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HYDROLOGIC EFFECTS OF FROZEN GROUND
Literature Review and Synthesis

S. Lawrence Dingman

March 1975

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This report, intended as a section of a future comprehensive monograph on cold regions hydrology, summarizes existing knowledge of the profound hydrologic effects of frozen ground. The general characteristics of seasonally and perennially frozen ground also are described, and the geographical distribution of frozen ground is discussed.

PREFACE

This report was prepared by Dr. S. Lawrence Dingman, formerly of DuBois and King, Inc., Randolph, Vermont, under Purchase Order No. DACA89-71-1441. It is published under DA Project 4A162121A894, *Engineering in Cold Environments*, Task 20, *Cold Regions Facilities Operations, Maintenance, and Engineering of Military Installations*, Work Unit 002, *Analysis of the State of the Art in Cold Regions Technology*.

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Literature Review and Synthesis

by

S. Lawrence Dingman

INTRODUCTION

Importance of hydrology in cold regions science and technology

Hydrology is the branch of earth sciences dealing with the quantities, states, locations, circulation, and quality of the waters of the earth. The central concept of hydrology is the hydrologic cycle, the continuous movement of water from the ocean, through the atmosphere, and over and through the land surfaces of the earth until its ultimate return to the ocean. This cycle is fundamental to virtually all aspects of earth science, including the biological aspects. It is intimately related to the thermal characteristics of the earth and the physical and chemical processes that modify the earth's land surfaces and produce the substrate that supports terrestrial life. And it maintains the circulation of water which is essential to all life. The central importance of the cycle has been well stated by L'vovich (1961):

> The hydrologic cycle is one of the great earth phenomena. In its process, large amounts of water pass through the atmosphere, over the surface of the earth and through the earth's crust; water does its erosional work, dissolves rock and takes part in the key functions of the life processes. The hydrologic cycle links water with all other components of nature and therefore opens the best possibilities for inter-disciplinary research into inter-related processes and phenomena. Finally, in the process of the cycle, water is also used for a wide range of practical purposes.

Hydrology is especially central to the fields of cold regions science and technology, as the freezing point of water, in one form or another, is the critical environmental boundary determining the cold regions.

Bates and Bilello (1966) attempted to define the geographical boundaries of the cold regions of the Northern Hemisphere on the basis of air temperature, snow depth, ice cover and frozen ground. The latter three phenomena fall directly within the purview of hydrology, and all the criteria are directly related to the freezing point of water.

It can be said, in fact, that it is the freezing of water at 0°C that defines and makes necessary the study of cold regions science and engineering. For this reason, virtually all the subjects in these fields can be thought of as branches of hydrology: the study of glaciers, freshwater and sea ice, snow, frozen ground and ice fog. A review of the titles of the technical publications of the U.S. Army Cold Regions Research and Engineering Laboratory (1972) reflects this fact.

More specifically, there are several hydrologic aspects of cold regions science and technology that are currently of critical environmental importance. There has been in North America, and there is now in the U.S.S.R., discussion and action toward the goal of tapping, by means of continental-scale engineering schemes, the water resources of the Arctic and Subarctic and diverting portions of them to temperate areas (Micklin 1969).

Obviously, such schemes require the evaluation of the quality, quantity, and spatial and temporal variations of these water resources. Of at least equal importance to this exploration is detailed knowledge of the interrelations among these water resources and the physical and biological environments of the cold regions. Not the least of these may be the influence of river waters entering the Arctic Ocean on the stability of the thin ice cover of that ocean (Antonov 1948, 1958, 1961, 1968). This highly reflective surface plays a crucial role in the radiation balance of the earth, and its diminution would almost certainly have very large, even catastrophic, effects on the earth's climate (Ewing and Donn 1956, 1959, Donn and Ewing 1966, Sellers 1969).

On a somewhat smaller scale, it now appears certain that gas and oil resources of high latitudes will be extensively exploited in the near future. The potential environmental effects of this exploration have received considerable attention, and many of these effects are hydrologic, relating to the thawing of frozen ground and subsequent movements of groundwater, effects of floods on structures, etc.

There is also a continuing and increasing need for human water supplies in cold regions. As surface and groundwater is frozen for long periods and over a wide horizontal and vertical extent, hydrologic knowledge must be brought to bear to locate and evaluate suitable sources.

A large number of additional practical problems of cold regions involving hydrology can also be enumerated, including runoff prediction, ice jams, frost heaving, dam construction, and road and culvert icing.

Definition of frozen ground and focus of report

This report, conceived as a section of a comprehensive monograph on cold regions hydrology, focuses on the hydrologic effects of seasonally and perennially frozen ground. "Frozen ground" is defined as earth material (soil, unconsolidated material and bedrock) that has a temperature less than 0°C, independent of the water content of the material. This definition seems most logical, as it uses the same temperature criterion as the now widely accepted definition of permafrost, originally given by Müller (1947). It is, however, apparently not the same definition used by many workers investigating seasonally frozen ground. Most of the latter studies at least implicitly define frozen ground as ground containing visible ice, and Willis et al. (1961, p. 117) stated that "A frozen soil is generally considered to be one in which the liquid phase has changed to solid state." As discussed below, the presence of temperatures below 0°C does not necessarily mean that soil water is frozen.

The occurrence of frozen ground is, of course, one of the most distinctive phenomena of cold regions. According to Bates and Bilello (1966) cold regions include "nearly all the land mass north of 40° (north latitude)." Where it occurs, frozen ground has important influences on hydrologic processes and the water balance, including: infiltration, and hence runoff from rain and snowmelt; movement of soil moisture, and hence evaporation, transpiration, and groundwater recharge; movement of groundwater, and hence the underground feeding of rivers.

This report begins by describing the general characteristics of seasonally and perennially frozen ground. Next, the geographical distribution of frozen ground is discussed, supplemented by a number of maps. Finally, the hydrologic effects of frozen ground are discussed, largely on the basis of observational data in published reports. Unfortunately, because of time and budgetary limitations, it was not possible to examine and translate a large number of Russian papers that treat various aspects of this topic. In the final section, the previous portions of the report are summarized, and gaps in knowledge are identified.

FROZEN GROUND

General characteristics

As noted above, frozen ground is defined as earth material that has a temperature less than 0°C. Material that has sustained a temperature below 0°C continuously for two or more years is called permafrost (Müller 1947, p. 3, Ferrians et al. 1969, p. 4, Williams 1970, p. 12); if temperatures below 0°C are less continuous, but recurring, the freezing is described as seasonal. Brown (1968, p. 1) has divided these intermittently frozen soils into: 1) "seasonally thawing" soils, which overlie permafrost, and 2) "seasonally freezing" soils, which overlie unfrozen ground. Figure 1 shows the thermal relationships characterizing these types of frozen ground.

It is important to note again that frozen ground is defined strictly on the basis of temperature, without regard for the state of the water in the soil.

Figure 1 also illustrates zones defined by the seasonal fluctuation of ground temperatures. At the surface, temperature is controlled by energy exchange with the atmosphere, and changes very rapidly. Over a year, the amplitude of surface temperatures is large, approximating the annual range in air temperatures. These fluctuations become smaller with depth due to the thermal diffusivity of the soil, and at some depth (depending on soil characteristics, soil moisture, climate, snow cover, and vegetative cover) reach zero. The depth of this level generally ranges from 10 to 30 m (Brown 1968, p. 3, Williams 1970, p. 12).

The temperature at the level of zero annual amplitude is taken as the mean annual ground temperature. Generally, this temperature is 2° to 7°C warmer than mean air temperature, due

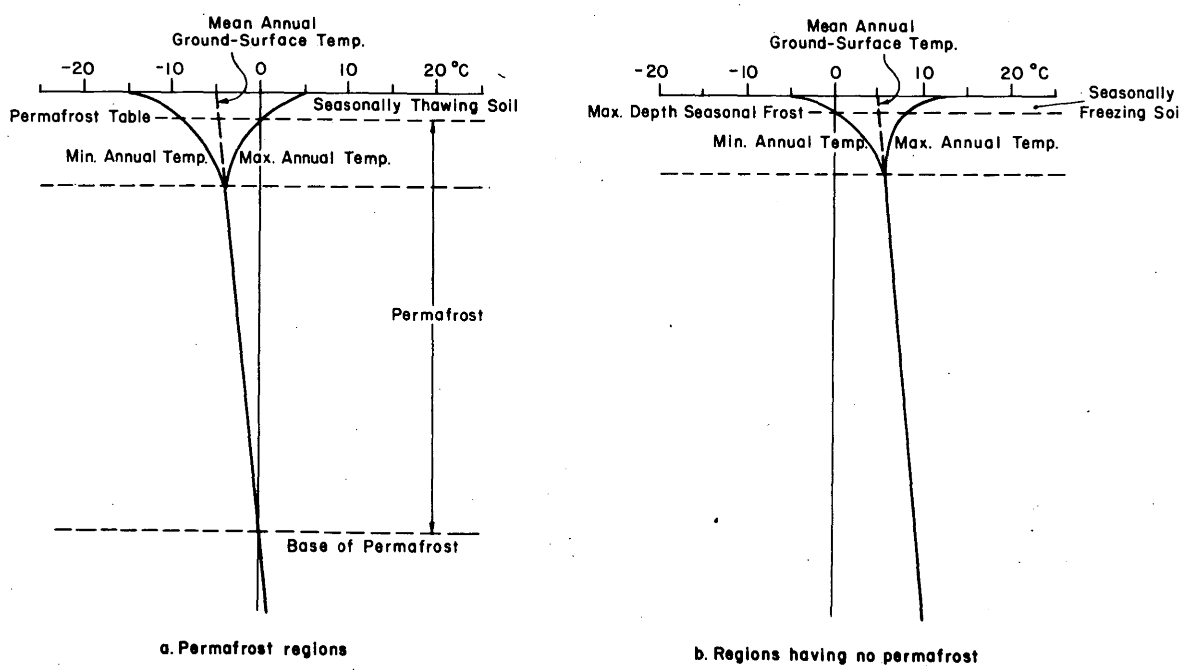

Figure 1. Ground-temperature changes in permafrost and non-permafrost regions. (After Williams 1970.)

largely to the insulating effect of snow in the winter, and the relationship between the two temperatures can be estimated as (Brown 1968):

$$T_G = T_A + A/2\,(1 - 1/f) \tag{1}$$

where

f = exp $(z\sqrt{\pi/kt})$

T_G = mean annual ground temperature

T_A = mean annual air temperature

A = annual amplitude of mean monthly air temperatures

z = thickness of snow cover

k = thermal diffusivity of snow

t = period of oscillation (= 1 year).

The permafrost table is defined as the level above which temperatures below 0°C do not persist from year to year; the layer above this is called the "active layer" or "seasonal-thaw zone." In seasonally freezing soils, the maximum depth of 0°C temperatures is generally called the "depth of frost penetration."

Above permafrost, cooling and freezing proceed both from the ground surface downward and from the permafrost table upward. Almost invariably, cooling and freezing are faster from the surface, as the stronger temperature gradient exists in that direction. It is possible under the right conditions that the seasonal freezing in each direction may not meet, leaving a thawed zone or *talik* between the two frozen zones. Thawing of the active layer always occurs downward from the surface.

In seasonally freezing soils, the situation is reversed. Freezing occurs from the ground surface only, while thawing takes place in both directions, and a frozen layer commonly exists at depth during the thawing process. Thawing is generally most rapid in a downward direction, though the opposite has been observed (Post and Dreibelbis 1943). Thawing from the bottom commonly begins before it starts from the top (Kienholz 1940, Atkinson and Bay 1940, Belotelkin 1941, Bay et al. 1952).

Virtually all near-surface earth materials contain some water, but this does not necessarily mean that ground at temperatures below 0°C contains ice. In many soils, ice does not form until the temperature falls considerably below 0°. This "freezing point depression" may be due to several causes: 1) the presence of dissolved ions in the water, 2) supercooling due to the absence of freezing nuclei, and 3) the existence of water that is tightly bound to soil particle surfaces, such that its intermolecular structure, and hence its thermodynamic properties, are altered. Although widely observed, this phenomenon has not been completely explained (Anderson 1967, p. 14). The freezing point is lowered as water content decreases for a given soil and, for a given water content, generally decreases as particle size decreases. Figure 2 summarizes some of the data illustrating the above tendencies; the phenomenon is more completely discussed in Tsytovich and Sumgin (1937, p. 20-29), Lovell and Herrin (1953, p. 112-115), Low et al. (1966) and Anderson (1967).

Intimately related to the freezing point depression phenomenon is the coexistence of unfrozen water and ice in equilibrium in soils. It is well established that unfrozen water films exist between the surfaces of soil particles and ice that has formed in pore spaces. Such films can exist due to the rejection of dissolved material as water freezes and the resulting increase in concentration of solutes in the unfrozen water, and due to the altered structure of water adsorbed on

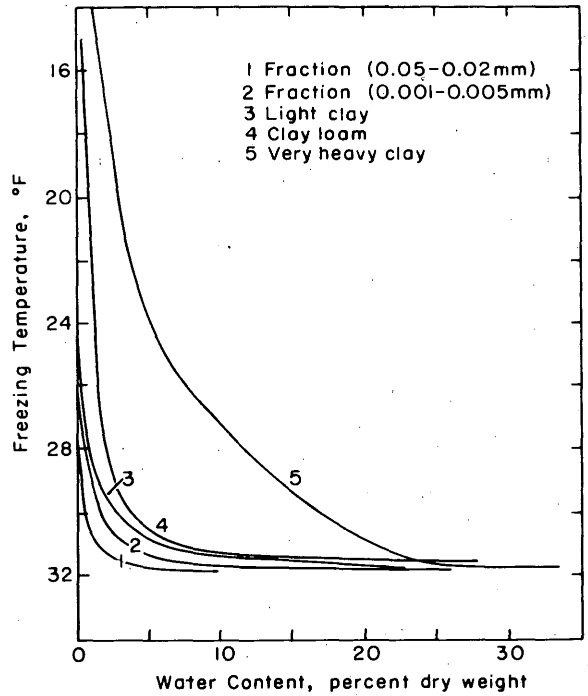

Figure 2. Relationship between freezing temperature and water content for different soils (from Lovell and Herrin 1953).

mineral surfaces. The latter phenomenon is probably the most important cause; experiments showing increasingly lowered freezing points of water in increasingly small capillary tubes are readily performed (Tsytovich and Sumgin 1937, p. 22). For a given soil, the amount of unfrozen water depends very largely on temperature (decreasing as temperature decreases), and is essentially independent of total water content (Anderson 1967, p. 15, Keune and Hoekstra 1967) (see Fig. 3). At a particular temperature, unfrozen water content is determined by the specific surface area of the soil, the nature of the soil particle, and the types and amounts of soluble substances present (Anderson 1967, p. 15). Dillon and Andersland (1966) were successful in predicting unfrozen water contents of a variety of soils based on grain size, Atterberg limits, freezing point depression and temperature. Koopmans and Miller (1966) demonstrated analogies between soil moisture characteristic curves and soil freezing characteristic curves, and Keune and Hoekstra (1967) developed a means for relating unfrozen water content to the moisture characteristic curve of a soil. Relations between clay contents (fraction finer than 0.002 mm) and unfrozen water content for some inorganic soils are shown in Figure 4.

The water in these unfrozen films is in general continuously connected and quite mobile, and experiments have shown extensive moisture migration in frozen soils under temperature (Hoekstra 1966) and electrical gradients (Hoekstra and Chamberlain 1964). These films play a major role in redistributing moisture during the freezing process, and are of crucial importance in soils engineering (frost heaving) as well as hydrology. As the 0°C isotherm reaches a given level during the freezing of a moist soil, any relatively solute-free "gravitational" water in the pore spaces will freeze. This freezing reduces the liquid water content at the freezing front and creates a water content (potential) gradient from the unfrozen soil to the freezing front. There is thus created a driving force for the movement of water to the front; this movement occurs principally in the liquid phase (Hoekstra 1966). As shown in Figure 5, the rate of flow decreases rapidly as temperature

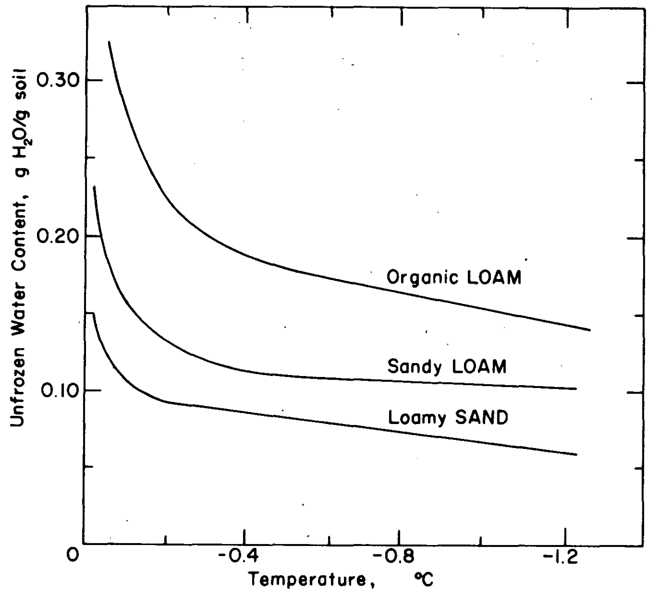

Figure 3. Unfrozen water as a function of temperature for three loam soils (from Keune and Hoekstra 1967).

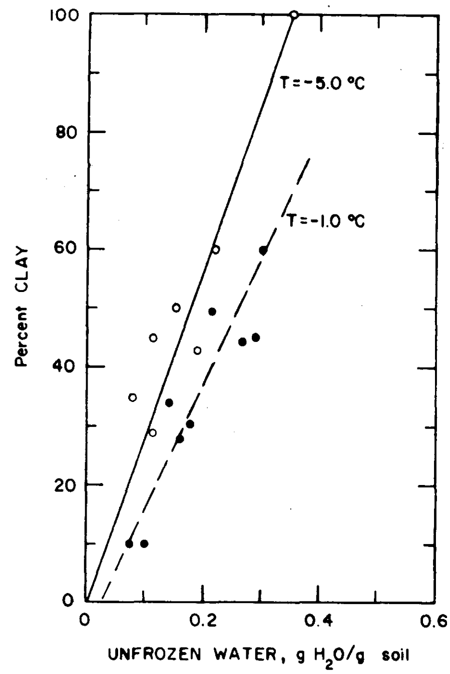

Figure 4. Unfrozen water as a function of percent clay (finer than 0.002 mm) for some inorganic soils at temperatures of -1 and -5°C (from Keune and Hoekstra 1967).

