THE PRAIRIE SOIL MOISTURE REGIME:
FALL TO SEEDING

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ABSTRACT

The paper focuses on the moisture regime of Prairie soils in the period from freeze-up in the fall to the time of seeding of annual cereal crops in the spring. It is shown that significant soil moisture changes can occur over winter -- the period extending from freeze-up to a time immediately preceding snowmelt -- because of moisture transfer across the soil-snow-air interfaces and migration of water to the freezing front. Field data on the magnitudes of, and factors affecting these changes are presented. The significance of the changes as they may affect the snowmelt infiltration potential of a frozen soil; the validity of "fall" soil moisture measurements for indexing "spring" soil water reserves, the amount of available root-zone water, fall irrigation practices and the upward migration of salts are discussed.

The problem of snowmelt infiltration to frozen soils is reviewed. It is suggested that for practical applications, frozen soils may be grouped into three general classes according to their infiltration potential, namely: Restricted - impervious; Unlimited - capable of infiltrating the snowcover water equivalent, and Limited - infiltration is governed by the snowcover water equivalent, and the moisture (ice) content of the 0-300 mm soil layer at the time of melt. An empirical relationship describing the amount of infiltration for the Limited case is presented. The implications of the results as they affect the application of snow management practices for increasing soil moisture reserves are discussed. It is suggested that to derive maximum benefit from the practices (in terms of soil water augmentation on soils of Limited potential), it is necessary to apply them in combination with a tillage or subsoiling practice that increases the macropore content of
the soil. Field data illustrating increased infiltration in cracked and subsoiled soils are listed.

Factors affecting the redistribution and disposition of water within a soil profile in the period from the end of snowmelt to the time of seeding (post-melt) are reviewed. Data are provided which demonstrate the relative effects of the net winter change in soil moisture (i.e., overwinter change plus snowmelt infiltration) and drainage, evapotranspiration and precipitation during the post-melt period (period following disappearance of the seasonal snowcover) on the availability of water within the root zone at the time of seeding of cereal grains.

INTRODUCTION

General

The importance of water to the production of cereal grains under dryland farming is well-recognized. For the Canadian Prairies, the Saskatchewan Advisory Council on Soils and Agronomy (1982) recommends that root-zone water reserves of 125 mm and 100 mm are needed in the Brown and Dark Brown soil zones, respectively, at the time of seeding to achieve average yields. The recommendation assumes normal, or close to normal rainfall amounts throughout the growing period. Staple and Lehane (1952, 1954) and Staple et al. (1960) have suggested that soil water additions above certain base levels increase yield in varying amounts; e.g., each 25 mm of water above a moisture reserve of 262 mm to some variable limit may increase wheat yields from 230-400 kg/ha. Similarly, de Jong and Rennie (1969) reported increases in the yield of spring wheat in the range of 200-275 kg/ha for each additional 25 mm of water above the long-term normal precipitation for the relatively-humid, east-central region of Saskatchewan. Suffice to say the results of research and practical experience evince the strong association
between yield and available water, thereby supporting the need for efficient water utilization and management practices.

Numerous studies have been conducted on consumptive use and evapotranspiration requirements and root-zone soil moisture withdrawal patterns by cereal grains during the growing period. Conversely, there has been a definite lack in research effort directed to the study of the soil moisture regime during the winter months. The seasonal difference in level of research effort is understandable given the inclement weather conditions under which both equipment and humans must operate during a Prairie winter. Further, there appears to be a general impression that moisture conditions remain relatively static over winter and the effects of the freeze-thaw sequence on the soil moisture status at the time of seeding are inconsequential, with the exception of the contributions by snowmelt infiltration. It is one purpose of this paper to demonstrate that the moisture regime of a Prairie soil can be extremely active during the winter months. Significant moisture exchanges, which may directly affect the amount of water available for crop growth, the snowmelt infiltration potential and other factors important to water conservation, land use management and farming practices, may occur.

The mean annual precipitation throughout a large part of the grain-producing area of the 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 use. Recently, there has been renewed interest in the potential of managing snow to increase soil water reserves. The basic premise underlying the use of any snow management
practice for increasing soil moisture is that an increase in snow water, the result of an increase in depth, density or both of these snowcover properties, will result in an increase in snowmelt infiltration. On flat areas, when the entry of meltwater to a soil is not limited and surface runoff can be neglected, the association between these two variables may be close to linear. The relationship can be expected to depart significantly from this trend on sloping terrain; when a late-lying snow (which can be related to a deep snowcover) incurs large evaporation losses because of increases in the net radiation flux, due to an increase in solar radiation and a decrease in albedo, and in the flux of advective energy from adjacent areas of bare ground; when a snowcover melts rapidly and increases surface runoff; and when the entry to and the downward movement of the meltwater in the soil is restricted, for example, as may result from refreezing of meltwater on the soil surface or at a shallow depth.

The role of snow as a water resource in agricultural production is not well understood. In an attempt to clarify certain aspects of the snow/soil-water interaction, the material presented below emphasizes the infiltration process to frozen Prairie soils; the relation between snowcover accumulation and snowmelt infiltration and the application of snow management practices for increasing soil water reserves in both uncracked and cracked soils.

The final question that must be addressed in a study of the moisture regime between freeze-up and seeding is the amount of the net winter change (overwinter moisture change plus snowmelt infiltration) which remains within the root zone and is available at seeding to support crop growth. The amount depends on the losses in the period due to evapotranspiration and drainage. Of course, these can be offset by infiltration from precipitation in the interval. Consideration is
given herein to the factors affecting the disposition of soil water gained by snowmelt infiltration following the disappearance of the seasonal snowcover and to those affecting the availability of root zone water at the time of seeding of cereal grains.

The discussions of soil moisture are presented as a time sequence covering three separate periods: Overwinter - the time from freeze-up in the fall to the time preceding the active melt of the snowcover and thawing of soil in the spring; Snowmelt - that period encompassing the ablation or melt of the snowcover in the spring, and Post-melt - the period extending from the disappearance of the seasonal snowcover to the time of seeding.

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 from 1978-84 in the Brown and Dark Brown soil zones of Saskatchewan; near Asquith, Bad Lake, Flaxcombe, Kerrobert, Outlook, Portreeve, Richlea and Saskatoon (see Fig. 1). The sites represent a range of soil textures (fine sand; -80% sand) to heavy clay (-63% clay), land use practices (fallow, grass and stubble) and climatic conditions encountered on the arable farm land of the Canadian Prairies. At Outlook, two sites in grass and two in stubble were located in a field under border dyke irrigation.

At most sites soil moisture content was monitored with a two probe gamma density meter. This system provides non-destructive sampling of the density of a soil volume approximately 50 mm wide, 250 mm long and 20 mm thick. By assuming the mass of the soil cube remains constant, i.e., no major structural changes occur, changes in density of the cube can be attributed to changes in the mass of water contained within the
Figure 1. Location of study sites within the Brown and Dark Brown soil zones of Saskatchewan.

The equivalent moisture change is calculated from the density readings assuming a density of water equal to 1000 kg/m$^3$.

Measurements were obtained at 20 mm increments of depth to 1 m and at 40 mm increments between 1 m and 1.6 m. Repeatability tests conducted with the equipment in the field showed a standard error of estimate in moisture content of about ±2.5 mm in a 1-m profile. All systems were extensively tested and calibrated to operate reliably under cold weather conditions (to ~ -20°C).

Soil moisture changes were also monitored at some sites with a neutron probe. With this equipment readings are taken at increments of
depth of 150 mm. Profile measurements were made in the fall prior to freeze-up, and continued throughout the winter, snowmelt and post-melt periods up to a date close to the time of seeding. Every effort was made to service each site at least once every three weeks.

