A MATHEMATICAL MODEL OF SOIL SURFACE LAYERS FOR USE IN PREDICTING SIGNIFICANT CHANGES IN INFILTRATION CAPACITY DURING PERIODS OF FREEZING WEATHER

by

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CHAPTER ONE. THE RUNOFF PROBLEM

Flood forecasting in major river basins of the United States is one responsibility of the National Weather Service (NWS). Various forecasting systems have been developed and are being used in various parts of the country. Each system requires some scheme for partitioning that portion of precipitation going into the soil from that going directly to runoff. The accuracy of predicted amounts of soil moisture recharge, volume of runoff, and height of flood crests is highly dependent upon this scheme.

Comparisons which have shown conceptual models to have improved accuracy and more consistency than the empirical models previously used led to development of a forecast system by NWS which combines necessary computer programs with hydrologic models to organize data, calibrate basins, and operationally forecast discharge (NOAA, 1972). Runoff estimations from this system (National Weather Service River Forecast System (NWSRFS)) have in general been very good; however, in some cases, some of the component models within the system have been inadequate.

In certain portions of the Midwest region, as much as 50 percent of the annual runoff may occur in a 2-month period during late winter and early spring. This runoff is associated with melting snow and spring rains running off nearly impervious, frozen, or saturated soils. Calibration of an upper Midwest watershed has revealed that, due to the variation in soil moisture and soil frost conditions from year to year, significant errors in estimation occurred (E. A. Anderson, personal communication, 1973).

Purpose of This Study

This study was undertaken to obtain a better understanding of the effects of frost and high soil moisture content on runoff during periods of freezing weather. Work of other investigators, presented in more detail later, has shown that:

1. differences in soil frost formation can result in differences in soil infiltration capacities that sifnificantly affect the accuracy of runoff forecasts; and

2. there is a runoff forecasting problem associated with changes in soil moisture distribution due to thermal gradients and with increased soil moisture retention occuring with low soil temperatures.

Specific objectives of this study were to:

1. incorporate available theories describing these phenomena into a frost formation model that can identify and warn of high runoff potentials due to restricted infiltration; and

4. test the frost formation model against actual field data to determine its accuracy in predicting high runoff conditions.

Constraints on Model Building

An ideal frost formation model would treat a large watershed and consider diverse soils, topography, vegetative cover, and land use pattern found within it. Such a model would be unnecessarily complex, however, and obscure the effects of individual variables on frost formation and penetration. If, on the other hand, a large watershed were divided into subunits having definable slope, aspect, cover, and limited variety of land uses, a simpler model could be developed to handle each of the subunits, This approach was followed in this study.

To be suitable for field application, input data required by the model <u>must be limited</u> to those already available or readily obtained. Most data currently available were derived from point measurements and are published in terms of daily totals, maximum and minimum values, or average values (e.g., solar radiation, air temperature, or wind velocity, respectively). This constraint makes it necessary to utilize both space and time averages in computations; and when conceptual equations describing essential relationships require unavailable information, empirical approximations will be substituted for the missing data.

Definition of Terms

Many of the terms that will be used repeatedly in this paper have been defined differently by different authors. Brief discussions of the more important terms are included below to clarify how they will be used herein.

Terms Related to Soil Moisture

Surface runoff is that portion of rain or melt water that does not infiltrate the soil but moves to rivers, streams, or lakes across the surface of the ground.

Field capacity has been variously defined as:

> "Soon after a rain when all the gravity water has been drained down to the water table, a certain amount of water is retained on the surfaces of the soil grains by molecular attractions . . . . The maximum depth of this water (if it were spread over the basin) that any soil can retain indefinitely against the action of gravity is called its field capacity." (Wisler and Brater, 1954)

> ". . . amount of water held in the soil after excess gravity water has drained away and the rate of downward movement materially decreased." (Linsley et al., 1949)

Bayer (1956) discusses findings which indicate that, in soils where drainage has become negligible, soil moisture tensions range from as much as -0.6 bar to -0.005 bar depending on the soil. Some point of reference is needed, however, to approximate the soil moisture range within a layer to some quality of the soil which can be measured and classified. For this reason, field capacity is taken as that moisture content when the soil moisture tension is one-third bar (Buckman and Brady, 1969). It is recognized that this concept must be used with caution. Field capacity has been shown to vary as a function of soil temperatures, viscosity, drying rate, and other factors (Stanhill, 1973).

Gravity water, sometimes called free water, is that water in the soil that is in excess of field capacity.

Wilting point is the moisture content at which permanent wilting occurs (Linsley et al., 1949). It is sometimes called the wilting coefficient. It has been found for most plants to occur at a tension of less than 15 bars (Baver et al., 1972). This is a minimum value in most areas. Under hot dry conditions, plant deaths would occur at higher percentages of soil moisture. Beyond this point, most plants cannot extract significant moisture from the soil.

Percolation is the passage of water within the soil.

Infiltration rate is the rate of passage of water into the soil in units of depth/time.

Infiltration capacity is the maximum rate of infiltration into the soil for a given condition.

Terms Related to Soil Frost

Several classes of soil frost have been recognized. Post and Dreibelbis (1942) described three classes based on their work at Coshocton, Ohio:

Concrete frost is composed of many ice lenses and small crystals and is very dense. Generally, it occurs in soil that has been "puddled" or settled by being saturated with water, soils that have been previously frozen and thawed, or in bare soils with little vegetal cover. It is found in most soils frozen more than 3 inches deep. Stoeckell and Weitzman (1960) defined a modification of this type frost. They said that when sand is concretely frozen, air can be readily blown through it, and, therefore, they called it "porous concrete frost." They noted that infiltration was still greatly reduced in the frozen sand.

Stalactite frost consists of many little icicles or needles of ice going between the ground and a layer of heaved soil. It is sometimes called needle ice. It occurs when nearly saturated soil is frozen rapidly by a sudden temperature drop. It also forms from soil which has thawed and is suddenly refrozen.

Honeycomb frost is loose and porous and easily broken into pieces. Generally, it is found in soils high in organic content which are highly aggregated.

Hale (1951) identified a fourth class which he called granular frost. It consisted of scattered granules of ice binding together litter, decomposed litter, and mineral soil. Granular frost may be an extension of honeycomb frost that occurs from the litter-soil interface throughout the litter.

CHAPTER TWO. RUNOFF ACROSS FROZEN AND SATURATED SOILS

Analysis of the March 1936 flooding in southern New York State showed that 100 percent of the rain that fell on concretely frozen bare fields became surface runoff but that little or no surface runoff came from nearby unfrozen forest covered soils. Data for these small watersheds revealed a correlation between extent of frost cover and the amount of surface runoff. Where soil frost was present across 25 percent of the watershed, runoff accounted for 12 percent of the total water equivalent in the rainfall and snowmelt. With soil frost present in 63 and 93 percent of the watershed, runoff increased to 41 and 53 percent of the total water equivalent, respectively (Storey, 1955).

Harrold and Roberts (1945) ran two instrumented watersheds for six winters. They show distinct differences in runoff between frozen and unfrozen ground. The amount of runoff varied from none for spring rains on unfrozen ground to 100 percent when rain fell on frozen ground.

Runoff with Different Classes of Soil Frost

The results just cited show that the presence or absence of frost does affect surface runoff. In addition, Shipak (1969) found that the relationship between runoff and soil frost varied inversely with the depth of frost penetration. Post and Dreibelbis (1942) also mentioned the effect of frost depth, and stated that percolation was " . . . reduced materially, or ceased, when frost depths were 3 inches or greater . . . ." Extension of frost front below the 3 inch depth may have been accompanied by the development of concrete frost.

In 1958, Trimble et al. found that the presence of granular frost increased infiltration rates significantly, as compared with the non-frozen condition, but that the same soils had a zero infiltration rate when frozen concretely. Stoeckeller and Weitzman (1960) reported similar results from similar tests.

Richardson et al. (1969) compared two heavy winter storms in Utah in 1962 and 1963. A rain on snow event in 1962 caused heavy flooding to a wide area of northern Utah and eastern Nevada, while the storm of 1963 caused only localized flooding at Heber, Utah. The 1962 storm dropped 1.15 inches during a 4-day storm. There were several inches of snow on the ground with the ground concretely frozen in many places to 3 feet. This caused severe flooding. In 1963 at the beginning of a storm, the ground surface was bare, not as cold as the previous year, but was frozen with a porous type of frost which was measured down to 26 inches. It was noted that "The existence of a porous frost to such depth is likely accounted for by the very dry fall and rapid rate of freeze." With these conditions, 4.56 inches of rain fell in a 3-day storm, but little flooding resulted.

Runoff when Soils are Unfrozen but Excessively Wet

Several investigators have described runoff events that occurred when soils were unfrozen, but excessively wet. Working in the Rock River Basin of southwestern Minnesota, Peck (1973) found that "average soil moisture for 18 stations was 44 percent, or about 12 percent greater than their average field capacity as reported in a Department of Agriculture report." He also noted that when soil samples were first removed from the ground they were soft and pliable like modeling clay but became very "soupy" when warmed up in the car.

Molchanov (1960) indicated that in Russia "Ground water lying at a depth of less than 1 meter will often rise toward the surface, in spring, at the time of snow thawing; then the water will run off the soil's surface."

Peck (1973) notes that relatively large bodies of water have been seen to form in agricultural fields where only a small average water equivalent had existed the day before in the immediate drainage area. The circumstances indicate that the source of moisture must have been soil moisture from the ground below rather than new precipitation.

The existence of any of these conditions during a significant storm could result in significant errors in forecasting flood crests and total volume of runoff with currently used hydrologic forecast models.

Theories of Soil Moisture Response to Cold Temperatures

The infiltration rate of a soil depends on the soil moisture present and how it is distributed. Soil moisture distributions during periods of cold temperatures can be significantly different than during the normal growing season. There are several processes involved and since those affecting unfrozen soils do not involve a change in state, they will be discussed first.

Soil Moisture in Cooled but Unfrozen Soils

The increase of soil moisture above field capacity noted by Peck (1973) raises several questions:

1. Can soil with temperatures which are near the freezing point hold more soil moisture than warmer soils?

2. If it can and there is a temperature gradient with the coolest temperature near the surface, is a significant transfer of soil moisture likely to take place?

3. If rain falls on the surface, how is the percolating water affected by soil temperature and the soil temperature gradient?

Hydraulic Conductivity, Viscosity, and Surface Tension

The answer to the first of the questions arising from Peck's work is suggested by Klock (1972). He noticed an effect of temperature on hydraulic conductivity and soil water retention. Hydraulic conductivity is defined by the following equation:

$$K = K'\rho g/\eta \qquad (1)$$

where:

K is the hydraulic conductivity in cm/sec;

K' is the intrinsic permeability of the soil;

ρ is the soil density in gm/cm^3;

g is the acceleration of gravity; and

η is the viscosity in dynes sec/cm^2.

The volume of water transferred through a soil is related to equation (1) by

$$V = Ki \qquad (2)$$

where:

V is the volume flux of water per unit cross sectional area; and

i is the hydraulic gradient.

Viscosity of fresh water is 1.798 dyne sec/cm^2 at 0°C, but drops to 0.8904 dyne sec/cm^2 at 25°C. If all other factors remain constant, the ratio of conductivities would be

$$\frac{K(0°C)}{K(25°C)} = \frac{K'\rho g/1.798}{K'\rho g/0.8904} = 0.4952 \quad .$$

Jensen et al. (1970) cited data from an unpublished PhD thesis by R. O. Meeuwig that the temperature dependence of soil water viscosity was two or three times that of free water. This was determined by tests on three different soils. If the effect were just twice as much as for free water then the ratio, whether squared or divided by two (depending on the meaning of two times in this effect), would result in a conductivity of about a fourth the warm season conductivity.

If a certain period is required for soil to drain to field capacity, and if hydraulic conductivity is significantly reduced, a longer period will be required for soil to reach this level where drainage due to gravity ceases. During a season with fairly frequent rainfall and little evapotransportation, soil may remain above field capacity for some time.

A second effect that Klock investigated was the temperature dependence of water surface tension. As water percolates into the soil, it goes through many small channels. To gain insight into this effect, consider a soil layer made up of many capillaries. The water pulled into the capillary fringe above the water table is proportional to surface tension. The height of water in the i^{th} capillary tube is given by this equation.

$$h_i = \frac{2\sigma \text{Cos}\theta}{r_i \rho g} \tag{3}$$

where:

σ is the surface tension of the liquid-vapor interface;

θ is the wetting angle of the water on the soil; and

r_i is the effective radius of the i^{th} capillary. The other symbols are as defined earlier. The volume of water held in the capillary fringe is

$$V = \sum_{i=1}^{n} h_i A_i = \sum_{i=1}^{n} \frac{2\pi\sigma r_i^2 \text{Cos}\theta}{r_i \rho g} = \frac{2\pi\sigma \text{Cos}\theta}{\rho g} \sum_{i=1}^{n} r_i \tag{4}$$

Klock noted that the surface tension changes from 71.97 dyne sec/cm^2 at 25°C to 75.6 dyne sec/cm^2 at 0°C. Again, if all factors but the surface tension are independent of temperature, then

$$\frac{V(0°C)}{V(25°C)} = \frac{\sigma(0°C)\frac{2\pi Cos\theta}{\rho g}\sum_{i=1}^{n}}{\sigma(25°C)\frac{2\pi Cos}{\rho g}\sum_{i=1}^{n} r_i} = \frac{75.60}{71.97} = 1.05$$

If the capillaries are now considered to be randomly oriented, there is still a certain amount of water which must be stored in them before water will pass through them under the influence of gravity. Neglecting the weight of water, it appears that water retained in soil capillaries would be the same as that retained in vertical capillaries (Kirkham, 1964).

With these two effects in mind, Klock ran a test using a simulated soil made up of glass beads. He found that the hydraulic conductivity at 25°C is about twice that at 0.3°C. A subsequent test on a prepared soil gave similar results. When this same soil sample was saturated and then drained at 0.3°C, he collected water in the amount of 15.5 percent of the saturated weight. When the soil was heated to 25°C, more water in the amount of 1.7 percent of the saturated weight drained out, indicating that up to 12 percent more water could come from soils which were warmed after having been drained to apparent field capacity when the soil was cold. Findings similar to those of Klock were reported by Jensen et al. (1960). They stated that the change in conductivity and retention could easily be accounted for by the temperature effects on viscosity and soil-moisture tension.

Moisture Transfer

The second question arising from Peck's work concerned transfer of soil moisture in cooled soils with temperature gradients. The excess moisture reported in soils by Klock, and reviewed above, can be partly attributed to an increased retention of rain or meltwater entering the soil surface; however, water is also transferred from lower warmer soil layers. Movement of this soil moisture probably takes place in the following way. Soil moisture is stored in films around soil particles. The closer a water molecule is to the soil particle, the more tightly it is bound. For instance, very little tension is required to move soil moisture when it is near field capacity;

but, as the film is reduced, the percent soil moisture nears the wilting point, and tensions in excess of 15 bars may be needed. In the Jensen study, it was found that tension required to move moisture at a given film thickness was temperature dependent. Cooler soil tended to require more tension to remove water than a warmer soil.

Consider a series of soil grains covered with a continuous moisture film at the same temperature. Neglecting gravitation effects, the film thickness should be about the same along the whole series. If a thermal gradient is now imposed, moisture will move from the warmer to the cooler granules until the various film thicknesses have the same tensions acting on them. The rate and the amount of moisture moved will depend on soil types, imposed gradients, and a number of other factors. At this time there seems to be no good way of quantitatively predicting this effect. However, the increase in moisture in the upper layer does affect the infiltration rate and it sets the stage for discussing soil moisture movement in the ground when freezing does occur.

To summarize, studies have shown that an increase of viscosity and surface tension do cause the soil to retain more moisture when cool, that transfer does occur from below because of a temperature gradient, and it is apparent that should a cool rain fall on cool ground, more moisture would be retained near the surface than would be expected in a warmer soil or one with an isothermal temperature distribution.

Effects of Freezing on Soil Moisture

While Peck (1973) found an increase of about 12 percent in the soil moisture held in cold non-frozen soils over the expected field capacity, Post and Dreibelbis (1942) found soil moisture values around 160 percent in frozen soil which, when unfrozen, had held a maximum of around 40 percent. During one severe winter they found that the soil moisture varied between 23 and 213 percent in the frozen surface soil and 25 to 40 percent in the unfrozen soil below.

Much work has been done on soil moisture movement in conjunction with freezing temperatures. Jumikis (1973) has developed a chart relating phenomena such as moisture transport, transmissibility, and frost penetration depth to porosity. He indicated that porosities of less than 27 percent are rare in soils and that maximum moisture transport occurs for soils with a porosity around 40 percent. He states that "The porosity is the primary factor controlling the amounts of heat flow, moisture transfer, and all related phenomena . . . ."

Should the soil be drier than some threshold value, chances for a continuous film are reduced; when the film is broken, moisture transport can only take place by vapor transport. Several investigators (Jumikis, 1972; Kapotov, 1971; and Ferguson et al., 1964) have felt that vapor transport was negligible in moving significant amounts of moisture. Ferguson et al. (1946) reported, "No water movement to the frozen zone occurred when the soil-water tension was greater than about 5 atm."

Harlan (1973) noted the effect of soil texture on transport:

> "First the magnitude of the effect of upward soil-water migration to the freezing front on ground water levels decreases from coarse textured soils to fine textured soils and with increase in depth to the water table. Second, soil water redistribution is increasingly restricted to the upper portion of the profile in closer proximity to the freezing front as the soil texture becomes finer and as initial moisture contents are reduced."

A model proposed to explain the movement of moisture to the freezing front is similar to that for film migration in unfrozen soil but with some additional features. As the free water freezes around a soil particle, the liquid film becomes thinner. Mineral concentrations increase in the remaining water setting up chemical gradients in addition to the thermal gradients already present. Expansion of the soil structure upon freezing increases pore volume for further moisture migration following subsequent thaw and refreeze cycles.

The effect of soil moisture transport on infiltration is difficult to deal with quantitatively. Harlan (1973) indicates that there is insufficient data available to make quantitative estimates of transported moisture even in the laboratory. He did, however, find that only the tight soils like clays, which would have a low infiltration capacity in any event, tended to pond water on their surface as a result of moisture transferred from below, and that looser soils tended to return moisture to lower levels fairly rapidly once temperature gradients were removed or reversed.

Problems which cold temperatures introduce to flood forecasting have been found to be significant. The movement of soil moisture by thermal gradients and the increased retention due to cold temperatures are partially understood so that they can be qualitatively considered; however, the measurements and theory are not sufficient to make quantitative estimates. Frost occurrence and penetration, which are also very significant in forecasting runoff, seem to be more readily and generally modeled.

CHAPTER THREE. MATHEMATICAL APPROACH TO FROST PENETRATION

The primary factors to consider in modeling frozen soil are:

1. the thermal properties determining the gain or loss of heat;
2. the moisture available to freeze and the effect that the moisture has on the thermal properties; and
3. the temperature at which freezing takes place.

The relationship between the various thermal properties, the required input data, and the initial and boundary conditions can best be understood if they can be related in some equation.

Frost Penetration Equation

There are several frost front penetration equations (Jumikis, 1956; Shannon, 1945) but some of these require variables which are not normally measured or for which the values are not widely published. The equation which seems best adapted to available data is one given by Van Wijk (1963), developed by the Corps of Engineers in 1949. It is given here in cgs units:

$$X = \left(\frac{8.64 \times 10^4 K\ F}{L + C(T_o + \frac{F}{2t})} \right)^{1/2} \qquad (5)$$

where:

X is the depth or thickness of the frozen layer in cm;

K is the thermal conductivity in cal cm^{-1} sec^{-1} $°C^{-1}$;

F is the frost index $\sum_{i}^{n}(T_i - T_{base})$ degree days where summation is over the days in the season;

L is the average latent heat in cal cm^{-3};

T_o is the mean annual air temperature in °C;

t is the duration of the freezing period in days; and

C is the average volumetric heat capacity in cal cm^{-3} $°C^{-1}$.

Where the equation is to be applied to layered soils, the latent heat and heat capacity must be computed using the forms:

$$L = \frac{L_1 d_1 + \cdot + L_i d_i + \cdot + L_n d_n}{\sum_{i=1}^{n} d_i} \qquad (6)$$

$$C = \frac{\sum_{i=1}^{n} C_i d_i}{\sum_{i=1}^{n} d_i} \qquad (7)$$

where L_i is the latent heat, C_i is the volumetric heat capacity, and d_i is the thickness of the i^{th} layer. The layers can be summed only down to the frost front so that the n^{th} layer is the one in which the frost front is currently found. If litter, snow, etc., form a layer above the mineral soil, then the latent heats and specific heats for these layers must be included.

Theoretical Basis for Equation

Theoretical support for the frost penetration equation is based on the principles of heat transfer. Basically, these are given by (1) the conservation of energy

$$\frac{\partial u}{\partial t} + \text{div } q = 0 \qquad (8)$$

where u is the internal energy per unit volume and q is the heat flux density; and (2) the defining equation for the coefficient of thermal conductivity

$$q = -K \text{ (grad T)} \qquad (9)$$

where T is the temperature field in the material through which the heat flows and K is the coefficient of thermal conductivity. The analytical equations determined from heat transfer have exact solutions, but since the necessary field data rarely are known, approximations and simplifications need to be made.

To freeze, a layer of thickness X, having an average temperature T_{av} (°C), the energy per unit area (u_{loss}) that the layer must lose is

$$u_{loss} = X(CT_{av}+L) \qquad (10)$$

where XCT_{av} is the energy to cool the layer to the freezing point and XL is the latent heat of fusion that must be given up. This equation estimates the frost penetration on a seasonal basis. The soil temperature used in the equation represents the heat that must be lost during the season to bring the soil moisture to the freezing temperature. It is known that soil temperatures fluctuate about some mean value during the year and the following holds for this mean value:

1. It is very close to the mean annual air temperature.

2. It is the temperature several feet deep in the soil where seasonal temperature changes cease to be measurable.

3. At some time during the fall, the mean daily temperature at the surface layer cools to the mean annual temperature.

4. At this time heat begins to be conducted from the lower levels of the soil up to the surface rather than into the ground.

This is the beginning time for our frost season. For a number of reasons, computed energy loss based on a frost index, $\sum_{i=1}^{n} (T_{base}-T_i)$, may not be sufficient to account for observed depth of frost when T_{base} is 0°C. To correct this situation, T_{base} may have to be adjusted by a degree or two above or below 0°C. This will vary from place to place and will have to be determined by calibration.

The frozen soil above the frost front must continue to lose heat to maintain the thermal gradient. This heat is considered in the term F/2t. The frost index F will be positive during periods of freeze. The term 2t tends to give the effect of a steep or shallow thermal gradient. The heat per unit area which must be lost, u', can be written

$$u'_{loss} = X[L+C(T_o+F/2t)] \tag{11}$$

The rate that heat flows from the ground is proportional to the thermal conductivity K and to the thermal gradient. The thermal conductivity as given in equation (5) is in units of heat flux per second. It is multiplied by $8.64\text{x}10^4$ to convert it to units of heat flux per day. Measurements of the thermal gradient are rarely available. Instead, a term F/X is substituted. The energy per unit area flowing from the soil is then roughly

$$u'_{loss} = 8.64\text{x}10^4 KF/X \tag{12}$$

Setting the energy that must be lost equal to the energy lost due to the thermal gradient gives

$$X[L+C(T_o+F/2t)] = 8.64\text{x}10^4 KF/X \tag{13}$$

which when solved for X gives equation (5).

While this equation was developed to be used on a seasonal basis, this study requires a model that computes penetration on a daily basis. This effect is achieved by recomputing the seasonal frost penetration each new day. It is assumed that conductivity within the soil remains fairly constant, but each day new cumulative values of F are used.

Effect of Energy Conducted from Below

At the same time that energy is being lost to the surface (equation(12)), energy is flowing from below to replace the lost energy. In more temperate climates the soil has more heat to lose, while in colder climates deeper penetration results from the same values of specific heat and frost index. Aldrich and Paynter (1953) adopted a correction coefficient to apply the Berggren Formula (Shannon,1945) showing smaller corrections in Kansas and Nebraska and larger in the Dakotas. The correction factor increased toward Alaska (page 30, Aldrich and Paynter, 1953). In Alaska, Aldrich and Paynter (1953) preferred an equation:

$$X = \left[\frac{12.22\text{x}10^{4}KF}{L+C(T_{\circ}+\frac{F}{2t})}\right]^{1/2} \tag{14}$$

The penetration computed in equation (14) is greater than that in equation (5) by a factor of $\sqrt{2}$, but, they were dealing with arctic applications.

Determination of Thermal Properties

Computation of frost penetration requires a determination of effective values of thermal conductivity, latent heat, and specific heat.

Thermal Conductivity

Since few soils are homogeneous, the composite value must be used for conductivity. De Vries (1963) gives the composite conductivity of a moist soil as:

$$\lambda = \frac{\Sigma\lambda_i X_i k_i}{\Sigma k_i X_i} \tag{15}$$

where λ_i is the conductivity in cal cm^{-1} sec^{-1} $^{\circ}$C^{-1} of the ith individual substance (soil, water, humus, etc.) by itself, X_i is the volume fraction in percent of the total volume filled by the ith material and k_i is the ratio of the thermal gradients in each minor substance to that through which most heat is transported, which, for simplicity, is taken to be water. This is given approximately by:

$$k_i = 1/3 \sum_{m = \alpha, \beta, \gamma} [1+(\frac{\lambda_p}{\lambda_o} - 1)g_m]^{-1} \tag{16}$$

where α, β, and γ are related to the axes of an ellipsoid approximating the soil particle. It is assumed that the soil particles are spheres so that $\alpha = \beta = \gamma = 1$ and $g_m = g_a$. The subscript "p" indicates the particles; and "o" the medium. It has been found that g_a is around 0.125 for dry soils and around 0.5 for soils near field capacity and above. It is called the depolarization factor and is related to the assumed shape of the soil granule. For dry soils, a correction factor of 1.25 must multiply the apparent conductivity. The conductivities of water, ice, and air are 1.35, 5.2, and 0.062 mcal sec^{-1} cm^{-2} °C^{-1} cm^{-1}, respectively. The value for mineral soils is about 7 mcal sec^{-1} cm^{-2} °C^{-1} cm^{-1} and for humus .08 mcal sec^{-1} cm^{-2} °C^{-1} cm^{-1}.

The logic behind these equations is that if only one substance were present, such as rock, the conductivity would be that for the rock; if it were all water the conductivity would be that for water; when they both are present and are the only things present, the conductivity must be somewhere between that for either one of them. However, if one substance has a much smaller conductivity than another, practically all heat transfer takes place in the medium of higher conductivity. For instance, the ratio of mineral soil to dry air is 7/0.062 = 113. Most of the heat transfer takes place in the soil, and transfer in the air can be neglected. The conductivity is just reduced by the amount of air present. If sufficient air is present, then it may begin to be significant; but in moist soils the bulk of the heat transfer is in the soil and water. In many soils, blocky soil particles are in poor thermal contact with each other. Water films tend to have their greatest thickness at soil contact points and can conduct most of the heat. Therefore, water is taken to be the more general medium of heat transport.

The different layers in soil tend to be composed of different amounts of water, air, soil material, humus, etc.; therefore, the thermal conductivity of the different layers also varies. The conductivity used in equation (5) must be an effective conductivity, which a homogeneous soil the same thickness would have. The effective conductivity can best be determined by summing the soil thermal resistances. The thermal resistance (R_i) of a layer is defined as

$$R_i = \frac{d_i}{K_i} \tag{17}$$

where d_i is the layer thickness and K_i the layer conductivity. The total resistance of a layer with a profile of n sublayers is

$$R_{total} \sum_{i=1}^{n} R_i = \frac{d_1}{K_1} + \frac{d_2}{K_2} + \cdots + \frac{d_i}{K_i} + \cdots + \frac{d_n}{K_n} \tag{18}$$

and the effective thermal conductivity to that depth is

$$K = \frac{d_{total}}{R_{total}} \quad . \tag{19}$$

Where snow and an organic layer are involved, they may exert the major effect on the conductivity. The final value for the effective conductivity should result from an iteration of successive estimates of frost depth (d_{total}).

Specific Heat

The dependence of the conductivity on the materials present applies equally to the heat capacity. De Vries (1963) gives the heat capacity as

$$C = X_s C_s + X_w C_w + X_a C_a \tag{20}$$

where X_s, X_w, and X_a are the fractions of the total volume made up by soil, water, and air, respectively. Kersten (1949) found that the specific heat of most soil minerals varied in a linear fashion from 0.16 cal cm^{-3} °C^{-1} at -18°C to 0.19 cal cm^{-3} °C^{-1} at 60°C. The specific heat at 0°C should be around 0.17 cal cm^{-3} °C^{-1}. The specific heat of air is 0.00030 cal cm^{-3} °C^{-1} and can be neglected. The heat capacity is then:

$$C = (1\text{-porosity})\ 0.17 + \text{volume fraction of soil moisture} \tag{21}$$

If there is organic material, an additional term $X_o C_o$ must be added. C_o averages about 0.6 cal cm^{-3} °C^{-1}.

Latent Heat

The latent heat of fusion is only exchanged by a substance when melting or freezing, and in the case of frozen soils, the latent heat of a layer depends on the moisture in the layer. At zero degrees, the latent heat of water is 79.71 cal/gm. Assuming a constant density of 1 gm/cm^3 for water, the thickness of a frozen soil layer in centimeters times the volume fraction of soil moisture times 79.7 gives the energy per cm^2 released as latent heat in that layer.

Soil Moisture Estimation

The effect of soil moisture on the thermal conductivity, specific heat capacity, and the latent heat has been made evident in the preceding sections. When freezing occurs, however, even the nature of the soil and water seem to change. Post and Dreibelbis (1942) noted that during the freezing process the water holding forces of the soil were overcome and the ice became the soil carrier rather than the soil being the carrier of the ice. The average volume weights of frozen Keene and Muskingum silt loams were 0.63 and 0.93 gm/cm^3, respectively, while the unfrozen values for these same soils were 1.27 and 1.36 gm/cm^3, respectively. To predict frost penetration, it is very important to have a good estimate of soil moisture and its distribution in the soil.

To formulate means of estimating quantity and distribution of soil moisture the following factors should be considered:

1. the means of gaining and losing moisture in the soil;
2. the nature of forces holding moisture to the soil;
3. the forces which take moisture from the soil and how these forces are distributed;
4. the continuous nature of soil moisture distributions, i.e., in fairly homogeneous soils sharp discontinuities in concentration tend to dissipate with time;
5. the existence of practical equations that can describe the nature of the soil moisture distribution; and
6. the corrections that are needed and available to fit the general theory with the variations from standard conditions that are postulated in its formulation.

Basic Assumptions

There are several generally accepted basic assumptions which serve as a foundation for estimating the percent soil moisture in the soil and how it is partitioned among the layers.

First, field capacity is defined as the maximum soil moisture that a given layer of soil can hold against gravity. It may temporarily be exceeded, but if no more input is received, the soil should drain to field capacity. When rains, snowmelt, or irrigation supply water to a soil in excess of field capacity, this water will be lost to runoff or groundwater recharge. Water loss from a layer to groundwater can take place only when the layer is at or above field capacity.

Second, layers with soil moisture contents below field capacity which are located some distance below the surface can only lose water by transpiration. Moisture movement during the growing season from lower layers to the surface by potential gradients not involving transpiration can be considered negligible in agricultural and forested areas (Buckman and Brady, 1969). Transpiration can occur only as long as the percent soil moisture is above the wilting point. This means that deeper layers are limited in the amount of moisture that they can lose, and they can only lose it when occupied by roots of transpiring plants. Layers on the surface can evaporate directly and can dry to the level of the hygroscopic coefficient. This can be done even when the soil is barren.

Third, significant moisture transfer takes place only where continuous films of water exist in the soil. This tends to limit the depth of soil subject to direct evaporation.

Fourth, evaporation and transpiration are greatly reduced during cold weather. It has generally been found that ground water recharge takes place during the winter and early spring periods of the year.

Fifth, naturally mobile substances such as liquid water do not maintain discontinuities in concentration or excessively steep concentration gradients for long periods.

Soil Moisture Budget

A fundamental means of estimating the total soil moisture in a soil profile is the maintenance of a soil moisture budget. Inputs are estimated by precipitation gages, irrigation volumes, or from estimates of snowmelt when snow is present. Output is measured in the form of runoff at stream gages, or estimated in terms of evaporation, transpiration, or groundwater recharge. Groundwater recharge can be ignored whenever the soil moisture content is below field capacity.

Evaporation and transpiration are dependent on the amount of soil moisture, the type of crop, and the soil texture. However, a maximum rate can be measured with an evaporation pan or through the use of a number of different equations. This maximum rate is assumed to be that from a thin film of free water. This evaporation rate is often called potential evaporation (PE) to differentiate it from the actual evapotranspiration.

Actual Evaporation

For many years studies have shown that most plants with plenty of water will transpire at a rate very similar to that of a lake or a free water surface. At the same time, soils which have been wetted evaporate at an initial rate close to the PE. After the soil begins to dry, this rate shows a marked drop. Some understanding of this reduction can be obtained by looking at a curve relating soil tension and percent soil moisture. In Figure 1 it can be seen that when the soil is rather moist the tension increases slowly for an incremental loss of soil moisture. For many soils the curve seems to break rather sharply at a certain point with the tension increasing sharply for a unit decrease in soil moisture. It stands to reason that increased energy is required to overcome the increased force holding the water to the soil.

Zahner (1965) discusses assumptions made by Thornthwaite and Mather (1955) and Penman (1956) in describing the reduction in the evaporation rate as soil moisture is reduced. Zahner (1965) modified their concepts to model evapotranspiration from forest soils of various textures. He proposed that evapotranspiration proceeds at the maximum rate so long as the tension is less than 2 bars. When the tension increases to 2 bars, the rate of evaporation is reduced to some other

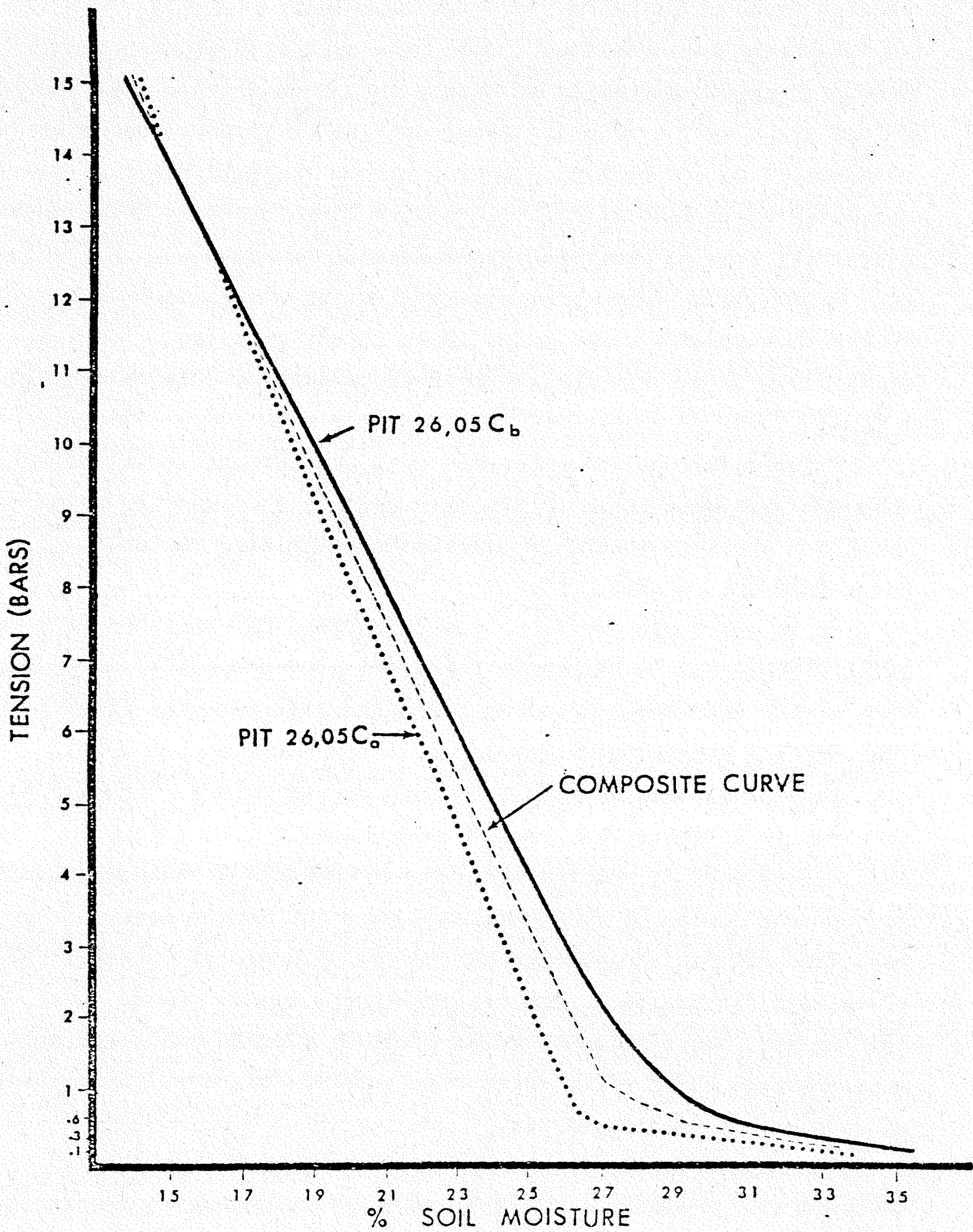

Figure 1.--Soil moisture tension curves for two pits dug in Wellston silt loam at Coshocton, Ohio. Dashed line is composite curve.

rate which is determined by the texture of the soil. This reduced rate is proportional to the fraction of the available water left in the soil. The available water is defined as that soil moisture between field capacity and the wilting point.

Mustonen and McGuiness (1968) have made studies of a similar nature on cultivated soils in Ohio. They show a graph on which several curves are drawn to describe the reduction of evaporation. They propose a reduction in evaporation proportional to the 0.25 power of the percent available water remaining in the soil. They developed their result using data from the weighing lysimeters at Coshocton.

Yeron (1973) found good agreement using a reduction curve similar to the Thornthwaite curve under wheat fields in Israel. Leaf and Brink (1973) similarly feel that in the subalpine forests of Colorado a linear curve is correct.

Denmead (1970), however, says that the reduction will be slight when the leaf area index of the crop gets up to around four indicating that soil moisture itself is not the only limiting factor.

Cover Crop and Seasonal Effects

Using the lysimeter data from Coshocton, Mustonen and McGuiness (1968) found differences in transpiration between meadow grass, corn, and wheat. Much of the difference resulted from different planting times. Once the root systems became fully developed, differences became small.

They also found at Coshocton that harvesting of pasture grass for hay resulted in a significant temporary decrease in transpiration. Similarly, the greater part of transpiration from deciduous forests takes place through the leaves of the trees, and when the leaves fall, moisture loss from all but the surface of the soil practically ceases. When trees leaf out in the spring, transpiration begins again. Leaf fall, leafing out, and root development of newly planted crops are not instantaneous events, but occur over a period of days. This effect can and must be accounted for.

When the effects of season, harvesting, ground cover, and soil moisture are considered, a fairly good approximation to actual evaporation should be obtained if the potential evaporation is known.

Potential Evaporation

There are many equations that estimate evaporation. One of the simplest is the aerodynamic equation which regresses pan evaporation against the product of wind and the vapor pressure difference between the water and the air. With a constant term to account for periods of no wind, the equation developed at Lake Hefner (see Lake Hefner Report, 1954) is

$$E_p = (e_o - e_a)(0.42+0.0040\ u_p) \tag{22}$$

where E_p is the evaporation in inches from the Class A pan, e_o is the vapor pressure in inches of Hg. of the water surface, e_a is the vapor pressure of the air, and u_p is the total daily wind movement in miles per day 8 inches above the pan.

It has been pointed out, however, that if a pan is available, there is little need to compute evaporation. Penman (1948) developed his equation which eliminated the need for pan water temperature by combining the aerodynamic equation with an energy budget concept. He did this by defining the following equations:

$$E = (e_s - e_a)f(u) = (e_s - e_a)(0.42+0.0040\ u_p) \tag{23}$$

$$E_a = (e_{sa} - e_a)f(u) \tag{24}$$

$$B = \gamma \frac{T_s - T_a}{e_s - e_a} = \frac{H_s}{H_E} \tag{25}$$

$$\Delta = \frac{e_s - e_a}{T_s - T_a}\ , \text{ and} \tag{26}$$

$$Q_n = E + H_s \tag{27}$$

where e_{sa} is the saturation vapor pressure at the temperature of the air; B is the Bowan ratio, the ratio of energy transfer by sensible heat H_s to that by evaporation H_E; γ is a constant; f(u) is a wind function such as in equation (23); T_s and T_a are the temperatures in °F of a water surface and the air respectively; Δ is the slope of the vapor pressure-temperature curve; and Q_n is the net radiation in langleys. All other terms are as previously defined. Manipulating these equations, Penman derived the equation:

$$E = \frac{\Delta Q_n + \gamma E_a}{\Delta + \gamma} \tag{28}$$

which describes the evaporation from a hypothetical pan. Kohler, Nordenson, and Fox (1955) multiplied this value by a coefficient 0.7 to get an equivalent lake or potential evaporation. This coefficient is the same as determined for a Class A pan and varies somewhat for different climates and during different seasons.

Lamoreaux (1962) adapted this equation for computer processing. The key to the adaptation was the use of the Clausius-Clapyron equation to express the vapor pressure:

$$e = \exp[-(\frac{K}{T+b})+C] \tag{29}$$

where e is the vapor pressure in inches of Hg, T is the temperature in °F, K, b, and C are constants with the values K=7482.6, b=398.36, and C=15.674; and an equation to fit an empirically derived curve to express the net radiation Q_n in langleys in terms of the incident short wave radiation (R):

$$Q_n = \exp[(T_a-212)(0.1024-0.01066 \ln R)-0.0001] \tag{30}$$

where R is the short wave radiation. Combining equations 23, 28, 29, and 30 gives

$$E = [\frac{\exp(T_a-212)(0.1021-0.01066 \ln R)-0.001+0.0105(e_s-e_a)^{0.88}f(u)}{0.015+(T_a+389.36)^{-2}(6.855\times10^{10}) \exp[-7452.6(T_a+398.36)]}] \tag{31}$$

This equation was used by Mustonen and McGuiness (1968) in their work and was considered satisfactory to describe the evaporation process. This equation has also been tested against several types of pans in several areas of the country with satisfactory results.

