SNOWMELT-FROZEN SOIL CHARACTERISTICS FOR A SUBARCTIC SETTING

Completion Report
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ABSTRACT

The pathways of soil water in cold climates are influenced, in addition to the normal forces, by the presence of permafrost and the temperature gradients in the soil system, whereas the infiltration of surface water into the soil system is a function of moisture levels, soil type and condition of the soil (whether it is frozen or not). Snowfall, with subsequent surface storage over a period of several months, typifies Alaskan winters. This snowfall often accounts for 50 per cent or more of the annual precipitation, with ablation occurring over a time span of 2 to 3 weeks in the spring. The melt period represents an event when large quantities of water may enter the soil system; the possibilities exist for recharging the groundwater system, or else generating surface runoff. The objective of this study was to determine the magnitude of potential groundwater recharge from snowmelt.

Instrumentation was installed and monitored over two winter seasons to quantify the accumulation and ablation of the snowpack. Thermal and moisture data were collected to characterize the snowpack and soil conditions prior to, during, and following the ablation. Lysimeters were installed at various depths to intercept soil water. The volume of potential areal recharge for 1976 was 3.5 cm and for 1977 was 3.0 cm, which represented about 35 per cent of the maximum snowpack content. It is concluded that permafrost-free areas can contribute significantly to groundwater recharge during snowmelt ablation.

ACKNOWLEDGMENT

The work upon which this completion report is based was supported by funds provided by the United States Department of the Interior, Office of Water Research and Technology under the Water Resources Research Act of 1964, Public Law 88379, as amended. The authors appreciate the assistance of Lalitha Rao, Roberta Jones, Ellen Kane, and Timothy Cordis in this project.

INTRODUCTION

The movement of snowmelt water is limited to three possible manifestations. It may evaporate back to the atmosphere, migrate into the subsurface system, or flow over the land surface. Snowmelt water which flows over the ground surface, as well as very shallow subsurface flow, is responsible for the generation of peak runoff events. That which flows into the deeper groundwater system is responsible for recharging the subsurface water resources.

The distribution of water among the three possible paths is not well established for northern regions. This is especially so for the water that flows into the deeper groundwater system. Since this infiltration is not easy to measure, it is often an estimated element in hydrologic balance studies. It is also one of the more important facets of the hydrologic cycle from the standpoint of human use, since it is a source of groundwater recharge.

A conceptual representation of the hydrologic cycle of cold regions is presented in Figure 1. There are two unique subarctic processes that are not observed or considered of interest in temperate climates. First, there is the process of aufeis development over the winter months. Excess pressures evolve in the stream channel under the ice cover forcing water up through cracks in the ice and onto the ice surface where it subsequently freezes. This ice, over the winter season, reaches thicknesses far in excess of what would be predicted for the normal depth of flow. The second process is the movement of water from a relatively warm area to a relatively cold area. During the winter months, moisture in the soil system moves from depth (warmer area) toward the ground surface (colder area) in response to a pressure gradient caused by the temperature gradient. This process is of importance in studying snowmelt infiltration because a redistribution of moisture is taking place during the winter season prior to the initiation of melt and ablation. It is possible that very thin layers of frozen soil that are relatively impermeable due to the high ice content can prevent or substantially reduce the amount of snowmelt infiltration. In general, processes occurring in cold regions are the same as those in

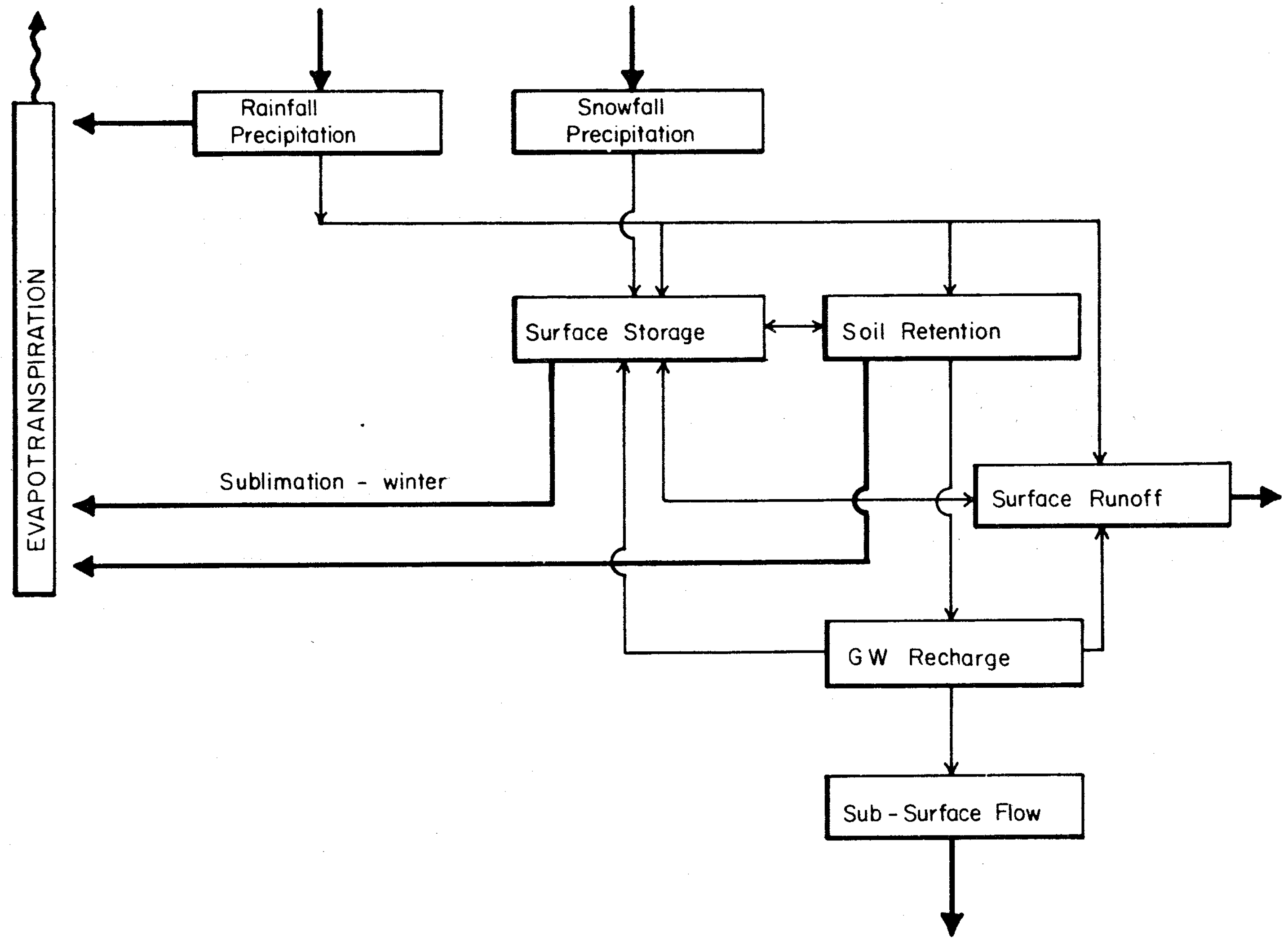

FIGURE 1: A Conceptual Representation of the Hydrologic Cycle of Cold Regions.

more temperate regions; however, of interest is the magnitude of variation that occurs in these physical processes in direct response to the extreme climatic conditions.

The major goal of this study was to define the amount of potential groundwater recharge from a melting snowpack in a subarctic setting; basically, this was a field study. The data were expected to aid in determining the magnitude of possible groundwater recharge, and serve as input in groundwater flow models to be implemented later. Two sites were selected for instrumentation: a permafrost-free site and a permafrost site. The permafrost site was underlain by relatively impermeable ice-rich permafrost. Because of the low hydraulic conductivity, there would be essentially no groundwater recharge at this site, and the water either would move downslope or return to the atmosphere by either evaporation or transpiration. The permafrost-free site would represent an intermediate area for groundwater recharge. Although, the soil is fairly well drained, seasonal frost of a thickness near 2 m does develop over the winter season, thus potentially reducing the hydraulic conductivity and subsequent groundwater recharge.

FIELD SITES

Two locations were selected for instrumentation: a permafrost site and a permafrost-free site. Both sites are located approximately 10 km north of Fairbanks, Alaska. Site one is a permafrost-free south-facing site at an elevation of 950 feet mean sea level (msl) (260 m) with a predominantly white birch-aspen (*Betula papyrifera-Populus tremuloides*) forest. This well-drained site has a litter layer of approximately 10 cm underlain by fine-grained silty loams. The thickness of the silt loam varies from zero where rock outcrops occur to several hundred feet, depending upon the local topography. Birch Creek schist is the predominating bedrock type in this area. The slope in the vicinity of the study area averages about 10 per cent.

Site two is a permafrost site located on a southeast-facing slope. Black spruce (*Picea mariana*) dominates the overstory vegetation. Unlike site one, this site is poorly drained with a thick organic layer of moss and lichens; maximum thicknesses are near 30-40 cm for this layer.

Permanently frozen silty loams are found below the moss layer. The thickness of the active layer is a function of the moisture content; however, the average late summer thickness in this area is approximately 1 m. The selected site was at an elevation of 800 ft msl (240 m) with an average slope of 5 per cent, draining in a southeast direction. Birch Creek schist is again the predominant bedrock type.

INSTRUMENTATION AND PROCEDURES

The procedure used in this study was to monitor the rate of ablation of the snowpack and follow changes in the soil system over the summer months until freeze-up of the next winter season.

Prior to the spring break-up, numerous snow surveys were made to determine the water content of the snowpack. An Adirondack snow sampler was used to determine the snow depth and water content. This sampler, with a 6.6-cm core diameter, works exceptionally well in shallow, light snowpacks. During the ablation period, these same surveys were continued; the only alteration was that the surveys were made on a daily basis and the number of sample points was increased. The difficulty of not recovering a complete snow column in the samples was solved by coring into the organic layer. A few centimeters of organic material were sufficient to ensure that the entire snow column was collected; the organic material was removed prior to weighing the sample. As the snow ablation proceeds, the first areas to become snow-free are at the bases of the trees. The diameter of the snow-free areas increase until they intercept other expanding snow-free areas. A substantial amount of melting occurs at the circumference of these circular areas.

While the other facets of the project were being studied, during the rainfall season, precipitation was measured by an 8-inch standard rain gage and a standard tipping-bucket gage connected to a continuous-event recorder.

Numerous measurements were made to detect changes in the soil system. During the winter and preceding the breakup, soil temperatures were monitored by a large number of thermistors installed the previous summer. The thermistors were installed in two vertical arrays, one spaced at 10 cm and the other spaced at 20 cm. These arrays were continued above the ground surface up through the accumulating snowpack at the same interval. Dowel rods were used to hold the thermistors in position and were painted white. The wiring was wrapped in a white tape to minimize the absorption of incoming short-wave radiation. The thermistor data was useful in assessing how rapidly the snowpack reached ripe conditions, the maximum extent of the seasonal frost, and the rate of thaw of the soil following ablation. The thermistors also allow an almost completely undisturbed measurement of soil temperatures, because they were placed during the summer season and were "weathered" into place. At the permafrost-free site additional thermistors were installed at 150 cm and 200 cm, since the maximum depth of seasonal frost was near that depth. Frost tubes were also used to assess the rate of development of the seasonal frost.

Although frost tubes are less reliable in indicating the depth of the freezing interface than are temperature measurements, they do provide a comparison and a duplication in the event of thermistor failure. Probing was used to obtain additional information on the rate of thaw. Probes were made with a stainless steel soil probe, which has a centimeter scale embossed on it. A series of 5 probes were made on each sampling date and averaged to give a mean depth to frost for that date.

Point measurements, such as the temperature measurements, present a good picture of the seasonal variation; however, spatial variations are often poorly described. Soil columns were recovered at various intervals to determine the variation in the moisture content. The difficulty of separating temporal variations from spatial variations for these samples is well understood. Since soil samples cannot be recovered at the same point, transects for sampling were established in a manner that would minimize the setting variation. Also, the samples were recovered in a manner that allowed future sites to remain undisturbed. This was a definite requirement if samples were recovered when a snow cover existed.

A 3-inch diameter motor-driven coring auger was used to obtain the soil samples. Samples were collected at an interval of 5 cm. Generally the samples were collected to a depth of 1 m; however, occasionally samples were collected to a depth of 2.5 m or more. Prior to the snowfall, screens were placed on the ground surface at points where soil cores were to be obtained. This facilitated the removal from the ground surface of the snowcover, so that very little snow sifted down into the surface organic layer at the time of sampling. The mineral soils were oven dried in the laboratory in the standard manner for moisture determinations while the organic soils were dried in a microwave oven. Average values of the bulk density at various depths were determined. This calculation makes it possible to express the moisture content either by weight or by volume. It should be noted that for the organic soils the bulk density was controlled by the depth from the surface as well as by the overall thickness of the organic layer. Considerable variation was found between samples.

Tensiometers were installed to measure pore pressures at various depths. These were intalled at 25-cm intervals at 25, 50, and 75 cm from the ground surface. These tensiometers were read every few days and gave a good indication of the pore pressure gradient. The tensiometers were checked by calibrated vacuum gages. Soil pore pressure measurements were initiated shortly after the snow ablation period in the unfrozen soils. As the depth of thaw increased, additional tensiometers were installed.

In order to measure the rate of infiltration from the snowpack into the soil system, several simple lysimeters were installed. These lysimeters were constructed in a manner similar to the technique outlined by Haupt (1969) and shown in detail in Figure 2. They have an approximate surface area of 1600 cm^2 and were filled with the organic and mineral soils native to the site. At each site, two of the lysimeters were filled with organic material only. The rationale for this approach was that the hydraulic properties of the two soils are drastically different. Since the hydraulic conductivity of the organic material is much greater than that of a mineral soil, saturated conditions may develop in the upper organic soil thereby inducing lateral flow.