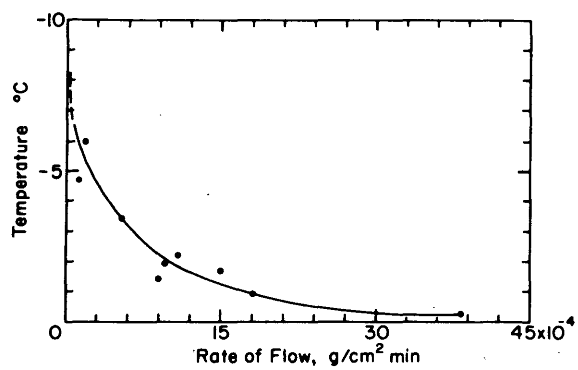

Figure 5. Rate of water flow in frozen Fairbanks silt as a function of temperature below 0°C (from Hoekstra 1966).

decreases, reflecting the decrease in thickness of the unfrozen films. Krumbach and White (1964) found that moisture contents and pore space increased while bulk density decreased as a result of freezing in the surface layer of loam soils, due to the migration of water to the freezing front from below.

If the unfrozen films are abundant (i.e. soils have a significant silt-clay fraction), if sufficient water is present in the unfrozen portion of the soil, and if freezing progresses slowly enough to allow significant accumulation of water at a particular level, "ice lenses" will form at the freezing point. These lenses are generally a few centimeters thick. In terms of total water content, the soil under such conditions is supersaturated, and its volume increases in proportion to the total amount of pure-ice lenses formed by water transported from unfrozen portions of the soil, plus the 9% volume increase that water undergoes upon freezing. Where grain sizes are large (sands and larger), water is limited, or freezing is fast, ice lenses will not form, and the pore water will freeze *in situ*, with only a small change in soil volume. Processes associated with the freezing of coarse- and fine-grained soils are discussed in detail by Yong (1962, p. 17-34).

Textures and structures of frozen soil

U.S. investigators of the hydrologic effects of seasonally frozen ground have generally recognized four types or textures of frozen ground: 1) "concrete frost," essentially saturated or supersaturated ground that is completely frozen (save for negligible amounts of unfrozen films on soil grains), and which may contain ice lenses; 2) "granular frost," with small ice crystals intermixed with soil particles and aggregated around them; 3) "honeycomb frost," which is similar to granular, but with a higher degree of connection among ice crystals and a lower porosity; and 4) "stalactite frost," apparently identical to "needle ice" or "pipkrake," which is characterized by small needle-like ice crystals aligned vertically and extending downward into the soil from a heaved surface. Three of these terms were originally used by Post and Dreibelbis (1943), while Hale (1950, p. 3) added the term "granular frost," and these terms have persisted (e.g. Storey 1955, Pierce 1956, Trimble et al. 1958, Megahan and Satterlund 1962). Stoeckeler and Weitzman (1960) recognized a "porous concrete" frost, similar to concrete frost, but permeable to air.

The effect of freeze-thaw cycles on the water stability and degree of aggregation of agricultural soils has been explored in several studies. Slater and Hopp (1949) found that repeated freezing and thawing markedly decreased the water stability of moist soils, and that the decrease was porportional to the number of freeze-thaw cycles and soil moisture. These conclusions were confirmed in field studies (Slater 1951, Slater and Hopp 1951). Willis (1955) also found that only a few freeze-thaw cycles reduced soil aggregation, and that most of the change occurred in the first cycle, with little change after the fifth. Logsdail and Webber (1959) and Sillanpää and Webber (1961) also found a general decrease in aggregation due to freeze-thaw, with larger decreases at higher soil water contents.

Detailed descriptions of permafrost in the literature are to a very large degree of saturated or supersaturated materials, very similar in texture to the concrete frost described above. However, permafrost commonly has larger and more spectacular pure ice bodies, for example as described by Sellmann (1967) in central Alaska. In addition to large, irregular ice lenses, perennially frozen ground commonly contains ice wedges, extending several meters down and forming, in plan view, a polygonal network. These structures have been described, and their mode of formation elucidated, by Lachenbruch (1963).

The soil structures associated with seasonally freezing and seasonally thawing ground are striking and varied, and have been subjects of much study and speculation (Troll 1958, Washburn 1956, Corte 1961, 1962a, 1962b). In addition to ice wedge polygons, these include various manifestations of soil sorting by frost action, such as stone stripes and nets, frost boils, down-slope movement of soils (solifluction), and smaller scale soil movement features caused by needle ice formation. Detailed discussion of these features is outside the scope of this report.

GEOGRAPHICAL DISTRIBUTION OF FROZEN GROUND

Global distribution

Permafrost. As will be discussed later, the occurrence of temperatures below 0°C in the soil depends upon a number of factors, including air temperature, solar radiation, cloud cover, snow cover, vegetative cover, soil moisture, and geothermal heat flow. These factors vary widely over short distances, making it difficult to map the depth, duration, or even occurrence of frozen ground. However, permafrost is by definition a relatively stable environmental feature, and a number of maps showing its distribution in the Northern Hemisphere have been published (e.g. Fig. 6). Generally, zones of spatially "continuous" and "discontinuous" permafrost are delineated.

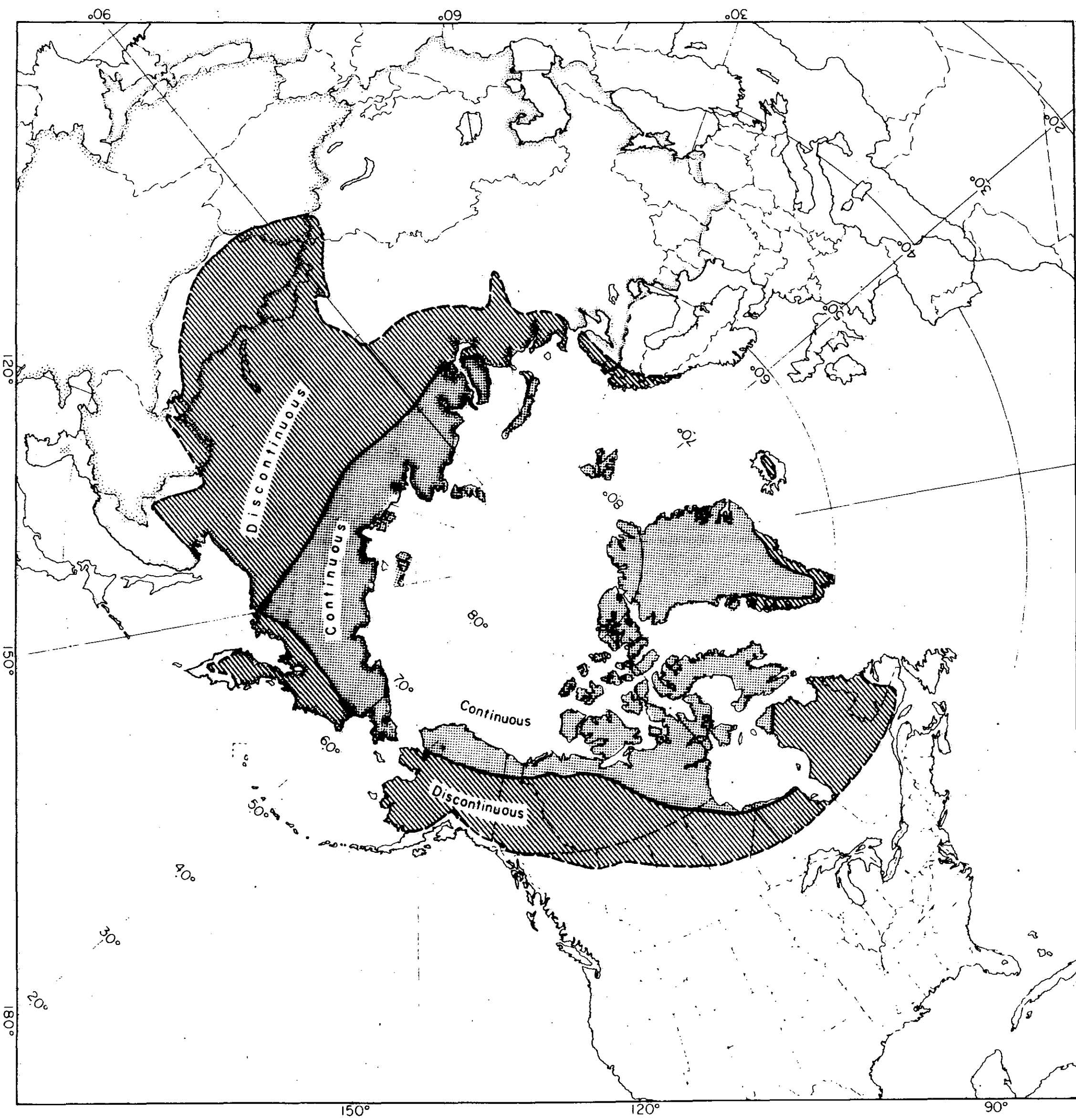

Figure 6. Permafrost zones in the Northern Hemisphere (from Ferrians et al 1969).

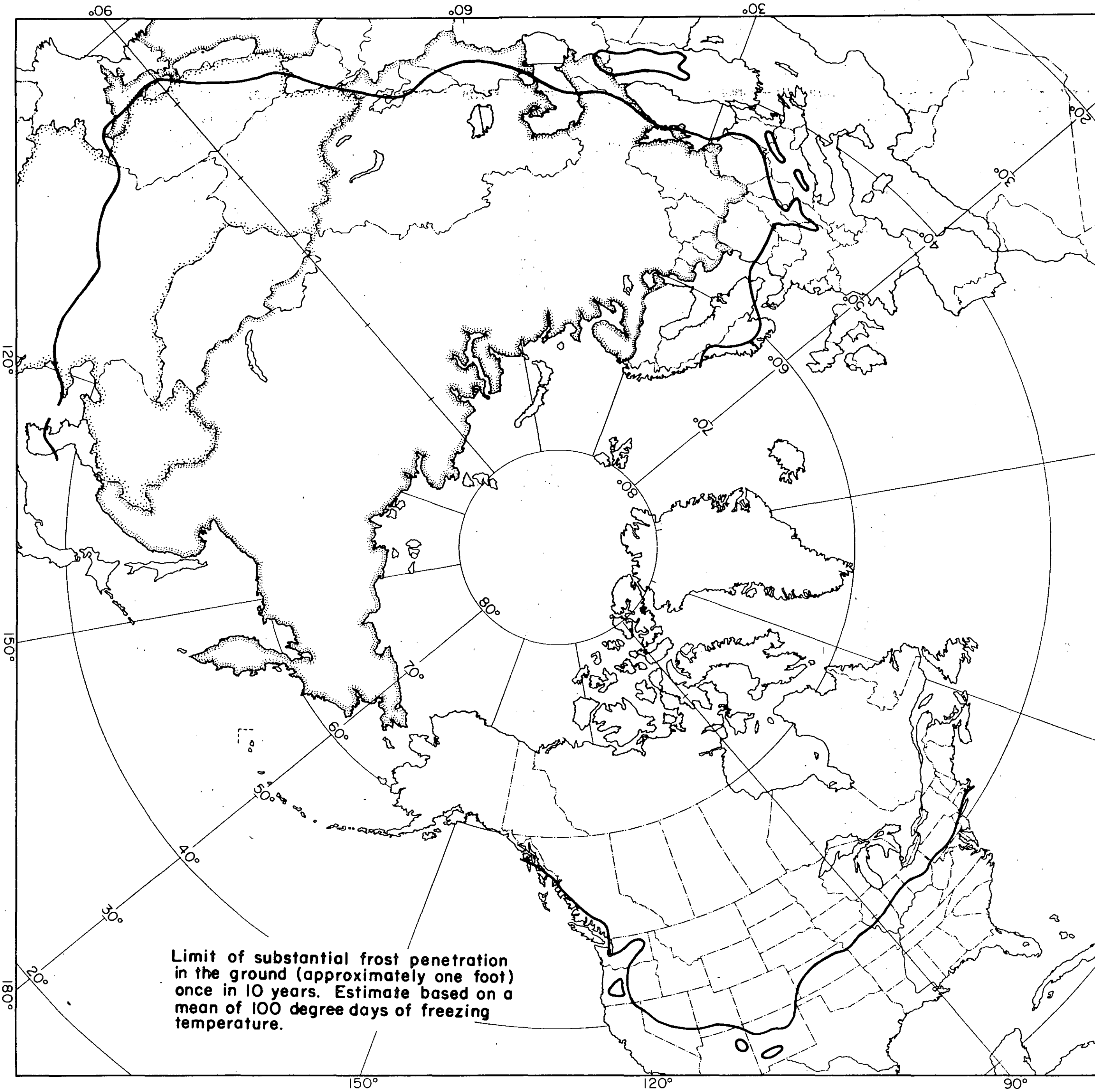

Figure 7. Seasonal freezing zones in the Northern Hemisphere. The limit of substantial frost penetration is virtually coincident with the limit of areas where the mean temperature of the coldest month is less than 0°C (from Bates and Bilello 1966).

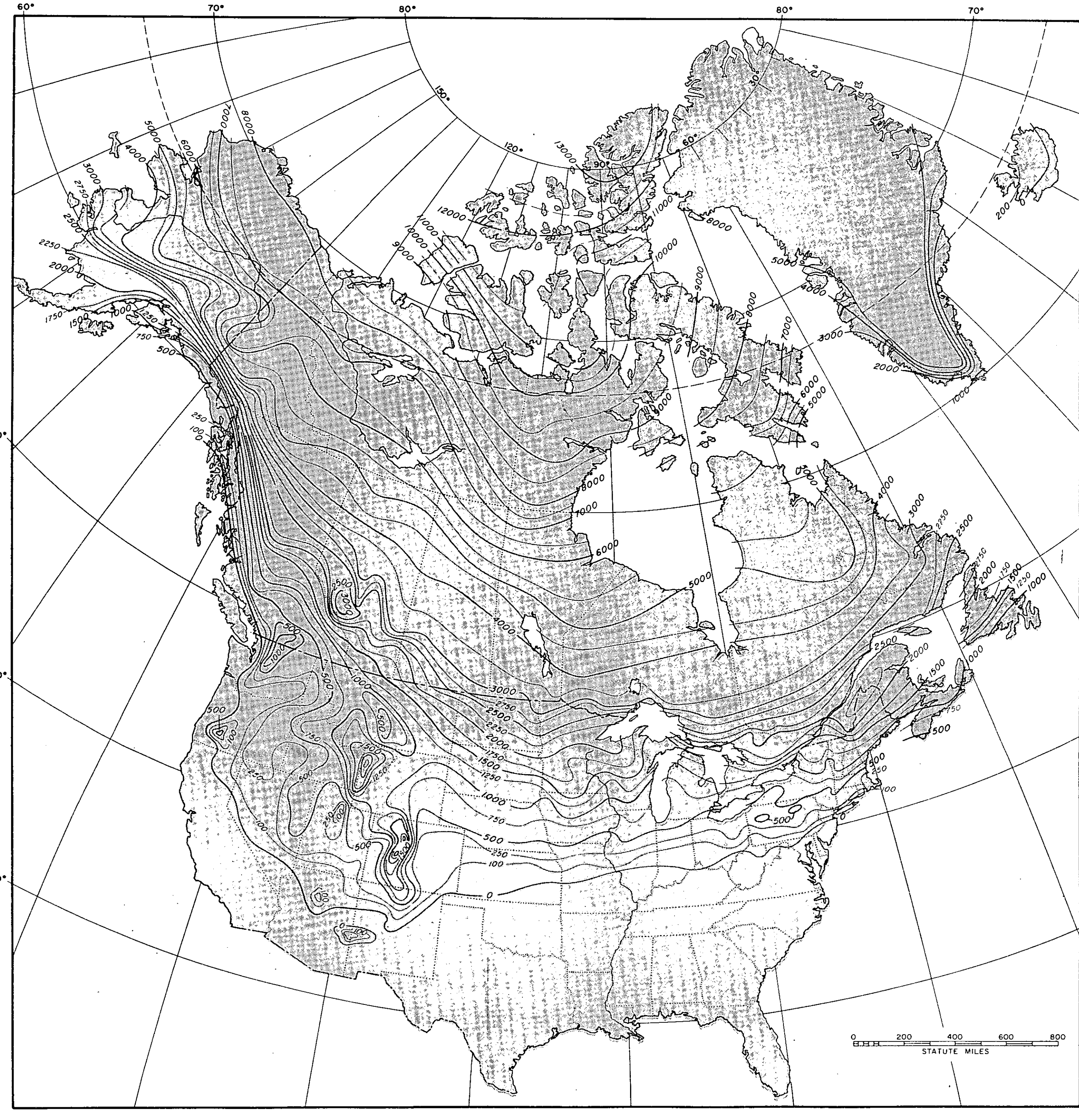

Figure 8. Distribution of mean air-freezing index values in North America (from Gilman 1968).