If radiation is not measured anywhere in the area of interest, the Thornthwaite equation can be used.

Distribution of Transpired Soil Moisture Loss with Depth

Mustonen and McGuiness (1968) quote a study made at Coshocton which showed that 85 percent of the total annual evapotranspiration was from the surface meter of soil and 50 percent was from the upper 18 centimeters. These results to some extent depend on the ground cover (which in this case was alfalfa-bromegrass), the soil texture, and the rainfall distribution, but the overall result should have general application.

Zahner (1965) quotes several studies where it was found:

> ". . . soil water is withdrawn from zones where it is most readily available and is removed where root density is highest. Forest root depth is inversely proportional to soil depth even in deeper well-aerated horizons and water is thus withdrawn in inverse proportion to depth."

To illustrate this situation, the soil moisture in various layers has been plotted as a function of time in Figures 2 and 3 using data published in the 1962 and 1963 volumes of Climatological Data (U. S. Weather Bureau, 1962, 1963). It can be seen that for these soils, the upper layer frequently has the soil moisture depleted down to the wilting point. The lower layers tend to lose moisture much more slowly with the bottom layers being depleted last and recharged last. Because of humus and because recharge enters at the surface, the upper layers were above field capacity in several cases. During early parts of 1962 the soil temperature and frost conditions could have been a factor.

Temperatures at Which Soils Freeze

It has been assumed in the previous discussions that soil moisture freezes at 0°C. This deserves some discussion. The temperature at which soils freeze is not always easy to predict. Post and Dreibelbis (1942) found that:

> ". . . with temperatures as low as 24°F. 1 inch below the frozen layer there was no evidence of freezing, as indicated by macroscopic ice crystal growth or changes in soil structure in well drained soils consisting of silt and decomposed shale and sandstone."

AVAILABLE SOIL MOISTURE FOR COSHOCTON, OHIO 1962

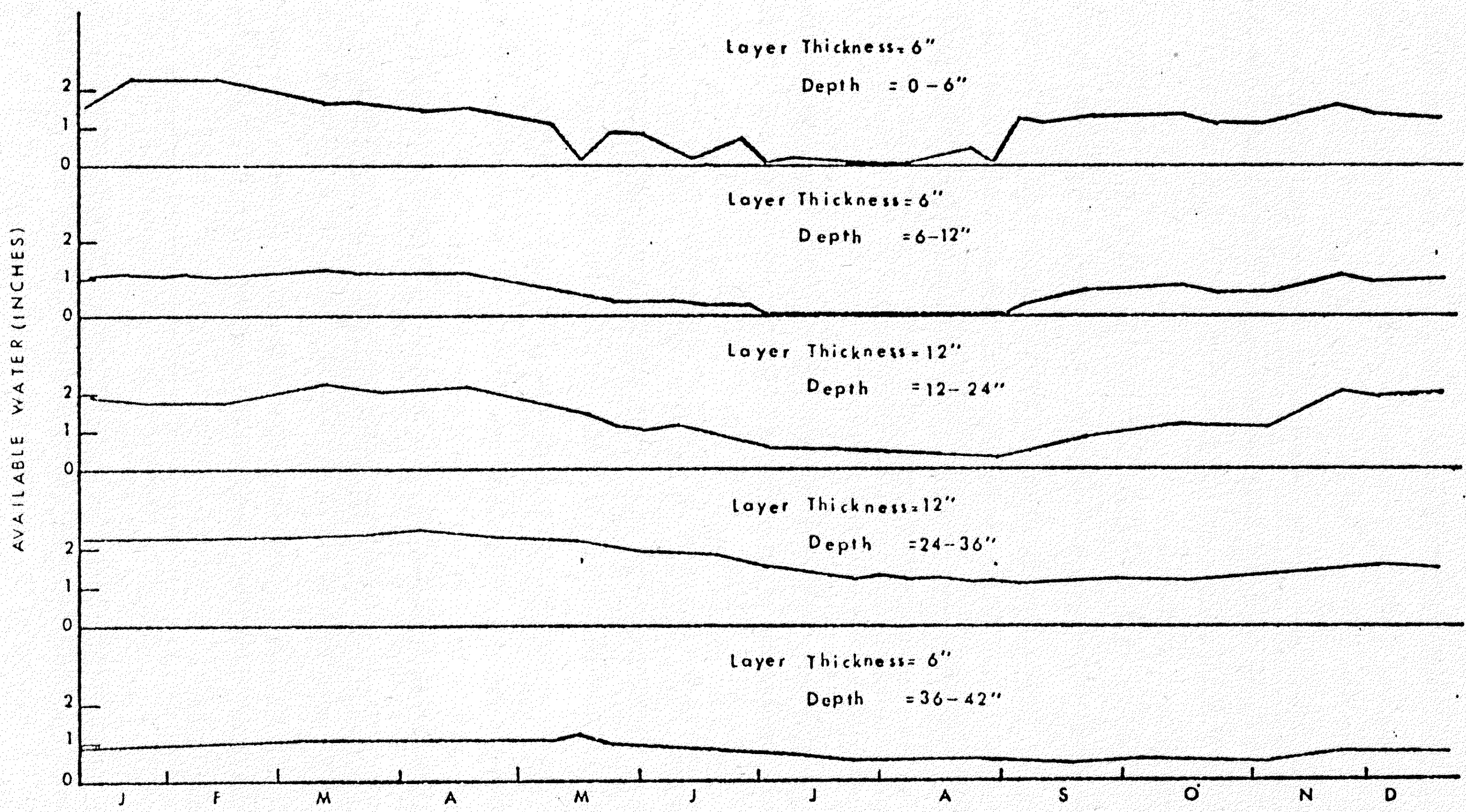

Figure 2.--Variation in soil moisture of five soil layers. Sampling point is near Watershed 109 at Coshocton, Ohio, Data are from 1962.

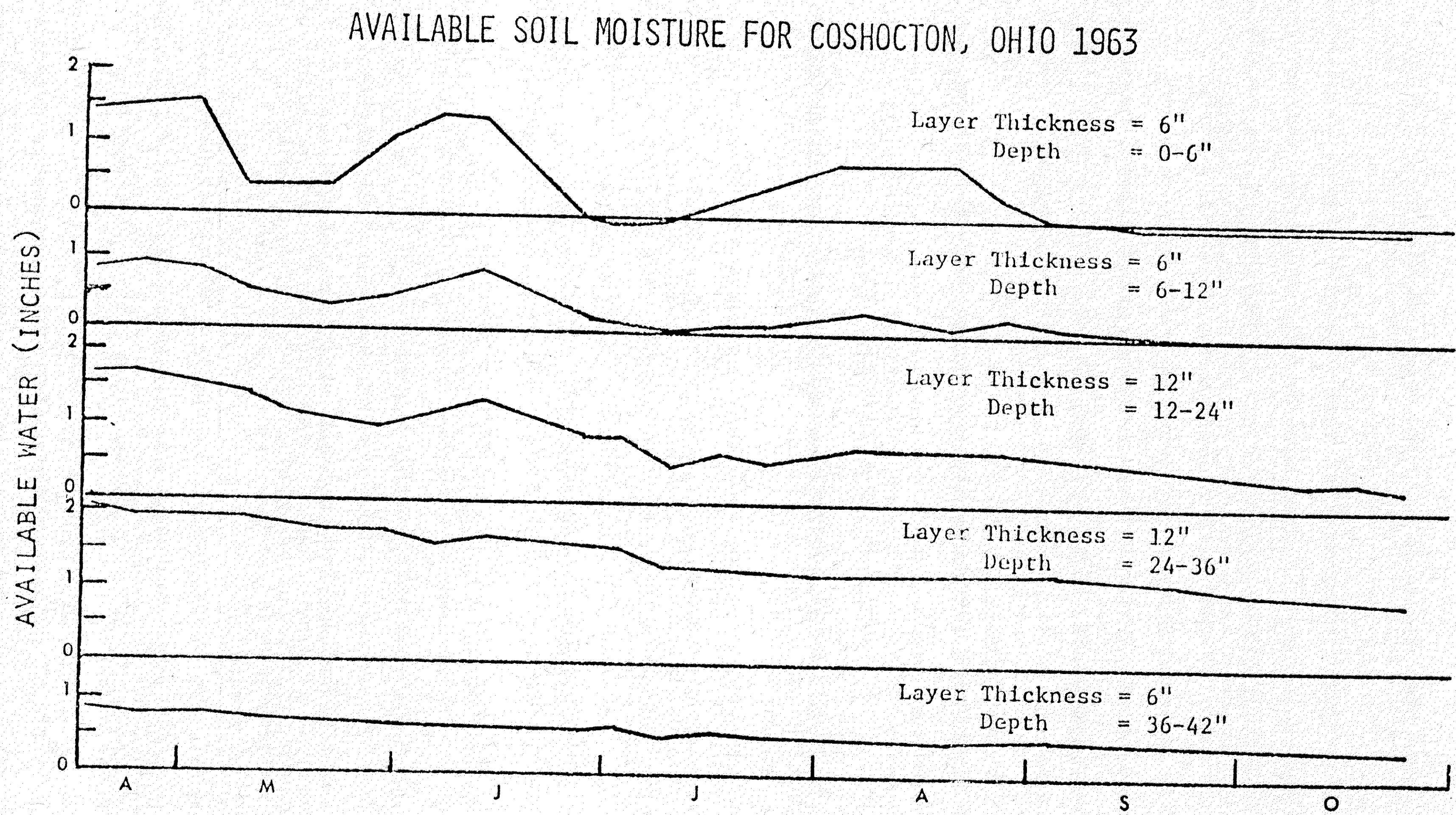

Figure 3.--Variation in soil moisture of five soil layers. Data were sampled at same point as in figure 2. Data are from 1963.

They found that freezing did take place as low as -3.3°C for 1/2 inch depth, -5.5°C for the 3 inch depth, and -4°C for the 6 inch depth. Bouyoucos (1921) reported that free water may freeze at -1.5°C capillary water at -4°C, but that hygroscopic water may not freeze until temperatures go as low as -78°C. Earlier he had found that soils could withstand a considerable amount of supercooling. Perfectly still water saturating sand, loam, or clay could be cooled to -4.2°C. Anderson et al. (1942) related freezing point to soil moisture in a curve plotted to show the change in the freezing point depression as a function of soil moisture. For Aiken clay loam, a second order equation which approximates the curve reasonably well in the range of soil moisture between wilting point at 19.5 percent and field capacity at 31 percent is

$$\Delta T = -7.4+0.448\ SMP-0.007(SMP)^2 \tag{32}$$

where SMP is the volume fraction of soil moisture and ΔT the freezing point depression (°C).

With limited temperature data available and variable field conditions, perhaps the most applicable description comes from Beskow (1947).

> "To summarize, then we can say that for coarse non-frost acting soils the temperature for freezing and thawing is practically the same, and is very slightly less than 0°C. For frost heaving soils there is quite a difference between the frost line temperature during freezing and thawing, the receding frost line temperature is 0°C. or very slightly less, while the frost line temperature during freezing is considerably lowered being greater the finer the soil is and the faster it freezes."

This statement requires some explanation of a difference between non-frost acting soils and frost susceptible soils. Linell and Haley (1952) of the Corps of Engineers gave this definition:

> "Frost susceptible soils are those in which significant ice segregation will occur when moisture is available and the requisite freezing conditions are present."

Summary of Soil Thermal Properties

To summarize this section, thermal properties necessary for frost computations are obtainable when the soil moisture can be determined. Soil moisture computations can be made in a general sense to sufficient accuracy if sufficient information is available about soils and weather. The freezing point, while not constant, can be treated as the freezing point of free water except in the case of frost susceptible soils. Frost-susceptible soils are greatly subject to stalactite ice in temperate climates which would not by itself aggravate the runoff problem.

Adjustments to Data

It is necessary to point out that there are several differences between the assumptions going into the equations and specific conditions in the field. For instance, in computing evaporation, the net energy used was based on the assumption that the amount of incident solar radiation that is reflected would be 5 percent or less, that the equivalent pan would be unshaded during the entire day, that the evaporating surface is always horizontal, and that temperatures are measured in the immediate vicinity for which evaporation is estimated. While these assumptions are logical for a pan, they can vary significantly from those of a watershed. Some set of rules are required to make corrections. This requirement leads to a brief review of energy exchange theory.

The principle source of energy to the earth's surface is the sun. The atmosphere, having been warmed by the sun, also supplies energy. Vegetation, structures, and the earth itself also exchange energy, being at times sources and at other times sinks.

Shortwave Radiation

Solar energy falls into two almost separate spectral bands. The band with the shorter wavelength will be discussed first. The shortwave energy includes visible light and is in the spectral band from 0.15 to 4 micrometers. The bulk of solar radiation at the top of the atmosphere is in this range.

Albedo

When this shortwave energy falls upon a surface, a certain fraction of it is reflected. The ratio of this reflected component to the total has been given the name albedo (A). If the short wave energy is represented by the symbol M_s, then the energy absorbed by a surface can be written $M_s(1-A)$.

The albedo is really a complex function of the wavelength of light, the characteristics of the absorbing surface, and even the angle of incidence of the radiation. Values change during the day and the season. Average or effective values have been measured, however, and are useful in adjusting the amount of solar radiation absorbed. Values vary from 0.02 for the ocean to 0.95 for some fresh snow (Munn, 1966). Grass is in the range of 15-40 percent while forests are 10-18 percent. During a growing season, a field is plowed bare, covered with growing crops, irrigated or moistened by rain, dried by the sun, harvested, and in the winter may be covered by snow. With every change, the albedo changes and some of these changes will be significant. These changes must be considered in the computation of energy transfer, especially when computing evaporation.

Slope, Aspect, and Season

Shortwave energy comes in two ways: a direct component, the part that causes shadows; and a diffuse component, the part that lights shaded areas in the daytime. The direct component is often the greatest and is affected significantly by the geometry of the receiving surface.

The irradiance of the direct component varies as the cosine of the angle of incidence. This results in cold winters and warm summers, warmer temperatures and drier soils on south facing slopes than on north facing slopes, and less radiation arriving in polar regions than in regions near the equator. A watershed with a slope having a component in the north or south direction requires a correction for this. Kimball (1919) gives a correction which, in effect, adjusts the latitude of a watershed to a location on the earth having a horizontal surface which receives direct sunlight with the same angle of incidence as the sloping watershed. A similar change could be computed for east and west facing slopes, but it is not considered sufficiently important even though east facing slopes receive the more intense radiation during the morning when the air is cooler, and west facing slopes are more directly radiated in the warmer afternoon. The equation for the slope and aspect correction to latitude is

$$\sin \Delta \phi = \cos(\measuredangle \text{ of azimuth})\sin(\measuredangle \text{ of slope}) \tag{33}$$

where $\Delta\phi$ is the latitude correction.

The seasonal change in the intensity of the incident radiation must also be considered. The equation for the angle of incidence (z) is

$$\cos z = \sin \phi \sin \delta + \cos \phi \cos \delta \cos h \tag{34}$$

where ϕ is the latitude, δ is the declination or the angle that a line from the center of the earth to the sun at solar noon would make with the plane of the equator, and h is the hour angle or the angle between a line from the center of the earth to the sun at any given time and a line from the center of the earth to the sun at solar noon.

The declination goes from 23°27' or 23.45° on June 21 to -23°26' or -23.43° on December 21 (Smithsonian Meteorological Tables). The variation can be computed fairly accurately by the equation

$$\delta = 23.5 \sin(\text{DAS}) \tag{35}$$

$$\text{DAS} = \frac{2\pi}{365} (\text{day of the year } -80) \tag{36}$$

The last term in the equation is just a phase shift to adjust the calendar year to when the sun crosses the equator on the vernal equinox which is on or near the eightieth day of the year.

A further correction must be made for the length of the day. Between the fall and spring equinox, the sun rises in the southeast and sets in the southwest so that any surface with a northern slope has fewer possible minutes of direct sunshine than a flat horizontal surface. A southern slope does not have a longer day than a flat plane since the sun doesn't shine on any surface until it makes a positive angle with the horizon. During spring and summer, a northern slope has the same day length as a horizontal surface while a southern slope does not receive direct rays until slightly later in the day because the sun now rises in the northeast and sets in the northwest. The length of the day is that period when cosine z is greater than zero, or when the hour angle is less than 90° or $\pi/2$ radians. The hour angle h for the beginning or ending of the day may be determined by setting cosine z=0 and solving the following equation

$$h = \cos^{-1}[-\tan(\phi+\Delta\phi)\tan\delta] \tag{37}$$

where $\Delta\phi$ is the latitude correction for slope. This gives the value of h for half of the day. The length of a day on a north or south sloping surface is less than 2h for half of the year by the ratio of h'/h where

$$h = \cos^{-1}(-\tan\phi\tan\delta) \tag{38}$$

and

$$h = \cos^{-1}[-\tan(\phi+\Delta\phi)\tan\delta] \quad . \tag{39}$$

Diffuse Shortwave Radiation

The other component of shortwave radiation is that which has been scattered by the atmosphere or reflected by other objects to such an extent that there is no defined direction of origin. While the intensity of both components is reduced by the cosine of the angle of incidence, the effect of shading does not reduce the diffuse radiation as much as it does the direct component.

When the sky is clear and the sun is at its zenith in the summer, the ratio of direct to diffuse radiation is about 6 to 1. This decreases to 2 to 1 when the sun is 20° above the horizon (Munn, 1966). When the sky is overcast and no shadows are visible most of the light is diffuse. During these times corrections for slope are unnecessary. Because cloudiness is difficult to estimate, the average effect of total cloudiness should be considered in making corrections for slope, albedo, and length of day.

If solar radiation is not measured, corrections are required to use the Thornthwaite equation for estimating evapotranspiration. These are adjustments for the variation in the optical path length, distance to the sun, precipitable water in the atmosphere, and the angle of incidence.

Longwave Radiation

The longwave component of the incident radiation is generally less variable and is not as widely measured as the shortwave component. While it is not used explicitly in any equations in this model, it is used implicitly in equations (5) and (30) and should be considered under vegetative influence on the energy budget. It can be noted that long wave radiation is being exchanged by the ground, vegetation and the atmosphere, and that there is no clear cut direct beam as there is with the shortwave radiation. Longwave radiation is continuous throughout the 24-hour period and is not confined to the daytime.

Energy Exchange by Sensible Heat Exchange

An invasion of cold arctic air or warm tropic air has a definite effect on the soil. The convection of heat to or from the air follows similar heat equation as conductivity within the soil, i.e.:

$$Q = \frac{\Delta T}{\Delta X} h_c \qquad (40)$$

where h_c is the heat transfer coefficient of the air and ΔX is the shelter height. All of these terms are less precisely defined for the air-soil interface than they are within the soil. If the ground is cold and colder heavier air is below relatively warm light air (inversion lapse rate), as it might be on a cold winter night, the exchange of energy takes place very slowly (h_c is small). If a strong pressure gradient exists in the atmosphere, the winds above the ground will tend to force turbulence at the ground surface with the result that much mixing takes place and h_c increases by orders of magnitude. Finally, on days when the ground is sufficiently heated by the sun and convective currents are developed, considerable mixing occurs with a resulting large h_c compared with that for stable air.

Effect of Vegetation on Energy Transfer

Vegetative cover which limits wind flow and surface heating causes conductive exchange of heat between the air and soil to take place much more slowly. Vegetation constantly exchanges longwave radiation with the ground, which tends to make temperatures in the air around the vegetation close to those of the soil. Moisture in the vegetation has a much higher specific heat than air, making those areas more resistant to temperature change. These effects on energy transfer explain why forest soils often freeze several days after soils in open pasture. In addition, freezing is not as great, but for the same reason forest soils also take relatively longer to thaw than open areas.

Energy Exchange by Condensation

Another important form of energy transfer is that due to condensation. The latent heat involved in the change of state has already been discussed. Condensation becomes important in this model when snow is on the ground. The energy released by condensation can melt a significant amount of snow. The ratio of the latent heat of vaporization to the latent heat of fusion is approximately 7.5. Thus, for each millimeter of water condensed out of the air, the heat released is sufficient to melt approximately 7.5 millimeters of water equivalent from the snow.

Energy Transfer from the Ground Below

The final source of energy transfer to be considered is that of heat flow from lower soil layers. There are two sources of heat available. The first is heat that has been received at the surface, from the sun, air, etc., during warm weather and which flows back to the surface when the gradient is reversed. The second is geothermal heat which is conducted from the center of the earth. In some places geothermal heat can be quite important, but in general it is small. Gieger (1966) estimates that the average flux density of geothermal heat is 0.0001 cal cm^{-2} min^{-1}.

The primary source of the energy which can be stored in a soil is radiation from the sun. This energy in the form of heat causes the soil temperature to vary with season. The variation in temperature is generally smaller in soils farther from the surface. At some point seasonal temperature changes are reduced to the fraction 1/e of the surface variation. This is called the seasonal damping depth (van Wijk and de Vries, 1963). The temperature at this depth remains very near the annual mean temperature. The importance of heat flow from the ground is important. It has been pointed out that in many cases thawing of frost takes place from below as well as from above. This effect can be taken into account by computing the damping depth or depth to relatively stable temperature, and then computing the heat flow from this point to the frost front.

To summarize the energy transfer discussion, radiation, conduction, sensible heat transfer, and changes of state are the means of losing and gaining the energy involved in the freezing and thawing processes. Radiation, evaporation, condensation, and sensible heat transfer are the important methods of energy exchange for the soil-air interface while conduction is the most important method of energy exchange below the surface.

Ground Surface Insulators

Energy exchange at the ground surface is significantly affected by natural ground insulators. The presence and extent of natural insulators is quite variable. There are two types of natural insulation that the ground can have. One is snow and the other is litter. The combination of the two is also important and will be discussed.

Snow

Many investigators have reported the insulating effect of snow. Post and Dreibelbis (1942) reported that in Ohio 2 inches of snow were enough to reduce daily changes in frost depth to less than 1 inch. Belotelkin (1941) noted:

> "Snow cover affects soil-freezing to such an extent that even slight trampling of the snow, thereby increasing its conductivity, was found to greatly affect the next remeasurement if taken too near the trampled area."

Atkinson and Bay (1940) made a test in which plots with 0, 6, 12, and 24 inches of snow were observed. The plots with 12 and more inches of snow, which had been frozen before the snowfall, thawed completely a few days following the snowfall. During later periods, temperatures fell below freezing and, although frost did form temporarily in the plot with 12 inches, it formed to a much lesser extent than the plots with shallower snow.

Molchanov (1963) quotes several Russian researchers:

> "According to the researches of A. P. Tel'shil the greatest differences of snow temperature are observed with the first five centimeters of the snow carpet ." (page 116)

> "...pointed out that a layer of snow as thin as 6 cm. decreases the cooling of the soil by 4°C., as compared with bare soil." (page 115)

Molchanov (1963) noted that with air temperatures as low as -20°C. a soil temperature only slightly below 0°C. was noted under a layer of 30-35 cm. snow.

Retention and Drifting of Snow

Because of the beneficial effects of snow both in limiting frost penetration and replenishing soil moisture from crops, shelter belts and snow fences have been erected. Hinman and Bisal (1973) write:

> "The favorable effect of freezing on soil properties may be due in part to the resultant increase in water stable aggregation in respect to the finer aggregate, which is reflected in increased percolation rate. Freeze drying, however, which is destructive to soil aggregates, leaves a thin mantle of fine soil on the surface. This would reduce infiltration of water and at the same time contribute to sheet erosion during rapid snowmelt in the spring. Because freeze-drying does not occur under snow cover it can be minimized by conserving the stubble to catch and hold snow during the winter."

In Russia it has been

> "...established that the snow lies most uniformly on fields fringed by forest strips without underbrush.... If, however, there is a second story, the thicker it is, the higher are the snow drifts at the forest's edge, and the closer to it." (Molchanov (1963) page 87)

He also noted that on

> "...treeless terrain (fields) the thickness of the snow is less by a factor of nearly 3, and the moisture reserve by a factor of 1.6."

The effect of windbreaks is different for different types of snow. A wet snow will be less susceptible to blowing and less likely to drift once it hits the ground, than dry snow. In spite of this, it would be useful to have some quantitative means to estimate snow holding value of stubble and vegetation. Smika and Whitfield (1966) found in Nebraska that fields with wheat stubble had 14 percent more moisture than plowed fields in the same area. Staple et al. (1960) found regression equations for overwinter conservation of water (C) as a function of the moisture in the fall (F), rain (R), and snow (S) for southern Canada. All values were recorded in inches. For a stubble field, the equation was:

$$C = 0.41S+0.35R-0.24F+0.20 \tag{41}$$

with an r value of 0.711 and for fallow fields:

$$C = 0.04S+0.50R-0.16F+0.30 \tag{42}$$

and r = 0.797. It can be seen that snow contributes a significant amount more to the soil moisture if the stubble can trap and hold it. Post and Dreibelbis (1942) wrote:

> "The principle functions which vegetation served were to provide standing, dormant, or dead plant materials for the retardation of drifting snow, and to form a supporting framework which tended to maintain the snow in a loose condition in which it fell."

In Russia Mikhel and Rudneva (1971) developed an equation for drifting snow:

$$I = Cu^3 \tag{43}$$

where I is the transport rate in m^3/running m/sec, u is the wind speed in m/sec at "the wind vane level," and C is a constant. They assumed that drifting snow was fresh with a density of 0.14 grams/cm^3. Using this equation they have developed maps of snow disposition. Granberg (1973) in Canada has taken a different approach and has regressed snow depth with features of topography; i.e., altitude, difference in altitude between a point and the generalized surrounding terrain, general slope, change in slope in upwind direction, and variation in slope in the downwind direction. Using variables made from a combination of these, he developed an equation with an r = 0.87. With this equation he developed a map which compared well with ground truth data, and with a sequence of aerial photographs.

Most of the literature indicates that there are techniques which can be calibrated and used in an area to approximate the snow distribution, but the main general rules are limited to the idea that snow drifting is proportional to wind velocities and snow densities. From this we can make rough estimates that (1) the denser vegetation is, the lower the surface wind speed will be and (2) snow is more likely to drift into this vegetation than out of it; but that, as snow gets deeper and rougher areas are filled with snow, the ability of ground vegetation to hold more snow against wind will be decreased. Studies discussed by Munn (1966) indicate that wind in forests is much smaller within the canopy than above it and may even be uncorrelated with the wind above.

In a large watershed of varying ground cover and topography, the importance of drifting on subsequent runoff would depend on the size of the areas of reduced and increased snow cover and where they are located. Should snow be gathered into north sloping forested ravines, the exposed snow free open areas would be susceptible to freezing, however, the snow having collected in protected areas would tend to melt slowly and have optimum opportunity to infiltrate if the ravine soil is porous. However, should a heavy rain fall on such a basin with the main portion of the land snow free and therefore frozen, it could be very dangerous if the frost were an impervious type.

Thermal Properties

Dry snow has a very low thermal conductivity. Conductivity of wet snow increases to that of fairly dry sand. Snow is composed of ice, air, and liquid water. It is not homogeneous and the percent by volume of each component tends to change with age, temperature, and compaction. Great extremes in density have been reported. Table 1 presents thermal properties for various densities.

TABLE 1

THERMAL PROPERTIES OF SNOW

(After Munn 1966 and Van Wijk 1963)

Density	Conductivity ($cal\ cm^{-1}\ sec^{-1}\ °C^{-1}$)	Volume heat capacity ($cal\ cm^{-3}\ °C^{-1}$)
0.05	0.00015	0.05
0.20	0.00032	0.2
0.50	0.0017	0.5
0.1	0.0018	
0.5	0.0015*	
0.9	0.54	

*Note that value in Munn differs slightly from value in Van Wijk.

Litter

The second basic insulator is litter or mulch. Mulch may be applied by farmers in cultivated fields. Litter may result from dead grass, weeds, leaves, needles, etc., in grasslands, pasture, and forested areas. The depth of the litter layer is not constant. Leaves normally accumulate in the fall, but some trees drop their leaves or needles in the spring. Decay and combination with the mineral soil take place most rapidly during the summer when warm temperatures favor decay activities. Equations have been developed for litter production, but for the purposes of estimating the insulating effect, an experienced estimate of leaves expected in the fall is sufficient. If the mass of leaf litter layer is known in the fall, it is likely to stay constant until spring, and its insulating effects can be adequately simulated.

Grass or brush litter may be quite variable. In the fall many grasses are green and healthy. Being fairly dense the grass restricts air movement near the ground and forms an insulating blanket of air and organic matter. Wind and snow tend to bend the grass down reducing the thickness of the layer. With the onset of freeze some grasses die back, their blades turn brown and lie down to form a thatch. When grass is not covered by snow or leaves, the sensible heat transfer is somewhat dependent upon the wind near the ground. Treating litter of this type will require estimates of effective layer thickness and will likely change during the winter.

De Vries (1963) estimates the conductivity of organic material around 0.0006 cal cm^{-1} sec^{-1} $°C^{-1}$ with a specific heat of 0.6 cal cm^{-3} $°C^{-1}$ and gives a value specifically for humus of 0.8×10^{-4} cal cm^{-1} sec^{-1} $°C^{-1}$. Water content of litter, like soil, has an important effect on the conductivity. Filippova (1956) gives data where highly frozen litter had 710 percent moisture on a dry weight basis. Unfrozen, however, Vil'yums (quoted in Filippova 1956) stated, "The permeability of litter is so great that any quantity of rain or snow water can penetrate into it." This suggests that the moisture content of the litter is to some extent dependent on the soil beneath it. Sozkin (quoted in Filippova (1956)) made experiments with sprinklers which confirmed that absorbtion into the ground with a litter cover is much greater than without it.

Thorud and Anderson (1969) found that litter from different trees varied fairly widely in their insulating qualities. Of the trees that they tested (red pine, white pine, and oak), white pine was the most effective. They also found that litter mixed with snow was a better insulator per unit of depth than either litter or snow alone.

Moisture in the litter caused an up to fourfold increase in the thermal conductivities. This effect again varied with species. Litter soaked and then drained for one half hour resulted in red pine litter having a moisture content of 30 percent of dry weight, while white pine and oak litter had 65 percent. Once freezing sets in, then moisture values can go as high as the 710 percent mentioned by Filippova (1956), (page 5), who noted that moisture in the litter layer varied from 300 percent to 700 percent, but at the decomposed layer it decreased to 99 percent.

Vegetation and Snow

The interaction of vegetation and snow is important but difficult to assess. Who has gone through a meadow with a shallow snow cover and not noticed how the grass tends to hold the snow off the ground, thus providing an air layer between the snow and the soil? This effect will be reduced as the snow becomes deeper and heavier. At the same time, closed canopies of evergreens may prevent the snow from reaching the ground, allowing it to sublimate while still in the trees. The resulting thinner snow blanket allows the cold air to have greater influence on the ground. The effect of soil temperatures under varying covers has been studied by Crabb and Smith (1953) and the variation of frost penetration and duration resulting from various covers has been reported by many such as Thorud and Anderson (1969).

From this discussion it can be seen that vegetation and resulting litter can be very important. Therefore, means for assessing the cover type, density of cover, litter producing characteristics, and snow trapping characteristics are desirable. For basins of any size and variability one of the easiest and quickest means of determining species and density would be by means of aerial photography. A limited number of visits or checks would need to ba made on the ground to determine age and general litter characteistics.

Discriminating Between Frost Types

The factors which appear to be most important in separating the different types of frost are soil type, moisture present, time scale of the temperature drop, organic matter content, and the depth of frost penetration.

Texture

The influence of texture is mainly felt in the water holding capacity of the soil. However, for a soil to be subject to ice segregation, that is for needle ice to form, it is felt that an inorganic soil must contain at least 3 percent or more by weight of grains finer than 0.02 mm in diameter (Casagrande, 1931).

Soil Moisture

Soil moisture content is also important. Kuznik and Bezmenov (1964) found that:

> "Freezing of the soil stops completely under any meteorologic conditions when the moisture content is less than one and one-half times the value of its maximum hydroscopicity."

They also found that soils with moisture content greater than field capacity tended to reduce infiltration to zero when frozen. These results were duplicated by Larin (1962). Mosiyenko (1958) found that soils with more than 50 percent of their total pore space free of ice are permeable to water. Byrnes (1958) found that at -17.8°C and at -9.5°C concrete frost formed in light and heavy soils, respectively, if they were at field capacity or above. Post and Dreibelbis (1942), however, stated that concrete frost "was the only type found with a moisture content of less that 50 percent dry weight." Perhaps this difference in findings is based on something other than moisture content alone. There appears, however, a general concensus that in mineral soils subject to frost, when frost forms in them it is of the concrete type, reducing their permeability to near zero when they are frozen with a soil moisture content at field capacity or above. These results are summarized in Figure 4.

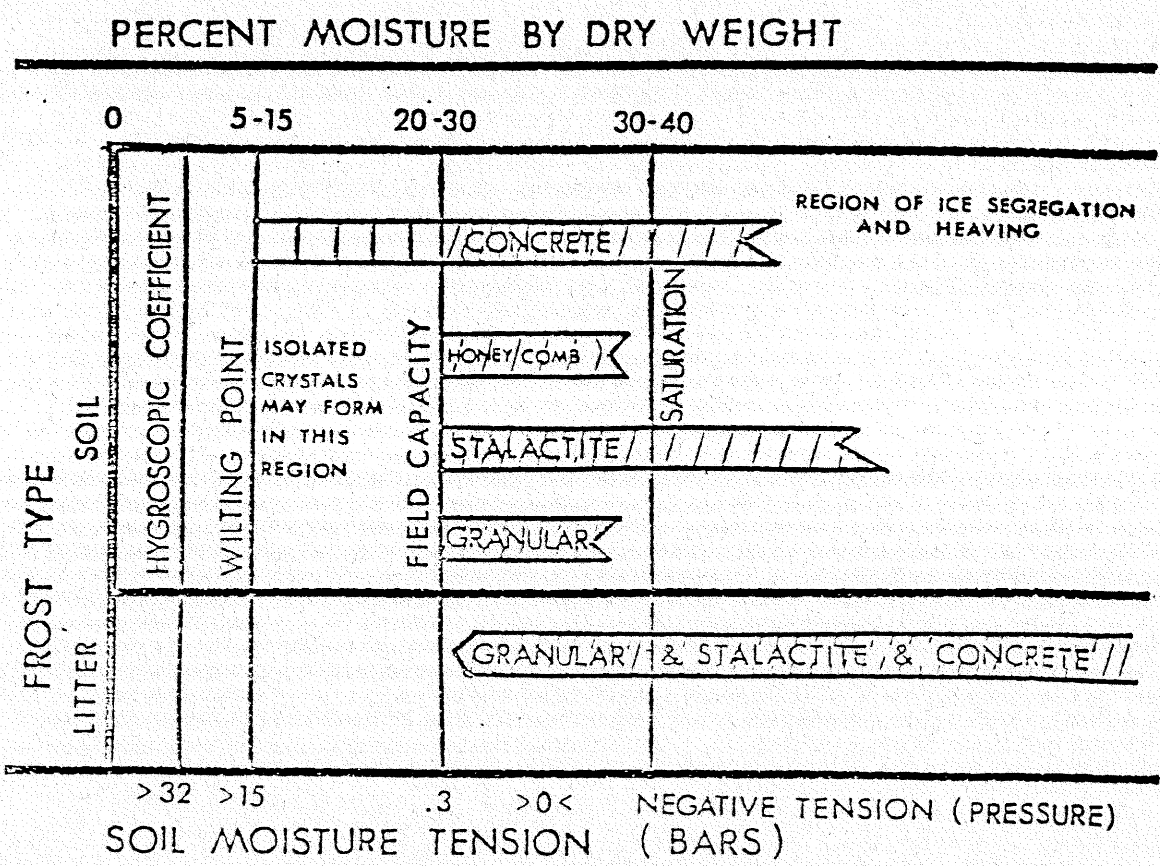

Figure 4.--Relationship of frost type to soil moisture.

Indirectly bearing on this subject is the ability of soils to drain when thawing has occurred before a new cycle of freezing can set in again. In Vermont, Benoit (1973) found that soils thawed and refrozen at high soil moisture content tended to decrease their hydraulic conductivities, while soil of low water content tended to increase their hydraulic conductivity. He derived an equation from 12 treatments of a silt loam

$$Y = 35.16\ e^{-0.081X} \tag{44}$$

where Y is the ratio of the final to the initial conductivity and X is the soil water content at freezing. This result was independent of aggregate size or freezing temperature. Hinman and Bisal (1973) found that alternate freezing and thawing tended to increase the hydraulic conductivity of clays with excess water and reduced the conductivity in loam soils. Alternate freezing reduced the conductivity less in loam soils, however, than a continuous freeze.

Rate of Penetration

The rate of frost penetration is important. Kirkham (1964) states that ice lenses form when the soil is unsaturated and freezes slowly. With a slow freeze the moisture has time to migrate to the frost front. In a rapid freeze the soil water freezes in place causing small separate crystals to form.

Organic Material

In their discussion of the difference between types of frost and their related permeabilities, Post and Dreibelbis (1942) mentioned that where there was very much organic material in the soil, aggregation tended to cause the soil to freeze in fairly large porous granules. The amount of the organic material necessary is that sufficient to prevent the soil from puddling.

Puddling is a process in which the particles of a soil reorient themselves under stress to reduce the apparent specific volume (Baver, 1956). Generally a soil must have enough soil moisture to be in what soil physicists call the moist or wet range for this process to occur. The way that organic material aggregates soil materials is not entirely understood but many interesting aspects are discussed by Baver (1956). He gives an example of a soil being suitable for cultivation at a moisture content up to 52.2 percent without puddling. After the organic matter was removed the soil puddled at 27.7 percent moisture. He also gave an example of a cultivated field where a virgin soil contained 3.9 percent organic matter and a cultivated field in the same area only 2.6 percent. This decrease had taken place over the 60 years of cultivation.

Byrnes (1958) reported that when the organic content was above 2 percent there was an increase in the size of the crystals formed. He also found that generally concrete frost formed in the mineral soils, honeycomb frost formed in the litter layer, and granular frost formed in humus layers, although humus layer generally had to be quite wet for frost to form at all.

The importance of this question at this particular point is to know what percent of organic material is required so puddling will not occur, and so frost occurring in a soil is more likely to be of a porous type than the impermeable concrete type. Undoubtedly this depends on the soils, but it can be said that in forests there is likely to be sufficient organic material to prevent concrete frost from forming in the upper layers, while in bare soils, which may have been heavily cropped, it is probably lacking. Marginal cases would be in meadows and in those cultivated fields where the organic matter is worked into the soil.

Implications

The type of frost depends on soil type, moisture present, and the time scale of the temperature drop. The permeable forms of frost, stalactite, granular, and honeycomb, form mainly near the surface while concrete frost occurs at all levels.

Granular and honeycomb frost form in soils having high organic content which helps the soil aggregate into granules. Therefore, formation of these types of frost is limited to areas where there is a good supply of organic material to be mixed with the soil such as in woodlands and grasslands. Frost penetration below the layers with significant organic content would tend to form ice lensing.

Summary

Equations have been presented which relate air temperature, soil types, and soil moisture with frost occurrence and depth of penetration. Equations have been given to determine the soil moisture both in total value and in distribution so that the thermal properties of soils needed in the frost penetration equation can be computed. Various corrections have been considered. The influence of the ground insulators snow and litter have been discussed. Finally, primary factors which determine the type of frost that will form were reviewed.

In the next chapter these equations will be combined to form the soil moisture and frost prediction model.

CHAPTER FOUR. THE MODEL

The ideas, theories, and observations so far presented must now be expressed in a set of simple assumptions so as to form a set of rules useful in a computer. These are outlined in the following paragraphs. Where appropriate, assumptions in the model will refer to equations in chapter 3. In some cases, no equation is referred to, but the development can be found in chapter 3.

Freezing and Frost Penetration

Frost penetration is proportional to the square root of the frost index, or degree days below freezing. It is inversely proportional to the square root of the energy which a soil layer must lose per unit of depth of penetration in order to cool to the freezing point and give up in latent heat as the soil moisture freezes (equation 5). The rate of heat loss will depend on the conductivities of the soil and their volume heat capacities.

The most important variable affecting the conductivities and heat capacities in the soil is soil moisture (equations 15, 19, and 20).

The presence of litter or snow on the surface will affect the rate of heat loss because of the added layers of insulation. The amount of snow that remains on the ground depends on the cover type and the wind since meadows and bare fields will tend to be blown free of snow, while wooded areas will tend to encourage greater catch because of diminished winds and the entrapment of snow from surrounding open areas. Drifting snow will be considered in more detail further on in the model description. Whenever snow is on the ground it will always be assumed to be the top layer even though some litter may fall on it.

In the normal course of events when cold air and snow occur during the same 24-hour period, the clouds and snow precede the arrival of cold air. In areas where large water bodies may modify the air mass, cold dry air would be warmed and moistened and could cause some snow. So long as the mean daily dew point is within four degrees of the mean daily temperature it is assumed that the snow blanket is added to the ground before the cold temperatures cause hard freezing of the ground.

If the dew point is very low it will be assumed that on this day freezing occurred prior to precipitation.

Estimating Soil Moisture

The soil moisture accounting system allows moisture to be collected by the soil or, as snow on the surface, or released as free water once the soil reaches field capacity. Generally, the ground water table is assumed to be below the active root zone and therefore not to affect infiltration.

If precipitation brings the soil moisture to above field capacity, the excess must eventually run off or go into groundwater recharge. No effort is made in this model to partition the portion going to runoff from that going to recharge, except that when the ground is concretely frozen then all precipitation is assumed to go to runoff.