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 the layers of soil where overwinter moisture changes occurred, periods of midwinter melt and infiltration, the times of freeze-up, snowmelt and thaw and to establish rates and depths of freezing and thawing.

Measurements were also made of the depth and density of the snowcover. Where possible these were taken throughout the accumulation period up to the time of active snowmelt. Unfortunately, at some sites, because of logistical problems, it was impossible to monitor these parameters on a regular basis. For these sites, when a snowfall event occurred after the snow survey, the "on-site" measurements were updated or revised using Nipher gauge readings recorded at a nearby climatological station and an assumed snow retention coefficient (see Gray et al., 1979). The writers are aware of the inaccuracies which may accrue in the estimate of snowcover water equivalent by this calculation because of the spatial variability in snowfall. In most cases, however, the adjustments to the data were small relative to the total snowcover water equivalent. Further problems arise in estimating water equivalent due to mixed rain and snow events, high wind speeds and snow transport (erosion and accumulation), condensation gains and sublimation losses, and the lateral flow of meltwater through the snowcover.

For additional details on different aspects of the experimental programs, for example, measurement procedures and equipment, the analysis
and interpretation of data, discussion of results and others, the reader is referred to the works of Granger et al. (1984) and Gray et al. (1984a; 1984b and 1984c).

OVERWINTER PERIOD

Moisture migrates in response to a temperature gradient during the freezing and thawing of unsaturated soils. The process is complex and involves the thermodynamics of the soil-ice-water regime. Numerous studies of the phenomenon are reported in the literature and these generally focus on: (a) the thermophysical properties of frozen porous soils; (b) the thermodynamics of the soil regime; (c) the effects of soil and water properties on moisture migration; and (d) the development and verification of models describing migration. Despite all the investigations, several questions on the fundamental aspects of the process remain unanswered, in particular a clear consensus is lacking on the relative importance of the different modes of moisture transfer, e.g., as a liquid or vapor.

Not unlike most investigations in earth sciences, the number of laboratory studies on freezing, frozen or thawing soils undertaken on small specimens in the laboratory vastly outnumber the field experiments. Because of the scarcity of field data and the difficulty in simulating field boundary conditions in laboratory experiments, only general agreement in the trends of the data from the two sources is evident.

Soil Moisture Changes

A general review of the field measurements of overwinter soil moisture changes indicated they can be discussed most conveniently by dividing the soil profile into two zones, herein referred to as an "Upper" zone and a "Lower" zone. The "Upper" zone is that soil layer extending from the surface to a depth of approximately 300 mm. In this
zone overwinter moisture transfer (other than infiltration) occurs primarily as a vapor and the changes in moisture content are greatly affected by those factors influencing the energy and vapor exchange processes within the soil and at the air/snow and snow/soil interfaces. Observations indicate that moisture losses from this zone are common. The "Lower" zone, located immediately below the "Upper", undergoes moisture changes by the upward migration or relocation of water in response to freezing or by drainage below the freezing front. Moisture may move either as a vapor or a liquid, and overwinter gains are common.

Figures 2 and 3 show typical plots of soil temperature and changes in soil moisture (includes water plus ice) regimes at selected dates throughout the period from freeze-up in the fall of 1982 through to the spring of 1983 measured in two different stubble fields. The Saskatoon site (Fig. 2) was a silty clay soil having a high fall moisture content (range 24-45% by volume) in which the depth to the water table was greater than 4 m. The soil moisture changes shown represent unusually high fluxes for the land use and are primarily the result of the high "fall" soil moisture. The Outlook site (Fig. 3) was 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. It is important to point out that the moisture fluxes shown for Outlook typify those that have been found during the past five years of measurement.

Figures 2 and 3 show:

1. Both sites exhibited a moisture loss from the Upper zone (0-200 mm) during the overwinter period. The losses were also evinced by the coarse, rounded, large-grained texture of the snowcovers at the soil surface - an indication of strong temperature-gradient metamorphism as a result of the flow of heat and moisture from the soil to the snowcover. Benson and Trabant (1973), Peck (1974) and Santeford (1976)
Figure 2. Soil temperature and soil moisture changes in a silty clay soil measured at Saskatoon, Sask. in the period from Nov. 16/82 to: (a) Dec. 2/82; (b) Dec. 17/82; (c) Jan. 13/83; (d) Feb. 8/83; (e) Mar. 8/83; (f) Mar. 29/83; (g) Apr. 18/83 and (h) May 31/83.
Figure 3. Soil temperature and soil moisture changes in a fine sandy loam soil measured at Outlook, Sask. in the period from Nov. 17/82 to: (a) Nov. 29/82; (b) Dec. 15/82; (c) Dec. 30/82; (d) Jan. 19/83; (e) Feb. 10/83; (f) Mar. 28/83; (g) Apr. 15/83 and (h) May 16/83. \( V( \) ) refers to the depth to the water table in metres; depth on Nov. 17/82 was 2.55 m (V2.55).
have recognized the coupling between moisture migration in a soil and an overlying snowcover as a result of thermal gradients and Santeford (1978) has reported moisture flow from a moss to a snowcover in the black spruce/permafrost region of interior Alaska.

2. Large fluxes of water migrate to the frozen Lower zone in response to the freezing action. Within a 1-m depth, these increases amounted to -47.8 mm at Saskatoon and -95.7 mm at Outlook. It can be observed that: (a) the accumulation of water/ice occurs immediately above the freezing front and (b) the elongation and extension of the zone-of-accumulation coincide with the downward movement of the 0°C isotherm into the soil.

At Outlook, as the rate of downward movement of the 0°C isotherm decreased, a zone of moisture loss developed in the unfrozen soil (Fig. 3d and 3c). This is attributed to the natural drainage of water, which could include water in transit to the frost front. The increase in the depth of the water table during freezing (Fig. 3a, 3d and 3e) and its reversal in the spring (Fig. 3h) suggests that the saturated zone is the primary source of the migrating water.

3. No pronounced soil moisture gradient in the unfrozen soil in the Lower zone. The absence of a hydraulic gradient below the freezing front suggests moisture transfer as a vapor; or if capillary (liquid) flow, the transport process is analogous to a coupled hand-to-hand exchange process.

4. A frozen depth in the silty clay soil at Saskatoon substantially greater than that in the sandy soil at Outlook. The accumulated degree-days of frost, calculated from measurements of soil temperature at 200 mm, was -300°C-days at each location after 80 days of freezing (up to ~Feb. 10/83). Assuming the index adequately reflects the freezing potential, the shallower depth-of-freezing at Outlook (1.1-1.2 m)
compared to Saskatoon (>1.6 m) can be attributed to the larger amount of water frozen.

The results demonstrate that significant fluxes of water may occur to the freezing front where conditions favor migration. They are in agreement with those 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 dry unsaturated soils.

Lacking fall rains, soils under stubble are often relatively dry at freeze-up. A soil, frozen in this condition can develop a moisture profile in the Lower zone which exhibits alternating layers of gains and losses throughout its frozen depth, but with little change in the total moisture content. Measurements of soil moisture changes made at Saskatoon in 1982/83 illustrate this condition (see Fig. 4). Figure 4a shows a zone of low moisture between 300-800 mm at freezeup; the result of withdrawal by the crop during the growing season. Figure 4b shows alternating zones of approximately equal gains and losses. Applying a moisture balance to the frozen zone (170-1520 mm) the net change was calculated as 1 mm (which is within the error of measurement of the moisture measurement equipment). The findings demonstrate the net transfer of water in a dry soil (excluding the surface layer) by freezing is negligible. Note, the Upper zone (0-170 mm) showed a loss of 26 mm in the period Nov. 18/82 - Feb. 8/83.

