
H = depth of penetration (mm)

In comparing the fall-cone apparatus with unconfined shear strength measurements, Towner (1973) found K_f to be dependent on soil texture and cone weight and constant with water content. Kullman et al. (1984) determined relationships between fall-cone penetration, water content and porosity for three soils. Al-Durrah and Bradford (1982) found detachment due to single raindrop impact to be a linear function of the ratio of water drop kinetic energy to fall-cone strength on a per soil basis, expressed as:

$$D = a + b*(K_e/\tau) \qquad (4)$$

where D = detachment (mg per drop)
K_e = water drop kinetic energy (J)
τ = fall-cone shear strength (kPa)
a,b = regression coefficients.

The slope of the line differed among the soils used in their experiment. Nearing and Bradford (1985) found the Swedish fall-cone to over-predict the resistance of a soil to detachment by raindrop splash due to over-prediction of the resistance to lateral jetting of water after impact. The above mentioned researchers have used a K_f value only dependent on soil texture. All samples used were remolded test cores, eliminating any effects of aggregate size and profile characteristics. Bradford and Grossman (1982) measured surface strength of soils in the field using a modified Swedish fall-cone device. They found the device to work well, stressing that soil water potential and cone weight influence the measurements. Heavier cones penetrate deeper and since it is common in the field to have a variation of consolidation in the upper 20 mm of the profile, they measure a different strength than the lighter cones. Franti et al. (1985) reported poor correlation between Swedish fall-cone shear strength and soil erodibility. In a field experiment with flow applied to rills they could not account for differences in erodibility between no-till and conventionally tilled fields. Different cone weights were used and they did not state whether cone weights were accounted for in the K_f value used. Elliot (1988) found Swedish fall-cone measurements, taken during rainfall simulation experiments for the WEPP project, to be highly variable. Formanek (Formanek et al., 1984) and Van Klaveren (Van Klaveren and McCool, 1987) used a fall-cone to measure strength of a thawing soil under laboratory conditions. Kok (1989) used a fall-cone to measure strength of a Palouse silt loam in-situ during 2 winters and found a very high variability in readings, which was attributed to spatial variability.

SOIL WATER CONTENT

Resistance blocks can be used to measure electrical resistance of the soil solution and hence determine soil water content. The method is sensitive to temperature changes and not feasible for year around water content determination.

The volume of soil probed with the neutron scattering method is dependent on the water content and bulk density of the soil: in wet soils about 100 mm, in dry soils about 250 mm around the probe. This makes the method infeasible to determine distinctive water changes in a profile such as wetting fronts. The method does not work in the top 200 mm of the soil due to neutrons lost into the atmosphere (NeSmith et al., 1986).

Gravimetric water content sampling is a labor intensive and time consuming method but it works for all weather and soil conditions. The method of sampling is important; some core samplers cause compaction and may not accurately represent profile conditions.

SOIL MATRIC POTENTIAL

Matric potential can be determined by tensiometers in the 0 to 80 kPa range. Freezing of water inside the tensiometers will destroy them. Wendt et al. (1978) used tensiometers filled with a methanol water mixture and found the matric potential readings very similar to readings with water filled tensiometers. No measurements on frozen or thawing soils were reported. Pikul and Allmaras (1985) used methanol as frost protection for tensiometers that were installed over-winter. Potential changes measured during diurnal freeze-thaw were obtained with water filled tensiometers, installed under the frost layer. Thermocouple psychrometers measure potential in the 2 to 50 bar range but do not withstand frost either (Hillel, 1982).

The filter paper method (Hamblin, 1981, Al-Khafaf and Hanks, 1974, McQueen and Miller, 1968) is a slow but reliable and inexpensive procedure. Soil samples are allowed to equilibrate with Whatman #42 filter papers in a constant temperature environment until the soil water matric potential is in equilibrium with the filter paper water potential. Filter paper matric potential is related to the mass water content of the filter paper by the relationship:

$$\psi = 10^{(3.5726-0.01458*M)} \quad \text{for } M > 41.22 \tag{5a}$$

and

$$\psi = 10^{(6.6711-0.08975*M)} \quad \text{for } M < 41.22 \tag{5b}$$

where ψ = filter paper matric potential (kPa)
M = (filter mass water / mass dry filter)*100 (percent by weight)

Tests by Kok (1989) showed that the disturbance of the very weak structure of thawing soils during field sampling caused erratic results in matric potential determination. Insitu equalibration of the soil with the filter papers is not practical as large temperature fluctuations would severely influence the results.

BULK DENSITY

Methods to determine bulk density were reviewed extensively by Erbach (1987), including compliant cavity, core sampling, radiation and cone index procedures. The sandcone method and the balloon method (Means and Parcher, 1963) are of the compliant cavity type. A section of soil is excavated and the soil is weighed and dried. The volume of the excavation is determined by filling the hole with a known volume of either sand or water, usually inside a membrane to prevent infiltration. All methods require establishing a reference level, which is hard to do when the soil is frozen or very weak after thaw (Kok, 1989).

Gamma ray transmission uses probes inserted in the soil about 200 mm apart. Under dry conditions inserting the pins can cause cracks, which will cause the method to underpredict bulk density (Henshall and Campbell, 1983).

Core samples of known volume will work under a wide range of weather and soil conditions. For frozen conditions specialized equipment is needed however**. Under very wet or dry conditions samples might not stay in the tube or inserting the tube may cause compaction of the soil (Casias, 1987; Doran and Mielke, 1984; Martin, 1986). Previous research on the Palouse silt loam, under laboratory conditions, was successfully done with double ring core samplers (Van Klaveren, 1987). Field sampling of thawing soils with double ring samplers resulted in very high variability, and was impossible under very weak soil conditions (Kok, 1989).

Radiation methods include gamma ray scattering, gamma ray attenuation and neutron scattering. Reports as to the success of their use in the top 100 millimeter of the soil profile are contradictory. The instruments are very expensive and must be calibrated for the soil on which they will be used (Erbach, 1987).

ARTIFICAL FREEZING AND THAWING OF SOILS

Soil samples can be frozen in several ways. Formanek (1983; Formanek et al., 1984) used a thermo-electric plate to freeze small soil samples at -4°C and -10°C. Freezing to

** R. A. Young. USDA-ARS-NCSC, Morris, MN. Personal communication, 1988.

a depth of 50 mm was obtained in 20 to 30 hours. The soils were thawed overnight by reversing the setting on the thermoelectric plate to $20^{0}C$.

Van Klaveren (1987, Van Klaveren et al., 1987) used radiative freezing to freeze large soil samples (2.74m long by 0.45 m wide and 0.6m deep). Automotive anti-freeze was cooled to between -12 and $-22^{0}C$ and circulated through a plate suspended over the sample surface. Freezing to a depth of 120 mm was achieved in 7 to 14 days. Thawing was obtained in 5 to 14 days by increasing the air temperature in the laboratory.

Earlier attempts to freeze a very large soil sample (the size of a railroad wagon) by convective freezing at the Palouse Conservation Field Station, Pullman, WA, were not successful. The large volume of cooled air blown over the sample caused the soil surface to dry substantially (unpublished internal report).

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A COMPARISON OF RUNOFF OCCURRING ON FROZEN AND UNFROZEN SOILS

Myron Molnau and J. G. Cherry[1]

INTRODUCTION

Various experimental watershed throughout the nation have obtained data for frozen soil events as part of their routine monitoring but rarely are experiments conducted to specifically obtain data from frozen soil events. As a part of a study to model frozen ground runoff events, a survey of hydrograph data from USDA watersheds was conducted to find as many frozen ground runoff events as possible. This paper presents frozen ground runoff data from six different locations in the United States which were found as a part of that study.

DATA COLLECTION AND ANALYSIS

Table 1 summarizes the data from each watershed. A more detailed description of just one location is given here as an example to show how the data can be used to analyze both frozen and unfrozen runoff events.

The runoff data for each watershed were examined to find frozen ground events. Data from East Lansing, MI, Coshocton, OH, Moscow, ID, Pullman, WA, Corvallis, OR and Pendleton, OR were found to have some frozen ground events. Other watersheds in Wisconsin, Nebraska and Vermont were also examined but it could not be definitely determined whether the events found were actually frozen soil or not and these were therefore eliminated.

The criteria for determining if the event did occur on frozen soil varied for each watershed. Frost depths or soil temperatures were actually measured at some of the sites while at others, the presence of frost had to be inferred by studying rainfall/runoff ratios, antecedent air temperatures and notes in the station logbooks. Freezing degree-days were computed and frozen soil runoff was assumed when the freeze index exceeded 83°C-days (McCool and Molnau, 1974).

One of the few watershed established specifically to study frozen soil runoff events was at East Lansing, Michigan. Rainfall, runoff, climatological and watershed records from 1941 to 1951 were available for analysis. Using these available data, comparisons between runoff on frozen and unfrozen ground were made. Watershed characteristics, such as detention storage and infiltration, were also determined. For this comparison, two watersheds were chosen. These watersheds had the following characteristics:

	Watershed A	Watershed B
Area	1.93 acres	1.345 acres
Slope	6.0 percent	6.5 percent
Soil	Fine sandy loam Loamy fine sand	Fine sandy loam Loamy fine sand
Cover and Rotation	Corn with cover crop, oats and 2 years of hay	Corn with cover crop, oats and 2 years of hay

Paper presented at the Frozen Soil Symposium, Spokane, WA, March 21-22, 1990.

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For each watershed, a pair of runoff events was chosen, one occurring on frozen soil and one on unfrozen soil. Events with similar antecedent conditions were chosen. From the available data it was concluded that all events occurred when little or no snow was on the ground. Runoff hydrographs were developed from the rainfall and runoff records. Data for these and all other hydrographs were stored in a database for ease of retrieval and updating. Hydrologic differences between the frozen and unfrozen events are evident.

The lag time for these events was determined and found to be much shorter for the frozen soil runoff events than the lag time on unfrozen soil. Because of the frozen ground, little infiltration occurs and the runoff appears at an earlier time than on unfrozen ground. The frozen soil runoff events also exhibited more runoff than the unfrozen events. Watershed A had a high of 98 percent of the rainfall appear a runoff on frozen ground as compared to 65 percent on unfrozen ground. Watershed B had a high of 89 pe:cent of the rainfall appear as runoff on frozen ground and 84 percent on unfrozen ground.

As seen from Tables 1 and 2, a significant difference exists between the amount of runoff on frozen and unfrozen ground. Further analysis of the unfrozen hydrographs was conducted to determine watershed characteristics. The detention storage and infiltration capacity of the watersheds were first examined. A method proposed by Dunin(1969) was used to calculate detention and infiltration. Dunin suggested that the recession limb of the hydrograph fit an exponential equation of the form

$$q(t)=q(0)k^t$$

$q(t)$ and $q(0)$ = flow at time t and time 0 respectively
t = time
k = hourly depletion rate

For two unfrozen ground hydrographs that met the requirement that the recession limb be rainless, the average depletion rate was 0.277 for Watershed A and 0.162 for Watershed B.

The detention storage of the watershed can now be determined since k is known. Holtan and Overton(1963 and 1964) derived an expression relating detention and the hourly depletion rate as

$$D_t = \frac{q_t}{\ln k}$$

where D(t) is the detention for flow q(t). Approximately 51 percent of the rainfall on Watershed A went into detention storage and approximately 46 percent of the rainfall on Watershed B went into detention storage. The infiltration rate could now be determined. Infiltration occurred during the first five hours of rainfall on Watershed A. For Watershed B, infiltration only occurred over the first three hours of rainfall.

With the infiltration characteristics determined for the watersheds, hydrographs could be generated. Unit hydrographs were developed for each watershed using the SCS methods.

For each watershed, the following unit hydrograph characteristics were determined:

	Watershed A	Watershed B
t(lag)	.058 hrs	.054 hrs
t(c)	.097 hrs	.090 hrs
D	.013 hrs	.012 hrs
t_p	.065 hrs	.060 hrs
Q_p	23.08 cfs	16.94 cfs

When using these unit hydrograph values to generate actual runoff hydrographs for these events, it was found that the SCS method overpredicted the runoff in the beginning and end of runoff but underestimateed the peak runoff. Part of this error is due to the small duration times.

SUMMARY

Although the watersheds are small, certain conclusions can be drawn about their hydrologic properties when comparing frozen and unfrozen runoff hydrographs. The SCS method illustrated the short lag time, time of concentration and peak runoff of the watersheds. Dunin's method showed the watersheds have a significant amount of detention storage. Infiltration was quite low and only occurred in the first few hours of the runoff events. Also, there is a major difference between runoff events on frozen and unfrozen soil. Frozen soils have a much shorter lag time and yield more runoff than unfrozen soil.

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Table 1. Summary of runoff from frozen and unfrozen soil for six locations in the United States.