Moisture is assumed to evaporate or transpire at the maximum rate until a critical value is reached, at which time the loss proceeds at a reduced rate proportional to the fraction of the water remaining. When the reduced amount of evaporation has been determined, moisture is removed from the profile with the surface layers contributing more actively until their moisture supply is reduced.

Moisture loss from the soil below the surface is generally restricted to the process of transpiration. During periods when evaporation is significant, diffusion resulting from thermal gradients is generally in the downward direction. Only on the surface can the soil evaporate directly to become as dry as the air. Thus, moisture will be lost from a layer beneath the surface only by plant transpiration. When the wilting point is reached, plants can extract no more moisture.

The maximum available water in a soil can be assumed to be the difference between the moisture present when the tension in the soil is 0.3 bars and the moisture present when the tension is 15 bars. This value can be determined in each layer of the soil and summed from the soil surface down to the bottom of the root zone. If the volume fraction of moisture is known for each of these tensions, then the moisture present in inches of depth can be computed by multiplying the volume fraction of moisture times the thickness of the layer.

This model is essentially one dimensional in that horizontal movement of soil moisture by concentration gradients or other means are neglected.

Field capacity is assumed to be a fixed value, but in order to consider the increase in retention with cooler temperatures, the field capacity is increased by 12 percent when mean daily air temperatures are at or below freezing. This adjustment corresponds with the observations of Peck (1973) and Klock (1972). When the ground has frozen concretely to some depth and thawing sets in but does not penetrate completely through the frozen layer, the field capacity of the upper layer is allowed to triple, which roughly allows soil moisture to reach values as high as those observed by Post and Dreibelbis (1942). As soon as mean temperatures rise to above freezing for 5 consecutive days or the ground thaws completely, all excess is lost to groundwater recharge or runoff. The figure of 5 days is based on the observations of Baker (1972).

Evaporation

Evaporation is proportional to the incident solar radiation, the humidity deficit, and the wind (equation 31). The humidity deficit is determined by taking the difference between the vapor pressure of water at the air temperature and at the dew point. The mean air temperature is taken as the average of the maximum and minimum temperatures. Point measured temperature data are adjusted by a coefficient to represent the mean temperature of the area of interest.

The amount of solar radiation absorbed and used in the evaporation process is reduced by some value of albedo. One average value is used during the summer, a different average value during the winter when there is no snow, and still a third average value is used when snow is on the ground. These values are determined by measurement or selected from tables as those best representing the cover type present. During the period of transition between summer and winter and winter and summer, there will be a linear interpolation from one albedo to the other over a period of time which comes from phenological estimates of periods of leafing out, or leaf fall, for the species and climate.

During this same period transpiration is assumed to be starting up or tapering off depending on whether the growing season is beginning or ending, and a transition from only surface evaporation to evapotranspiration from throughout the full root zone is computed by a linear interpolation similar to that for the albedo.

The evapotranspiration loss equals the potential evaporation corrected for seasonal values of albedo and solar geometry so long as the available soil moisture is above some critical value. To determine this value where data are available, a tension-soil moisture curve is drawn similar to Figure 1 for each layer. On each curve is found a break point or point at which the soil tension shows a significant rate of increase for further reduction in soil moisture. The available moisture in each layer at this critical point in the tension curve is summed and averaged for the whole profile to determine the profile value (see Figure 5). Where data are not available, curves from similar type soils may be used to estimate the critical moisture below which reduced evaporation takes place. This reduced evaporation is taken as the fraction of the total potential available water remaining times the potential evaporation. It may be, however, that the breakpoint comes at very low tensions. It is assumed that plants can transpire at near the maximum rate until the tension is at least 3 bars.

Distribution of Soil Moisture within the Soil Profile

Once the actual evapotranspiration loss for a day has been determined, the loss is partitioned to each layer on a weighted basis. Suppose that when partitioning is begun all layers are under equal tension. This is a condition that would be approached in spring when each layer is at field capacity. Losses are distributed inversely with depth, to correspond with the findings reported by Zahner (1965).

Consider a sample of soil having six layers above the bottom of the root zone, as shown in the schematic.

The moisture content of each layer is indicated by the X axis and the depth by the Y axis. The moisture held more tightly than 15 bars (wilting point) is indicated by hatched lines. A line marks the moisture at 0.3 bar (field capacity). From Figure 1 it can be seen that moisture held at tensions less than 0.6 bars is likely to be released almost entirely at the maximum rate, while moisture held more tightly is released at some reduced rate according to a function of the remaining soil moisture.

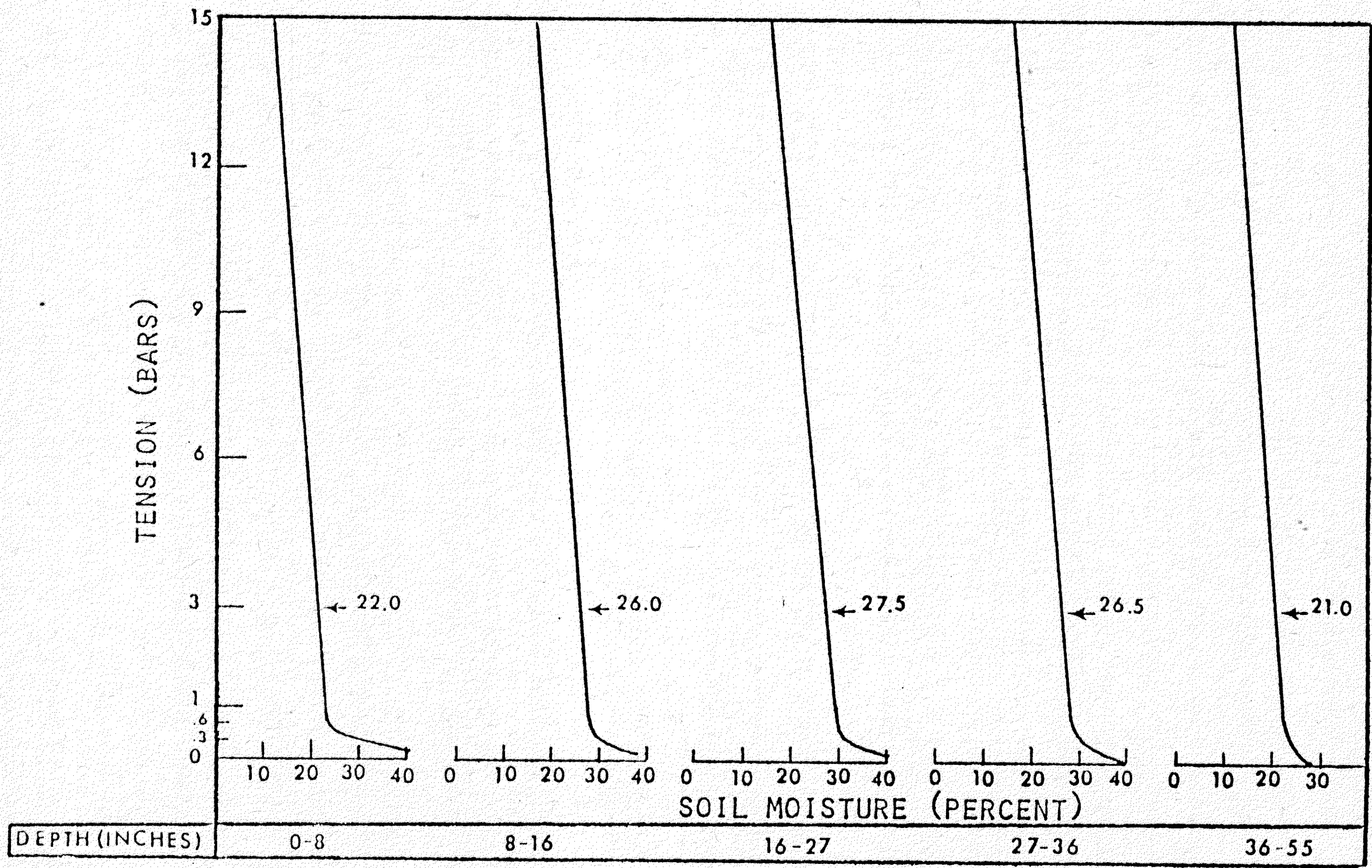

Figure 5.--Composite soil moisture-tension curve for five soil layers for Wellston silt loam. Critical percent is given as the percent at three bars of tension. Note how the bottom layer has a much shorter tail for high moisture content than the surface layer.

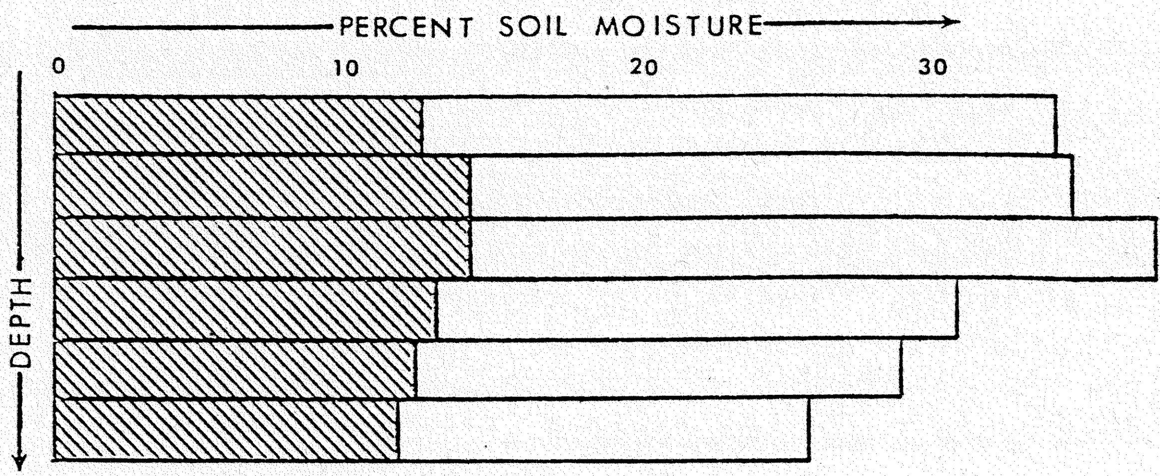

Figure 6.--Soil layer schematic.

To distribute the loss in this particular example, let the surface layer lose six times what the bottom layer might lose. Let the second from the top lose five times the bottom, etc. Then the total loss at each layer would be as in Table 2.

TABLE 2

EXAMPLE OF SOIL LOSS FROM LAYERS IN A SOIL PROFILE

Layers numbered from the surface down	Fraction of loss
1	$6/(1+2+3+4+5+6)=6/21=2x3/21=2x\frac{1}{(N+1)}$
2	5/21
3	4/21
4	3/21
5	2/21
6	1/21

Since the surface is the active region of replenishment, there is much opportunity for soil moisture to be higher than field capacity for certain periods, for instance on days with continuing rain. The surface layer is also subject to both evaporation and transpiration. For this reason, let the surface lose to direct evaporation an amount equal to that lost by transpiration (Buckman and Brady, 1969). Thus, the functional loss for the upper layer would be $2N/(1+2+...+N+N)=4N/(N^2+3N)$ rather than $2/(N+1)$, as shown in Table 2, where N is the number of layers. (Surface evaporation loss in wooded areas could be much less.)

As the moisture is lost the tension in the upper layers increases and the moisture loss increasingly comes from lower layers. The weighting factor governing this will be a ratio of

$$\frac{\text{current available moisture in the layer}}{\text{total available moisture capacity for layer}} \quad .$$

The loss from all the layers must still equal the total amount to be lost. Defining F_i as the constant fraction each layer would lose by reason of its depth, r_i as the fraction of the total available water left in the layer, twl as the total water loss for the period, and x as the unit loss then

$$\sum_{i=1}^{n} F_i r_i x = twl \tag{45}$$

and

$$x = twl / \sum_{i=1}^{n} F_i r_i \quad . \tag{46}$$

x is recomputed each day for the changing twl and r_i.

Normally, in nature, layers are not of equal thickness, nor even regular multiples of each other. Because of the irregularity of each layer, their depths are used only to the nearest inch. It is then assumed that each inch contributes according to some inverse function of depth. For example, should a profile consist of a 5-inch layer above a 2-inch layer, the 5-inch layer would have an F_1 of

$$F_1 = 14/35+6/35+5/35+4/35+3/35 = 32/35 \tag{47}$$

and the bottom layer F_2

$$F_2 = 2/35+1/35 = 3/35 \quad . \tag{48}$$

The variation in layer thickness thus being taken into consideration, as well as the extra potential for loss from the surface inch.

Recharge

Recharge is assumed to enter only at the surface. When moisture enters the soil the maximum gradient that can be maintained in the soil moisture content between layers for a period of 24 hours would depend on many things, such as the permeability of the soil, the percent moisture in adjacent layers, the thermal and chemical gradients, etc. An analysis of data for Coshocton for 1962 from Climatological Data (U. S. Weather Bureau, 1962) showed that when the ratio of available water present in a layer to the maximum possible available water in a layer was computed, the differences between layers in this ratio rarely exceeded 0.4 (see Figures 7 and 8). Testing of Coshocton soil moisture data for 1963 showed that good fit was given when 20 was the maximum difference in percent available water that was allowed between two adjacent layers. Based upon this information, moisture is passed down layer by layer using the following assumptions:

1. No layer can hold more than its currently assessed field capacity. (It may be remembered that the field capacity of the surface layer is adjusted for temperature and frost effects.)

2. The ratio of current layer moisture to layer potential available water cannot exceed that of the next lower layer by 0.2

The question of which process, recharge or evaporation, to apply first is important because if rain is entered before evaporation, there is a chance that field capacity could be reached and some water be found excess and assigned to runoff. Then when the evaporation loss is taken from the soil, the net result is somewhat lower than if the evaporation loss were applied before the recharge. Using only daily means and totals there is no way to avoid this error. Therefore, it was decided to apply only the net effect of precipitation minus the loss. This approximation treats recharge and evapotranspiration as if they were identical but reverse processes and took place at the same time, which is not true, but they cannot be differentiated with the available information.

Snow Accumulation and Loss

Snow has an important place in the model. As snow, water can reside on the surface without being lost to evaporation in significant amounts, changing the soil moisture distribution, or being lost to surface flow or groundwater recharge. Snow makes a significant change in the rate of heat exchange between the soil and the air.

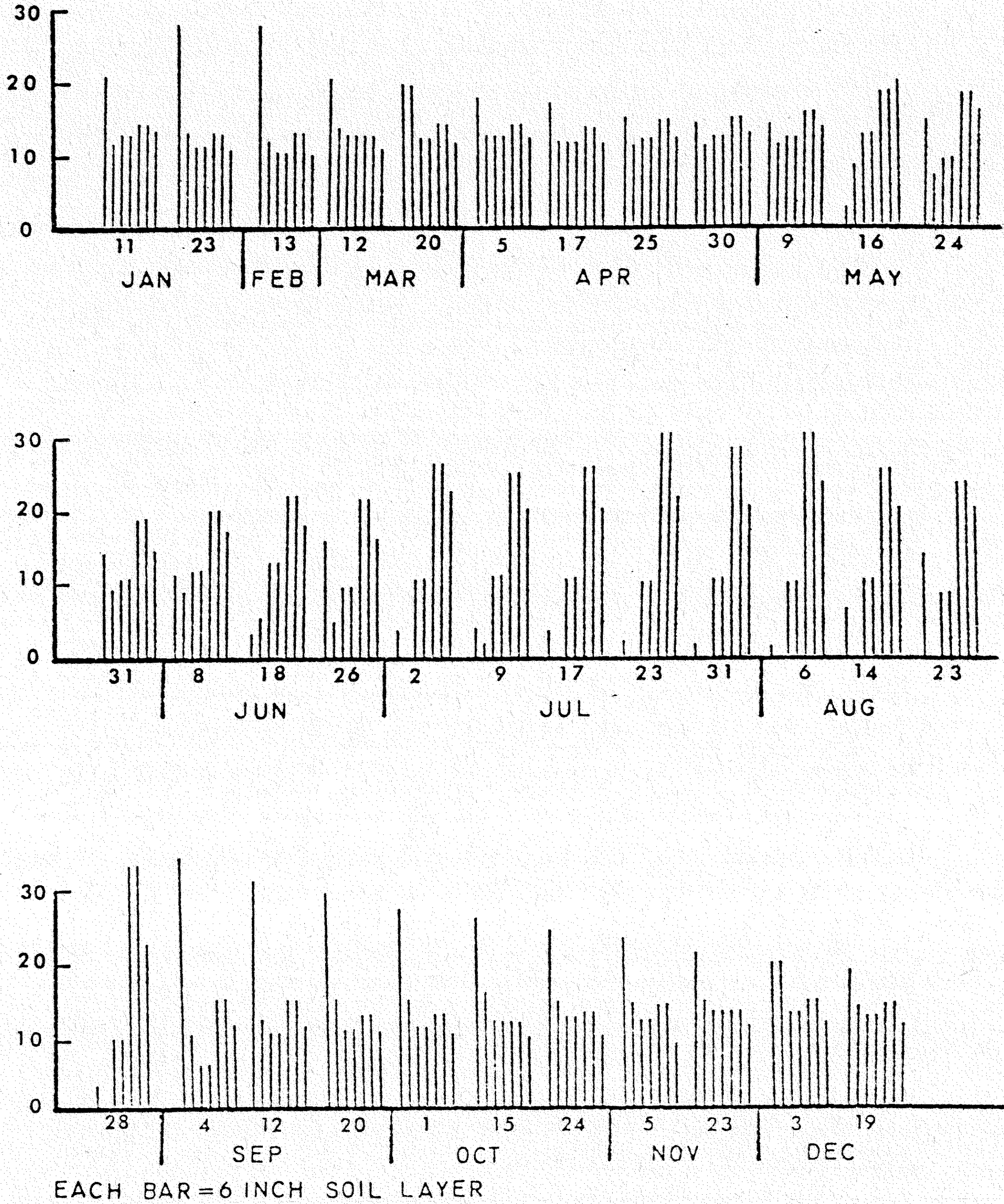

Figure 7.--Percent of available water found in each 6-inch layer at Coshocton, Ohio, in 1962.

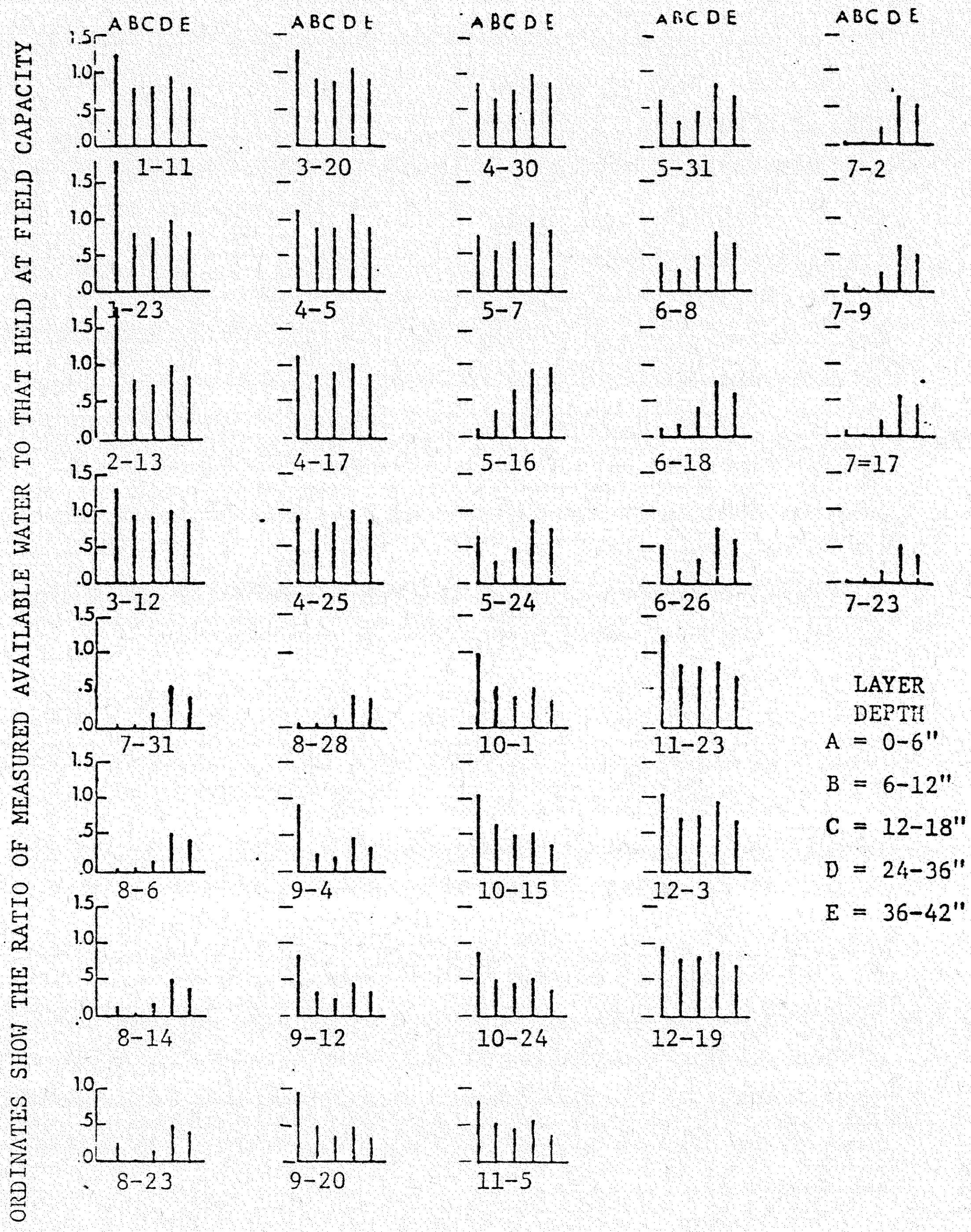

Figure 8.--Ratio of plant available water in each layer to that held in that layer at field capacity. The source data are the same as that shown in figure 2. Figures were derived from figures given in Climatological Data which gave one figure for soil moisture in the depths from 12-24 in. and 24-36 in. These increments were divided in half and the moisture placed equally in the upper and lower halves to give an indication of the moisture in each.

Snow Versus Rain

It is sometimes difficult to differentiate between precipitation that is snow and will remain on the ground and rain that will infiltrate or run off. This is especially difficult for a storm which changes from rain to snow.

With data limited to: (1) daily observed snowfall at a point and (2) extremes in air temperature, it was decided to declare any precipitation falling to be rain when the mean air temperature was 1°C or above and snow when the mean temperature was below this value. The observed snowfall is then used to verify or correct the division of total precipitation into solid and liquid precipitation.

At Coshocton, Mease (1970) determined a discriminant temperature for each month to separate rain and snow. It was not felt, however, that for this model the increase in accuracy justified the increased complexity.

Once the amount of precipitation in inches of water is partioned as snow, the following factors must be considered.

Snow Density and Depth

Density is very difficult to model with the data available. Because of this three values were used to approximate the snow as it settled. When all the precipitation is snow, the ratio of observed snowfall to observed precipitation can be computed as the density. Otherwise, the density is assumed to be 0.1 on the day of snowfall and increases to 0.15, and finally 0.2 with the passage of time. Should temperatures be very cold no settling occurs.

Were the pack to get deeper, a loading factor would be needed to consider the compression of the lower levels of snow as a result of the weight of the snow above. However, because shallow packs are the type of snow associated with soil frost, no loading correction was developed.

Snow depth is computed by dividing inches of precipitation declared to be snow by the density.

Drifting

Snow is often redistributed by the wind. It is logical to assume that snow will be picked up from regions where the wind blows the fastest and deposited in places where the wind is slowed down. For a uniform wind field several feet above the ground, it is assumed that the surface wind speed will be affected in such a manner that snow redistribution follows this equation

$$SNODPT = SNODPT-(WNDFACxSNODPTxK) \tag{49}$$

where SNODPT is the snow depth, K is a constant, and WNDFAC is a wind adjustment factor. The wind factors to be used here really depend on the relative roughness of the watershed under consideration to the roughness of the surrounding areas. For example, a forested area surrounded by a forested area would not be expected to gain or lose s snow to or from surrounding areas, whereas a cultivated field surrounded by wood lots might lose considerable snow to the forested land. On the other hand, relatively smooth areas intersected by ditches or ravines might lose much snow, with its associated insulating qualities, from the smooth areas to the ravines while losing little of the water equivalent of the snow from the watershed. Because of the almost infinite variations between watersheds, no constant functions can be assigned, but numbers would best be determined by calibration. Because data are very limited, it was decided to use a relative roughness factor. The initial value of the factor would be estimated using aerial photographs of the watershed and following examples by Kuz'min (1960).

When a watershed is covered by stubble, the capacity to catch and hold snow will be decreased as the snow builds up. This is accomplished in the model by the equation

$$WNDFCT = 0.1+0.0061\ SNODPT-0.00065\ SNODPT^2 \tag{50}$$

where WNDFCT represents a function for reducing the ability of stubble and vegetation to retain snow against the wind as the depth builds up, and SNODPT is the snow depth. These coefficients assume that should the snow depth reach 18 inches, the retentive effectiveness of the stubble to trap more snow would be reduced to zero. At 12 inches, the retentive effectiveness is reduced to 80 percent of its zero snow depth value. Should the effective stubble heights be found to be significantly different, these values would have to be changed. This function is only used in the program when the cover is coded as a stubble or low vegetation.

The K in equation (49) is strictly a constant of proportionality and should be determined from a calibration or by an experienced estimate of the total wind movement required to blow away a unit depth of the types of snow generally received in the area. Kuz'min (1963) found great variation in a relationship between the wind and snow removal, but it seems logical that 4 hours of a 25-mile per hour wind would remove at least an inch of loose snow from an area where an input of snow drifting in did not replace the snow blowing out. For a first approximation, a value of 0.01 does not seem unreasonable. The matter of snow drifting in as fast as it drifts out is taken care of by the relative roughness factor which indicates the net difference.

Snowmelt

Four factors are considered in melting snow. First, the sensible transfer of heat from the air which is proportional to the degree days. Anderson (1973) indicates that melt is more correlated with degree days if the coefficient relating the melt is varied as the length of the day. Using coefficients determined for the area around Rock Rapids, Iowa gives the equation

$$\mathrm{SNOMLT} = (T_{av}-T_{fr})(0.216+0.072\ \sin[\frac{2\pi}{365}(\text{day of the year}-80)] \qquad (51)$$

where T_{av} is the mean air temperature, T_{fr} is the freezing temperature, both in degrees Celsius, and SNOMLT is the amount of snow melted in inches of water equivalent. Because temperatures often swing above and below the freezing point, the mean temperature is sometimes slow to show melt that would naturally take place. For this reason, a daytime temperature is estimated as

$$T_{day} = (3 \times \text{max. air temperature} + \text{min. air temperature})/4 \qquad (52)$$

and the night time temperature as

$$T_{night} = (3 \times \text{min. air temperature} + \text{max. air temperature})/4 \qquad (53)$$

Snowmelt is divided between the day hours and the night hours as

$$\mathrm{SNOMLT}_{day} = (T_{day}-T_{fr})[0.108+0.036\ \sin\ (\mathrm{SAF})] \qquad (54)$$

and

$$\mathrm{SNOMLT}_{night} = (T_{night}-T_{fr})[0.108+0.036\ \sin\ (\mathrm{SAF})] \qquad (55)$$

where SAF is $\frac{2\pi}{365}$ (day of year-80). Negative values of snowmelt are set to zero. Daily values are then the sum of the two melts.

The second factor is the snow melted by rain. For every inch of water that falls as rain at a temperature T°C, there are 2.54xT calories of energy that can be released as rain water is cooled to 0°C. For every inch of meltwater produced from snow, approximately 2.54x80.0 calories of energy are absorbed. Thus the melt produced by rain is approximately

$$\text{RANMLT} = \text{PCP} \times \text{DYMNTC}/80.0 \tag{56}$$

where PCP is the amount of rain in inches, DYMNTC is the mean air temperature in °C, and RANMLT is the water in inches produced from melting snow. When rains occur in the winter it is very possible that temperatures could rise or drop just before, during, or after the rain. For this reason, the day was again broken into a warm and a cold period with temperatures T_{day} and T_{night} as determined in equations (52) and (53), respectively. The two melt values are summed to determine the daily total. Any negative values are set to zero before the summation.

When rain is occurring, the air temperature, dew point, and wet bulb temperature are all assumed to be very nearly the same. Falling rain is generally at the wet bulb temperature, and it will be assumed that the air temperature represents the temperature of the rain.

The third factor is the melt resulting from absorbtion of longwave energy from the clouds and atmosphere. When rain occurs on snow, it can be assumed that the clouds and atmosphere are near the air temperature and that the clouds and the snow have an emissivity near one. The net energy for snowmelt (RADMLT) is

$$\text{RADMLT} = \sigma(T^4_{air} - T^4_{snow}) \quad . \tag{57}$$

where σ is the Stephen-Boltzman constant. The slope of the T^4 curve can be approximated by a straight line in the region between 0°C and 25°C with less than a 5 percent error.

$$T^4 \approx 8.13 \times 10^7 \times T \tag{58}$$

where σ is 5.78×10^{-10} inches of melt day^{-1} °K^{-4}. When snow is melting,

T_{snow} is 0°C so that equation (57) can be written

$$RADMLT = 0.047 \times (T_{air}-T_{snow}) = 0.047 \times DYMNTC \qquad (59)$$

where DYMNTC is the daily mean temperature in °C.

The fourth factor is melt due to energy released by water vapor condensing on the snow. For each inch of water condensed out on the snow, enough energy is released to melt about 7.5 inches of water equivalent from the snow. Condensation is proportional to: (1) the excess water vapor pressure of the air over that of the snow and (2) the wind that creates the turbulence needed for mixing. The equation for melt resulting from condensation (CNDMLT) is

$$CNDMLT = wind \times (e_{sat}-0.18) \times 3.54 \times 10^{-4} \qquad (60)$$

where e_{sat} is the saturated water vapor pressure of the air and 0.18 is the water vapor pressure of the snow in inches of Hg. at 0°C (Wilson, 1941), and wind is the daily wind travel in miles day^{-1}. The wind travel is measured near the ground, such as the class A pan wind travel at 18 inches above the ground.

Because frost penetration is computed on each day of freezing as a result of cumulative values of the frost index and the number of days of freezing rather than as in incremental change in the frost depth, whenever the snow depth decreases, the conductivity of the effective layer is changed and the cumulative frost index must be reset to a value consistent with the reduced snow depth. The new effective frost index is found by solving equation (5) for F using the most recently computed value for the frost penetration.

Effective Litter Depth

When the cover type is mainly trees, the litter layer is assumed to remain fairly constant in its thickness, and thermal properties are changed only by the effect of soil moisture. If the cover is pasture or winter wheat, then an initial effective layer is assumed in the fall. The thickness of this layer needs to be based on a calibration but can be estimated based on grass height and density. Winter wheat in Coshocton is generally about 4 inches high during the winter season and seems to form an insulating layer about 2.5

inches thick. Under the weight of snow, grasses are pressed down and seem to remain so for the remainder of the dormant season. In the model the litter layer of grass in pasture or wheat was automatically reduced when the snow exceeded 2 inches in depth. When grasses freeze, many completely die down forming a dense thatch. When frost penetration reaches 1 inch in depth, the litter layer is reduced to the estimated thatch thickness. The effective depths of grass in its initial autumn state, when pressed down by snow and when matted after freezing, are all estimated and read as input variables to the model.

When snow falls into grass, the combined insulating effect is increased because of the dead air spaces formed by the lattice work of the grass. The conductivity of the litter is reduced to half its value by the model when the snow depth reaches the effective litter layer. It is increased by a linear interpolation as the snow is reduced to zero. When the snow is deeper than the effective litter layer, the conductivity of the litter layer remains constant at one-half the value of the litter alone. The figure of one-half was chosen to agree with the qualitative findings of Thorud and Anderson (1969).

Thawing and Freeze-Thaw Cycles

The same equation used for frost penetration is used for thawing. When seasonal freezing is computed it is assumed that the soil layer must be cooled from the mean annual temperature to freezing. When thaw occurs the soil layer need only be brought to 0°C with enough additional heat to maintain a thermal gradient. The primary energy requirement in this process is the latent heat. Thawing from beneath is also calculated based on the soil conductivity and the damping depth.

When a thaw sets in after freezing has occurred, the maximum computed depth of penetration is recorded. This maximum depth is subject to thawing from below by soil heat flow. Should thawing reach this point, the maximum depth of frost is set to zero.

Should a new cycle of freezing begin before the ground is completely thawed, both the maximum computed frost depth and the maximum computed thawing depth are recorded. If freezing in the new cycle reaches the computed depth of thawing, the depth of maximum thaw is set to zero, and penetration begins at the bottom of the total frozen layer. This discussion is illustrated by Figure 9.

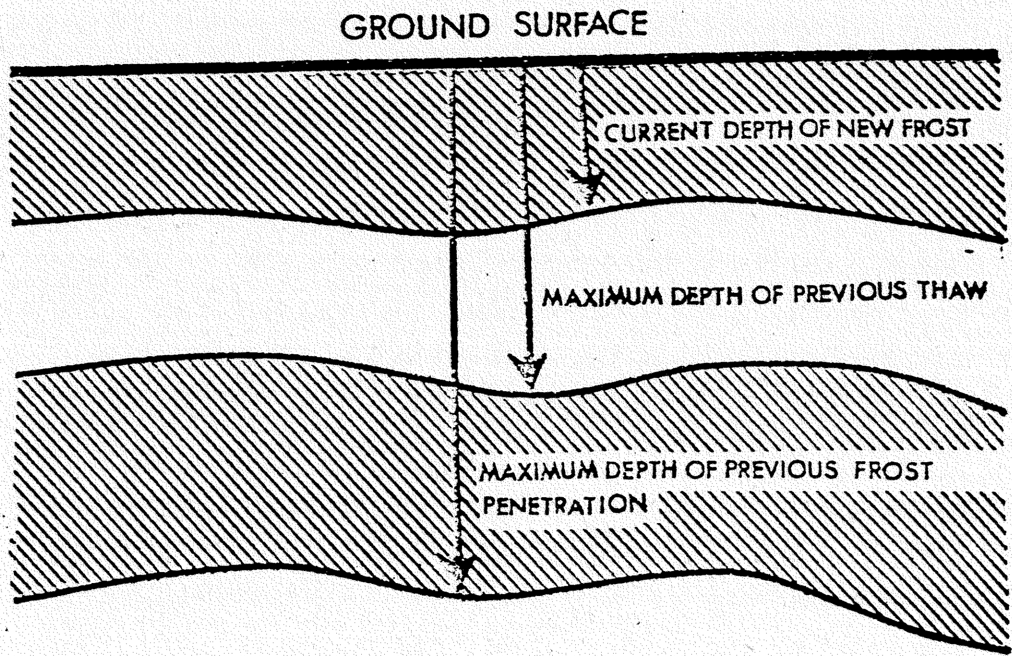

Figure 8. Illustration of frost-thaw depth bookeeping requirements.

Cover

Densities within cover types are assumed to be uniform. Soil and soil profiles are assumed to be similar within subareas. Crop or cover types used are those which predominate. Areas with significant variation in cover will be further broken into smaller subdivisions. The extent of variation that can be accepted before a reclassification is made would have to be subjectively determined based on the type of cover, topography, soil type, tendency for the ground to freeze, and the seriousness of the situation when it does.

Type of Frost Determination

The assumptions concerning conditions under which different types of frost form are:

1. If the ground is very moist or has thawed and the minimum temperature drops to at least 5°F below freezing, it is assumed that porous stalactite frost will form. By "very moist" is meant that

light snow has melted, the upper layer is at least at field capacity, or rain has preceded the freeze.

2. If the moisture conditions for stalactite frost are not present and the organic content of the soil is above some threshold value, it is assumed that granular or honeycomb frost will form. It is assumed that under any of the conditions given to this point the soil is still porous. If the rate of frost penetration exceeds an inch per day, it is assumed that the moisture is frozen in position into separate granules of ice leaving the soil porous.

3. If slow freezing occurs, organic content is too low, or freezing exceeds 3.25 inches into the soil, it is assumed that concrete frost has formed and that infiltration rates for any but forested areas are brought to near zero.

4. Should thawing occur above concretely frozen soil and then freezing reoccur, the ground will likely be saturated and concrete frost will form in the thawed region.

5. Finally, if the ground is initially warm, freezing temperatures must occur for 2 days in a row before sufficient ground will be frozen to materially change the basin infiltration capacity.

General Applicability

The area to which this model may be applied is determined by the input data. An area of any size that can be represented by a single slope, aspect, mean daily temperature, precipitation, and cover type can be used.

At the edge of cover types, variations in most of the variables used will increase. The analysis of this problem can be looked at in two phases.

If the cover type is over a fairly large area, such as a forest or large cultivated area, the boundary strip can be considered small in relation to the total area, and such boundary areas would tend to favor less frost because drifts form more in the edge of forests in the lee of trees. With deeper snow, frost penetration would not be as deep in these marginal areas and the errors that would result would be on the side of safety from a flood potential viewpoint.

If the watershed is quite diverse with bunches of trees spread throughout open meadows, the diversity becomes the rule and the cover would have to be estimated to follow a pattern in between the forest class and the meadow class. The diversity in itself would tend to cause irregular frost patterns and infiltration for a certain range of input intensities and would probably be sufficient in the unfrozen or porously frozen areas to compensate for areas that are imperviously frozen. Danger of high runoff would appear only as the total percent of frost covered ground neared 70 percent or more and then the location of the impervious and pervious areas would become very important.

The basic plan is to:

1. Determine and maintain the soil moisture distribution in known soil profiles.

2. Using the soil moisture estimate, determine the thermal conductivity, specific heat, and latent heat of each layer, including litter and snow layers, if present.

3. Using the thermal properties computed in step 2, compute the frost penetration.

4. When penetration occurs, determine the type of frost that will form and note whether formation of this type of frost leaves the soil permeable or impermeable.

5. When the soil thaws and refreezes, if there is sufficient time for excess moisture to drain away, then the frost type is determined by factors other than its past frost condition. If there is insufficient time to drain then it is assumed to freeze imperviously.

a. The quantity of water estimated to drain during a thaw is the smaller of the total excess water or $V = i \times t$; where V is the volume drained, i is the assumed infiltration rate, and t the duration of thaw.

b. The infiltration rate is taken from a warm weather infiltration capacity computed for the soil at field capacity. Because preparation (evacuation, dikes, etc.) costs are generally less than flood damage costs, the process is weighted to the worst case by multiplying this warm weather infiltration capacity by 0.25 to conform to the findings of Jensen et al. (1970) and Klock (1972). This term is called MNINFL.

c. The effective drain time is estimated as that fraction of time between the occurrance of the maximum and minimum temperature that the air temperature is above the freezing temperature T_f. The temperature is assumed to decrease linearly.

Expressed mathematically

$$\text{drain time} = \left(\frac{Tmax - T_f}{Tmax - Tmin}\right) 12 \text{ hours} \tag{61}$$

$$\text{or drain time} = \left[1 + \frac{Tmin - T_f}{2(Tave - Tmin)}\right] \times 12 \text{ hours} \quad . \tag{62}$$

Computer Model

In this section a brief description of the function of the main program and each of the subroutines will be given. Listings and a glossary of terms are included in Appendix C.

The main program handles reading in of the initial conditions, setting indices and sums to zero, listing headings, reading in daily values, making tests, and calling necessary subroutines.

LAYER uses soil profile information to compute the field capacity, available water, wilting point, threshold soil moisture, and saturation value for the total profile.

COVER uses the cover code that is read in at the start of the model and selects word descriptions of the cover type; i.e., PASTURE or HARDWOODS WITH LEAVES.

THNWTR computes the constants I and A which would be needed if data were not sufficient to use the Penman equation to compute evaporation. In this case, the Thornthwaite equation would have to be used (see Palmer and Havens, 1958).

EQNOXR computes for the days of the equinox the values of the following:

1. the angle of incidence of direct solar radiation,
2. the climatological average of the precipitable water (Kimball (Kimball, 1919).
3. the optical path length,
4. a scattering coefficient for this path length, and
5. an approximate attenuation for extraterrestrial radiation falling on a plane surface.

This computation is done for the equinox because the adjustment factor for the Thornthwaite equation is unity for this day (Palmer and Havens, 1958).

CALNDR computes the day of the year from the month and the day.

DALNCR

1. corrects the watershed day length for slope and aspect and the day of the year,

2. computes the mean incident angle for direct solar radiation on a horizontal plane and on a plane with the same slope as the watershed, and

3. computes necessary factors needed to correct the Thornthwaite evaporation equation if it is needed should solar radiation be unavailable.

EVPNR

1. checks to see if there is any snow on the ground,

2. adjusts the effect on wind in the watershed by the cover type,

3. adjusts for the shading of non-transpiring vegetation,

4. checks to see if the air temperature is above freezing,

5. checks the difference between the mean air temperature and the dew point temperature, and

6. if all factors are right for evaporation, it computes potential evaporation using equation (31).

EVPADJ tests to see if the amount of evapotranspiration loss is greater than the threshold value of the soil at which actual evaporation is reduced from potential evaporation. If the loss is greater, the program reduces evaporation to an amount proportional to the fraction of available water remaining in the profile.

The program also tests whether evapotranspiration is coming from the whole profile or only the surface layer. When evaporation is only from the surface layers, the program checks the threshold soil moisture for that portion of the profile and adjusts actual evaporation accordingly.

HAYCUT reduces evaporation by 20 percent on the first day of a harvest and by 40 percent on the second in accordance with the observations of Mustonen and McGuiness (1968). It then restores evaporation to normal by a linear interpolation during the following 28 days.

It also computes the fraction of available water in the part of the profile subject to direct evaporation and that in the part of the profile subject only to transpiration. It then computes the portion of evaporation to be lost by each region during the time when vegetation is coming out of or going into a dormant period.

TRNSLS takes the losses for the surface region and for the lower region which were computed in HAYCUT and partitions them among the various soil layers. It is only called during the spring and fall transition periods.

ESTEVP is called only when measured solar radiation is missing. It computes evaporation based on the Thornthwaite equation corrected for seasonal and climatic conditions.