For the remaining lysimeter, a layer of 20 cm of mineral soil was placed under the organic layer. A layer of pea-gravel and a coarse screen were used to support the soils in the lysimeter. The outlets of

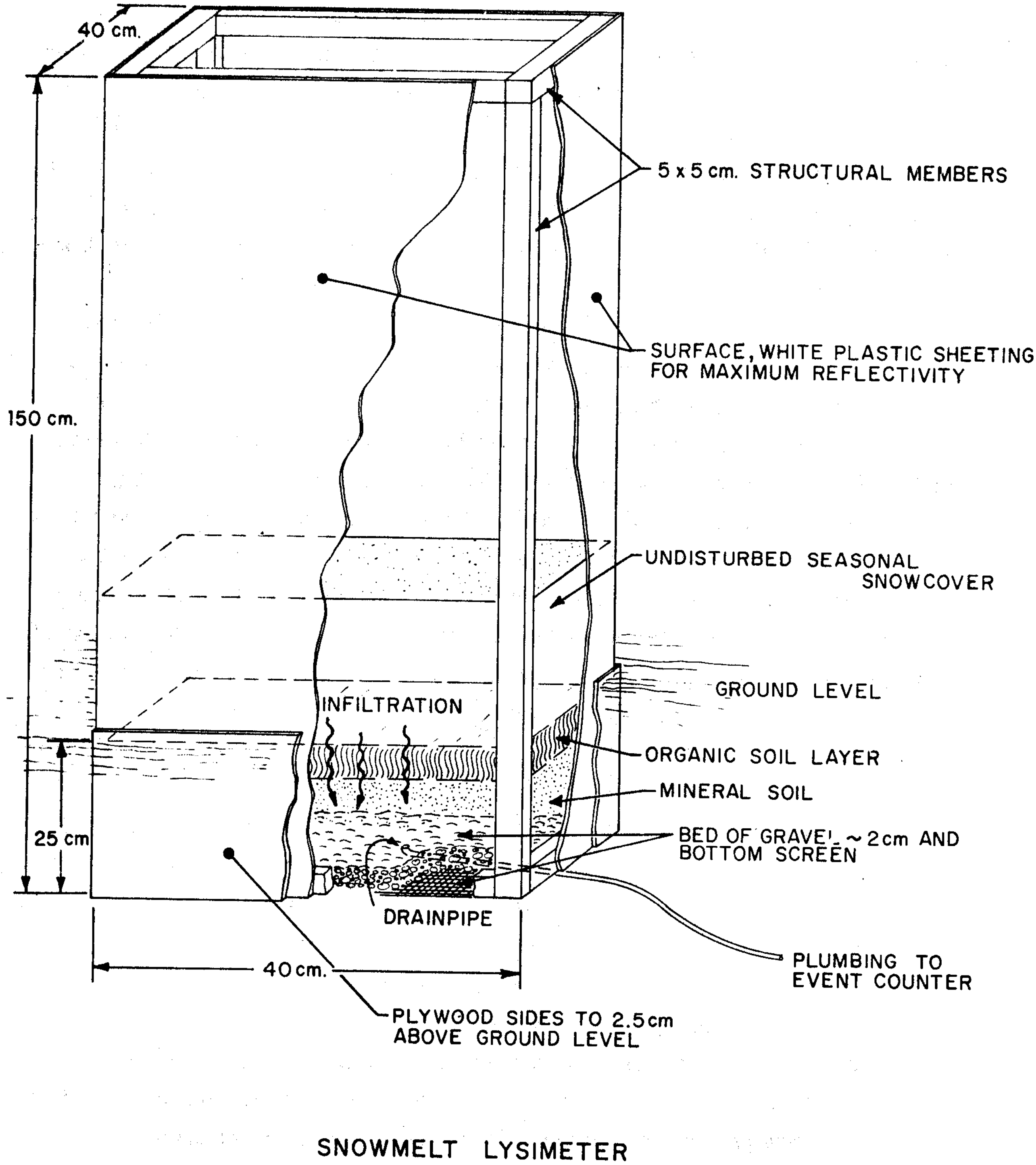

FIGURE 2: Snowmelt Lysimeter Used to Measure the Flow of Snowmelt Water Through Various Soils.

the lysimeters were connected to a tipping-bucket rain gauge. The tipping bucket was in turn attached to an event recorder. This arrangement allowed a calibrated and very accurate measurement of the amount of infiltration water which reached the bottom of each lysimeter. The total measured outflow from the lysimeters with the organic layer can be compared with the total snowpack composition, the difference being water lost to evaporation (or transpiration) and retention in the organic layer. Outflow from the lysimeters containing both organic and mineral soils reflect the additional water lost to soil retention and also indicate the additional time needed for the flow to travel through the thicker layer.

During the second winter when some further data were collected, a fourth lysimeter was installed at the ground surface and contained no soils. The difference between the maximum snowpack content and the volume of flow from this lysimeter would represent water lost to evaporation (or sublimation) only. Although it was intended to collect another year of data from all the lysimeters, the nature of the snowpack ablation precluded it. After a brief warm spell that initiated flow through the lysimeter, it became rather cold causing water to freeze in the plumbing to the tipping-bucket recorders as well as in the bottom of the lysimeter. This cold spell lasted several days and therefore water ponded in the bottom of the lysimeter where it also froze. Once the cold spell was over and water was again moving through the snowpack, the lysimeter was not operating.

A hygrothermograph was used to measure the relative humidity and the ambient temperatures. Also, a solar radiation recorder was installed under the canopy cover to measure incoming solar radiation. A plexiglass dome transmits 90 per cent of the incoming radiation between 0.36 and 2.0 microns. Measurements were continued throughout the summer under a canopy cover of birch and aspen.

During the second year of the investigation, a dye study was initiated to see if some indication of the rate of movement and the flow pattern of the meltwater through the snowpack could be obtained. Various organic dyes were used in this study: malachite green, crystal violet, azure A, and basic fuchsin. The procedure used in this part of the study was to spread the dye on the surface early in the morning prior to significant melt. Then approximately 12 and 24 hours later, a trench would be excavated to obtain a cross-section through the dye. This allowed the cross-section to be photographed as well as permitting a visual inspection of the dye movement. Since these dyes when dissolved are dark in color and readily absorb incoming radiation, duplicate plots were set up each day with the second set of plots being covered with brown cardboard. The purpose of this cover was to intercept all the incoming radiation before it could penetrate the snowpack. Obviously the rate of melt could be increased by the dye on the surface, since the albedo was altered substantially.

RELATED RESEARCH

Several diverse investigations relate both directly and indirectly to this snowmelt-soil interaction study. Most of the large number of related papers concentrate on examining just one aspect of this process, whereas, in this study, the net result of flow from a snowpack into a frozen soil was examined. Studies on such particular aspects as the importance of climatic variables on snowmelt, evaporation and condensation at snow surfaces, flow through snowpacks, and infiltration into frozen soils are a few of the areas wherein numerous detailed studies have been made. Gray and O'Neill (1974) used a lysimeter and made comprehensive radiation measurements in order to use an energy budget concept to determine which energy input factors were critical to the snowmelt process. Net radiation was found to be the most critical source of energy for snowmelt in their case. The resulting flow of water through a snowpack following surface melt is also an important consideration to runoff and infiltration. Colbeck (1974) determined that, for runoff prediction, the important factor to consider is thickness of the saturated snow layer. It is recommended that, for long-term determination of this saturation depth, measurement be accomplished by measuring the dielectric constant by remote sensing.

A problem of snowmelt runoff prediction is inability to measure the evaporation rate of the snow cover. A technique was used by Lemmelä (1972) similar to the approach used in this study, with necessary additional measurements made in order to determine evaporation and condensation from a melting snowpack. From measurements using regression analysis, Lemmelä showed that the most useful variables for understanding snowmelt are degree-day factor and net radiation. For evaporation processes, the variables are relative humidity and saturation deficit.

A comprehensive review of the physical and thermodynamic properties of frozen ground is given in Anderson and Morgenstern (1973). This is a state-of-the-art review paper and contains much reference data.

Many researchers have investigated the physical parameters of soil moisture movement. Jumikis (1973) examined the effect of porosity on the amount of soil water transferred in a freezing silt. Porosity was found to be a controlling factor in determining heat flow, moisture transfer, and all the phenomena related to these factors. Miller (1969) published a treatise on the hydraulic conductivity of soils and their water retention characteristics. Soil texture was mentioned as an important factor in this treatise in affecting the water retention properties. Williams and Burt (1974) describe a method which they used to determine the hydraulic conductivity of frozen soils. Lactose was used as an interface medium to apply pressure to a frozen soil sample in a permeameter, with the hydraulic conductivity being measured as a function of temperature and hydraulic gradient.

Langham (1974) investigated the movement of liquid water in the snowpack by using dyes to trace water movement, a method attempted in the study reported herein. The volume of unfrozen water was determined by centrifuging snow samples.

Numerous mathematical models have been developed for examination of the movement of moisture through frozen soils. This is of interest primarily because this redistribution of moisture influences the hydraulic conductivity. Guymon and Luthin (1974) developed a coupled heat and moisture transport model specifically for arctic soils. The model was tested in trial runs and results indicate it to be a feasible model. Harlan (1973) also formulated a mathematical model incorporating soil properties, and determined the relation of soil type of those soils affected by freezing to the soil-water redistribution. He found that although the general pattern of soil-water redistribution as affected by freezing is similar for all soils in which initial water table depth was 100 and 200 cm, there are significant differences in both the rate and the magnitude of the response. Secondly, soil-water redistribution is increasingly restricted to the upper portion of the profile in closer proximity to the freezing front as the soil texture becomes finer and as initial moisture contents are reduced.

Differences in volume of surface runoff during the snowmelt period at Yellowknife, Northwest Territories, were studied by Landals and Gill (1972). They found that the most important factors in determining the volume and timing of snowmelt runoff in this northern setting are: 1) the various amounts of storage afforded by exposed bedrock and mineral-organic surface materials; and 2) the amount of water restored to this material the previous autumn by rain that immediately precedes active layer freezeback. Brown (1973) also studied the influence of climatic and terrain factors on subarctic soils. In particular, he measured ground temperatures at three locations in the permafrost regions of Canada. It was found that the relationship of snow depth and density to ground temperatures is not straightforward. However mean annual temperatures in permafrost regions were shown to vary over a 2°C range among various types of terrain at any one location. The variations are also reflected in different depths of active layer that may range from less than 0.5 m to more than 2 m.

Snow structure is similar in some ways to soil structure and an analogy can be made between snow and soil in relation to pore pressures in snow during melting. Colbeck (1976a) has investigated this possibility at Barrow and concluded that tensions of meltwater in snow can indeed be used as a predictor for snow state. Peck (1974) examined the effect of snow cover on the upward movement of soil moisture. The important point made in this paper is that the layer of soil near the surface can retain additional moisture above field capacity when subjected to the temperature gradients that are observed under the winter snow cover. Soil moisture under snow cover in temperate regions usually does increase near the surface during winter.

Popov (1972) discusses several soil characteristics as they relate to several hydrologic problems in the USSR, including water storage, water losses, infiltration, surface retention, and water adsorption.

Recognizing the importance of the permeability of frozen soils, Popov states "A well-moistened frozen soil at a temperature of 2-3° below zero (°C) becomes practically impermeable to meltwater. Conversely when it is dry it remains highly permeable. Even impermeable frozen soil can retain from 5 to 15 mm of meltwater due to the presence of free non-capillary pores in the upper layer. These losses, however, should be attributed to surface retention."

Popov (1972) uses an integral equation incorporating capacity type for water adsorption to arrive at total snowmelt runoff as a function of the active melt, the depth of water needed to fill all "capacity types," and antecedent soil moisture conditions. The important parameter to determine is the water adsorption.

Komarov and Makarova (1973) investigated the effect of ice content, temperature, cementation, and freezing depth on meltwater infiltration in a basin. These authors maintain that it has now been established that slightly frozen soil absorbs meltwater. Curves are given relating the variation of soil freezing depth of less than 15 cm as a function of the average depth of freezing. This provides an index of the probability of meltwater infiltration. The importance of the spatial distribution of moisture in the upper soil layer is also a strong influence on meltwater infiltration.

Many studies in Alaska and Canada relate to this work as well. A black spruce taiga ecosystem study managed by the U. S. Forest Service at a site 17 miles north of Fairbanks is presently proceeding. The study includes a comprehensive ecosystems study and is intended to investigate thoroughly the effects of fire in the taiga. Included is a study of water movement in the top 60 cm of soil (the active layer), as well as measurements of the soil tensions in the organic soil layer.

DISCUSSION OF DATA

Temperature Regimen

Thermistors were installed at the two sites to define the location of the 0°C isotherm both for either identifying frozen or unfrozen zones and monitoring temperature variations. At the permafrost site, both the permafrost table and the base of the seasonal frost were determined during the freezing stage. If the active layer completely froze during the prior winter, then thawing proceeded down from the ground surface reaching its maximum depth by early August. In our case, the active layer was completely refrozen by late February. In the non-permafrost site, thermistors, frost tubes, and probing were used to define the frozen and thawed zones. The thickness of the active layer at the permafrost site was about 100 cm, whereas the thickness of the seasonal frost at the non-permafrost site was about 220 cm (Figure 3). Although the onset of freezing occurs at the same time at both sites, the rate of freezing is much greater at the non-permafrost site. This is directly related to the moisture content at each site. This topic will be discussed further in the section entitled "Soil Moisture Regimen." During the first year of data collection, the maximum depth of freeze was attained by late February, the same date that the active layer was completely refrozen. During the second year, the depth of freeze was slightly less at the non-permafrost site (Figure 4).

The primary interest in taking soil temperature measurements at the time of ablation was twofold; first, the raw temperature data would give an excellent description of the condition of the soil system prior to the snow ablation period and, second, the temperature at various depths would approach 0°C as the infiltration water moved downward through the soil system. The soil temperature in conjunction with the soil water levels is responsible for influencing the hydraulic conductivity under freezing conditions. Selected temperature profiles are plotted in Figures 5 and 6 for the two sites studied. At the permafrost site, it can be seen that there is very little variation ($\sim$1°C) in the soil temperatures at the 100-cm depth. At the same depth at the non-permafrost site it can be seen that the temperature varies by 12°C from late March to early September. Temperature fluctuations of the same magnitude are shown in the Appendix for 1977 at the non-permafrost site.

In general, at the non-permafrost site the soil is much colder initially when the snow ablation begins, however these soils warm up very rapidly and reach much higher summer temperatures than do those at

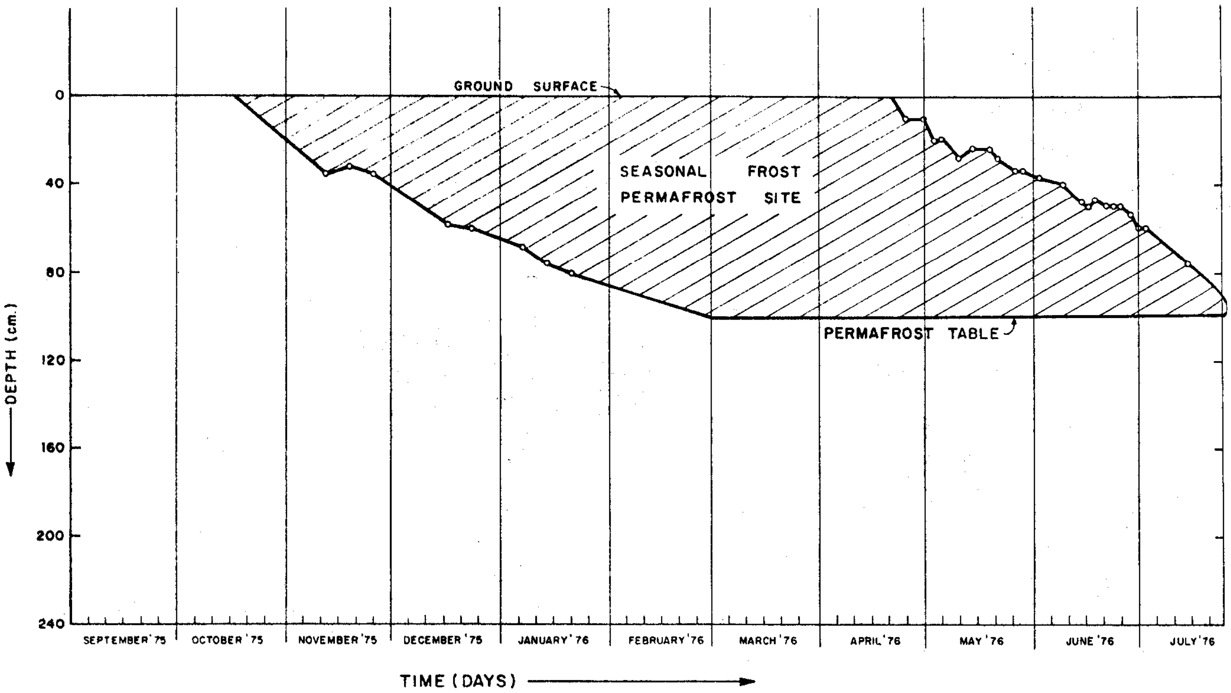

FIGURE 3a: Seasonal Frost Depth vs. Time, Permafrost Site, 9/75 - 7/76.