Average Statistics

Table 1 gives some average statistics of overwinter moisture changes measured at fifty uncracked sites under dryland farming conditions in the years 1980-1983 inclusive. These data demonstrate a wide range in values. The magnitudes of the changes in the soil layer (0-
Figure 4. (a) Initial soil moisture profile on Nov. 18/82 and (b) Soil moisture change during the period Nov. 18/82 - Feb. 8/83 measured in a relatively-dry silty clay soil at Saskatoon.

300 mm) of -11.1 to 5% by volume are greater than those in the 1-m depth of the root zone, i.e., -3.4 to 4.3% by volume. The differences in the overwinter changes between land use practices can be assumed to be due to differences in snowcover and soil moisture and temperature regimes.

A significant value in terms of the total winter change (overwinter soil moisture change plus snowmelt infiltration) is the ratio of snowmelt infiltration (INF) to the total change in moisture in a 1-m profile for the period from freeze-up to the end of melt (TC). The mean value of INF/TC for all sites was 0.79; for those sites where migration could be identified it was 0.50, that is 50% of the total winter change was due to migration.
Table 1. Average overwinter moisture changes in uncracked silty clay and clay soils under dryland farming (mm).

<table>
<thead>
<tr>
<th>ITEM</th>
<th>DEPTH INCREMENT</th>
<th>0-300 mm</th>
<th>0-1000 mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Range Fallow</td>
<td>-27.2 + 11.8</td>
<td>-16.7 + 37.4</td>
<td></td>
</tr>
<tr>
<td>Range Stubble</td>
<td>-33.3 + 15.1</td>
<td>-33.7 + 43.3</td>
<td></td>
</tr>
<tr>
<td>Mean Fallow</td>
<td>-7.5</td>
<td>10.6</td>
<td></td>
</tr>
<tr>
<td>Mean Stubble</td>
<td>1.3</td>
<td>11.5</td>
<td></td>
</tr>
<tr>
<td>INF/TC</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Average b)</td>
<td></td>
<td>0.79</td>
<td></td>
</tr>
<tr>
<td>Migration c)</td>
<td></td>
<td>0.50</td>
<td></td>
</tr>
</tbody>
</table>

a) INF/TC = ratio of the amount of snowmelt infiltration (INF) to the total change in soil moisture in a 1-m profile in the period freeze-up to the end of snowmelt (TC).

b) Average = mean value calculated using data from all sites.

c) Migration = mean value for those sites where moisture migration could be identified.

Practical Applications

A few practical applications of the results which are of direct interest to agriculturalists are:

1. Because the snowmelt infiltration potential of a frozen Prairie soil is inversely related to the soil moisture (ice) content of the soil layer 0-300 mm at the time of melt (see Granger et al., 1984 and later discussions), overwinter soil moisture changes directly affect the amount of soil water recharge from a snowcover.

2. The large fluxes of moisture due to freezing in lighter-textured soils that contain 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 retard thawing in the spring.
3. The amount of water that migrates in response to freezing and is retained within the crop root zone depends on the moisture content and drainage properties of the soil. It would appear that established crops, such as biennials or perennials, would always benefit to some extent from these additions, particularly in the early part of the growing season.

4. The soil moisture change calculated as the difference in fall and spring moisture content may not accurately reflect snowmelt infiltration.

5. The overwinter change in the moisture regime of a soil must be included in any model designed for forecasting spring soil water reserves that is based on the fall soil moisture, snowfall and snowcover water equivalent.

Liquid or Vapor Transfer

A special concern of agriculturalists is whether water migrating to a freezing front contains salts which may be deposited and subsequently move to the soil surface in response to evaporation. This depends on the mode of transfer of the water, i.e., as a liquid or a vapor. Obviously if moisture moves primarily as a vapor, it would be impossible to rationalize a build-up of cations which would contribute to soil salinity.

As pointed out in an earlier section, there is general consensus that the migrating water moves as a liquid. However, this consensus is not without controversy. For example, only recently Gray et al. (1984a) compared the amount of soil water frozen with the degree-days of frost at Outlook and Saskatoon in 1980/81, a winter when snowcover conditions were comparable at the two sites. The analyses made use of the fact that the latent heat for freezing of water is 333.4 kJ/kg compared with 2708.7 kJ/kg for water vapor. Water frozen in place in a soil changes
directly from liquid to ice; water migrating to a freezing front may change from liquid or vapor to ice, depending on the mode of transport. On the assumption the degree days of frost is an accurate index of the energy used for freezing, the plot of frozen water versus degree days of frost should be linear (straight line) and the amounts of water frozen at the same value of the freezing index at Saskatoon and Outlook should be equal when all the soil water is frozen as a liquid or the relative proportions of liquid and vapor have been defined correctly. The results of the analyses suggested that in 1980/81 upward movement occurred principally as a vapor. Similar calculations on data for the same year at other sites were less clear insofar as identifying the predominant transfer mechanism; that is, both liquid and vapor transfer were likely to have occurred.

During 1983/84 soil cores were obtained from two sites, one at Outlook and the other at Saskatoon, and analysed for salinity (electrical conductivity, pH and ion content). Three sets of samples were taken: (1) in the fall prior to freeze-up at the time the access tubes used to monitor soil moisture changes were installed; (2) in mid-February when the soils were frozen, and (3) in May, when the soils had thawed. Figures 5 and 6 show profiles of electrical conductivity and soil moisture in the fall (Figs. 5a, 6a) and changes in these profiles from fall to mid winter (Figs. 5b and 6b) and mid winter to spring (Figs. 5c and 6c). In Figs 5b, 5c, 6b and 6c the soil temperature on the sampling date is also plotted. As the electrical conductivity provides a direct measure of the total soluble salt of a sample it was used to index changes in salinity.

At Outlook, by mid February (Feb. 17) the depth of penetration of frost was approximately 950 mm and it can be observed in Fig. 5b that a significant amount of moisture had accumulated above this depth, the
Figure 5. Profiles of soil moisture, electrical conductivity and soil temperature on selected dates and changes in soil moisture and electrical conductivity in the period from fall to the time of seeding measured at Outlook, 1983/84: a) Fall (Nov. 9/83); b) Fall to mid February (Feb. 9/84) and c) mid February to May (May 14/84). Note the temperature profiles given in Figs. 5b and 5c were made on Feb. 9/84 and May 14/84 respectively.
Figure 6. Profiles of soil moisture, electrical conductivity and soil temperature on selected dates and changes in soil moisture and electrical conductivity in the period from fall to the time of seeding measured at Saskatoon, 1983/84: a) Fall (Nov. 10/83); b) Fall to mid March (Mar. 13/84 and c) mid March to May (May 16/84). Note the temperature profiles given in Figs. 6b and 6c were made on Mar. 13/84 and May 16/84 respectively.
result of moisture migration to the freezing front. The figure also shows a corresponding increase in electrical conductivity (and hence salt content) above the frost front which extends to the bottom of the "Upper" zone (0-200 mm). The fact that the zones of salt and moisture accumulation coincide indicates that moisture movement has occurred primarily in the liquid phase. The Upper zone, however, shows essentially no change in conductivity indicating no accumulation of salts. This finding suggests that moisture migration in the zone occurred as a vapor, probably the result of very large temperature gradients at the onset of freezing.

At Outlook the average electrical conductivity of the soil in the zone of salt accumulation increased over winter from 3.5 ms/cm in the fall to 5.7 ms/cm in February. In that the threshold conductivity for saline soils is generally accepted as 4 ms/cm, the soil profile has become saline as a result of the upward movement of salts (principally sodium and magnesium sulfates) during the freezing period.