SMP partitions the net loss or gain in soil moisture among the various layers of the soil. It is called whenever moisture goes into the ground except during transitions when TRNSLS performs the same function. It also readjusts the field capacity of the upper layer when cold temperatures increase surface tension and lower the hydraulic conductivity, or when rain or snowmelt occurs over concretely frozen soil.

FREZR tests for the freezing conditions and based on the result, records the number of days in the current freezing or thawing cycle, the maximum number of days in previous freezing or thawing cycles, computes thermal properties based on estimated conductivity values combined with estimated values of soil moisture, estimates the penetration of frost or thawing, and determines the type and permeability of the frost which forms.

The Computer Requirements

The model was written to run on either an IBM 360 or CDC 6600 and with minor modifications should run on any machine compatible with these.

Data Acquisition

The acquisition of data falls into roughly three categories: static, variable, and special.

Static Data

The static data are those input quantities relating to location, soil types, and normal vegetation. Many of these data can be obtained from readily available maps and tables. They are:

1. location: latitude and longitude;
2. topography: slope and aspect;

3. soil types and profiles;

4. moisture tension curves for given soil layers; and

5. local vegetation affected by climatic conditions such as: depth of roots, traditional crops in the area, general advent of leaf fall and onset of leafing out of trees, general organic content of untreated soils, and whether pasture grass is subject to revival in late summer after undergoing drought conditions.

<u>Variable Data</u>

The variable data are those values that change frequently and must be measured nearly continuously. Air temperature, precipitation, wind travel, solar radiation, dew points, and harvesting data are some of these. Most of the weather oriented values are available from a network of NWS stations. Harvesting data could be obtained from county agents or other agricultural services.

<u>Special Data</u>

Variables in the third category are things such as litter depth in the fall, estimated soil moisture as an initial condition when the model is begun, stubble, precise date of the onset of leaf fall or leafing out, and actual extent of cover type at the beginning of fall. These are variables which change slowly in nature in the fall but are subject to rapid change by man, climatic variations from normal, or unusual events such as fire.

This information is effectively determined and displayed by remote sensing techniques. If recent flights have been flown for a government agency, the photography will be available at very low cost, and in it the land use and farming practices will be fairly obvious. Recently multispectral data from the Earth Resources Technology Satellite have become available on an 18-day interval when the area is not hidden by clouds. Species have been recognized, and the onset of leaf out and of leaf fall have been seen (Dethier et al.). Corroboration of cold precipitation can be done with imagery from the NOAA satellites and the GOES satellite.

For the data listed in the first category, topographic maps and geologic maps are available for much of the country from the USGS. Soils maps for much of the country have been made by the Soil Conservation Service. Soil profiles for many experimental sites have been dug and data are available from the Agricultural Research Service of USDA. Many of the soils maps are made on aerial photographs which provide vegetation information as well.

Considerable data are also available from watersheds run by universities throughout the country.

CHAPTER FIVE. DEVELOPMENT, TESTS, AND RESULTS

To develop and test the model, data were obtained from the Agricultural Research Service Hydrologic Station near Coshocton, Ohio. This region was chosen because previous frost studies had been made there and the following data were available:

1. soil frost penetration and snow depth observations at four different sites each having a different cover type and generally a different soil type,

2. lysimeter data measuring the runoff, percolation, evapotranspiration, and precipitation,

3. runoff records from small watersheds,

4. pan evaporation,

5. soil moisture measurements at five different depths in the soil at frequent time intervals for 1962 and 1963, and

6. weather data sufficient to compute evaporation and frost penetration.

Model Input Data

Variable Data

Daily weather records as well as some climatological data were taken from published U.S. Weather Bureau (now National Weather Service) records (1958, 1959, 1962, 1963). Other data was obtained from the Coshocton station records. During earlier years, solar radiation and wind had been extrapolated from measuring points some distance away. The solar radiation was measured at Wooster, Ohio, and the wind at the Akron-Canton, Ohio, airport. Test runs at Coshocton using these data have compared well with lysimeter data (Mustonen and McGuiness, 1968).

Static Data

The Coshocton Hydrologic Research Station is located northeast of the City of Coshocton, Ohio, at 40°22' North and 81°30' West. The area is unglaciated and well disected with elevations varying from 700 to 1300 feet above sea level.

Maps of the area are available from the USGS in the 7.5 minute quadrangle series. The watershed is on the Coshocton and New Bedford map sheets, which have a contour interval of 20 feet and show the forested areas as of 1960. A map was made of the watershed in 1938 by the Soil Conservation Service. This map has a 5-foot contour interval and has the soil types, slopes, and erosion types classified. Buildings, wooded areas, and watershed instrumentation are shown.

The frost data that were collected came from numbered watersheds (see Figure 10) 109, 103, 131, and in the region of 134. Watersheds 131 and 134 were forested and rarely showed much frost penetration or runoff and so stand as examples of cover types that are not flood prone and hold little interest for frost penetration modeling. Watersheds 109 and 103 are the watersheds that are modeled.

Watershed 109 is a small watershed of 1.69 acres. Runoff occurs only during rainfall or snowmelt. The ground slopes to the southeast, with a slope of 13.5 percent or an angle of 7.72 degrees. The aspect is 135±45 degrees. Watershed 103 has an area of 0.65 acre and is at a slightly lower elevation than 109. It slopes to the west with a gradient around 12 percent or 7 degrees.

The general classification of soils in the area of the station was Muskingum silt loam in the upland areas and Keene silt loam on the more gently sloping lands. Both of these have been reclassified into several more refined classes. A fairly detailed description of the soils is given by Dreibelbis and Bender (1953) and the percolation rates are discussed by Harrold and Dreibelbis (1945). Both soils belong to the gray brown podzolic group of soils developed under deciduous forests in a humid temperate climate. The Keene soil is developed on clay shales and the Muskingum on sandstone, siltstone, and shale. The Muskingum is medium textured and well drained. The Keene is a moderately heavy silt loam and slowly permeable. When wet it swells and slows water movement.

The soil profile and tension data used for watershed 109 are for the Wellston silt loam which was formerly mapped as Muskingum silt loam. The soil pits used to get data for ARS publication 41/144 (Holtan et al., 1968) were dug within a few hundred feet of the watershed and at about the same elevation. The root zone is around 55 inches deep. There are five layers (A, B_{21}, B_{22}, B_3, and C) above the bottom of the root zone. The soil is well drained. The data for the profile representing watershed 109 are shown in Table 3.

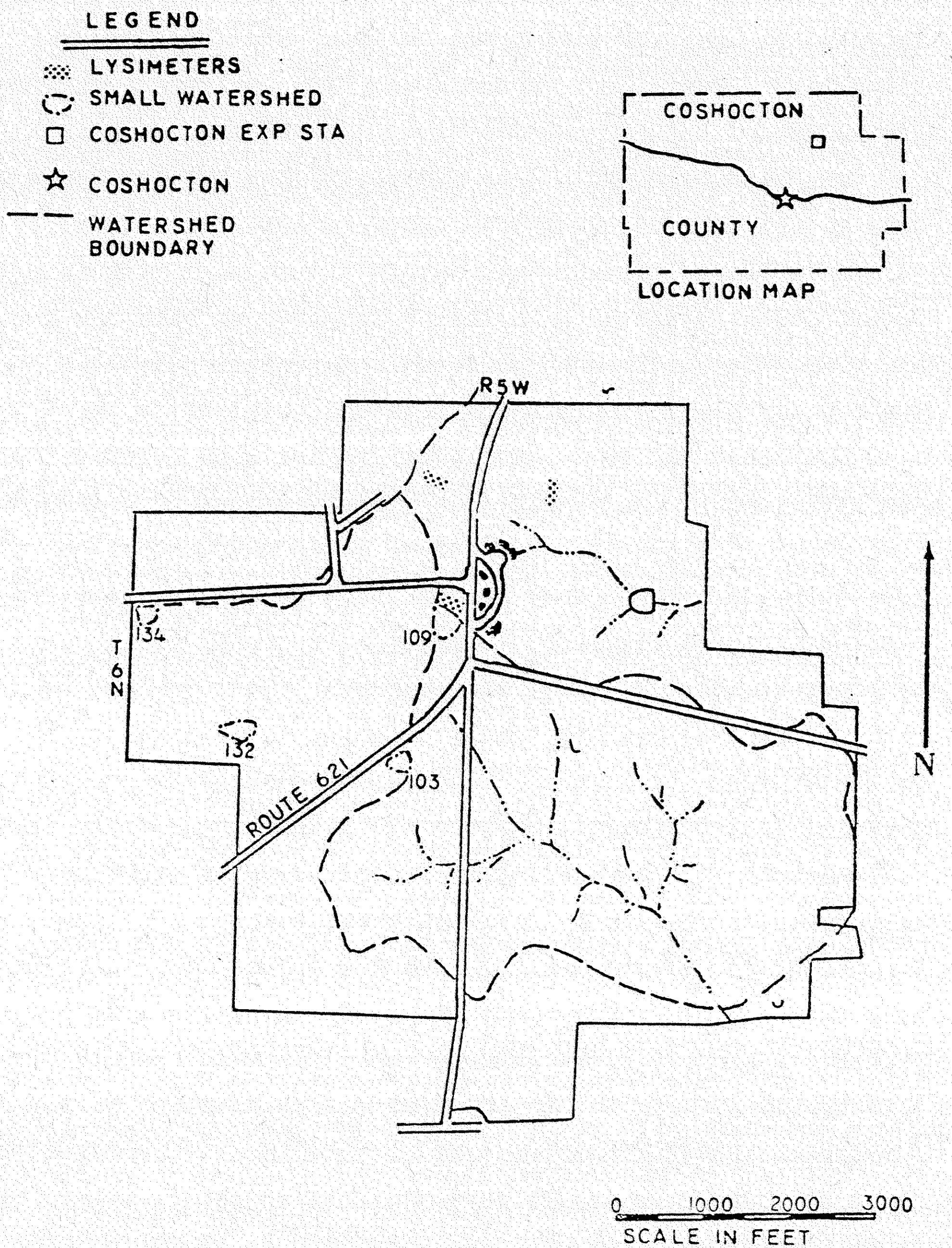

Figure 10.--Map of Coshocton hydrologic experiment station showing location of lysimeter batteries and other equipment used in obtaining hydrologic data.

TABLE 3

SOIL PROFILE DATA FOR WELLSTON SILT LOAM USED WITH WATERSHED 109

Depth (in.)	Thickness (in.)	Porosity (percent)	Wilting point (in. of water)	Field capacity (in. of water)	Critical moisture (percent)	Roots	Estimated organic content (percent)
0-8	8	46.06	1.00	2.80	22.0	Common	2.3
8-16	8	44.7	1.28	2.61	26.0	Common	1.8
16-27	11	41.9	1.64	3.92	27.5	Common	1.5
27-36	9	38.9	1.42	2.96	26.5	Common	1.3
36-55	19	44.3	2.15	4.77	21.0	Occasional	0.7

The soil on watershed 103 is Keene silt loam. The data for the profile is in Table 4.

TABLE 4

SOIL PROFILE DATA FOR KEENE SILT LOAM USED WITH WATERSHED 103

Depth (in.)	Thickness (in.)	Porosity (percent)	Wilting point (in. of water)	Field capacity (in. of water)	Critical moisture (percent)	Roots	Estimated organic content (percent)
0-9	9	49.1	0.78	3.53	25.90	Abundant	3.3
9-12	3	48.3	.52	1.01	23.52	Common	2.8
12-15	3	40.38	.67	1.14	22.85	Common	2.0
15-20	5	41.89	.97	1.905	23.88	Common	1.9
20-25	5	37.36	1.14	1.89	22.36	Common	1.8
25-39	14	33.96	3.28	4.82	18.62	Occasional	0.5

Special Data

The climate of the area is typical of the hardwood forest areas of central and east central United States. The rainfall is approximately 36-42 inches annually, with 50 percent falling between April and September. Approximately 2 inches of the precipitation, on the average, fall as snow. The average length of the growing season is around 160 days, with the last frost in early May and the first frost in early October.

Phenological data used in tests were taken from the records of Lamb (1914) and Smith (1914). Climatic values of water vapor and their effect on radiation were taken from Kimball (1919). The albedos used were from tables in Munn (1966). Litter was estimated based on descriptions of Post and Dreibelbis (1942). Litter is a very important variable and should be estimated as accurately as possible. Aerial photographs can help to improve estimates of litter.

Aerial photographs taken of this area in the summer of 1965 and in February of 1968 are available from the Agricultural Conservation and Stabilization Service. The scale of the photographs is approximately 1:12,000. The 1965 photographs are in a 15-inch format and the 1963 in a 9-inch format. For 1968, photographs with a scale of 1:6,000 are also available in a 9-inch format. In these photographs it is quite apparent which fields have standing stubble. No method was found, however, to determine the height of standing stubble or the thickness of the litter layer. The type of cover, insofar as the classes being used here, is easily discernible from the aerial photographs and in most cases from ERTS satellite imagery with the help of a map.

The average organic content of the soils cultivated under "improved conditions" in this area was around 2.34 percent. Both watersheds 103 and 109 fall into this category. The fields cultivated under the prevailing practices which tend to deplete the soil of organic material had around 2.23 percent (USDA, 1962). Organic material content shown in Tables 3 and 4 is assumed.

In 1955, the land use in the area was 40 percent cropland, 20 percent pasture, 25 percent forested, and 15 percent other (Ohio Forestry Association, 1955). The cover on the two watersheds tested varied through a corn, wheat, meadow, meadow sequence. The cover for a particular year is given in the USDA-ARS Miscellaneous Publication Series "Hydrologic Data for Experimental Agricultural Watersheds in the United States." The aerial photographs and ERTS satellite imagery mentioned earlier are also a good source of information on cover type.

Verification Data

In the course of arriving at a determination of a possible change in the permeability the model estimates the following: potential evaporation, actual evapotranspiration, soil moisture distribution, snow depth, frost occurrence and penetration, and, finally, changes in permeability which would not be expected were cold or frozen ground not a factor. Data for testing the accuracy of estimates comes from the following places:

1. Class A pan evaporation is published in Climatological Data (formerly published by U.S. Weather Bureau, now by the Environmental Data Service of NOAA). It has been found that potential evaporation is usually in the range of 0.7 times the pan evaporation (Lake Hefner Report, 1954). The observed pan data is not continuous. No observations were taken on weekends, but the weekday readings are sufficient to verify average potential evaporation values.

2. Actual evapotranspiration can be determined from the weighing lysimeter records which are available for 1948-1965 (Mustonen and McGuiness, 1968). Net changes in the upper 42 inches can be estimated from soil moisture measurements. Attention has been called to this data for 1962 and 1963 in the Annual Summaries of Climatological Data (U.S. Weather Bureau, 1962, 1963).

3. These soil moisture measurements (from item 2) also were used to verify the distribution of the soil moisture and the loss within the profile.

4. Snow depth and frost depth observations were made once a day during the frost season on the watersheds mentioned previously starting in the early 1940's and continuing through the 1950's. Frost depth is best determined from test pits but can be inferred from soil temperatures. Soil temperatures were measured near watershed 109 and the daily maximum and minimum values for depths of 1/2, 3, 6, 12, and 24 inches are published in Climatological Data.

5. Runoff data are published in "Hydrologic Data for Experimental Agricultural Watersheds in the United States," which includes monthly sums of precipitation and runoff for each watershed, and in the volume containing the data for earlier years of a given watershed, a large-scale map is given. Unusual runoff events are documented in more detail. These data were used to verify increased runoff during a condition of impervious frost.

In Appendix A, the precipitation and runoff and the ratio of runoff to precipitation for December, January, February, and March for the years 1956 through 1967 are shown. From these data, it can be seen that the periods of abnormally high runoff are January of

1959, January and February of 1962, and March of 1963. February of 1963 was very high on watershed 103, but there was no runoff measured on 109.

Soil Moisture Tests and Results

Soil Moisture Profiling

Estimations of frost penetration and frost type are dependent on a good estimate of the total soil moisture and its distribution in the soil profile. Based on the available data, it was decided to use the 1962 soil moisture data to develop the profiling process. It could then be tested using the 1963 soil moisture data.

The limitations in the data should be noted before the results are discussed. First, pan evaporation was limited to weekdays. Second, the available soil moisture data published in Climatological Data was measured for layers between 0-6, 6-12, 12-24, 24-36, and 36-42 inches rather than as a function of natural layers or texture changes. Finally, precipitation, one of the major factors, was for only one gage. During the summer when most of the precipitation occurs during thunderstorms, there is a great variability in the point to point catch.

To test the soil moisture partitioning procedure without introducing errors caused by estimating evaporation, the precipitation and available soil moisture data given in the listings in Climatological Data were used with a water loss term. The water loss term was calculated with the following equation.

$$\text{Waterloss} = \text{old soil moisture} - \text{new soil moisture} + \text{precipitation} \quad (62)$$

where old soil moisture is the previously observed value, new soil moisture is the currently observed value, and the precipitation is that recorded between soil moisture observations. This water loss term combines runoff, groundwater recharge, and evaporation in one term; however, since runoff or recharge is only expected when the soil moisture is above field capacity, loss will be primarily evaporation. Harrold and Dreibelbis (1945) had found that at Coshocton:

> "The smallest of the depletion factors is runoff which amounted to one percent or less of the total of all the depletion processes."

Percolation, they noted, ranged from 5 to 8 inches a year during a 2-year study and occurred mostly in the late winter or early spring. Percolation on the Muskingum soil is up to 50 percent greater than from the Keene soil.

Preliminary tests with the 1962 data indicated that the inverse relationship for water loss discussed in Chapter 4 seemed to draw too much water from some layers and not enough from others. An improved fit of the simulated to the observed data resulted when the inverse loss relationship was limited at some depth. In this study, 24 inches seemed best. It was therefore concluded that the profiling procedure would assume the following:

1. The surface inch is subject to both transpiration and evaporation.

2. When the layers are under equal tension water is extracted inversely proportional to depth down only to a certain depth. Below that point the loss in each unit thickness of soil is independent of depth A possible justification for this is that when soils have been depleted to the extent that most of the moisture left is in the lower layers, it is likely that greater root development will take place in those regions and that plants with shallower roots may have wilted leaving only those species with deeper roots.

A plot of observed and simulated data for 1962 is shown in Figure 11. Several differences can be seen between the simulated and observed data. During the middle of the summer too much moisture is taken from the third layer down while not enough is taken from the bottom layer. On certain days such as May 24 the simulation significantly underestimates the top layer and overestimates the third layer down. It will be noted however that going into the frost season all layers are well represented.

These data represent observation periods of from 3 days to over 2 weeks. Many storms of varying magnitude can occur during this time and, depending on the transpiration demand, affect certain layers more than others. With the observed data it may be seen that in certain periods, one layer is decreasing by a significant amount, while a different layer is increasing by a significant amount on the same date. A simplification made in this model allows only a decrease

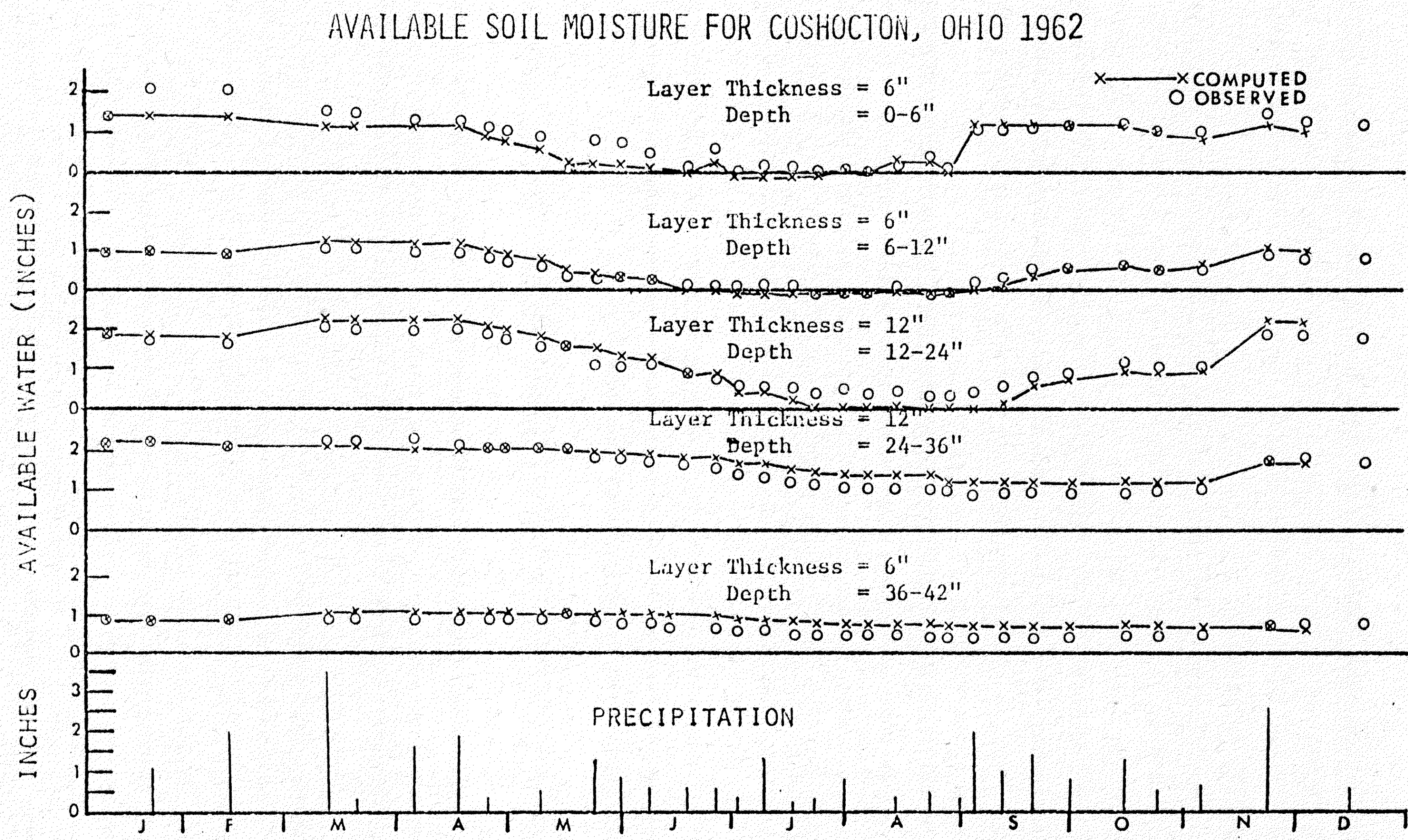

Figure 11.--Comparison of observed and computed soil moisture in five layers for Coshocton, Ohio, in 1962.

or an increase to be made on any given day. This should be adequate on a daily basis, but scatter about observed values is likely to become larger when the periods between observations are increased.

The model must approximate to some degree of accuracy the soil during dry years and wet years. There will be many times during the part of the year with greater evapotranspiration demand when one layer or another will not be simulated too closely. However, it is felt that the inverse extraction of soil moisture to a certain depth is adequate.

The model was run next on the 1963 available moisture data and the results are shown in Figure 12. The fit shows the same strengths and weaknesses shown with the 1962 data. It can again be seen that going into the frost season the soil moisture is represented very well throughout the profile.

Test of Generality

The "generality" of the model was tested with available moisture data from the Ohio State University Farm station at Columbus, Ohio, published in U.S. Weather Bureau records (1962, 1963, Figures 13 and 14). It might be noted that these were years of extremes. In 1963 the spring was extremely wet with much of the soil being above field capacity into late May. By late summer, however, many areas of the state were very dry. The model does very well with the 1962 data with all layers closely simulated going into the winter. In 1963, because the ground was above field capacity from frequent rains, the model was not even started until late May. The model does not allow soils to be above field capacity unless they are frozen. It seemed as though much of the soil was in a capillary fringe and being affected by a water table rather than draining to field capacity. During the rest of the year, the model did well for the upper 12 inches but underestimated by about a half inch the available water in the 12-24 inch level and overestimated by a lesser amount the moisture in the 24-26 inch level. These errors caused differences of up to 2 percent in the percent soil moisture by volume in the layer (there could be over 100 percent error in percent available moisture). The underestimation of soil moisture would indicate a lower heat capacity, latent heat, and a lower conductivity. The overall effect would be to calculate a deeper frost penetration than would be calculated were there no error. For example, Figure 15 is a plot of frost penetration

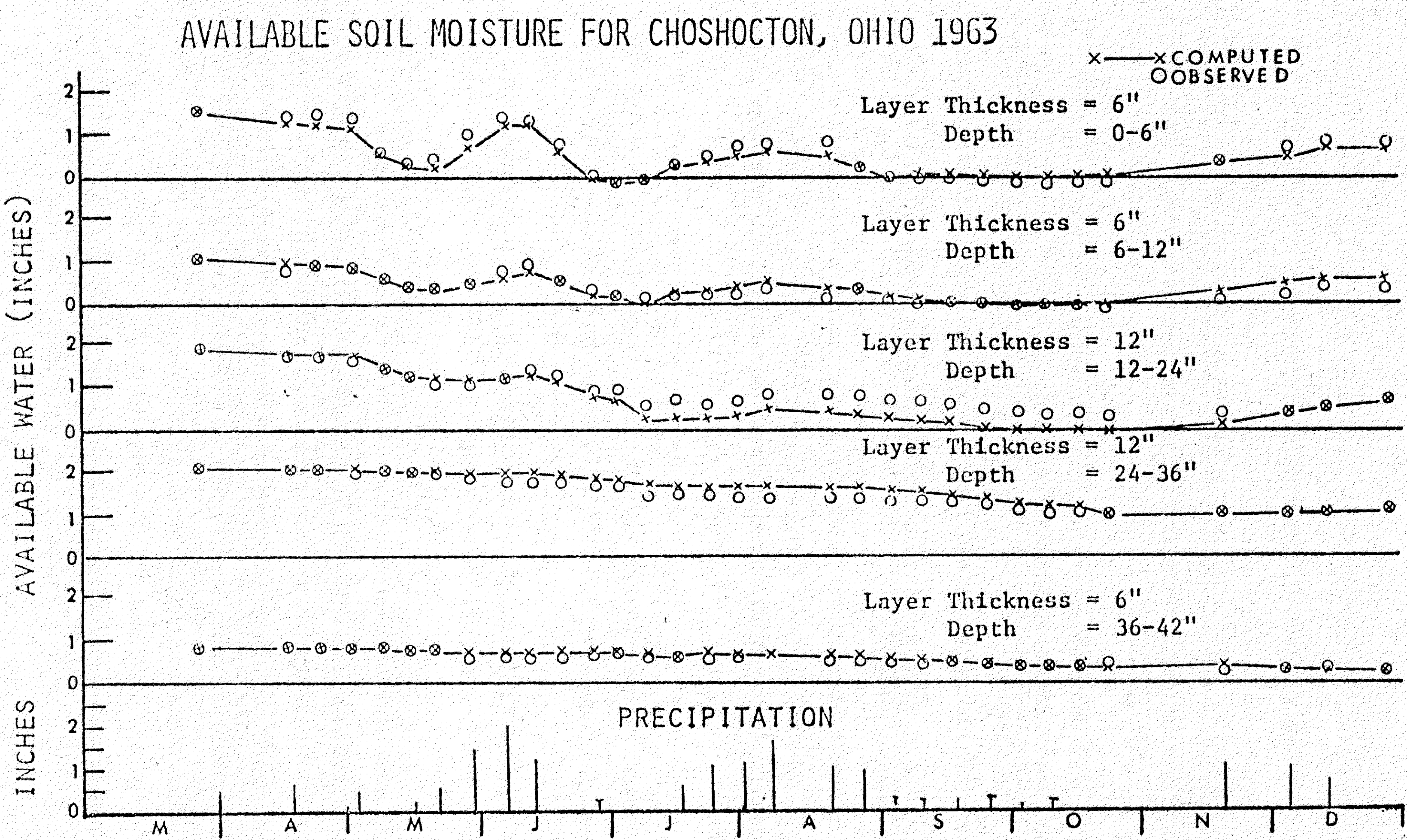

Figure 12.--Comparison of observed and computed soil moisture in five layers for Coshocton, Ohio, in 1963.

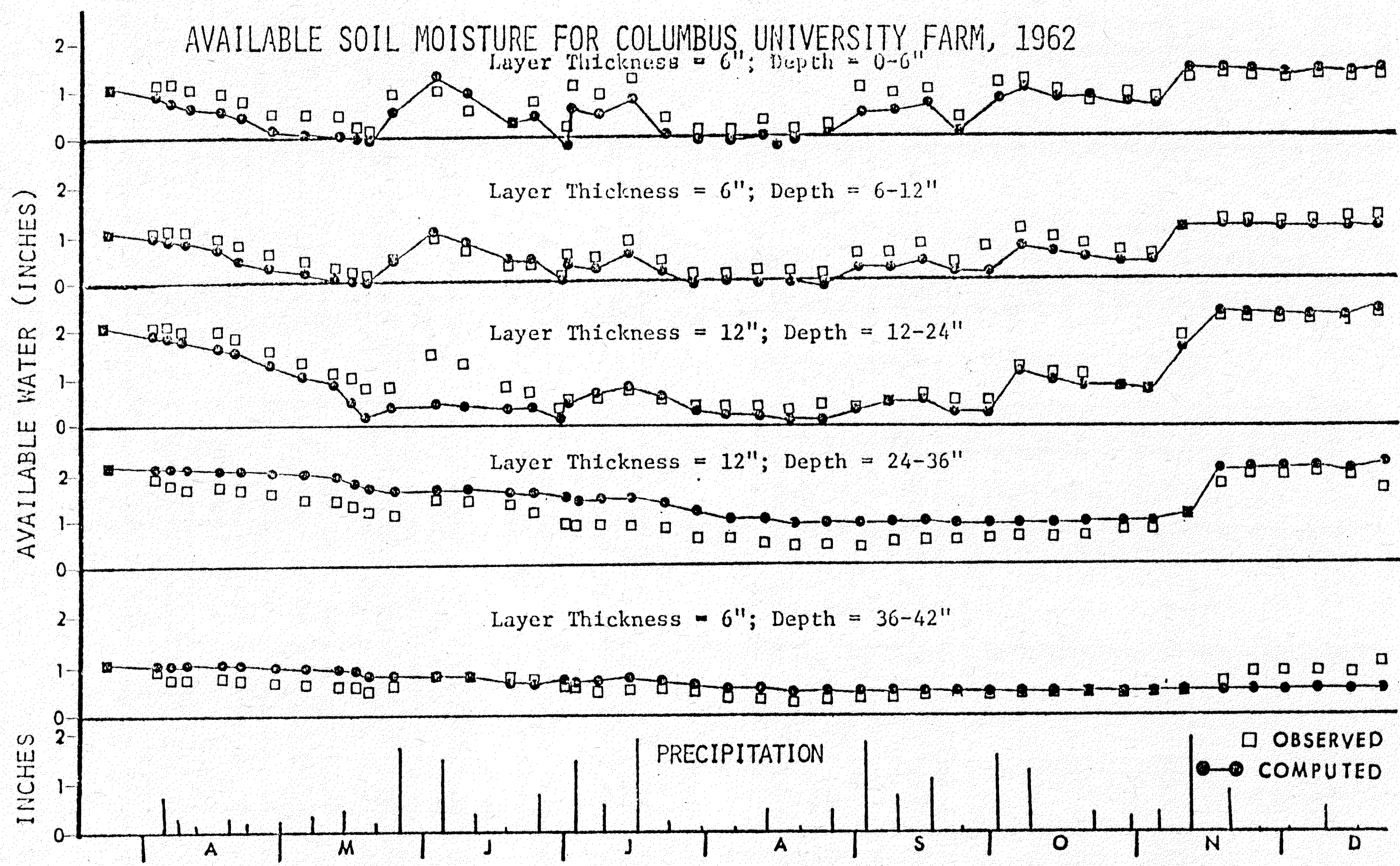

Figure 13.--Comparison of observed and computed soil moisture in five layers for Columbus University Farm, Columbus, Ohio, for 1962.

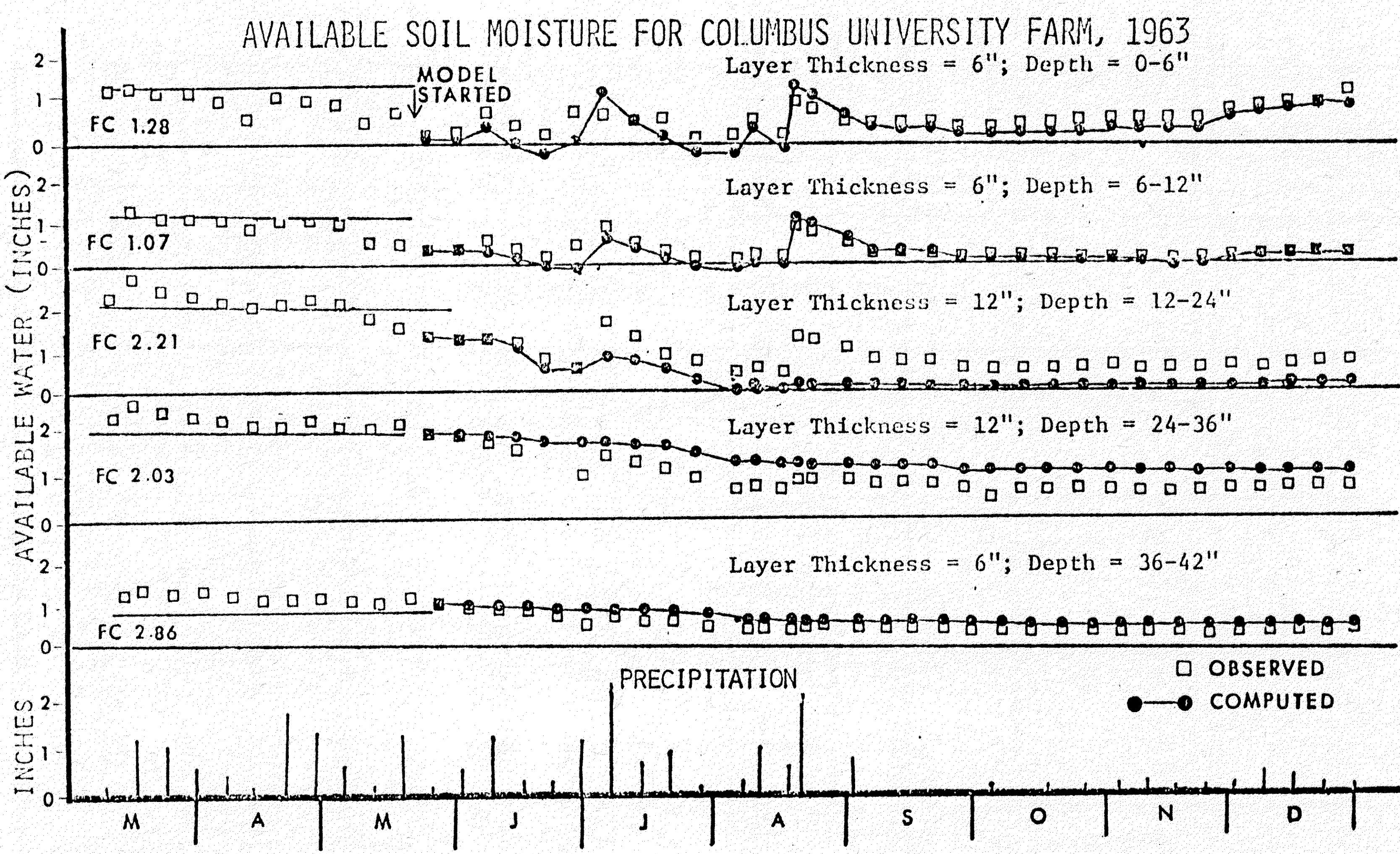

Figure 14.--Comparison of observed and computed soil moisture in five layers for Columbus University Farm, Columbus, Ohio, for 1963. Note how long the lower layers remained above field capacity during the spring.

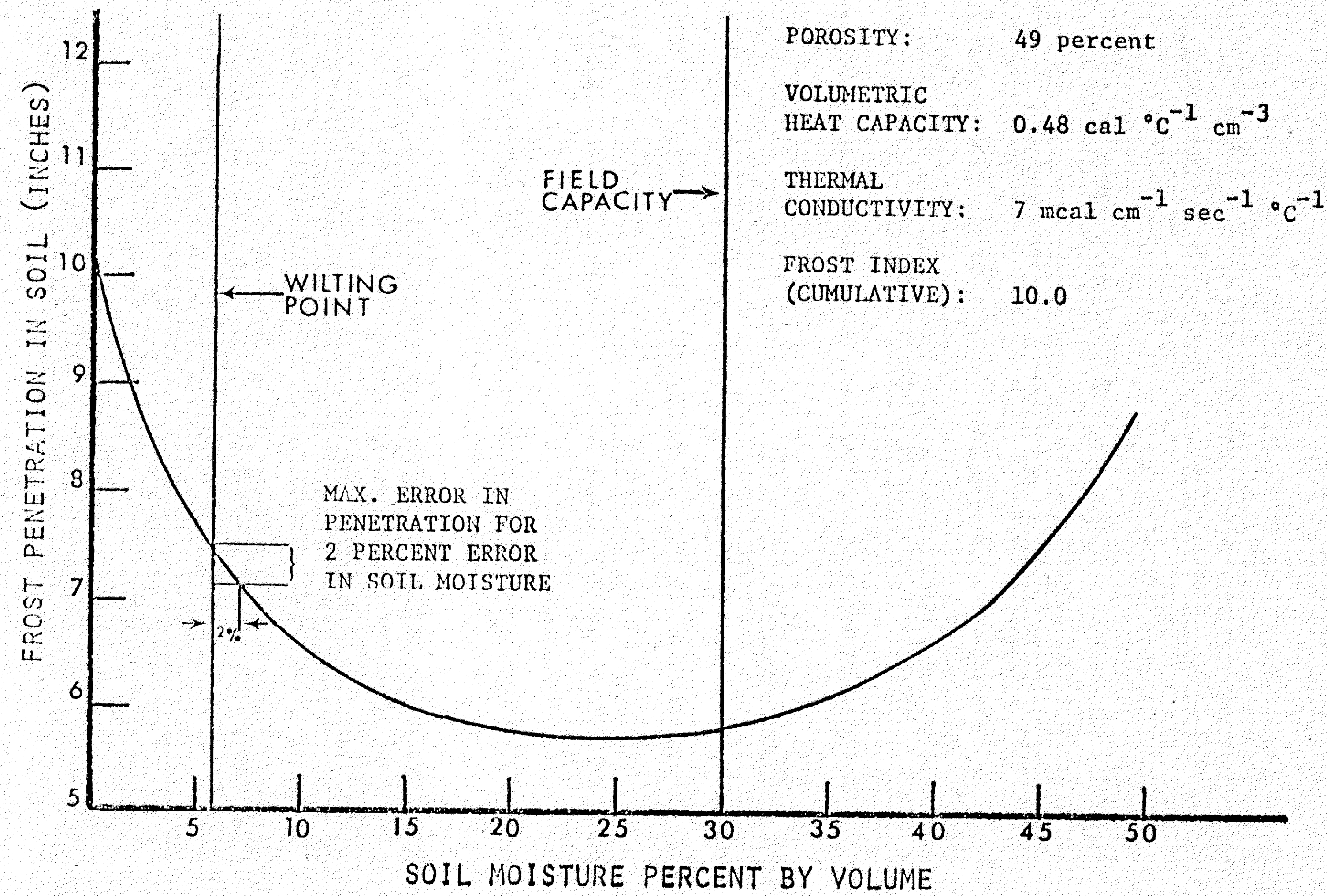

Figure 15.--Variation of frost penetration in the soil as a function of soil moisture. The curve is for a soil having the qualities and frost index as shown. Note how flat the penetration curve is in the soil moisture range immediately below field capacity.

as a function of soil moisture for a soil having a porosity of 49 percent, a volumetric heat capacity of the mineral in the soil of 0.48 cal $°C^{-1}$ cm^{-3}, a thermal conductivity of the mineral of 7 mcal cm^{-1} sec^{-1} $°C^{-1}$ and a cumulative frost index of $(T_{av}-2)°C = 10.0$. It can be seen that the frost penetration exhibits a minimum near 24 percent. For a soil with these thermal properties, it can also be seen that between the wilting point and field capacity an error of 2 percent in the soil moisture makes at most a difference of 0.4 inch in the frost penetration; and this maximum difference occurs on the very driest end of the soil moisture moisture range. This error is proportional to the square root of the product of the conductivity and the frost index accumulation (equation 5) and so would increase as the cold season progresses. However, for the percent soil moisture range between 20-30 percent the slope of the curve is fairly flat and a 2 percent error in soil moisture estimation would not cause significant error in estimating soil frost penetration.

The error in computing frost penetration caused by an error in soil moisture estimation would only occur were the penetration computation to reach that soil layer in which the inaccurate soil moisture was estimated. At Coshocton most observed depths were less than 6 inches and fell into regions where soil moisture was well estimated. In regions of very dry soils, penetration during cold temperature is likely to be very rapid - a condition favoring porous frost. An underestimation or an overestimation in this type region would not greatly change the infiltration estimate. Should a soil for some reason be more moist than field capacity, chances that an impervious frost would form might increase markedly. This condition is however outside the range of this model.

The surface regions which are the most critical seem to be adequately simulated by the model, given that the water budget is correctly computed.