Watershed	Area (acres)	Crop	Summary of Events: Date of Storm	*F/NF	Precip (inches)	Runoff (inches)	% of Precip as Runoff	Comments
East Lansing, MI Watershed A	1.93	Corn with cover crop, oats and 2 years of hay	01/08/42	F	1.35	1.40	101.4	Some thawing
East Lansing, MI Watershed B	1.35	Corn with cover crop, oats and 2 years of hay	01/08/42	F	1.38	1.19	86.2	Ground bare of snow
East Lansing, MI Watershed B	1.35	Corn with cover crop, oats and 2 years of hay	01/23/49	F	0.12	0.11	91.7	Ground bare of snow
East Lansing, MI Watershed A	1.93	Corn with cover crop, oats and 2 years of hay	01/18/49	NF	1.70	1.11	65.3	Trace of snow on ground
East Lansing, MI Watershed B	1.35	Corn with cover crop, oats and 2 years of hay	01/18/49	NF	1.70	1.43	84.1	Trace of snow on ground
Corvallis, OR Watershed E4F2	3.46	Grass seed, small grains, hay, pasture, orchard and forest	01/09/79	F	1.92	0.40	20.3	Trace of snow on ground
Corvallis, OR Watershed E4F2	14.8	Grass seed, small grains, hay, pasture, orchard and forest	01/09/79	F	1.92	0.61	31.2	Trace of snow on ground
Corvallis, OR Watershed E4F1	3.46	Grass seed, small grains, hay, pasture, orchard and forest	2/27/89	NF	0.58	0.20	34.5	Ground bare of snow
Corvallis, OR Watershed E4F2	14.8	Grass seed, small grains, hay, pasture, orchard and forest	03/15/79	NF	0.29	0.02	7.9	Ground bare of snow

Watershed	Area (acres)	Crop	Date of Storm	*F/NF	Precip (inches)	Runoff (inches)	% of Precip as Runoff	Comments
Moscow, ID Naylor Watershed	177.9	Wheat, alfalfa, sweet clover and peas	01/24/41	F	0.49	0.45	91.8	Ground bare of snow
Moscow, ID Naylor Watershed	177.9	Wheat, alfalfa, sweet clover and peas	02/01/41	NF	0.25	0.06	22.0	Ground bare of snow
Pullman, WA Watershed GS2	68.2	Wheat, peas and clover	01/23/35	F	0.23	0.31	135	Snow--depth unknown
Pullman, WA Watershed GS5	14.4	Winter Wheat	01/23/35	F	0.23	0.25	109	Snow--depth unknown
Pullman, WA Watershed GS6	15.2	Peas and clover	01/23/35	F	0.23	0.26	113	Snow--depth unknown
Pullman, WA Watershed GS8	762	Wheat, peas and fallow	01/23/35	F	0.23	0.31	135	Snow--depth unknown
Pullman, WA Watershed GS2	68.2	Wheat, clover and fallow	03/08/36	NF	0.35	0.05	14.8	Ground bare of snow
Pullman, WA Watershed GS5	14.4	Peas	03/08/36	NF	0.35	0.09	25.4	Ground bare of snow
Pullman, WA Watershed GS6	15.2	Peas and clover	03/08/36	NF	0.35	0.06	16.3	Ground bare of snow
Pullman, WA Watershed GS8	762	Wheat, peas and fallow	03/08/36	NF	0.35	0.08	21.7	Ground bare of snow
Pendleton, OR Conlie Watershed	0.014	Newly seeded winter wheat	01/03/83	F	0.19	0.20	105	Trace of snow on ground
Pendleton, OR Neal Watershed	0.014	Newly seeded winter wheat	01/02/83	F	0.42	0.32	76.2	Trace of snow on ground
Pendleton, OR Conlie Watershed	0.014	Newly seeded winter wheat	02/17/83	NF	0.72	0.40	55.0	Ground bare of snow

Watershed	Area (acres)	Crop	Date of Storm	*F/NF	Precip (inches)	Runoff (inches)	% of Precip as Runoff	Comments
Pendleton, OR Neal Watershed	0.014	Newly seeded winter wheat	03/08/83	NF	0.43	0.19	43.0	Ground bare of snow
Coshocton, OH Watershed 106	1.56	Meadow and pasture	01/24/79	F	0.44	0.35	79.5	Snow--depth unknown
Coshocton, OH Watershed 110	1.27	Pasture	01/25/76	F	0.70	0.57	81.4	Snow--depth unknown
Coshocton, OH Watershed 129	2.71	Pasture	01/13/76	F	0.46	0.43	93.5	Snow--depth unknown
Coshocton, OH Watershed 106	1.56	Meadow and pasture	03/21/80	NF	0.86	0.08	8.8	Ground bare of snow
Coshocton, OH Watershed 110	1.27	Pasture	03/21/80	NF	0.92	0.09	9.3	Ground bare of snow
Coshocton, OH Watershed 129	2.71	Pasture	03/21/80	NF	0.84	0.01	1.2	Ground bare of snow

*F - Frozen
NF - Not Frozen

Table 2. Rainfall/runoff percentages for selected events at the East Lansing, MI watershed.

Watershed	Conditions	Number of Events	Average percent of Rainfall as Runoff
A	Frozen	4	71
B	Frozen	3	88
A	Unfrozen	6	31
B	Unfrozen	6	27

FROST DATA IN THE PACIFIC NORTHWEST

Myron Molnau D. K. McCool G. E. Formanek[1]

INTRODUCTION

The depth of frozen soil is of interest in many aspects of engineering work. Utility lines need to be buried below the frost line while foundations must be deep enough to avoid problems with heaving. Frozen soils are of particular interest to hydrologists because of the influence of soil frost on runoff and erosion.

The effects of frozen soils on human activities have been extensively documented. These studies represent hydrologic effects (Dingman, 1975; Molnau and Bissell, 1983), construction of roads, buildings and pipelines (Jumikis,1977; Sanger, 1959) and agriculture (Wall, 1987; Cary and others, 1978). These and other papers give ample justification for studying frozen soil but frost depths are not a commonly measured variable in non-research data collection programs. Therefore, the few data sets that are available need to be used by many people in obtaining a solution to their problem.

This paper describes frost depth data sets that have been systematically collected in the Pacific Northwest.

FROST DEPTH DATA SETS IN THE PACIFIC NORTHWEST

Several types of data sets have been collected in the Pacific Northwest. These may be characterized as reconnaissance, systematic, (such as climatic networks) and research or special purpose data sets.

Reconnaissance

The Soil Conservation Service(SCS) and the Agricultural Research Service(ARS) collected a reconnaissance dataset in Idaho, Oregon, Utah and Washington. A total of 608 station years of data were collected in 20 counties. This network was started to obtain data over a large area with a wide spatial distribution (Table 1).

The objective of operating this network was to obtain data under a wide range of tillage conditions and climate in order to better understand the role of freeze/thaw cycles in the erosion process.

These data were collected by farmers and SCS cooperators using frost tubes (McCool and Molnau, 1984). Maximum and minimum temperatures and snow depths were also measured, usually on a daily basis. Many of the cooperators read the instruments only during periods of deep frost so that period of little or no frost are underrepresented. To some extent, the number of stations shown in Table 1 reflects the severity of frost in a particular year. Often, the observer will abandon or not read the station if it appears that little or no frost occurs. Portions of this network are still being operated by the local Soil Conservation Districts with the overall coordination being done by the ARS at Pullman, WA.

Paper presented at the Frozen Soil Symposium, Spokane, WA March 21-22, 1990.

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Table 1. Total Station Years for Frost Sample Sites By State and Water Year

State	WY79	WY80	WY81	WY82	WY83	WY84	WY85	WY86	WY87	TOTAL
ID	0	0	10	12	6	11	11	7	0	39
OR	0	0	0	0	2	2	0	0	0	4
UT	0	0	3	4	4	3	2	1	0	14
WA	20	26	110	115	95	65	78	85	32	551
Total	20	26	123	131	107	81	51	37	32	608

Systematic

In 1977, the western states' Agricultural Experiment Stations Regional Research Committee on climate instituted a program of systematic collection of frost depth data at climatological stations. Frost tubes were installed in six states. Good data were collected from Idaho, Montana, Wyoming and Colorado. From two to eight frost tubes were installed in each state, usually at the agricultural experiment stations where a National Weather Service climatic station was also present. Data are currently being collected in Idaho and Montana. The data sets from Idaho and Montana are currently available and have been used for various frost depth analyses throughout the area. A description of most of these stations is given by van Rooij (1987).

Research

The ARS maintains several frost depth networks in the Pacific Northwest. These are used for research purposes and, as such, have much more intensive data available. Often there are many meteorological elements collected along with the basic frost, snow depth and air temperature. In the PNW, there are two large networks and several smaller ones. The large networks are located at the Reynolds Creek Experimental Watershed of the Northwest Watershed Research Center, ARS at Boise, Idaho. A second large network is operated by the Palouse Conservation Field Station (PCFS) of the Land Management and Water Conservation Research Unit, ARS at Pullman, Washington.

The Reynolds Creek data have been collected for hydrologic purposes since at least 1975 as part of the data collection program (Hanson and others, 1988). Forty eight station years of record have been put into a single database for public distribution. The period of record is 1976-77 to 1983-84 with additional years now being added. Table 2 is a summary of the records available from Reynolds Creek (Hanson and others, 1988). These data have been used in a variety of applications (Bullen, 1982; Flerchinger and Hanson,1990).

Table 2. Frost data inventory for the Reynolds Creek Experimental Watershed

Station	Elev	Aspect	Period of Record
*** sites at weather stations ***			
76x59	3715	Flat	1970 - present
127x07	5410	N	1970 - present
176x14	6880	NW	1972 - present
*** sites at non-weather stations ***			
various	3885-7100	various	114 station-years

The PCFS plots at Pullman, Washington are used to determine the effect of tillage and management practices on erosion and runoff. These are standard runoff plots with two frost tubes on each plot (Table 3). Two tubes are necessary because the plots are on a steep hillside and the soil varies with location on the slope. A complete meteorological station is located near the plots to enable the data to be used in the derivation and validation of detailed frost models such as that of Flerchinger and Saxton (1989) and Kok (1989).

Table 3. Data inventory for the Palouse Conservation Field Station Plots

Water year	Number of sites
1979	18
1980	22
1981	29
1982	33
1983	32
1985	35
1986	36
1987	30

A Comment on Quality Control

The quality of data directly governs the possible applications of that data. Here data quality is defined to mean not just the data accuracy but also the frequency and the auxiliary information that is available. If data are to be used merely for a broad view of frost depths over a large area, then spatial coverage is important with many reading surrounding the period of interest. On the other hand, if one were interested only in the maximum frost depths, then good readings around that period of time are needed while the onset and thaw periods are of less importance.

In order to meet these varying requirements in a general purpose database, it is important that data be screened before they are accessed by the public. A screening procedure is needed that will find the periods of poor or suspect record and then either correct or flag them. Since most of the data sets to be screened are not very large, the initial screening is usually done by using a graphical technique. In a graph, the missing and poor records are usually quite obvious particularly when graphed along with multiple elements such as temperature, snow-on-ground and precipitation.

SUMMARY

Frost depth data sets are relatively rare, especially over a large area and for any significant length of time. The Pacific Northwest is fortunate in that several data sets ranging from reconnaissance data sets to detailed research data sets are available. These are now being applied to a variety of projects ranging from frost depth mapping to detailed frost depth modelling.

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SOIL THAW AS INFLUENCED BY SMALL GRAIN RESIDUE COLOR

by

B.S. Sharratt and G.S. Campbell[1]

INTRODUCTION

Soil conservation cropping practices necessitate maintaining crop residue cover on the soil where there exists potential for erosion. Crop residues not only reduce wind and water speeds at the soil surface, and thus reduce soil movement and loss, but also influence the soil water and heat balance. The residues can affect soil freezing and thaw and growing season soil temperature and water.

Crop residues placed on the soil surface affect the heat and water balance by reflecting incoming global radiation and insulating the soil. Aase and Siddoway (1980) found the albedo of wheat stubble was 1.5 times higher than a bare soil surface, resulting in cooler stubble-covered soils. Gupta et al. (1981) reported that maximum surface soil temperatures were as much as 15°C higher over bare surfaces than corn residue-covered surfaces. Similarly, Unger (1978) found higher maximum and mean daily 10cm soil temperatures under a bare surface compared to residue-covered surfaces. Crop residues on the soil surface suppress vapor flow from the soil to the atmosphere, thereby conserving water. Water may also be conserved by enhancing infiltration in residue-covered soils (Van Doran and Allmaras, 1978).

Bidlake (1988), working in interior Alaska, found that early season soil temperatures were higher in conventional-till (bare surface) than no-till (residue-covered) soils. Soil water was lower in the conventional-till soil in the early season. Thus, in interior Alaska, residues slow soil warming and may delay planting.

Interior Alaska is characterized by a short growing season (about 90 to 100 days) that is generally mild (July maximum air temperature of 15°C) and moist (precipitation from May through August of 12cm). These climatic factors are constraints to the diversity and production of agricultural crops grown in Alaska. Thus crop residue management methods of conservation cropping systems are sought that enhance global radiation absorption and heat movement to the soil, thereby promoting early planting and improving the soil and aerial environment of plants. The purpose of this study is to determine the effect of coloring small grain residue on soil temperature and spring thaw.

MATERIALS AND METHODS

The study was conducted at the University of Alaska Agricultural and Forestry Experiment Station near Fairbanks, Alaska (64°51′N, 147°52′W). The relief of the research site was level and 145m above mean sea level.

Barley residue color treatments were randomized within a block with 3 replications. Plots were 6m on a side and established after harvest in August of 1987 and 1988. Barley was harvested by cutting at a 15 and 30cm height in the respective years and raking and removing loose residue. The amount of stubble remaining was determined on a dry weight basis by clipping the stubble at the soil surface from 10 areas of $0.1m^2$ in 1987 and from 4 areas of $0.5m^2$ in 1988. The amount of stubble was 300 and 1600 kg/ha in the respective years. Stubble was painted using an exterior paint and applied with an airless sprayer. Boards were placed between rows of stubble to minimize overspray and coloring of the soil surface. Barley residue plots were painted either black or white, or remained unpainted for the third color treatment (yellow).

Presented at the Frozen Soils Symposium, March 21-22, 1990; Spokane, Washington.

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Loose residue (excluding stubble) remaining following threshing was simulated by hand applying a uniform layer of colored barley residue. Loose residue laid on a tarp was painted in an attempt to apply a uniform and complete coat of paint. The amount of loose, colored residue applied to plots was determined to achieve 2500 and 3500 kg/ha of total barley residue in 1987 and 1988, respectively. Loose, colored residue was applied to the corresponding stubble-colored plots.

Plots were instrumented following the application of colored residue to all plots to monitor soil temperature, heat flow, and reflected radiation. Concurrent measurements were made of incoming global radiation, air temperature, relative humidity, and wind speed. Soil temperatures at 1 and 5cm were averaged from 6 locations in each plot by wiring thermocouples in series. Temperatures at a depth of 10, 20, and 40cm were measured at one location in each plot. Soil heat flux was measured in one replication of each treatment with a heat flux transducer placed at a 3cm depth. Reflected global radiation was monitored in one replication of each treatment using a pyranometer mounted on a boom orientated in a southerly direction. Air temperature, relative humidity and wind speed were monitored at a height of 2m.

The instruments were wired to a micrologger. Hourly data were recorded prior to snowmelt and into June of both years. Following snowmelt, soil water, thaw depth and barley surface residue temperatures were measured on a weekly basis. Thaw depth was determined using a pointed rod and inserting into the soil until resistance prevented further insertion. Periodic comparisons were made with a soil core sampler. Soil water was determined gravemetrically and surface residue temperatures measured with an infrared thermometer.