At the Saskatoon site (see Fig. 6) there was an increase in moisture content in the 150 to 600 mm depth range over the fall to mid-March period, however, there was no increase in electrical conductivity down to a depth of approximately 400 mm. It would appear that the increase in moisture content in the top 400 mm was due to vapor movement. Below 400 mm there was a substantial increase in the electrical conductivity even where there was little increase in moisture content. Apparently water moved upward in the liquid form until it reached a depth between 400 to 800 mm; from there it moved as vapor. Consequently the salts were deposited in the 400-800 increment even though there was little change of moisture content below 600 mm.

Although the moisture content in the fall of each soil at Outlook and Saskatoon fell in the range between 25-30% by volume, the texture
of the soils (silty clay at Saskatoon and fine sandy loam at Outlook) are sufficiently different to cause the moisture flow characteristics in the particular moisture range to differ considerably at the two sites. The soil at Outlook would be near field capacity at 25% moisture content. Therefore it is reasonable that moisture movement would be by liquid transfer unless a high temperature gradient was present. This would explain the electrical conductivity profiles and would indicate that vapor movement was only important in the relatively shallow Upper zone. At Saskatoon, the heavier-textured soil could appear quite "dry" at 30% moisture content with respect to the transfer of water. Water movement in the top 400 mm could easily be due to vapor movement rather than liquid movement. This is borne out by the data.

From the above results and discussions two important propositions can be presented regarding moisture transfer and salt migration in the overwinter period. Namely;

1. Soil moisture changes in a profile occur as a result of moisture transfer in both liquid and vapor phases, and

2. Liquid water moving to a freezing front is an effective transport for the upward migration of soluble salts.

PERIOD OF SNOWMELT

Infiltration to Frozen Soils

During the period of active snowmelt the moisture regime of the soil below the depth of penetration of infiltrating meltwater remains reasonably stationary and unchanged. Ablation of the snowcover in the spring is the time of major recharge of the soil water reserve, particularly to the soil layer adjacent to the surface, by the snow water resource. The amount of recharge is related to the snowcover water equivalent and the infiltration characteristics of the frozen soil at the time of melt.
Based on approximately fifteen years of study of the snow hydrology of the Prairie region, the results of infiltration studies under similar climatic regions of the USSR reported in the literature (Motovilov, 1978, 1979; Popov, 1973) and the findings of the field study of infiltration to frozen soils, it is postulated that frozen Prairie soils may be grouped into three broad categories in respect to their infiltration potential, namely: Restricted, Limited and Unlimited (see Fig. 7).

**Restricted** - infiltration is impeded by an impermeable layer, such as an ice lense, at the soil surface or within the soil at a shallow depth. For practical purposes the amount of meltwater infiltration can be assumed to be negligible and most of the snowcover water equivalent goes to evaporation or direct runoff. Occurrences promoting this condition include rainfall or snowmelt late in the fall near the time of freeze-up and melt/rainfall-freeze events during winter or prior to the time of active continuous melt.

**Limited** - infiltration is governed primarily by the snowcover water equivalent and the frozen water (ice) content of the shallow layer of soil, 0-300 mm.

**Unlimited** - a soil in this condition contains a high percentage of large, air-filled, non-capillary pores or macropores at the time of melt and most or all the snow water will infiltrate. Examples of soils having these properties are dry, heavily-cracked clays and coarse, dry sands.

In the above classification it is evident that in a practical soil water modelling scheme when evaporation and storage losses are neglected, the volumetric infiltration coefficient (ratio of the amount of infiltration to the snowcover water equivalent) to be assigned to a
Restricted: Infiltration is low, high runoff potential.

(b) Limited: Infiltration is governed primarily by ice content of the soil layer 0-300mm at the time of melt.

(c) Unlimited: Soil has the capacity to infiltrate all or most of the snowcover water equivalent.

Figure 7. Conceptual model for classifying the infiltration potential of frozen soils: (a) Restricted; (b) Limited and (c) Unlimited.
soil whose infiltration potential is classed as Restricted is 0; for the Unlimited case the coefficient is equal to 1. Thus, the problem remaining is one of defining the relationship between infiltration, snowcover water equivalent and frozen soil moisture content for the Limited case. This can be done using the results reported by Granger et al. (1984). They found in uncracked frozen Prairie soils: (a) the average depth meltwater penetrated a soil during the snowmelt was 260 mm (standard deviation = 100 mm) and (b) the amount of infiltration was inversely related to the average moisture content of the frozen soil layer, 0-300 mm, at the time of melt. Using these findings it can be shown that the interrelationship between snowmelt infiltration (INF), snowcover water equivalent (SWE) and the premelt moisture content of the 0-300 mm soil layer ($\theta_p$), for cases where SWE>INF, can be described by the equation:

$$\text{INF} = 5(1 - \theta_p)\text{SWE}^{0.85}, \quad \ldots \ (1)$$

in which INF and SWE are in mm and $\theta_p$ is the degree of pore saturation cm$^3$/cm$^3$. A comparison of measured infiltration amounts with those calculated by Eq. 1 gave a correlation coefficient of 0.85. Figure 8 shows the family of curves defined by Eq. 1.

Equation 1 was developed from measurements made in soils with different vegetative cover, i.e., fallow, stubble and grass. The weak association found between snowmelt infiltration to frozen soils and surface cover is in contrast to the strong relationship between these variables in unfrozen soils. This can be attributed primarily to refreezing of meltwater when it enters a frozen soil. The major effects of land use and/or vegetative cover are they influence the soil moisture content at the time of freeze-up, hence the premelt ice content, and the amount of snow water accumulated on the soil surface. Both factors affect the soil temperature regime but in differing amounts. Unfortu-
SNOW WATER EQUIVALENT (mm)

Figure 8. Relation between infiltration (INF), snowcover water equivalent (SWE) and the premelt moisture content of the soil layer, 0-300 mm, ($\theta_p$) for frozen, uncracked Prairie soils. Unfortunately, the effect of soil temperature on infiltration to frozen soils is less clear than soil moisture and the absolute effect of a difference in temperature of a frozen soil, e.g., -3°C versus -5°C, on the process cannot be quantified. Also, an important role of a snowcover is that it reduces the overwinter soil moisture loss from the surface.

Equation 1 can be solved for INF when SWE and $\theta_p$ are known. It is expected that any person interested in estimating snowmelt infiltration would have snowcover data available. Hence the major obstacle restricting its use is it requires an estimate of the premelt moisture content, $\theta_p$. In the previous section it was reported that changes may occur in the moisture regime of a Prairie soil over winter because of moisture transfer as a vapor across the soil/air or soil/snow interfaces, the infiltration of water from mid-winter snowmelt or rain events and the migration of water in response to the freezing action. In the absence of mid-winter infiltration, the soil moisture changes in the Upper zone are relatively small (particularly when the soil is snowcovered through-
out most of the winter) and the soil moisture of the 0-300 mm layer in
the fall ($\theta_f$) can be used to index the moisture content at the time of
melt ($\theta_p$). The "best-fit" regression equations describing the relationships for fallow and stubble are:

Fallow: $\theta_p = -5.80 + 1.05 \theta_f$, and ... (2a)

Stubble $\theta_p = 0.294 + 0.957 \theta_f$ ... (2b)

in which $\theta_p$ and $\theta_f$ are expressed as a percent moisture by volume. Each
expression has a correlation coefficient of approximately 0.9 with a
standard deviation from regression of approximately 3.3% by volume.