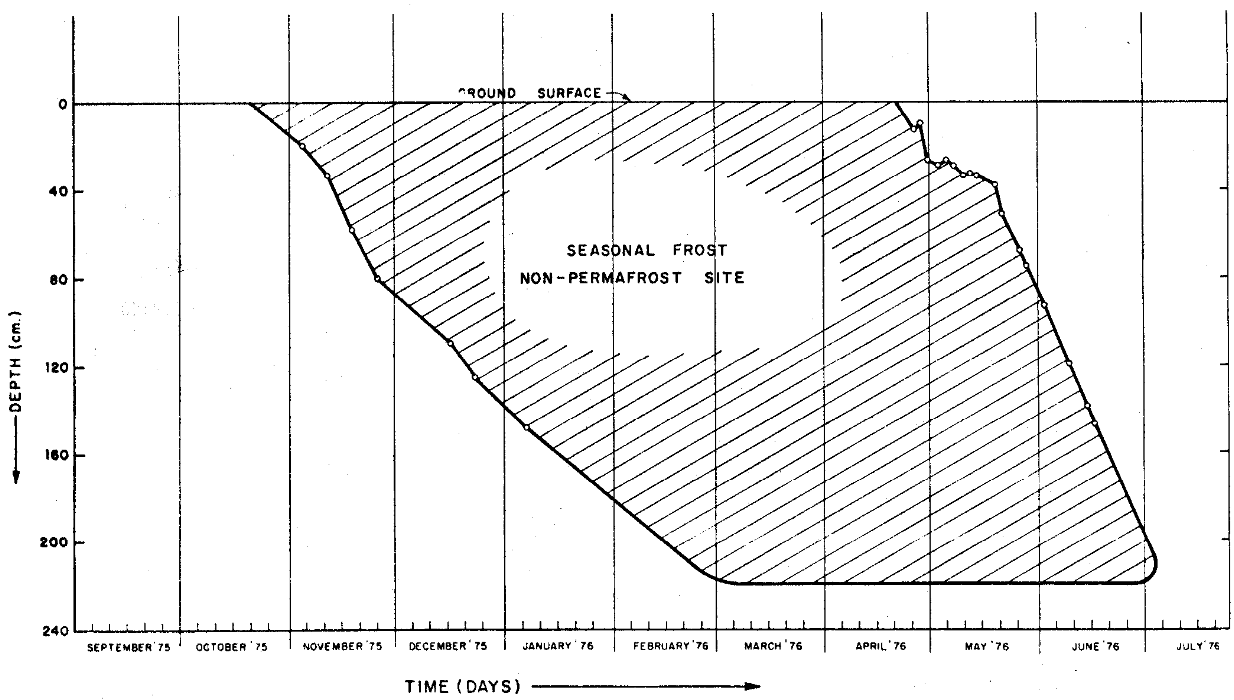

FIGURE 3b: Seasonal Frost Depth vs. Time, Non-Permafrost Site, 9/75 - 7/76

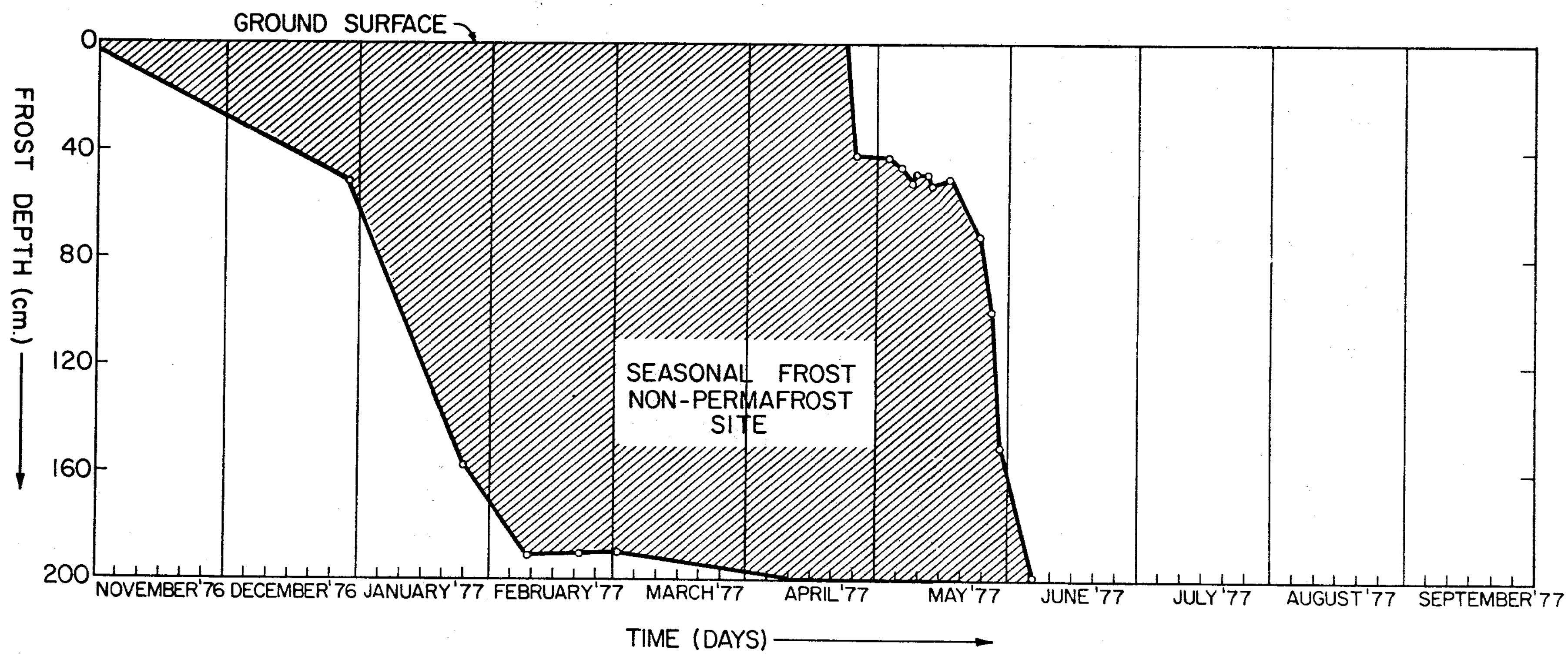

FIGURE 4: Seasonal Frost Depth vs. Time, Non-Permafrost Site, 11/76 - 9/77

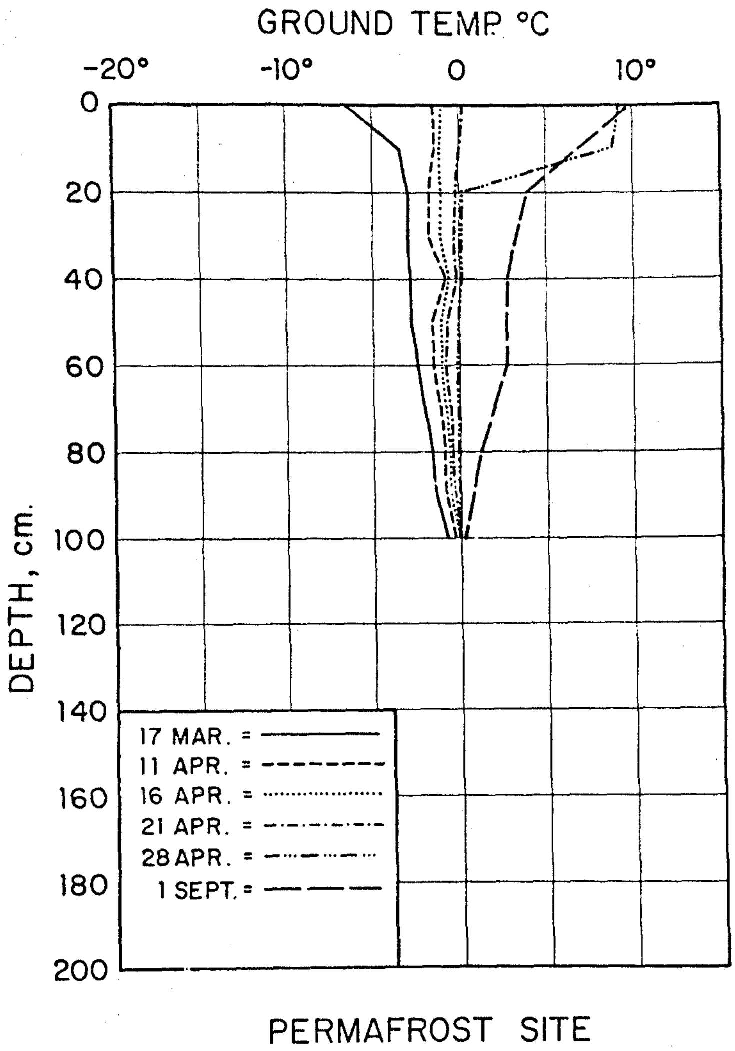

FIGURE 5: Selected Ground Temperature Profiles, Permafrost Site

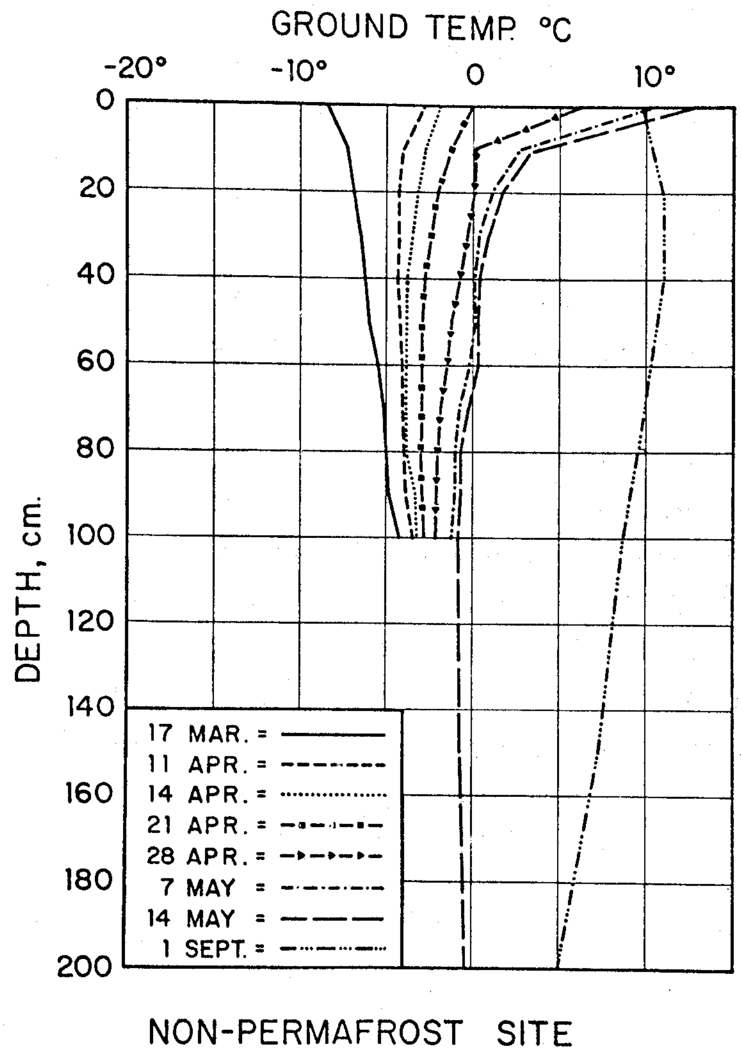

FIGURE 6: Selected Ground Temperature Profiles, Non-Permafrost Site

the permafrost site. The main reason for this is that the amount of energy required to initiate phase change (ice to water) and to warm up the soil at the non-permafrost site is much less because the soil moisture levels are substantially lower. This results in moist soils having both a shallower frost depth and a narrower range in fluctuations of soil temperatures whereas dry soils have just the opposite properties. Another important factor is the insulating quality of the organic soil layer which strongly inhibits downward conduction of summer heat. The buffering influence of this organic layer is somewhat reduced during the winter months because, under the weight of the snow load, this layer is compacted to a thickness substantially less than for summer conditions. Reduction in thicknesses of 50 per cent have been measured.

Climatic Input

In order to determine the amount of water moving through the system, it is essential to know the quantities of input from either snowmelt or summer precipitation. The accumulation of the snowpack for 1975-1976 at the two sites is illustrated in Figure 7a and a similar curve is shown in Figure 7b for 1976-1977 at the non-permafrost site. When the snowpack reached its maximum depth in late March the water content was approximately 10 cm of water in 1976 and 8.8 cm of water in 1977. This data was obtained by taking snow surveys. It can be seen that the snowpacks reached their greatest depth by early March; at that time settlement of the pack took place as the temperatures increased. The data in Figure 8 show the daily incremental values of snowmelt once the ablation started. For 1976 ablation season, it can be seen that melt started on 17 April at a very slow rate and reached a maximum of 2.45 cm/day on 27 April, at which time there was essentially no snow cover remaining. In a period of 11 days, the ablation was initiated and completed without any breaks. It would not be uncommon for cold weather to delay the melt; this occurred in the 1977 breakup. For 1977, the first period of ablation was responsible for reducing the snowpack from about 8.8 cm of water to 4.3 cm of water from April 20 to April 25. From April 19 until April 25 the temperature did not drop below the freezing level and then from April 25 to April 29, the maximum temperature

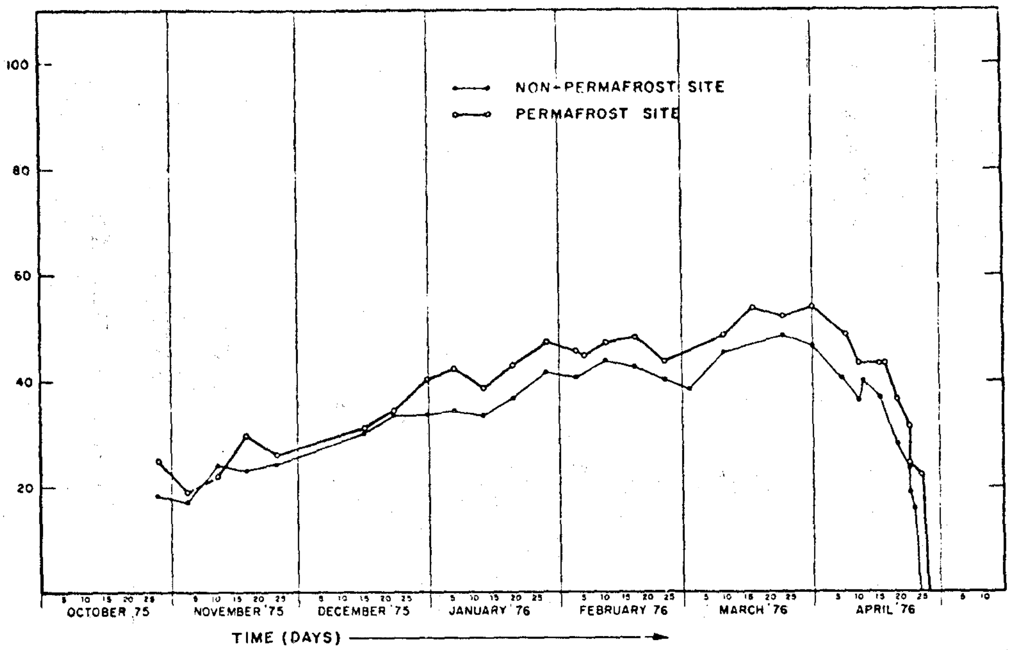

FIGURE 7a: Depth of Snow, 1975-1976.