Evapotranspiration

The overall soil moisture budget was mainly dependent on the estimate of evapotranspiration. Potential evapotranspiration was compared against the class A pan using a pan coefficient of 0.7. The differences between the computed values and the pan values were recorded, and Figure 16 shows a histogram of the data for 1958. The differences for the 1963 data

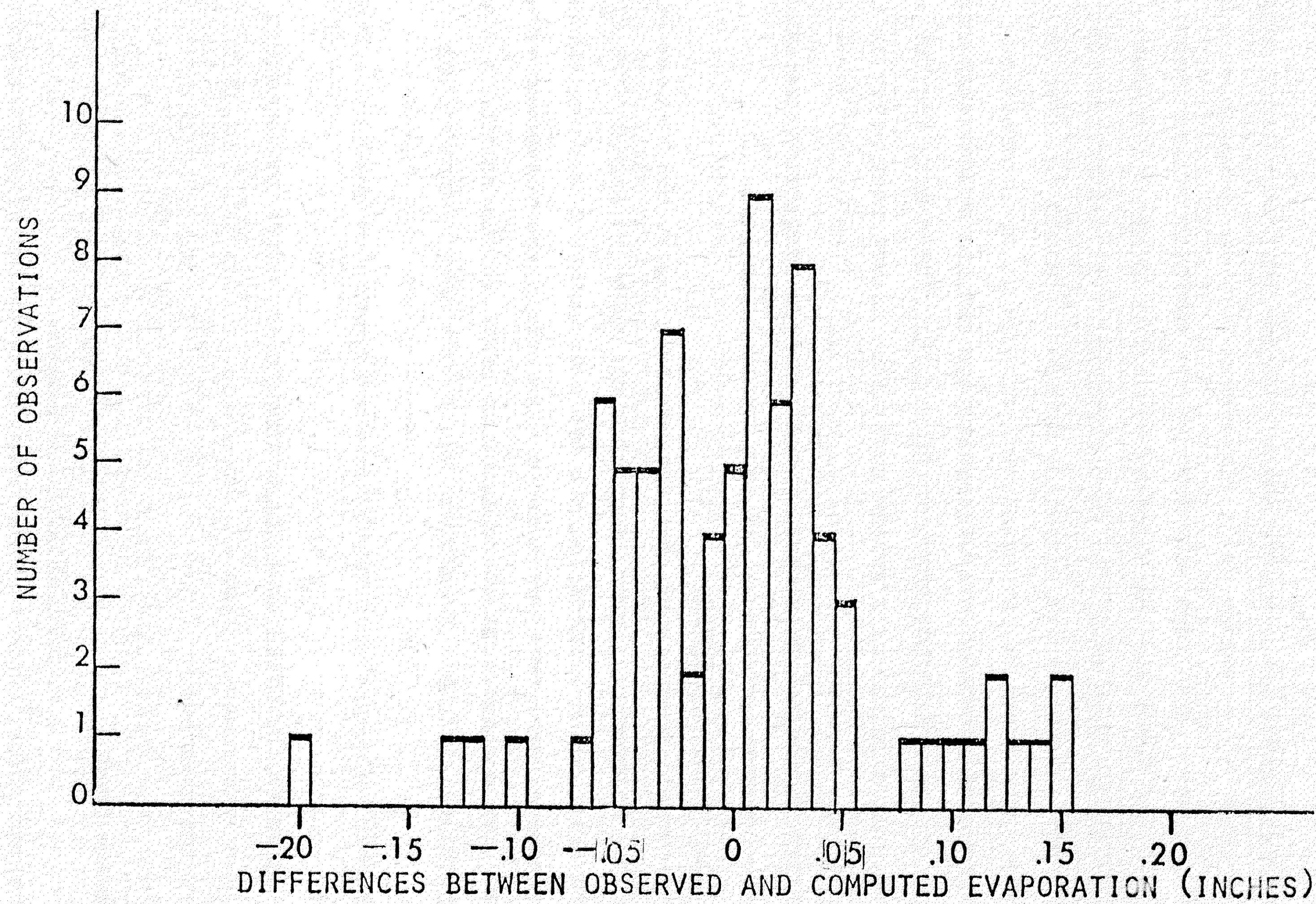

Figure 16.--Histogram of daily differences between computed potential evaporation and 0.73 times the observed class A pan evaporation for the 1962 evaporation season.

are shown in Figure 17. In 1963 there were 113 days on which a comparison could be made. The average daily difference between computation and observation was 0.0044 inch, the mean of the absolute values (mean deviation) was 0.055 and the standard deviation of the differences was 0.074 inch. This indicates good agreement over longer periods such as weeks or months but a scatter around 0.05 inch on a day by day basis. The level of agreement is indicated in the comparison of the computed potential evapotranspiration and the evapotranspiration measured by the lysimeters shown in Figure 18. These data are for 1958 and were affected by a grease seal which caused some error in daily values and was modified because of that in 1962. The values for longer periods are considered good, and in spite of the variability in the data, the agreement of day by day data seems quite good.

Total Soil Moisture Estimate

The overall accuracy of the soil moisture budget can be seen in Figure 19. Agreement is good except for the month of August when the model underestimated evaporation and the soil moisture was overestimated until October. An effort was made to find out if the disagreement resulted from a failure of the theory, an error in the data or a mistake in computations. It was found that there were many storms during August. It is likely that a 24-hour period is too long during a thunderstorm season to give really good agreement. A cumulative double mass plot of the computed and observed losses is shown in Figure 20. It may be noted, however, that the spring period tends to deviate as much as late summer. Even so, the computed seasonal loss is very close to the observed.

These data suggest that the model adequately fulfills its evaporation need.

Snow Depth Verification

The temporary nature of snow makes verification of average areal estimates difficult. The computed depth results from consideration of observed snowfall, drifting, settling, and melting. Adjustment is made in one increment each day. This adjustment uses mean temperatures, and cumulative winds.

Snow depth was observed once each day at watersheds 109 and 103. The observed snowfall for 1958-59 data was measured once each day

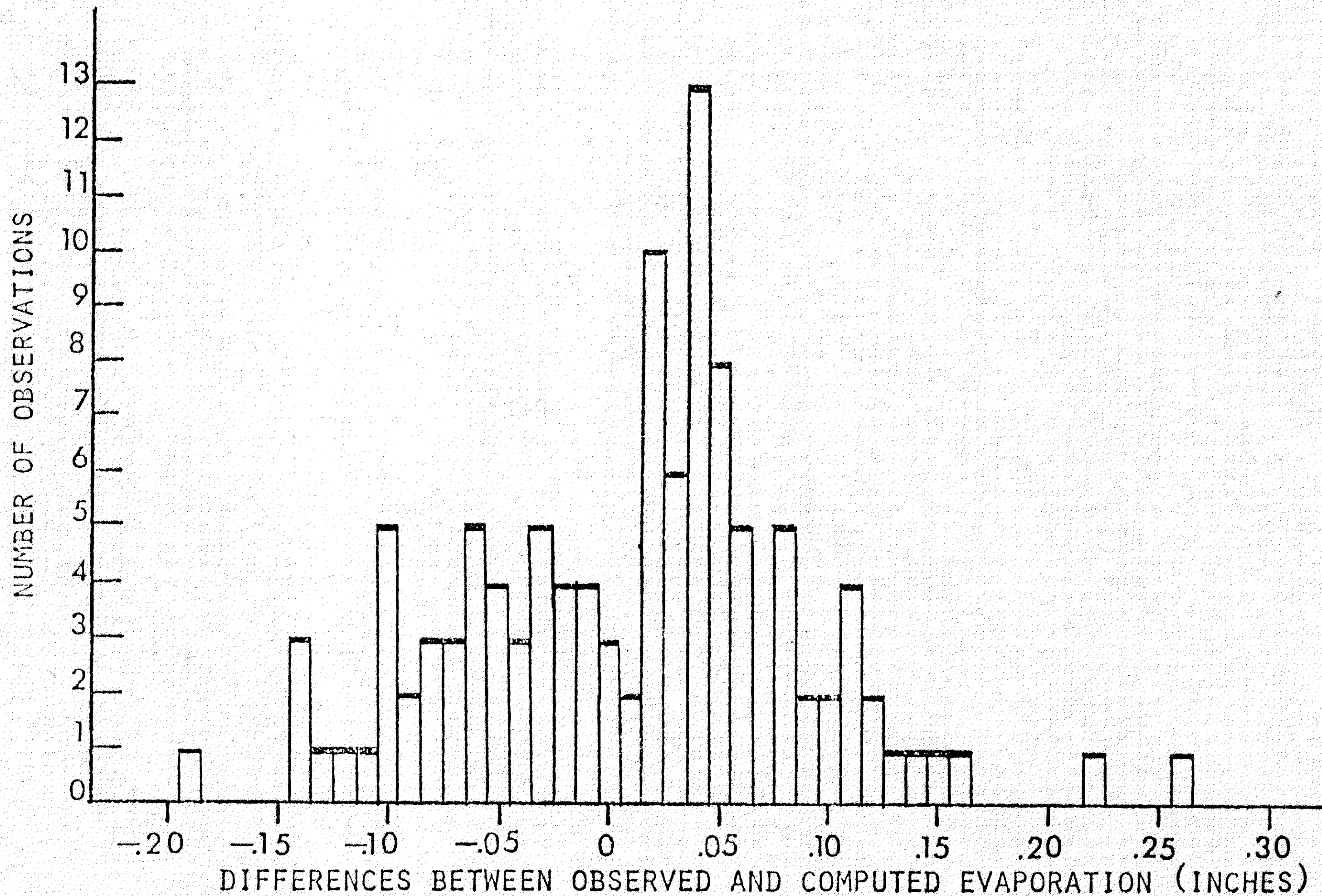

Figure 17.--Histogram of daily differences between computed potential evaporation and 0.73 times the observed class A pan evaporation for the 1963 evaporation season.

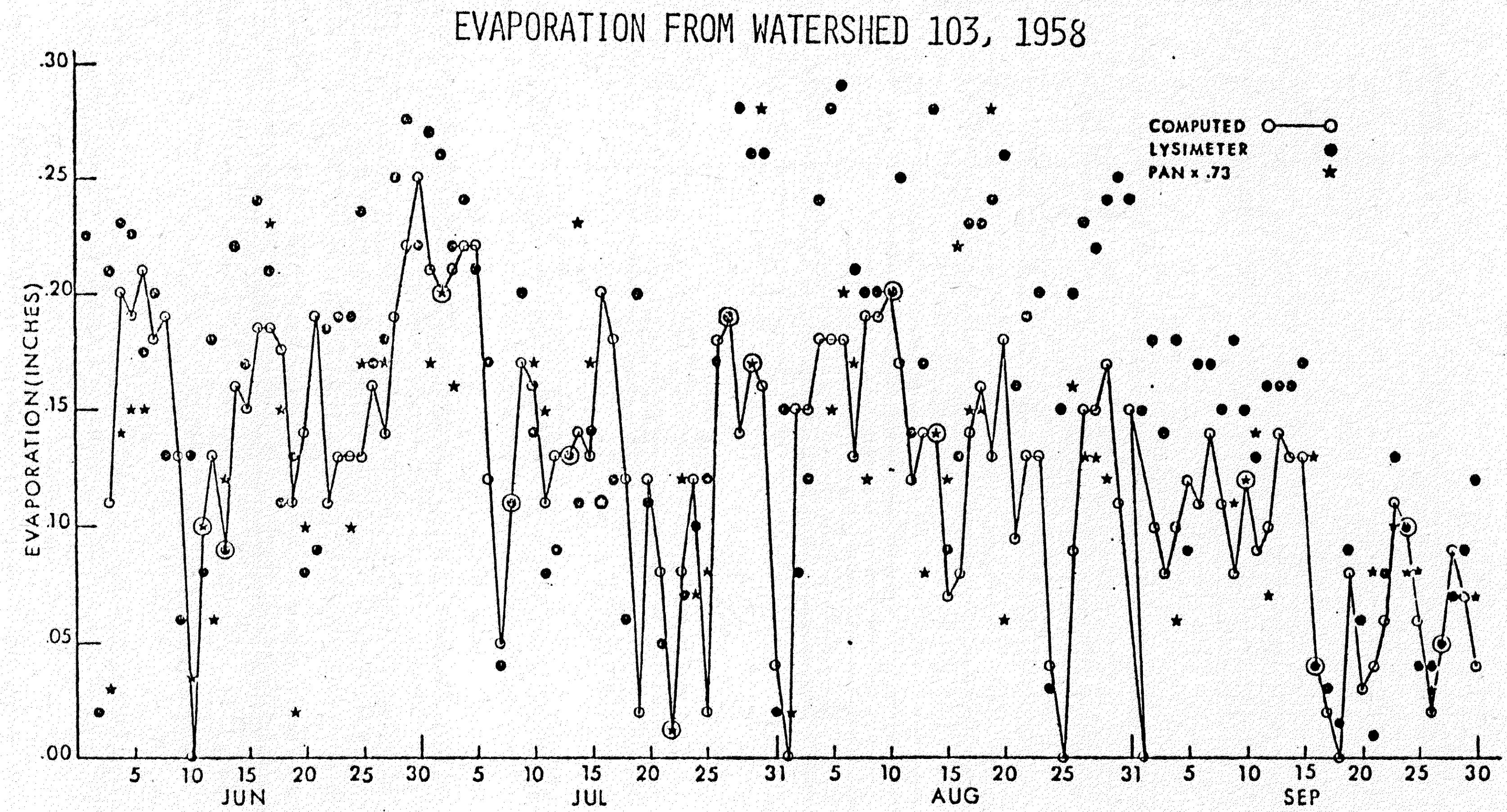

Figure 18.--Time series plot of evaporation from Watershed 103 for the 1958 evaporation season. Points are plotted for computed, pan, and lysimeter data.

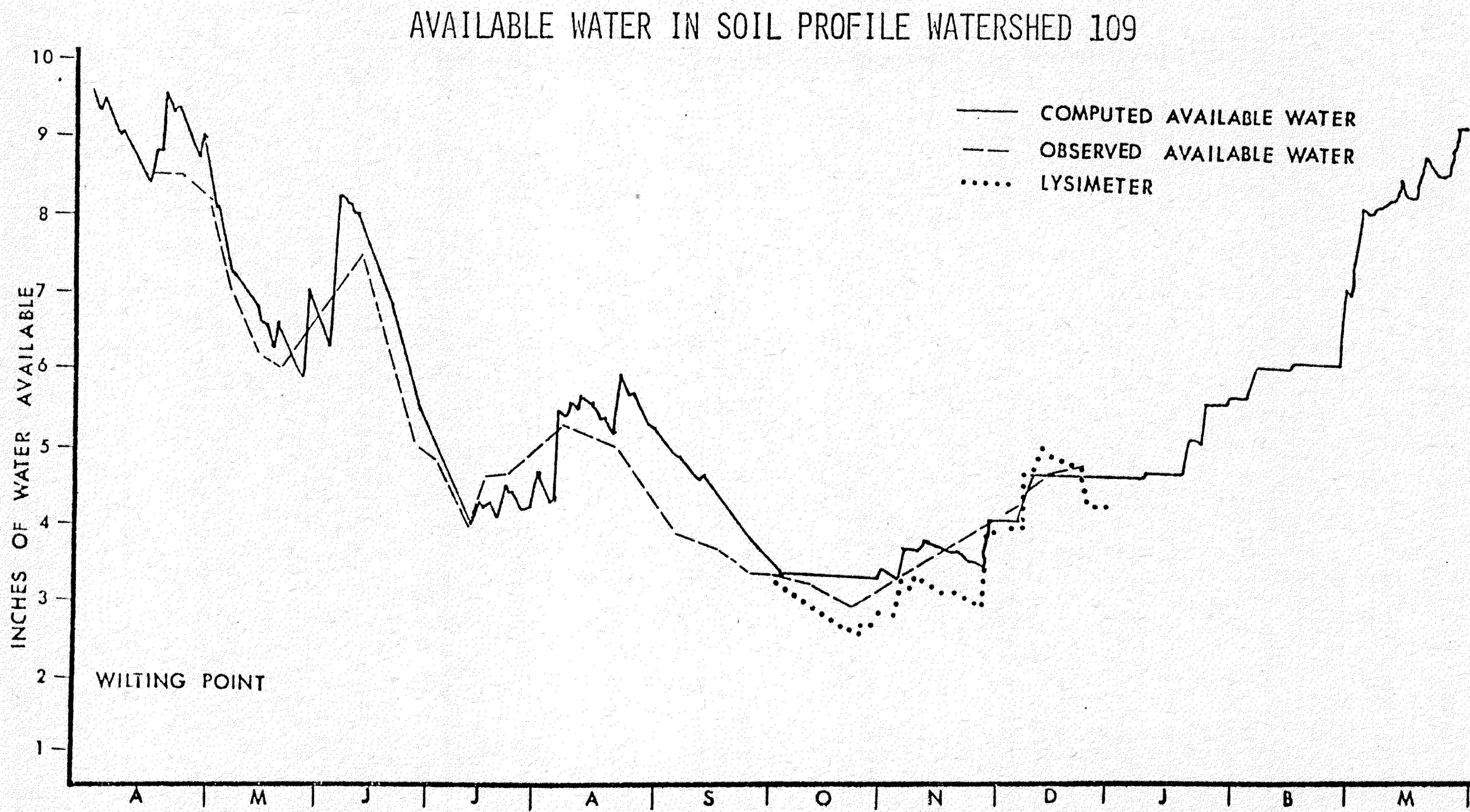

Figure 19.--Time series plot of soil moisture in the root zone of Watershed 109 at Coshocton, Ohio, for April 1963 through March 1964. The computed soil moisture is plotted on a daily basis for the entire period. The observed data from a nearby sampling site are for periods ranging from 2 or 3 days to 3 weeks from April through December 1963. The lysimeter data run on a daily basis from October to December 1963.

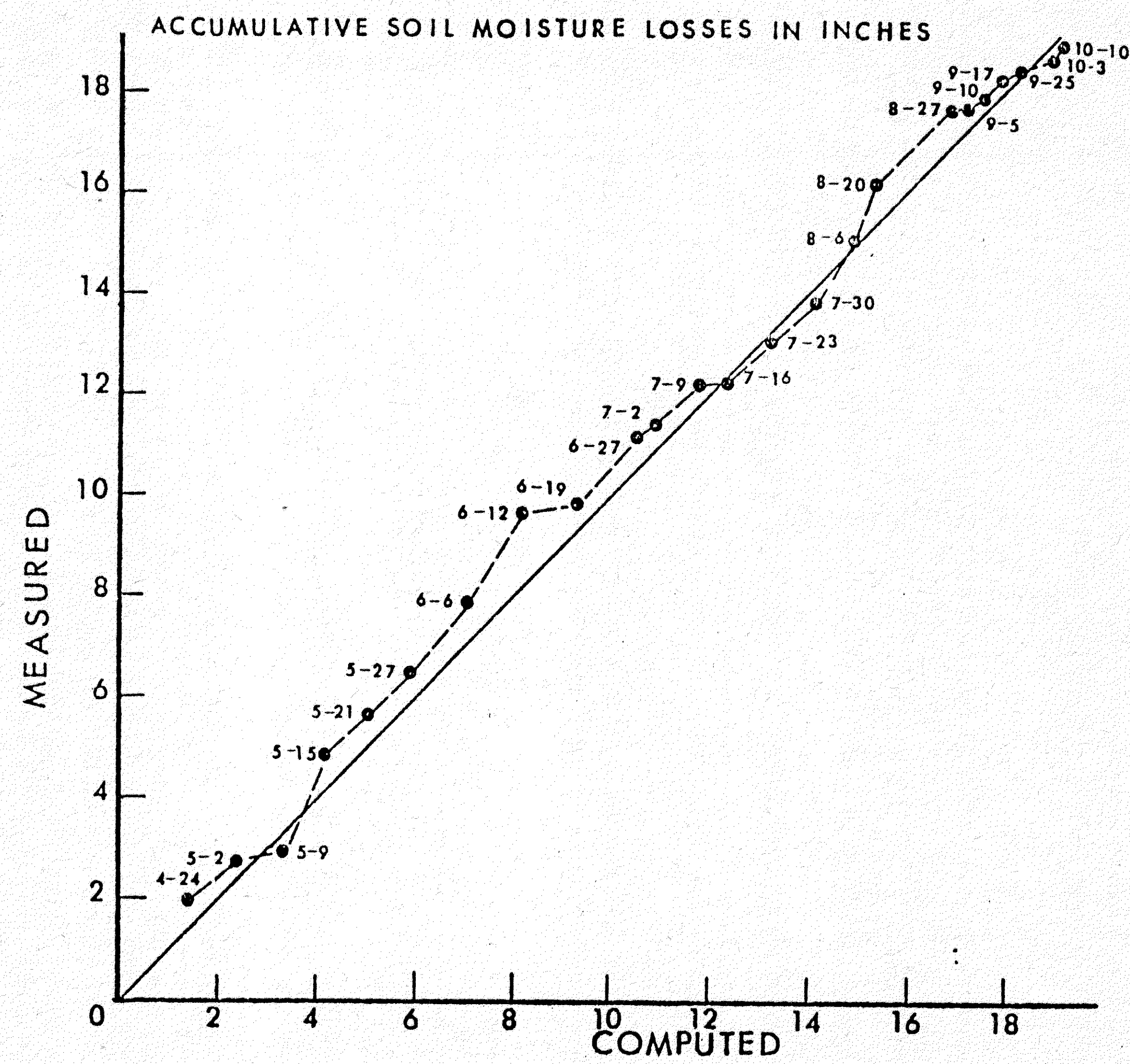

Figure 20.--Double mass plot of measured soil moisture losses versus computed soil moisture losses for Coshocton, Ohio, 1963.

at one point on the station several hundred feet from both of these watersheds. The published snowfall used for verification for the 1963-64 data comes from Mohawk Dam, about 16 miles west southwest of the station, and from the Zanesville FAA airport station, about 28 miles south of the station. The snow depth comparison is a comparison of instantaneous point measurements with computed values based on temporally and spatially averaged values.

This comparison allows the following problems to occur and should be acknowledged. Consider a day where 3 inches of snow are observed to fall but with a maximum temperature sufficient to melt 3 inches of snow. The following possibilities exist. The early part of the day can be warm with no snow on the ground. In the latter part of the day, temperatures can drop and snow fall. The observed change in snow depth is a net increase of 3 inches. Alternately, if snow falls at the first of the day and then temperatures rise, the snow can melt before an observation is taken. Other variations of these conditions would allow some fraction of the snow to be observed. However, since it is normal for clearing skies and colder temperatures to follow precipitation, the program normally computes melt before new snow is added.

The density of the snowfall also poses a complication. Density is not generally observed and tends to change with time depending on the temperature. Snow depth is dependent on the density and can decrease with ripening snow without loss of water equivalent. This is considered by procedures described in the chapter on modeling.

Figures 21-24 show plots of the observed snow depth. It can be seen that in the 1958-59 data that snow depth on the two watersheds differed by more than 1 inch on three occasions. Whenever this much difference occurred, the snow depth on watershed 109 with the meadow cover was greater than on watershed 103 with the winter wheat cover.

The time of observation must have contributed some variation. On January 2, 1958, no snowfall was reported at the station, but the snow depth increased from no snow to 1 inch on both watersheds. A total of 0.63 inch of precipitation was measured on the first 2 days of January.

Figure 25 shows a comparison of the observed snowfall at Mohawk Dam and at Zanesville FAA station with that computed on watershed 109.

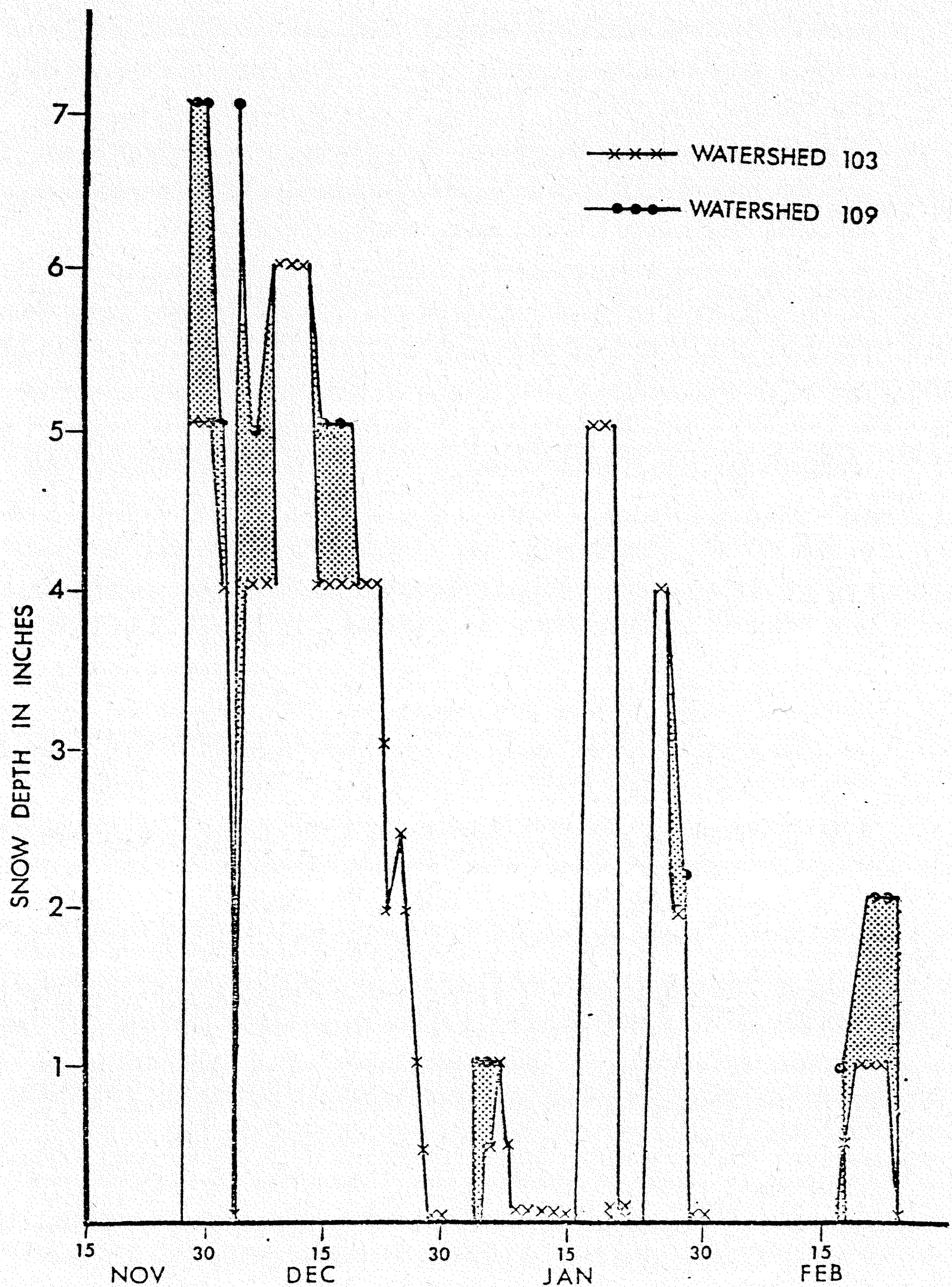

Figure 21.--Plot of observed snow depth on Watersheds 109 and 103 during December 1958 and January and February of 1959. Shaded areas indicate when Watershed 109 had deeper snow than Watershed 103.

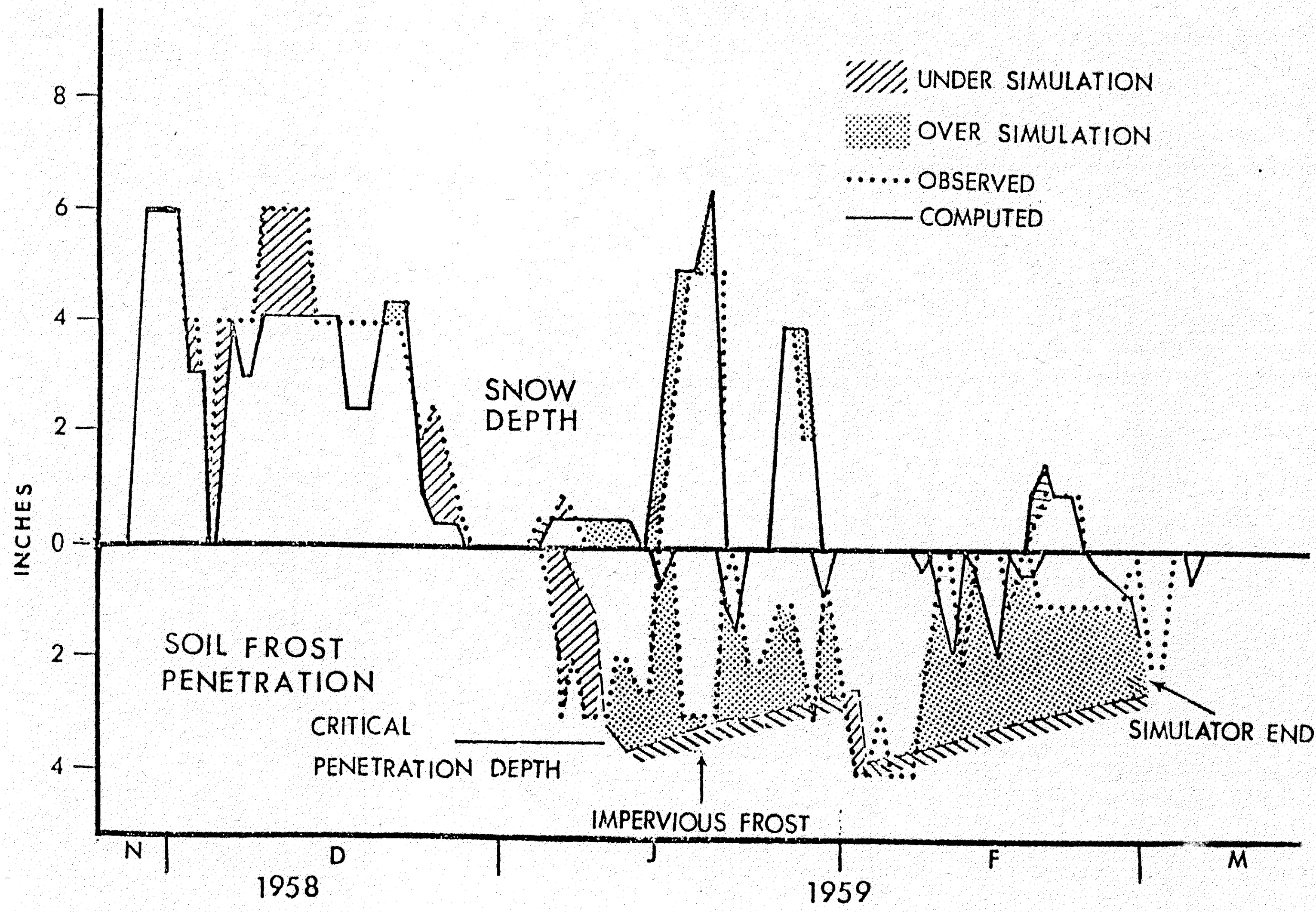

Figure 22.--Time series plot of snow depth and frost penetration and thaw showing measured-and simulated plots for Watershed 103 for 1958-59.

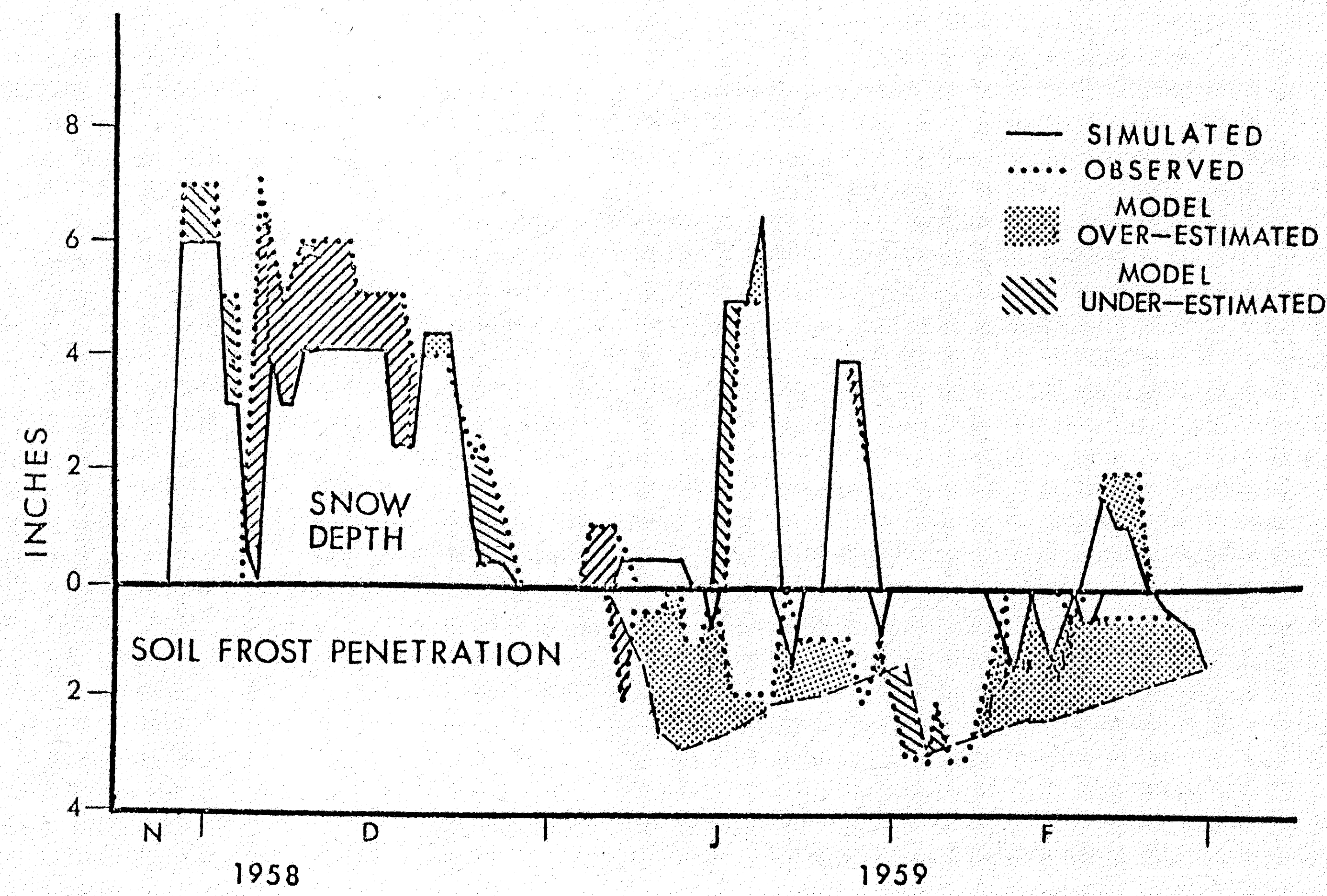

Figure 23.--Time series plot of snow depth and frost penetration and thaw showing measured and simulated plots for Watershed 109 for 1958-59.

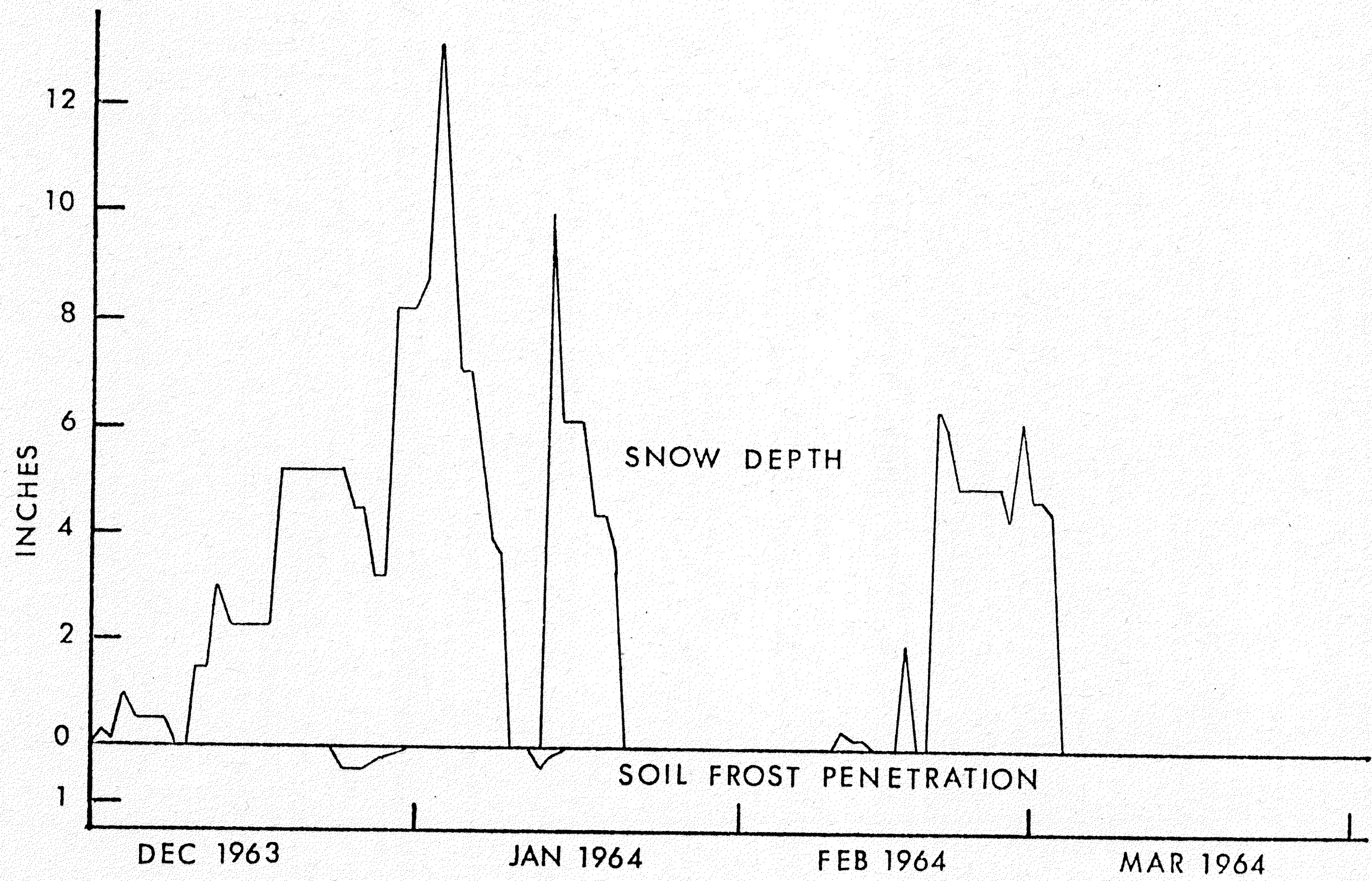

Figure 24.--Simulated time series plot of snow depth and frost penetration for Watershed 103 for 1963-64.

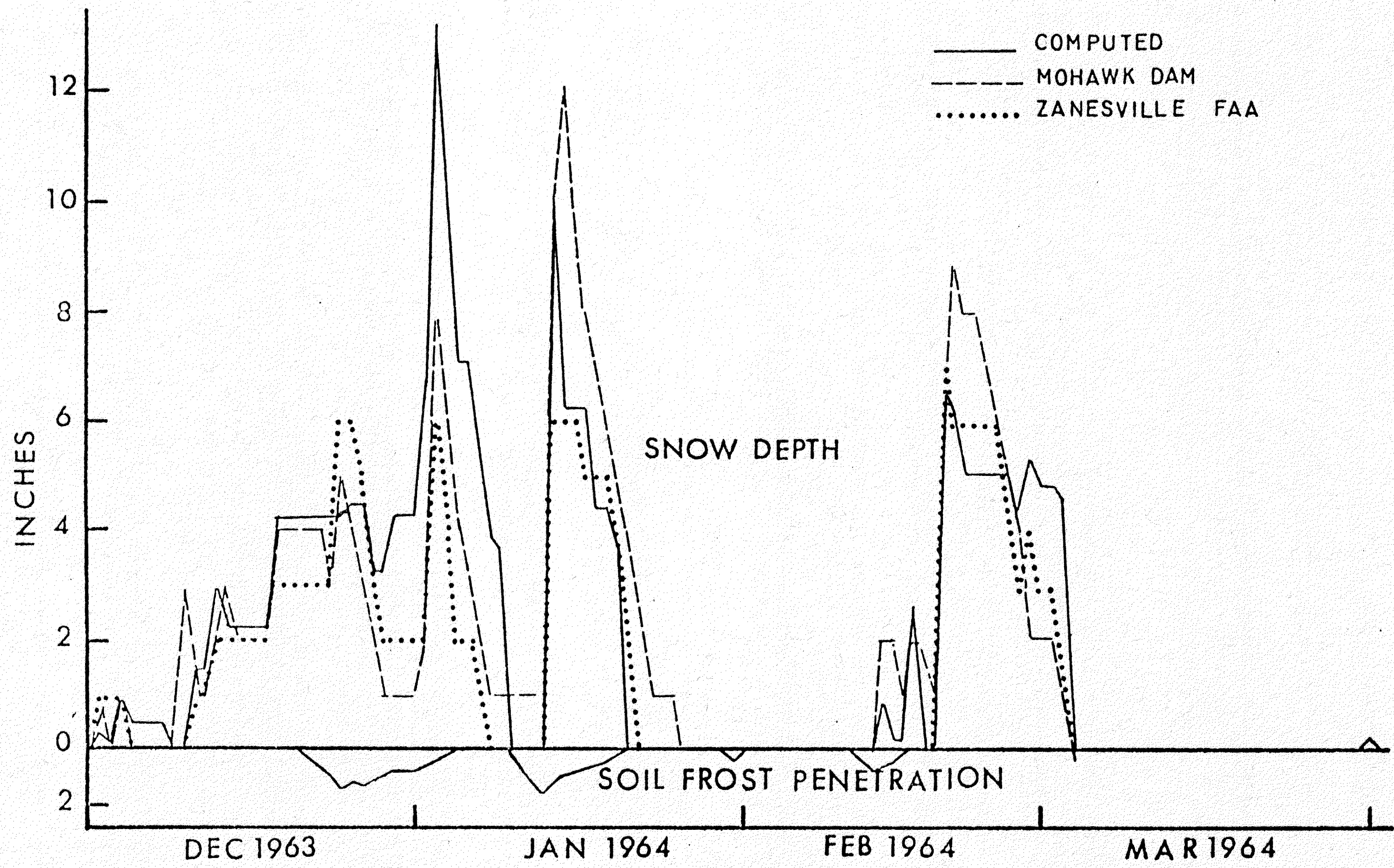

Figure 25.--Simulated time series plot of snow depth and frost penetration for Watershed 109 and reported snow depths at Mohawk Dam and at Zanesville FAA station for 1963-64.