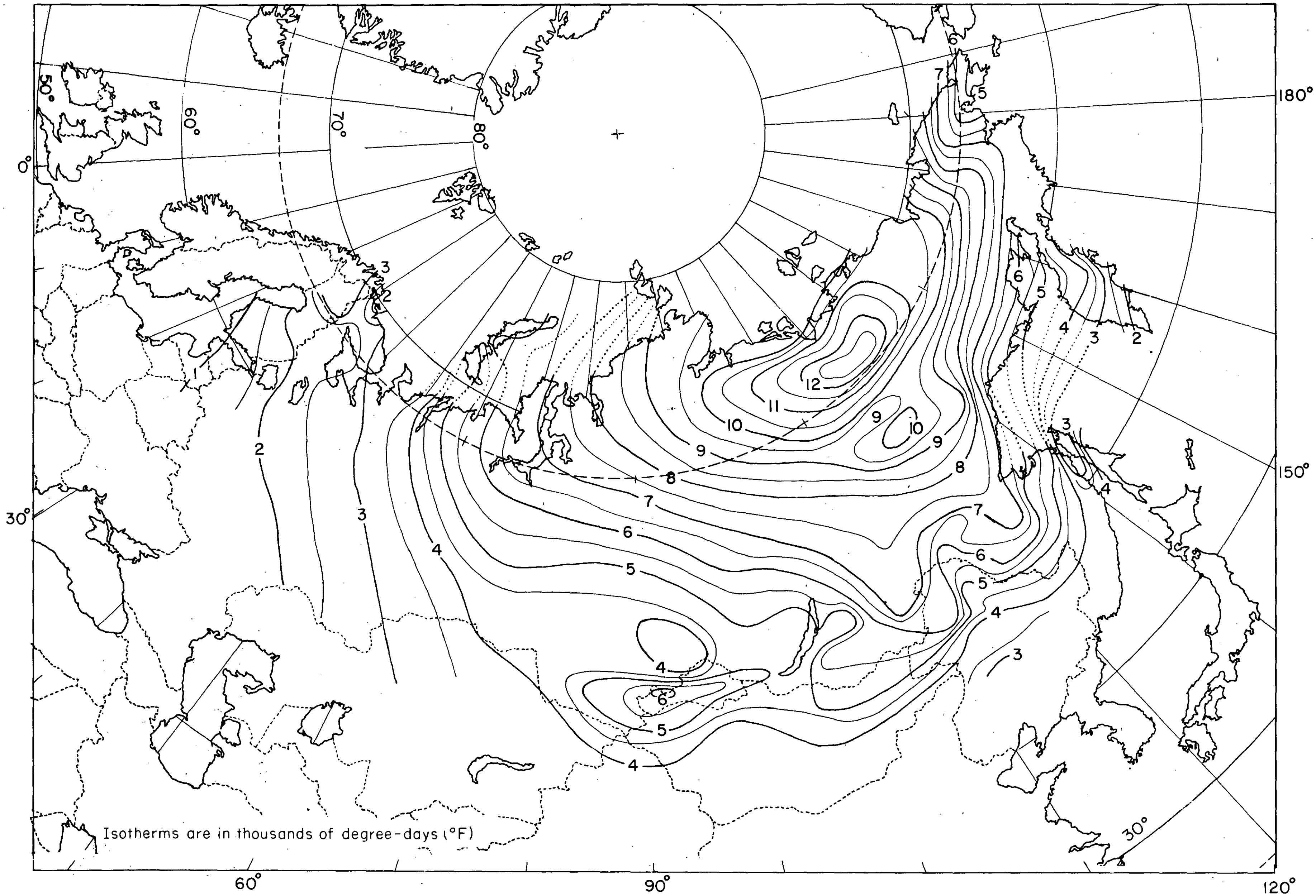

Figure 9. Distribution of mean air-freezing index values in northern Eurasia (from Scott 1963).

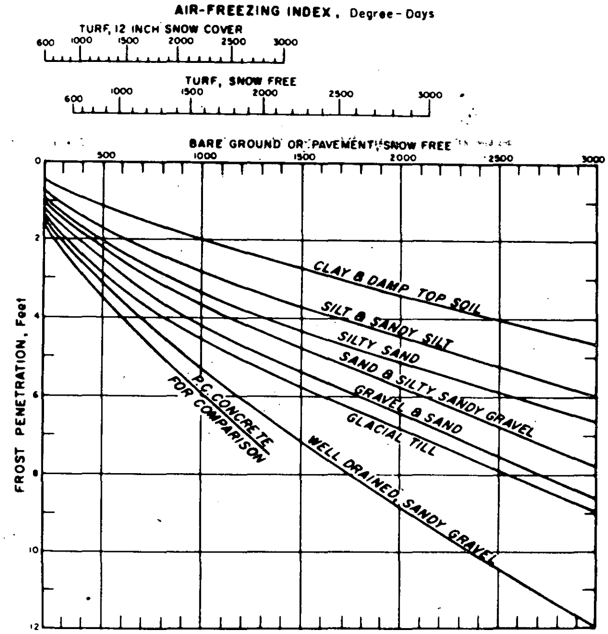

Figure 10. Relations among air-freezing index, surface cover, and frost penetration for various soil types (from Departments of the Army and the Air Force 1966).

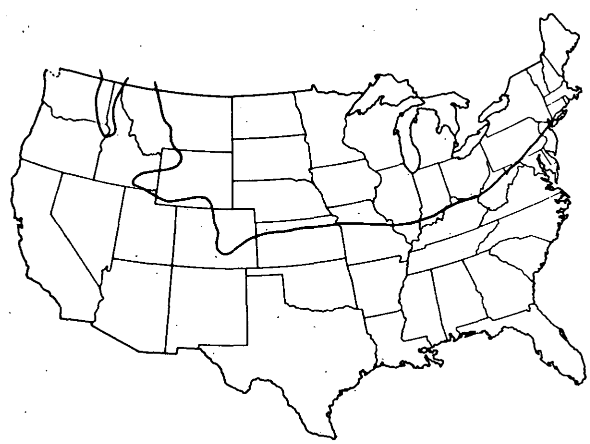

Figure 11. Southern limit of "hydrologically important" soil freezing in the U.S. (from Storey 1955).

Seasonally frozen ground. Seasonally frozen ground is highly variable in time as well as space, and no maps showing its extent were encountered. However, it is possible to provide some indication of the global extent of this phenomenon using air temperatures as an index. Bates and Bilello (1966, Fig. 1 and 4) presented maps of the Northern Hemisphere showing the limits of the areas with a mean temperature of the coldest month less than 0°C and "the limit of substantial frost penetration in the ground (approximately one foot) once in ten years." The latter limit is actually the 100 freezing degree-day isopleth (i.e. the line connecting those points at which $\sum_{i=1}^{n} (32 - T_i) = 100$, where T_i is average daily temperature for days when this is less than 32°F, and n is the number of days when $(32 - T_i) > 0$. The two limits are essentially coincident, as shown in Figure 7. Bates and Bilello (1966) did not discuss the other conditions that affect ground freezing, and it is assumed that the frost penetration they refer to is for bare ground.

Maps of mean freezing degree-days in North America were given by Gilman (1968) (Fig. 8) and for Eurasia by Scott (1963) (Fig. 9). Approximate relations between air freezing index (= freezing degree-days) for various soil types and ground conditions were given in Departments of the Army and the Air Force (1966) (Fig. 10). For the United States, Storey (1955) described the southern limit of "hydrologically important" seasonal ground freezing; this is illustrated in Figure 11. This is very close to Bates and Bilello's (1966) line of "substantial frost concentration" (Fig. 7) east of the Rocky Mountains.

Local variations

Permafrost. In the continuous permafrost zone, permafrost is present nearly everywhere, but may be absent beneath large lakes and rivers, as illustrated in Figure 12. Permafrost may also be lacking where there are thermal springs or high geothermal heat flow. In the discontinuous permafrost zone, permafrost is generally limited to north-facing slopes and/or fine-grained surficial deposits (Ferrians et al. 1969, p. 7), and this tendency becomes stronger as one proceeds southward. Dingman (1971) found that the presence or absence of permafrost in the discontinuous zone in central Alaska was related to "equivalent latitude," an index of insolation that depends on slope and aspect. The depth of the active layer was also found to be significantly correlated with this index. Immediate geological history and ground and snow cover may also strongly influence the occurrence of permafrost within the discontinuous zone. Detailed descriptions of the relations between local conditions and permafrost can be found for Alaska in Brewer (1955), Hopkins and Karlstrom (1955), Cederstrom (1963), Ferrians (1965), Péwé (1966), Ferrians et al. (1969), and Williams (1970); for Canada in Lindsay and Odynsky (1965), Owen (1967), Bird (1967), and Brown (1970); and for the U.S.S.R. in Stepanova (1956), Efimov (1957), Popov (1958), Ponomarev and Tolstikhin (1959) and Piguzova (1965).

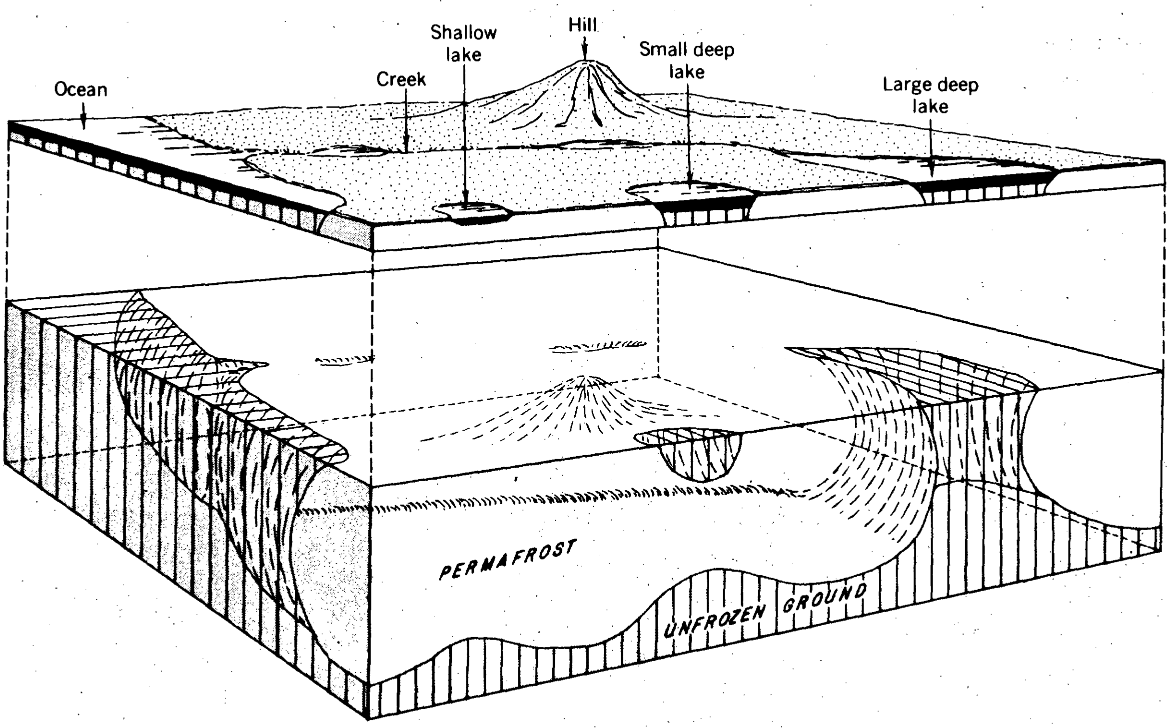

Figure 12. The effect of surface features on the distribution of permafrost in the continuous permafrost zone (from Ferrians et al. 1969).

Seasonally frozen ground. There are a large number of studies in the United States and Russian literature dealing with local variations in seasonal freezing and thawing of the ground. It seems useful to review and summarize this literature as a basis for understanding these variations, as this has not been attempted for the U.S. since Storey's (1955) synopsis. It was not possible to obtain and translate the extensive Russian literature on this, though a few translations were available and are discussed below.

The first discussion in the American literature is Hough's (1877) report that European investigators found considerably greater frost penetration in open fields than in forests. Zon (1912) also cited results of German research showing that forest soils may remain unfrozen while nearby soils in open fields are frozen to significant depths, and that when freezing did occur in forest soils, it was less deep than in the open. The difference was attributed mostly to the insulating effect of the litter layer on the forest floor.

Patten (1909) carried out laboratory studies that showed that mineral soil froze more solidly (that is, with lower porosity) than did humus. Pearson (1911), studying western yellow (ponderosa) pine forests, found that frost penetration beneath the forest litter was one-half of that beneath grass in natural parks. Similar results in the same forest type were reported by Jaenicke and Foerster (1915). MacKinney (1929), working in Connecticut, compared soil temperatures and freezing characteristics in undisturbed plots beneath white/red pine stands with those beneath areas from which the litter had been removed. He found that litter delayed the penetration of frost by one month, and that impermeable concrete frost formed in undisturbed plots.

In a study in Wisconsin, Scholz (1938) reported frost penetration to a depth of 4 in. in a hardwood woodlot as compared to 10 in. in a nearby bluegrass pasture. A similar study in England showed frost depths of 8.5 cm (3.3 in.) beneath bare sandy loam, 3.5 cm (1.4 in.) beneath grass, and no frost beneath undisturbed hazel brush (Salisbury 1939).

Diebold (1938) reported on snow cover, soil frost and runoff conditions near Ithaca, N.Y., during the very severe flooding of March 1936. He found that only the exposed, bare fields were frozen during the floods, and that these produced a large amount of runoff; nearby forested areas and meadows with herbaceous cover did not have frozen soil. The contrast was attributed to the insulating properties of the snow cover, which was retained best in forest, but was eroded from exposed, bare areas.

Goodell (1939) undertook a systematic study of the effects of vegetative cover and of slope aspect on soil freezing in Illinois. Using paired plots with similar aspects, he found that freezing occurred frequently and to an appreciable depth in pastures and cornfields during weather that produced no freezing in oak forests. The effect of aspect was also pronounced; freezing was considerably deeper on north- as opposed to south-facing slopes with identical vegetative cover (pasture). Apparently there was no snow cover during the study.

The effect of vegetative cover on frost depth and duration was also investigated by Kienholz (1940) in Connecticut. He found frost penetration to be about twice as deep in open fields as in six forest types (including conifers and hardwoods), with white pine having the least penetration. Figure 13 compares the progression of freezing and thawing in three cover types. Since snow cover was similar on all plots studied, Kienholz (1940) considered other factors responsible for the variations in frost penetration, principally the litter of the forest floor. "Under accumulations of leaves in depressions the frost barely penetrated to the mineral soil, whereas nearby in the same forest type, where the wind had swept most of the leaves away, frost penetrated into the mineral soil nearly as deep as in the plowed field" (Kienholz 1940, p. 349). His observations suggested that pine needles and hardwood leaves were equally effective insulators. Litter layers also retarded the time of ground thawing by as much as 10 days (see Fig. 13).

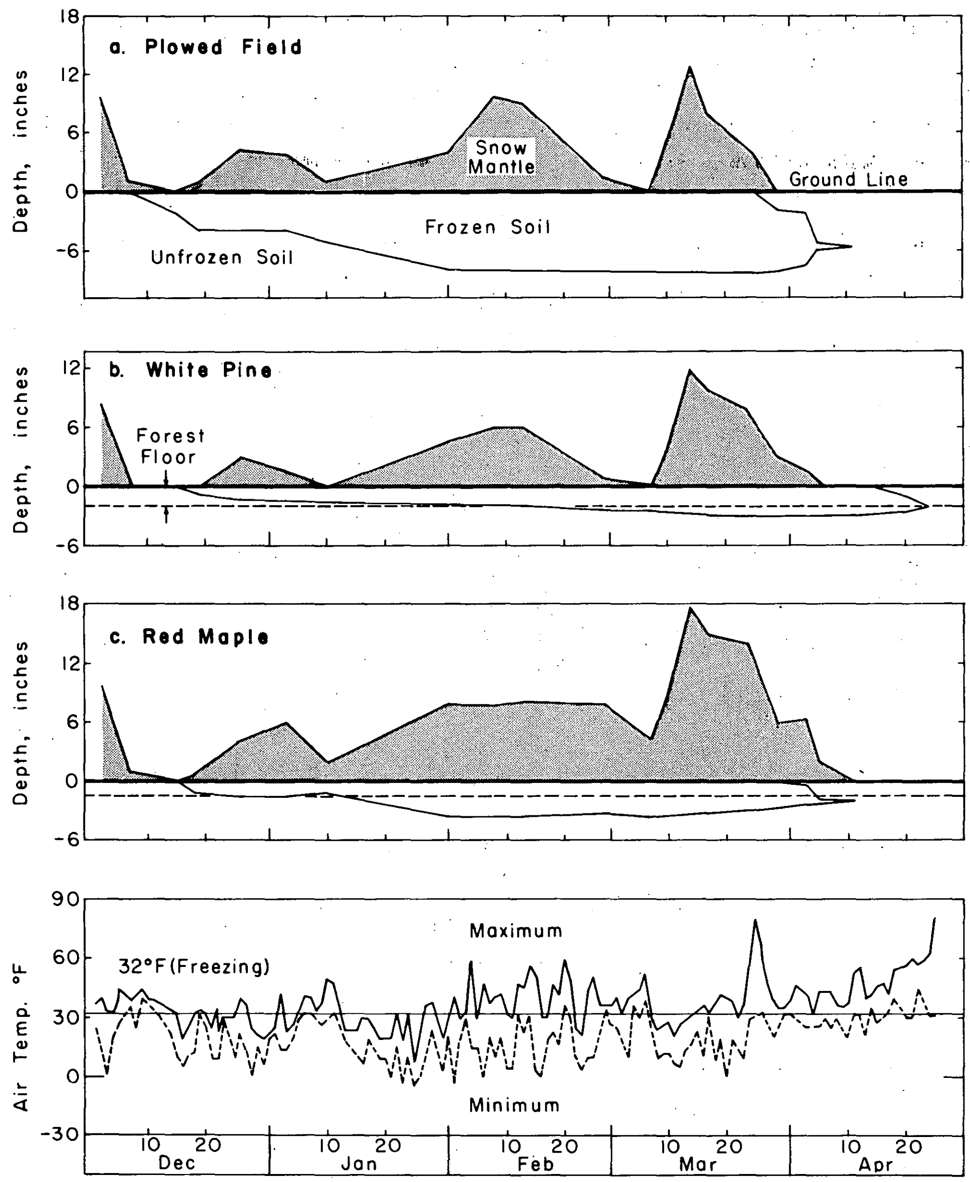

Figure 13. Snow depth, frost depth and litter thickness under three land-use conditions, and air temperatures, Connecticut (from Kienholz 1940). (Reprinted from the Journal of Forestry, April 1940.)

Dreibelbis and Post (1940), working in Ohio, also found that depth of frost penetration was much greater in cultivated soils than in pastures or forests. Further experiments and observations relating soil freezing and thawing behavior to environmental characteristics were made by Atkinson and Bay (1940) in Wisconsin. These field experiments showed a definite inverse relation between snow depth and frost penetration on a number of vegetative types. They even placed snow on a plot that had been kept snow-free through half the winter, and found a subsequent decrease in frost depth. They also examined the effect of forest litter, and reported that, with the same snow depths, frost depth was twice as great in pasture, and five times as great in plowed land, as compared to forests. These results are summarized in Figure 14.