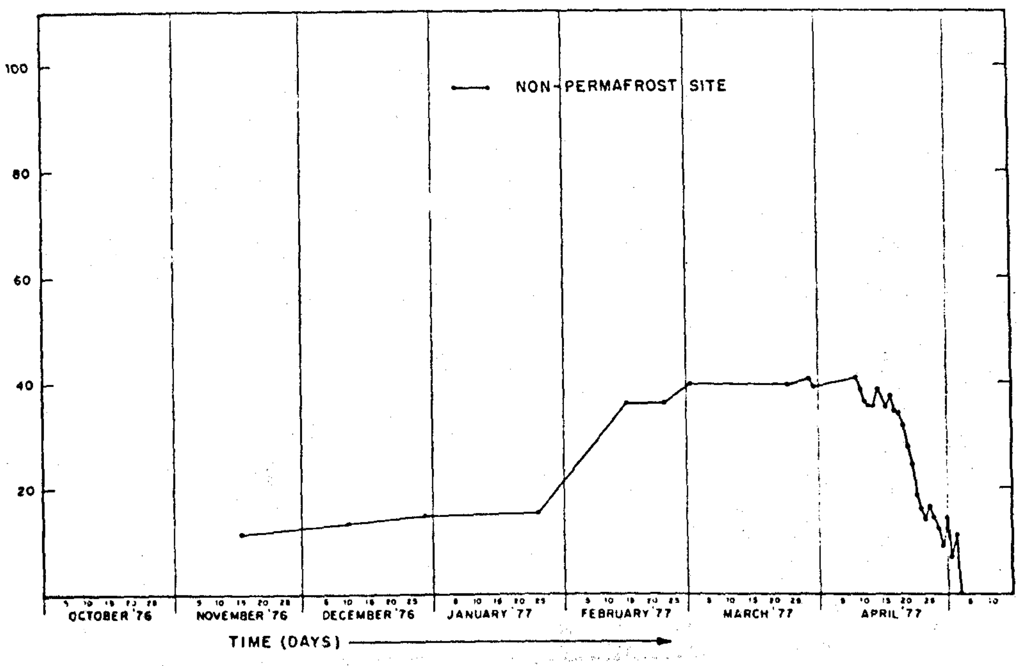

FIGURE 7b: Depth of Snow, 1976-1977.

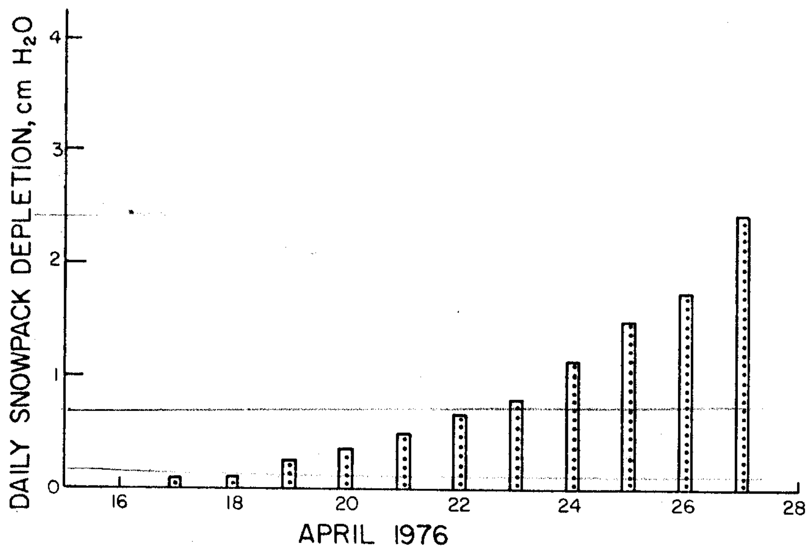

FIGURE 8a: Daily Snowpack Depletion for 1976

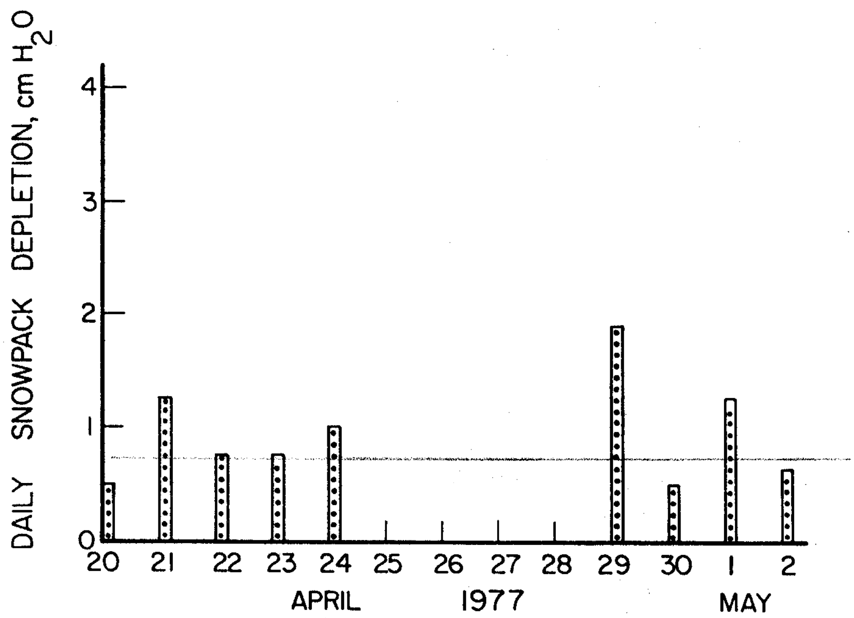

FIGURE 8b: Daily Snowpack Depletion for 1977

hovered around or below the freezing point. The temperature reached 5°C on April 29 and did not drop below the freezing level until the snow cover was completely absent. Once the temperatures consistently stayed above the freezing point, the remaining equivalent 4.3 cm of water disappeared quickly.

In 1977, the overall disappearance of the snowpack was uniform; that is, about 1 cm of water per day was lost from the snowpack as long as above-freezing temperatures existed. This contrasts with 1976 when the melt began very slowly and then the daily melt rate gradually increased toward the end of the ablation period. Temperature graphs during the ablation period are shown in the appendix along with the incoming solar radiation for the late spring and summer months. The radiation data was collected under a canopy cover, therefore, the peak incoming radiation was measured in late May just prior to the emergence of the leaves.

Moisture Regimen

The emphasis of this study was to examine the infiltration of the water from the snowpack into the seasonally frozen ground. Numerous soil columns were collected over the winter months prior to the snowmelt ablation to monitor changes in the moisture levels in response to the temperature gradients that existed. Numerous samples were collected during the ablation period and the interval between sample collection increased as the changes in the moisture contents became minimal. During the winter, soil samples were collected monthly; as the ablation period approached, they were collected weekly; and, finally, during the ablation period, they were collected every 2-3 days. For the remainder of the summer they were collected weekly. The results of this data are published in the appendix, with selected profiles plotted in Figures 9 and 10. It should be noted that the moisture content as defined for frozen solids includes both the ice and water.

Considerable variation exists in moisture regimen between the two sites studied even though the mineral soil is of the same composition. The initial moisture content at the non-permafrost site in the mineral soil is much lower, primarily due to the vertical drainage that occurs in these soils. Where permafrost impedes vertical drainage, such as at

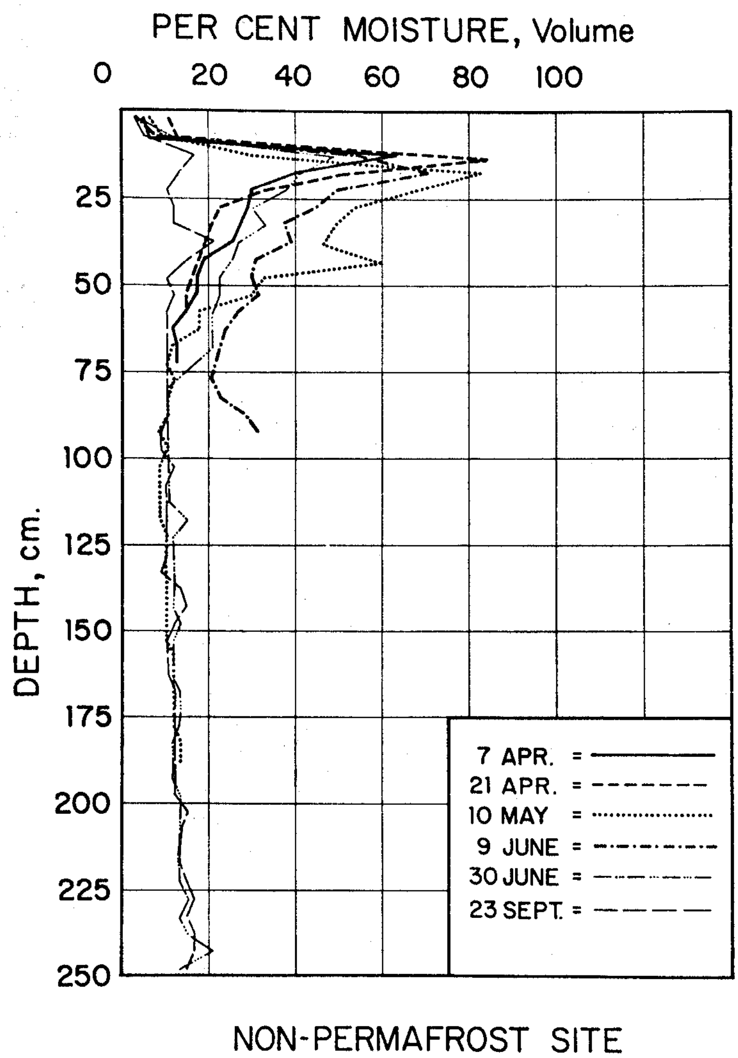

FIGURE 9: Selected Soil Moisture Profiles at Non-Permafrost Site.

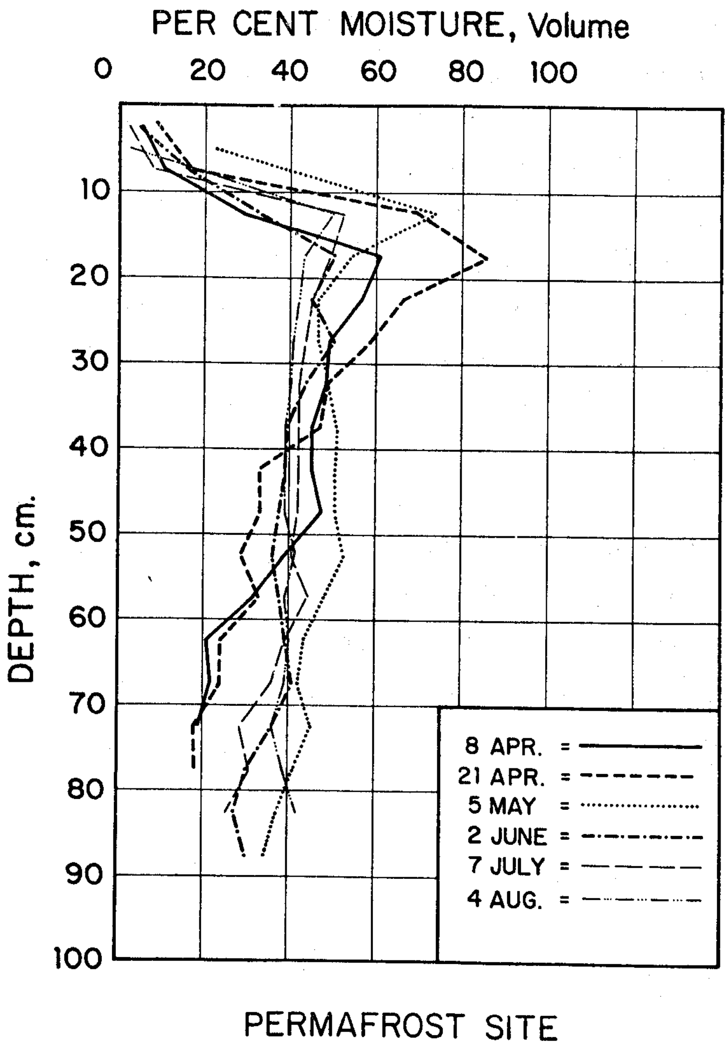

FIGURE 10: Selected Soil Moisture Profiles at Permafrost Site.

the permafrost site studied, higher initial moisture contents prevail prior to snow ablation. These higher moisture levels above the permafrost also prevent the infiltration of meltwater into the seasonally frozen mineral layer because when this water is frozen, large volumes of the pores are filled with ice. Subsequently, the water moves downslope at the interface of the organic-mineral soils.

At the permafrost-free site, the moisture contents are much lower (12-20 per cent vs. 20-40 per cent) and, although frozen, these moisture contents do not prevent the infiltration of meltwater. For the 1976 season, with an equivalent of 10 cm of water, and the 1977 season, with an equivalent of 8.5 cm of water in the snowpack, no water was ever observed ponded on the surface or moving downslope on the ground surface or at the interface between the different soil layers. So either the water moved vertically down through the soil column (organic and mineral), evaporated from the snow or ground surface, or was retained in the soil column.

As can be seen from the plotted profiles, the major changes in the permafrost site occurred at a depth of about 20 cm, this is about the depth to the interface between the organic layer and the mineral soil. Initially, at the permafrost-free site, this same cycle is observed at a shallower depth (near 10 cm). However, later an increase in the moisture content at depth was observed.

Due to spatial variations in soil moisture content, the exact determination of the volume of water movement can be very difficult from field samples. Because of this difficulty, other field variables are measured that result in the calculation of moisture movement. The one-dimensional Darcian equation of steady flow is:

$$q = -K(\psi)\Delta H \qquad (1)$$

where

q = flow/unit area

$K(\psi)$= unsaturated hydraulic conductivity (L/T)

ψ = pore pressure (L)

H = hydraulic head, both pore pressure and gravitational component (dimensionless)

To solve this equation for a unit area, the relationship between hydraulic conductivity and the pore pressure is needed if unsaturated conditions exist. The field pore pressure can be measured using tensiometers. Often, to get the range of values of interest quickly, it is convenient to examine the relationship between pore pressure and moisture content in the laboratory.

The determination of the unsaturated hydraulic conductivity from soil retention curves is discussed in a number of papers (Jackson et al., 1965; Green and Corey, 1971; Kunze et al., 1968; Marshall, 1958; and Millington and Quirk, 1959, 1960, 1961). The modified form of the Millington and Quirk equation was used in this study to determine the unsaturated hydraulic conductivity:

$$K(\theta)_i = \frac{K_s}{K_{sc}} * \frac{30\ \gamma^2}{\rho\ g\ \eta} * \frac{\varepsilon^P}{n^2} \sum_{j=i}^{m} [(2j + 1 - 2i)h_j^2] \qquad (2)$$

where

$K(\theta)_i$ = the calculated unsaturated hydraulic conductivity for a specified water content or pressure (L/T)

θ = the water content by volume (dimensionless)

i = the water content class, i=1 defines the water content class corresponding to saturated conditions and i=m defines the lowest water content class for which the unsaturated hydraulic conductivity is determined (dimensionless), $i = 1, 2, \ldots m$.

K_{sc} = the calculated saturated hydraulic conductivity in the above equation where $K_{sc}/K_s = 1$ (L/T)

K_s = measured saturated hydraulic conductivity (L/T)

γ = the surface tension (M/T^2)

ρ = the density of water (M/L^3)

g = the gravitational constant (L/T^2)

η = the viscosity of water (M/LT)

ε = the water saturated porosity

P = parameter $1 \leq P \leq 2$

n = total number of water content classes between $\theta=0$ and θ_s (saturated water content)

m = total number of water content classes over range of interest, θ_i to θ_s, $n \geq m$

h_j = pressure for a given water class (L)

Once the relationships between the pore pressure and the unsaturated hydraulic conductivity and the pore pressure and the moisture content are obtained, solutions to equation (1) can be obtained. From tensiometers placed at two vertically separated points, the hydraulic head can be determined; and, utilizing the pore pressure data from the tensiometers, the unsaturated hydraulic conductivity can be determined. The product of these two values over a given time increment defines the moisture flux occurring for given conditions. Changes in the moisture content generally are not great over a 1-day period; so, in order to determine the net vertical water movement following the ablation period, average daily values can be used.