Recently Gray et al. (1984b) developed a snowmelt infiltration
model based on the above concepts and empirical relationships and used
it to calculate streamflow from snowmelt on the Creighton Tributary, a
small watershed (11.4 km$^2$) located in the semi-arid region of western
Saskatchewan. They compared the areally weighted snowcover water equivalent less infiltration ($Q_{cal}$) with the measured streamflow from snow-
melt ($Q_{meas}$) in 1974 and 1975; two years of widely different snowcover
and soil moisture conditions. The winter of 1973/74 was a year of near
record snowfall producing an average depth of snowcover on the watershed
of 556 mm having a water equivalent of 143 mm. The fall soil moisture
conditions over the watershed were very dry (average -15% by volume in
the 0-300 mm soil layer) due to an extended warm, dry fall. Snowcover
and soil moisture conditions on the area in 1974/75 were closer to
normal in which the average snowcover depth and water equivalent were
299 mm and 72 mm respectively and the average "fall" soil moisture
content was 27.4% by volume. The ratio, $Q_{cal}:Q_{meas}$ was 1.02 in 1974
and 1.16 in 1975. The results support the concept of the infiltration
model and further verification and testing in streamflow and soil mois-
ture forecasting systems is recommended.
Snow Management Practices

As mentioned in the Introduction of this paper, in recent years there has been renewed interest in managing snow for better utilization of this water resource for crop production. For cereal grains, stubble management is probably the most viable snow management practice. The most popular methods include tall stubble, alternate height stubble, hi-low stubble and leave strips.

**Tall Stubble** - the term "tall" stubble is used in a "relative" sense to infer that the crop is cut at the highest height possible during the harvesting operation, thus leaving a field of "high" stubble. Normally this practice is used when the crop is straight-combined. The actual height of a "tall" stubble is a function of the crop variety, plant density and crop height, but usually will fall in the range from 300 to 600 mm.

**Alternate-Height Stubble** - this practice involves harvesting the crop to leave bands of stubble of alternate height (low and high) within a field. The width of each band is normally the width of the swather. By adjusting the swather height on each round of a field, a series of bands of "low" (150-300 mm) and "tall" stubble (300-600 mm) are formed.

**Hi-Low Stubble** - this practice, also referred to as deflector strips, involves leaving narrow strips of tall stubble in a field. The strips, which are usually 400 to 600 mm wide, are made with a swather or combine header attachment called a deflector. Several deflector units have been developed; a common type is the 'V'-shaped deflector consisting of a V-shaped divider having lifter fingers on either side which are fastened to the bottom of the guards of the cutter head of the swather. As the swather travels through the crop, the deflector
separates the crop and forces the stems to bend sideways. The heads of the grain are lifted vertically by the lifter fingers and cut by the knife leaving a strip having the shape of an inverted 'V' which is generally 250-350 mm higher than the adjacent stubble. The use of single or double deflectors is common.

**Leave Strips** - in this practice, 300-400 mm wide strips of crop, spaced 1, 2 or 3 swather/combine widths apart are left unharvested to act as barriers to trap snow. The width of a swather usually is in the range from 4.6 to 7.6 m. With a 300 mm wide strip placed on a 15-m spacing, approximately 2% of the area of a field is left unharvested.

A particular advantage in cutting a field to leave stubble at different heights is that the vegetative surface formed is aerodynamically rougher than that of a uniform stubble; consequently, its snow trapping efficiency is better. A stubble of variable height would be favored where crop conditions, principally density and height, are insufficient to provide a "tall" dense stubble and on highly-exposed topographic facets, e.g., the tops of ridges and knolls.

Figures 9 and 10 are typical snowcover accumulation patterns found in fields harvested with "alternate-height" and "hi-low" practices. The data in Fig. 9 were obtained in March of 1983 at a site where snowcover, wind and other conditions favored accumulation and it can be observed that the snowcover filled the vegetation. The average density of the snow in the low stubble was 207 kg/m³ and in the tall stubble, 248 kg/m³; the higher density in the tall stubble was likely due to greater packing by wind. Taking an average snowcover depth in the low stubble of 200 mm and 400 mm in the high stubble, the difference in snow water equivalent on the two strips would be -58 mm which represents an average increase over the field attributable to the tall stubble of -29 mm. The data in Fig. 10 were collected in February, 1983 and show larger accumulations in a field with "hi-low" strips than in an adjacent
Figure 9. Snowcover accumulation pattern on a field cut with "alternate-height" stubble management practice, Eston, Sask., 1983.

Figure 10. Snowcover accumulation pattern on a field cut with "hi-low" stubble management practice at Saskatoon, Sask., 1983. W = width of deflector strip, S = distance between strips.
field of "tall" stubble. In the figure note the change in the accumu-
lation pattern in the strips caused by reducing both the width and spacing
of the strips by a factor of about two. The accumulation on the wide
spacing exhibits an undulating pattern as a result of scour or erosion
of snow from between the strips. Conversely, on the area with the
strips placed on the narrow spacing, the depth of snowcover is more
uniform and deeper, the average increase in depth being approximately
50 mm. Further work is needed to define the interaction between snow-
cover accumulation, the physical properties (spacings, density, height)
of strips and upwind fetch conditions in order to establish design
criteria for this stubble management practice.

Frozen Soils of Limited Infiltration Potential

The evidence showing an increase in depth of accumulated snow with
a snow management practice is indisputable (see Steppuhn, 1981); how-
ever, the effects of increased depth on snowmelt infiltration, soil
water augmentation and crop yield are unclear. Referring to Fig. 8 it
can be observed that the rate of increase in infiltration per unit
increase in snowcover water equivalent decreases with increasing SWE.
This trend is also exemplified by the increase in departure of a curve
of constant moisture content ($\theta_p$) from the 45° line which describes the
condition, $\text{INF} = \text{SWE}$. The curves flatten out and tend toward a reason-
ably constant value; the SWE-value ($\text{SWE}_c$) at which this trend occurs
varies with $\theta_p$. The higher $\theta_p$, the lower $\text{SWE}_c$, and vice versa. For $\theta_p$
in the range 0.20-0.90, $\text{SWE}_c$ appears to fall between 5 mm and 100 mm.

A practical aspect of the shape of the infiltration curves is that
it can be used to establish an upper limit in the amount of snow water
which should be accumulated on a soil of a given premelt moisture con-
dition. The optimum value of $\text{SWE}_c$ for a given premelt moisture condi-
tion will depend on many factors: the interaction of snowmelt infiltra-
tion and crop yield, especially the incremental increases in yield per
unit increase in infiltration; the interactions between direct runoff and erosion and drainage problems; crop conditions at the time of harvest and others. No single value will satisfy all cases, hence only general guidelines and criteria can be presented to assist with the selection. Table 1 was prepared for this purpose. It lists the recommended snow water equivalent (RSWE) and corresponding height of stubble (D), which when filled by a snowcover having a density equal to 250 kg/m$^3$ would provide the RSWE, for different soil moisture levels and infiltration efficiencies; i.e., the relative amount of RSWE that would infiltrate (equal to the ratio INF/RSWE). To illustrate the application of these data, assume the case where it is desirable to have 70% (or more) of the snowcover water infiltrate a soil having a premelt moisture content of 30% of saturation. Entering Table 1 with $\theta_p = 0.30$ and $x = 0.70$, the recommended value for D is $\approx 190$ mm. It is obvious from the data presented in the table that for a given $\theta_p$, D increases and the incremental increase in the amount of infiltration per unit increase in snowcover water equivalent decreases as INF/SWE decreases.

The horizontal lines in Table 2 define minimum values of RSWE and D and correspond to a stubble height of 100 mm. Practicing farmers suggest that rarely would a crop be harvested at a lower height unless this was unavoidable. The values for D in the table can be adjusted for a different snow density ($\rho$) by direct calculation using RSWE-values given; e.g., $D = (\text{RSWE} | \rho) \times 250$. Because the data presented were derived from measurements made at sites on relatively flat topography, one may wish to consider increasing the value of D given by the table to account for the differences in infiltration time and exposure conditions on mid- and upper-slope topographic facets.