It can be seen that there is about as good agreement between the observed and computed values as there is between the two observations themselves. The agreement between observed and computed snow depth in 1958 compares very favorably with the agreement between the two observed stations in 1963.

Frost Penetration Verification

The observed and computed frost depths are shown in Figures 22-25. There were no frost observations made in 1963-64. The model indicates that penetration was limited to less than 1 inch on both watersheds during that season. In 1958-59 the frost appears to come completely out of the ground during several periods while the model shows thawing occurring a significant distance down from the surface but not reaching the frost front. The general trend is quite good. The model lags the observed penetration by a day or more. This is partly expected when point measurements are compared with an areal estimate.

Frost Type Verification

No observations were made of frost type; however, a hydrologic event occurred in January of 1959 which tends to indicate the infiltration conditions of the two watersheds. During the month of January, 5.21 inches of precipitation were recorded. Watershed 103 had 2.17 inches of runoff, or 42 percent. During the same period, watershed 109 recorded 5.89 inches with 1 inch of runoff, or 17 percent. Much of this runoff occurred on January 20-21.

The short term record which is available for watershed 103 shows 2.40 inches of rain on January 21 and 2.15 inches of runoff. There were 5 inches of snow on the ground when the rain started. If the density of the snow was 0.2 when the rain began then the water available for runoff from both rain and snow was 3.40. The percent runoff was 63 percent. While there is no short term record available for this storm on watershed 109, if we assume that the entire inch of runoff for the month occurred on January 20 and that the precipitation was generally the same for January 21 on watershed 109 as on watershed 103 then the runoff percentage for the event would be as high as 30 percent. Much higher than normal but less than one-half of the runoff percentage on watershed 103. For comparison, in 1956 during the month of February, with 5.34 inches of measured precipitation on watershed

103, 0.76 inch, or 14 percent, ran off. During the same period watershed 109 had less than 1 percent runoff. These data suggest that infiltration in January 1959 was greatly reduced on both watersheds but especially on watershed 103.

The model indicates porous soil conditions existed during both seasons on watershed 109 although frost penetration was approaching that depth where impervious frost is nearly always expected to form. Watershed 103 was porous during the months of November and December and the first third of January of 1959. Frost penetration on this watershed exceeded a depth of 3.50 inches around January 12, 1959, and was classified impervious. It remained in an impervious condition through February. In this sense the model shows excellent agreement with observation.

Table 5 shows precipitation, runoff, and percent of precipitation which ran off for November, December, January, February, and March of 1958-59 and 1963-64.

TABLE 5

PRECIPITATION, RUNOFF, AND PERCENT RUNOFF BY MONTH FOR NOVEMBER-MARCH 1958-59 AND 1963-64 ON WATERSHEDS 103 AND 109 AT COSCHOCTON, OHIO

	SEASON	NOV.		DEC.		JAN.		FEB.		MAR.	
		#103	#109	#103	#109	#103	#109	#103	#109	#103	#109
Precipi-	58-59	1.81	2.15	.51	.90	5.21	5.89	2.84	3.49	2.33	2.51
tation	63-64	1.63	1.61	1.44	1.57	2.75	2.57	1.94	1.93	7.31	8.17
(in.)											
Runoff	58-59	0	0	0	0	2.17	1.00	.88	.30	T	T
(in.)	63-64	0	0	0	0	0	0	0	0	2.99	.18
Percent	58-59	0	0	0	0	42	17	31	9	0	0
	63-64	0	0	0	0	0	0	0	0	41	2

March 1964 shows a high runoff similar to the runoff of January 1959. The difference in the precipitation patterns of the two seasons is shown in Figures 26 and 27. The high runoff of March 1964 is easily attributable to high precipitation. These data indicate normal in infiltration in all months but January and February of 1959 and that watershed 103 was more affected by a reduction of infiltration than 109.

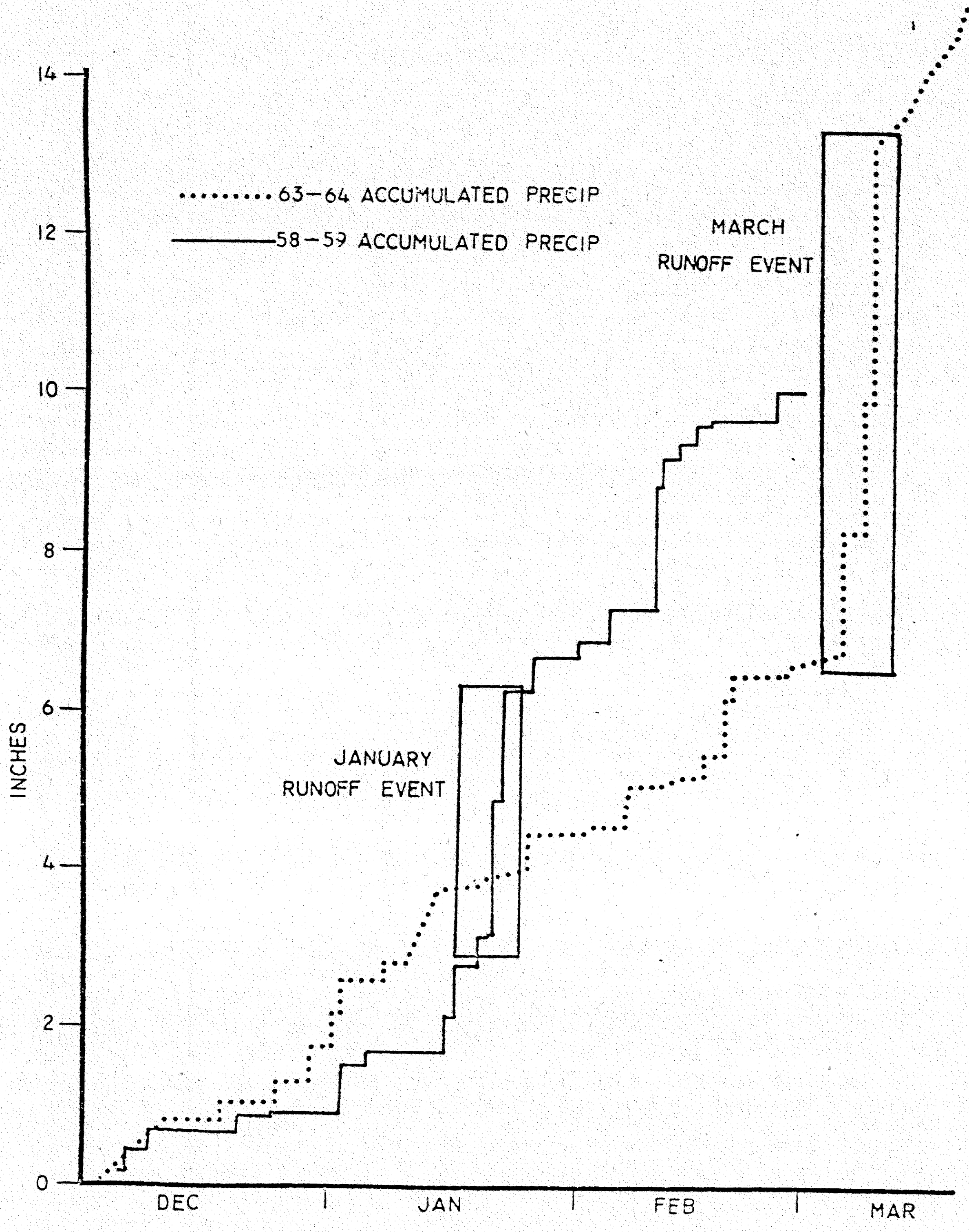

Figure 26.--Accumulated rainfall for winter 1958-59 and for winter 1963-64 at Coshocton, Ohio.

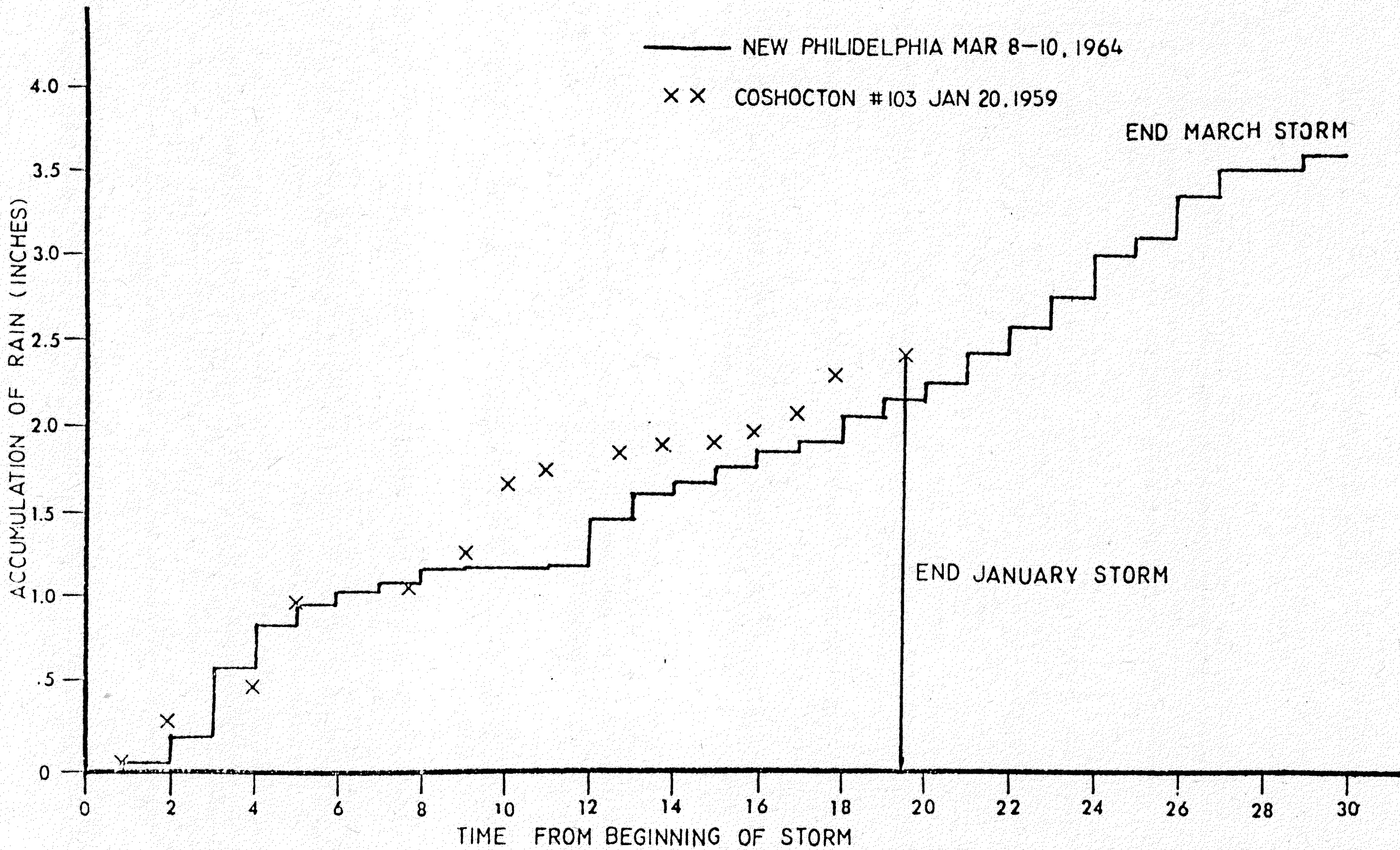

Figure 27.--Accumulated rainfall for storm of January 20, 1959, at Coshocton, Ohio, on Watershed 103 and for New Philadelphia for the storm of March 8-10, 1964.

Stalactite conditions were predicted only twice. They were on watershed 109 and occurred during the period that 103 was classified impervious.

Discussion

The estimation of depth of penetration and type of frost results from the consideration of many different measurements and estimates. Each was considered sufficiently important to be required in the model, and it is conceivable that cases exist where any one of them might be the governing factor. In this study, however, it appears that the effective thickness and conductivity of the litter layer and the depth, density, and the thermal conductivity of the snow were most important. These include variables which are rarely measured and subject to significant changes during the daily time step.

Sensitivity studies were made for the data of January 12, 1959, and are shown in Figures 15, 28, 29, 30, and 31. For soil moistures between the wilt point and field capacity the porosity seems to be more important than soil moisture or soil type. This agrees with the conclusions of Jumikus (1973). This conclusion is further illustrated by comparing the depth of frost penetration on the two watersheds during the season 1963-64. Although watershed 103 has the less well-drained soil, reference to Tables 3 and 4 show that for the upper layer in which most of the freezing occurred, 103 had the greater porosity. Assuming similar thermal conductivity of the mineral in the two soils, 109 with a lower porosity would have a higher thermal conductivity and was simulated to freeze deeper than 103 (see Figure 32) in a season when they both had similar types of shallow ground cover.

Figure 29 shows frost penetration as a function of litter for soils with two different porosities and various soil moistures. When even a thin layer of litter is on the surface, the decrease in conductivity makes the increase in latent heat and specific heat of soil moisture the governing factor for frost penetration. When the ground is bare, soil frost follows the lower curve shown in Figure 30, but when the ground is insulated by litter or snow, the shape of the curve changes with reduced frost depths found in moister soils. Thus, it is expected that the effect of litter and vegetation

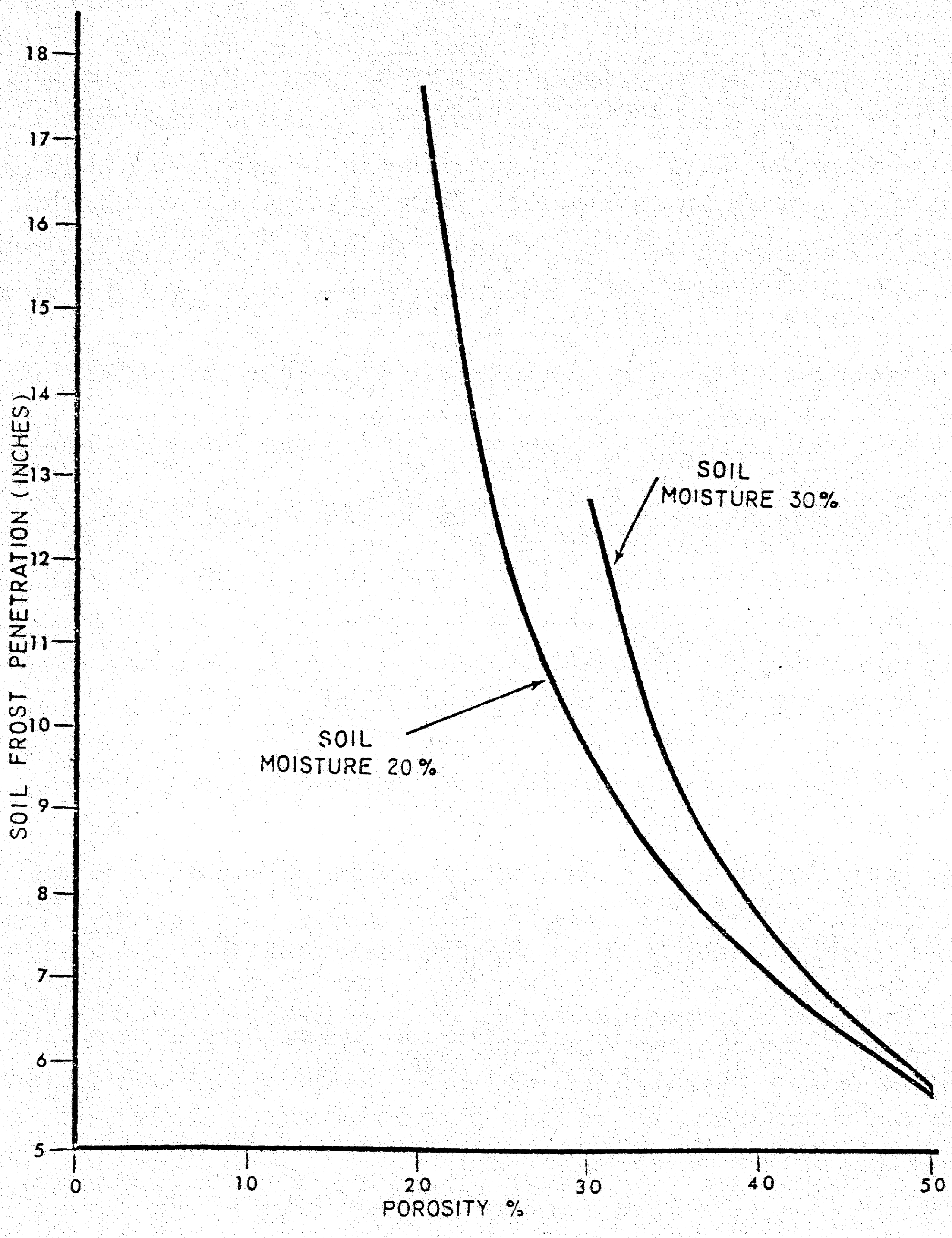

Figure 28.--Variation of soil frost penetration as a function of porosity and soil moisture using data from January 12, 1959.

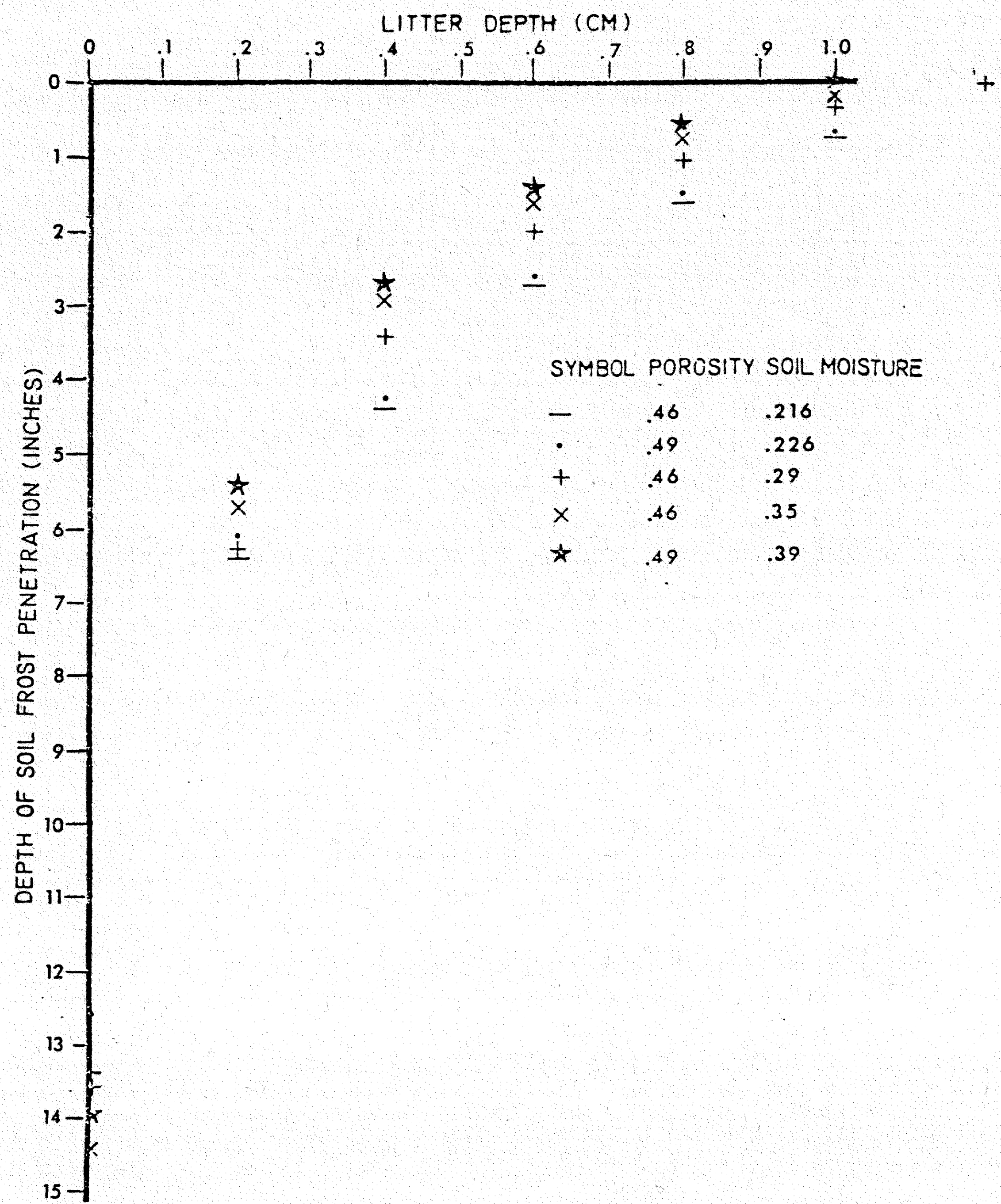

Figure 29.--Variation in frost penetration as a function of litter depth for soils with two different porosities and varying soil moisture concentrations. The soil with a porosity of 49 percent is similar to Watershed 103, while the soil with a porosity of 46 percent is like the upper layer of Watershed 109. Antecedent temperature conditions are similar to those of January 12, 1959.

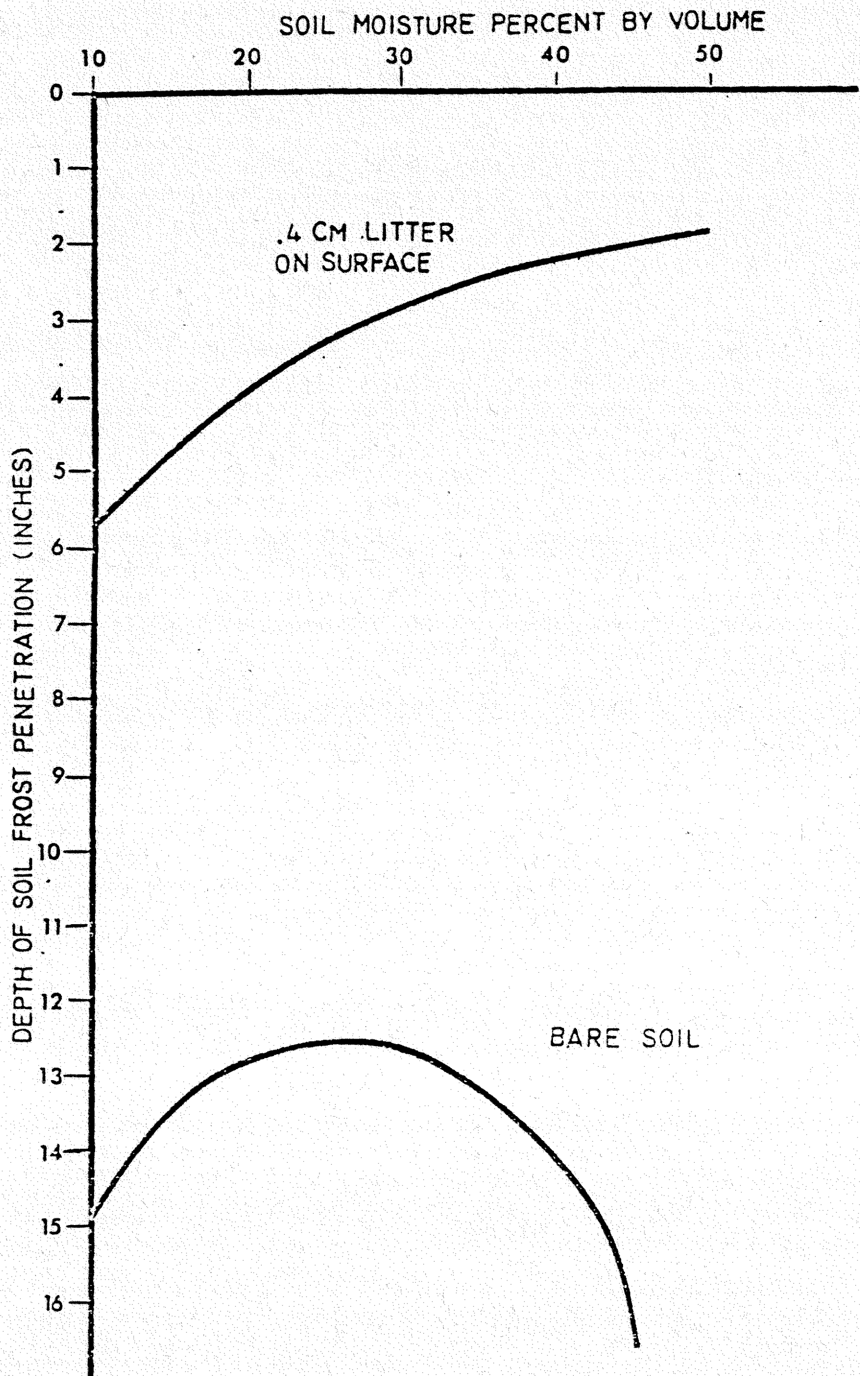

Figure 30.--Soil frost penetration as a function of soil moisture for bare soil in a 0.4-cm cover of litter. The porosity of the soil is 0.49. Other data are the same as for figure 29.

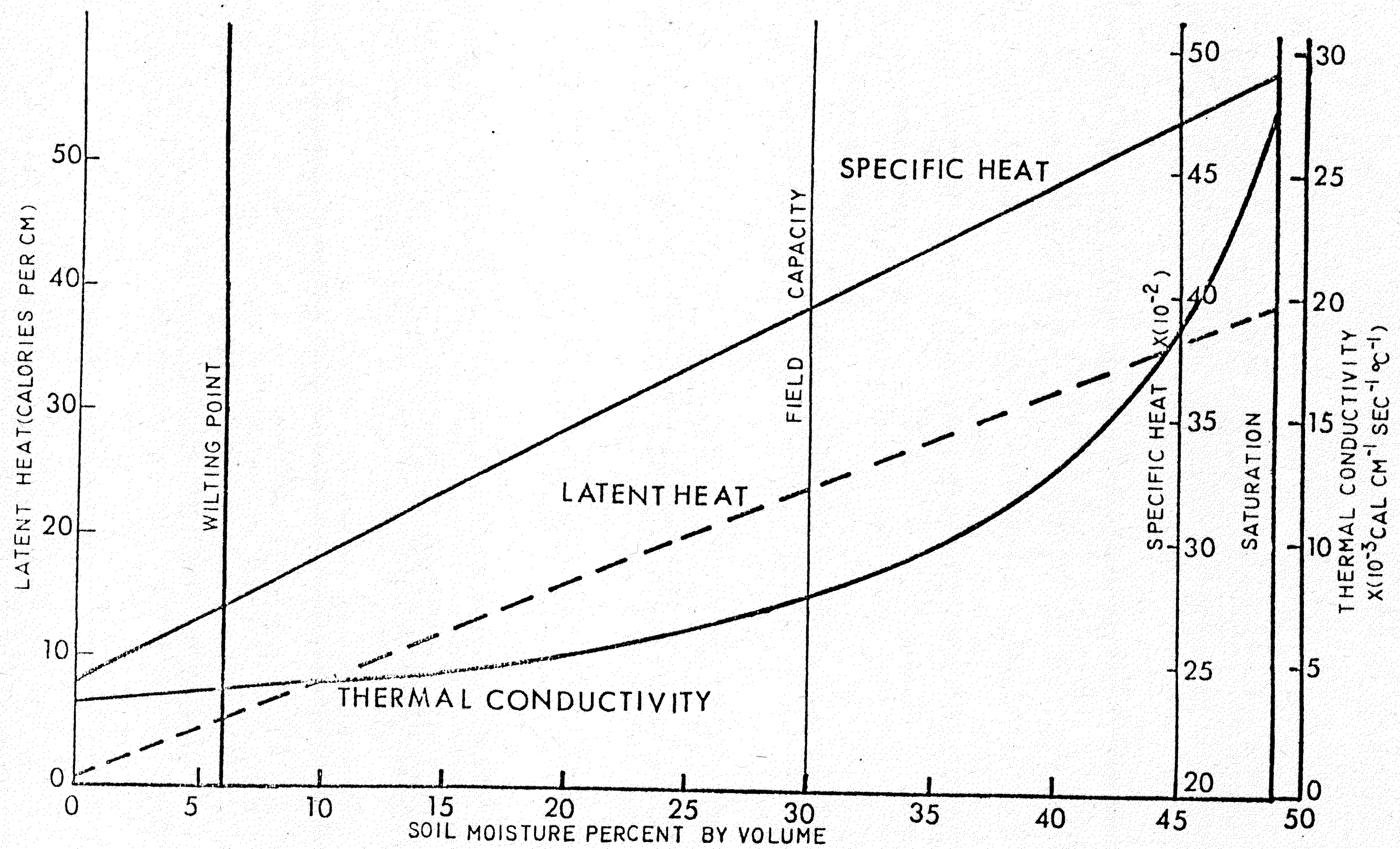

Figure 31.--Variation of thermal conductivity, latent heat, and volumetric heat capacity as a function of soil moisture. The soil is the same used in making figure 15.

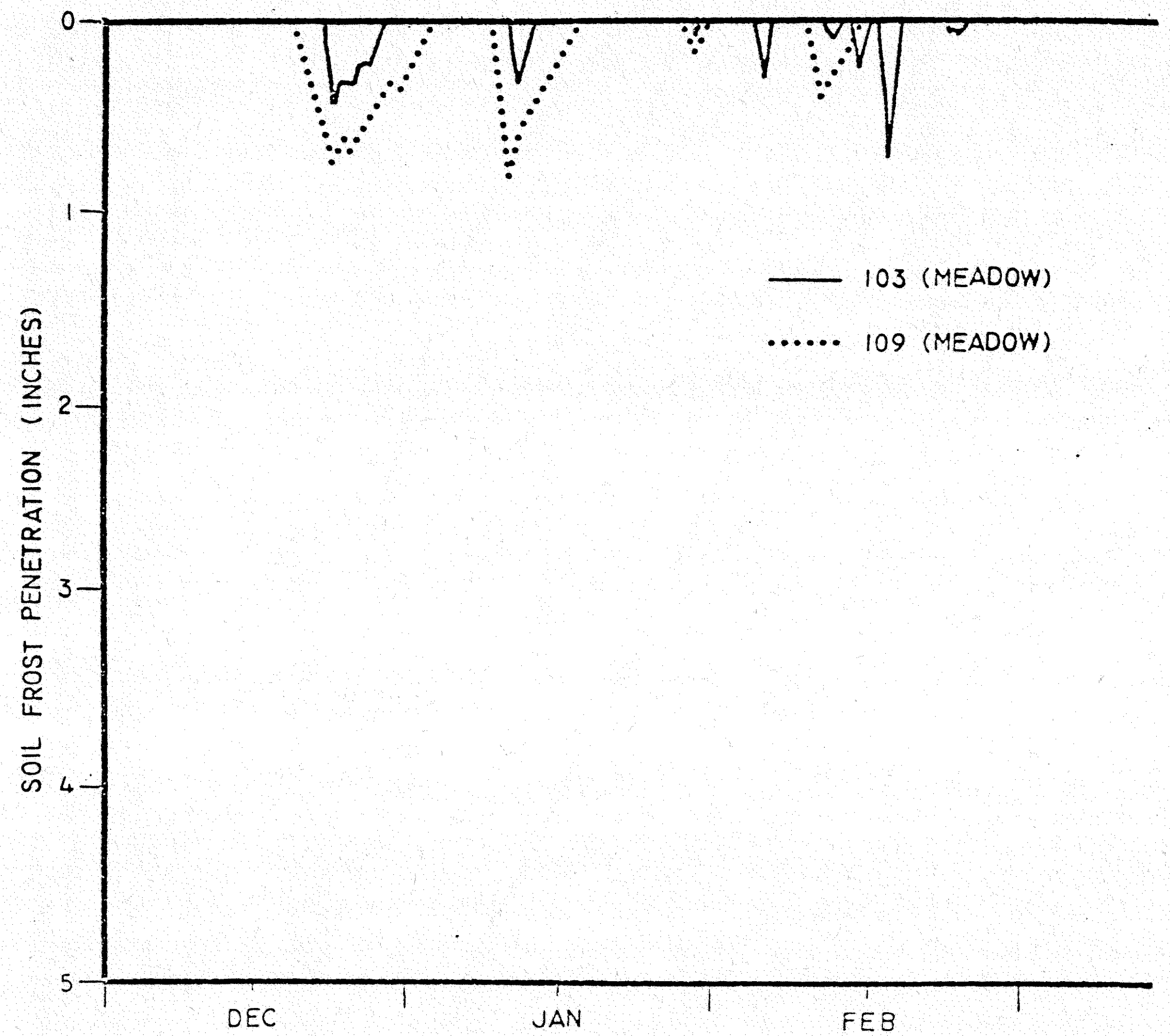

Figure 32.--Frost penetration on Watersheds 103 and 103 at Coshocton, Ohio, during the winter of 1963-64.

will mask many of the variations caused by differences in wheat. The litter cover on the meadow areas was in the range of four to five times the weight per unit area of the winter wheat areas. During this season 103 was both simulated and observed to freeze deeper than 109 (see Figure 33). The effect of the vegetation and litter on the surface had masked the effect of the porosity. This agreed with the findings of Post and Dreibelbis (1942).

The interaction of snow, litter, and temporal changes in snow and litter has been previously discussed. These effects make sufficient data difficult to obtain for determining calibration constants. With the current meteorological data available, conceptual modeling of how the litter layer insulation changes based on agricultural practices, land use surveys, and practical experience combined with snow surveys seems the best approach to making frost predictions. Caution and logic must be used in interpreting results. An erroneous boundary layer assumption could cause the model to be incorrect for an extensive period. Once the ground has frozen imperviously it seems unreasonable to assume that the ground will again become porous until it has completely thawed and has had time to drain away the excess water that has been trapped above it. In all but the very sterile soils, the portion of the frozen soil most likely to be impervious is the bottom. There is less organic material there to cause aggregation and fewer plant roots to provide openings in ice lenses. The most active boundary for energy exchange is at the surface. With this in mind, it is clear that for imperviously frozen ground to rapidly return to a porous condition, the ground must thaw from the surface down to the bottom of the frozen layer. Thawing of the ground from below will occur and is shown in Figures 22-25. It occurs however fairly slowly. Should the lower portion of the frost layer which first became impervious become thawed from below, it is assumed that, because the impervious layer will likely have caused moisture concentrations to build above the lense and freeze imperviously there, the frost is still impervious until the entire profile has thawed and drained.

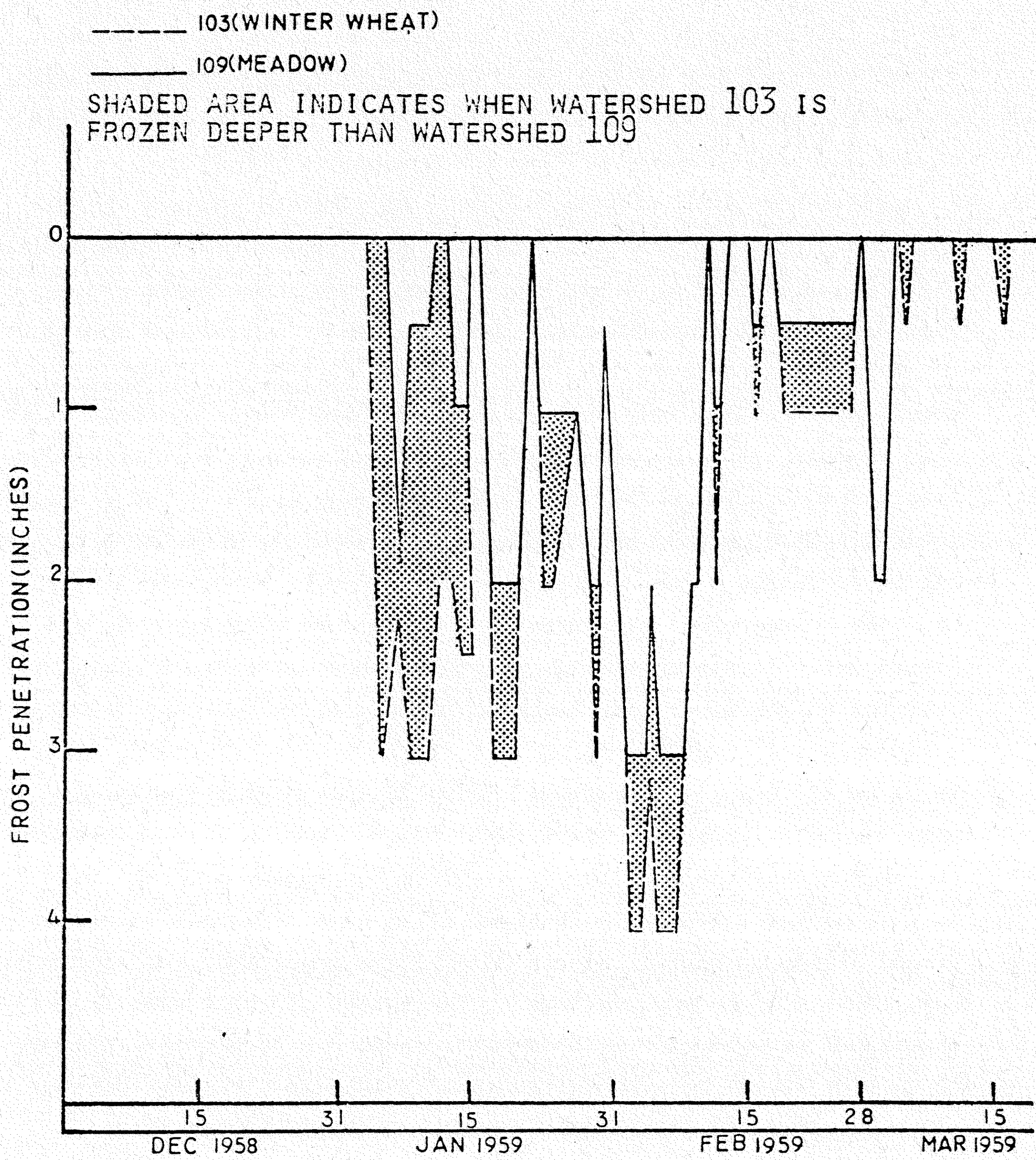

Figure 33.--Comparative plot of frost penetration for the winter of 1958-59 at Coshocton, Ohio, on Watersheds 103 and 109.

There is some leeway built into the system by the variability of nature. As a watershed begins to approach conditions when an impervious estimate would be given, patches of impervious frost would already have formed in wet spots or possible areas of tighter soil. When the condition develops to where the model predicts impervious ground then a significant portion of the area should be impervious. There are likely several areas which yet remain porous. It is the effective permeability which will cause changes in the runoff. The conversion from porous to impervious is not likely to occur in a single instant either, but the model should be oriented to give warning when the reduction or increase of infiltration changes significantly.

CHAPTER SIX. CONCLUSIONS

Several investigators have found that infiltration capacity changes when the soil is cold or the ground frozen. Such changes do influence runoff events and have an impact on flood forecasting. The objectives of this study were to identify the magnitude and occurance of frozen ground and warn of deviations from normal rainfall-runoff relationships.

The effect has been examined in two categories: soil with temperatures near, but not at, or below, the freezing point; and soils in which frost occurs. There is good evidence of moisture transport to the surface in a cold but unfrozen soil. The moisture is moved by a potential gradient arising from a temperature gradient. However, no quantitative results have been found with sufficiently general application to be incorporated into quantitative models. The results do indicate however, a direction for corrections.

During this study it has been demonstrated that frost penetration can be computed and the type of frost and its resulting effect on infiltration can be qualitatively estimated with the relatively limited data from generally available sources. These data can be processed through simple formulations in relatively little time. Two years of data for one watershed can be run on a large computer in less than 4 seconds.

This model can be used for sensitivity testing to determine the importance of cover type, thermal conductivity of litter and soil, the density of snow, etc. It serves to point out the importance of the sequence of natural events, such as snowfall, rain, and freezing temperatures, in their effect on frost penetration. It helps to educate the user to the relationship between snowmelt and ground thaw. It serves as an estimator of evapotranspiration and soil moisture distribution. It brings those with an interest in this problem to the point where varying land use and its effect on frost can now be used to check out larger basins. There is now a system where land cover measurements available from remote sensing can be used to estimate the effect on infiltration in multiple land use basins.

The model has inherent weaknesses because of the daily time step which doesn't resolve the sequence of precipitation, evapotranspiration, and freezing temperatures. Dependence on a frost a frost index does not take into account the effect of energy absorption when the ground is heated by solar radiation which may not always be reflected in an increase of the mean air temperature. The model is not closely tied to ground temperatures, which is an asset in the light of the sparsity of data but a liability in that the results tend to be smoothed by the averaging effect of the frost index and there the model does not duplicate the rapid changes in the observed data.

Further work should be considered in testing the model under more severe winters and in other geographical areas. The forested regions near Coshocton rarely have much frost or runoff.

The importance of snow and ground cover require that measurements be gathered and evaluated rapidly for large areas. The value of remote sensing measurements should be assessed in this repect.

APPENDIX A

COVER, PRECIPITATION(P), DISCHARGE(Q), AND RATIO Q/P FOR DEC., JAN., FEB., AND MAR. FOR WATERSHEDS 103 AND 109.