RESULTS AND DISCUSSION

Climatic conditions during the early season of 1989 were cooler and wetter than the previous year. Air temperature for a 20 day period from April 19 (approximately snowmelt) to May 9 was 5.9 and 6.6°C in 1989 and 1988, respectively. Global radiation was 13.0 and 16.0 MJ/day and precipitation 1.5 and 0.9cm in the respective years.

Thaw in 1989 was more rapid than the previous year. The soil thawed 20 cm between days 114 and 126 in 1988. Over this same time period the soil thawed 40cm in 1989 (Table 1). The rapid recession of the freezing front in 1989 may have been influenced by the relatively warmer subsoil. At the time of snowmelt, soil temperature at 40cm was about -4°C in 1988 and -2°C in 1989.

Table 1. Depth of soil thaw under black, white and yellow colored-barley residue following snowmelt on day 108 in 1988 and day 113 in 1989 at Fairbanks, AK.

Year	Julian day	Depth of soil thaw Black	White	Yellow	LSD (0.1)
		------------ cm	------------		
1988					
	109	6.4	5.6	5.5	0.7
	112	18.8	15.0	15.6	4.0
	113	19.5	19.2	18.8	4.2
	118	31.5	30.5	31.0	2.0
	126	46.1	43.5	42.5	4.8
	140	39.8	39.6	39.4	4.4
1989					
	114	8.9	7.2	7.3	0.5
	116	23.0	22.8	23.1	5.3
	117	24.7	23.3	26.3	4.0
	121	41.0	40.7	45.3	7.5
	126	51.3	53.3	54.3	3.5

The color of barley residue had an effect on the global radiation balance as indicated by the daily trend in albedo in Figure 1. The differences in albedo became apparent before complete snowmelt (Figure 1). Visual observations in the field indicated the emergence of stubble on day 105 in 1989 and day 104 in 1988. Complete snowmelt occurred on days 113 and 108 in 1989 and 1988, respectively. The albedo remained fairly constant following snowmelt. The albedo of black residue was 0.05 and for the yellow and white residue was 0.2 and 0.3. A change in the albedo of the residue color plots occurred at the time the plots were seeded. Following seeding on day 131 in 1989, the trend in albedo tended to converge as the albedo of the black colored residue increased and of the white colored residue decreased (Figure 1).

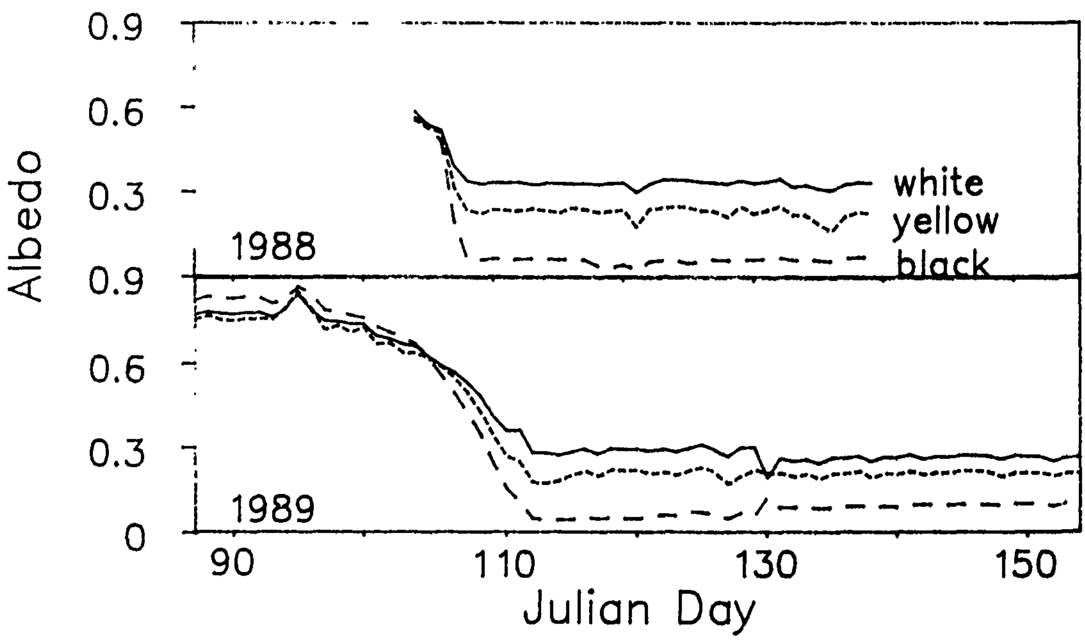

Figure 1. Albedo of white, yellow, and black colored-barley residue in the early growing seasons of 1988 and 1989 at Fairbanks, Alaska.

With an approximate 15 to 25% greater absorption of global radiation due to blackening the barley residue, greater energy may be available for thawing and warming the soil. Immediately following snowmelt in 1988, soil thaw was greater on the black colored-residue plot (Table 1). Two days following snowmelt the soil was thawed to 6.5cm under the black residue and 5.5cm under the white and yellow residue. Differences in thaw were not apparent at five days after snowmelt (Table 1). Similar observations of differences in soil thaw due to residue color were made following snowmelt in 1989, but did not persist thereafter (Table 1).

Clear day diurnal soil temperature trends at the 5cm depth indicated thaw occurred earlier in the day under black residue, followed by thaw under yellow, then white residue (Figure 2). Maximum temperatures under black residue were higher than under white and yellow residue. This was exemplified by the diurnal trend in 5cm soil temperatures in Figure 2. On day 109 in 1988, black residue maximum soil temperatures were significantly higher (1.3°C) than white residue temperatures. No differences were found between black and yellow residue soil temperatures on this day. On day 137 in 1989, soil temperatures were significantly higher (2.0°C) under black residue than under white residue with no differences between black and yellow residue temperatures. However the similarity of night temperatures dampened the effect of barley residue color with no difference in daily temperatures.

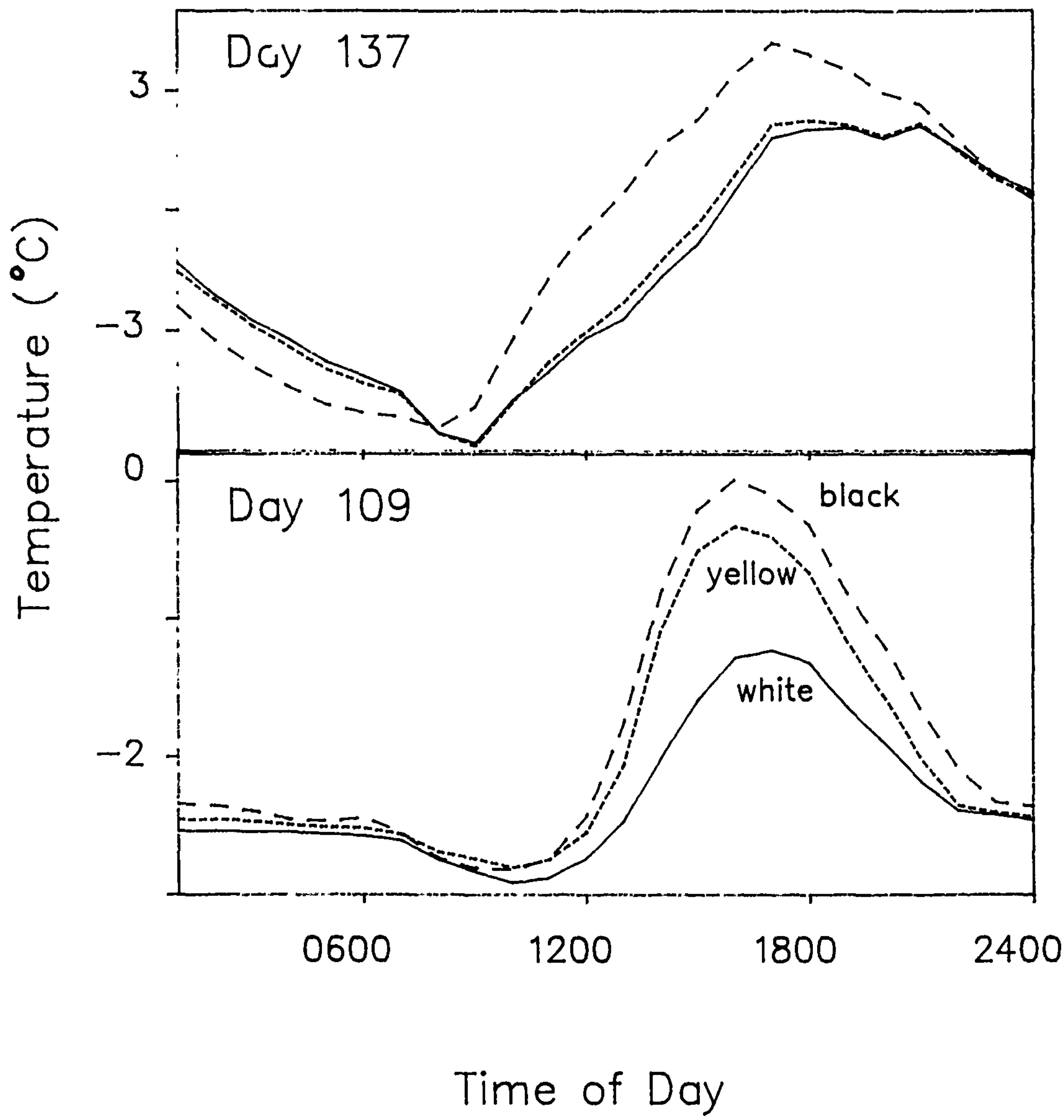

Figure 2. Diurnal soil temperature trend at 5cm under black, white, and yellow colored-barley residue on days 109 and 137 in 1988 and 1989, respectively, at Fairbanks, Alaska.

CONCLUSIONS

Small grain residue color affects absorption of global radiation. Black colored residue absorbs 15 to 25% more radiation than white or yellow residue. The resultant higher energy absorption on the black residue only slightly affected soil temperatures and soil thaw. Therefore other means of dissipating the additional radiant energy on the black residue appear more important than conduction into the soil. A mechanical device, such as plastic, to trap the greater radiant energy on black residue may result in elevated soil temperatures and enhance soil thaw.

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SOIL STRUCTURE AND FROST DEPTH AS AFFECTED BY SOIL COMPACTION

by

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INTRODUCTION

Soil compaction caused by wheel traffic from ordinary farming operations can have a significant effect on several state soil properties that affect movement of heat and water through the soil profile. These relationships have been reviewed by Barnes et al. (1971). One of the major concerns with respect to soil compaction is the increasing evidence from long-term field experiments that soil compaction can persist much longer than previously thought. In latitudes where freezing temperatures commonly extend through the surface 30 cm of soil, earlier field research showed little evidence that soil compacted by wheel traffic during the growing season would persist over the winter (Phillips and Kirkham, 1962). But within the past two decades several researchers have documented and reported persistence of soil compaction for several years in spite of annual freezing and thawing. Eriksson (1976) measured compaction from heavy military equipment extending to depths of 1 m and persisting for several years. Blake et al. (1976) reported persistence of an artificially induced plow pan in Minnesota for 10 years.

Reasons for this change in soil compaction persistence are varied. In the northern Corn Belt of the U.S., three factors have probably dominated: (1) row crops have replaced crop rotations that included cereal crops, grasses and legumes; row crops conventionally require more field operations, and thus more wheel traffic; (2) the size and weight of farm machinery have increased tremendously since World War II, and thus both the depth and intensity of the compaction process has increased; (3) use of reduced tillage has increased. Fall moldboard plowing was the norm in the northern Corn Belt, and this intensive type of tillage was sufficient to ameliorate the annual compaction problem, if there was one. Inadvertently, freezing and thawing during the non-growing season were singularly credited with preventing any soil compaction problems from accumulating.

This paper will review and discuss recent field research in Minnesota concerning some relationships between soil compaction and soil freezing, as well as address some practical implications of these findings.

METHODS

Two sets of field experiments will be discussed. The first experiment was conducted on a Ves clay loam (formerly Nicollet clay loam, Aquic Hapludoll) located on the Southwest Minnesota Experiment Station in Lamberton, Minnesota. One of the experimental objectives was to quantify the effect of wheel traffic from normal field operations on various soil physical properties in the surface 30 cm. Field machinery typical for corn production (tractors with gross weights up to 7 Mg using 6-row equipment covering a width of 4.6 m) was used. Heavy harvesting equipment was not used on this experiment. Replicated compaction treatments consisted of no interrow wheel traffic at any time during the course of the experiment and wheel traffic controlled to occur within the same interrows within and across seasons. There was an average of 5 wheel passes per year. Bulk density, penetrometer resistance and aggregate size were periodically measured to assess the extent to which annual freezing and thawing ameliorated compacted soil conditions in the surface 30 cm. Detailed description of the experiment can be found in Voorhees et al. (1978).

Another experiment to assess the persistence of subsoil compaction was conducted on a Webster clay loam (Typic Haplaquoll) at the Southern Minnesota Experiment Station in Waseca, Minnesota, and on a Ves clay loam at the Southwestern Minnesota Experiment Station in Lamberton, Minnesota. In the fall of 1981 subsoil compaction was induced by trafficking

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Presented at Frozen Soil Symposium, March 21-22, 1990, Spokane, Washington.

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the entire plot surface with the wheel traffic from farm equipment with axle loads of 9 and 18 Mg, typical for harvest and transport equipment. Previous research had shown that axle loads in excess of 9 Mg could cause compaction below the normal depth of tillage (Danfors, 1974). A control treatment with no heavy wheel traffic was also included at both sites. The Webster soil was relatively wet at time of trafficking while the Ves soil was relatively dry. The intensity and depth of compaction were assessed by measuring bulk density, penetrometer resistance, and hydraulic conductivity immediately after trafficking. All treatments were then moldboard plowed to a depth of about 25 cm to alleviate surface compaction. Thereafter, fall tillage consisted of either no-till on the Webster soil or reduced tillage on the Ves soil. All subsequent wheel traffic normal for growing corn and soybeans was limited to 4.5 Mg/axle or less.

Frost tubes (Ricard et al., 1976) were installed in all 4 replications of the control and 18 Mg/axle treatments, and readings were taken weekly once the soil started to freeze until spring thaw. Average monthly air temperatures and snowfall were measured in a nearby Class A weather station. Snow cover on the plots was also measured but was of limited value because of random and uneven drifting. The plots were void of any plant residue cover the first winter because of the initial plowing. There were plant residues the second and third winters. Experimental details are discussed in Voorhees et al. (1986).