As a guide it is recommended that for the "broad" soil moisture
Table 2. Recommended snowcover water equivalent (RSWE) and corresponding stubble height (D), which when filled with a snowcover having a density of 250 kg/m³ will give the RSWE, for different premelt soil moisture conditions ($\theta_p$) and infiltration efficiencies (x).

<table>
<thead>
<tr>
<th>$\theta_p$</th>
<th>x = 1.0</th>
<th>x = 0.9</th>
<th>x = 0.8</th>
<th>x = 0.7</th>
<th>x = 0.6</th>
<th>x = 0.5</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>RSWE (mm)</td>
<td>D (mm)</td>
<td>RSWE (mm)</td>
<td>D (mm)</td>
<td>RSWE (mm)</td>
<td>D (mm)</td>
</tr>
<tr>
<td>.2</td>
<td>28.0</td>
<td>112</td>
<td>36.1</td>
<td>144</td>
<td>47.9</td>
<td>192</td>
</tr>
<tr>
<td>.3</td>
<td>20.3</td>
<td>81</td>
<td>26.2</td>
<td>105</td>
<td>34.7</td>
<td>139</td>
</tr>
<tr>
<td>.4</td>
<td>14.0</td>
<td>56</td>
<td>18.1</td>
<td>72</td>
<td>24.0</td>
<td>96</td>
</tr>
<tr>
<td>.5</td>
<td>9.0</td>
<td>36</td>
<td>11.7</td>
<td>47</td>
<td>15.5</td>
<td>62</td>
</tr>
<tr>
<td>.6</td>
<td>5.3</td>
<td>21</td>
<td>6.8</td>
<td>27</td>
<td>9.0</td>
<td>36</td>
</tr>
<tr>
<td>.7</td>
<td>2.7</td>
<td>11</td>
<td>3.4</td>
<td>14</td>
<td>4.5</td>
<td>18</td>
</tr>
<tr>
<td>.8</td>
<td>1.0</td>
<td>4</td>
<td>1.3</td>
<td>5</td>
<td>1.7</td>
<td>7</td>
</tr>
<tr>
<td>.9</td>
<td>1.2</td>
<td>2</td>
<td>1.2</td>
<td>2</td>
<td>3.3</td>
<td>1</td>
</tr>
</tbody>
</table>

a) x = infiltration efficiency, i.e. ratio snowmelt infiltration: snowcover water equivalent.
classes: "very dry" - near the wilting point ($\theta_p \approx 0.2 + 0.35$); "medium" - between the wilting point and field capacity ($\theta_p \approx 0.35 + 0.45$) and "moist" - near field capacity ($\theta_p \approx 0.45 + 0.60$), the stubble should provide a snowcover depth in the ranges of $250 \div 320$ mm, $150 \div 220$ mm and $100 \div 120$ mm respectively.

Frozen Soils of Unlimited Infiltration Potential and Subsoiled Soils

Within the framework of the infiltration model presented herein (see Fig. 7) it is evident that those soils which are likely to benefit most in terms of water augmentation from snowmelt are those classed as having Unlimited infiltration potential. These soils contain a large number of air-filled macropores which extend from the soil surface to considerable depth. Under natural conditions the arable soils that most commonly exhibit these properties are those heavy, lacustrine clays which severely crack on drying (either the result of moisture withdrawal for evapotranspiration or dessication on freezing). The macropore content of soils that do not crack by natural causes can only be increased substantially by some tillage or subsoiling operation such as deep cultivation or cultivation.

Table 3 lists some representative values of snowmelt infiltration to uncracked, cracked and subsoiled soils measured in the same fields at five different locations in Saskatchewan during snowmelt of 1983 and 1984. The following points should be noted when interpreting these data.

1. The values given for infiltration and the depth infiltrating water penetrates a frozen soil were obtained from measurements made with a twin probe density meter. On the cracked sites and those subsoiled with the a deep cultivator (Killefer plow) the crack or opening was located between the access tubes used to measure soil
Table 3. Representative values of the amount of infiltration (INF) and the depth infiltrating meltwater penetrates into a frozen soil during the melt period (d) for some uncracked, cracked and subsoiled Prairie soils.

<table>
<thead>
<tr>
<th>Location and Soil Texture</th>
<th>Year</th>
<th>Landuse</th>
<th>Snowcover Water Equivalent (mm)</th>
<th>Snowmelt Infiltration</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Uncracked Cracked Subsoiled</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>INF^a (mm) d^b (mm)</td>
<td>INF (mm) d (mm)</td>
</tr>
<tr>
<td>BAD LAKE (heavy clay)</td>
<td>1983</td>
<td>Fallow</td>
<td>46.2 19.5 220</td>
<td>61.2 24.7 360</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Stubble</td>
<td>69.1 17.9 160</td>
<td>86.7 27.3 160</td>
</tr>
<tr>
<td></td>
<td>1984</td>
<td>Fallow</td>
<td>28.6 7.6 140</td>
<td>60.4 1.0 100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Stubble</td>
<td>49.9 1.0 100</td>
<td>54.3 27.8 560</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>86.8 27.8 560</td>
<td>99.9 113.8 920</td>
</tr>
<tr>
<td>KERROBERT (clay loam)</td>
<td>1984</td>
<td>Fallow</td>
<td>19.7 4.7 120</td>
<td>39.1 3.8 100</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Stubble</td>
<td>23.0 4.3 60</td>
<td>78.2 14.6 90</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>55.4 26.7 400</td>
<td>150.0 26.7 400</td>
</tr>
</tbody>
</table>

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Table 3 - continued

<table>
<thead>
<tr>
<th>Location and Soil Texture</th>
<th>Year</th>
<th>Landuse</th>
<th>Snowcover Water Equivalent (mm)</th>
<th>Snowmelt Infiltration</th>
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<tr>
<td></td>
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<td>Uncracked Cracked Subsoiled</td>
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<tr>
<td></td>
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<td></td>
<td>INF^a d^b INF d INF d</td>
<td>(mm)</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>Fallow 60.7 15.1 180</td>
<td>83.1 35.9 460</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>60.7 40.8 330</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Stubble 63.2 94.3 800</td>
<td>59.7 21.0 320</td>
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<tr>
<td>PORTREEVE (heavy clay)</td>
<td>1983</td>
<td>Fallow 60.7 15.1 180</td>
<td>83.1 35.9 460</td>
<td>135.5 1080</td>
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<tr>
<td></td>
<td></td>
<td>Stubble 120.0 106.1 1160</td>
<td>25.1 200</td>
<td>99.4 880</td>
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<tr>
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<td>63.2 94.3 800</td>
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<td></td>
<td></td>
<td></td>
<td>120.0 106.1 1160</td>
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<td>RICHLEA (heavy clay)</td>
<td>1983</td>
<td>Fallow 59.2 21.5 200</td>
<td>60.2 107.4 1440</td>
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<tr>
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<td>Stubble 60.2 107.4 1440</td>
<td>73.3 21.2 190</td>
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<td></td>
<td>1984</td>
<td>Fallow 33.4 15.4 200</td>
<td>73.3 21.2 190</td>
<td>77.7 38.2 300</td>
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<td>Stubble 77.7 38.2 300</td>
<td>112.2 777</td>
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Table 3 - continued