These same relationships can be used in the two dimensional form of the unsteady flow equation:

$$\frac{\partial}{\partial x} K(\psi) \frac{\partial (H)}{\partial x} + \frac{\partial}{\partial y} \frac{K(\psi)\, \partial (H)}{\partial y} = \frac{\partial \theta}{\partial \psi} \frac{\partial \psi}{\partial t} \qquad (3)$$

where

t = time (T)

all other terms identified in earlier equations

With relationships between moisture content (θ) vs pore pressure (ψ) and hydraulic conductivity $K(\psi)$ vs pore pressure (ψ), this equation, once initial conditions are identified will yield the pore pressure for grid points of interest at various specified times. Using the ADI (Alternating Direction Implicit) method with measured field values of initial moisture content and measured values of daily snowmelt ablation as boundary flux, pore pressures were determined. The moisture content of the seasonally frozen soil was quite low, so initially it was assumed that the characteristic curve (moisture content vs pore pressure) would be only slightly influenced and therefore the unsaturated hydraulic conductivity would be quite similar to the relationship developed for unfrozen soils. This would not be true for soils that are just moderately wet prior to freezing; in this case, ice lenses in the soil pores would alter the hydraulic conductivity.

The pore pressures determined from the above equation with the snowmelt input on a non-permafrost site were all less than atmospheric. This means that saturated conditions never developed in the simulated run. The same conclusion was observed in the field at the non-permafrost site; saturated conditions were never observed in the field. This was determined by the field soil moisture profiles and the fact that no flow was ever observed flowing into trenches cut into the soil that were perpendicular to the slope.

By nonlinear, least-squares curve-fitting, the following relationship was determined for relating the unsaturated hydraulic conductivity to the measured pore pressure:

$$K(\psi) = Ae^{-B\psi} + Ce^{-D\psi} \tag{4}$$

where

$A = 5.67 \times 10^{-3}$

$B = 1.75 \times 10^{-2}$

$C = 1.44 \times 10^{-2}$

$D = 3.89 \times 10^{-3}$

When the pore pressure (ψ) equals zero, A + C equals the saturated hydraulic conductivity.

Several tensiometers were installed as the seasonally frozen ground thawed following the ablation of the snow. These measurements were continued through the summer. Due to horizontal variations in moisture content at the same depth, it is often difficult to readily recognize when the tensiometers are not working properly. Also, early in the spring, the temperature may drop below freezing at night for several weeks following ablation. This tends to make the continuous operation of water-filled tensiometers very difficult.

As the ground thawed, the tensiometers were installed at intervals of 25 cm. At the 25-cm depth, from late April until early June, near-saturation conditions prevailed (Figure 11). In early June, the tensions started to increase rapidly from less than 100 cm of H_2O to nearly 800 cm of H_2O by 23 June. Some of the water during this time was lost to vertical infiltration and a large percentage was probably lost to

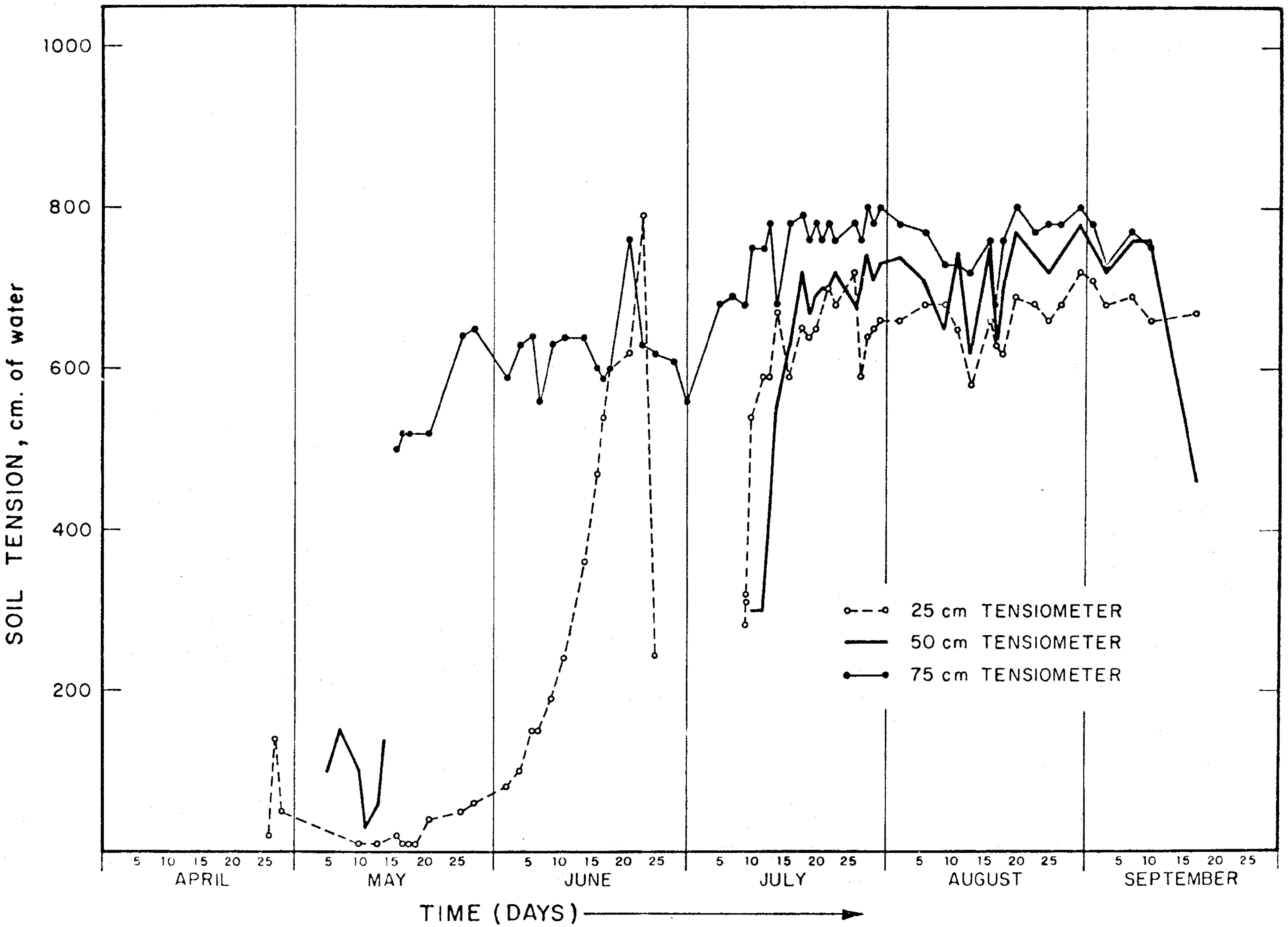

FIGURE 11: Field Soil Tension vs. Time

evapotranspiration. The measured summer precipitation is illustrated in the appendix. As can be seen from the total accumulation over the summer (May through August), only 11 1/2 cm of water was measured in the rain gauge located under the canopy cover. The general pattern of precipitation consisted of events of light rainfall; only three times did the precipitation approach a rate of 1 cm/day. Because of the nature of this summer rainfall, very little change was observed in the soil moisture level from late June through the middle of September.

Soil tensions at a depth of 50 cm were initially low (≈100 cm of H_2O) immediately following the ablation period. Gradually the soil at this depth released water until the soil tensions approached those of the 75-cm depth. Soil tensions were initially much higher at the 75-cm depth compared to the other depths. Values of 500 cm of H_2O were measured in mid-May at 75 cm; these increased to 600 cm of H_2O through June. In June, the levels quickly jumped up to 800 cm of H_2O where they remained for the rest of the summer.

Lysimeter Results

The rationale for utilizing lysimeters was that a good indication of the potential groundwater recharge from the snowpack ablation could be obtained. The status of the snowpack was monitored by daily snow surveys that were made early each morning. The difference between the maximum snowpack accumulation and the existing snow content was determined and is plotted in Figures 12 and 13. In 1976, the snowpack had an initial water content of approximately 10 cm and, in 1977, the content was slightly less, 8.8 cm. Channel 3 on both the above figures represents the lysimeter with only a layer of organic material for the meltwater to flow through. Prior to the melt, the moisture content of the organic material was very low. It is reasonable to expect a percentage of the water to be retained in this layer as well as lost to the atmosphere. The percentages of the initial snowpack content flowing through the channel 3 lysimeter were 55 per cent in 1976 and 62 per cent in 1977.

Channel 4 lysimeter was installed the second year to determine the immediate losses to the atmosphere; for 1977, the measured losses were 17% of the total snowpack. These figures would indicate that 21 per

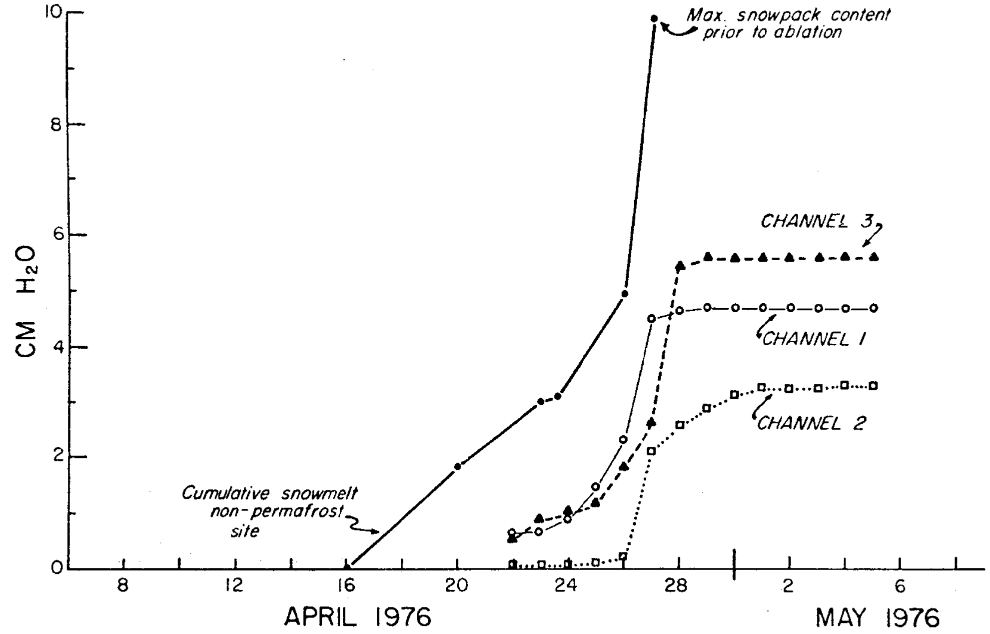

FIGURE 12: Curves of the Snowpack Ablation Rate and the Outflow Rate from Lysimeters, 1976.

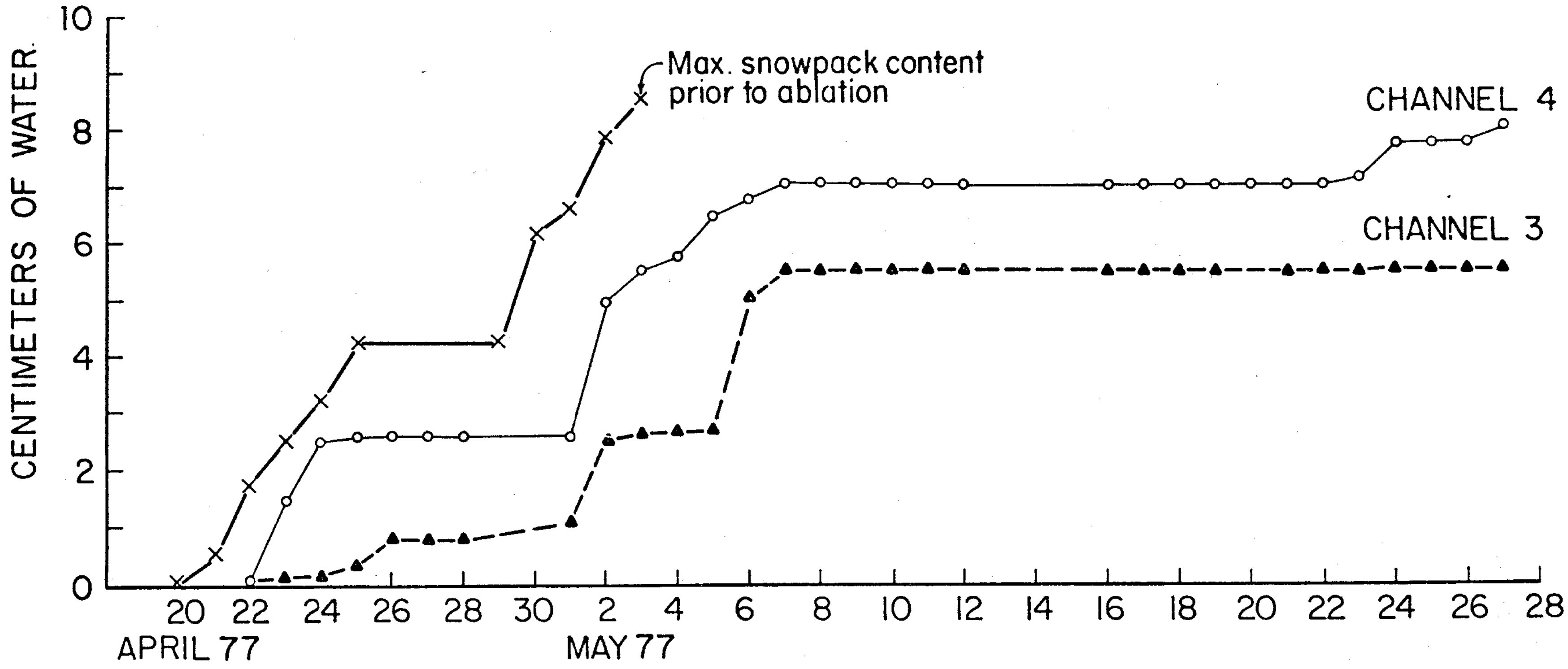

FIGURE 13: Curves of Snow Ablation Rate and Outflow Rate from Lysimeters, 1977.

cent of the snowpack was retained in the organic layer and was lost later to either evapotranspiration or, possibly, to downward flow through the unsaturated soils.

The results from lysimeter 1 for 1976 suggest that only 47 per cent of the melt water filtered downward through the lysimeter. It should be recalled that this lysimeter was constructed in the same manner as lysimeter 3 (organic material only).

The remaining lysimeter had, in addition to the organic layer, a 20-cm layer of mineral soil, and is denoted as channel 2 in Figure 12. About 35 per cent of the initial snowpack in 1976 infiltrated down through the seasonally frozen soil in the lysimeter. The difference between the lysimeters with just the organic material and the lysimeter with the mineral soil ranges from 12 to 20 per cent.

Water movement through the seasonally frozen soils was observed in the soil cores collected. A saturated or near-saturated zone developed in the top of the mineral layer shortly after the ablation season and, as this water moved downward, the moisture difference between the melted soil and unmelted soil gradually declined until a visual difference could not be detected. This was never more than a depth of 75 cm.

It was assumed, because of the shallow root system, that, once the water reached a depth of 50 cm, it was likely to continue its downward movement. Measured pore pressures during the summer indicate that the general hydraulic gradient is downward for most of the summer. Reverse gradients may exist above the 25-cm depth during very dry periods, but from the 25-cm depth to the 50-cm depth, measurements indicate that the predominant direction of soil water movement is downward.

Dye Studies

Very little information is available on the rate of movement of snowmelt water through the cold snowpacks. So, during the spring of 1977, several plots were set up at various times with dye applications on the surface. It was anticipated that, with the large snow grain size that existed and especially the recrystallization known as hoar-frost, water retention capabilities of the snowpack would be low. Colbeck (1976a) had made pore pressure measurements in a shallow snowpack near

Barrow, Alaska. The low snow tensions that he reported lead him to conclude the gravity gradient, not the pore pressure gradient, was the main force in vertical drainage. Positive pore pressures were measured by Colbeck just above the ground surface in the snowpack during what he describes as a heavy melting period. It is improbable that saturated conditions could exist in a snowpack unless the flow were restricted either by an impermeable layer at the snowpack-soil interface or at the ice lenses within the snowpack.