RESULTS

Surface compaction

The persistence of surface compaction over winter is shown in Figure 1 where bulk density is plotted at various times over a period of one year. Beginning with an initial bulk density in the surface 30 cm ranging from 1.3 to 1.4 Mg/m^3, one season of wheel traffic increased the values to about 1.6 Mg/m^3 throughout the surface 30 cm (compare curves E and A in Fig. 1). By the following spring, with no fall tillage, natural weathering forces decreased the bulk density in the surface 10 cm of the wheel track by only 0.1 Mg/m^3 (compare curves A and B). Fall tillage (especially plowing) in combination with natural weathering was quite effective in ameliorating surface compaction (compare curves C and D with A). Moldboard plowing may be more effective than reduced tillage because it exposes more of the surface soil to freezing temperatures.

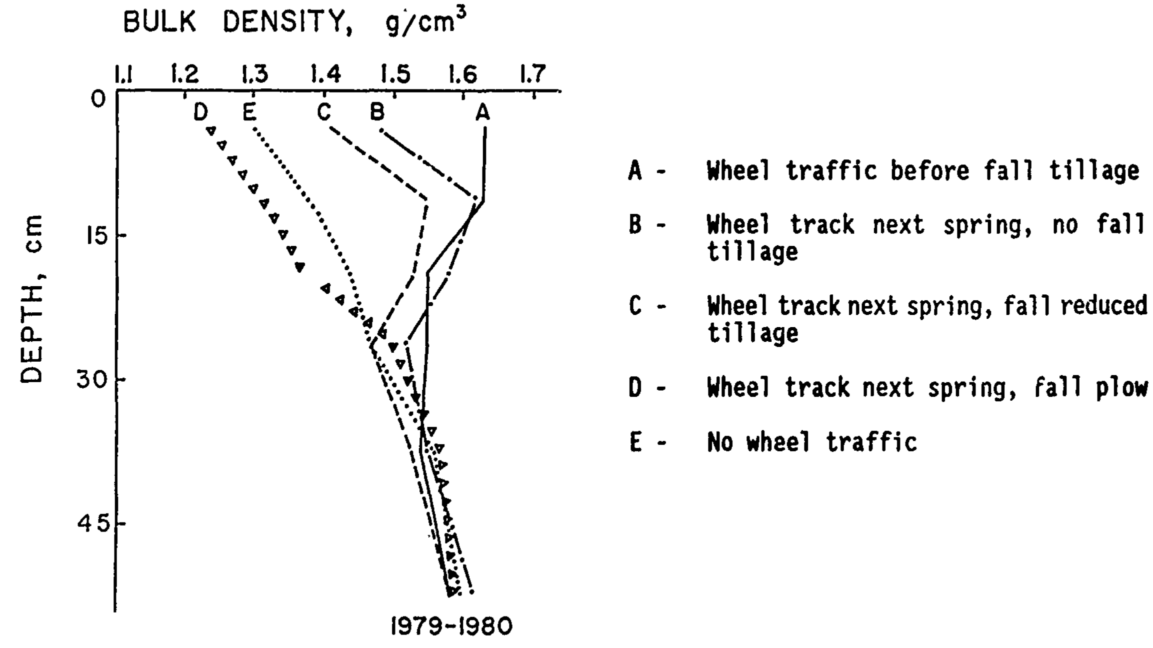

Figure 1. **Bulk density as affected by wheel traffic, weathering, and fall tillage (Voorhees, 1983).**

In addition to long-term effects of traffic on bulk soil properties such as bulk density, there were also significant effects on individual soil structural units. Data in Table 1 show that density and strength of individual aggregates can be affected by compaction and that this effect can be quite resistant to degradation by freezing and thawing. Normally, climatic conditions are such that this soil is not subjected to more than one freeze-thaw cycle each winter with the exception of the surface few mm, suggesting that several freeze-thaw cycles are needed to ameliorate compacted soil.

Table 1. Density and crushing strength of wheel-tracked and nontracked soil clods. Lamberton, MN (Voorhees et al., 1978).

Sampling date	Wheel-tracked	Nontracked	Significance
	Clod density, g/cm^3		
May 1975	1.72	1.56	**
October 1975	1.73	1.47	**
May 1976	1.59	1.49	**
	Crushing strength, g/cm^2		
October 1975	572	134	**

**Wheel-tracked values significantly higher than nontracked values at the 0.01 level as determined by Student "t" test for difference between means.

<u>Subsoil compaction</u>

Figure 2 shows the increase in subsoil bulk density in response to applied loads of 9 and 18 Mg/axle for the Ves and Webster soils. The Webster soil (trafficked when wet) showed a significant increase in bulk density to a depth of almost 60 cm under the 18 Mg/axle load. The Ves soil, trafficked when dry, showed an increase in bulk density to only about 35 cm. These differences persisted across more than one winter (Voorhees et al., 1986; Voorhees et al., 1989; and Johnson et al., 1989).

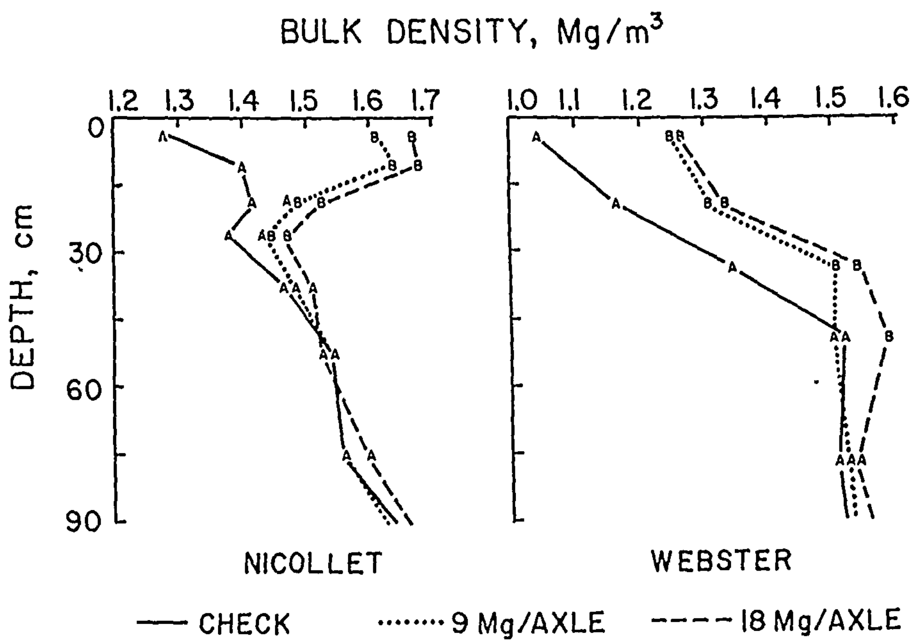

Figure 2. Bulk density in response to high axle loads on Ves (Nicollet) and Webster soils. October, 1981 (Voorhees et al., 1986).

Patterns of frost penetration in response to subsoil compaction are illustrated in Figures 3-11. Differences between compacted and noncompacted are seldom significant but are generally consistent within a year, but not necessarily across years. This is partially due to uneven snow drifting across the plots. If comparison is made between two individual plots having approximate equal snow cover, then compaction effects are generally significant. But the variability in snow cover (timing and amount) appears to cause a correspondingly high variation in frost penetration such that compaction treatments averaged over replications are not significant. The effectiveness of snow cover in influencing frost penetration has been well documented (Benoit et al., 1988). In addition, soil water content at time of freezing has a large influence on frost penetration (Benoit and Mostaghimi, 1985).

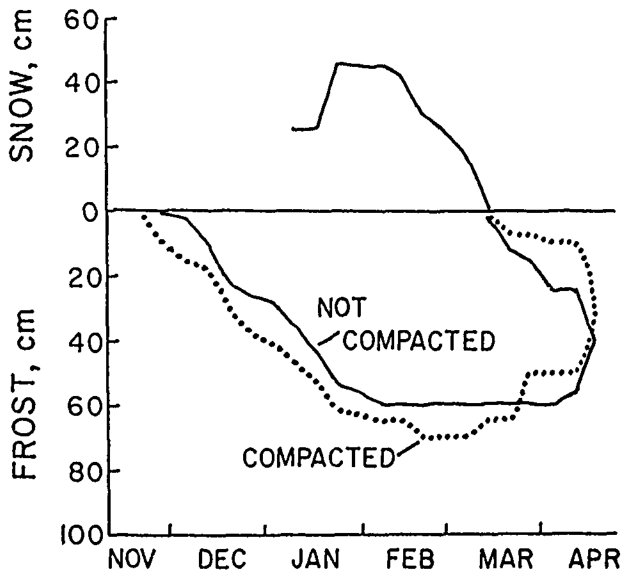

Figure 3. Snow and frost depth on Webster soil, 1981-82. Bare soil surface.

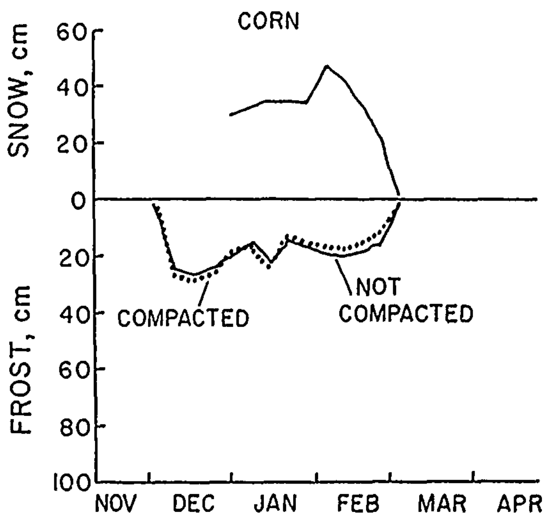

Figure 4. Snow and frost depth on Webster soil, 1982-83. Corn residue, no fall tillage.

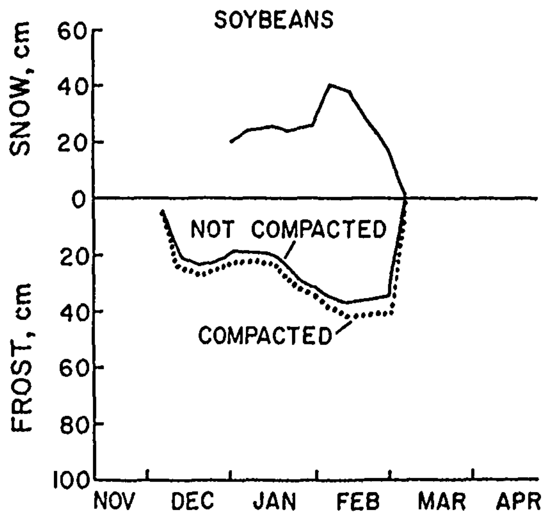

Figure 5. Snow and frost depth on Webster soil, 1982-83. Soybean residue, no fall tillage.

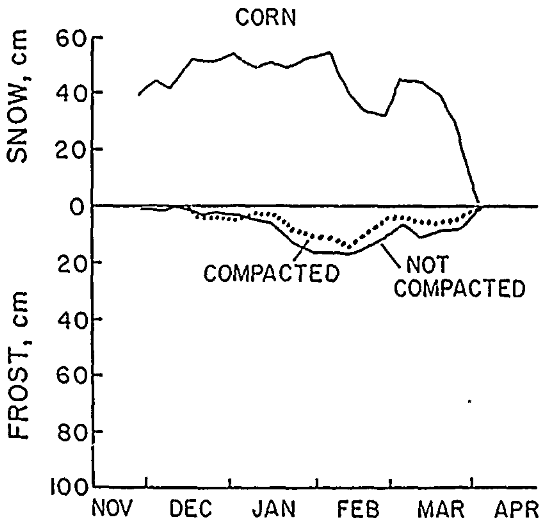

Figure 6. Snow and frost depth on Webster soil, 1983-84. Corn residue, no fall tillage.

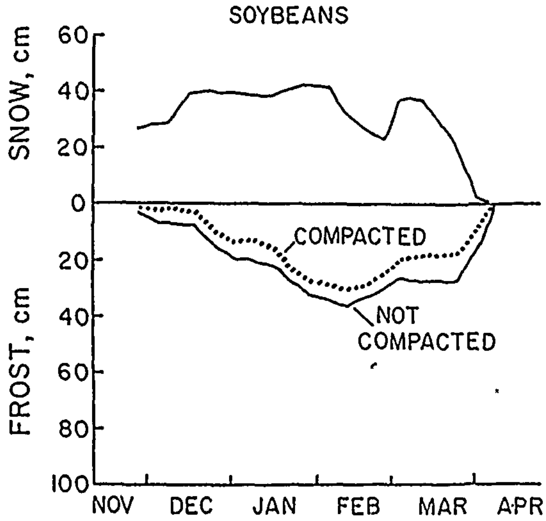

Figure 7. Snow and frost depth on Webster soil, 1983-84. Soybean residue, no fall tillage.

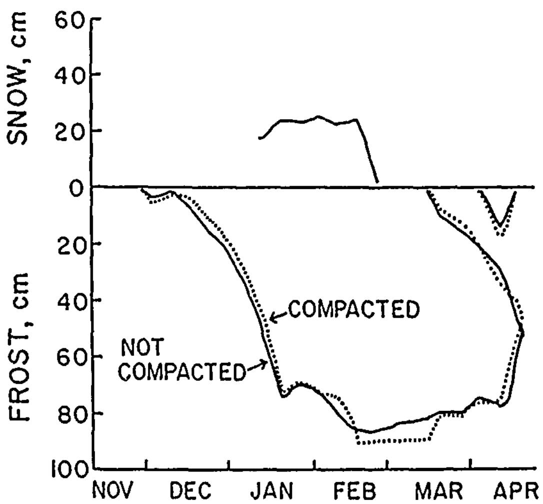

Figure 8. Snow and frost depth on Ves soil, 1981-82. Bare soil surface.