<table>
<thead>
<tr>
<th>Location and Soil Texture</th>
<th>Year</th>
<th>Landuse</th>
<th>Snowcover Water Equivalent (mm)</th>
<th>Snowmelt Infiltration</th>
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<tr>
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<td></td>
<td></td>
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<td>Uncracked</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>INF&lt;sup&gt;a&lt;/sup&gt;</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>(mm)</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>d&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(mm)</td>
</tr>
<tr>
<td>SASKATOON (heavy silty clay)</td>
<td>1982</td>
<td>Fallow</td>
<td>51.7</td>
<td>10.0</td>
</tr>
<tr>
<td></td>
<td></td>
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<td>57.2</td>
<td>21.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>61.2</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Stubble</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>1984</td>
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<td></td>
<td></td>
<td>23.3</td>
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</tr>
</tbody>
</table>

a) INF = depth (amount) of infiltration.
b) d = depth meltwater penetrated the frozen soil during the melt period.
c) Subsoiling treatments: single shank deep cultivator used at Kerrobert (Killefer plow); Paraplow used at Saskatoon.
density. As mentioned previously, the twin probe measures soil density and density changes between successive dates of measurement are attributed to soil moisture changes. Thus, if the mass of soil confined between the tubes changes during the period of infiltration, for example because of swelling, dessication or sloughing of the soil, this will produce an erroneous reading in the soil moisture content. Because an effective procedure of partitioning the density changes to soil density and soil moisture (if indeed required) has not been developed it is recommended that the snowmelt infiltration amounts listed for the cracked and subsoiled soils be accepted as index values.

2. The amount of infiltration measured in a crack can be much larger than the snowcover water equivalent. This occurs when a crack receives snowmelt water generated on areas adjacent to the crack either by direct surface runoff or interflow as a result of differences in elevation. In reality these factors cause the drainage area of a crack to be much larger than the area exposed at the surface.

3. The subsoiling treatments included a deep cultivator (Killefer plow) used at Kerrobert and a Paraplow used at Saskatoon. The Killefer plow was developed in the mid western United States and one of its principal uses was as a subsoiler to break up hard pans underlying coarse, medium-textured loessial soils so as to improve their drainability. It is simply a large wheel-mounted cultivator with a shank approximately 25 mm in thickness, 200 mm in width and 970 mm in length having which a wedge-shaped chisel (~75 mm in width) attached to the bottom. When drawn through the soil it produces a 30 mm slit (opening) to a depth of about 600 mm. Fracturing of the soil, caused by the chisel extends to a depth of ~800 mm.
The paraplow was developed in the United Kingdom and only recently introduced to the North American market. Its physical features are similar to those of a moldboard plow; however, rather than rotating and inverting a furrow it simply lifts, shears and fractures the soil and leaves a standing stubble or trash cover on the surface intact. The width and depth of cut of each plow is approximately 500 mm.

The data in Table 3 confirm the substance of previous discussions, namely the amount of snowmelt infiltration and the depth of penetration of meltwater are greater in a cracked or mechanically-disturbed soil than an uncracked soil. The average ratio of snowmelt infiltration to cracked and uncracked soils under similar snowcover conditions was calculated as 3.72.

The actual amount of infiltration to a field depends not only on the amount of snow water but also on the density (spacing) and physical dimensions of the natural cracks. Measurements made with the two probe density meter show a depth-of-cracking usually in the range 400-800 mm; the deeper depths being associated with soils under stubble following a very dry summer and fall and undergo dessication by freezing. Direct "ruler" measurements of the surface width of cracks usually gave a value in the range of 30-50 mm. By direct calculation, assuming a crack of rectangular shape with an average width and depth of 40 mm by 600 mm, would result in a volume equal to 8% of the volume of a block of soil 0.3 m x 0.3 m x 1 m and 1.8% of a block 1 m x 1 m x 1 m. The storage potential of the cracks for infiltration in terms of the snowcover water equivalent on the two areas (0.09 m² and 1 m²) would be 80 mm and 18 mm respectively. As a first guess, however crude, it is postulated that under natural conditions cracking of a soil probably leads to average increase in the "field" infiltration potential at least of the order of 35-65 mm of water. Note, it is expected the
uncracked parts of a cracked field, despite an increase in soil density, would still infiltrate some meltwater although perhaps in a smaller amount than they would in the natural condition. Additional research is needed toward quantifying the absolute increase of cracks on the soil infiltration potential.

**Practical Method of Estimating Snowmelt Infiltration**

In the above discussions a method of estimating the amount of snowmelt infiltration to uncracked frozen soils from the snowcover water equivalent and the premelt soil moisture (ice) content has been presented. Frequently however a person will not have access to these data, hence the problem is one of indexing the quantity from other parameters.

On review of the data given in Table 3 an association between snowmelt infiltration (INF) and the depth of penetration of meltwater into a frozen soil (d) can be observed, i.e., INF increases with d. This trend was also noted in other data. Granger et al. (1984) showed that following snowmelt the moisture content of the soil layer which infiltrating meltwater penetrated tended toward a limit of saturation. The limit value was highest in the wet soils and lowest in dry soils. In other words the amount of infiltration is strongly associated with the available soil moisture storage, to the extent that INF will be directly related to d in soils with the same premelt moisture content. Similarly in cracked soils one would expect a similar trend provided the cracks filled completely with water.

Figure 11 shows a plot of INF versus d from measurements obtained at 138 cracked and uncracked sites. The best-fit regression line for the data is,

\[ INF = -5.89 + 0.1139d \]

... (3)

in which INF and d are in mm. The standard error of estimate of the
regression is 20.5 mm with a correlation coefficient of 0.86. Despite the scatter of the data, Eq. 3 allows a simple direct means of obtaining a rough estimate of infiltration from measurements of the depth of penetration of the infiltrating water. For practical purposes $\text{INF} \approx 0.10 \text{ d}$, i.e. following melt if one were to find a wetted depth (usually the depth to the frozen layer) of 100 mm, $\text{INF}$ would be estimated to be $0.10 \times 100 = 10 \text{ mm}$. Obviously in cracked fields, both the cracked and uncracked portions should be probed and the values weighted in proportion to the relative areas each represent so as to obtain an average value for infiltration.

![Graph showing infiltration vs. depth of penetration]

Figure 11. Scatter diagram showing the relationship between infiltration and the depth meltwater penetrates a frozen soil during snowmelt.

POST MELT

Following the disappearance of the seasonal snowcover up to the time of seeding, which usually spans a six-nine week period, a soil moisture profile undergoes changes because of withdrawal of water from
the surface layer of soil due to evapotranspiration, thawing of the soil followed by the downward movement of water, and infiltration of precipitation (rain and/or snow). Figures 2h, 3g and 3h show typical profiles of changes in the soil moisture regime following melt. At Outlook the site was snowfree by March 28/83; at Saskatoon by April 18/83. The curves show:

1. A loss of moisture from the Upper zone (0 - ~200 mm).

2. A frozen soil thaws from both the top and bottom. During the thawing process evaporation occurs from the soil surface and water below the frost line drains downward. Note however (Figs. 2h and 3h) there is no pronounced change in the soil moisture profile due to drainage until the soil temperature throughout the frozen zone reaches 0°C. The increased rate and amount of drainage at Outlook, as compared with Saskatoon, can be attributed to the larger amount of water which had migrated and to the better drainage properties of the sandy soil.

Regarding the above (items 1 and 2) Table 4 shows that at 124 sites where soil moisture changes were monitored all showed evaporation losses during the post-melt period whereas at 66 sites there was measurable drainage from the root zone (~1 m). Calculations by Dyck and Granger (1979; 1980) of evaporation in the post-melt period using Morton's evapotranspiration model (Morton, 1978) applied to climatological data from Bad Lake suggest that 20-60 mm of water may be lost to areal evapotranspiration during the period. Similar values have also been reported by Martin (1977). That is, upwards of 20% of the total amount of water which would produce an average crop may be lost in the period.