Within this study, a ripe snowpack was not observed until the last 2 to 3 days of melt. Before these snowpack conditions existed, water was measured flowing out of the lysimeters indicating that a substantial amount of meltwater was moving through the snowpack.

As stated in the procedure section, dye was applied to numerous plots early in the morning with observations being made after about 12 hours and 24 hours. One-half of the plots were covered with cardboard to reduce the impact of solar radiation absorbed by the dye which would otherwise have accelerated melt in the pathway of the dye. Cross-sections were cut through the dyed areas to measure the penetration of the dye.

A fairly uniform dye front was observed below the surface from 2 to 6 cm. There was no significant difference in the covered sites compared to the uncovered sites. Early in the melt season, the maximum depth of uniform penetration was measured. This may have been in response to pore (tension) gradients. Readings from the plots that were observed after 24 hours (instead of 12) showed only a slight change in the position of the uniform dye front.

Of more interest were the less uniform paths of travel through the snowpack. When the cuts through the snowpack were made, small pathways could be observed. These pathways are depicted in the conceptual drawing in Figure 14. Many of these pathways appeared to be only columns the size of a pencil. However, if one traced a pathway back into the snow, it could eventually be traced through a tortuous route to the surface or further down into the snowpack. These preferred pathways appeared to drain areas on the surface of the snowpack.

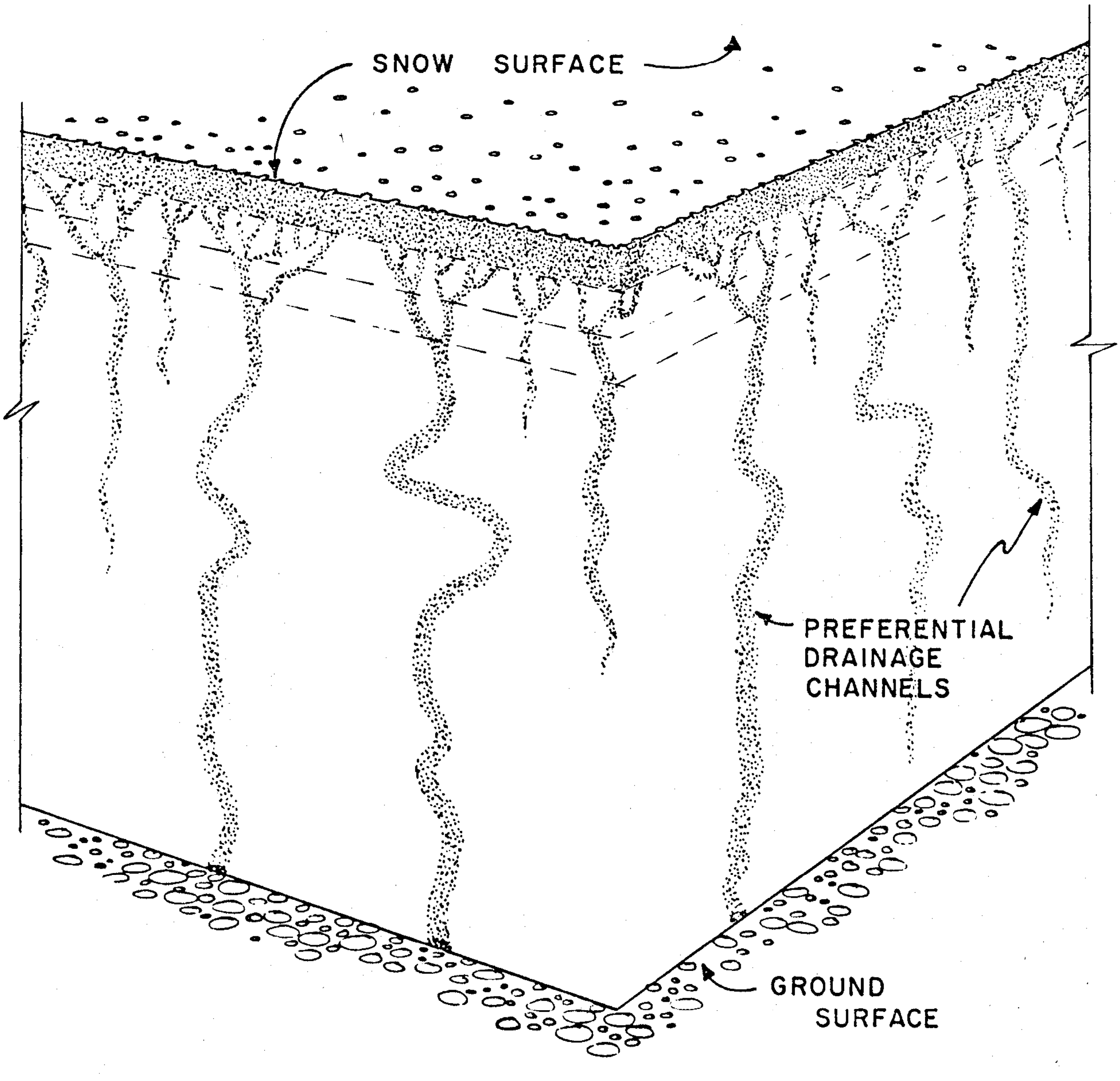

FIGURE 14: Conceptual Representation of Water Movement Through a Relatively Dry Snow Pack

No attempt was made to quantify these observations, but more careful observations were made once it was realized that a "drainage network" was forming. After 5 days, a cold spell brought a complete halt to the melt. During this time the snowpack refroze with sufficient moisture to cement the grains together and intact snow cores could be obtained. In these cores, open channels the diameter of pencils were observed.

The results of this phase of the study lead to the conclusion that although some water may move through the snowpack in a uniform manner once ripe conditions exist, prior to this time almost all of the water moved through the lower snowpack by way of these preferred flow channels. These observations are not new nor are they unique for cold, dry snowpacks. Gerdel (1954) describes a similar process:

> Both vertical and horizontal internal drainage channels develop rapidly in a seasonal snowpack under the action of solution and hydrostatic pressure. The vertical channels are similar to the pipes or "ice glands" which develop in the firm snow on a glacier.

These observations were made while water was artificially applied to a deep snowpack at rates far greater than the natural ablation rates observed in this study. Surface subsidence was observed where vertical drainage was occurring leaving a pock-marked surface. Similar observations were made by Wankiewicz (1976) in a deep, wet snowpack in British Columbia:

> Both kinds of experiment point out that liquid water flow in snow is not homogeneous but follows horizontal structures laterally and often flows in separated vertical zones. The large local variability of liquid water content plus experimental difficulties in distinguishing the phases in snow made representative measurements of the small changes in liquid water content very difficult.
>
> Tiny ice glands, a centimeter or less across could be found at some depths. The ice glands seemed to have a relatively high ice content, to be roughly cylindrical in shape and to follow a somewhat wandering route vertically through a snow layer. The layer of snow between ice sheet #33 and #34 was unusual in having a "close forest" of 1-cm

> diameter glands, about 1.5 cm apart (between adjacent gland walls) with each gland extending the width of the layer. These glands were sufficiently solid that individuals could be physically separated from the surrounding snow, just like ice sheets could be. All sheets and glands which were so isolated were quite permeable, air could be blown through them and the large pores could be seen (and heard) draining after they had been immersed in water.
>
> Between the ice layers can be seen two kinds of flow patterns. The first kind is the finger flow pattern. When a large amount of water enters snow, especially if that snow is dry or contains ice glands, the water will flow as narrow trickles. The trickles have a characteristic wandering appearance, often branching. Indeed the ice glands themselves may have been formed from partial freezing of trickles entering a cold pack, since they look like frozen trickles of flow.

These statements all lead to the same conclusion: that, regardless of the condition of the snowpack, preferential pathways of vertical drainage exist in the snowpack. Although we did not observe any ice glands in the snowpack, it is likely that sufficient cooling time did not exist in the shallow snowpacks to induce a phase change; however, an ice mound was observed at the base of one of these drainage features. Apparently, from the observations of Wankiewicz (1976), these ice glands appear to be very permeable. These observations tend to make spurious the application of mathematical flow models where homogeneity of the porous media is assumed. If, in fact, phenomenological equations are used to model this sytem, the use of more complex equations will be needed to model the real processes.

CONCLUSIONS

The need to build an understanding of groundwater dynamics for permafrost environments is present. Many generalities have been made from the few studies performed, however, the need exists to more precisely quantify related processes. Where permafrost is continuous, usable groundwater aquifers are nonexistent and where permafrost is discontinuous, the amount of recharge which actually reaches the aquifer in use needs to be determined. Approximately one half of the precipitation in interior Alaska falls as snow and accumulates during the winter months. The snowmelt occurs over a period of about 10-16 days with large volumes of water being available for runoff, evaporation, and groundwater recharge. Because of the apparent potential for groundwater recharge during this period, a field study was performed in an effort to define the volume of recharge with the following conclusions:

1. If it is assumed that all of the snowmelt water that flowed through the deeper lysimeter eventually reached the groundwater table, for 1976, the areal recharge would be 3.5 cm (1.35 inches or 36,000 gallons/acre). Since lysimeter 2 was inoperable in 1977, the same estimate cannot be made. The snowpack was lighter in 1977, but the percentage of water loss flowing through lysimeter number 3 was also less--45 per cent for 1976 vs. 36 per cent for 1977.
2. No snowmelt water was ever observed flowing downslope on the ground surface or at the organic-mineral soil interface for the nonpermafrost site. Although the seasonal frost extended to a depth of 2 m or more, the hydraulic conductivity of this well-drained frozen soil was sufficient to accept all the meltwater from the snowpack. Contrary to these observations, at the permafrost site large quantities of water were observed moving downslope through the organic layer. Higher initial moisture contents were responsible for sufficiently reducing the hydraulic conductivity in the mineral soil such that saturated conditions existed above the interface.
3. Results from the lysimeters for the summer of 1976 revealed no groundwater recharge whatsoever occurred from summer rainfall. Although summer precipitation as rain approached 11.5 cm (≈4.5 inches) individual rainfall events were light and uniformly spaced over the summer months. By mid-June the soil tensiometers indicated that relatively dry conditions existed in the soil system. Rainfall during the summer period only temporarily altered the soil tensions near the surface. Evapotranspiration was responsible for removing this water, therefore the soil tensions remained relatively high.

4. The movement of meltwater through the snowpack was shown not to be uniform, instead preferred pathways were delineated by the dyes. The same results had been observed in a variety of snowpacks in more temperate climates. It is likely that this type of drainage is typical of all snowpacks. Due to the colder climate, it would be expected that the snowpacks examined in this study would be more homogeneous than temperate ones.
5. Also from the lysimeter data, it was found that the snowpack exhibited very poor water retention properties. The length of time for the meltwater to move through these shallow snowpacks was quite short. Outflow from the lysimeters was observed for several days before the snowpack could be described as ripe. The development of large snow grains (hoarfrost) is probably responsible for the poor retention properties in the lower snowpack.

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APPENDIX

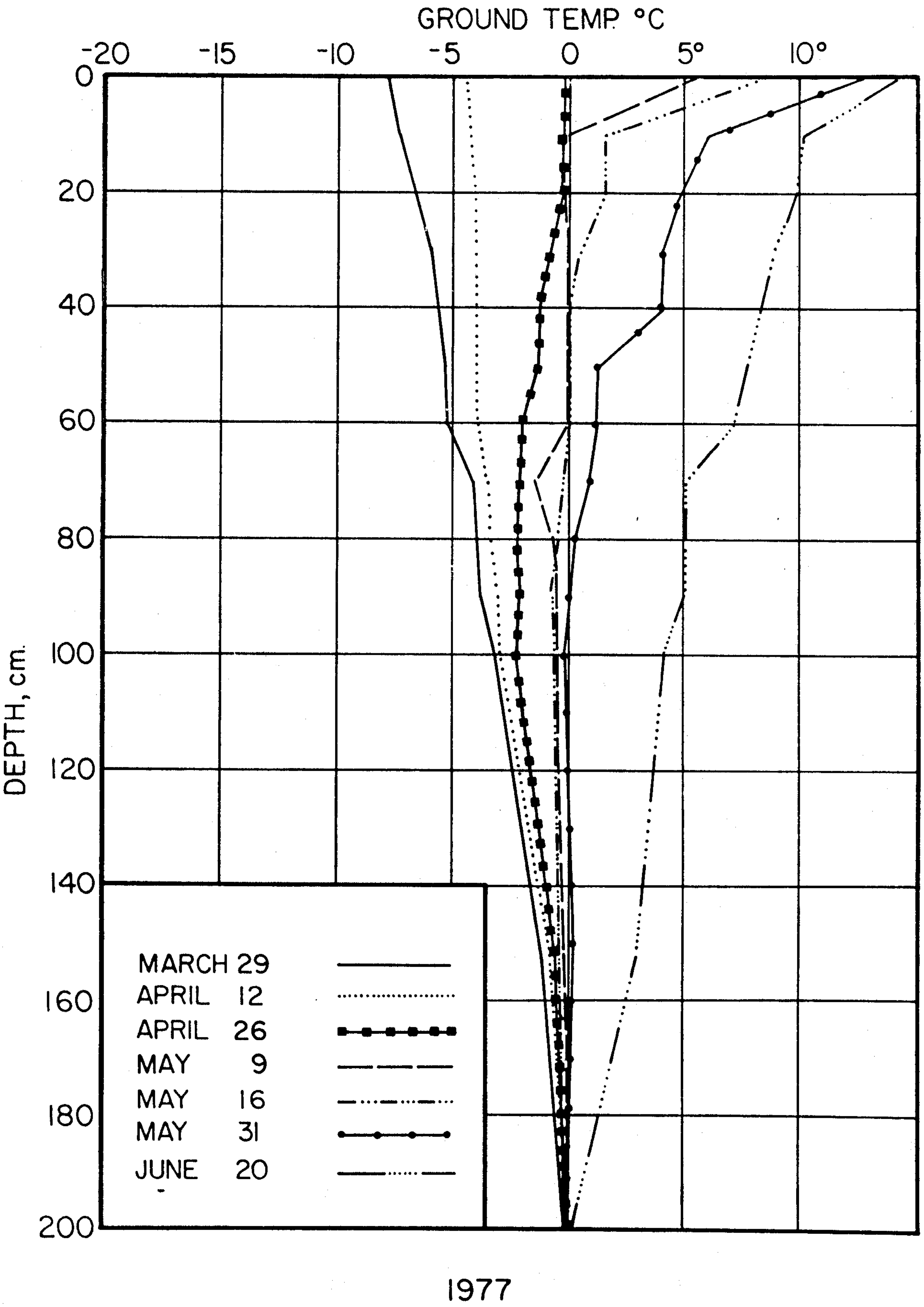


FIGURE A-1: Selected Soil Temperature Profiles at Non-Permafrost Site.