M..Meadow C..Corn W..Wheat M/W..Meadow then planted to Wheat

WATERSHED	MONTH		1956	1957	1958	1959	1960	1961	1962	1963	1964	1965	1966	1967
		COVER VALUE	M	M	C	W	M	M	M/W	W/M	M	M	C	W
103	Dec	P	3.61	3.99	.51	2.11	1.50	2.52	2.09	1.44	4.36	.55	2.09	2.61
		Q	.0	.38	.0	.23	.0	.0	.0	.0	.04	.0	.0	T
		Q/P	.0	.10	.0	.11	.0	.0	.0	.0	.01	.0	.0	.0
	Jan	P	1.60	1.59	1.48	5.21	2.87	.62	2.83	2.07	2.75	2.58	3.39	.89
		Q	.0	.11	.05	2.17	.29	.0	.39	.02	.0	.11	.11	.0
		Q/P	.0	.07	.03	.42	.10	.0	.14	.01	.0	.04	.03	.0
	Feb	P	5.34	1.34	.58	2.84	3.19	3.83	3.37	1.13	1.94	3.64	2.32	1.64
		Q	.76	.0	.0	.88	.09	.26	2.38	1.50	.0	.33	.39	.0
		Q/P	.14	.0	.0	.31	.03	.07	.71	1.32	.0	.09	.17	.0
	Mar	P	3.49	1.80	.98	2.33	1.00	3.40	3.10	5.95	7.31	2.28	1.00	4.38
		Q	T	.0	.0	T	1.37	.74	.41	4.59	2.99	.02	T	.46
		Q/P	.0	.0	.0	.0	1.37	.22	.13	.77	.41	.01	.0	.13
			M	C	W	M	M	C	W/M	M	M	C	W	M
109	Dec	P	2.97	4.24	.90	2.33	1.52	2.44	2.16	1.50	4.51	.82	2.41	2.72
		Q	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0
		Q/P	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0	.0
	Jan	P	1.52	1.58	1.40	5.89	3.24	.84	2.79	2.06	2.57	2.67	3.50	.91
		Q	.0	.01	.02	1.00	.0	.0	.31	.0	.0	.0	.01	.0
		Q/P	.0	.01	.01	.17	.0	.0	.11	.0	.0	.0	.0	.0
	Feb	P	5.48	1.32	.75	3.49	3.43	4.01	3.38	1.13	1.93	3.55	2.63	1.43
		Q	.01	.0	.06	.03	.0	T	2.08	.02	.0	.10	.15	.0
		Q/P	.0	.0	.08	.09	.0	.0	.61	.02	.0	.03	.06	.0
	Mar	P	3.65	1.89	.96	2.51	1.17	3.50	2.85	6.14	8.17	2.40	1.44	4.64
		Q	.0	.0	.0	T	.0	.01	.47	2.67	.18	.0	.0	.05
		Q/P	.0	.0	.0	.0	.0	.0	.17	.43	.02	.0	.0	.01

APPENDIX B

AVAILABLE WATER FOR COSHOCTON, OHIO, 1962 AND 1963

1972 A=AVAILABLE WATER D=DEFICIT

DATE	DEPTH (ins.)	0-6	6-12	12-24	24-36	36-42	TOTAL	PRECIP
JAN 11	A	1.49	.96	1.88	2.11	.89	7.33	
	D	- .29	.24	.42	.06	.20	.63	
JAN 23	A	2.16	1.00	1.74	2.19	.89	7.98	1.09
	D	- .96	.20	.56	- .02	.20	- .02	
FEB 13	A	2.16	.98	1.70	2.17	.90	7.91	2.07
	D	- .96	.22	.60	.00	.19	.05	
MAR 12	A	1.56	1.14	2.15	2.21	.96	8.02	3.67
	D	- .36	.06	.15	- .04	.13	- .06	
MAR 20	A	1.56	1.08	2.01	2.31	.99	7.95	.31
	D	- .36	.12	.29	- .14	.10	.01	
APR 5	A	1.35	1.05	2.05	2.33	.98	7.76	1.68
	D	- .15	.15	.25	- .16	.11	.20	
APR 17	A	1.35	1.05	2.10	2.26	.97	7.73	.94
	D	- .15	.15	.20	- .09	.12	.23	
APR 25	A	1.14	.89	1.91	2.17	.95	7.06	.22
	D	.06	.31	.39	.00	.14	.90	
APR 30	A	1.05	.80	1.78	2.17	.95	6.75	.15
	D	.15	.40	.52	.00	.14	1.21	
MAY 9	A	.94	.68	1.63	2.11	.94	6.30	.53
	D	.26	.52	.67	.06	.15	1.66	
MAY 16	A	.14	.46	1.49	2.10	1.08	5.27	.05
	D	1.06	.74	.81	.07	.01	2.69	
MAY 24	A	.81	.39	1.12	1.94	.83	5.09	1.25
	D	.39	.81	1.18	.23	.26	2.87	
MAY 31	A	.74	.38	1.05	1.83	.75	4.75	.87
	D	.46	.82	1.25	.34	.34	3.21	
JUN 8	A	.48	.36	1.10	1.80	.73	4.47	.58
	D	.72	.84	1.20	.37	.36	3.49	
JUN 18	A	.13	.22	.96	1.71	.66	3.68	.57
	D	1.07	.98	1.34	.46	.43	4.28	
JUN 26	A	.64	.20	.76	1.63	.65	3.88	.65
	D	.56	1.00	1.54	.54	.44	4.08	
JUL 2	A	.05	- .01	.55	1.48	.60	2.67	.28
	D	1.15	1.21	1.75	.69	.49	5.29	
JUL 9	A	.12	.04	.59	1.40	.55	2.70	1.23
	D	1.08	1.16	1.71	.77	.54	5.26	
JUL 17	A	.10	- .01	.51	1.29	.48	2.37	.12
	D	1.10	1.21	1.79	.88	.61	5.59	
JUL 23	A	.04	- .07	.40	1.18	.43	1.98	.27
	D	1.16	1.27	1.90	.99	.66	5.98	
JUL 31	A	.03	- .02	.47	1.22	.45	2.15	.94
	D	1.17	1.22	1.83	.95	.64	5.81	
AUG 6	A	.01	- .07	.37	1.17	.48	1.96	0.00
	D	1.19	1.27	1.93	1.00	.61	6.00	

1972 A=AVAILABLE WATER D=DEFICIT

DATE	DEPTH (ins.)	0-6	6-12	12-24	24-36	36-42	TOTAL	PRECIP
AUG 14	A	.16	.05	.43	1.18	.44	2.26	.38
	D	1.04	1.15	1.87	.99	.65	5.70	
AUG 23	A	.31	- .01	.37	1.10	.46	2.23	.51
	D	.89	1.21	1.93	1.07	.63	5.73	
AUG 28	A	.06	- .06	.32	1.11	.40	1.83	T
	D	1.14	1.26	1.98	1.06	.69	6.13	
SEP 4	A	1.13	.28	.42	1.02	.37	3.22	2.05
	D	.07	.92	1.88	1.15	.72	4.74	
SEP 12	A	1.01	.41	.58	1.03	.37	3.40	1.00
	D	.19	.79	1.72	1.14	.72	4.56	
SEP 20	A	1.14	.62	.82	1.06	.38	4.02	1.41
	D	.06	.58	1.48	1.11	.71	3.94	
OCT 1	A	1.22	.66	.96	1.16	.41	4.41	.88
	D	- .02	.54	1.34	1.01	.68	3.55	
OCT 15	A	1.26	.78	1.15	1.17	.43	4.79	1.31
	D	- .06	.42	1.15	1.00	.66	3.17	
OCT 24	A	1.09	.65	1.08	1.18	.41	4.41	.51
	D	.11	.55	1.22	.99	.68	3.55	
NOV 5	A	1.03	.66	1.09	1.26	.43	4.47	.69
	D	.17	.54	1.21	.91	.66	3.49	
NOV 23	A	1.49	1.02	1.92	1.94	.76	7.13	2.72
	D	- .29	.18	.38	.23	.33	.83	
DEC 3	A	1.29	.90	1.81	2.04	.77	6.81	0.00
	D	- .09	.30	.49	.13	.32	1.15	
DEC 19	A	1.19	.96	1.88	1.93	.77	6.73	1.30
	D	.01	.24	.42	.24	.32	1.23	

1973 A=AVAILABLE WATER D=DEFICIT

DATE	DEPTH (ins.)	0-6	6-12	12-24	24-36	36-42	TOTAL	PRECIP
MAR 27	A	1.53	1.04	1.87	2.04	.83	7.31	
	D	- .33	.16	.43	.13	.26	.65	
APR 17	A	1.41	.84	1.66	2.08	.88	6.87	.65
	D	- .21	.36	.64	.09	.21	1.09	
APR 24	A	1.49	.92	1.67	1.95	.83	6.86	1.92
	D	- .29	.28	.63	.22	.26	1.00	
MAY 2	A	1.42	.85	1.57	1.94	.83	6.61	.46
	D	- .22	.35	.73	.23	.26	1.35	
MAY 9	A	.62	.56	1.40	1.97	.79	5.34	0.00
	D	.58	.64	.90	.20	.30	2.62	
MAY 15	A	.40	.40	1.18	1.87	.74	4.59	.20
	D	.80	.80	1.12	.30	.35	3.37	
MAY 21	A	.46	.35	1.05	1.81	.73	4.40	.55
	D	.74	.85	1.25	.36	.36	3.56	
MAY 29	A	1.04	.45	1.00	1.78	.71	4.98	1.42
	D	.16	.75	1.30	.39	.38	2.98	
JUN 6	A	1.40	.73	1.17	1.64	.66	5.60	1.99
	D	- .20	.47	1.13	.53	.43	2.36	
JUN 12	A	1.37	.83	1.35	1.73	.66	5.94	1.16
	D	- .17	.37	.95	.44	.43	2.02	
JUN 19	A	.72	.55	1.19	1.72	.65	4.83	0.00
	D	.48	.65	1.11	.45	.44	3.13	
JUN 27	A	.01	.20	.89	1.64	.64	3.38	T
	D	1.19	1.00	1.41	.53	.45	4.58	
JUL 2	A	- .07	.11	.88	1.59	.67	3.18	0.00
	D	1.27	1.09	1.42	.58	.42	4.78	
JUL 9	A	- .03	.02	.50	1.38	.56	2.43	0.00
	D	1.23	1.18	1.80	.79	.53	5.53	
JUL 16	A	.25	.10	.65	1.36	.62	2.98	.53
	D	.95	1.10	1.65	.81	.47	4.98	
JUL 23	A	.50	.13	.56	1.34	.56	3.09	1.02
	D	.70	1.07	1.74	.83	.53	4.87	
JUL 30	A	.70	.21	.65	1.27	.56	3.39	1.09
	D	.50	.99	1.65	.90	.53	4.57	
AUG 6	A	.80	.33	.76	1.27	.54	3.70	1.57
	D	.40	.87	1.54	.90	.55	4.26	
AUG 20	A	.79	.10	.72	1.29	.50	3.40	.97
	D	.41	1.10	1.58	.88	.59	4.56	
AUG 27	A	.29	.23	.72	1.26	.49	2.99	.86
	D	.91	.97	1.58	.91	.60	4.97	
SEP 3	A	.00	.09	.64	1.29	.52	2.54	T
	D	1.20	1.11	1.66	.88	.57	5.42	
SEP 10	A	- .03	.06	.57	1.25	.48	2.33	T
	D	1.23	1.14	1.73	.92	.61	5.63	
SEP 17	A	- .12	.00	.52	1.21	.49	2.10	.14
	D	1.32	1.20	1.78	.96	.60	5.86	
SEP 25	A	- .12	- .04	.43	1.12	.45	1.84	T
	D	1.32	1.24	1.87	1.05	.64	6.12	
OCT 3	A	- .12	- .05	.38	1.05	.43	1.69	.03
	D	1.32	1.25	1.92	1.12	.66	6.27	

1973 A=AVAILABLE WATER D=DEFICIT

DATE	DEPTH (ins.)	0-6	6-12	12-24	24-36	36-42	TOTAL	PRECIP
OCT 10	A	- .12	- .06	.33	1.03	.43	1.61	T
	D	1.32	1.26	1.97	1.14	.66	6.35	
OCT 17	A	- .12	- .06	.34	.97	.39	1.52	0.00
	D	1.32	1.26	1.96	1.20	.70	6.44	
OCT 24	A	- .12	- .08	.26	.92	.39	1.37	0.00
	D	1.32	1.28	2.04	1.25	.70	6.59	
NOV 19	A	.38	.05	.37	.98	.36	2.14	1.03
	D	.82	1.15	1.93	1.19	.73	5.82	
DEC 4	A	.72	.19	.44	1.00	.39	2.74	.96
	D	.48	1.01	1.86	1.17	.70	5.22	
DEC 13	A	.89	.35	.57	1.00	.37	3.18	.65
	D	.31	.85	1.73	1.17	.72	4.78	
DEC 27	A	.83	.42	.70	1.02	.38	3.35	.29
	D	.37	.78	1.60	1.15	.71	4.61	