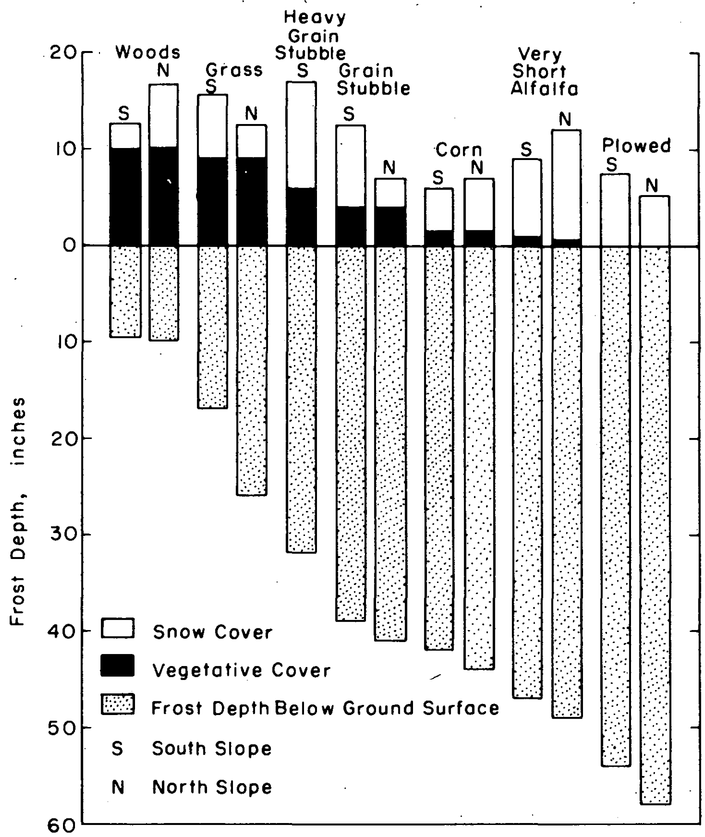

Figure 14. Snow depth, frost depth and vegetative cover at 13 locations, Wisconsin. (From H.B. Atkinson and C.E. Bay, EOS: Transactions of the American Geophysical Union, p. 935-948, 1940, copyright by American Geophysical Union.)

Similar observations on the effect of vegetation on soil frost were made by Belotelkin (1941) during three winters in northern New Hampshire. In addition to open fields, and hardwood and softwood forest, he examined a spruce swamp. The latter showed the greatest frost penetration, which exceeded that in the open field, while the softwood (spruce) forest had about the same penetration as the field. Hardwood forest had the least frost penetration. Belotelkin (1941) attributed these results to the effect of snow depth, noting that conifers intercept more snow and thereby reduce the insulating layer on the ground. However, he also noted effects of soil drainage and texture, and stated that frost penetrates deeper and remains longer in poorly drained and fine-textured soils than in well drained and coarse-textured soils. He also noted, as had earlier writers, that the litter layer is an effective insulator.

In observations of freezing of silt-loam soils at Coshocton, Ohio, Post and Dreibelbis (1943) found that ultimate frost penetration increased in the order: woodland, meadow, pasture, cropland (wheat). The thickness of the litter layer decreased in the same order, again suggesting its marked effect on soil freezing.

Sakharov (1945) reviewed the Russian literature on soil freezing and discussed the factors influencing it. He classified these factors as follows:

1. Climatic
   a. Regional climatic characteristics
   b. Seasonal weather variations

2. Orographic
   a. Altitude
   b. Relief
   c. Aspect
   d. Degree of slope

3. Edaphic – determining the thermal properties of soils
   a. Chemical composition, quantity of humus, concentration of soil solutions
   b. Physical properties, structure, color, moisture, etc.

4. Biotic
   a. Vegetation
   b. Fauna – surface and soil animals
   c. Human activity

The results of many years of observation and experimentation at forest sites in the European U.S.S.R. were presented. The insulating effect of snow cover was discussed, and its effect in maintaining a small spatial variation in soil temperature despite wide variations in air temperature was illustrated. He noted that temperatures generally decrease with elevation, while snowfall increases up to some point and then decreases, leading to widespread soil freezing at high elevations. Increased humus content leads to increased soil heat capacity and decreased conductivity, while increased sand content reduces heat capacity. Litter is also discussed, not only as an insulator, but also in relation to the physical properties of the soil beneath it. A number of experiments on the effects of litter on freezing are described that show a pronounced inverse relation between litter depth and freezing depth. In one experiment, even one year's leaf-fall measurably reduced frost depths as compared to a litter-free area, and in another a 3.5-in. layer of litter was found to have three times the effect of a 4.5-in. moss layer. Sakharov (1945) also emphasized that the depth to the groundwater table is a significant edaphic factor, with higher water tables associated with shallow or no freezing. The net effects of several of the edaphic and climatic factors, particularly snow depth, litter depth and depth to water table, associated with various forest types of the northern European U.S.S.R. were described. Also discussed were the effects of several vegetative factors, such as stand density and species composition; freezing was deeper beneath spruce than beneath hardwoods, due to snow interception by the conifers. Human activity, including logging practices, grazing, fires, and swamp drainage has significant effects on the many factors influencing soil freezing.

Sakharov (1945) notes that deep-freezing soils thaw only after the snow cover has melted, while shallow-freezing soils thaw before the beginning of the spring melt. He also characterized the thawing of forest types, as follows:

Type of thawing	*Type of forest*
Winter thawing (before beginning of spring thaw).	Oak, birch, aspen, linden, and mixed stands of these species without admixture of conifers.
Thawing during snow melt; frost disappears with the complete disappearance of snow.	Pine and pine-hardwood mixtures.
Thawing ends after disappearance of snow.	Spruce, pine with spruce understory.

Finally, Sakharov (1945) discussed the positive and negative effects of soil freezing in forests and means by which freezing can be reduced.

The influence of brush vegetation on soil freezing in the Sierra Nevada of California was studied by Anderson (1946). In this area, freezing air temperatures frequently occur at night, with daytime thawing. The natural low brush vegetation effectively prevented soil freezing, while adjacent plots that were stripped of vegetation froze frequently, to depths of 0.1 to 0.7 in., at night. Anderson (1947) described subsequent studies in the same area, in which average maximum frost depths under bare ground, grass and brush were found to be 2.6, 0.5 and 0 in., respectively.

Lassen and Munns (1947) pointed out that humus depth was a critical determinant in the formation of concrete frost. Factors influencing the formation of concrete frost were also studied by Bethlahmy and Reigner (1949). They examined forest, field and croplands on a large number of plots scattered throughout New England, New York and Pennsylvania. Many of their conclusions confirmed those of earlier workers: snow tends to prevent soil frost formation (no concrete frost was found beneath snow covers exceeding 27 in. depth), more concrete frost forms in poorly drained areas than well drained areas, depth of litter is inversely related to frost depth, and only 26% of the observations in wooded areas had frost, compared to 73% in agricultural open lands and 93% on bare ground. It was also noted that aspect had no effect on frost formation in the areas they studied, although frost persisted longest under north-facing slopes. Interestingly, these authors reported that the type of humus layer, as well as its thickness, affects freezing, but they did not elaborate. Burned areas had high amounts of concrete frost, while recently logged areas in which the forest floor had not been disturbed had little frost formation, again emphasizing the importance of litter.

Soil freezing conditions across a climatic-vegetational transect from the west slope of the Cascade Mountains to the rangelands of eastern Oregon and Washington were studied by Hale (1950, 1951). In the Douglas fir-lodgepole pine forests of the west slopes of the Cascades, no soil freezing was observed, as heavy snowfall preceded extended freezing temperatures. Snow deposition also prevented freezing, with the exception of short-lived granular frost in the litter, in the lodgepole pine forests of the upper east slopes. At lower elevations, in ponderosa pine, frost was found only in places where the snow was shallow (e.g. near the bases of trees), and was generally granular and confined to the litter. In the rangelands, impermeable frost occurred at one time or another in all cover types: open ridge tops, open grasslands, ponderosa pine and lodgepole pine, reaching up to 7.5 in. depth. Snow depth was one of the most important determinants of frost, and considerable thawing from the bottom took place when snow began to accumulate. Aspect was a factor in the timing of soil freezing and thawing, but not the depth or duration. Concrete frost was never observed to be continuous over large areas.

Byrnes (1951) reported the results of laboratory studies on the freezing of natural loamy sand, silt loam and clay loam. When frozen for 48 hours at 0°F and 15°F, all soils developed concrete frost, whether saturated with moisture or at field capacity. At 30°F, little or no concrete frost formed in any samples. Granular and honeycomb frost consistently occurred in the litter layers. Experimental work showed that litter and humus layers are effective in preventing concrete frost formation.

The effects of air temperature, snow cover and exposure on frost penetration in agricultural soils with various crop covers near Madison, Wisconsin, were studied by Bay et al. (1952). All three factors were important. Snow depth proved to be the major determinant of frost depth, with up to 3 ft of penetration in winters with little snow. Snow depths of 24 in. prevented frost penetration at −21°F, while an 18-in. snow cover maintained frost less than 12 in. deep at −12°F, but not at −21°F.

Factors affecting the occurrence of soil frost in the northwestern United States were reviewed by Bullard (1954). In that region, soil freezing is not continuous over wide areas; snow 90 cm (35 in.) deep is sufficient to prevent frost penetration, and such depths are common in the mountains. Previously frozen soil will begin to thaw from below beneath 25 cm (10 in.) of snow. Where snow is not deep, the forest canopy and litter are effective preventers of soil freezing; in fact, "litter on the soil surface and organic matter incorporated into the soil body appear to have the greatest effect on soil freezing ... litter 10 cm (4 in.) deep is generally sufficient insulation to protect the soil from freezing" (Bullard 1954, p. 130). Frost in the region generally melts before the end of snow melt, except in forest swamps.

Pierce (1956) and Pierce et al. (1958) reported the results of a study of the effect of a wide range of land use conditions on the frequency and depth of concrete frost. The study was conducted over two winters on 174 plots in south-central Maine, northeast New York, east-central New Hampshire, northwest Massachusetts, south-central New York, and northwest Pennsylvania. Results are summarized in Tables I and II. Greatest frost depths and highest rates of frost occurrence were found for non-forest, due to differences in organic matter, compaction, vegetative cover and snow depth. "When frost was present in open land it was observed at all points of measurement about three-fourths of the time. However, frost in the forest occurred at all points only about one-third of the time" (Pierce et al. 1958, p. 260). This frequency of complete frost distribution in the forest is higher than had been previously realized. A wide difference between frost conditions in hardwood and conifer forests was also noted: frost depth in hardwoods averaged about one-half that in conifers, and when frost did occur, it was found at all points in conifer stands 46% of the time as opposed to 23% of the time in hardwoods. This difference is attributed to lesser amounts of snow interception, and perhaps to differences in the structure of the litter and humus layer.

Table I. Comparative depths of concrete frost, northeastern U.S. (Pierce, 1956).

	Concrete frost depth (in.)					
Comparisons	*Penobscot S. Cent. Me.*	*Paul Smiths N.E. N.Y.*	*Bartlett E. Cent. N.H.*	*Hopkins N.W. Mass.*	*Deposit S. Cent. N.Y.*	*Pocono N.E. Pa.*
1. Open land	6.3*	6.4*	6.7*	3.5*	4.8*	4.2*
Forest land	3.8	3.3	1.4	2.7	3.2	2.1
2. Open land	6.3*	6.4*	6.8*	3.5*	4.8*	–
Forest reproduction	4.0	2.2	2.8	2.2	3.2	–
3. Bare cultivated	7.2*	7.1	7.2	5.7*	5.1	4.8*
Grass land	6.0	6.2	6.4	2.7	4.7	4.0
4. Conifer pole timber	–	5.2*	1.4	2.9	4.6*	2.4*
Hardwood pole timber	–	1.8	1.2	2.2	1.8	1.6
5. Conifer sawtimber	–	2.7*	1.7*	3.3**	4.8**	1.7
Hardwood sawtimber	–	1.0	0.0	2.3	1.8	2.1
6. Hardwood reproduction	–	1.2	2.8*	2.4	3.0*	–
Hardwood sawtimber	–	1.0	0.0	2.3	1.8	–

*Difference is highly significant.
**Difference is significant.

Table II. Average frost depth, occurrence, accumulated depth and snow depth for various land use conditions, northeastern U.S. (Pierce 1956).

Land-use condition	Frost: Depth (in.)	Frost: Percentage occurrence (%)	Frost: Accumulated depth (inch-days)	Snow depth (in.)
Non-forest				
1. Bare cultivated	6.2	86	624	5.6
2. Pasture	4.8	74	476	5.1
3. Hayland	4.9	77	456	6.6
Forest				
Disturbed				
4. Coniferous sawtimber logged	2.9	49	187	8.6
5. Hardwood sawtimber grazed	3.4	63	282	6.7
6. Hardwood sawtimber logged	1.6	32	66	10.0
7. Hardwood reproduction, brush	2.8	45	167	10.3
Undisturbed				
8. Coniferous plantation (12-18 yr)	3.5	68	264	4.9
9. Coniferous sawtimber	3.2	50	204	7.4
10. Hardwood sawtimber	1.4	27	55	10.5

Willis et al. (1961) studied the effects of fall soil moisture level on soil freezing and spring runoff in unvegetated agricultural soils in North Dakota. In contrast to many earlier workers, they reported that dry soils freeze faster and deeper than wet soils, and also begin to thaw two to three weeks earlier. These observations are consistent with the physics of heat flow in the soil, as water tends to increase the diffusivity and latent heat of the soil. Earlier workers reporting the opposite relation between frost depth and soil moisture were probably misled by their purely visual, rather than thermometric, identification of soil frost. Willis et al. (1961) also noted the insulating effect of snow depth on frost depth, and reported that dry soils thaw from the bottom, while wet soils thaw in both directions. For both soil moisture types, complete thawing was coincident with the disappearance of the snow cover.

In a study of the influence of forest type on soil frost in the Adirondack and White Mountains of the northeastern U.S., Lull and Rushmore (1961) found only sporadic occurrences of concrete frost. Concrete frost was present only 15% of the time beneath hardwood and conifer saw timber, 30% of the time beneath conifer saplings, and 40% of the time beneath hardwood saplings. Frost depths were shallow under all types, averaging 2 in., and frost was concentrated beneath conifer crowns, a direct effect of snow interception (Fig. 15).

Megahan and Satterlund (1962) studied infiltration in frozen soils beneath fields, hardwood forests, and conifer forests near Syracuse, N.Y. Their observations of frost occurrence were generally similar to those of many earlier reports: concrete frost was found most commonly in grassed fields, while forests had other frost types. Frost persisted for the entire winter beneath softwoods, but only for shorter periods under hardwoods and in fields (Fig. 16). Snow was deeper in fields than in conifer forests and the observed differences seem largely due to differences in snow depth in the three cover types.

Schumm and Lusby (1963) observed granular frost in the upper zones of unvegetated lithosols developed on shales in western Colorado. In spite of the lack of vegetation and a thin snow cover,

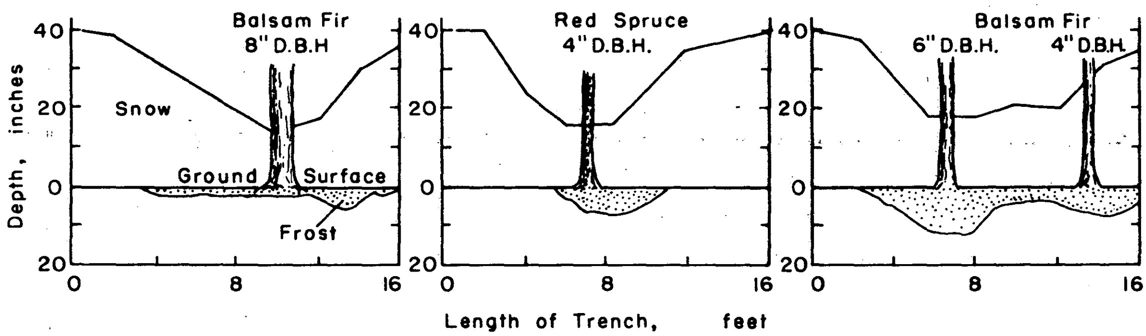

Figure 15. Snow and frost profiles under selected forest cover, New York (from Lull and Rushmore 1961).

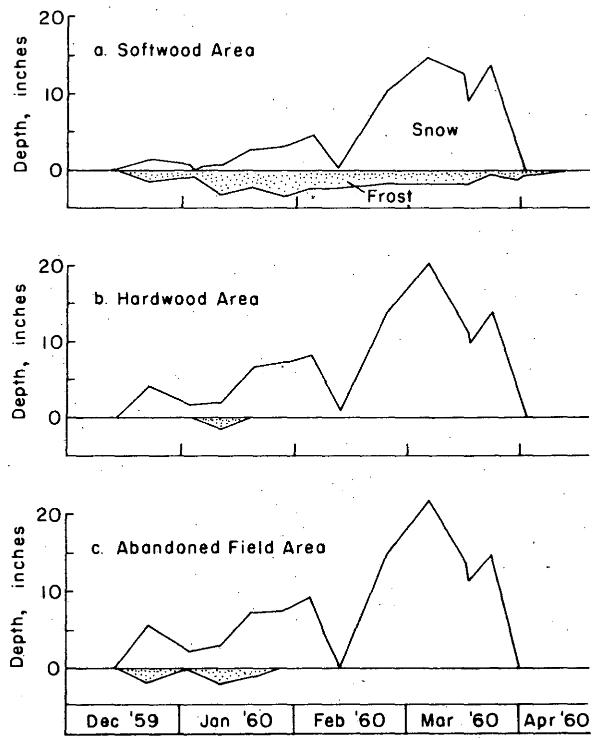

Figure 16. Frost depth as related to snow cover and vegetative type, New York (from Megahan and Satterlund 1962).

concrete frost was not reported, but rather an increased infiltration capacity developed as a result of the formation of granular frost.