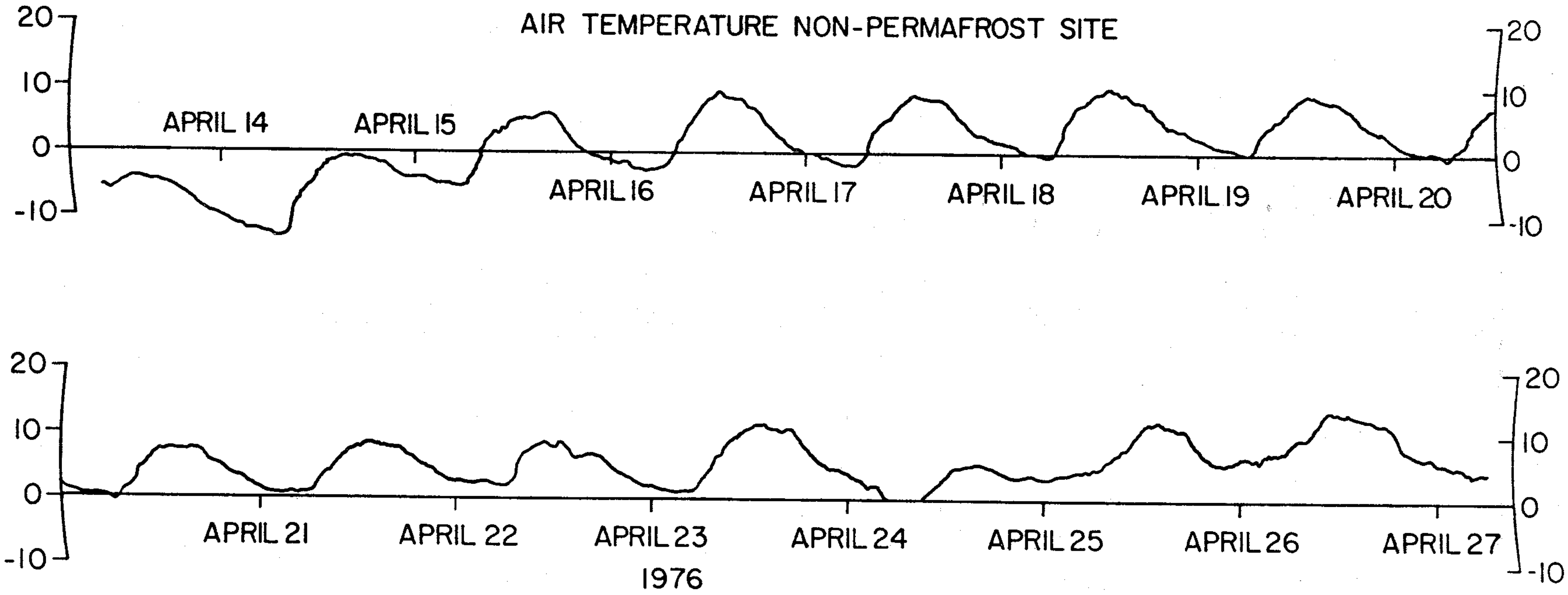


FIGURE A-2: Air Temperatures During Snow Ablation Period, 1976, °C.

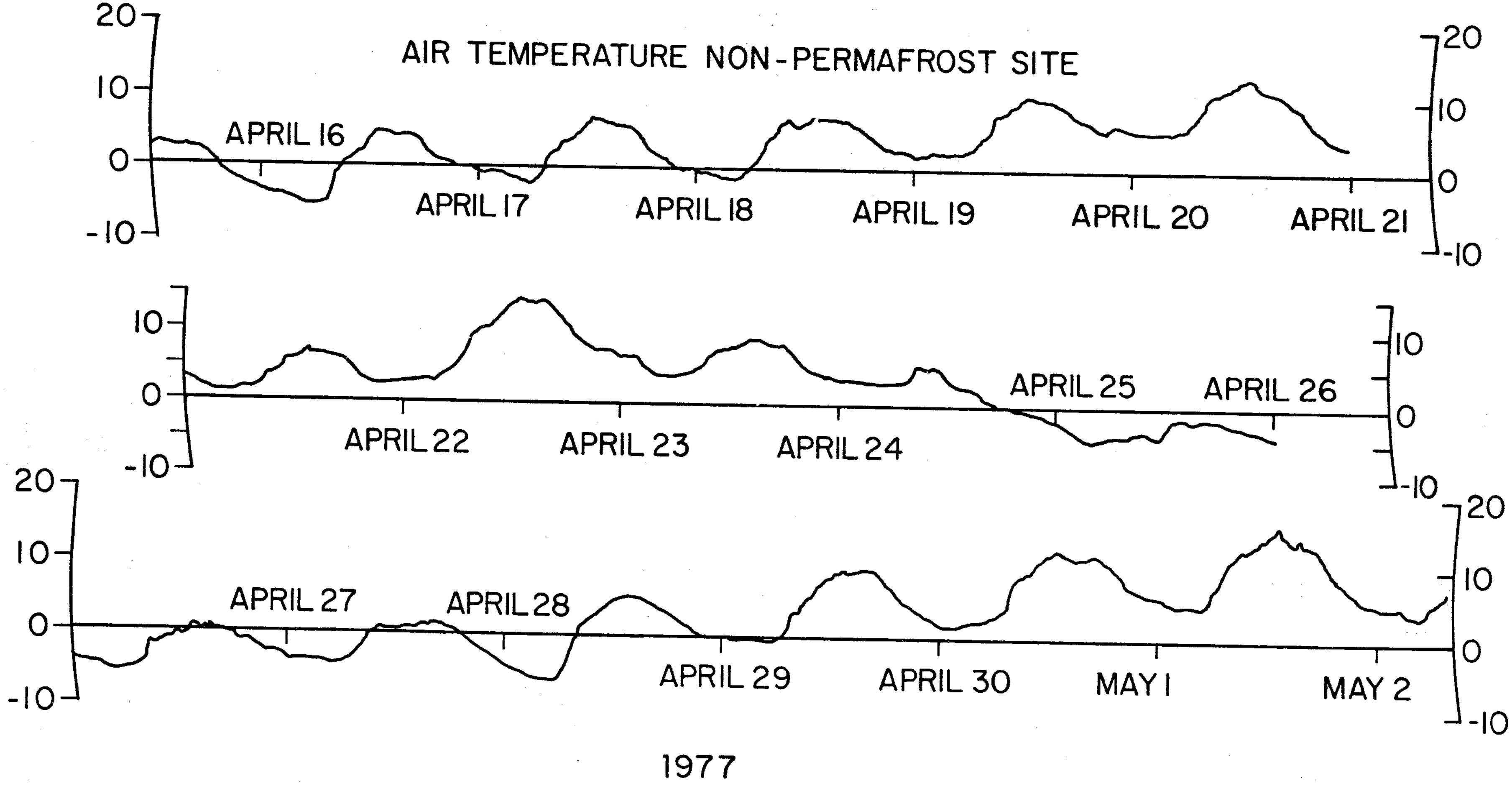


FIGURE A-3: Air Temperatures During Snow Ablation Period, 1977, °C.

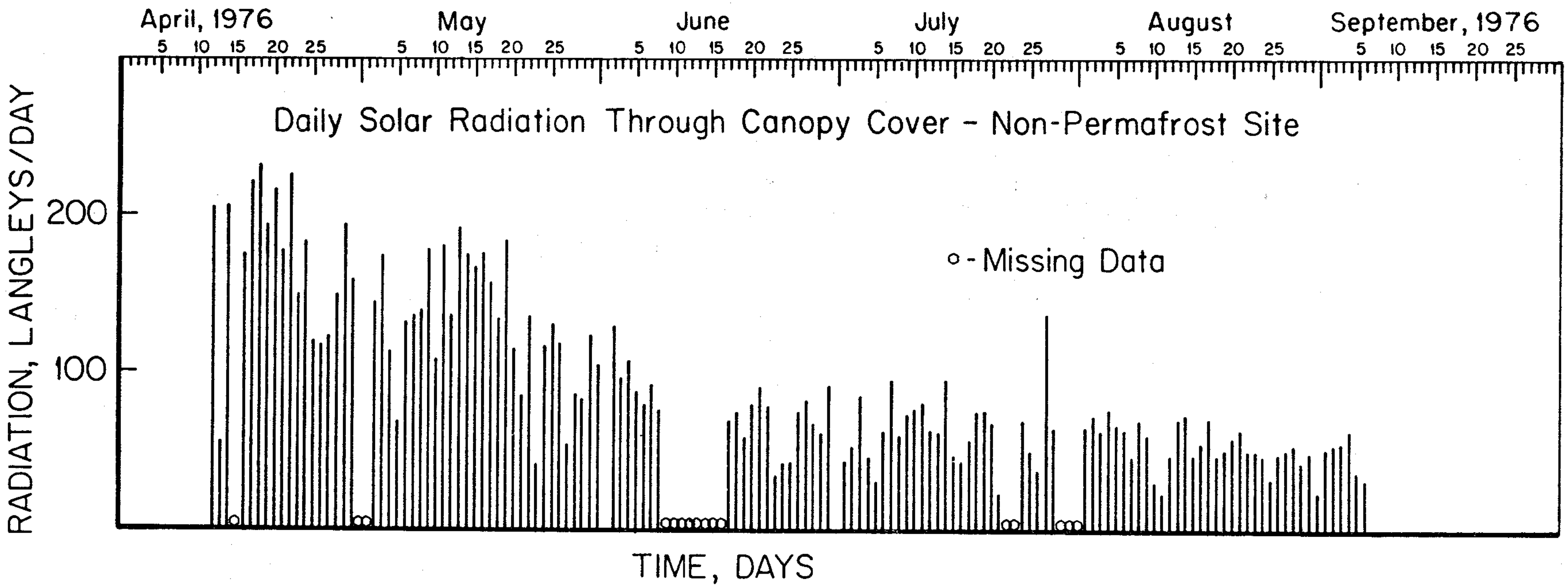
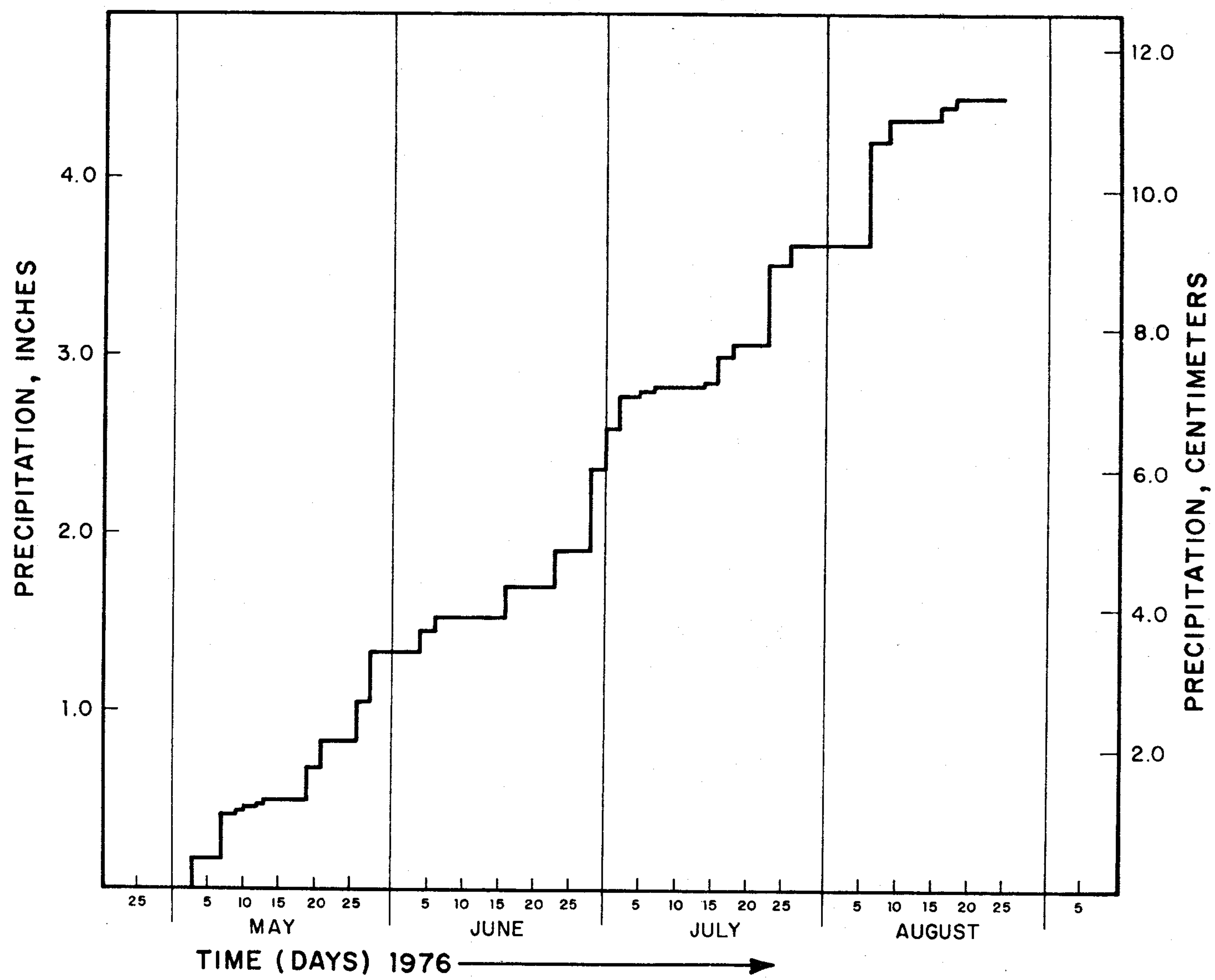


FIGURE A-4: Incoming Daily Solar Radiation Through Canopy Cover of White Birch and Aspen (Betula papyrifera-Populus tremuloides).



SUMMER PRECIPITATION AT NON-PERMAFROST SITE

FIGURE A-5: Summer Precipitation Through Canopy Cover, 1976.

TABLE A-1

MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE

Depth (cm)	11/20/75	12/16/75	1/20/76	2/18/76	3/10/76	4/7/76
0- 5	*168*	*147*	*233*	*148*	*120.4*	*104.2*
5-10		*144*	*108*			*63.4*
10-15	*9.8*	*37.5*	48	24.7	19.5	42.2
15-20	18.5	18.6	40.3	15.7	15.0	26.7
20-25	13.6	22.2	28	16.1	14.5	20.7
25-30	12.6	23.4	20.8	16.2	14.3	19.5
30-35	11.7	15.0	18.9	14.5	13.0	18.3
35-40	10.5	13.3	17.3	12.6	10.1	16.9
40-45	9.3	12.3	16.2	13.1	10.7	12.8
45-50	8.9	11.7	14.3	12.3	9.8	11.9
50-55	8.1	12.3	14.5	9.5	9.0	11.6
55-60	7.4	10.3	11.9	8.8	8.3	10.0
60-65	7.2	10.9	9.8		8.3	8.0
65-70	8.3	11.0	9.4			8.5
70-75	8.0	10.3				8.5
75-80	9.1	9.1				
80-85	8.4	7.5				
85-90	6.6					
90-95	7.6					

Note: Italicized numbers represent organic samples

Continued

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (Cont.)

Depth (cm)	4/16/76	4/21/76	4/27/76	4/29/76	4/30/76	5/5/76
0- 5	601	440	399*	503	153	220
5-10	325	133	141*	321		94
10-15	49	83	58	128	73	82
15-20	28	34	49	65	46	67
20-25	17	22	24	44	35	67
25-30	17	15	20	31	29	64
30-35	16	14	21	26	29	42
35-40	10	13	16	17	21	36
40-45	8	12	11	13	16	24
45-50		11	10	10	10	26
50-55		10	9	7	8	30
55-60		10	8	7	7	27
60-65		8	13	7	7	17
65-70			10	8	7	10
70-75			9	7	7	8
75-80			8	7	7	7
80-85			8	7	--	7
85-90			7	7	7	8
90-95			7	7	7	7

*organic samples taken on April 24

Continued

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (Cont.)