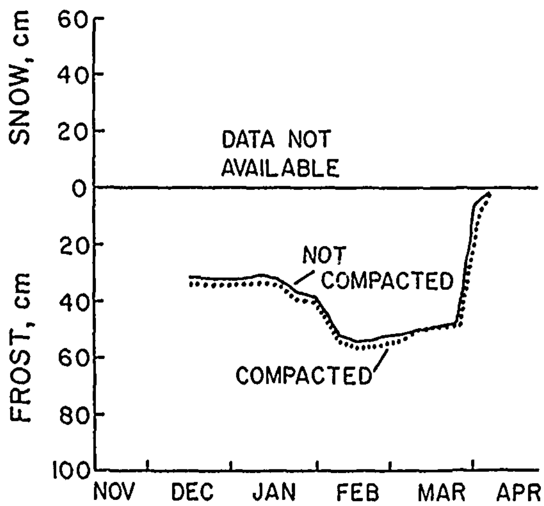

Figure 9. Snow and frost depth on Ves soil, 1982-83.

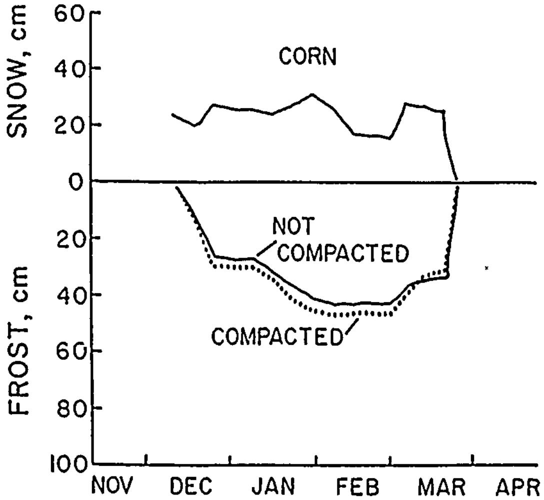

Figure 10. Snow and frost depth on Ves soil, 1983-84. Corn residue, interrow chisel tillage.

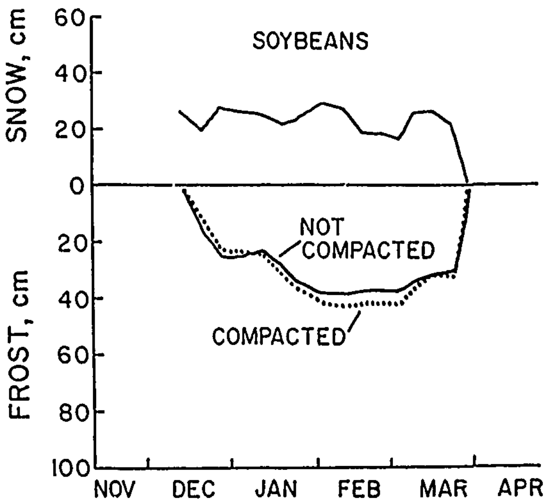

Figure 11. Snow and frost depth on Ves soil, 1983-84. Soybean residue, no fall tillage.

There is merit, however, in discussing frost penetration trends within a year and comparing these trends with the climatic conditions for that year. The average monthly air temperature and accumulated snowfall are shown in Figures 12 and 13 for the Webster and Ves soils, respectively. For both soils, the winters of 1981-1982 and 1983-1984 were relatively cold while 1982-1983 was relatively warm. Both sites received the highest snowfall in 1983-1984.

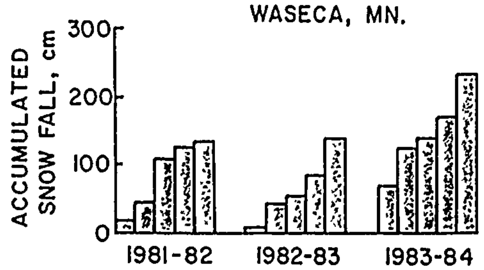

Figure 12. Winter air temperature and snowfall at Waseca, MN, 1981-84.

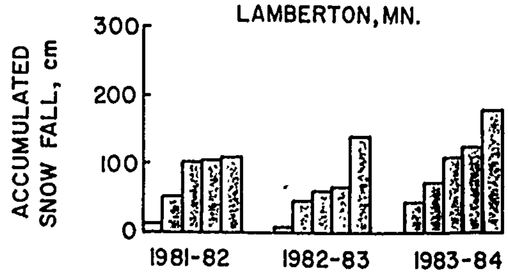

Figure 13. Winter air temperature and snowfall at Lamberton, MN, 1981-84.

For the Webster soil in 1981-1982, the treatments with subsoil compaction started to freeze about a week sooner than the noncompacted plots (Fig. 3). The rate of frost penetration remained equal throughout the winter with the noncompacted plots lagging the compacted plot by about 1 week. The compacted plots froze about 5-10 cm deeper than the noncompacted plots. The cessation of downward frost movement coincided with the timing of maximum snow cover on the plots and minimum air temperatures (Figs. 3 and 12). The compacted plots also started thawing from the bottom sooner than the noncompacted plots but thawed slower from the surface. The higher bulk density in the compacted subsoil (Fig. 2) likely increased the thermal conductivity to cause faster and deeper frost penetration. The reasons for decreased thawing from the surface are not obvious, but they were concomitant with observed delayed drying of the surface soil in early spring (Voorhees and Lindstrom, 1983).

Frost penetration was limited to only 30-40 cm on the Webster soil during the winter of 1982-1983 (Fig. 4 and 5) due to relatively warm temperatures (Fig. 12). The only real difference in frost penetration was related to differences in plant residues on the soil surface. Plots with soybean residues accumulated about 10 cm less snow (Fig. 5) than plots with corn residues, (Fig. 4) and froze about 10 cm deeper. Even though differences in snow accumulation were measurable early in the winter season, differences in frost penetration did not occur until the cold temperatures of January. There were no differences in frost penetration due to subsoil compaction.

The winter of 1983-1984 on the Webster soil was very cold early (December) but also had the greatest snowfall (Fig. 12). Consequently, frost penetration was the shallowest of the three years that measurements were made (Figs. 6 and 7). There were very consistent differences in frost penetration, with the noncompacted plots freezing about 5 cm deeper than the compacted plots; this was opposite the effect seen in 1981-1982. These differences were not statistically real because of the uneven snow drifting discussed earlier. The differences between corn (Fig. 6) and soybean (Fig. 7) residues in accumulating snow were again apparent as in 1982-1983 and were reflected in differences in frost penetration.

Average monthly air temperatures at the Ves site were almost identical to the Webster site but snowfall tended to be less at the Ves site (Fig. 13). More important perhaps is the fact that about 40-50 percent of the snowfall accumulated on the Webster plots but only 20-25 percent accumulated on the Ves plots due to more wind. Consequently, frost penetrated about 20 cm deeper every year in the Ves soil compared with the Webster soil (compare Figs. 3 and 8; 4 and 5 with 9; 6 and 7 with 10 and 11). Similarly, there was much less effect of crop residue on frost penetration in the Ves soil compared with the Webster soil (compare Figs. 6 and 7 with 10 and 11) because there was some fall tillage on the Ves soils that buried some plant residues and decreased their effectiveness for catching drifting snow.

There were no differences in frost penetration due to subsoil compaction on the Ves soil. This should be expected because the degree and depth of subsoil compaction on the Ves soil was less than on the Webster soil (Fig. 2). There was a slight trend for compacted subsoil to freeze deeper than noncompacted subsoil for both soils most years.

DISCUSSION

The relationship between movement of heat and water through the soil is very complex. Models have been developed that can predict movement of frost through the soil profile given various soil, snow, and residue parameters (Benoit and Mostaghimi, 1985). The effect of soil compaction alone on frost movement can be calculated if it is known how corresponding basic soil characteristics are altered by compactive forces. But under field conditions, compaction effects rarely, if ever, are the only active factors. The data above clearly show the overriding influence of snow cover. If compaction affects crop growth, then the amount of residue on the soil surface may affect snow catch. Compaction can also affect the depth of snow cover by influencing the amount of upward heat loss early in the winter season when soil initially begins to freeze. Thus, it is difficult to delineate cause-and-effect relationships between soil compaction and snow cover.

The frost data presented above have significant implications in several important aspects of agriculture. The relative ineffectiveness of freezing and thawing to ameliorate the surface soil layer, as shown in Figure 1 and Table 1, emphasizes the need to reevaluate some of our previous soil management assumptions developed under cultural practices that existed a few decades ago. Soils compacted by wheel traffic under a no-till system can have significantly slower infiltration rates (Lindstrom et al., 1981), and can be an important factor in soil erosion and water runoff (Voorhees et al., 1979). The density and stability of compacted soil aggregates can affect root growth and nutrient uptake (Voorhees et al., 1971).

Perhaps the most important myth to dispel concerns subsoil compaction under today's farming methods and the ability for natural forces, especially freezing and thawing, to ameliorate these conditions. Using the growing plant as an integrator of the various ways that subsoil compaction can affect soil properties over time, Voorhees et al. (1989) showed significant corn yield reductions on compacted plots for the soils and years over which frost penetration data were reported in Figures 3 through 11. Furthermore, during the drought stress year of 1988, corn yields were decreased by 15 percent on plots that had been compacted in the fall of 1981 with the 18 Mg/axle load (Voorhees, unpublished data, 1988); seven years of annual freezing and thawing were not sufficient to completely ameliorate subsoil compaction.

The persistence of subsoil compaction in the northern Corn Belt may progressively become a more serious problem with the trend towards reduced tillage. The data presented in Figures 3 through 11 illustrate the importance of snow cover in moderating the rate and depth of frost penetration. No-till characteristically leaves more residue on the soil surface than conventional tillage. Thus, the more plant residue on the surface, the less ameliorative action of frost in the subsoil. Kay et al. (1985) reported that soil pores created by ice lenses in a no-till system were unstable compared with those created in a plowed situation. During subsequent thawing, the no-till soil more easily consolidated, resulting in a higher bulk density. Compacted soil below the tilled layer can profoundly affect the hydraulic conductivity of the tilled layer (Blake et al., 1976), and thereby affect soil erosion.

The data discussed here illustrate the complexity of soil compaction-frost interactions and point out the need for continued research to delineate cause-and-effect relationships.

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APPENDIX: ABSTRACTS

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

FROZEN SOIL IMPACTS ON AGRICULTURAL, RANGE AND FOREST LANDS— AN INTRODUCTION

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 1-3
Spokane, Washington, March 21-22, 1990

Soil freezing and thawing is a natural phenomena of any landscape in which the climatic conditions and physical setting are such that heat transfer results in a soil water phase change. Soil freezing impacts are particularly important on agricultural, range and forest lands where mankind gleans valuable food and fiber while desiring to preserve the sustainability of the land. Of particular interest is the management of agricultural soils where tillage is an option for manipulating the structure and porosity for liquids and air. Empirical evidence often has been obtained but the physical description and predictability are still quite lacking. The papers found in this publication present a broad summary of the current understanding about frozen soils and their impact on the production activities found on these lands. The compiled information of these papers will provide a sound base from which to develop the current status of frozen soil impacts and launch additional research. With improved knowledge of both occurrence and impact of soil freezing, appropriate management choices can be made for resource utilization and conservation on our farms, rangelands, and forests.

SUBJECT INDEX

1. Freezing
2. Thawing
3. Landscape
4. Susceptibility
5. Management
6. Knowledge
7. Resource

AUTHOR INDEX

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Saxton, et al.
(cont'd.)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

FROZEN SOIL, RUNOFF AND SOIL EROSION RESEARCH IN NORTHEASTERN OREGON

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 4-10
Spokane, Washington, March 21-22, 1990

This paper summarizes research conducted in northeastern Oregon over the past 10 years. The relationships between frozen soils, snowmelt, rain on snow and runoff and erosion are discussed using field data. The data indicate that a major cause of accelerated runoff and soil erosion is rapid snowmelt associated with high dewpoint temperatures, high windspeeds and rain on snow while the soil is frozen. The results of experiments on the effects of tillage and residue on the depth and duration of soil frost are discussed. In general, surface residue reduces frost penetration while thawing rate is a function of atmospheric conditions during the thawing period. Water infiltration rates into frozen soils were investigated using a rainfall simulator. Frozen soil reduced infiltration for chiseled and standing stubble treatments when compared to the unfrozen rates. No infiltration occurred on winter wheat while the soil was frozen.

SUBJECT INDEX

1. Modeling
2. Tillage
3. Infiltration
4. Probability
5. Frequency analysis
6. Snowmelt
7. Rain on snow.

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

NATURE OF THE CRYIC THERMAL REGIME OF AGRICULTURAL SOILS IN THE YUKON TERRITORY, CANADA

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 11-20
Spokane, Washington, March 21-22, 1990

Soils which support forage-based agriculture near Dawson and Whitehorse, Yukon Territory, were monitored over five years to determine soil thermal regime. Using monthly values, the normal annual soil temperature cycle at 10, 20 and 50 cm depths was fit as a function of time to a Fourier series curve with two harmonics. Based on the estimated daily values generated by the analyses, the mean annual soil temperature, mean summer soil temperature, the length of season and accumulated degree days above 0, 5 and 15°C were calculated. The Dawson soil had a mean annual soil temperature at 50 cm of 0.9°C and a mean summer temperature of 7.4°C giving it a very cold or cryic thermal regime. The Whitehorse soil had a mean annual soil temperature at 50 cm of -0.1°C and a mean summer temperature of 4.1°C giving it an extremely cold or pergelic thermal regime. The plow layer of these soils does not thaw until late April and returns to the frozen state by mid October. Soil temperatures seldom exceed 15°C.

A well-drained soil from a second site near Whitehorse, was monitored daily through a nine-month period to establish seasonal trends

SUBJECT INDEX

1. Soil temperature
2. Thermal regime,
3. Brunisol
4. Cryocrept
5. Growing degree days

AUTHOR INDEX

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in freezing and thawing. Cultivated soils near Whitehorse subject to cooling rates between 0.02 and 0.06°C/day were observed over a four-month period. The difference in rates of warming and cooling and in minimum observed temperatures can be attributed to snow cover.