Thorud (1965) observed the important insulating effects of litter and snow in a Minnesota oak forest by means of field experiments. His results are summarized below:

Soil treatment	*Freezing depth (in.)*
Natural oak forest	34
Soil compacted	36
Litter removed	39
No snow	47
No snow, soil compacted	52

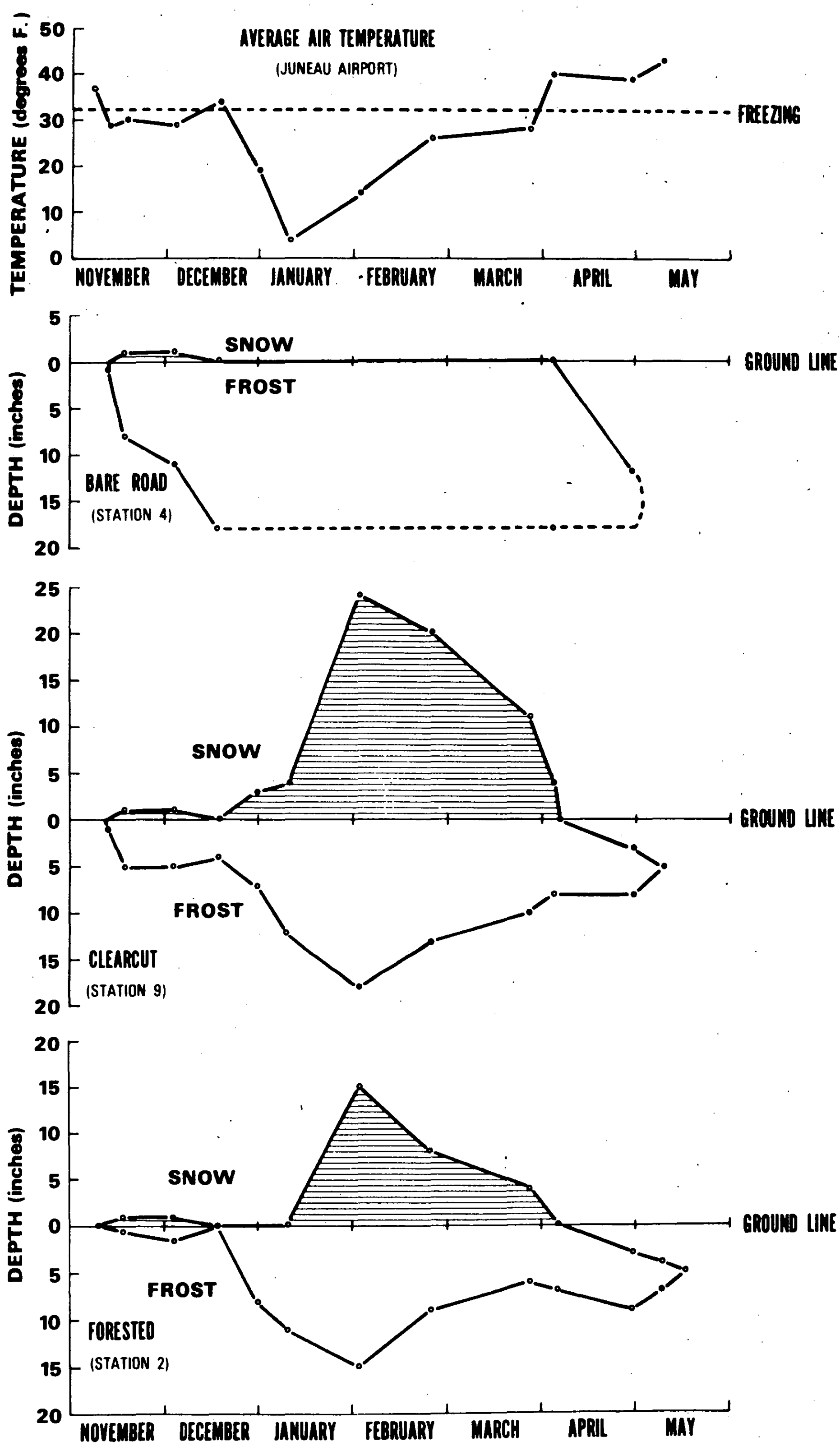

Figure 17. Air temperature, snow accumulation and depth of soil freezing under three conditions of soil cover (from Patric 1967).

In southeastern Alaska, Patric (1967) observed frost depths over one winter in:

1. Typical mature western hemlock (80%) and Sitka spruce (20%) forests
2. An area where this forest had been clear-cut without disturbance of the forest floor
3. A logging road

Figure 17 shows the results. In the early winter, concrete frost formed in the road, with honeycomb frost in the forested and clear-cut areas; concrete frost was found everywhere after 30 December. Maximum frost depths were very similar in all three plots, but the insulating effect of litter was manifested by more rapid freezing and thawing on roads than on clear-cut and forested plots. The effects of the canopy were to produce the slowest rates of freezing and thawing and a slightly smaller maximum thaw depth.

Mace (1968) examined soil freezing in the grasslands, mixed conifer forest, and aspen forest of the White Mountains of Arizona. Concrete frost occurred throughout the snow season in the silt loam soil of the open grassland, and was present even under 2 ft of snow for short periods. Granular frost occurred in the sandy loam soils of the conifer and aspen forests, but only intermittently, and to shallow depths (average 1.5 in.). Concrete frost increased soil moisture and decreased bulk densities, while granular frost did not affect those properties. Thawing in these soils occurred largely from the top.

In a study of the ablation of late-lying snowpacks in the Colorado Rockies (DeWalle 1969), concrete frost was frequently encountered under snow in exposed sites and above rocks near the surface in forested areas. This frost was generally less than 8 cm (3 in.) thick. Granular frost was primarily confined to pockets in the litter and humus layers in the forest. Ice layers at the soil surface, apparently formed by freezing of slush layers where snow had melted above saturated soils, were also observed locally.

From the above review of the literature on local variations of seasonal freezing, it is clear that, at least in the United States, a sufficient number of observational studies have been done in a number of geographical areas to fully establish certain generalizations: first, vegetative cover type is a major determinant of soil freezing characteristics, with the depth and rapidity of freezing increasing in the sequence: hardwood forests $\leq$ conifer forests < brush or fields < bare ground. Further, a large number of investigations, including some field experiments, have indicated that this sequence is largely a result of the combined insulating effects of litter and snow depth, the latter being dependent upon canopy interception of snow and exposure to eroding winds. In a few studies, the effects of slope aspect have also been investigated.

The field observations have also produced a number of statements relative to the effects of soil grain size, organic content and moisture content. However, in most cases the effects of these parameters were not carefully separated or quantified. A further confusing factor when reviewing the numerous studies on seasonal freezing is the general lack of a careful definition of what is meant by "frozen ground"; in most cases the presence of frost has been determined visually or by probing, rather than by reference to soil temperature.

Thus, although quantitative observational data have been given on the effects of these parameters in several geographic areas, none of the above studies provides a sufficiently general insight into the dynamics of soil freezing and thawing to allow quantitative application of results to other areas. They do not provide answers to the general question: "Given a particular soil, litter layer, snow cover, canopy, and slope aspect, how fast and how deep will the soil freeze, and how fast will it thaw?"

It would seem, then, that further purely observational studies of local variations in seasonal freezing should have low priority, and that attention should be turned to mathematical models of soil temperature incorporating layers of varying thermal properties and the surface energy balance. A first step in this direction is the conceptual algorithm for needle ice formation described by Outcalt (1971).

One of the more recent and promising attempts at modeling temperatures in layered soils is that of Nakano and Brown (1972). Other approaches are described in Scott (1963), Sanger (1966), Departments of the Army and the Air Force (1966), Aldrich and Paynter (1966) and Aitken and Berg (1968).

The importance of using an objective definition of frozen soil, which is generally lacking in the literature on seasonal freezing, should be emphasized again. Most logically, this definition should be based on soil temperature, as in the permafrost literature. Observations based on visual inspection or probing are not adequate. An important advantage of this, in addition to a reduction in confusion, is that all modeling attempts must be based on heat flow and hence on temperatures. Rickard and Brown (1972) have described an inexpensive and simple frost tube that allows the position of the 0°C isotherm to be readily determined, non-destructively and with an accuracy acceptable for most field studies. This appears to be an attractive alternative to installation of thermocouple and thermistor strings in many instances.

HYDROLOGIC EFFECTS OF FROZEN GROUND

Introduction

Conditions at the ground surface and immediately below it are the major determinants of the amounts and rates of water entering the various pathways of the land phase of the hydrologic cycle. Thus, these conditions, and in particular the presence of frozen ground, either directly or indirectly affect rain and snow interception, evapotranspiration, overland flow, infiltration, snowmelt, streamflow, groundwater recharge and movement, water quality, sheet erosion and bank erosion. Some of these influences have been the object of considerable research, which is reviewed below, while others are little known and must be discussed largely by inferential application of general principles.

Infiltration and overland flow

Importance of infiltration. Since the work of Horton (1939), infiltration (the movement of water downward across the soil surface) has been considered to be a critical hydrologic process. Horton developed the concept that, if the rainfall intensity exceeds the infiltration capacity of the soil, overland flow (surface runoff) will occur. This overland flow will be generated at a rate equal to the difference between rainfall intensity and infiltration capacity. The infiltrating water enters the unsaturated zone of the soil, to replenish soil moisture and either to be taken up as evapotranspiration or to percolate to the water table as groundwater recharge.

While this concept is based on flawless logic, it has become increasingly recognized that, in many places, infiltration capacities are almost always many times greater than any except extremely improbable rainfall intensities, and other mechanisms of storm runoff generation must occur. As Freeze (1972, p. 1273) pointed out:

> In large areas of the semi-arid western United States, [Horton's] concept is quite acceptable, but in the vegetated humid basins of the east, it is not. Rather, storm hydrographs originate from small but consistent portions of upstream source areas that constitute no more than 10%, and usually only 1-3%, of the basin area, and even on these restricted areas only 10-30% of the rainfalls cause overland flow.

Freeze (1972) went on to show that, in humid areas, storm runoff arises from the infiltration of rain in areas near streams where the water table is usually close to the surface. When sufficient rainfall has infiltrated in these limited areas to cause the water table to reach the surface, overland flow to the channel begins. Because these storm-runoff source areas generally enlarge as rainfall continues (Ragan 1968, Betson and Marius 1969, Rawitz et al. 1970, Dunne and Black 1970a, b, Hills 1971, Dingman 1971), this has been called the variable-source area, or partial-area- contribution mechanism.

Clearly, the presence of frozen ground near the surface would be expected to have a direct effect on infiltration capacity, which would have a major effect on the generation of storm runoff by the Horton mechanism, and on the amount of recharge to the soil moisture and groundwater. This effect has been the focus of a number of investigations, which are reviewed below. However, because the understanding of the variable-source area mechanism is so recent, and because the influence of frozen ground on that mechanism is somewhat indirect (see Dingman 1973a), few studies exist exploring this process.

The infiltration process. The physical processes by which water infiltrates unfrozen soil, and the factors affecting this infiltration, are presented in most texts on hydrology, and a complete review of the subject is outside the scope of this report. A good qualitative discussion is given by Ward (1967, p. 180-193), while Wesseling (1961) and Eagleson (1970, p. 291-299) review the quantitative aspects.

When liquid water is present at the soil surface, and the soil is unsaturated, infiltration generally occurs due to gravitational and capillary (surface tension) forces. The rate at which infiltration occurs depends on a number of factors, which have been classified by Ward (1967, p. 181-189) as follows:

- Rainfall intensity
- Soil surface conditions
  - Compaction
  - Inwashing of fine particles
  - Depth of surface storage
  - Sun cracks
  - Slope
  - Frost
  - Cultivation
- Surface cover conditions
  - Vegetation type
  - Litter
  - Snow
  - Urban areas
- Transmissibility of the soil mass
  - Non-capillary porosity
  - Biotic factors
  - Soil moisture content
- Characteristics of the infiltrating water
  - Temperature
  - Water quality

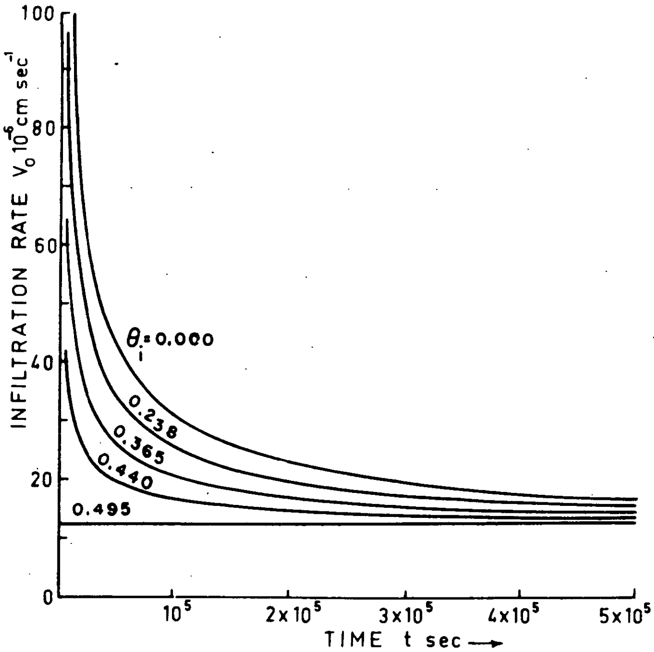

Figure 18. The influence of initial soil moisture content, θ_i, on infiltration rate (Wesseling 1961).

A number of experiments have shown that infiltration capacity, the maximum rate at which infiltration can occur, decreases with time as infiltration proceeds. An example of this is shown in Figure 18, which also illustrates the effect of the initial moisture content on infiltration capacity. The principal cause for this decrease with time is that the capillary potential gradient at the surface decreases as the soil is wetted. Other factors, such as compaction of the surface soil by raindrops, clogging of surface pores by inwashing of fine particles, and the swelling of clays, may also contribute to this phenomenon. Figure 19 shows the moisture content profile at successive times during infiltration of a clay soil.

The process of infiltration can be described by the following relation (Eagleson 1970, p. 282):

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[D(\theta)\frac{\partial\theta}{\partial z} + k_z(\theta)\right] \tag{2}$$

where θ is volumetric water content, t is time, z is distance measured vertically from the ground surface, D is the diffusivity of water in the soil (which can include vapor as well as liquid flow), and k_z is the vertical hydraulic conductivity. The first term in the brackets represents the capillary sorption process, and the second the effect of gravity. As indicated in eq 2, k_z is a function of θ; a curve relating these two parameters is shown in Figure 20. Diffusivity D is the product of the hydraulic conductivity k_z and the slope of the moisture tension (ψ) vs water content curve (the " moisture characteristic" curve of the soil) at the water content of interest. An example of the moisture characteristic curve is also shown in Figure 20.

Exact solutions to eq 2 have not been found, but a useful simplification can be made by omitting the effects of gravity. This term is negligible when k_z tends to be uniform or when the gravitational potential gradient is small relative to that of the capillary gradient. These conditions are met when saturation is low, as in the early stages of infiltration, and experiments show that neglecting the

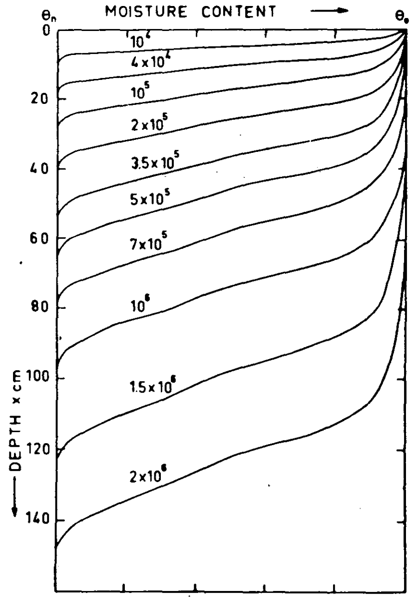

Figure 19. Computed moisture profiles during infiltration into Yolo light clay. (Numbers on curves are times in seconds after beginning of infiltration, Wesseling 1961.)

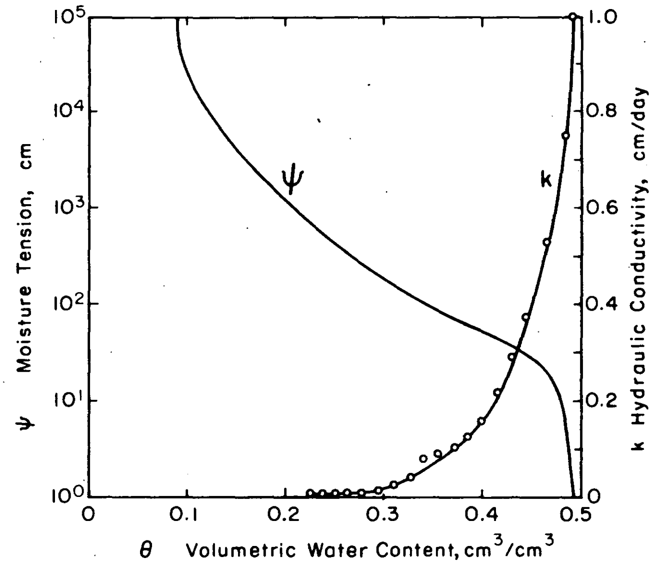

Figure 20. Relation of moisture content θ to moisture tension ψ and hydraulic conductivity k for Yolo light clay (from Wesseling 1961).

gravitational term gives results which are qualitatively correct (Eagleson 1970, p. 292). The experiments of Horton and Hawkins (1965) indicate that even where large pores occur in a relatively fine-grained soil, the capillary gradient from the large pores to the small pores is sufficient to cause horizontal flow to the finer pores. Thus the entire soil column is wetter during infiltration, and gravity drainage in the larger pores does not dominate during infiltration.

It has also been found that good approximations to measured infiltration capacity can be obtained when a constant diffusivity is assumed in eq 2. Including the effect of gravity, this results in the relation

$$f = (\theta_i - \theta_0)(D/\pi t)^{1/2} - k_0 \tag{3}$$

where f is infiltration capacity, θ_i is initial water content of the soil (assumed uniform), θ_0 is the water content at the surface, and k_0 is the hydraulic conductivity at the surface (see Eagleson 1971, p. 292-293 and p. 412-416 for mathematical development). Equation 3 may be compared with infiltration equations proposed by Kostiakov (1932):

$$f = bt^a \tag{4}$$

where b is assumed to depend on the saturated hydraulic conductivity of the soil and $a = -\frac{1}{2}$, and with the well-known Horton (1939) equation,

$$f = f_\infty + (f_i - f_\infty)\, e^{-\beta t} \tag{5}$$

where f_∞ is the final infiltration capacity, f_i is initial infiltration capacity, and β is a constant that depends on the diffusivity of the soil.

Experiments have shown (see Wesseling 1961, p. 14) that eq 3 describes the infiltration process very well, eq 4 moderately well, and eq 5 quite poorly, in spite of requiring specification of three parameters.

Infiltration in frozen ground. As with the topic of local variations in seasonal freezing, the literature on infiltration in frozen ground has not been thoroughly reviewed, heretofore, at least in English. It therefore also seems useful to attempt such a chronological review herein.

Shalabanoff (1903) was one of the earliest to discuss the permeability of frozen soil. In studies in western U.S.S.R., he found that almost 80% of the precipitation falling between September and May, when the ground was frozen, did not run off. On the basis of these observations and data on the thermal and water table conditions in the soil, he concluded that most of this water infiltrated through the frozen ground.