Depth (cm)	5/10/76	5/19/76	5/26/76	6/2/76	6/9/76	6/16/76
0- 5	263	127	196	216	--	115
5- 10	98	64	70	62	--	[illegible]
10- 15	60	33	38	38	30	28
15- 20	55	32	30	33	27	25
20- 25	45	32	29	32	26	24
25- 30	36	31	29	29	26	24
30- 35	33	26	28	28	27	24
35- 40	31	28	28	26	26	23
40- 45	40	28	29	26	25	25
45- 50	22	28	29	24	25	18
50- 55	20	28	36	22	14	20
55- 60	12	8	20	23	8	22
60- 65	12	7	12	22	7	13
65- 70	8	7	8	20		9
70- 75	7	7	7	15		7
75- 80	8	8	7	8	8	7
80- 85	7	6	7	10	7	7
85- 90	7	8	7	9		7
90- 95	6	9	7		7	6
95-100	7	7			8	
100-105	6	6			7	
105-110	6	6			6	
110-115	6	6			6	
115-120	6	6			6	
120-125	7	6			6	
125-130	7	6			5	
130-135	7	9			6	
135-140	7	9			6	
140-145	7	7			6	
145-150	7	7			8	
150-155	7	7			7	
155-160	8	7			7	
160-165	8	8			7	
165-170	8	8			8	
170-175	8	8			8	
175-180	8	8			8	
180-185	9	9			9	
185-190	9	9			10	
190-195		9			10	
195-200		8.6			9	
200-205		8.4			9	
205-210		9.0			9	
210-215		9.9			10	
215-220		9.1			11	
220-225		9.8			11	
225-230		9.7			10	
230-235		9.6			10	
235-240		12.6			11	
240-245		9.6			11	
245-250		4.9			11	
250-255		10.2			12	
255-260		11.4			12	
260-265		11.4			12	
265-270		14.2			12	
270-275		18.3			12	
275-280		14.5			12	

(continued)

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (cont.)

Depth (cm)	6/23/76	6/30/76	7/7/76	7/14/76	7/21/76	7/28/76
0- 5	88	185	147	96	138	[illegible]
5- 10	85	119	136	26	79	[illegible]
10- 15	68	33	39	15	13	22
15- 20	41	27	19	12	8	16
20- 25	32	25	14	13	7	11
25- 30	27	20	15	13	7	10
30- 35	24	22	15	14	8	10
35- 40	20	18	16	15	9	8
40- 45	20	17	16	15	10	10
45- 50	22	15	16	15	10	12
50- 55	22	15	15	13	8	9
55- 60	19	14	15	12	8	9
60- 65	19	14	15	12	8	8
65- 70	17	14	14	12	7	8
70- 75	17	11	13	10	6	8
75- 80		8	12	9	7	7
80- 85		7	10	8	6	7
85- 90		7	8	8	6	7
90- 95		7	7	8	6	7
95-100		7				
100-105		7				
105-110		7				
110-115		7				
115-120		10				
120-125		8				
125-130		8				
130-135		8				
135-140		8				
140-145		8				
145-150		9				
150-155		8				
155-160		8				
160-165		8				
165-170		9				
170-175		9				
175-180		9				
180-185		8				
185-190		8				
190-195		8				
195-200		9				
200-205		9				
205-210		9				
210-215		9				
215-220		9				
220-225		9				
225-230		10				
230-235		9				
235-240		10				
240-245		14				
245-250		9				
250-255		8				
255-260		11				
260-265		11				
265-270		11				
270-275		12				
275-280		11				

(continued)

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST (cont.)

Depth (cm)	9/17/76	9/23/76	11/16/76	12/29/76	1/25/77	3/1/77
0- 5	*47*	*136*	*250*	*186*	*305*	*514*
5- 10	10	*58*	*46*	*83*	*133*	*73*
10- 15	12	11	19	48	67	16
15- 20	10	9	18	22	33	17
20- 25	10	7	13	13	17	13
25- 30	11	8	11	12	17	10
30- 35	7	8	10	9	10	9
35- 40	7	14	9	8	10	9
40- 45	9	10	9	8	9	9
45- 50	9	7	9	9	9	7
50- 55	10	8	8	14	8	8
55- 60	7	7	7	12	9	9
60- 65	7	7	7	9	6	9
65- 70	6	7	6	9	6	7
70- 75	5	7	6	9	6	7
75- 80	5	7	6	10	6	8
80- 85	5	7	7	7	6	8
85- 90		7	7	7	6	8
90- 95		6	7	8	7	8
95-100		6	11	7	7	8
100-105		8	11.3	9	9	
105-110		7	11.9	8	8	
110-115		7	10.6	8	7	
115-120		7	7.6	8	6	
120-125		7	7.0	8	7	
125-130		7	6.7	9	7	
130-135		6	6.3	7	6	
135-140		9	8.2	8	7	
140-145		10	6.9	8	6	
145-150		8	6.7	7	6	
150-155		7	7.2	7	7	
155-160		7	8.6	7	7	
160-165		7	15.8	8	7	
165-170		8	10.6	8	12	
170-175		8	9.2		12	
175-180		8	8.8	9	12	
180-185		8	10.1	12	15	
185-190		8	13.3	12		
190-195		8	14.4			
195-200		8				
200-205		10				
205-210		9				
210-215		9				
215-220		9				
220-225		10				
225-230		11				
230-235		10				
235-240		11				
240-245		11				
245-250		10				
250-255		11				
255-260		12				
260-265		10				
265-270		10				
270-275		12				

(continued)

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (Cont.)

Depth (cm)	8/4/76	8/12/76	8/17/76	8/25/76	9/1/76	9/7/76
0- 5	--	90	49	68	49	
5- 10	--		9	11		7
10- 15	13	8	8	11	8	7
15- 20	9	8	8	8	10	6
20- 25	8	9	7	8	8	6
25- 30	8	8	7	8	8	7
30- 35	9	8	8	9	8	7
35- 40	9	8	8	9	8	7
40- 45	8	11	8	8	7	7
45- 50	7	10	8	9	7	7
50- 55	7		6	8	8	7
55- 60	7		7	7	9	7
60- 65	7		7	9	7	8
65- 70	8		5	10	6	8
70- 75	8		7	11	6	8
75- 80	8		8	9	6	9
80- 85	7		9	8	6	8
85- 90	7		6			7

Continued

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (Cont.)

Depth (cm)	3/30/77	4/13/77	4/22/77	4/29/77	5/6/77	5/10/77
0- 5	*354*	*265*	*314*	*1225*	*136*	*104*
5- 10	*108*	*80*	*87*	*101*	*75*	*84*
10- 15	32	29	81	127	63	61
15- 20	22	19	66	89	50	55
20- 25	22	14	19	61	40	34
25- 30	15	12	16	50	36	33
30- 35	13	10	17	35	28	34
35- 40	11	9	13	14	29	41
40- 45	11	7	8	7	22	32
45- 50	9	7	7	9	22	21
50- 55	9	8	7	7	10	16
55- 60	9	7	8	6	7	7
60- 65	9	7	8	6	7	15
65- 70	10	7	7	6	8	7
70- 75	8	6	6	6	9	6
75- 80	7	6	6	6	9	6
80- 85	7	7	6	6	8	7
85- 90	6	7	7		8	7
90- 95	7		7		7	7
95-100	7					

Continued

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (Cont.)

Depth (cm)	5/13/77	5/17/77	5/20/77	5/24/77	6/1/77	6/6/77
0- 5	*120*	*106*	*105*	*179*	*131*	*114*
5- 10	*86*	*52*	*67*	*74*	*58*	*52*
10- 15	38	42	38	54	48	30
15- 20	37	36	32	32	32	27
20- 25	35	38	31	31	31	29
25- 30	31	36	33	29	28	25
30- 35	32	26	28	29	27	26
35- 40	27	22	28	26	--	26
40- 45	21	22	27	28	27	27
45- 50	14	23	30	29	26	27
50- 55	8	24	11	28	25	22
55- 60	7	8	13	11	23	22
60- 65	7	8	8	11	23	18
65- 70	7	7	7	7	13	13
70- 75	7	6	7	6	7	8
75- 80	7	7	7	7	7	7
80- 85	7	7	7	6	6	6
85- 90	7	8	7	5	7	6
90- 95	7	7	7	6		6
95-100		6	7	6		

Continued

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (Cont.)

Depth (cm)	6/10/77	6/15/77	6/19/77	6/24/77	6/29/77	7/5/77
0- 5	*91*	*129*	*184*	*155*	*228*	*153*
5- 10	*56*	*77*	*151*	*89*	*198*	*85*
10- 15	29	32	47	36	36	20
15- 20	27	30	40	24	21	21
20- 25	27	27	36	24	20	20
25- 30	25	25	35	23	21	20
30- 35	25	25	25	22	20	21
35- 40	24	25	26	23	19	20
40- 45	25	24	26	22	19	18
45- 50	18	50	26	23	19	19
50- 55	9	26	54	13	20	16
55- 60	6	23	21	10	21	16
60- 65	6	7	18	10	21	17
65- 70	6	6	16	8	22	15
70- 75	6	7	7	8	15	15
75- 80	5	7	7	7	12	11
80- 85	6	6	7	6	8	9
85- 90	7	6	7	6	8	8
90- 95	7	6	7	6		7
95-100	6					6

Continued

A-2: MOISTURE CONTENT BY WEIGHT AT PERMAFROST SITE (Cont.)

Depth (cm)	8/25/76	9/1/76	9/7/76	9/15/76
0- 5	*42*	*45*	*65*	*76*
5-10	26		*51*	*43*
10-15	24	30	26	28
15-20	26	29	27	24
20-25	25	33	26	23
25-30	25	26	24	23
30-35	26	27	23	22
35-40	25	25	23	22
40-45	24	25	23	22
45-50	25	25	--	22
50-55	23	25	--	22
55-60	24	25	23	24
60-65	23	24	22	24
65-70	24	24	23	24
70-75	24	24	23	24
75-80	26	25	24	24
80-85	27	26	28	
85-90			24	

A-2: MOISTURE CONTENT BY WEIGHT AT PERMAFROST SITE (Cont.)

Depth (cm)	6/23/76	7/2/76	7/7/75	7/14/76	7/21/76	7/28/76	8/4/76	8/12/76	8/17/76
0- 5	*112*		*83*	*38*	*62*			*105*	*118*
5- 10	34	*189*	*83*	*52*	*86*		*43*	*81*	*63*
10- 15	24	*107*	35	35	67	38	33	*72*	28
15- 20	22	44	33	29	36	34	29	31	25
20- 25	24	31	30	27	28	29	28	29	25
25- 30	26	30	29	28	28	27	27	25	24
30- 35	18	28	28	27	28	27	27	25	24
35- 40	18	28	28	27	29	27	26	26	24
40- 45	13	32	28	27	25	27	26	26	24
45- 50	10	25	28	27	26	27	26	26	25
50- 55	9	20	27	27	26	27	27	26	25
55- 60	8	16	30	28	32	28	26	24	24
60- 65	7	14	26	27	31	26	26	24	25
65- 70	6	17	24	26	26	22	26	27	27
70- 75	7	16	19	25	26	23	25	29	25
75- 80	7	15	20	20	21	18	26		25
80- 85	8	17	17	22	18		28		24
85- 90		20		17	14				
90- 95		24			15				
95-100		26							

Continued

A-2: MOISTURE CONTENT BY WEIGHT AT PERMAFROST SITE (Cont.)

Depth (cm)	4/29/76	4/30/76	5/5/76	5/12/76	5/19/76	5/26/76	6/2/76	6/9/76	6/16/76
0- 5	226	242	358	312	229	157	221	159	68
5-10	287				98	169	162	73	82
10-15	133	58	148	118	37	49	65	36	47
15-20	42	35	36	53	34	29	34	47	32
20-25	34	35	31	34	36	29	30	33	27
25-30	33	32	31	40	34	44	33	30	29
30-35	32	33	32	37	32	33	29	25	25
35-40	32	35	34	32	29	28	26	26	24
40-45	32	37	34	34	27	27	26	21	21
45-50	31	35	34	31	27	24	--	20	21
50-55	30	33	35	31	27	25	24	21	16
55-60	30	32	32	29	24	25	25	18	16
60-65	28	30	29	27	23	23	26	16	16
65-70	27	27	28	24	18	18	27	14	17
70-75	25	27	30	21	13	17	24	15	17
75-80	20	26	27	25	13	17	20	19	19
80-85	16	19	25	23	15	16	18	21	
85-90	16	18	23	23		18	20		
90-95				23					

Continued

TABLE A-2
MOISTURE CONTENT BY WEIGHT AT PERMAFROST SITE

Depth (cm)	11/20/75	12/16/75	1/20/76	2/18/76	3/10/76	4/8/76	4/16/76	4/21/76	4/27/76
0- 5	*65.8*	*144*	*315*	*244*	*119.2*	*215.7*	*313*	*346*	*358**
5-10						*104.6*	*173*	*166*	*636**
10-15	49	60.4	145	48.2	*85.6*	*58.3*	*97*	*139*	*143**
15-20	41	44.6	273	51.8	54.2	40.6	49	76	51
20-25	40.9	34.5	43.9	36.7	35.1	37.7	39	44	43
25-30	39.7	33.4	113.0	31.3	44.1	32.6	34	39	32
30-35	32.9	31.7	34.9	34.2	34.0	32.0	32	32	32
35-40	18.2	32.3	41.0	28.4	33.5	30.2	31	31	28
40-45	38	27.4	35.9	25.7	31.9	30.3	33	22	27
45-50	22	27.1	33.6	28.2	30.0	31.6	30	22	27
50-55	20.9	24.4	31.4	28.5	21.6	25.5	30	19	28
55-60	22.5	19.3	28.3	27.6	19.4	21.0	28	22	30
60-65	20	19.7	27.5	25.7	18.2	13.8	26	16	28
65-70	17.7		24.6	26.0	18.5	14.2	28	16	19
70-75	24.1		23.4	25.4	14.9	12.2	25	12	19
75-80	25.6		18.3	32.9	14.6		25	12	20
80-85	31.1		22.4				22		23
85-90	30.6						19		
90-95	34.8								

*organic samples taken on April 24
Note: italicized numbers represent organic samples

Continued

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (cont.)

Depth (cm)	8/31/77	9/7/77	9/15/77	9/27/77	10/3/77	10/12/77
0- 5	171	55	256	320	[illegible]	137
5- 10	43	46	113	104	81	[illegible]
10- 15	11	10	13	39	60	31
15- 20	12	9	12	27	38	27
20- 25	11	8	15	25	34	26
25- 30	10	8	10	29	27	26
30- 35	9	8	9	24	27	25
35- 40	7	8	9	24	22	26
40- 45	7	7	9	23	22	26
45- 50	8	8	7	16	22	15
50- 55	8	7	7	7	10	9
55- 60	9	8	7	7	8	8
60- 65	7	7	7	7	7	7
65- 70	7	7	7	7	7	8
70- 75	7	7	7	6	7	8
75- 80	7	6	7	6	7	8
80- 85	6	6	6	6	7	7
85- 90	6	6	5	6	7	7
90- 95	6	6	5	6	6	8
95-100	6	6	6	6	6	6
100-105						6
105-110						6
110-115						6
115-120						6
120-125						6
125-130						6
130-135						6
135-140						6
140-145						7
145-150						7
150-155						7
155-160						7
160-165						7
165-170						8
170-175						7
175-180						8
180-185						8
185-190						8
190-195						8
195-200						9
200-205						10
205-210						12
210-215						11
215-220						12
220-225						12
225-230						11
230-235						10
235-240						11
240-245						11
245-250						11

A-1: MOISTURE CONTENT BY WEIGHT AT NON-PERMAFROST SITE (Cont.)

Depth (cm)	7/10/77	7/24/77	8/1/77	8/7/77	8/16/77	8/22/77
0- 5	*101*	*236*	*98*	*127*	*94*	*40*
5- 10	*87*	*62*	*60*	*107*	*70*	*40*
10- 15	22	23	14	24	22	11
15- 20	21	22	15	16	17	8
20- 25	21	17	17	12	11	7
25- 30	16	17	12	10	8	7
30- 35	17	13	10	9	10	7
35- 40	15	11	11	7	9	8
40- 45	14	11	10	7	8	8
45- 50	14	11	10	7	9	7
50- 55	15	11	10	7	7	7
55- 60	10	12	10	7	8	7
60- 65	8	11	9	7	8	7
65- 70	8	10	9	7	8	6
70- 75	8	10	7	7	8	5
75- 80	7	10	6	6	7	5
80- 85	7	10	6	6	7	6
85- 90	7	10	6	7	7	6
90- 95	8	8	6		7	6
95-100	7	7	8		7	6

Continued