Smith (cont'd)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

SOIL FREEZING IN A SUBARCTIC DECIDUOUS FOREST

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 21-30
Spokane, Washington, March 21-22, 1990

We monitored processes related to the freezing of a moist subarctic soil in a forested setting to facilitate a comparison of methods of measuring and calculating the soil freezing front. After fall irrigation of the study site, the thermal and moisture states of the soil profile were measured frequently throughout the winter using nondestructive methods. The advance of the freezing front was monitored using thermistors, the unfrozen moisture content of the soil was determined through time domain reflectometry, and the total soil moisture and soil density were determined using a two probe nuclear density/moisture gage. The rates and depth of freeze were calculated using the Stefan equation formulated for layered media (St. Paul equations) and with a finite element computer model incorporating heat conduction and phase change. These simulation results were compared to the measured values of frost depth.

SUBJECT INDEX

1. Soil freezing
2. Modeling
3. Time domain reflectometer
4. Neutron moisture probe

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

TILLAGE AND CROP RESIDUE EFFECTS ON SOIL FROST DEPTH

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 31-35
Spokane, Washington, March 21-22, 1990

Crop residues can influence soil frost depth through their insulating value and their ability to catch snow. This study was designed to determine the effects of crop residue management, in three farming systems, on soil freezing and thawing. The Alternate (A) farming system consisted of a four year crop rotation (oat/alfalfa - alfalfa - soybean - corn) with no synthetic fertilizer or pesticide and no moldboard plow. Primary fall tillage was chisel plow after alfalfa and disk after corn. The conventional (C) and Ridge-till (R) systems both had a corn - soybean - wheat rotation. The C tillage was fall disk after corn and moldboard plow after wheat. The R tillage was chisel plow after spring wheat. Systems C and R used recommended rates of fertilizer and pesticides. During the 1986-87 winter, with no snow cover, soil frost depth was oat/alfalfa < corn < spring wheat < soybean. Fall tillage which reduced crop residue generally reduced snow catch. In 1987-88 and 1988-89, moderate snow catch (4-8 cm) reduced soil frost depth beyond moderate levels eliminated differences in frost depth due to residue management.

SUBJECT INDEX

1. Soil frost depth
2. Residue management
3. Farming systems
4. Tillage

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

COMPARISON OF NUMERICAL SIMULATIONS WITH EXPERIMENTAL DATA FOR A PROTOTYPE ARTIFICIAL GROUND FREEZING

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 36-44
Spokane, Washington, March 21-22, 1990

This project numerically simulates the use of artificial ground freezing as a means of containing and consolidating toxic chemical spills. This innovative treatment technology can cleanse the soil insitu through the use of freeze/thaw cycles. Artificial ground freezing has been in practice for over a century in civil engineering applications. Its ability to form impermeable barriers, dewater sludge, and consolidate solids has been demonstrated. Additionally, the environmental dangers of the treatment process are virtually nonexistent even in populated regions. However, predicting or controlling the location of the frozen barrier as a function of time for arbitrary geometries is unsolved. This control is paramount for successful implementation of the process.

The numerical simulations herein are compared to existing, large-scale experimental data for situations involving the solidification of saturated Lebanon silt soil. The successful agreement of the numerical simulations and experimental data supports the use of numerical modeling as a necessary tool for deployment of refrigeration systems for the treatment of toxic spills.

SUBJECT INDEX

1. Frozen soil
2. Hazardous waste containment
3. Artificial ground freezing

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

EFFECT OF FREEZE-THAW ACTIVITY ON WATER RETENTION, HYDRAULIC CONDUCTIVITY, DENSITY, AND SURFACE STRENGTH OF TWO SOILS FROZEN AT HIGH WATER CONTENT

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 45-53
Spokane, Washington, March 21-22, 1990

The effect of freeze-thaw cycles on soil physical properties was determined on a Barnes loam (Udic Haploboroll) and a Hamerly cl (Aeric calciaquoll, fine-loamy, frigid) soil. Soil water-holding capacity, hydraulic conductivity, bulk density, and surface strength were measured before and after 1, 5, and 10 freeze-thaw cycles. Water content at freezing and at time of physical measurement was established by saturating all samples and then allowing free drainage to the atmosphere. Tests were made on samples prepared by packing soil aggregates 0-0.5, 1-3, or 5-12 mm in diameter into 76- by 76-mm plastic cores to target bulk densities of 1, 1.2, 1.4, and 1.6 Mg/m^3. Actual bulk densities achieved after soaking and free drainage and before freeze treatments were applied were determined and used as initial values for all calculations involving soil density. Results indicate that freeze-thaw activity improves the physical condition of dense samples but decreases the condition of samples with low initial bulk density. Thus, freezing and thawing samples with an initial bulk density of 1.0 increases bulk density, surface strength, and water-holding capacity but decreases hydraulic conductivity. The reverse is true for samples with initial target densities of 1.6. In addition, the magnitude of the freeze-thaw density relation is modified by initial aggregate size.

SUBJECT INDEX

1. Freeze-thaw
2. Soil water content
3. Hydraulic conductivity
4. Bulk density
5. Soil aggregates

AUTHOR INDEX

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Benoit and Voorhees (cont'd)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

PREDICTING UNFROZEN WATER CONTENT BEHAVIOR USING FREEZING POINT DEPRESSION DATA

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 54-60
Spokane, Washington, March 21-22, 1990

It has been recognized since the pioneering work of Taber and Beskow that the existence of a continuous liquid layer separating the soil matrix from the ice in frozen soil is the controlling factor in the dynamics of soil freezing. All constitutive relationships for hydraulic, mechanical and thermal properties of frozen soil are functions of the quantity of unfrozen water present. Successful agricultural solutions to the adverse effects of seasonal freezing on plant growth therefore depend upon the ability to predict or measure the amount of unfrozen water present. Unfortunately, most laboratory and field researchers lack a simple and inexpensive method to determine unfrozen water content. This brief paper presents unfrozen water contents determined by the freezing point depression method and nuclear magnetic resonance. It is found that within the experimental limitations, these two methods give similar results, which suggests that the simple freezing point depression method is a viable method of inexpensively determining unfrozen water contents.

SUBJECT INDEX

1. Freezing point depression
2. Frozen soil
3. Nuclear magnetic resonance
4. Unfrozen water content

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

EFFECTS OF FREEZING ON AGGREGATE STABILITY OF SOILS DIFFERING IN TEXTURE, MINERALOGY, AND ORGANIC MATTER CONTENT

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 61-69
Spokane, Washington, March 21-22, 1990

Aggregate stability, an important property influencing a soil's response to erosive forces, is affected by freezing. The objectives of two laboratory studies were to determine the effects of constrainment, number of freeze-thaw cycles, water content at freezing, and the addition of potential inorganic bonding agents on the aggregate stability of six continental U.S. soils differing in texture, mineralogy, and organic matter. Moist aggregates, after being frozen and thawed either 0, 1, 3, or 5 times, were vapor-wetted to 0.30 g g and analyzed by wet sieving. Aggregate stability decreased with increasing water content at freezing but, for some soils particularly when unconstrained, at first increased with increasing number of freeze-thaw cycles but then decreased. After thawing, aggregates at gravimetric water contents of 20% or more that were constrained when frozen were less stable than aggregates that were unconstrained when frozen.

SUBJECT INDEX

1. Constrainment
2. Water content
3. Bonding agents
4. Calcium carbonate
5. Calcium sulfate
6. Wet sieving

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CREEL
Special Report 90-1
Catalogue No.

FREEZE THAW EFFECTS ON SOIL STRENGTH

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 70-76
Spokane, Washington, March 21-22, 1990

Research on the effects of freezing and thawing on soil properties has been conducted at the Palouse Conservation Field Station, Pullman, Washington. One field study and two laboratory experiments are discussed. A brief summary of the results obtained in these studies is given.

SUBJECT INDEX

1. Soil
2. Shear strength
3. Erosion
4. Freeze-thaw
5. Modeling

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

SOIL FREEZING AND THAWING SIMULATION WITH THE SHAW MODEL

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 77-86
Spokane, Washington, March 21-22, 1990

Frozen soil plays a significant role in the hydrology and management of watersheds in northern climates. Soil freezing can inhibit infiltration and increase the potential for flooding and erosion. Many methods have been developed for predicting frost depth in soils, but most require surface temperature or surface heat flux for input, and none include the effects of snow, residue, and tillage on heat and water flow. The Simultaneous Heat and Water (SHAW) model uses hourly weather data to simulate the interrelated heat, water, and solute transfer through snow, residue, and soil for a wide range of conditions. This paper illustrates the SHAW model's applicability to widely varying site and climatic conditions. Field data and model results are presented for: shallow, transient freeze-thaw cycles prevalent in the Pacific Northwest grain lands around Pullman, Washington; elevational effects in the mountainous rangelands of the Reynolds Creek Watershed in southwestern Idaho; and deep, soil-frost conditions in interior Alaska. Maximum frost depth was simulated within 3 cm for all plots at the Pullman and Reynolds Creek sites. (Maximum frost depth at the Alaska site exceeded monitoring depth.) Soil temperature, snow depth, and rate of freezing and thawing were successfully simulated for all sites.

Flerchinger, et al.
(cont'd)

SUBJECT INDEX

1. Frozen soil
2. Snow
3. Tillage
4. Hydrologic modeling
5. Rangeland

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

SIMULATING THE FREEZING AND THAWING OF ARABLE LAND IN SWEDEN

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 87-98
Spokane, Washington, March 21-22, 1990

The impact of freezing and thawing on arable land can be seen on the structural changes of the soil and on the paths used by the snow melt water. During spring, the leakage of nutrients will be influenced as well. A physically based numerical model was used to simulate the frozen soil as well as the atmosphere and the snow cover. Simulations were made on typical Swedish agricultural soils as a heavy clay soil and a layered loamy soil. The snow cover had a large influence on the simulated heat flow. Difficulties arose when the snow cover was simulated using daily mean values as input data. As an upper boundary condition measured snow depth was only of limited use; the water equivalent values contained more information. Simulating frost with "hydraulic models" necessitate the use of an "impedence factor" to limit upward flow of water during freezing. The variation of this impedence factor and the sensitivity for different soil are discussed. In heavy clay soils freezing contributes to the development of a crack system and an increased percentage of voids after thawing. On a virgin soil these changes are irreversible. For a soil experiencing seasonal frost most of the voids created by the frost

SUBJECT INDEX

1. Coupled heat and mass transfer
2. Hydraulic conductivity of frozen soil
3. Infiltration
4. Soil-moisture characteristic curve

AUTHOR INDEX

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Lundin (cont'd)

will disappear as the soil consolidates. Comparing measured and simulated water contents during spring, quantitative values of the structural changes could be given. In a heavy clay soil the decrease in water content in spring because of structural changes was as high as 15 % by volume. During cold winters the crack system contributed to the infiltration of snow melt water, whereas during warmer winters the crack system often was filled with ice. Further development of the model was needed to account for the infiltration in the crack system.

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

MODELING THE EFFECTS OF SOIL FROST AND SNOWMELT ON RUNOFF AND EROSION

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 99-107
Spokane, Washington, March 21-22, 1990

The development of soil frost is the result of complex interactions of several factors, including soil characteristics, type of tillage and residue management, surface roughness, type of vegetative cover, duration and extent of freezing temperatures, and the extent and timing of snow cover. The freezing process itself modifies those soil physical properties that, along with temperature, determine the depth and duration of soil frost. The magnitude of soil changes that takes place as a result of soil freezing depends on freezing temperature, soil water content at freezing, initial soil aggregation, and the number of freeze-thaw cycles that take place over winter. As a result, land management combined with over winter frost action determines a soil's erodibility during winter thaw periods and from spring snowmelt to planting.

This paper describes a routine to model overwinter processes so that snowmelt runoff and changes in soil erodibility can be predicted. The routine is divided into three separate components which interact with each other on a daily basis. The components deal with soil frost formation, snowmelt, and snowdrifting. The frost component estimates the extent of frost development and thawing over the winter period as well as changes in soil water content and infiltration capacity. The snowmelt component estimates the amount of snowmelt occurring and how much snowmelt water is available for runoff in the spring. The snowdrift component estimates the depth, density, and distribution of snow cover over a watershed. Interaction of the three components provides a method to predict the effect of soil frost and snowmelt on runoff and soil erosion.

SUBJECT INDEX

1. Frost
2. Snowmelt
3. Soil erodibility
4. Erosion
5. Snowdrift

AUTHOR INDEX

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Young, et al. (cont'd)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

CONSERVATION APPLICATION IMPACTED BY SOIL FREEZE-THAW

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 108-112
Spokane, Washington, March 21-22, 1990

Soil freeze-thaw impacts about half the agricultural land in the United States of America. Estimates of land use are made for impacted area in land resource regions of the United States. Much information on climate and potential frost action for soils is contained in local soil surveys published by the National Cooperative Soil Survey. Although the effects of soil freeze-thaw are generally known, they are quantifiable for only a few locations and practices. A need exists to quantify the effects for design of individual conservation practice application in conservation systems.

SUBJECT INDEX

1. Frozen soil
2. Land use
3. Conservation application
4. Land resource region

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

HEAT AND WATER FLUX IN A DIURNALLY FREEZING AND THAWING SOIL

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 113-119
Spokane, Washington, March 21-22, 1990

This paper reports on field experiments and simulation of water redistribution and temperature profiles during diurnal freezing and thawing of a silt loam soil. Depth of soil freezing, water content, temperature, and hydraulic potential, of unfrozen soil, were measured on treatments having a bare surface (B) and residue covered surface (C). Soil on the B treatment froze to about 1.5 cm each night during 7 consecutive diurnal freezing and thawing cycles. Soil on the C treatment did not freeze. A finite difference numerical model was used to simulate soil temperature, depth of freezing, and water movement during 7 hr of freezing followed by 12 hr of thawing on the B treatment. Experimental results show that surface cover provides thermal insulation of the soil thereby decreasing the incidence of soil freezing. Simulation results support the use of this modeling approach for diurnal simulations of heat and water flux, near the surface, in freezing and thawing soil.

SUBJECT INDEX

1. Numerical model
2. Hydraulic potential

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

INFLUENCE OF MANAGEMENT PRACTICES ON SNOWMELT RUNOFF

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 120-124
Spokane, Washington, March 21-22, 1990

Agricultural management practices can significantly alter the amount of snowmelt water that infiltrates and thus the amount that appears as runoff. Fall cultivation is a common practice on agricultural fields to reduce the springtime workload and dramatically alters the infiltration characteristics of the soil surface. Research plots were established in an agricultural field near Grande Prairie, Alberta to quantify snowmelt runoff from fall cultivated, stubbled, and grassed plots. Runoff flow rates were measured during two spring melt events and total volumes of runoff determined. Fall cultivation did not appear to reduce the snow water equivalent. Highest runoff was from the fescue plots and lowest from the fallow (fall cultivated). Total runoff was significantly correlated with snow water equivalent and fall cultivation. Between 70 and 90% of the snow water equivalent occurred as runoff each spring. Fall cultivation reduced runoff and thus may increase soil water content.