Table 5 summarizes the soil moisture changes $\Delta$SM, precipitation (Ppt) and evaporation (EVAP) recorded in the post-melt period at
Table 4. Post-Melt disposition of infiltrated meltwater.

<table>
<thead>
<tr>
<th>Location</th>
<th>Number of Observations</th>
<th>Observed Evaporation Losses After Melt</th>
<th>Observed Downward Movement of Meltwater</th>
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<tr>
<td>Saskatoon</td>
<td>37</td>
<td>37</td>
<td>25</td>
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<tr>
<td>Outlook</td>
<td>18</td>
<td>18</td>
<td>7</td>
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<tr>
<td>Bad Lake</td>
<td>26</td>
<td>26</td>
<td>16</td>
</tr>
<tr>
<td>Portreeve</td>
<td>16</td>
<td>16</td>
<td>6</td>
</tr>
<tr>
<td>Richlea</td>
<td>6</td>
<td>6</td>
<td>2</td>
</tr>
<tr>
<td>Kerrobert</td>
<td>11</td>
<td>11</td>
<td>5</td>
</tr>
<tr>
<td>Flaxcombe</td>
<td>10</td>
<td>10</td>
<td>5</td>
</tr>
<tr>
<td>Total (percent)</td>
<td>124 (100%)</td>
<td>124 (100%)</td>
<td>66 (53%)</td>
</tr>
</tbody>
</table>

selected sites at Bad Lake, Kerrobert, Portreeve and Saskatoon. The table also contains information on the relative magnitude of EVAP in terms of the snowmelt infiltration (INF) and the sum INF + Ppt, i.e., the total available water. Note, EVAP is taken equal to the sum of ∆SM + Ppt. It can be observed that the calculated evaporation amounts fall in the range from 14–54 mm, a range of values which is in agreement with that predicted by the evapotranspiration model. No consistent pattern was found for the moisture loss to be significantly greater or smaller from fallow than stubble. This would suggest that evaporation rather than transpiration is the dominant process affecting the moisture loss. On uncracked soils, 83% of the sites showed a post-melt evaporation loss greater than the snowmelt infiltration; the average value of EVAP/INF was calculated as 2.41. Similarly at 42% of the sites, evaporation exceeded the sum of snowmelt infiltration plus precipitation. The average value of EVAP/(INF + Ppt) was 1.31. A person should exercise considerable caution in applying these average ratios as they are heavily biased to conditions in 1984, a year of low snowmelt infiltration and post-melt precipitation. For example, at Saskatoon the average ratios of EVAP/INF and EVAP/(INF + Ppt) in 1984 were 3.41 and 0.97 respectively compared with average values of 1.23
Table 5. Soil moisture changes ($\Delta$SM), evaporation (EVAP) and precipitation (Ppt) in the post-melt period and comparison of the relative magnitudes of EVAP to snowmelt infiltration (INF) and the sum INF + Ppt for uncracked and cracked soils.

<table>
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<tr>
<th>Location</th>
<th>Year</th>
<th>Site</th>
<th>Land Use</th>
<th>Post-Melt Period</th>
<th>$\Delta$SM $^a$</th>
<th>Ppt $^b$</th>
<th>EVAP $^c$</th>
<th>INF $^d$</th>
<th>EVAP INF</th>
<th>EVAP INF + Ppt</th>
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<td>02</td>
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<td>21/3 - 9/5</td>
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... continued
### Table 5. ... Continued

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<th>Location</th>
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<th>ΔSM&lt;sup&gt;a&lt;/sup&gt; mm</th>
<th>Ppt&lt;sup&gt;b&lt;/sup&gt; mm</th>
<th>EVAP&lt;sup&gt;c&lt;/sup&gt; mm</th>
<th>INF&lt;sup&gt;d&lt;/sup&gt; mm</th>
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<td>9.7</td>
<td>4.06</td>
<td>1.04</td>
<td></td>
</tr>
</tbody>
</table>

Mean: Uncracked 2.41 1.31
Cracked 0.28 0.24

a) ΔSM = change in soil moisture in post-melt period.
b) Ppt = precipitation in post-melt period.
c) EVAP = evaporation in post-melt period = ΔSM + Ppt.
d) INF = snowmelt infiltration.
e) designates cracked soil.
and 0.85 for the two years, 1979 and 1980.

The implications of the results important to agriculturalists are:
1. In many years the amount of evaporation in the post-melt period in uncracked soils exceeds the gain from snowmelt infiltration. In these cases snowmelt infiltration is used to compensate the evaporation loss and the moisture status of a soil, particularly to seeding depth, at the time of seeding will depend primarily on precipitation received in the post-melt period.
2. The effect of post-melt precipitation events on the moisture content at seeding are twofold: (a) they reduce evaporation demands during the period and thereby conserve soil water derived from snow and (b) they contribute moisture directly to the soil water reserve through infiltration.

The data in Table 5 also show that in cracked soils the evaporation loss was much less than the infiltration gain. In the cracked soils the average values of the ratios EVAP/INF and EVAP/(INF + Ppt) were 0.28 and 0.24 respectively. The increased benefit of snowmelt infiltration to these soils, as reflected by the low values of the ratios, results from increases in both the amount of infiltration and the depth of penetration of the meltwater. Water stored deep in the root zone is less susceptible to evaporation.

The fact that much of the water that migrates in response to freezing is retained within the root zone until the frost goes out of the ground, which can be as late as the date of seeding, and that much, if not all, of the snowmelt infiltration may be lost to evaporation during the post-melt period suggests that established crops would benefit most from root-zone water derived from these sources. In this regard a case may be made for the use of winter wheat for more efficient moisture conservation and utilization in the northern Prairies.
As discussed above as item 2, following thawing of the frozen soil water moves downward due to gravity. Drainage is important, as it leaches these salts which have moved upward during the freezing cycle from the root zone. The effect is illustrated in Figs. 5c and 6c in which decreases in both the moisture content and electrical conductivity (salt content) from the frozen zone during the post-melt period can be observed. The change in salt content roughly followed the moisture change. By May, at both Outlook and Saskatoon the soil moisture and salt distributions to a depth of about 700 mm have returned to approximately the same levels found in the fall. There is some indication however, that the salt content of the soil at Outlook below 1 m and the soil at Saskatoon below 700 mm may have increased slightly in the Fall-May period. Conceivably, further drainage would reduce the quantities.

SUMMARY

The paper describes, in the form of a time sequence, changes in the moisture regime of a Prairie soil in the interval from fall to the time of seeding cereal crops. It is emphasized that the moisture regime is very active over winter because of moisture losses from the surface soil layer and the migration of water to the freezing front. The relocation of water in a soil profile during winter can directly affect the snowmelt infiltration potential of a soil; soil moisture measurement and budget procedures; the feasibility of fall irrigation practices and the migration of soluble salts.

Attention is also given to the interaction between snowmelt infiltration, snowcover water equivalent, and the premelt soil moisture (ice) content of frozen soils. An empirical relationship describing the relationship between these parameters and a method of estimating infiltration from simple, direct field measurements are presented. It is also demonstrated that maximum benefits from snow management prac-
tices for increasing soil water reserves in uncracked soils will occur when they are applied in combination with some practice that increases the macropore content of a frozen soil at the time of melt.

The paper also discusses the disposition of soil water from the end of snowmelt to the time of seeding of cereal grains. It is shown that evaporation losses and precipitation in this period strongly affect the soil moisture status to the seeding depth. Evaporation losses in the period often exceed snowmelt infiltration.

ACKNOWLEDGEMENTS

The results presented herein are based on data acquired in a comprehensive field study undertaken in the six-year period 1978-84. This program has involved a large number of human, financial and physical resources and we wish to acknowledge the contributions of:

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