Until the 1940's, it was apparently generally held by American hydrologists that frozen ground is completely impermeable. This belief is implied by Diebold's (1938) analysis of spring floods in the northeast. However, Kienholz (1940, p. 349), working in Connecticut, noted that " A frozen layer of duff (litter) is still very porous and allows winter rains and the water from winter thaws to penetrate through the duff into the mineral soil." The first direct investigation of infiltration in frozen ground in the U.S. was that of Augustine (1941), who carried out infiltrometer experiments on a gravelly silt loam with covers of forest, pasture and corn stubble. He found that initial infiltration rates in forest soils frozen to a depth of 4 in. were little different from those in the same soils in the unfrozen state. However, the rates decreased markedly after the initial wetting; this was due to the formation of concrete frost upon freezing of the infiltrated water. No infiltration occurred on corn and pasture areas frozen to 3 to 4 in., except around frost-heaved stones and in animal burrows, but with only 2 to 3 in. of frost, pastures allowed measurable infiltration. He also reported that little or no thawing occurred during the infiltrometer runs, but rather there was frost penetration and the formation of less permeable frost.

Based on field observations in New Hampshire over several winters, Belotelkin (1941) concluded that the formation of honeycomb frost in the surface layers of forest soils maintains a high infiltration capacity. In open fields, however, freezing reduced permeability. A heavy December rain falling on frozen soils in both forested and open areas produced less than ½ in. of thaw in the open, but infiltrated and completely thawed the forest soils. This latter observation is contrary to those of Augustine (1941), reported above.

In a review of the state of knowledge concerning the role of snow, ice and frost in the hydrologic cycle, Horton (1941) noted that there were almost no data on the influence of frozen ground on infiltration. There were indications that frozen soils could absorb significant amounts of water from rain and melting snow.

Post and Dreibelbis (1943) reported on lysimeter experiments on agricultural soils, pasture, and hardwood forests in Ohio, and found that percolation rates were often significantly reduced, sometimes to zero, when frost depth exceeded 3 in. More complete hydrologic studies of a similar range of cover conditions in Michigan (Garstka 1944) showed that the presence of soil frost in agricultural lands significantly reduced infiltration and increased runoff, while forested areas generally had very little runoff.

Both permeable and impermeable frost were observed in the rangelands of eastern Oregon by Hale (1950, 1951), and it was noted that 1 in. of concrete frost was sufficient to cause surface runoff.

Although the question of infiltration into frozen ground was not extensively covered in Straub's (1950) broad discussion of arctic and subarctic hydrology, he did note that the permeability of any frozen saturated soil is small, but that intense freezing may cause surface cracks that increase infiltration rates in the spring. It is also interesting that Straub (1950) remarked on the very extended hydrograph recessions that are characteristic of arctic and subarctic streams, and attributed this phenomenon to "infiltration into a surface layer of high absorptive capacity" (p. 28). However, he did not discuss the nature of the surface, i.e. whether frozen or unfrozen soil, moss, etc.

In the course of experiments relating soil texture and type of soil frost, Byrnes (1951) noted that litter and humus "are capable of absorbing a certain amount of water, while concrete frost in the mineral soil is completely impermeable." Without presenting any data on infiltration, Bethlahmy (1953) stated that concrete frost is impermeable, while infiltration occurs readily in granular frost. He went on to point out that soil resistance blocks (and, by implication, temperature sensors) do not differentiate among the types of ground frost and therefore do not provide critical information to the hydrologist. He also noted that, because water is drawn to the freezing front, granular frost may gradually become concrete frost as freezing progresses.

In a review of the effects of soil frost on runoff and erosion in the northwestern United States, Bullard (1954) again noted that concrete frost is impermeable, while other types permit some infiltration. Because concrete frost occurs only locally, generally being restricted to areas of sparse plant cover and artificially disturbed soil, it only promotes increased surface runoff in small upland watersheds where those conditions exist.

Storey (1955) gave a more extensive review of the hydrologic effects of seasonal ground freezing throughout the United States, and noted that concrete frost is impermeable, while other types of frost have little effect on infiltration. This conclusion was based on earlier studies that showed up to 100% runoff from bare fields where there was concrete frost, and progressively lower runoff percentages as the area of concrete frost decreased in other cover types.

Beginning in 1957, a series of papers reporting the results of investigations into the permeability of frozen ground in the U.S.S.R. became available in English translation (Tsykin 1956, Komarov 1957, Mosienko 1958). As indicated below, this appeared to spur interest in similar investigations in the United States, as until this time there had been no serious study of the problem except for the brief report of Augustine (1941).

Tsykin (1956) made careful observations of soil moisture and soil temperature during infiltrometer experiments and natural snowmelt conditions in an agricultural area near the Volga River, at about 52°N. The soils were heavy clay loams, with five different cover types: stubble field, unplowed field, fall-plowed field, orchard, and young forest. Soil temperatures were around −7°C (19°F), and frost depths were about 2 m (6 ft) in the stubble and unplowed fields. In infiltrometer tests, these sites absorbed very little water, as infiltrating water immediately froze, preventing further infiltration even in dry soils. However, fall-plowed soils absorbed water even at −5° to −7°C (19° to 23°F), and could continue to allow infiltration until saturated. This difference in behavior was attributed to the presence of large water-free pores in the plowed soil. The experiments also showed that some absorption of water into the upper layers of unplowed and fall-plowed soils is possible, but the water freezes in the upper 30 cm (1 ft), and this eventually limits further infiltration.

In the forest soils, at temperatures of −1° to −2°C (28° to 30°F), the infiltrometer tests showed infiltration through the entire 40-cm (1.3-ft) frozen depth, and warming of the soil to 0°C (32°F) in this layer. An infiltration velocity of one-quarter of the thawed value was observed in these soils.

It was noted that, in the infiltrometer experiments, the release of latent heat as some freezing occurs allows water to enter thin capillaries ahead of the wetting front, and this process continues until saturation is reached and the saturated layer is at 0°C. There is then a sharp decrease in temperature as well as moisture below the soaked layer.

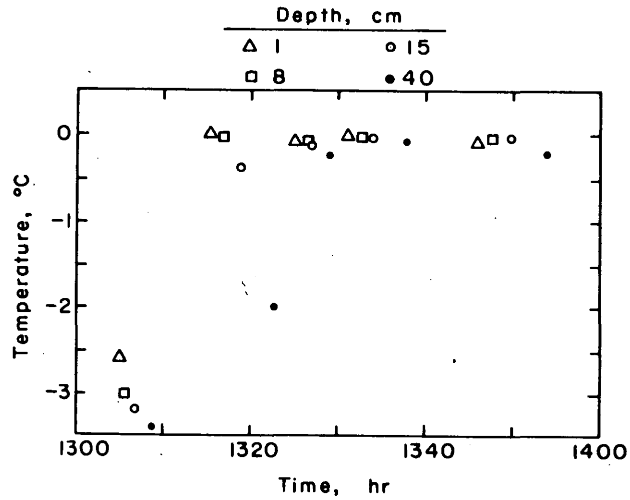

Figure 21. Temperature as a function of time at several depths during infiltration into frozen sand (from Komarov 1957).

Under natural conditions, the large increase of soil moisture and temperature in the wetted layer does not occur, because much smaller amounts of water are infiltrating as compared to the infiltrometer experiments. With smaller and intermittent influxes of water, enough heat is conducted downward from the wetting front to cause freezing in the early stages of infiltration.

Tsykin (1956) concluded that the permeability (i.e. the presence of water-free pores) and the temperature of the soils are the basic factors influencing the infiltration of water into frozen soil. These conclusions confirm the general observations of American writers, that concrete frost is impermeable, while other types of frost permit infiltration.

A more controlled series of investigations into the permeability of frozen soils was reported by Komarov (1957). In these, uniform sand (grain size characteristics: 20% 1.00-0.25 mm; 69% 0.25-0.05 mm; 1% 0.05-0.01 mm; 1% less than 0.01 mm) was frozen in a cylinder in which thermocouples were placed at depths of 1, 8, 15 and 21 cm. Initial temperatures in the sand ranged from about -3°C (26.5°F) to -5°C (21°F). Water at a temperature of 0°C (32°F) was maintained at a constant level on the surface, and the volume of infiltrating water was measured as a function of time. The initial water contents of the sand were 2 to 17% (porosity was 40%). In almost every case, the temperatures at a given level rose abruptly from about -3°C (26°F) to about -0.2°C (31.6°F) as soon as water infiltrated to that level (Fig. 21). This was due to the release of latent heat as the infiltrating water froze. The infiltration rate decreased with time, and reached a nearly constant value, just as in unfrozen soils (Fig. 22). Komarov (1957) stated that this decrease in infiltration with time was due to a decrease in the capillary potential, as indicated by the infiltration theory discussed earlier. The final constant rate of infiltration ("coefficient of filtration") reached in all the experiments was found to be closely related to the initial ice content of the medium (Fig. 23). Komarov pointed out that, with an initial moisture content approaching the capillary moisture (field capacity), the coefficient of filtration becomes very low, even though there are still a large number of air-filled pores. He also found that the total water content of sand (liquid plus ice) remained considerably above field capacity of the unfrozen sand for several hours after completion of the experiments, indicating a mechanism for supersaturating frozen soils. It should be noted that Komarov's data are for much faster rates of infiltration than could occur in nature.

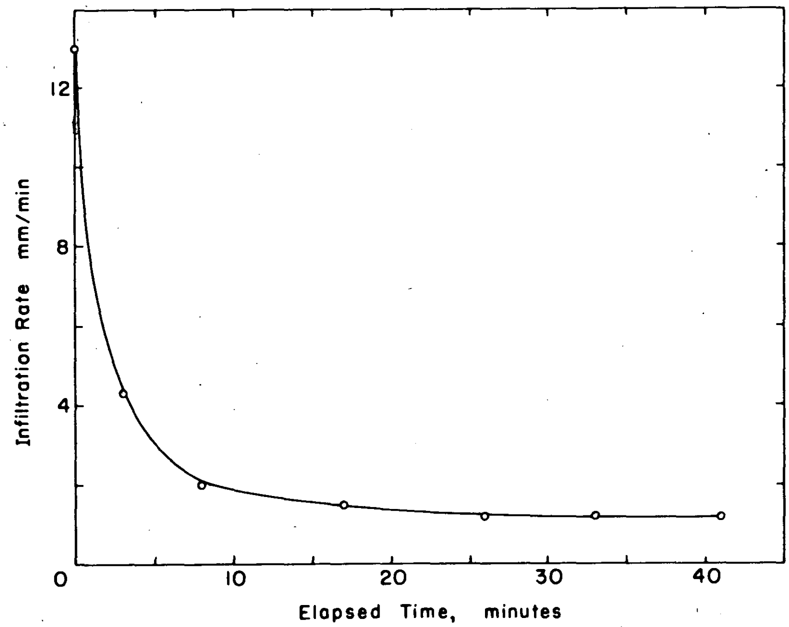

Figure 22. Infiltration rate into frozen sand as a function of time (data from Komarov 1957).

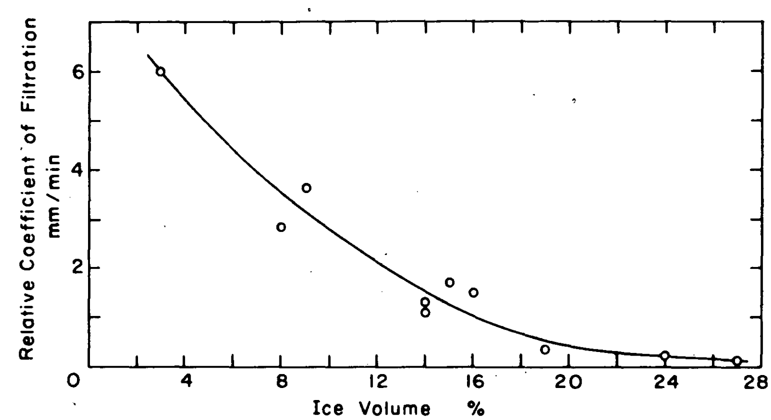

Figure 23. Coefficient of filtration as a function of ice volume for a frozen sand (from Komarov 1957).

Mosienko (1958) briefly reported on the permeability of frozen loamy chernozem soils from the steppe region of the U.S.S.R. He concluded that, when a soil has no ice-free pores (concrete frost), it is impermeable, and if the soil has more than 50% of its pores free of ice, infiltration occurs regardless of temperature.

The first extensive field infiltration experiments on frozen soil in the U.S. were reported by Trimble et al. (1958). They carried out about 100 ring infiltrometer tests in forests and fields in New Hampshire to discover the effects of frost type on infiltration. Only concrete and granular frost types were investigated, as other types seemed to be rare in the region. These tests were carried out on soil from which the litter layer had been removed, and which had been thoroughly wetted before testing. The results are indicated in Table III.

Table III. Results of field infiltrometer tests of Trimble et al. (1958).

Cover	Soil	Mean infiltrometer rates (in./hr)		
		No frost	Granular frost	Concrete frost
White pine	Sandy loam	50	204	0
Hardwood	Loam	14	16	0
Abandoned pasture	Sandy loam	9	–	0
Hardwood	Sandy loam	30	–	0
Red spruce	Sandy loam	14	–	0
Lawn	Sandy loam	11	–	0

Clearly, concrete frost was impermeable, except for root holes and burrows. The importance of these larger pores could not be assessed in the experiments, but they occur only in forest soils. Interestingly, granular frost significantly increased infiltration rates over unfrozen values in the two cases where they could be compared. Trimble et al. (1958) also noted that the occurrence of concrete frost in forests is very spotty, so any water running off such areas quickly reaches a place where it can infiltrate.

Relative to the discussion of mechanisms of runoff production given earlier in this report, it is worth noting that all the results in Table III (except, of course, for concrete frost) show infiltration rates much higher than any reasonably expected values of rainfall or snowmelt intensity.

Noting the scanty existing information on infiltration in frozen ground, Stoeckeler and Weitzman (1960) ran a series of infiltrometer tests on three frost types in northern Minnesota. The frost types were concrete, "porous concrete" (similar in appearance to concrete frost, but permeable to air; confined to sandy soils), and "partially frozen" (much like granular frost, except that many parts appeared unfrozen). Soils were silt loams and loamy sands, and frost depths ranged from 0 to 15 in. Results are given in Table IV. Concrete frost had a low, but not zero, infiltration capacity in both soil types, and all soils had significantly lower capacities when frozen than when unfrozen. Loamy sands had considerably higher rates than silt loams, both frozen and unfrozen. It was also noted that the silt loams had essentially zero infiltration for almost one hour. The soils were not pre-wetted in this study, and the rates in the second ten minutes were generally lower than in the first ten minutes, with the hourly rates substantially lower than both.

Megahan and Satterlund (1962) studied infiltration in frozen soils near Syracuse, N.Y. They randomly selected plots on fine sandy loam soils and, using infiltrometers, applied water until a constant rate of infiltration was achieved and compared these constant rates for different frost types through the winter. Several different land-use types were included in the study. In hardwoods, concrete frost was found on only one day, and allowed no infiltration. In a field, thin porous-concrete frost was found on three days, and infiltration rates under these conditions were about the same as those measured in unfrozen conditions. Frost occurred most often in a spruce forest, where infiltration rates in granular frost were much higher than in unfrozen soil, while rates in concrete frost were very low. They concluded that, with all cover types, granular and porous concrete frost give significantly more rapid infiltration capacities than unfrozen soil, while concrete frost gives consistently lower capacities. Depth of frost did not affect infiltration rate. The major factor controlling infiltration in frozen soils was the number and size of ice-free pores, as Tsykin (1956) had concluded.

Table IV. Infiltration on loamy sands and silt loams.
From Stoeckeler and Weitzman 1960.

Period of tests and frost type	*Infiltration for period indicated (in.)* 1st 10 min	2nd 10 min	60 min total	*Number of trials* 10 min	60 min
		Loamy sands			
Mar-Apr 1957					
Concrete	0.12	0.12	0.47	44	44
Porous concrete	0.78	0.71	2.19	33	23
Partly frozen	1.13	0.93	3.97	38	22
Unfrozen	2.37	2.77	–	8	0
Sept 1957					
Unfrozen	2.73	1.65	13.22	46	46
Least sig difference at 5% level	0.55	0.45	0.58	–	–
		Silt loams			
Mar-Apr 1957					
Concrete	0.05	0.03	0.09	50	50
Unfrozen	0.32	0.15	0.88	5	5
Sept 1957					
Unfrozen	1.19	0.56	4.20	18	18
Least sig difference at 5% level	0.24	0.19	1.11	–	–

Larin (1963) developed a method for measuring the air permeability of frozen soils, and for relating air permeability to water permeability. The latter was measured by flooding small plots. The soils studied were plowed agricultural soils (clay loams), with some plots covered to bring about low moisture contents, some left open to achieve a natural moisture level, and others artificially wetted to high water contents. Both water and air permeability varied directly with moisture content, and this was the major factor controlling infiltration. In every case when the frozen soil was relatively dry, it was permeable to air and water. On the basis of the air permeability data, which were measured at various levels in the soil, Larin (1963) concluded that the major determinant of water entry into the soil is the permeability in the upper 5- to 10-cm layer.

In a study of erosion processes and sediment yield on shale lithosols in the semi-arid climate of western Colorado, Schumm and Lusby (1963) found that infiltration capacity varied markedly with the seasons. In this case, the growth of granular frost crystals in the surface layers of the clay soil, and the accompanying desiccation, loosened the soil and formed aggregates, resulting in infiltration capacity increasing through the winter and reaching a maximum in the spring. During the summer, rain compaction and swelling of clays reduced the infiltration capacity. The authors suggested that this cycle may be common where fine-grained, largely unvegetated soils exist in an arid or semi-arid climate.

The long-term study of infiltration into thawed and frozen soil by Subbotin and Dygalo (1963) in the U.S.S.R. indicated that soil freezing did not decrease infiltration over wide areas sufficiently to cause extensive surface runoff from snowmelt or spring rain. Interestingly, their work suggests that the partial-area contribution mechanism of runoff generation was operating, but unfortunately few details of the study were given in the English summary.

Haupt (1967) reported the results of one winter's study of infiltration, overland flow, and soil erosion on frozen and snow-covered experimental plots on the east slope of the Sierra Nevada in Nevada. The soils were a cobbly sandy loam, and the ground covers were bitterbrush, sparse grass, pine litter, burned pine litter, rocky bare soil and rocky furrowed soil. The three frost types observed were concrete, porous concrete and stalactite. The latter type absorbed rainfall and snowmelt "literally like a sponge" (Haupt 1967, p. 159), and very little overland flow occurred where this was present. Porous concrete frost also improved infiltration over the unfrozen condition where there was appreciable plant and litter cover, and where frost cracks formed around exposed rocks. However, concrete and porous concrete frost reduced infiltration and increased overland flow under other ground conditions. The study also indicated that soil frost deteriorates rapidly during rain if there is no snow cover. As stalactite frost melted, infiltration capacity decreased, but the melting of porous concrete frost caused a progressive increase in infiltration capacity.