SUBJECT INDEX

1. Frozen soil
2. Snow
3. Management practices
4. Runoff
5. Sediment yield

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

ENVIRONMENTAL CONDITIONS AND PROCESSES ASSOCIATED WITH RUNOFF FROM FROZEN SOILS AT REYNOLDS CREEK WATERSHED

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 125-134
Spokane, Washington, March 21-22, 1990

Frozen soil conditions have been associated with severe flooding and erosion events in the Pacific Northwest. Instrumented subwatersheds within the Reynolds Creek watershed representing different elevations were used to investigate the distribution, frequency and associated conditions of runoff from frozen soils. Frozen soil related runoff occurred in event-specific elevational bands. The environmental conditions proscribing those bands include frozen soil, shallow (<35 cm) snow cover, and rainfall. These conditions were found most frequently at intermediate elevations where low sagebrush was the dominant vegetation. Soil water content data indicate that infiltration occurs during frozen soil related runoff. This infiltration may result from thawing during events or infiltration through the frozen soil itself, both of which may vary spatially due to soil and snow variability.

SUBJECT INDEX

1. Frozen soil
2. Snow
3. Runoff
4. Flooding

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

THE EFFECT OF FROZEN SOIL ON EROSION-A MODEL APPROACH

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 135-144
Spokane, Washington, March 21-22, 1990

Nutrient losses and erosion are the main causes of eutrophic fresh waters in agricultural areas in Norway. A model was developed to predict non-point source pollution and to evaluate changes in agricultural management. The model presented is a combination of the hydrology model SOIL, and the erosion module from the CREAMS model. The model was applicated on two fields with clay soils. Simulation results were satisfactory, but showed that the situation with freezing and thawing has to be improved. During snowmelt a high conductivity was requested to simulate a macro-pore system. The effect of frozen soil on the infiltration of meltwater depends on the temperature regime during the foregoing winter. Soil erodibility has to be related to both temperature and water content in the top soil.

SUBJECT INDEX

1. Frozen soil
2. Snow
3. Erosion model
4. Hydrologic modeling

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

EFFECT OF FREEZE-THAW CYCLES ON THE PERMEABILITY AND MACROSTRUCTURE OF SOILS

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 145-155
Spokane, Washington, March 21-22, 1990

The effect of freeze-thaw cycling on the permeability and macrostructure of fine-grained soil was investigated in the laboratory. Five soils of different texture and index properties were compacted in a permeameter and subjected to freeze-thaw cycling. The temperature gradient and freezing rate were automatically controlled using a computer programmed power supply and data logging system. Soil permeabilities were measured prior to and after freezing and thawing using a falling head permeability device.

The permeability of compacted clayey soil is increased by factors higher than two orders of magnitude by freeze-thaw cycling. Changes in soil permeability are associated with changes in the soil macro- and micro-structure. Macro-cracks were observed with a standard microscope and micro pore size changes were observed with a scanning electron microscope (SEM). The smallest changes in permeability occur when clayey soils are compacted to densities where the saturated moisture content is equivalent to the plastic limit water content. An initial liquidity index of less than one is necessary to prevent detrimental permeability increases.

SUBJECT INDEX

1. Clay
2. Freeze-thaw
3. Permeability
4. Soil
5. Structure

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

INFILTRATION INTO A SEASONALLY FROZEN CLAY SOIL

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 156-160
Spokane, Washington, March 21-22, 1990

The infiltration capacity of thawing soil influences many processes and properties. Low infiltration capacity may cause surface runoff leading to severe loss of plant nutrients. It may also cause persistent high water contents in the superficial soil layers, leading to a deterioration of soil structure. A study was made on the effects of the amount of ice in the soil on infiltration rate in a seasonally frozen clay soil. Measurements of soil temperature and total water content down to a depth of 1 m and infiltration rate in small field plots were analyzed using a numerical model. Different amounts of water were supplied to the plots before freezing and during winter. Infiltration rate was subsequently measured during the thawing period. Maximum infiltration rates were as high as 8 mm/min, which could be interpreted as water flow in the crack system. The infiltration rate was related to the soil temperature, which points at an inverse relationship between ice content and infiltration rate. Plots with a

SUBJECT INDEX

1. Infiltration rate
2. Water content
3. Ice content
4. Soil temperature
5. Soil structure
6. Coupled water
7. Heat flow model

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

SNOWMELT INFILTRATION INTO COMPLETELY-FROZEN, SUBSOILED SOILS

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 161-170
Spokane, Washington, March 21-22, 1990

Altering the macropore content of a "completely-frozen" soil-a soil whose frozen profile does not thaw to appreciable depth during snowcover ablation-greatly improves its ability to absorb meltwater. Infiltration into soil cracks and rips averaged 3.8 and 5.8 times the amounts into the same soils in uncracked or undisturbed condition. A model for estimating the depth of infiltration into subsoiled fields using line spacing. Snowcover water and soil moisture (ice) content is presented. Subsoiling to depths between 400 and 500 mm on spacings between 0.9 and 1.3 m should enhance the infiltration capacities of frozen soils in the semi-arid prairie region so that they are capable of absorbing most of the snow water accumulated by stubble management practices in a normal snow year.

SUBJECT INDEX

1. Snowmelt infiltration
2. Frozen soils
3. Subsoiling
4. Model
5. Yield (wheat)

AUTHOR INDEX

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Over a 4-yr period. increases in the annual yield of spring wheat from subsoiled areas over those from undisturbed plots of continuous stubble averaged 1033 kg/ha on plots with snow management and 240 kg/ha on plots without snow management. The useful life of the treatment effect is estimated of the order of 4 or 5 years.

Gray, et al. (cont'd)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

CROP MANAGEMENT EFFECTS ON RUNOFF AND SOIL LOSS

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 171-176
Spokane, Washington, March 21-22, 1990

Frozen and thawing soil runoff events account for a significant portion of the runoff and erosion in the Palouse region of the Pacific Northwest. Crop management can significantly affect soil loss from frozen and thawing soil, as shown by ten years of data collected from runoff plots with different crop management near Pullman, Washington. Runoff events were segregated into frozen and thawing and non-frozen and treatment comparisons were made. The data showed a significant impact on both runoff and erosion for events when there was frozen soil, or if the soil was only frozen to a shallow depth and thawed rapidly For runoff events, when the soil was deeply and impermeably frozen, crop management had little effect on runoff but retained a major impact on erosion. These results are important for successful application of erosion models in meeting soil loss tolerance requirements of the 1985 farm bill.

SUBJECT INDEX

1. Crop
2. Tillage
3. Frozen soil
4. Runoff
5. Erosion

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

EFFECT OF FREEZING ON MASS AND HEAT TRANSFER IN POROUS MEDIA

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 177-185
Spokane, Washington, March 21-22, 1990

Frost action in soils results from lowering the soil water temperature below its freezing point. The study investigated the effect of artificial salting of the ground on the mass flow and heat transfer through a frost susceptible soil.

Open-system unidirectional freezing tests on 5x15 cm specimens simulated actual field conditions in a laboratory environment. The tests were conducted under a temperature gradient of -10°C air temperature and +4°C bottom temperature of specimens. The effect of the type of cation and its concentration in the pore fluid on the temperature profile, depth of freezing, moisture transfer, and magnitude of frost heaving was studied.

The presence of salts reduced the rate and magnitude of frost heave and the depth of frost penetration. The effectiveness of the treatment was inversely proportional to the concentration of the salt ion in the pore water.

SUBJECT INDEX

1. Artificial salting
2. Salts as frost modifiers
3. Soil additives
4. Moisture movement in frozen soil
5. Frost heaving mechanism
6. Frozen soil\
7. Soil freezing
8. Moisture transfer
9. Moisture freezing
10. Thermal properties of soil

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

APPLICATION OF TIME DOMAIN REFLECTOMETRY TO MEASURE SOLUTE REDISTRIBUTION DURING SOIL FREEZING

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 186-194
Spokane, Washington, March 21-22, 1990

The redistribution of solutes during soil freezing was investigated in the laboratory. Uniaxial freezing tests were performed on twelve columns of Guelph loam. The columns differed in average bulk density and concentration of solutes added. Volumetric (unfrozen) water content and bulk electrical conductivity (BEC) were measured in situ using time domain reflectometry (TDR). Temperature profiles were also monitored. A regression model was developed to explain BEC in terms of solute concentration, unfrozen water content and temperature; the coefficient of determination (R^2) was 0.985. The relationship between BEC and unfrozen water content was nonlinear. The temperature effect could be related to the viscosity of liquid water. After correcting for unfrozen water content and temperature, residual

SUBJECT INDEX

1. Frozen soil
2. Solute redistribution
3. Bulk electrical
4. Conductivity
5. Unfrozen water content
6. Time-domain reflectometry

AUTHOR INDEX

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van Loon, et al.
(cont'd)

peaks in BEC were assumed to indicate an increase in solute concentration. The peaks advanced downwards with a velocity of 6 mm/day ahead of the 0°C isotherm, suggesting redistribution of solutes as a result of freezing.

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

MODELING OF SOLUTE REJECTION IN FREEZING SOILS

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 195-201
Spokane, Washington, March 21-22, 1990

Quantification of the transport and concentrations of solutes dissolved in groundwater is necessary to determine the extent of a contamination problem. Soil water quality is an important factor affecting plant uptake of water, as well as other agricultural processes. Complex processes are incorporated into the soil system, during the freezing of soil water. The solute is excluded from the freezing water to the unfrozen liquid phase. The increase in solute concentration thereof, depresses the freezing point of the soil solution further, hence affecting the amount of unfrozen water remaining at a given subzero temperature. Furthermore, flow of unfrozen water is induced by temperature gradients towards the freezing zone. To estimate these processes quantitatively, a system of governing equations describing the physical processes is developed.

First, an energy equation is developed, which incorporates the heat convected and conducted into the system to raise the internal energy of the system. This includes temperature changes and the latent heat of phase change. A mass balance equation for liquid water states that the change in water content is due to flow and the phase changes, while a mass balance equation for dissolved solutes determines the solute concentration changes. Thermodynamic equilibrium relations determine the freezing point of the soil solution, in the presence of solutes. Capillary pressures are determined using moisture retention curves, while the gradient in pore water pressure is used as the driving force for unfrozen water flow. With boundary and initial conditions prescribed, the system of equations is complete and a solution is sought by a numerical technique.

SUBJECT INDEX

1. Freezing soils
2. Solute rejection
3. Modeling
4. Heat transfer

AUTHOR INDEX

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Pandy and Corapcioglu
(cont'd)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

FATE AND TRANSPORT OF CONTAMINANTS IN FROZEN SOILS

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 202-211
Spokane, Washington, March 21-22, 1990

In order to understand and predict any contaminant fate and transport in frozen soils, measurements of physical and chemical solute transport properties of soil/ contaminant systems are necessary. Some of these properties, such as the sorption diffusion coefficient and effective porosity, can be determined from breakthrough curves (BTC) generated in the soil column studies. BTC analysis was conducted as part of an investigation to evaluate the use of artificial freezing to move organic explosive compounds in Lebanon silt. Freezing is being explored at CRREL as an alternative innovative technique for soil decontamination and immobilization. Explosives studied included 2,6-dinitrotoluene (2,6-DNT), ortho-nitrotoluene (O-NT), and meta-nitrotoluene (M-NT) which represent some of the residues commonly found at U.S. Army ammunition plants.

On the basis of the laboratory column studies, the breakthrough curve characteristics appeared to be dependent on the type of the explosives. When Lebanon silt contaminated with these explosives was frozen halfway from bottom up at an average freezing rate of 0.5 cm/sec, statistically significant movement was observed for O-NT and M-NT but not for 2,6-DNT. For one freeze cycle, about 40% reduction in contaminant concentration was measured for O-NT and M-NT and less than 20% for 2,6-DNT within the frozen zone. As a result of this reduction in the frozen zone, an increase in concentration just above the freezing front was expected but not observed. Soil spatial variability and contaminant losses due to sorption and/or biodegradation were assumed to be responsible for not observing contaminant buildup ahead of the freezing front. Mass balance was not obtained in this experiment. Hence, factors, such as inherent soil variability and physical/ chemical processes, are not yet fully understood. They represent some of the sampling problems encountered in the use of artificial freezing as a potential soil decontamination method for explosive organic compounds.

SUBJECT INDEX

1. Contaminants
2. Chemical contamination
3. Soils
4. Solutes
5. Solubility
6. Freezing
7. Explosives

AUTHOR INDEX

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Ayorinde and Perry
(cont'd)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

EFFECTS OF FREEZING ON SULFATE SALTS IN NORTH DAKOTA SOILS AND WETLANDS

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 212-215
Spokane, Washington, March 21-22, 1990

Sulfate minerals frequently contain waters of hydration in their crystal structure. The hydrogen bonds associated with these waters of hydration are exceedingly temperature dependent; therefore, differential solubility occurs in saline, high sulfate soils regards their mineral composition. The mineral precipitation sequence that occurs on or near freezing segregates ions in the soil profile. A similar segregation is noted in pond water in wetlands. Mirabilite ($Na_2SO_4 \cdot 10H_2O$) in particular changes solubility with temperature. The effect of freezing concentrates the soil solution and may approach the levels that precipitate mirabilite. This mineral is common in sublimated salts on brackish and saline wetlands. Magnesium minerals, however, do not form as readily. The result is that magnesium is concentrated in the soil solution or in pond water during freezing.

We noted also that gypsum ($CaSO_4 \cdot 2H_2O$) tended to precipitate out in pond ice during concentration by freezing. The effect is expected to occur also on pond edges in soils. The expected interaction between calcite edges in soils. The expected interaction between cal-

SUBJECT INDEX

1. Frozen soil
2. Sulfates
3. Gypsum
4. Thenardite
5. Espomite

AUTHOR INDEX

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Richardson, et al. (cont'd)

cite and gypsum between summer and winter account for large pH swings between summer and winter observed in wetlands.

We hypothesize that in saline soils high in sulfate, the freezing effect tends to maximize the sodic influence; the result is that a natric or sodium affected soil can occur with lower amounts of sodium ions present than in an unfrozen soil or in a soil high in chloride salt.

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CRREL
Special Report 90-1
Catalogue No.

AN SCS PERSPECTIVE ON USING RESEARCH MODELS IN PLANNING AND APPLYING CONSERVATION MEASURES

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 216-219
Spokane, Washington, March 21-22, 1990

In the 1990s, powerful microcomputing capability will make possible onsite utilization of software based on research models in ways heretofore considered impractical. Such "real-time" analysis of soil and water conservation problems and opportunities will provide the Soil Conservation Service (SCS) and others an invaluable tool in providing better service to decision makers in protecting and enhancing the Nation's natural resources. Over the next decade, much attention will be focused on reducing damage caused by excessive soil erosion on rural lands. AB the provisions of the 1985 Food Security Act are carried out, up-to-date knowledge of planning soil and water treatment measures must be translated into effective application of conservation systems on about 140 million acres of highly erodible cropland. Much of this cropland is potentially subject to freezing and thawing. Therefore, knowledge of the complex phenomena associated with this aspect of erosion is an important element in planning these treatment measures. It is necessary, in translating basic scientific knowledge incorporated in research models into effective, efficient technological tools useful at the field level of user agencies, for

SUBJECT INDEX

1. Research models
2. Technological tools
3. Water erosion prediction project (WEPP)

AUTHOR INDEX

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Herndon (cont'd)

research organizations, academic institutions, and user agency personnel to work closely together. An excellent "template" for such interagency, interdisciplinary cooperation is emerging from work being done on the Water Erosion Prediction Project (WEPP), a project led by the Agricultural Research Service (ARS) which will eventually replace the Universal Soil Loss Equation (USLE). Cooperative efforts on this project may one day lead to use of this technology in fashioning conservation systems that better address the overall needs associated with complex agricultural ecosystem

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CRREL
Special Report 90-1
Catalogue No.

FROZEN SOIL IMPACTS: RESEARCH NEEDS

Proc. Int. Symposium, Frozen Soil Impacts on Agricultural, Range, and Forest Lands: Pgs. 220-223
Spokane, Washington, March 21-22, 1990

Frozen soil is a natural occurrence on a large percentage of the world's land areas which are under cultivation, in grasslands, and forested. Frozen soil research is needed to develop new knowledge and technologies that can provide management strategies for overcoming the adverse effects of freezing and thawing on production and the environment while enhancing its beneficial effects. Priorities of frozen soil research must be aligned with problem-solving research on existing and emerging major issues for the improved management and use of agricultural, range, and forested lands. Twelve of these high priority issues and their relation to frozen soil impacts are described. With the recent knowledge and predictability advances described in this symposium and the new directions for future research, frozen soil phenomena will rapidly become more clearly understood for further improvements in management and understanding of our worlds producing lands.

SUBJECT INDEX

1. Soil
2. Freezing
3. Research
4. Priorities
5. Advances
6. Production

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

RUNOFF AND EROSION DURING SIMULATED RAINFALL ON FROZEN FIELD PLOTS WITH DIFFERENT DEPTHS OF SURFACE THAW AND LEVEL OF ERODIBILITY

Proc. Int. Symposium, Frozen Soil Impacts on Agricultural, Range, and Forest Lands: Pgs. 224-230
Spokane, Washington, March 21-22, 1990

In northern climates, rainfall and snowmelt often occur on soils that are frozen or partially thawed. Runoff and erosion can be accelerated by these conditions, resulting in off-site pollution and loss in soil productivity. Erosion from frozen or partially thawed soil has long been recognized as a problem—one that has been poorly documented and quantified. Our work here documents the runoff and erosion that occur under simulated rainfall on three soils (Hattie clay, Barnes loam, and Sverdrup loamy sand) that had been fall plowed or fall chiseled and had corn residue left or removed before tillage. All plots were surveyed, established, and instrumented in the fall of each year. Initial random roughness, aggregate stability, and surface bulk density data were recorded for each treatment. Weekly determinations were made over winter of soil moisture (neutron method), soil temperature, frost depth, thaw depth, and snow depth. Simulated rainfall was applied to one plot of each tillage-residue combination at four successive stages of soil thawing in the spring. All soil measurements were repeated for each simulated rainfall event. In addition, time to first runoff, total runoff, and sediment yield were determined. Results indicate that applied water tends to infiltrate the soil and flow downslope on the surface of still frozen soil to reappear at the soil surface at downslope positions. Time to runoff was related to depth of thaw. Relationships observed concerning rainfall, runoff, sediment yield, tillage-residue treatment, depth of thaw, aggregate stability, surface roughness, and soil density are discussed.

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SUBJECT INDEX

1. Runoff
2. Erosion
3. Frozen soil
4. Thaw depth
5. Soil moisture
6. Soil density
7. Random roughness

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

SEED ZONE TEMPERATURE AND MOISTURE CONDITIONS IN A PARTIALLY FROZEN SOIL-CROP RESIDUE SYSTEM

Proc. Int. Symposium, Frozen Soil Impacts on Agricultural, Range, and Forest Lands: Pgs. 231-238
Spokane, Washington, March 21-22, 1990

A study was conducted to examine effects of tillage and crop residue management practices on seed zone moisture and temperature conditions during spring in interior Alaska. A model is presented to simulate heat and water transport in a partially frozen soil-crop residue system. Soil freezing and thawing are simulated. Model simulations were implemented and tested using data collected at the site. Seed zone temperature, soil water matric potential, and depth to frozen soil were measured and simulated for conventional-till and no-till plots. Both simulated and measured results indicated that the no-till soil was cooler and wetter, and thawed more slowly than the conventional-till soil. It is concluded that cool seed zone temperatures under no-till systems may be ameliorated by decreasing planting depth. The simulations also suggested that thawing soil at depth is a source to replenish water lost from upper soil layers through evaporation.

SUBJECT INDEX

1. Transport model
2. Frozen soil
3. Crop residue
4. Conservation tillage

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

TILLAGE AND RESIDUE MANAGEMENT SYSTEMS AFFECT ON WATER SOIL TEMPERAT RES

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 239-246
Spokane, Washington, March 21-22, 1990

The purpose of this study was to evaluate the effect of tillage and residue management systems on winter soil temperatures in the subarctic over a three year period. Maximum tillage plots with all straw and stubble removed after combining and no tillage plots with all straw and stubble remaining were instrumented with thermocouples at depths from 1 to 100 cm. Soils in interior Alaska generally have one to two freeze/thaw events before freezing with no subsequent thaw in early October. The 0°C isotherm penetrates the soil, reaching the active permafrost layer at a 2-m depth in late October. Results from tillage and residue management studies indicate that winter soil temperatures are higher under no tillage than maximum tillage. The beneficial effect of straw and stubble trapping snow in high wind areas, retarding heat loss from the soil during winter, was greater than its adverse effects preventing soil warming in the spring. Winter survival of small grains may be more prevalent for no-till plots since soil temperatures did not go below -20°C at crown depth as did temperatures in maximum-till plots.

SUBJECT INDEX

1. Frozen soil
2. Snow
3. Tillage
4. Residue management
5. Soil temperature

AUTHOR INDEX

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Cullum. et al.
(cont'd)

PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

THE IMPACT OF FROZEN SOIL ON PRAIRIE HYDROLOGY

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 247-256
Spokane, Washington, March 21-22, 1990

This paper examines: (a) the magnitudes of soil moisture changes occurring over winter due to moisture transfers at the soil surface and migration of water to the freezing front; (b) the infiltration characteristics of "completely-frozen" soils; and (c) the redistribution and disposition of soil water following ablation of the seasonal snowcover in a Prairie environment. These factors are discussed in respect to their effects on the movement and redistribution of soluble salts, the use of snow management practices for augmenting soil water reserves, the prediction of runoff rates and volumes from snow. and the amount of soil water available for crop growth.

SUBJECT INDEX

1. Frozen soil
2. Moisture migratior
3. Snowmelt infiltration
4. Evaporation

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

COMPARISON OF THREE METHODS FOR MEASURING DEPTH OF SOIL FREEZING

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 257-262
Spokane, Washington, March 21-22, 1990

Three systems were used to measure the depth of frozen soil on the Reynolds Creek Experimental Watershed in southwest Idaho. The systems include thermistors to measure soil temperature, gypsum soil-water-resistance blocks (gypsum blocks), and tubes filled with sand and a fluorescein dye solution (frost tubes). Frost tube readings were within 5 cm of the depth of the 0°C isotherm for most of the 1988-1989 winter season. The greatest differences between the 0°C isotherm and the frost tubes were during thawing periods when field observers were unable to determine the frost line because the color in the frost tube was not distinct. Gypsum block readings followed the same pattern of soil freezing and thawing as indicated by the other two measuring methods. However, the gypsum blocks tended to lag when the soil was freezing. The gypsum blocks indicated essentially the same maximum depth of frozen soil when all types of sensors indicated frozen soil conditions. Results from the study show that all three systems can be used in reconnaissance-type soil frost measuring networks.

SUBJECT INDEX

1. Hydrology
2. Frozen soil
3. Instrumentation
4. Frost

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

REDISTRIBUTION OF SOIL WATER AND SOLUTE IN FINE AND COARSE TEXTURED SOILS AFTER FREEZING

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 263-270
Spokane, Washington, March 21-22, 1990

Freezing unsaturated Fargo silty clay (Vertic Haplaquoll) and Hecla loamy fine sand (Aquic Haploborolls) in undisturbed columns in the presence of a simulated watertable induced migration of water and solutes towards the freezing front as it advanced. Water and electrical conductivity redistribution was inversely related in the silty clay and a fine sandy loam with a simulated watertable low in salts but directly proportional in these two soils when the simulated watertable was high in salts. Increases in solute concentration were evident in the frozen soil after prolonged freezing. Mg and Na were the most mobile cations and were correlated to increases in EC. Differences between soluble and exchangeable salts occurred with Na concentrating below the freezing front and Mg exchanging with Ca. The inherent physical and chemical spatial variability characteristics within the undisturbed soil columns resulted in the inability to detect significant differences between freezing depth treatment.

SUBJECT INDEX

1. Freezing soils
2. Solute movement

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

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SUBJECT INDEX

1. Freezing soils
2. Solute movement

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

FIELD AND LABORATORY TECHNIQUES FOR FROZEN AND THAWING SOIL

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 271-278
Spokane, Washington, March 21-22, 1990

Methodology and instrumentation for conducting research on frozen and thawing soils are presented. Both field and laboratory techniques to determine frost depth, soil strength, soil water content, soil matric potential, bulk density and methods for artificial freezing and thawing of soil samples are discussed. References will provide additional information.

SUBJECT INDEX

1. Frozen soil
2. Thawing soil
3. Measurement
4. Techniques

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

FROST DATA IN THE PACIFIC NORTHWEST

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 286-289
Spokane, Washington, March 21-22, 1990

Frost depth is not a commonly measured variable at climatological stations. Thus these data are usually collected using special purpose networks. In the Pacific Northwest, there are three types of frost depth data sets available. The first is a reconnaissance data set consisting of 608 station-years of data collected over a period of nine years and over four states. The second data set was gathered at NWS climatological stations in four states over a period of ten years resulting in 93 station-years of record. A third data set was collected for research purposes by the Agricultural Research Service. These research data include intensive measurements for hydrologic purposes and for runoff-erosion experiments. These data have been used for various projects are generally available for use by other researchers.

SUBJECT INDEX

1. Frost depth
2. Data sets
3. Networks

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

SOIL THAW AS INFLUENCED BY SMALL GRAIN RESIDUE COLOR

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 295-303
Spokane, Washington, March 21-22, 1990

Crop residues generally suppress heat and water flow between the soil and atmosphere. Residue management methods of conservation cropping systems are needed in subarctic regions that enhance global radiation absorption and heat movement into the soil. This report investigated the effect of residue color on global radiation absorption and soil thermal characteristics at Fairbanks, Alaska. Barley stubble and loose residue remaining after harvest on 36 m^2 plots was colored black, white, or remained unpainted (yellow). The plots were instrumented to monitor reflected global radiation and soil temperature and thaw. Black-colored residue absorbed 15 to 25% more radiation than yellow and white-colored residue. The greater energy absorption resulted in higher maximum soil temperatures (2°C at 5 cm) and greater thaw (1 cm) in plots with black residue. Differences in thaw depth were only evident immediately following snowmelt. A mechanical means of trapping the greater radiant energy on black residue may enhance soil thaw in the early season.

SUBJECT INDEX

1. Frozen soil
2. Tillage
3. Soil temperature
4. Albedo

AUTHOR INDEX

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PROC. SYMP. FROZEN SOIL IMPACTS

CRREL
Special Report 90-1
Catalogue No.

SOIL STRUCTURE AND FROST DEPTH AS AFFECTED BY SOIL COMPACTION

Proc. Int. Symposium, Frozen Soil Impacts on
Agricultural, Range, and Forest Lands: Pgs. 295-303
Spokane, Washington, March 21-22, 1990

Field research several decades ago suggested that annual freezing and thawing were sufficient to ameliorate seasonal compaction in the northern U.S. Recent field research in Minnesota conclusively shows that ordinary farm wheel traffic can cause soil compaction that persists for more than one year in spite of freezing soil temperatures throughout the tilled layer. Bulk density of the tilled layer was increased by about 0.3 $Mg.m^3$ by one season's wheel traffic. Freezing and thawing in the absence of fall tillage decreased the bulk density of the surface 10 cm by only 0.1 Mg/m^3. Traffic from heavy harvest equipment caused increased bulk density to a depth of 60 cm. Subsequent freezing and thawing have not ameliorated this deep compaction after seven years of annual freezing and thawing Depth and duration of frost penetration were not significantly affected by soil compaction, although there were definite trends within a year. Depth of snow cover was more important but was not uniform over the experimental site, which caused high variation in frost depth within compaction treatments.

SUBJECT INDEX

1. Wheel traffic
2. Bulk density
3. Snow cover
4. Plant residue
5. Freezing

AUTHOR INDEX

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U. S. GOVERNMENT PRINTING OFFICE 1990/700-057/02071