Runoff plot studies were used by Mace (1968) to study soil freezing and its effect on infiltration in the White Mountains of Arizona. Concrete frost formed in silt loams beneath open grasslands, and significantly reduced infiltration and increased runoff. In the sandy loams of the forest areas, the granular frost melted during the beginning of snowmelt, and did not affect infiltration. Mace (1968) noted that granular frost may tend to increase infiltration capacity because of the aggregation that freezing sometimes produces.

Bloomsburg and Wang (1970) measured the air permeability of frozen soil columns in the laboratory, and established a relation between air permeability and porosity n and initial moisture content θ_i. When $n(1 - \theta_i) < 0.13$, permeability approached zero in all soil types tested. It was suggested that this criterion should be tested for a wider range of soils and might be useful for flood prediction in conjunction with a field sampling program.

In a detailed study of snowmelt runoff production on a steep hillside in Vermont, Dunne and Black (1971) observed the effects of soil frost. Concrete frost was widespread at the beginning of the spring melt to depths ranging from 0.20 to 0.83 ft, and its presence was attributed to the infiltration and subsequent freezing of meltwater produced during winter thaws. A surface ice layer that had formed in the same way was also found in places. The concrete frost had a very low permeability, and significantly affected the timing of snowmelt runoff. This frost was not completely impermeable, however, as some larger pores (worm holes) were found in it which were conducting water.

As with the literature on local variations in seasonal freezing, a review of the above studies indicates that certain generalities have been well established. Almost every study has indicated that concrete frost is virtually impermeable. Other types of frost, particularly granular, either had little effect on infiltration capacities or increased them, in some cases markedly. Bloomsburg and Wang (1970) were the only authors to suggest a quantitative operational criterion for estimating the permeability of frozen ground.

It would seem that the hydrologic effects of frozen ground depend on the areal distribution of various frost types in a particular drainage basin. However, there is a further consideration, which many of the studies reviewed have not dealt with: the subsequent movements and phase changes of the infiltrating water. Where this has been investigated, the water has usually been observed to freeze in the soil (at least under conditions approximating natural rates of rainfall and snowmelt), eventually reducing the infiltration capacity and, of course, not immediately contributing to subsurface flow, available soil moisture or groundwater recharge. In a few cases, rain or infiltrating water has thawed the soil.

However, none of the above studies has critically examined the complicated heat and moisture relations of water in a frozen soil. Clearly, some kind of "ripening" process must go on, as in a snowpack. Water infiltrating into ground that is below the freezing point must freeze, with the liberation of latent heat. This continues until the soil has warmed to 0°C, and then thawing begins. The two papers reviewed subsequently in this section, both very recent contributions to the literature, elaborate further on these considerations.

Harlan (1971) has suggested a means of describing the process of infiltration into frozen ground mathematically. This involves the straightforward formulation of a heat flow equation for a differential volume of soil, including conduction and advection due to water flow and vapor flow. For the one-dimensional vertical case, this equation is:

$$\lambda(T, \theta) \frac{\partial^2 T}{\partial z^2} - c_0 \rho_0 \frac{\partial(v_z T)}{\partial z} - c_g \rho_g \frac{\partial(u_z T)}{\partial z} = c\rho \frac{\partial T}{\partial t} + L_0 \rho_s \frac{\partial \theta_s}{\partial t} + L \rho_v \frac{\partial \theta_v}{\partial t} \tag{6}$$

where

λ = thermal conductivity of the soil element

T = temperature of the soil element

θ = total (liquid plus ice) water content of the soil element

z = distance in the vertical direction

c_0 = specific heat of water (liquid)

ρ_0 = density of water (liquid)

v_z = vertical velocity of liquid water flow

c_g = specific heat of soil air

ρ_g = density of soil air

u_z = vertical velocity of soil air

c = bulk specific heat of soil element

ρ = bulk density of soil element

L_0 = latent heat of freezing of water

ρ_s = density of ice

θ_s = volumetric ice fraction

L = latent heat of vaporization

ρ_v = density of water vapor

θ_v = volumetric water vapor fraction.

Equation 6 is then used in conjunction with a flow equation for water (eq 2) and air (analogously to eq 2). This system of equations cannot be solved analytically, and Harlan (1971) suggests a finite difference or finite element approach to eq 2 and its analog for the gas phase. This provides an estimate of v_z and u_z over the region of interest, and these estimates can then be used as inputs to eq 6, which is then also solved numerically to estimate the temperature field at the predetermined points. An iterative computation algorithm must be used because the coefficients used to calculate the flow velocities are functions of the final temperature. When a phase change is involved, the heat liberated or consumed in each volume element of the soil is accounted for on the basis of the "apparent" specific heat, which must in turn account for changes in the volumetric ice, liquid and gas fractions with time and temperature. If the soil does not swell or shrink, the sum of the volumetric ice, liquid and gas fractions is constant and equal to the soil porosity.

Harlan (1971) concludes that our understanding of the processes of soil water freezing, heat flow and mass flow is sufficient to allow mathematical simulation of infiltration in frozen and partially frozen soils. However, he points out that a present limitation to accomplishing this is the lack of quantitative information about these processes to allow verification of a simulation model.

A different approach to the quantification of the process of infiltration into frozen ground, and one more closely related to the results of the several field and laboratory studies described above, was recently presented by Alexeev et al. (1972). Their approach is directed at the "ripening" process described earlier. First qualitatively, these writers consider the process of infiltration in frozen ground in four phases:

I. Phase of decreasing infiltration rate
II. Phase of zero infiltration rate
III. Phase of increasing infiltration rate
IV. Phase of steady infiltration

In Phase I, infiltrating water initially freezes, liberating latent heat. Since the specific heat of soil is low (0.5-0.7 cal/cm^3), the soil temperature rises quickly to 0°C, and subsequent water does not freeze. During this phase, the infiltration rate (assuming constant water supply) decreases due to: 1) decrease of piezometric head, 2) decrease of capillary head, and 3) decrease of pore cross sections due to formation of ice crystals.

If the temperature of the frozen ground at the beginning of Phase I is only slightly below freezing, the "cold content" of the soil is small, ice crystals only partly fill the pores, and complete stoppage of infiltration does not occur. Otherwise, all empty pores at a certain level become filled with ice and the pores above that level fill with water, at which time infiltration ceases (Phase II). Phase III cannot begin until the blocking ice layer melts, which usually does not occur until complete disappearance of the snow cover. Phase IV represents the steady-state infiltration rate characteristics of the unfrozen soil.

Alexeev et al. (1972) give a number of formulas for calculating parameters related to the above phases. At a given initial water content, there is a critical temperature below which the cold content of the soil is sufficient for ice to completely block all the free pores, giving rise to Phase II. For a given soil, the relation between the critical temperature and the initial water content at that temperature is given by

$$\theta_c = \frac{(L + c_0 T_0)n\rho_s + [c + \theta_u(T_c)(c_0 - c_s)]\rho T_c + L\rho\theta_u(T_c)}{\rho_0(2L + c_0 T_0) - c_s\rho_0 T_c - (\rho_0 - \rho_s)(L + c_0 T_c)} \qquad (7)$$

where

θ_c = critical initial volumetric water content (dimensionless)

L = latent heat of freezing (cal/g)

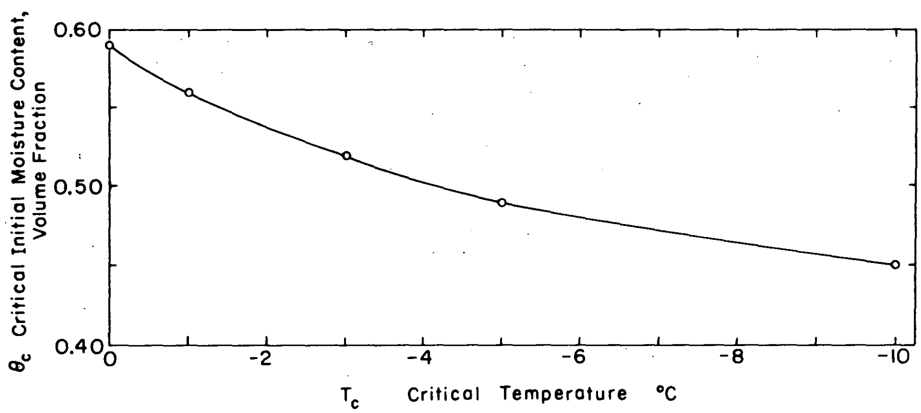

Figure 24. Relation between critical temperature and critical initial moisture content for a typical chernozem soil.

c_0 = specific heat of water (cal/g °C)

T_0 = initial temperature of infiltrating water (°C)

n = porosity (dimensionless)

ρ_s = density of ice (g/cm^3)

c = specific heat of soil (cal/g °C)

$\theta_u(T_c)$ = unfrozen volumetric water content at critical temperature T_c (dimensionless)

c_s = specific heat of ice (cal/g °C)

ρ = soil density (g/cm^3)

T_c = critical temperature (°C)

ρ_0 = density of water (g/cm^3).

Equation 7 can be used to construct tables or curves of θ_c as a function of T_c; one such curve is shown in Figure 24.

The blocking layer of ice is formed at the soil depth z_b, where the initial soil temperature T_i equals the critical soil temperature T_c (both T_i and T_c in general vary with depth, the latter because soil properties, particularly water content, vary with depth). The volume of infiltrating water that can be held in a unit volume of frozen soil (above the blocking layer) θ_v is found by

$$\theta_v = \frac{\rho_s n - \theta_i \rho_0 + \rho\theta_u(0)(1 - \rho_s/\rho_0)}{\rho_0} \tag{8}$$

where θ_i is the initial water content at that level and the other symbols are as in eq 7. To determine the total amount of water that can be held above the blocking layer V, eq 8 is integrated from the surface to z_b, the depth at which blocking occurs:

$$V = \int_0^{z_b} \theta_v dz. \tag{9}$$

Several examples of the variation of soil properties with depth are given, and the authors stress the importance of the depth profiles of initial temperature and moisture content, as opposed to bulk values, in determining the depth of the blocking layer and the amount of infiltrating water that can be held above that depth. The above approach was successfully used to reproduce the results of some of Tsykin's (1956) experiments.

Alexeev et al. (1972) also present a simplified dynamic model of infiltration into frozen ground based on the equations

$$q = n_f \frac{dy}{dt} \tag{10}$$

and

$$q = k(1 - \psi/y) \tag{11}$$

where

q = rate of infiltration (cm/sec)

n_f = free porosity

y = depth of wetting front (cm)

t = time (sec)

k = effective permeability of wetted soil

ψ = capillary head (cm).

This model assumes a wetted zone through which infiltrating water has penetrated and in which temperature is assumed to be 0°C, separated by a wetting front from a dry zone in which the temperature is equal to the initial soil temperature. Freezing occurs at the wetting front. Analytical solutions to the system of eq 10 and 11 can be found giving relations of q, y and V to time:

$$t = \frac{n_f \psi}{k}\left[\frac{k}{q-k} - \ln\left(\frac{q}{q-k}\right)\right] \tag{12}$$

$$t = \frac{n_f \psi}{k}\left[\frac{y}{\psi} - \ln\left(1 + \frac{y}{\psi}\right)\right] \tag{13}$$

$$t = \frac{n_f \psi}{k}\left[\frac{V}{\psi n_f} - \ln\left(1 + \frac{V}{\psi n_f}\right)\right]. \tag{14}$$

These equations have the approximations

$$q = k + \left(\frac{k\psi n_f}{2t}\right)^{1/2} \tag{15}$$

$$y = \frac{kt}{n_f} + \left(\frac{2k\psi t}{n_f}\right)^{1/2} \tag{16}$$

$$V = kt + (2k\psi n_f t)^{1/2}. \tag{17}$$

These equations hold if it is assumed that the initial temperature is constant over the depth of soil under consideration.

Alexeev et al. (1972) go on to point out that the decrease in permeability of frozen ground as the degree of ice saturation increases is explained by two factors: 1) the increase of degree of ice saturation decreases the pore space saturated with liquid water, and 2) the presence of more ice increases the specific surface area of the soil. Further, the ice formation at the wetting front depends on: 1) the loss of heat to the cold soil minerals and ice (specific heat effect), and 2) the heat loss due to the melting of ice at negative temperatures between the initial soil temperature and 0°C (latent heat effect). The former is by far the dominant process in sandy soils where there are few fine particles (clays) and therefore very little water subject to freezing point depression. In this case, the ice saturation occurring at the wetting front θ_{sw} is given by

$$\theta_{sw} = \frac{\theta_{si} - T_i(c_s\rho_s\theta_{si} + c_s\rho)}{\rho_s L} \tag{18}$$

where θ_{si} is initial ice saturation and other symbols are as in previous equations. Equation 18 was found to give "perfect results" for infiltration into sandy soils, for which it was found that soils in which T_i = −5° to −10°C have sufficient cold content to form only a small degree of ice saturation. This explains the high permeability of frozen sands noted by many earlier writers.

Where fine materials are present and bound water is subject to freezing point depression, another expression for θ_{sw} holds:

$$\theta_{sw} = \frac{-T_i[c\rho + c_s\rho_s\theta_{si} + c_0\rho_0\theta_u(T_i)] + L\rho_0[\theta_b - \theta_u(T_i)] + \rho_s L\theta_{si}}{\rho_s L} \tag{19}$$

where $\theta_u(T_i)$ is the unfrozen water content at the initial temperature T_i, θ_b is the maximum bound water content of the soil, and the other symbols are as previously defined. It is interesting that, in the case of infiltration into a frozen fine-grained soil, melting and freezing are going on simultaneously, with freezing in the larger, free-water pores and melting of the bound frozen water in very small pores.

In comparing the work of Harlan (1971) and Alexeev et al. (1972) with the observations of earlier workers, it is clear that the former, and particularly Alexeev et al. (1972), have significantly advanced our understanding of the process of infiltration into frozen ground. The basis now exists for the development of simulation models that could improve our understanding of hydrologic and ecologic systems where freezing occurs and perhaps provide improved operational models for flood and water supply predictions.

Overland flow and frozen ground. Since, in the Hortonian concept, overland flow is the complement of infiltration, the previous section has treated that aspect of the relations between frozen ground and the generation of overland flow. Other aspects of overland flow will be discussed later under the topic of "erosion."

It remains to discuss the influence of frozen ground on overland flow as generated by the partial-area contribution mechanism. As noted earlier, this mechanism involves infiltration of rain or snowmelt into areas adjacent to upland streams in which the water table is close to the surface. If the infiltration capacity f is less than the precipitation or melt rate p, then the rate at which the water table rises r in these areas is

$$r = \frac{f}{n - \theta} \tag{20}$$

where n is the average porosity above the water table and θ is the volumetric water content in the unsaturated zone. If f is greater than p,

$$r = \frac{p}{n - \theta} . \tag{21}$$

Thus, at water contents close to saturation (θ almost as large as n), the rate of water table rise could be rapid. When the water table reaches the ground surface, overland flow begins. Equations 20 and 21 assume that no groundwater flow from upslope contributes to the water table rise. Freeze (1972) indicated that such contributions are usually significant.

As noted above, this mechanism seems to be the principal means of storm runoff generation in humid areas (Freeze 1972). However, no studies of the effects of seasonally frozen ground on this mechanism were found. An impermeable surface layer due to the presence of ice (concrete frost) would, of course, prevent the process from operating, and overland flow would occur immediately. If θ in eq 20 and 21 is considered to include liquid and solid water, effects of ice are included in that term. From the work of Alexeev et al. (1972), it would seem that, in some cases, both the Hortonian and water table rise mechanisms might operate during a single storm where near-surface ground frost is present. Initially, if the near-surface soils are close to saturation and frozen, infiltration might be less than rainfall rate, and Hortonian overland flow could occur. As infiltration progressed, however, there could be considerable melting, and infiltration could increase and reach the water table more rapidly. Eventually, melting would be complete, and the water table could reach the surface to produce overland flow by the second mechanism. If this sequence of events were to happen in the basin of a small stream, its hydrograph might show a rapid initial rise, a decrease as melting progressed and infiltration increased, and then a second rapid rise once the water table reached the surface.

Dingman (1973a) has discussed the effect of shallow permafrost on the partial-area mechanism in a small (0.7-mile^2) watershed near Fairbanks, Alaska. In this watershed, permafrost was present at depths of 1 to 3 ft beneath the stream and beneath a wide, gently sloping area adjacent to the stream. Hydrograph analysis indicated a runoff source that responded quickly to rainfall events under all conditions except when the watershed was extremely dry. Overland flow from the valley bottom area by the partial-area contribution mechanism was inferred to be operating, and the principal effect of permafrost here was in supporting the high water table beneath the source area.

Seasonal thawing

It is the writer's impression that many hydrologists believe that the thawing of the active layer in permafrost regions is a significant source of streamflow. However, only three published reports were found which explicitly discuss this possibility. Dingman (1966) arbitrarily separated a "base flow" hydrograph from the total runoff measured from a 0.7-mile^2 watershed near Fairbanks, Alaska, during one summer (1964). This "base flow" was very low (0.14 ft^3/sec mile^2) and constant through June and most of July, rose to a peak of 0.43 ft^3/sec mile^2 in late August, and then receded gradually to zero at freeze-up on 25 October. Dingman (1966, p. 754) suggested that "a hydrograph such as this may be in part due to the thawing of the active layer, but this has not been established."

A more definite statement as to the importance of runoff contributions from the thawing active layer was made by Sommer and Spence (1968). They noted that the month of maximum flow for the Yellowknife and Snare Rivers (N.W.T., Canada), whose basins contain widespread permafrost, coincided with the month of maximum temperature, rather than snowmelt or maximum precipitation. The Hay River, which is in the same general area but drains an area of only scattered permafrost, did not show this correspondence. Their explanation was as follows (Sommer and Spence 1968, p. 63-64):

> During the spring snowmelt period the temperatures in the active layer of the ground remain below freezing. This layer is also relatively dry and receptive to moisture as a result of water deficiency in the late summer and fall of the preceding year. Therefore, the snowmelt waters are able to pass into the ground where they become refrozen in the active layer. This refrozen moisture is then slowly released in summer as the active layer thaws. Thus, the maximum streamflow would be expected during the summer corresponding to the warmest period.

This explanation must be regarded as possible, but conjectural. First, it is difficult to understand why the process described would not operate as well in nonpermafrost areas subjected to seasonal freezing as in permafrost areas. In fact, one might expect it to be more important where deep seasonal freezing takes place, as there would be a deeper layer of soil to absorb and release meltwater. Further, Dingman (1971) made an analysis similar to that of Sommer and Spence (1968) for three Alaskan rivers with permafrost basins, but found no correspondence between month of maximum temperature and month of maximum runoff.

In analyzing longer-term data for the same Alaskan watershed he discussed previously (Dingman 1966), Dingman (1971) concluded that water released by the thawing of the active layer did not contribute significantly to runoff. About 70% of the watershed is underlain by permafrost, yet during extended rainless periods in two summers flows decreased to zero or near-zero while relatively rapid thawing was taking place. It now appears that the base-flow hydrograph described in Dingman (1966) did not represent runoff from a separate source, but a coincidental result of the method of hydrograph separation applied to the streamflow pattern of that particular year.

The small (0.62-mile^2) watershed at Barrow, Alaska, studied by Brown et al. (1968) is completely underlain by permafrost, yet hydrographs for four summers show zero or near-zero flow during periods when rapid active-layer thawing is occurring. This is particularly clear in 1964, when flow ceased entirely from about 10 July to freezeup in early September, indicating no significant runoff contributions due to thawing ground.

The few other detailed studies of permafrost hydrology that exist (e.g. Likes 1966, McCann and Cogley 1972) do not suggest the operation of this mechanism of runoff production. While it would be premature to state that it does not occur anywhere, recent work by Freeze (1972) suggests that it is rare. That writer developed a simulation program combining saturated-unsaturated subsurface flow and channel flow, based on fundamental physical principles. The results indicate that except in watersheds with convex hillslopes, a deeply incised channel, and highly permeable soils, subsurface flow is not an important contribution to upland streams. Where shallow permafrost exists, then, one would not expect subsurface flow generated by any mechanism to be an important source of runoff.

Groundwater

The topic of groundwater in permafrost areas has received a great deal of attention, particularly in the Soviet literature. Williams (1965) has compiled an annotated bibliography of the world literature in this field through 1960 that contains 862 entries. Another annotated bibliography, that of Dingman (1973b), also contains a number of entries for articles concerned with the occurrence of groundwater in permafrost areas. Because of the existence of these and of several general articles on this topic (for example, Brandon 1963, Ponomarev and Tolstikhin 1959, Suslov 1961, Williams 1970, Owen 1967), it will not be treated in detail herein.

Hydrogeologically, permafrost behaves as an aquiclude, i.e. as an impermeable barrier to groundwater flow and to recharge. The contact between permafrost and saturated unfrozen material must be a mobile one, constantly adjusting to vertical heat flow changes, changes in the rate of flow past the boundary, and chemical and thermal conditions of the groundwater.

Unfrozen groundwater can exist above the permafrost (suprapermafrost water), within the general permafrost body in unfrozen zones called *taliks* (intrapermafrost water), and below the permafrost (subpermafrost water). All of this water can contribute to stream flow and runoff. Figure 12 shows that, even in the continuous permafrost zone, permafrost tends to be absent beneath large, deep lakes and beneath the ocean. This is also true under larger rivers, so they may receive base flow from groundwater. A number of Soviet articles estimate the degree of "underground feeding" of northern rivers (L'vovich 1961, 1962, Kupriyanova 1963); where permafrost is widespread, this makes up 3% to 10% of total runoff, as compared to more than 30% groundwater contribution where permfrost is limited or absent.

Anderson (1970) studied the hydrology of the Tanana River basin, an area of 44,500 miles^2 in the discontinuous permafrost zone of Alaska, and estimated that 10% of the runoff leaves the basin as groundwater. However, in portions of the basin this may be as high as 50%, and Dingman et al. (1971) estimated that 32% of the runoff from the Delta River, a tributary of the Tanana, occurred as groundwater. These Alaskan figures are not equivalent to the data from the U.S.S.R. cited in the preceding paragraph; the Alaskan figures are for water leaving the basin as groundwater, not for the proportion of streamflow coming from the groundwater. However, the data do indicate extensive unfrozen zones and much groundwater movement in the discontinuous permafrost zone. Unfortunately, no data on groundwater contributions to runoff were found outside of the U.S.S.R.

Freezing ground in and near stream valleys can, by confining or blocking the groundwater draining toward the stream, create *icings*, or *naleds*. This process has been described by Waller (1961), and the general topic of icings has been reviewed by Carey (1970, 1973). This process generally occurs only in permafrost areas, where freezing from the surface downward confines suprapermafrost water. When the confining pressure gets high enough, the water forces its way through the overlying ice to break through at the surface and freeze. When this occurs repeatedly, great thicknesses of ice may build in the stream, causing flooding and other damage.

Outside the permafrost zone, seasonally frozen ground has a direct influence on groundwater to the extent that it controls infiltration. In addition, some more indirect influences have been suggested. Dreibelbis (1949), studying intensively instrumented watersheds and lysimeters in Ohio, indicated that freezing near the surface keeps water from percolating to the water table, and makes it more available for evaporation and transpiration. Thus frost could reduce groundwater recharge by reducing percolation of water that has previously infiltrated as well as by limiting infiltration. Another mechanism producing the same effect, but operating over the entire soil column above the water table rather than just in the freezing zone, was also postulated by Dreibelbis (1949). He suggested that the presence of a surface layer of impermeable frost could prevent percolation by forming an airtight cap. This essentially suspends the water column by preventing gravity drainage, just as one can suspend a column of water in a straw by sealing the upper end with his finger. This phenomenon might occur where concrete frost was quite widespread, but it has not been further discussed in the literature.

Schneider (1958) observed a close relation between air temperature and groundwater levels during the winter and spring in Minnesota. In two wells, the water table declined during the winter as long as air temperatures were below freezing, but within a few days of the onset of positive air temperatures, a rise in groundwater levels was observed. If freezing temperatures recurred and persisted, the decline would begin again. The decline was attributed to the migration of water to

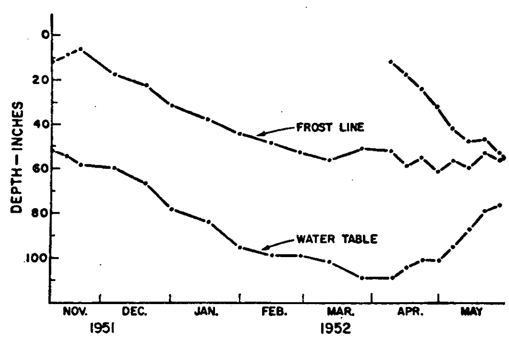

Figure 25. Depth of frost and water table in North Dakota. (Willis et al. 1964.) (Copyright 1964, The Williams and Wilkins Co., Baltimove, Md.)

the freezing front, and the phenomenon was observed when water tables were as much as 9 ft below the surface. The rise in water table was apparently due to release of water by melting at the bottom of the frozen layer and subsequent downward percolation, as the surface layer of concrete frost prevented infiltration from melting snow.

Willis et al. (1964) reported essentially the same phenomena from a study in North Dakota. Beginning in November, the water table was at a depth of about 4 ft and the frost line at 1 ft. Both declined at the same rate throughout the winter (see Fig. 25), and the moisture content in the surface soils increased concomitantly. When thaw began in early April, the water table began to rise, even though thawing of the soil occurred almost entirely from the top. However, moisture contents in the frozen zone, which were near saturation, decreased, and the frost apparently became somewhat permeable, allowing percolation. As in Schneider's (1958) study, infiltration was ruled out as a source of recharge during the thaw.

In southwestern Wisconsin, Sartz (1967) observed a marked increase in the flow of a spring and a rise in the groundwater table during winter thaws. However, the thickness of the frozen soil layer (2.0 to 2.6 ft) did not change, suggesting that the mechanism described by Schneider (1958) was not operating. Even though about 90% of the area above the spring was frozen, Sartz (1967) suggested that infiltrating snowmelt during the thaws may have been responsible for the groundwater fluctuations.

The studies of Schneider (1958), Willis et al. (1963) and Sartz (1967) indicate that seasonally freezing ground may have rather profund influences on the groundwater regime, through the operation of several mechanisms. If these phenomena are widespread, marked effects on stream baseflow could also occur. Unfortunately, observations of these processes are too scarce to allow a full assessment of their importance.

Runoff and streamflow

A number of possible direct effects of frozen ground on runoff and streamflow are readily apparent from the above reviews of the relations between frozen ground, infiltration and groundwater, and will not be restated here. Instead, the rather meager literature on general geographic relations between frozen ground and runoff will be reviewed.

The overall effects of frozen ground on streamflow and runoff have not been extensively studied, in part because of the complex interrelationships that exist between frozen ground and other climatic conditions that also influence runoff. L'vovich (1961) stated that data were insufficient for comparing runoff in various regions of the U.S.S.R., but gave some values suggesting that permafrost increases runoff (Table V). Data from Kupriyanova (1963) (Table VI) suggest the same trend.

This is generally what one would expect if permafrost were acting largely as a barrier to prevent infiltration and groundwater recharge. One would also expect streams in permafrost areas to be "flashier," that is to have steeper flow-duration curves, with higher floods and lower low-flows (on a unit area basis) than nonpermafrost streams. However, these effects can be explained largely on the basis of climate alone: in progressing from the taiga without permafrost to the taiga with permafrost to the tundra, evapotranspiration decreases (Hare 1950), which would tend to increase runoff. The flashiness could also be largely due to the shorter summer period available for runoff as one moves north.

Table V. Runoff from various geographic zones, northern U.S.S.R. (L'vovich 1961).

Zone	*River basin*	*Runoff* (mm)	*Runoff* (% of precip)	*Runoff* (% surface runoff)
Tundra	Much'ya	340	76	97
	Amguema	255	73	97
Taiga	Pinega	344	69	73
	Vym	314	56	76
Taiga with permafrost	Vilyuy	210	58	92
	Olenek	195	49	95

Table VI. Runoff from various geographic zones, northern U.S.S.R. (Kupriyanova 1963).

Zone	*Total runoff* (mm)	*Total runoff* (% of precip)	*Surface runoff* (mm)	*Surface runoff* (% of total runoff)
Tundra	250	70	225	90
Northern taiga	260	54	180	69
Typical taiga	230	46	130	57

One of the independent variables used by Mustonen (1967) in his multiple regression analyses of annual runoff in Finland was seasonal frost depth. Specifically, the frost depth for each of 33 experimental watersheds was taken as the average depth measured on 31 March. All watersheds were outside the permafrost zone. Frost depth had an inverse relation to average runoff, and explained a significant portion of the runoff variance.

> In Finnish conditions it is generally known that annual runoff is small in years with thick frost. Of course, this is mostly owing to the circumstance that frost depth is thick because of low winter precipitation, thus indicating that thin snow cover causes small runoff. But frost depth apparently has a primary effect of its own, because it is significant even after removal of the effect of winter precipitation from runoff in orthogonal analysis. (Mustonen 1967, p. 128.)

Frost depth was, in fact, highly inversely correlated ($r = -0.60$) with winter precipitation, as well as annual precipitation ($r = -0.56$).

Mustonen (1967) does not discuss further what the independent inverse effect of frost depth on runoff might be. It might involve the migration of water to the upper soil layers and immobilization there during the winter, making more water available for evapotranspiration after thawing, as suggested by Anderson (1946) (see p. 45) and Dreibelbis (1949).

Evapotranspiration

Studies by Anderson (1946) on the west slopes of the Sierra Nevada in California have indicated the degree to which evapotranspiration can be increased by diurnal soil freezing. In these studies, soil moisture profiles were monitored on two plots, one bare of vegetation and subject to almost nightly soil freezing and the other covered with brush which prevented freezing. Surface freezing occurred on the bare plot to depths of 0.1 to 0.7 in. per night, and drew water from depths as great as 36 in. During the day, complete thawing occurred. Thus, the surface soil in this plot remained very wet during extended rainless periods, and evaporation during periods when diurnal freezing occurred was three times as great as that from the bare plot when there was no freezing, nearly four times greater than from the non-freezing plot, and about 12 times greater than free water evaporation during the same periods. Clearly, under the right conditions, soil freezing could have an important effect on soil moisture distribution and evapotranspiration.

Bethlahmy (1952) pointed out that frozen soil is, from a soil moisture point of view, a dry soil. Thus, when freezing occurs in the root zones of plants subject to a transpiration stress, wilting may occur. The effect of ground freezing in this case is to reduce evapotranspiration.

Another effect of soil freezing that may have a minor effect on soil evaporation under certain conditions was discussed by Goll (1964). He found that wet soils, on freezing, showed an increase in albedo: a bright yellow soil from 23% albedo when thawed to 26.4% when frozen, and a brown forest soil from 9% to 12%. This effect occurred only on bare soils, and affected only the evaporation from the frozen soil, which is undoubtedly quite small.

Erosion

The susceptibility of frozen ground to erosion by water has been widely observed in permafrost regions. Whenever flowing water is in contact with ice-rich sediments, rapid and extensive degradation occurs. The resulting landforms, similar to karst topography, have been given the name *thermokarst*. These features are most common where ice wedges or other thick ice bodies are present.

Both Straub (1950) and Walker and Arnborg (1963) describe the undercutting that occurs during flooding at the water line of a river flowing through frozen materials. Following the Russian terminology, the result of this action is called a "thermoerosional niche." The formation of niches, and their subsequent collapse, along with other details of erosion of frozen ground, were described by Walker and Arnborg (1963). Other observations of riverbank erosion in frozen soils are found in Eardley (1938), Williams (1952), Waller (1957) and Stefansson (1910).

It could be expected that the operation of thermal-erosional processes might give rise to hydraulic geometry relations quite different from those described for temperate regions where alluvial processes operate alone. However, no observations reflecting this were found.

Several studies have been directed to determining the effect of seasonal frost on runoff by overland flow. Atkinson and Bay (1940), examining freezing and runoff on agricultural soils in Wisconsin, noted that concretely frozen soil is very susceptible to erosion in the early stages of thawing. At this stage, supersaturated soil is present at the surface, and any rain that falls cannot infiltrate, so a slurry is produced that can carry away a high proportion of the thawed topsoil. They even observed that the supersaturated thawed layer can flow by itself under the right conditions, without rain.

Similar observations were made by Tigerman and Rosa (1949) in the mountains of Utah. The surface soil had been frozen to a 6-in. depth, and thawing of the top few inches produced a mud slurry that could flow downhill if slopes were greater than 50%. These miniature mudflows were most common in clay soils, but were observed in sandy and even gravelly material as well.

Bay et al. (1952) also found high runoff and high soil losses from agricultural lands in Wisconsin when rainfall and snowmelt occurred at the same time on partly frozen soil. The high water content, often greater than saturation, due to upward moisture migration during freezing, again was apparently responsible for the increased susceptibility to erosion.

The effects of stalactite frost in detaching soil particles so that they are readily eroded by overland flow, and in moving particles downslope by a step-like action, were noted by Bullard (1954). He also stated that frost action breaks down the structure of the surface layer, making it readily erodable.

Schumm and Lusby (1963) studied the erosion of lithosols on shales in western Colorado. Repeated episodes of freezing during the winter caused a cracking and loosening of the soil surface that greatly increased the infiltration capacity and reduced runoff and sheet erosion. However, soil creep, due to frost action, was the dominant erosive process during the winter.

Haupt's (1967) study of the hydrologic effects of frozen ground on the east slope of the Sierra Nevada in California demonstrated a "strong antierosive influence" (Haupt 1967, p. 160) of soil frost. In this area, two types of soil frost formed: stalactite frost, which absorbed snowmelt and rainfall and prevented overland flow and sheet erosion, and concrete frost, which protected the soil particles from detachment by raindrops.

The studies reviewed above concerning the effects of seasonal frost on soil erodability give somewhat conflicting results. Several indicate severe erosion by overland flow and miniature mudflows when concrete frost thaws and is subjected to rains, while Haupt's (1967) work indicates that concrete frost prevents sheet erosion. Part of the reason for the difference may lie in the fact that Haupt's soils were in cobbly sandy loam, while the others were in finer-grained soils. In any case, it would seem that soil freezing and thawing can have a marked effect on erosion, at least for bare and sparsely vegetated soils. The nature of the effect is largely determined by soil type and frost type.

SUMMARY

Frozen ground is defined as earth material at a temperature below 0°C. Perennially frozen ground underlies about 21% of the land area of the Northern Hemisphere, and seasonal ground freezing is a common occurrence over about 48% of the Northern Hemisphere (Bates and Bilello 1966).

The hydrologic effects of permafrost are profound. However, these effects can be considered static unless one is concerned with long-term climatic changes. Permafrost acts as an aquiclude, restricting the movement and recharge of groundwater. However, the effects of this on streamflow are not necessarily the expected straightforward ones of increased runoff and flashier streamflow. Several studies have indicated extremely long hydrograph recessions for streams draining permafrost areas, and some large basins have a high proportion of their runoff occurring as groundwater outflow. There are a few indications that runoff percentages increase northward in the U.S.S.R., but the role of permafrost in controlling this has not been established. Although runoff percentages estimated for small central and northern Alaskan watersheds are moderately high (around 50%), most of this is snowmelt, and the thaw season runoff percentages are much lower (as low as 5% at Barrow). Much more remains to be learned about the processes of runoff generation in permafrost areas and, indeed, in temperate regions as well. The susceptibility of frozen ground to erosion by running water has a profound effect on the geomorphology of permafrost areas.

Seasonal freezing and thawing also have important hydrologic effects, and these are far from static. In addition to affecting infiltration in complicated ways, sometimes increasing it and sometimes decreasing it, and temporarily immobilizing soil moisture and groundwater, soil freezing also involves the transport and redistribution of moisture. A few studies have suggested quite significant effects of this on the groundwater regime and on evapotranspiration. One regional study showed an inverse relation between frost depth and runoff. Ground freezing also affects the erosion process, both through its control of infiltration and overland flow and its physical effects on the soil surface.

The environmental variables affecting the type and distribution of frozen ground are well known qualitatively, but work is only beginning on the construction of quantitative models that will allow simulation and prediction of the presence and depth of permafrost and the seasonal freezing and thawing of soils. Such models are required to evaluate man-made impacts on the natural environment.

The freezing of earth material is a complicated physical process that includes heat and mass transfer at the earth's surface and below, and involves three phases of water and phase changes among them, plus air and a heterogeneous mixture of mineral and organic matter with an irregular geometry. In spite of this complexity, the process appears to be quite well understood and its various components can be expressed mathematically. The foundations needed to combine the various processes and couple them to the surface heat and moisture balance and the geothermal heat flow in a simulation model are just beginning to be formulated, and continued work in this area is a major research need.

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