APPENDIX C

PROGRAM LISTING AND GLOSSARY OF TERMS

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C      ******DATA REQUIRMENTS FOR FROST AND SOIL MOISTURE PROGRAM*******
C      **********************************
C      FIRST CARD REQUIRES              ****
C      WATERSHED NUMBER OR DESIGNATION,SOIL TYPE(NAME),LATITUDE IN
C      DEGREES TO THE NEAREST THOUSANDTH.SLOPE OF WATERSHED IN DEGREES TO
C      THE NEAREST WHOLE DEGREES, THE ASPECT ORIENTATION TO THE NEAREST
C      DEGREE.THE NUMBER OF SOIL LAYERS.THE NUMBER OF SOIL LAYERS SUBJECT
C      DIRECT EVAPORATION.COVER CODE, A NUMBER TO INDICATE THE PRINCIPLE
C      TYPE OF LAND USE--SEE GLOSSARY FOR FURTHER EXPLANATION OF COVER
C      CODE---.YEAR.DEPTH AT WHICH EVAPOTRANSPORATION LOSS BECOMES
C      INSENSITIVE TO DEPTH. THE FORMAT OF THE FIRST CARD IS
C      (2A4.5A4.F6.3.F3.0.3X.F3.0,2X,4I3,F5.0)  AN EXAMPLE IS
CNO 109 WELLSTON SILT LOAM   40366 11    135     5  1  4 63    30
C
C      SECOND CARDS ***FOR EACH LAYER READ A CARD****LISTING
C      WILTING POINT.FIELD CAPACITY.LAYER THICKNESS.POROSTY.CRITICAL
C      POINT AT WHICH WATER LOSS IS REDUCED FROM THE MAXIMUM RATE TO A
C      LESSER RATE PROPORTIONAL TO THE SOIL MOISTURE REMAINING.AND THE
C      PERCENT ORGANIC CONTENT.  THE FORMAT IS (2F10.2,F10.0.3F10.2)
C
C      NEXT CARD *** FOLLOWING LAYER DATA
C      THE MINIMUM OF THE MEAN MONTHLY WATER VAPOR PRESSURES,INCREMENT
C      BETWEEN MAX AND MIN MEAN WATER VAPOR PRESSURES,LATEST DAY OF
C      GROWING SEASON WHEN DROUGHT AFFECTED GRASSES STILL WILL REVIVE,
C      THE LENGTH OF THE SPRING PERIOD REQUIRED BY VEGETATION TO GO FROM
C      A DORMANT CONDITION TO GROWING SEASON TRANSPIRATION,AND THE LENGTH
C      OF THE PERIOD IN THE FALL TO GO FROM TRANSPIRATION TO DORMANCY.
C      THE FORMAT IS (2F10.2,I4,2F3.0)
C
C      NEXT CARD  ***
C      LONG TERM MONTHLY MEAN TEMPERATURES (ONE VALUE FOR EACH MONTH)
C      FORMAT(12F5.2)
C
C      NEXT CARD ***
C      EXPECTED DAY OF THE YEAR WHEN SPRING BEGINS, EXPECTED DAY OF ONSET
C      OF DORMANT CONDITIONS IN FORMAT(2I3)
C
C      NEXT CARD ****
C      THE CURRENT SOIL MOISTURE IN EACH LAYER IN PERCENT OF TOTAL VOLUME
C      THE FORMAT IS(8F10.2)
C
C      NEXT CARD ****
C      THE MAXIMUM EFFECTIVE DEPTH OF LITTER (OR HEIGHT OF GRASSY VEGETA-
C      TION).THE EFFECTIVE THICKNESS OF THE LITTER LAYER WHEN COMPRESSED
C      BY SNOW, THE EFFECTIVE THICKNESS OF THE LITTER LAYER AFTER IT HAS
C      BEEN SUBJECTED TO FREEZING TEMPERATURES, THE RELATIVE ROUGHNESS
C      FACTOR FOR GAIN OR LOSS OF SNOW DUE TO DRIFTING, AND THE MINIMUM
C      INFILTRATION CAPACITY FOR THE GIVE SOILS OR WATERSHED UNDER
C      FROZEN CONDITIONS.  FORMAT(3F5.3,F10.5,F5.3)
C
C      NEXT CARD****TEMPERATURE MIXING FACTOR TO ADJUST TO STATION TEMP-
C      ERATURE TO THE EFFECTIVE TEMPERATURE ON THE WATERSHED.AVERAGE
C      HEAT CAPACITY OF SOIL,SUMMER ALBEDO,WINTER ALBEDO,SNOW ALBEDO,
C      AVERAGE CONDUCTIVITY,THRESHOLD FOR ORGANIC MATERIAL BELOW WHICH
C      PUDDLING OF THE SOIL IS LIKELY TO OCCUR,THRESHOLD FOR ORGANIC
C      MATERIAL ABOVE WHICH PUDDLING IS UNLIKELY TO OCCUR, AND THE MINI-
C      MUM VALUE OF SOIL MOISTURE AT WHICH GRASSES CONTINUE TO TRANSPIRE-
C      THE FORMAT IS (F6.4,F12.8,3F4.2,F8.6,2F6.4.F6.2)
C
C      DAILY DATA ******
C      MONTH.DAY.YEAR.MAXIMUM DAILY AIR TEMPERATURE.MINIMUM DAILY AIR
C      TEMPERATURE, PRECIPITATION,OBSERVED EVAPORATION, MEAN DAILY
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C     DEW POINT, WIND. SOLAR RADIATION, HAYCUTTING OR HARVEST INDICATOR-
C     THE VALUE IS 1 ON A DAY OF HARVEST AND 0 THE REST OF THE TIME,
C     AND THE OBSERVED SNOW FALL IN INCHES OF SNOW.  THE FORMAT IS
C     (3I3,2F3.0,F4.3,24X,F3.2,4X,3F4.0,I1,F4.1)
C
      DIMENSION CCODE5(4), CCODE6(4), CCODE7(4), CCODE0(4)
      DIMENSION WSNO(2), SOILTP(5)
      DIMENSION CCODE(4), CCODE1(4), CCODE2(4), CCODE3(4), CCODE4(4)
      DIMENSION LABEL(8),FC(8),CRTPCT(8),SCOND(9)
      DIMENSION DLTAH(8),PROSTY(8),ORGCNT(8),WP(8),FRACJ(8)
      REAL LGTMMN(12),LAT,LATRAD,MNVPRS,LITTER
      REAL LITMAX,LITSNO,LITFRZ,LITLOS
      REAL MNASD,MNSCAT,MNSTG,NFLTRN,MNTPCY,MAXPEN,MXPNSL
      REAL LYRSMP(8),NETNYD,LWRLOS,LYSMO(8),LITMOS
      INTEGER DAOFFZ,PNCODE,TPOFFT,FREEZE,SURFAC,SUMGRS,QUAD
      INTEGER CC,DA,YR,FROST,FIRST,DAOFYR,DASINC,HC,SPRING,FALL
      INTEGER FRZFST,DAOFTW
C     *************FUNCTION STATMENTS*************************
      ANGRAD(DEG)=DEG*.01/4533
C     ***********************INITIALIZE******************************
      DATA SCOND/4HPORO,4HUS  ,4H    ,4HIMPE,4HRVIO,4HUS  ,4H    ,
     14H    ,4H    /
      DATA PI/3.1416/
      DATA LABEL/3HNOR,3HTH ,3HEAS,3HT  ,3HSOU,3HTH ,3HWES,3HT  /
      DATA CCODE/4HBARE,4H GRO,4HND  ,4H    /
      DATA CCODE1/4HSTUB,4HBLE ,4HFROM,4H CRP/
      DATA CCODE2/4HHARD,4HWOOD,4HS W/,4HO LV/
      DATA CCODE3/4HHARD,4HWOOD,4HS W ,4HLEVS/
      DATA CCODE4/4HPAST,4HURE ,4H    ,4H    /
      DATA CCODE5/4HOPEN,4HCROP,4HS(CO,4HRN) /
      DATA CCODE6/4HCLOS,4HED C,4HROPS,4H    /
      DATA CCODE7/4HEVER,4HGREE,4HN   ,4H    /
C     DISPLAY OFF
      AVHTCY=.181
C     AVHTCY IS THE AVERAGE HEAT CAPACITY IN CAL/GM DEG C
      DENSIT=2.65
      AVHTCY=DENSIT*AVHTCY
C     CONVERT TO VOLUMETRIC HEAT CAPACITY
    1 ADDPTH=0.0
      APRTFC=0.0
C     CDSNFA=16E-3
      DAOFFZ=0
      DASINC=0
      DENS=0.2
      FIRST=0
      RUNOFF=0.0
      FALDAY=0.0
      FREEZE=0
      MLTWTR=0
      FROST=0
      IRTN=0
      FRZFST=0
      IPRINT=0
      ITHW=0
      LITLOS=0.0
      LITDEX=0
      LITMOS=0.0
      PEN=0.0
      MAXPEN=0.0
      MXPNSL=0.0
      SNOMLT=0.0
      ITRNST=0
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      NFLTRN=0.0
      NOINFL=0
      RSUM=0.0
      SFCSTG=0.0
      SIGMA=1.17E-7
      SIGSQR=6500
      SNOLOS=0.0
      FALFRC=0.0
      SPRFRC=0.0
      SNODPT=0.0
      AOFSMC=0.0
      SNOW=0.0
C.    SPHTAV=.48
      SPRDAY=0.0
      THW=0.0
      SURFAC=0
      READ(5,410) (WSNO(J),J=1,2),(SOILTP(J),J=1,5),LAT,SLOPE,ASPECT,
     1NOLAY,NDLAY,CC,YR,ROOTBK
C      **********READ IN SOIL LAYERS******************************
      J1=NOLAY
      DO 4 J=1,NOLAY
    4 READ(5,411)WP(J),FC(J),DLTAH(J),PROSTY(J),CRTPCT(J),ORGCNT(J)
      CALL LAYER(NOLAY,WP,FC,DLTAH,LYRSMP,FLDCPY,WILTPT,TOTSTG,DCEN,
     1ACTLAY,NDLAY,ESFCLY,CRTPCT,CRTPNT,SFCWLT,DPTAV,SFCPOT,SFCCRT,
     2POTSTG,PROSTY,FRACJ,ROOTBK)
      DTOT=ACTLAY
      READ (5,420) MNVPRS,DLTAVP,SUMGRS,SPRPRD,FALPRD
      POTSTG=FLDCPY-WILTPT
      WRITE (6,430) (WSNO(J),J=1,2),(SOILTP(K),K=1,5)
      QUAD=1
      QUAD=IFIX((ASPECT+45.)/90.)+QUAD
      QUAD=QUAD*2-1
      IF (QUAD.GT.8) QUAD=1
      WRITE (6,435) SLOPE,LABEL(QUAD),LABEL(QUAD+1),ACTLAY
      CALL COVER (CC,CCODE,CCODE0,CCODE1,CCODE2,CCODE3,CCODE4,CCODE5,CCO
     1DE6,CCODE7)
      WRITE (6,440) FLDCPY,WILTPT
      WRITE (6,445) TOTSTG,(CCODE(J),J=1,4)
      READ (5,450) (LGTMMN(J),J=1,12)
C     CALCULATE I IN THORNETHWAITE EQUATION.
      CALL THNWTR (LGTMMN,I,A,ANMENC)
C      ***CORRECT FOR SLOPE AND ASPECT*****************
      LATRAD=ANGRAD(LAT)
      SLPRAD=ANGRAD(SLOPE)
      ASPRAD=ANGRAD(ASPECT)
      SACOR=ARSIN(COS(ASPRAD)*SIN(SLPRAD))
      IF (SACOR.GT.PI/2.0) SACOR=PI/2.0
      CORLAT=LATRAD+SACOR
C     DEPTH TO STABLE TEMPERATURE ARBITRARILY SET AT 2 FEET UNTIL THE
C     DIFFUSIVITY IS COMPUTED.
      DPDEST=24.0
C      **FIND EQUINOX VALUE OF MEAN COS Z.
      CALL EQNOXR (CORLAT,MNVPRS,DLTAVP,SIGSQR,MNSCAT,MNASD,PCPWEQ,COSZE
     1Q,PLNCO2)
      READ(5,455)SPRING,FALL
      READ(5,460)(LYSMO(J),J=1,NOLAY)
      READ(5,461)LITMAX,LITSNO,LITFRZ,RRF,MNINFL
      LITTER=LITMAX
      READ(5,470)TMIXFC,AVHTCY,ALBDOS,ALBDOW,ABDOSN,AVSCDY,ORGMN1,
     1ORGMN2,GRASMN
      GRASMN=GRASMN-WILTPT
      AOFSMC=0.0
```

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      DO 6 J=1,NOLAY
      AOESMC=AOESMC+LYSMO(J)-WP(J)
      IF(J.GT.NOLAY)GO TO 6
      SFCSTG=AOESMC
    6 LYSMO(J)=LYSMO(J)-WP(J)
      MNSTG=AOESMC
      IF (CC-3) 10,15,20
10    WRITE (6,475) ESFCLY
      HC=0
      GO TO 25
15    MNSTG=AOESMC
      GO TO 45
20    IF (FROST.LT.1) GO TO 15
      IF (CC.LT.7) GO TO 10
      GO TO 15
25    SURFAC=1
      IF(IRTN.EQ.1)GO TO 124
   45 READ(5,480,END=600)MO,DA,YR,AIRMAX,AIRMIN,PCP,EVAP,DP,WIND,SRAD,
     1HC,SNOFAL
      IF(MO.GT.12)GO TO 1
C     FOR MORE WATERSHEDS TO BE RUN WITHOUT RELOADING CODE CARD W/MO=13
   46 EVAP=EVAP*1.0
      PCP=PCP*1.0
      IF(PCP.LT.0.001)PCPW=0.0
C     ****COMPUTE DAILY MEAN TEMP AND CONVERT TO DEGREES CENTIGRADE.
      DYMNTF=(AIRMAX+AIRMIN)/2.0
      DYMNTC=(DYMNTF-32.)*.55555
   49 IF (FIRST.EQ.0.0) GO TO 80
      ADPT=(2.0*YDP+DP)/3.0
C     ***THIS STEP UNIQUE FOR COSHOCTON DATA TO ADJUST MIDNIGHT TO MID-
C     NIGHT MEAN DEW POINT TO 7AM-7AM MEAN DEW POINT.
      YDP=DP
   50 IF(DYMNTF.LE.ADPT)GO TO 85
C      CALCULATE THE DAY OF THE YEAR
      IF(AIRMAX.LT.32.0)GO TO 89
   51 CALL CALNDR(DA,MO,DAOFYR,YR)
      IF(FIRST.LT.1)GO TO 79
   52 IF(DAOFYR.EQ.FALL)FALDAY=0.0
      IF (DAOFYR.EQ.SPRING) SPRDAY=0.0
      IF (DAOFYR.GE.SPRING.AND.DAOFYR.LT.FALL) GO TO 60
      FROST=1
      FALDAY=FALDAY+1.0
55    IF (DA.EQ.1.AND.FIRST.EQ.0) GO TO 65
      IF (DA.EQ.1) GO TO 70
      IF(FIRST.EQ.0)GO TO 65
      GO TO 95
60    FROST=0
      IF(DAOFYR.GT.SPRING+SPRPRD)GO TO 55
      SPRDAY=SPRDAY+1.0
      GO TO 55
65    WRITE (6,485) MO,YR
      GO TO 75
70    WRITE (6,490) MO,YR
75    WRITE (6,495)
      WRITE (6,500)
      GO TO 95
   79 IF(DAOFYR.GT.SPRING.AND.DAOFYR.LT.FALL)GO TO 81
      GO TO 82
80    ARMXYD=AIRMAX
      ARMNYD=AIRMIN
      YDP=DP
      ADPT=DP
```

```
      GO TO 50
   81 SPRDAY=DAOFYR-SPRING
      FALDAY=FALPRD
      GO TO 52
   82 FALDAY=DAOFYR-FALL
      SPRDAY=SPRPRD
      GO TO 52
C     ADJUST MEAN DEWPOINT TO BE EQUAL TO OR LESS THAN MEAN AIR TEMP.
85    FPOT=0.0
      APRTEC=(ADPT-AIRMIN)/(AIRMAX-AIRMIN)
      DYMNTF=ADPT
      GO TO 51
C     TEST TO ESTIMATE WHETHER COLD WEATHER PRECEDES OR FOLLOWS SNOW
C     ACCUMULATION.
   89 FPOT=0.0
      IF(PCP.GT.0.1.AND.ADPT.LT.DYMNTF-5.0)FRZFST=1
      IF(DA.EQ.1)GO TO 90
      IF(FIRST.EQ.0) GO TO 91
      IF(PCP.GT.0.01.AND.AIRMAX.GT.33.0)GO TO 51
      IF(FRZFST.LT.1)GO TO 51
      GO TO 180
90    WRITE (6,490) MO,YR
      GO TO 92
   91 WRITE(6,485)
   92 WRITE (6,495)
      WRITE (6,500)
      IF(FRZFST.GT.0)GO TO 180
      GO TO 160
C      ***CORRECT FOR DAY LENGTH.
   95 CALL DALNCR(DAOFYR,CORLAT,MNVPRS,DLTAVP,PLNCO1,SIGSQR,R,DAY,SRAD
     1,LATRAD,SRCOR,SARG,UNCDAY)
      SRAD=SRAD-SRCOR
      IF(SACOR.GT.0)GO TO 94
      IF(DAOFYR.LT.80.OR.DAOFYR.GT.264)GO TO 98
      GO TO 99
   94 IF(DAOFYR.GT.80.AND.DAOFYR.LT.264)GO TO 98
      GO TO 99
   98 UNCDAY=DAY
   99 SR=(DAY/UNCDAY)*SRAD
      IF (FROST.EQ.1) GO TO 105
      IF(SPRDAY.GT.0.0.AND.DAOFYR.LT.SPRING+SPRPRD)GO TO 100
      ALBDO=ALBDOS
      GO TO 120
100   ALBDO=ALBDOS-(ALBDOW-ALBDOS)*SPRDAY/SPRPRD
      IF(SPRDAY.EQ.1)GO TO 175
      GO TO 120
  175 IF(SURFAC.LT.1)GO TO 104
      WRITE (6,580) ACTLAY
      HC=3
      SURFAC=0
  104 FROST=0
      LITTER=LITMAX
      LITDEX=0
      GO TO 120
105   IF (SNOW.GT.0.0) GO TO 115
      IF (FALDAY.GT.0.0.AND.FALDAY.LT.FALPRD) GO TO 110
      ALBDO=ALBDOW
      GO TO 120
110   ALBDO=ALBDOS+(ALBDOW-ALBDOS)*(FALDAY/FALPRD)
      GO TO 120
115   ALBDO=ABDOSN
120   SR=SR*(1-ALBDO)
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      IF (SR.EQ.0.0) GO TO 140
C     *****COMPUTE EVAPORATION
      IF(SNOW.GT.0.0)GO TO 1200
      CALL EVPNR (DYMNTF,ADPT,WIND,SR,EPOT,TMIXFC,CC,DYMNTC,DPVPRS,
     1SNOW)
      GO TO 121
 1200 EPOT=0.0
  121 EVPPOT=EPOT
      IF(HC.EQ.1.AND.FROST.EQ.0.AND.SPRDAY.GT.SPRPRD)GO TO 125
      IF(FALDAY.EQ.1)GO TO 130
      IF (EPOT.LT.0.0) GO TO 145
      IF(FROST.EQ.1.AND.FALDAY.LT.FALPRD)GO TO 124
      IF(HC.GT.1.AND.SURFAC.EQ.1) GO TO 135
      IF(FROST.EQ.0.AND.DAOFYR.LT.SPRING+SPRPRD)GO TO 124
      IF(SNOW.GT.0.001)GO TO 145
      CALL EVPADJ(EPOT,SURFAC,MNSTG,SFCSTG,CRTPNT,SFCWLT,SFCPOT,SFCCRT,
     1POTSTG,PCP)
      GO TO 145
  124 ITRNST=1
  125 CALL HAYCUT (EPOT,DASINC,HC,FALDAY,SFCSTG,FALPRD,MNSTG,SPRFRC,I
     1TRNST,POTSTG,SFCPOT,FROST,LWRLOS,SFCLOS,ACTLAY,SFCCRT,CRTPNT)
      IF(ITRNST.GT.0.AND.EPOT.GT.0.0)CALL TRNSLS(LYSMO,LWRLOS,SFCLOS,
     1NOLAY,NDLAY,LYRSMP,POTSTG,SFCSTG,MNSTG,SFCPOT)
      GO TO 145
  130 IRTN=1
      GO TO 10
  135 WRITE (6,475) ESFCLY
      GO TO 124
  140 CALL ESTVAP(PLNCO1,PLNCO2,DAY,R,DYMNTC,I,A)
      GO TO 121
  145 IRTN=0
  146 IF(SNOW)160,160,155
  150 MNSTG=0.0
      EVPDIF=0.0-EVAP
      GO TO 165
C     ******ADJUST PRECIPITATION FOR SNOW MELT
  155 CALL SNOWR(DYMNTC,PCP,SFCSTG,SNOMLT,SNOW,DPVPRS,WIND,CC,ADPT,
     1SARG,SNOLOS,AIRMAX,AIRMIN,SNODPT,APRTFC)
  160 IF(MNSTG.EQ.0.0.AND.PCP.EQ.0.0)GO TO 150
      IF(SNOMLT.GT.0.01)GO TO 1639
  165 IF (EVAP.GT.0.0) GO TO 260
      EVPDIF=0.0
  164 IF(PCP.LT.0.001)GO TO 170
 1639 IF(FRZEST.GT.0)GO TO 180
 1640 CALL PRECPR(PCP,DYMNTC,SNOW,CC,WIND,EPOT,FROST,ADDPTH,RRF,SNODPT,
     1SNOFAL,OSNDPT,SNOLOS,PCPW,AIRMAX,DENS)
      IF(SNODPT.GT.2.0.AND.FROST.GT.0.AND.LITDEX.LT.1)GO TO 1700
      GO TO 170
 1700 LITLOS=LITTER-LITSNO
      LITTER=LITSNO
      LITDEX=1
      GO TO 170
  170 IF(FRZEST.GT.0)GO TO 1801
      IF(DYMNTC.LT.1.0)GO TO 172
  169 IF(FREEZE-1)172,171,171
  171 IF(PNCODE-2)173,172,172
  173 IF(TRUEFT.NE.3)GO TO 172
      RUNOFF=PCP+SNOMLT
      IF(SNOW.GT.0.0)GO TO 174
      WRITE(6,561) PCP
      GO TO 179
  174 IF(LITTER.GT.0.0)GO TO 176
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      WRITE(6,562)PCP
      MLTWTR=1
      GO TO 179
  176 LITMOS=LITMOS+PCP
      WRITE(6,563)PCP
      IF(FRZFST.GT.0)GO TO 1801
      GO TO 130
  172 IF(ITRNST.GT.1.AND.EPOT.GT.PCP)GO TO 180
      CALL SMP(NOLAY,MNSTG,LYRSMP,LYSMO,EPOT,PCPW,SFCSTG,DA,POTSTG,
     1ADESMC,DCEN,DTOT,AIRMAX,FIRST,SURFAC,NDLAY,DPTAV,DLTAH,SFCPOT,
     1MLTWTR,SNOMLT,RUNOFF,FRACJ)
      IF(MNSTG.LT.GRASMN.AND.DAOFYR.GT.SUMGRS)GO TO 350
      GO TO 180
  179 EVPDIF=0.0
180   MNTPCY=(((ARMXYD+ARMNYD)/2.0)-32.0)*.5555
      FREZTP=(MNTPCY+DYMNTC)/2.0
      IF (FREZTP.GT.0.0.AND.FREEZE.LT.1)GO TO 380
      CALL FRFZR(FREEZE,DYMNTC,MNSTG,POTSTG,SFCSTG,SNOW,NOLAY,AVHTCY,
     1PEN,THW,DPOFST,ORGMN1,DLTAH,SFCPOT,ORGCNT,PNCODE,TPOFFT,PROSTY,
     2NDLAY,MNTPCY,ANMENC,RUNOFF,DAOFFZ,LITTER,ORGMN2,MLTWTR,LYSMO,
     3DCEN,AIRMIN,AVSCDY,DAOFYR,LITMOS,NFTNYD,MAXPEN,MXPNSL,ADDPTH,
     4SNOLOS,THWMAX,WP,NOINFL,SNODPT,RSUM,PENONM,MNINFL,DAOFTW,THWSUM,
     5FSTSUM,LITLOS,DENS)
      IF(MXPNSL.GT.1.0.AND.FROST.GT.0.AND.LITDEX.LT.2)GO TO 1705
 1803 IF(FRZFST.GT.0)GO TO 1640
 1801 IPRINT=1
C     IPRINT IS AN INDEX TO HAVE PROGRAM PRINT DAILY DATA BEFORE MESSAGE
C        ON FROST
      IF(FRZFST.GT.0)WRITE(6,1802)MXPNSL,SNODPT,LITTER
 1802 FORMAT(20H SOIL PENETRATION    ,F6.2,13H SNOWDEPTH    ,F6.2,10H LITT
     1ER    ,F6.2)
      FRZFST=0
      GO TO 381
 1705 LITLOS=LITTER-LITFRZ
      LITTER=LITFRZ
      LITDEX=2
      GO TO 1803
  181 IF (PNCODE-2) 190,185,240
185   WRITE (6,505)
      GO TO 380
190   IF (TPOFFT.EQ.0) GO TO 380
      IF (TPOFFT-2) 195,205,210
195   WRITE (6,510)
  200 WRITE(6,515)DAOFFZ,MXPNSL
      FREEZE=1
      GO TO 380
205   WRITE (6,520)
      GO TO 200
210   IF (TPOFFT-4) 215,220,225
215   WRITE (6,525)
      WRITE (6,530)
      GO TO 200
220   WRITE (6,535)
      WRITE(6,540)MXPNSL
      GO TO 380
225   IF (TPOFFT.GT.5) GO TO 230
      WRITE(6,545)THW,MXPNSL
      GO TO 380
230   IF (TPOFFT.GT.6) GO TO 235
      WRITE (6,550)
      FREEZE=0
      PNCODE=0
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      TPOFFT=0
      GO TO 380
  235 IF(TPOFFT.GT.7)GO TO 236
      WRITE (6.555)
      GO TO 200
  236 IF(TPOFFT.GT.8)GO TO 237
      WRITE(6.556)
      GO TO 200
240   IF (PNCODE-8) 245.250.250
  237 WRITE(6.557)
      WRITE(6.558)THWMAX.MXPNSL
      GO TO 380
245   WRITE (6.560)
      GO TO 380
  250 IF(PNCODE.LT.20)GO TO 255
      WRITE(6.565)PNCODE
      GO TO 380
  255 WRITE(6.570)PNCODE
      GO TO 380
260   CALL EVAPR(EVPPOT.EVAP.EVPDIF)
      GO TO 164
350   IF (CC.GE.4.AND.CC.LT.6) GO TO 355
      GO TO 180
355   IPRCC=CC
      CC=0
      SURFAC=1
      WRITE (6.585)
      WRITE(6.475) ESFCLY
      IF(SFCSTG.LT.0.0)SFCSTG=0.0
      GO TO 180
  380 IF(IPRINT.GT.0)GO TO 386
  381 ARMXYD=AIRMAX
      ARMNYD=AIRMIN
      NETNYD=RUNOFF
  385 J1=NOLAY
      IF(NOLAY.GT.5)J1=5
      J2=1
      IF(FREEZE.LT.1)J2=7
      IF(NOINFL.EQ.1.AND.PNCODE.EQ.1)J2=4
      J3=J2+2
      IF (FALDAY.GT.FALPRD.AND.SPRDAY.GT.SPRPRD) ITRNST=0
      WRITE(6.595)MO.DA.AIRMAX.AIRMIN.PCP.EVPPOT.EVAP.EVPDIF.MNSTG.
     1(LYSMO(J).J=1.J1).WIND.SNOW.ADPT.SRAD.(SCOND(J).J=J2.J3)
      IF(IPRINT.GT.0)GO TO 181
  386 FIRST=1
      SNOMLT=0.0
      APRTEC=0.0
      PCPW=0.0
      IPRINT=0
      RUNOFF=0.0
      IF(DYMNTC.LT.-3.0)GO TO 390
C     GENERAL FREEZE CONDITION WHICH SEVERELY DAMAGES VEGETATION.
      IF (SPRDAY.EQ.1.0.AND.CC.EQ.2) GO TO 405
      GO TO 45
390   IF (CC.EQ.5.OR.CC.EQ.6) GO TO 400
      IF (CC.EQ.3.AND.FALDAY.EQ.1.) GO TO 395
      GO TO 45
395   IPRCC=CC
      GO TO 10
400   IPRCC=CC
      CC=2
      GO TO 10
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405    IF (SPRDAY.LT.SPRPRD) GO TO 45
       CC=IPRCC
       SURFAC=0
       WRITE (6,580) ACTLAY
       GO TO 45
  600  STOP
C
420    FORMAT (2F10.2,I4,2F3.0)
  410  FORMAT(2A4,5A4,F6.3,F3.0,3X,F3.0,2X,4I3,F5.0)
  411  FORMAT(2F10.2,F10.0,3F10.2)
425    FORMAT (4(4A4))
430    FORMAT (35H1                SOIL MOISTURE ON WATERSHED ,2A4,20H HAVING A
      1 SOIL TYPE ,5A4)
435    FORMAT (46H0     THIS AREA OF THE WATERSHED HAS A SLOPE OF,F7.2,2X,
      117H DEGREES. FACING ,2A3,30H WITH AN ACTIVE ROOT LAYER OF ,
      2F8.2,8HINCHES. )
440    FORMAT (42H FIELD CAPACITY FOR THIS LAYER OF SOIL IS ,F6.2,36H INC
      1HES. THE WILTING POINT IS ABOUT ,F6.2,8H INCHES.)
445    FORMAT (27H THE LAYER IS SATURATED BY ,F6.2,20H INCHES.  THE COVER
      1 ,10H TYPE IS  ,4A4)
450    FORMAT (12F5.2)
  455  FORMAT(2I3)
  460  FORMAT(8F10.2)
  461  FORMAT(3F5.3,F10.5,F5.3)
  470  FORMAT(F6.4,F12.8,3F4.2,F8.6,2F6.4,F6.2)
475    FORMAT (42H THE ACTIVE EVAPORATION LAYER IS NOW ONLY ,F6.2,9H INCH
      1ES. )
  480  FORMAT(3I3,2F3.0,F4.3,24X,F3.2,4X,3F4.0,I1,F4.1)
485    FORMAT (1H0,30X,7H MONTH ,I3,6H OF 19,I2)
490    FORMAT (1H1,30X,6HMONTH ,I3,6H OF 19,I2)
  495  FORMAT(55H0DATE AIR TEMPERATURE  PRECIP-  EVAPORATION   SOIL MOI,
      157HSTURE       LAYERS                                 DEW  SOLAR)
  500  FORMAT(55H MO DA    MAX    MIN     ITATION COMP  OBS  DIF  TOTAL    ,
      161H  1       2       3       4       5    WIND  SNOW POINT RADIATION,
      113H PERMIABILITY)
505    FORMAT (52H  THE GROUND IS PROTECTED FROM FREEZING BY THE SNOW ,27
      1H  COVER, OR LITTER OR BOTH.)
510    FORMAT (55H  THE GROUND IS FREEZING BUT CONDITIONS FAVOR FORMATION
      1,31H OF MAINLY GRANULAR TYPE FROST.)
515    FORMAT (32H  FREEZING HAS BEEN PROGRESS FOR,I4,8H DAYS.   ,23HPENET
      1RATION IS DOWN TO ,F6.1,9H  INCHES.)
520    FORMAT (54H  CONDITIONS FAVOR THE FORMATION OF NEEDLE ICE.  FROST,
      124H HEAVING IS VERY LIKELY.)
525    FORMAT (51H  CONDITIONS FAVOR THE FORMATION OF IMPERVIOUS ICE ,55H
      1LENSES.  ***SHOULD RAIN OCCUR BEFORE THE GROUND THAWS.)
530    FORMAT (51H  LITTLE INFILTRATION WOULD OCCUR IN THIS AREA AND ,53H
      1HIGH RUNOFF PEAKS MAY BE EXPECTED.*********************)
535    FORMAT (54H  SOME MELTING OF THE SNOW HAS TAKEN PLACE AND THAWING,
      161HHAS OCCURRED FROM THE BOTTOM OF THE FROZEN ZONE.  PENATRATION)
540    FORMAT (15H NOW IS ONLY TO,F6.1,8H INCHES.)
545    FORMAT (55H  SOME THAWING HAS TAKEN PLACE BUT THE GROUND IS STILL
      1,12HFROZEN FROM ,F6.1,8H DOWN TO,F6.1,8H INCHES.)
550    FORMAT (44H  THE GROUND IS NOW GENERALLY FREE OF FROST.)
555    FORMAT (53H  THE GROUND IS FREEZING BUT CONDITIONS MAKE THE KIND,2
      14H OF FROST INDETERMINATE.)
  556  FORMAT(57H THERE IS STILL FROST IN THE LITTER BUT NONE IN THE SOIL
      1.)
  557  FORMAT(123H SNOW AND LITTER HAVE COMBINED WITH AIR TEMPERATURES T0
      1 ALLOW LITTLE CHANGE IN THE FROST.  THE GROUND IS STILL FROZEN FR0
      2M )
  558  FORMAT(F6.2,4H TO ,F6.2,8H INCHES.)
560    FORMAT (56H THERE IS SOMETHING WRONG IN THE FREEZE-THAW SUBROUTINE
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     1.)
 561 FORMAT(F6.2.85H OF RAIN OR WATER FROM MELTED WET SNOW HAS NOT BEEN
     1ABLE TO INFILTRATE AND HAS RUNOFF.)
 562 FORMAT(F6.2.90H OF RAIN HAS GONE INTO THE SNOW AND WILL PROBABLY D
     1RAIN AWAY. THE GROUND BEING IMPERVIOUS.)
 563 FORMAT(F6.2.60H INCHES OF PRECIPITATION IS ADDED TO MOISTURE IN TH
     1E LITTER.)
565   FORMAT (50H  SOIL MOISTURE PERCENTAGE CALCLATIONS UNSUCSESFL.,I4)
 570 FORMAT(40H  PENETRATION EQUATION DID NOT CONVERGE.,I4)
575   FORMAT (50H WATER FOR RUNOFF OR GROUND WATER RECHARGE EQUALS ,F6.2
     1,10H  INCHES. )
 580 FORMAT(49H TRANSPIRATION IS BEGINNING AGAIN FROM THE ACTIVE,16H R
     1OOT ZONE OF   ,F6.2,8H INCHES.)
585   FORMAT (46H PASTURE GRASSES OR CROPS HAVE WILTED AND THIS,54H LAT
     1 IN THE SEASON ARE NOTE EXPECTED TO SIGNIFICANTLY,29H TRANSPIRE AN
     2YMORE THIS YEAR.)
590   FORMAT (43H THERE IS AN ERROR IN THE SOIL DESCRIPTION.)
 595 FORMAT(2I3,2F6.1,6X,F6.2,3F5.2,6F7.2,F7.0,3F7.2,3X,3A4)
      END
      SUBROUTINE EVAPR(EVPPOT,EVAP,EVPDIF)
      EVAP=EVAP*.7
      IF(EVPPOT.LT.0.0)GO TO 5
      EVPDIF=EVPPOT-EVAP
      GO TO 10
5     EVPDIF=0.0-EVAP
10    RETURN
      END
      SUBROUTINE SNOWR(DYMNTC,PCP,SFCSTG,SNOMLT,SNOW,DPVPRS,WIND,CC,
     1ADPT,SARG,SNOLOS,AIRMAX,AIRMIN,SNODPT,APRTFC)
      INTEGER CC
      VAPRS(DPT)=EXP(-7482./(DPT+398.36))*6.414E6
      SNOMLT=0.0
      RADMLT=0.0
      RANMLT=0.0
      YSNDPT=SNODPT
      YSNOW=SNOW
      CNDMLT=0.0
      SNOLOS=0.0
      DPVPRS=EXP(-7482/(ADPT+398.36))*6.414E6
      IF(AIRMAX.LT.32.)GO TO 50
      IF (CC-3) 5,10,15
5     IF (CC.EQ.2) DYMNTC=DYMNTC-1
      GO TO 20
10    DYMNTC=DYMNTC-3
      GO TO 20
15    IF (CC.EQ.7) DYMNTC=DYMNTC-3
  20 DYMXTC=(AIRMAX-32.0)*0.55555
      DMINTC=(AIRMIN-32.0)*.555555555
      IF(APRTFC.GT.0.5)GO TO 21
      WMHFTM=(3.0*DYMXTC+DMINTC)/4.0
      CDHFTM=(3.0*DMINTC+DYMXTC)/4.0
      GO TO 201
  21 ADPTC=(ADPT-32.0)*0.5555555
      WMHFTM=(DYMXTC+ADPTC)/2.0
      CDHFTM=(DYMNTC+DMINTC)/2.0
 201 IF(PCP.GT.0.01)GO TO 22
      DASNMT=(0.108+0.036*SIN(SARG))*WMHFTM
      EVSNMT=(0.108+0.036*SIN(SARG))*CDHFTM
      IF(EVSNMT.LT.0.0)EVSNMT=0.0
      SNOMLT=DASNMT+EVSNMT
      IF(SNOMLT.LT.0.0)SNOMLT=0.0
      GO TO 23
```

```
   22 RANMLT=0.5*PCP*WMHFTM/80.0
      RANNIT=0.5*PCP*CDHFTM/80.0
      IF(RANNIT.LT.0.0)RANNIT=0.0
      RANMLT=RANMLT+RANNIT
      IF(RANMLT.LT.0.0)RANMLT=0.0
      RAD1=0.24*WMHFTM
      RAD2=0.24*CDHFTM
      IF(RAD2.LT.0.0)RAD2=0.0
      RADMLT=RAD1+RAD2
      IF(RADMLT.LE.0.0)RADMLT=0.0
      SNOMLT=SNOMLT+RANMLT+RADMLT
   23 IF(ADPT.GT.AIRMIN)GO TO 24
      CNDMLT=4.89E-2*WIND*(VAPRS(ADPT)-0.18)
      IF (CNDMLT) 30,30,25
   24 WMHFDT=ADPT+(ADPT-AIRMIN)
      CDHFDT=AIRMIN
      CNDWMT=2.45E-2*WIND*(VAPRS(WMHFDT)-0.18)
      CNDCMT=2.45E-2*WIND*(VAPRS(CDHFDT)-0.18)
      IF(CNDCMT.LT.0.0)CNDCMT=0.0
      CNDMLT=CNDWMT+CNDCMT
25    SNOMLT=SNOMLT+CNDMLT
      IF(SNOMLT.LT.0.0)SNOMLT=0.0
      GO TO 35
   30 CNDMLT=0.0
35    IF (SNOMLT.GT.SNOW) GO TO 40
      SNOW=SNOW-SNOMLT
      GO TO 45
   40 SNOMLT=SNOW
      SNOW=0.0
   45 SNODNS=YSNOW/YSNDPT
      SNOLOS=SNOMLT/SNODNS
      SNODPT=SNODPT-SNOLOS
      GO TO 54
   50 SNODNS=SNOW/SNODPT
      IF(DYMNTC.LT.-6.0)GO TO 55
   54 SNODNS=0.2
   55 IF(SNODNS.LT.0.09)GO TO 58
      SNODPT=SNOW/SNODNS
      GO TO 60
   58 SNODPT=10.0*SNOW
   60 OSNDPT=SNODPT
      IF(YSNDPT.GT.SNODPT)GO TO 61
      SNOLOS=0.0
      GO TO 62
   61 SNOLOS=YSNDPT-SNODPT
   62 RETURN
      END
      SUBROUTINE EQNOXR (CORLAT,MNVPRS,DLTAVP,SIGSQR,MNSCAT,MNASD,PCPWEQ
     1,COSZEQ,PLNCO2)
      REAL MNSCAT,MNASD,MNVPRS
      COSZEQ=COS(CORLAT)*0.63662
      PCPWEQ=MNVPRS+DLTAVP*EXP(-((120)**2)/SIGSQR)
      MNASD=-0.089*(1/COSZEQ)**.75-.174*(PCPWEQ/(COSZEQ*20.0))**.6
      MNSCAT=-0.083*((1/COSZEQ)**.9)
      PLNCO2=EXP(MNASD+MNSCAT)+.5-.5*EXP(MNSCAT)
      RETURN
      END
      SUBROUTINE DALNCR (DAOFYR,CORLAT,MNVPRS,DLTAVP,PLNCO1,SIGSQR,R,DAY
     1,SRADW,LATRAD,SRCOR,SARG,UNCDAY)
      REAL MNVPRS,MNCSZ,LATRAD
      INTEGER DAOFYR
      SARG=((6.2831/365.)*(DAOFYR-80.))
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```
      RARG=((6.2831/365.)*(DAOFYR-90.))
      DFC=23.5*SIN(SARG)
      ADEC=DEC/57.29557
      R=1.0+.01673*SIN(RARG)
      ANG=ARCOS(-TAN(CORLAT)*TAN(ADEC))
      HAFDA=ANG*24./6.2832
      DAY=2.0*HAFDA
      UNCANG=ARCOS(-TAN(LATRAD)*TAN(ADEC))
      UNCDAY=2.0*UNCANG*24./6.2832
      MNCSZ=SIN(CORLAT)*SIN(ADEC)+COS(CORLAT)*COS(ADEC)*0.63662
      COSZ=SIN(LATRAD)*SIN(ADEC)+COS(LATRAD)*COS(ADEC)*.63662
      SRCOR=MNCSZ/COSZ
      IF (SRADW.GT.0.0) GO TO 10
      PCPWTR=MNVPRS+DLTAVP*EXP(-((DAOFYR-200.)**2)/SIGSQR)
      ABSRBD=-0.089*(1/MNCSZ)**.75-.174*(PCPWTR/(MNCSZ*20.0))**.6
      SCTRD=-.0834*((1/MNCSZ)**.9)
C      COMPUTE PATH LENGTH CORRECTION.
      PLNCO1=EXP(ABSRBD+SCTRD)+.5-.5*EXP(SCTRD)
10    RETURN
      END
      SUBROUTINE CALNDR(DA.MO.DAOFYR,YR)
      INTEGER DA.DAOFYR
      IF (MO.LT.7) GO TO 35
      IF (MO.EQ.9.OR.MO.EQ.11) GO TO 25
      IF (MO-10) 5,15,20
5     IF (MO.GT.7) GO TO 10
      DAOFYR=DA+181
      GO TO 70
10    DAOFYR=DA+212
      GO TO 70
15    DAOFYR=DA+273
      GO TO 70
20    DAOFYR=DA+334
      GO TO 70
   25 IF(MO.EQ.11)GO TO 30
      DAOFYR=DA+243
      GO TO 70
30    DAOFYR=DA+304
      GO TO 70
   35 IF(MO.EQ.4.OR.MO.EQ.6)GO TO 60
      IF (MO.EQ.2) GO TO 55
      IF (MO-3) 40,45,50
40    DAOFYR=DA
      GO TO 70
45    DAOFYR=DA+59
      GO TO 70
50    DAOFYR=DA+120
      GO TO 70
   70 A=(YR+1900.)/4.
      I=A
      IF((A-I).LT.0.1)GO TO 72
      IF(MO.LT.3)GO TO 72
   71 DAOFYR=DAOFYR+1
      GO TO 72
55    DAOFYR=DA+31
      GO TO 70
60    IF (MO.EQ.6) GO TO 65
      DAOFYR=DA+90
      GO TO 70
65    DAOFYR=DA+151
      GO TO 70
   72 RETURN
```

```
      END
      SUBROUTINE PRECPR(PCP.DYMNTC.SNOW.CC.WIND.EPOT.FROST.ADDPTH.RRF.
     1SNODPT.SNOFAL.OSNDPT.SNOLOS.PCPW.AIRMAX.DENS)
      INTEGER CC.FROST
      WNDFCT(SNODPT)=0.1+0.0061*SNODPT-0.00065*SNODPT*SNODPT
      IF(SNOFAL.LE.0.001)GO TO 89
      IF(SNOW.LE.0.0)GO TO 50
       YSNDPT=SNODPT
      IF(DYMNTC-1.0)40.79.79
   40 OSNDSN=SNOW/SNODPT
      IF(OSNDNS.LT.0.2.AND.DYMNTC.LT.-8.0)GO TO 42
      OSNDNS=0.2
      GO TO 45
   46 ADDPTH=SNOFAL
      IF(SNOW.LT.0.001)OSNDPT=0.0
      IF(SNOFAL-0.1.GT.PCP)GO TO 4610
      SNOW=SNOW+0.1*SNOFAL
      PCPW=PCP-SNOW
      GO TO 75
 4610 PCPW=0.0
      SNOW=SNOW+PCP
      GO TO 75
   42 OSNDNS=0.15
   45 OSNDPT=SNOW/OSNDNS
   47 ADDPTH=SNOFAL
      IF(DYMNTC.LT.-8.0)GO TO 48
      IF(AIRMAX.LT.32.0)GO TO 48
      SNOW=SNOW+0.2*SNOFAL
      PCPW=PCP-0.2*SNOFAL
      IF(PCPW.LT.0.0)PCPW=0.0
      IF(0.2*SNOFAL.GT.PCP)GO TO 475
      GO TO 75
  475 SNOW=SNOW+PCP-0.2*SNOFAL
      GO TO 75
   48 SNOW=SNOW+PCP
      PCPW=0.0
      GO TO 75
   50 OSNDPT=0.0
      GO TO 47
   75 SNODPT=OSNDPT+ADDPTH
      WNDFAC=RRF
      DENS=SNOW/SNODPT
   76 SNODPT=SNODPT-(WNDFAC*SNODPT*WIND*0.01)
      SNOW=SNODPT*DENS
      EPOT=0.0
      OSNDPT=SNODPT
      IF(YSNDPT.GT.SNODPT)GO TO 80
      GO TO 81
   80 SNOLOS=SNOLOS+YSNDPT-SNODPT
      ADDPTH=0.0
      GO TO 90
   81 ADDPTH=SNODPT-YSNDPT
      GO TO 90
   79 IF(SNOFAL.GT.0.0)GO TO 46
   89 PCPW=PCP
   90 RETURN
      END
      SUBROUTINE HAYCUT (EPOT.DASINC.HC.FALDAY.SFCSTG.FALPRD.MNSTG.SPRHR
     1C.ITINST.POTSTG.SFCPOT.FROST.LWRLOS.SFCLOS.ACTLAY.SFCCRT.CRTPNT)
      REAL LWRFRC.LWRLOS
      REAL LWRWTR
      INTEGER DASINC.HC.FROST
```

```
      HAYCTF(DASINC)=.565+.015*DASINC
      IF (HC.GT.1.OR.ITRNST.EQ.1) GO TO 5
      IF (DASINC.GT.0.AND.DASINC.LE.30) GO TO 10
      EPOT=EPOT*.80
      DASINC=DASINC+1
      GO TO 20
5     ITRNST=1
      LWRFRC=(MNSTG-SFCSTG)/(POTSTG-SFCPOT)
      SFCFRC=SFCSTG/SFCPOT
      FALFRC=FALDAY/FALPRD
      SFCWTR=SFCSTG-SFCCRT
      IF(SFCWTR.LE.0.0)SFCWTR=0.0
      LWRWTR=MNSTG-SFCSTG-(CRTPNT-SFCCRT)
      IF(LWRWTR.LE.0.0)LWRWTR=0.0
      IF(LWRFRC.GT.1.0)LWRFRC=1.0
      IF(SFCFRC.GT.1.0)SFCFRC=1.0
      IF(FALFRC.GT.1.0)FALFRC=1.0
      IF(SPRFRC.GT.1.0)SPRFRC=1.0
      IF (FROST.GT.0) GO TO 15
      LWRLOS=EPOT*SPRFRC*0.55
C     CONSTANT OF 0.55 ASSUMES A 6 LAYER PROFILE WITH THE TOP LAYER
C     TAKING AN EXTRA AMOUNT OF EVAPORATION LOSS, UNDER FIELD CAPACITY
C     LEAVING 0.55 OF THE LOSS TO LOWER LAYERS.
      SFCLOS=EPOT-LWRLOS
      DIF=SFCLOS-SFCWTR
      IF(DIF.LE.0.0)GO TO 7
      SFCLOS=SFCWTR+DIF*SFCFRC
    7 DIF=LWRLOS-LWRWTR
      IF(DIF.LE.0.0)GO TO 8
      LWRLOS=LWRWTR+DIF*LWRFRC
    8 IF(SPRFRC.EQ.1)GO TO 25
      GO TO 20
   25 ITRNST=0
      GO TO 20
10    EPOT=EPOT*HAYCTF(DASINC)
      DASINC=DASINC+1
      IF (DASINC.EQ.30) DASINC=0
      GO TO 20
   15 LWRLOS=EPOT*FALFRC*0.55
      SFCLOS=EPOT-LWRLOS
      DIF=SFCLOS-SFCWTR
      IF(DIF.LE.0.0)GO TO 17
      SFCLOS=SFCWTR+DIF*SFCFRC
   17 DIF=LWRLOS-LWRWTR
      IF(DIF.LE.0.0)GO TO 18
      LWRLOS=LWRWTR+DIF*LWRFRC
   18 IF(FALFRC.EQ.1)ITRNST=0
   20 EPOT=SFCLOS+LWRLOS
      RETURN
      END
      SUBROUTINE EVPNR (DYMNTF,ADPT,WIND,SR,EPOT,TMIXFC,CC,DYMNTC,
     1DPVPRS,SNOW)
      INTEGER CC
      IF (CC-1) 20,5,10
5     WIND=.9*WIND
      GO TO 20
10    IF (CC-4) 15,20,20
15    SR=SR*.6
      DYMNTF=DYMNTF-1.0
      IF (DYMNTF.LT.ADPT) ADPT=DYMNTF
      DYMNTC=(DYMNTF-32.)*.555555
20    IF (DYMNTC.LT.0.0) GO TO 25
```

```
 21   TA=DYMNTF+398.363
      TSAT=ADPT+398.363
      WINDFC=0.37+.0041*WIND
      SATVP=1./EXP(7482.6/TSAT)
      AIRVP=1./EXP(7482.6/TA)
      IF(AIRVP.LE.SATVP)GO TO 25
      DELTVP=(6.41334E6*(AIRVP-SATVP))**.88
      IF(SNOW.GT.0.0)GO TO 25
      PRTNUM=.0105*DELTVP*WINDFC
      ARGNUM=(DYMNTF-212.)*(0.1024-0.01066*ALOG(SR))
      FSTNUM=EXP(ARGNUM)-.0001
      DENOM=0.015+(1./(TA*TA))*(6.8554E10)*SATVP
      EPOT=((FSTNUM+PRTNUM)/DENOM)*TMIXFC
 25   RETURN
      END
      SUBROUTINE THNWTR (LGTMMN,I,A,ANMENC)
      DIMENSION TEMNC(12)
      REAL LGTMMN(12),I
      TSUM=0.0
      DO 5 J=1,12
C      CONVERT TO DEG C
      TEMNC(J)=(LGTMMN(J)-32.0)*.55555
 5    TSUM=TSUM+TEMNC(J)
      ANMENC=TSUM/12.0
      TSUM=TSUM/5.0
      I=TSUM**1.514
      A=(0.675*I*I*I-77.1*I*I+492390)*.000001
      RETURN
      END
      SUBROUTINE COVER(CC,CCODE,CCODE0,CCODE1,CCODE2,CCODE3,CCODE4,
     1CCODE5,CCODE6,CCODE7)
      DIMENSION CCODE(4),CCODE0(4),CCODE1(4),CCODE2(4),CCODE3(4)
      DIMENSION CCODE4(4),CCODE5(4),CCODE6(4),CCODE7(4)
      INTEGER CC
      IF(CC.GT.5)GO TO 509
      IF(CC-1)501,502,503
 501  DO 601 J=1,4
 601  CCODE(J)=CCODE0(J)
      GO TO 488
 502  DO 602 J=1,4
 602  CCODE(J)=CCODE1(J)
      GO TO 488
 503  IF(CC-3)505,506,507
 505  DO 605 J=1,4
 605  CCODE(J)=CCODE2(J)
      GO TO 488
 506  DO 606 J=1,4
 606  CCODE (J)=CCODE3(J)
      GO TO 488
 507  IF(CC-5)511,512,509
 511  DO 611 J=1,4
 611  CCODE(J)=CCODE4(J)
      GO TO 488
 512  DO 612 J=1,4
 612  CCODE(J)=CCODE5(J)
      GO TO 488
 509  IF(CC.GT.6)GO TO 513
      DO 614 J=1,4
 614  CCODE(J)=CCODE6(J)
      GO TO 488
 513  DO 613 J=1,4
 613  CCODE(J)=CCODE7(J)
```

```
  488 RETURN
      END
      SUBROUTINE FREZR(FREEZE,DYMNTC,MNSTG,POTSTG,SFCSTG,
     1SNOW,NOLAY,AVHTCY,PEN,THW,DPOFST,ORGMN1,DPOFLY,SFCPOT,
     2ORGCNT,PNCODE,TPOFFT,PROSTY,NOLAY,MNTPCY,ANMENC,
     3NFLTRN,DAOFFZ,LITTER,ORGMN2,MLTWTR,LYSMO,DCEN,AIRMIN,AVSCDY,
     4DAOFYR,LITMOS,NFTNYD,MAXPEN,MXPNSL,ADDPTH,SNOLOS,THWMAX,WP,NOINFL,
     5SNODPT,RSUM,PENDNM,MNINFL,DAOFTW,THWSUM,FSTSUM,LITLOS,DENS)
      DIMENSION DPOFLY(NOLAY),PCTMOS(8),PROSTY(NOLAY)
      DIMENSION ORGCNT(NOLAY),DCEN(NOLAY),WP(NOLAY)
      INTEGER DAFYRY
      INTEGER  DAOFYR
      REAL LITMOS,NFTNYD,LITK
      REAL LYSMO(NOLAY),MAXPEN,MXPNSL
      REAL KWATSL,KSOLAR,KSOLIC,NFLTRN,LITTER
      REAL KSOHUM,LITLOS
      REAL LNUM,L,KSUM,MNSTG,MNTPCY
      INTEGER DAOFFZ,TPOFFT,PNCODE,FREEZE,DAOFTW
      INTEGER DATWMX,DAFZMX,YTPOFT
C     *********CALCULATE THE SOIL MOISTURE FOR EACH LAYER**********
      DO 5 J=1,NOLAY
    5 PCTMOS(J)=(LYSMO(J)+WP(J))/DPOFLY(J)
      IF(FREEZE.EQ.1.AND.DYMNTC.GT.0.0)GO TO 1290
      IF(DYMNTC.GT.-1.0.AND.PNCODE.NE.1)GO TO 115
      IF (FREEZE.EQ.1) GO TO 45
      CDYDS=1.0E-2
C     CDYDS IS THE CONDUCTIVITY OF DRY SOIL AND IS EQUAL TO 7.0 E-3 CAL
      MLTSM=0
      DAOFFZ=0
      KSOLIC=.9
      KWATAR=1.4678
      KSOHUM=.0335
      KSOLAR=.0253
      KWATSL=.322
      CDYWTR=1.42E-3
      CDYAIR=0.062E-3
      CDYICE=5.2E-3
      CDYHUM=.00006
      CHTHUM=.6
      MXPNSL=0.0
      MAXPEN=0.0
      DAOFTW=0
      NOINFL=0
      IF(AVSCDY.GT.0.0)CDYDS=AVSCDY
      DATWMX=0
      MLTWTR=0
      DAFZMX=0
      THWMAX=0.0
      TPOFFT=0
      FTSMMX=0.0
      THWSUM=0.0
      FSTSUM=0.0
      DPTTSM=0.0
      RSUM=0.0
      YTPOFT=0
C     ***CALCULATE THE FROST PENATRATION PARAMETERS**********************
   45 DATWMX=DAOFTW
      IF(SNOLOS.GT.0.0)GO TO 46
   47 THWMAX=THW
      THW=0.0
      THSMMX=THWSUM
      THWSUM=0.0
```

```
      DAOFFZ=DAOFFZ+1
      IF (SNOW.GT.0.0) GO TO 70
      MAXPEN=MAXPEN-SNOLOS
      SNOLOS=0.0
      RSUM=0.0
      SNODPT=0.0
      IF (LITTER.GT.0.0) GO TO 120
      KSUM=0.0
      CNUM=0.0
      FSTSUM=FSTSUM-DYMNTC-2.0
      PNDNMK=ANMENC+(FSTSUM/(2.0*DAOFFZ))
      DPTTSM=0.0
      GO TO 49
   46 MXPNSL=(MXPNSL+LITTER+SNODPT)*2.54
      IF(MXPNSL.LT.0.0)GO TO 48
      IF(KSUM.LT.0.0)GO TO 48
      FSTPRM=MXPNSL*MXPNSL*PENDNM/(KSUM*8.64E4)
      IF(FSTSUM.GT.FSTPRM)FSTSUM=FSTPRM
      FTSMMX=FSTSUM
      MAXPEN=MAXPEN-SNOLOS
      MXPNSL=MAXPEN-SNODPT-LITTER
      SNOLOS=-0.001
      GO TO 47
   48 FSTSUM=0.0
      FTSMMX=0.0
      GO TO 47
   49 LNUM=0.0
   50 J=0
   51 J=J+1
      JLAY=0
  501 JLAY=JLAY+1
      IF(JLAY.LT.DPOFLY(J))GO TO 502
      IF(J.LE.NOLAY)GO TO 51
      PNCODE=9
      GO TO 115
  502 DPTH2=DPTTSM
      SLCYFT=PCTMOS(J)*CDYWTR+KWATSL*(1.-PROSTY(J))*CDYDS
      DPTTSM=DPTTSM+1.0
      SLCYSD=(PROSTY(J)-PCTMOS(J))*CDYAIR*KWATAR
      S1LDNM=PCTMOS(J)+(1.-PROSTY(J))*KWATSL
      SOLDNM=S1LDNM+(PROSTY(J)-PCTMOS(J))*KWATAR
      SOLCNY=(SLCYFT+SLCYSD)/SOLDNM
      SPCHET=(PCTMOS(J)+AVHTCY*(1-PROSTY(J)))
      LNUM=LNUM+PCTMOS(J)*80.0
      CNUM=CNUM+SPCHET
      RSUM=RSUM+1.0/SOLCNY
      KSUM=DPTTSM/RSUM
  512 L=LNUM/DPTTSM
      C=CNUM/DPTTSM
C     THERE ARE 8.64E4 SEC/DAY
      IF(DPOFST.NE.24.0)GO TO 52
C     3 TIMES THE DAMPING DEPTH IS IN THE REGION OF .04 PERCENT OF THE
C     SURFACE VARIATION OR ABOUT 1 DEGREE C.
      DFSVTY=SOLCNY/SPCHET
      DPOFST=3.*SQRT(2.*DFSVTY/1.99E-7)/2.54
   52 PENNUM=KSUM*FSTSUM*8.64E4
      PENDNM=L+C*PNDNMK
C     THERE ARE 929.03 CM.2SQ./FT.SQ.
      PEN=0.394*SQRT(PENNUM/PENDNM)
      WRITE(6,513)L,C,RSUM,KSUM,FSTSUM,PNDNMK
  513 FORMAT(5H L    ,F6.3,5H C    ,F6.3,8H RSUM    ,F12.3,8H KSUM    ,F9.5,
     110H FSTSUM    ,F6.1,10H PNDNMK    ,F8.3)
```

```
      IF(PEN.LE.DPTTSM)GO TO 60
      GO TO 501
   60 IF(PEN.LT.DPTH2)GO TO 605
      IF(J.EQ.1.AND.SNOW.EQ.0.0.AND.LITTER.EQ.0.0)GO TO 80
      DPDIF=PEN-DPTTSM
      GO TO 61
  605 DPTTSM=DPTH2+0.25
      DPDIF=-0.75
      GO TO 62
   61 DPTTSM=DPTTSM+DPDIF
   62 LNUM=LNUM+PCTMOS(J)*DPDIF*80.0
      CNUM=CNUM+SPCHET*DPDIF
      RSUM=RSUM+DPDIF/SOLCNY
      KSUM=DPTTSM/RSUM
      L=LNUM/DPTTSM
      C=CNUM/DPTTSM
      PENNUM=KSUM*FSTSUM*8.64E4
      PENDNM=L+C*PNDNMK
      PEN=0.394*SQRT(PENNUM/PENDNM)
      ITHW=ITHW-1
      IF(PEN.GT.MAXPEN)MAXPEN=PEN
      IF(PEN.GT.THWMAX)GO TO 65
      IF(PEN.LT.THWMAX.OR.PEN.LT.MAXPEN-2.0)GO TO 130
      GO TO 80
   65 IF(THWMAX.LE.0.0)GO TO 66
      THWMAX=0.0
      THW=0.0
      THSMMX=0.0
      PEN=MAXPEN
      FSTSUM=FTSMMX
      GO TO 80
   66 YMXPN=MAXPEN
      SOLCN1=PCTMOS(J)*CDYWTR+KWATSL*(1.-PROSTY(J))*CDYDS
      SOLCN2=(PROSTY(J)-PCTMOS(J))*CDYAIR*KWATAR
      S1FDNM=PCTMOS(J)+(1.-PROSTY(J))*KWATSL
      SLFDNM=S1FDNM+(PROSTY(J)-PCTMOS(J))*KWATAR
      SOLCNY=(SOLCN1+SOLCN2)/SLFDNM
      ACZONE=(DPOFST-MAXPEN)*2.54
      LDEEP=PCTMOS(NOLAY)*80.0
      DUMMY=(ANMFNC/ACZONE)*SOLCNY*8.64E4/LDEEP
      MAXPEN=MAXPEN-(DUMMY/2.54)
      IF(PEN.GT.MAXPEN)PEN=MAXPEN
      IF(MAXPEN.LT.YMXPN)FSTPRM=MAXPEN*MAXPEN*PENDNM/(KSUM*8.64E4)
      FSTPRM=FSTPRM*6.45
      IF(FSTPRM.LT.FSTSUM)FSTSUM=FSTPRM
      FSTPRM=FSTSUM
      GO TO 80
   70 FSTSUM=FSTSUM-DYMNTC-2.0
      MAXPEN=MAXPEN+ADDPTH
      IF(ADDPTH+(LITTER/4.0).GE.MAXPEN)MAXPEN=0.0
      ADDPTH=0.0
      SNOLOS=-0.001
      IF(FSTSUM.LT.0.0)GO TO 130
      IF(MAXPEN.LT.0.0)MAXPEN=0.0
      IF (MNTPCY.GT.0.AND.SNODPT.GT.4..AND.DYMNTC.GT.-5.0) GO TO 75
      CNUM=SNODPT*DENS
      PNDNMK=ANMFNC+FSTSUM/(2.0*DADFFZ)
      RSUM=SNODPT/((0.07+1.25*DENS)*1.0E-3)
      KSUM=SNODPT/RSUM
      PNNMSM=FSTSUM*8.64E4*KSUM
      PENDNM=0.2*PNDNMK
      PEN=0.394*SQRT(PNNMSM/PENDNM)
```

```
      IF(PEN.LT.SNODPT)GO TO 74
      DPTTSM=SNODPT
      IF (LITTER.GT.0.0) GO TO 125
      GO TO 49
   74 IF(MAXPEN.GT.SNODPT)GO TO 129
      GO TO 75
   75 IF(NOINFL.EQ.1)GO TO 81
      IF(PNCODE.EQ.1)GO TO 76
      PNCODE=2
   76 TPOFFT=8
      FREEZE=1
      PEN=0.0
      GO TO 115
   80 IF(NOINFL.EQ.1)GO TO 81
      WRITE(6,800)NFTNYD,SFCSTG,SFCPOT,NFLTRN
  800 FORMAT(10H NFTNYD    ,F8.2,10H SFCSTG    ,F8.2,10H SFCPOT    ,F8.2,
     110H NFLTRN    ,F8.2)
      IF(NFTNYD.GT.0.0.OR.SFCSTG.GT.SFCPOT)GO TO 95
   86 IF(MLTWTR.GT.0)GO TO 84
      IF (ORGCNT(1).GT.ORGMN1) GO TO 85
      IF(PEN/DAOFFZ.GT.1.0)GO TO 105
      IF(PEN/DAOFFZ.GT.0.5)GO TO 90
   84 IF(MXPNSL.LT.1.0.AND.TPOFFT.NE.3)GO TO 85
      TPOFFT=3
      NOINFL=1
      GO TO 110
   81 IF(YTPOFT.EQ.TPOFFT)TPOFFT=3
      GO TO 115
85    IF (ORGCNT(1).GT.ORGMN2) GO TO 105
   90 IF(MXPNSL.GT.3.5)GO TO 84
      TPOFFT=7
      GO TO 110
   95 IF(PEN.LT.SNODPT+LITTER)GO TO 110
      AMNC=(32.-AIRMIN)*0.555555555
      DEPPRC=(1-(AMNC/(2.0*(DYMNTC-AMNC))))*12.0*MNINFL
      IF(2.*DYMNTC-AMNC.LT.0.0)GO TO 96
      IF(NFTNYD.GT.DEPPRC)GO TO 84
      IF(NFLTRN.GT.DEPPRC)GO TO 84
   96 IF(AIRMIN.GT.27..AND.DYMNTC.LT.0.0)GO TO 84
      IF(MLTWTR.GT.0)GO TO 84
      IF(MXPNSL.GT.3.5)GO TO 84
      IF(ITHW.NE.1)GO TO 86
      TPOFFT=2
      GO TO 110
105   TPOFFT=1
      MXPNSL=MAXPEN-LITTER-SNODPT
      IF(PEN-LITTER-SNODPT.LT.MXPNSL-2.0)GO TO 130
      IF(MXPNSL.GT.3.5.AND.MNSTG.GT.0.8*POTSTG)GO TO 84
110   PNCODE=1
      IF(PEN.GT.MAXPEN)MAXPEN=PEN
      IF(MAXPEN.LE.0.0)GO TO 116
      FREEZE=1
  115 YTPOFT=TPOFFT
      MXPNSL=MAXPEN-SNODPT-LITTER
      IF(MXPNSL.LT.0.0)MXPNSL=0.0
      IF(THW.GT.MXPNSL)GO TO 160
      WRITE(6,101)MXPNSL,SNODPT,LITTER,THW
  101 FORMAT(20H SOIL PENETRATION    ,F6.2,13H SNOWDEPTH    ,F6.2,10H LITT
     1ER    ,F6.2,7H T-W    ,F6.2)
      RETURN
  116 PEN=0.0
      NOINFL=0
```

```
      IF(MXPNSL.GT.0.0)GO TO 117
      PNCODE=2
      GO TO 115
  117 TPOFFT=9

      GO TO 115
  120 MAXPEN=MAXPEN-LITLOS
      IF(MAXPEN.LT.0.0)MAXPEN=0.0
      LITLOS=0.0
      IF(-2.0.LT.DYMNTC)GO TO 130
      FSTSUM=FSTSUM-DYMNTC-2.0
      PNDNMK=ANMENC+FSTSUM/(2.0*DAOFFZ)
      DPTTSM=0.0
      CNUM=0.0
      RSUM=0.0
125   CNUM=CNUM+0.6*LITTER+LITTER*LITMOS
      MAXPEN=MAXPEN-LITLOS
      LITLOS=0.0
      DPTTSM=DPTTSM+LITTER
      LNUM=(PCTMOS(1)+LITMOS)*80.0*LITTER
C     LITMOS IS MOSTURE STORED IN THE LITTER LAYER ABOVE WHAT IS STORED
C     THE LAYER BELOW.
C     MLTTIM TESTS TO SEE IF WHEN TODAY IS ADAY OF THAWING, YESTERDAY
C     WAS ALSO ADAY OF THAWING.
C     MLTSM=SUM OF DAYS OF CONSECUTIVE MELT--IN THE MIDWEST FIVE DAYS OU
C     TO REDUCE EXTRA MOISTURE HOLDING CAPACITY OF UPPER LAYER.
      IF(SNODPT.GE.LITTER)GO TO 126
      LITK=0.6E-3/(1.+(SNODPT/LITTER))
C     THIS ACCOUNTS FOR REDUCTION IN EFFECTIVE CONDUCTIVITY WHEN GRASS,
C     LITTER AND SNOW COMBINE.
      GO TO 127
  126 LITK=0.6E-3/2.0
  127 RSUM=RSUM+LITTER/LITK
      KSUM=DPTTSM/RSUM
      L=LNUM/DPTTSM
      C=CNUM/DPTTSM
      PENNUM=KSUM*FSTSUM*8.64E4
      PENDNM=L+C*PNDNMK
      WRITE(6,513)L,C,RSUM,KSUM,FSTSUM,PNDNMK
      PEN=0.394*SQRT(PENNUM/PENDNM)
      IF (PEN.LT.DPTTSM) GO TO 129
      GO TO 50
  129 IF(MAXPEN.GT.SNODPT+LITTER)GO TO 66
      ITHW=ITHW-1
      NOINFL=0
      GO TO 75
 1290 MAXPEN=MAXPEN+ADDPTH
      ADDPTH=0.0
  130 IF(MAXPEN.LT.0.0.AND.PEN.LE.0.0)GO TO 160
      MAXPEN=MAXPEN-SNOLOS
      SNOLOS=0.0
      BTMSMP=PCTMOS(NOLAY)
      MLTTIM=DAOFYR-DAFYRY
      DPTTSM=0.0
      IF(MLTWTR.GT.0)GO TO 131
  132 MLTSM=MLTSM-1
      IF(MLTSM.LT.0.0)MLTSM=0.0
      DAFYRY=DAOFYR
      IF (DAOFTW.LT.1) THWSUM=0
      IF (THWSUM.LT.0.0) THWSUM=0.0
      DO 1320 J=1,NOLAY
```

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      IF(MXPNSL.LT.DPOFLY(J))GO TO 1330
 1320 CONTINUE
      PCTMST=PCTMOS(NOLAY)
      PRSTAV=PROSTY(NOLAY)
      GO TO 1325
 1330 WTMOS=0.0
      WTPROS=0.0
      DPTLFT=0.0
      J30=J+1
      DO 1334 K=J30,NOLAY
      WTMOS=WTMOS+PCTMOS(K)*DPOFLY(K)
      WTPROS=WTPROS+PROSTY(K)*DPOFLY(K)
 1334 DPTLFT=DPTLFT+DPOFLY(K)
      PCTMST=WTMOS/DPTLFT
      PRSTAV=WTPROS/DPTLFT
 1325 SOLCN1=PCTMST*CDYWTR+KWATSL*(1.-PRSTAV)*CDYDS
      SOLCN2=(PRSTAV-PCTMST)*CDYAIR*KWATAR
      S1FDNM=PCTMST+(1.-PRSTAV)*KWATSL
      SLFDNM=S1FDNM+(PRSTAV-PCTMST)*KWATAR
      SOLCNY=(SOLCN1+SOLCN2)/SLFDNM
      ACZONE=(DPOFST-MAXPEN)*2.54
      LDEEP=KIMSMP*80.0
      DUMMY=(ANMFNC/ACZONE)*SOLCNY*8.64E4/LDEEP
      MAXPEN=MAXPEN-(DUMMY/2.54)
      MXPNSL=(MAXPEN-SNODPT-LITTER)
      RSUM=(SNODPT/0.32E-3)+(LITTER/0.6E-3)+(MXPNSL/SOLCNY)
      KSUM=MAXPEN/RSUM
      MXPNSL=MAXPEN*2.54
      FSTSUM=MXPNSL*MXPNSL*PENDNM/(KSUM*8.64E4)
      MXPNSL=MAXPEN-SNODPT-LITTER
      IF(MAXPEN.LE.0.0)GO TO 160
      IF(PEN.GT.MAXPEN)PEN=MAXPEN
C     ********************PNCODE
C     1..........FROST DOES EXIST IN THE SOIL.
C     2..........SOIL PROTECTED FROM FROST BY SNOW.
C     3......... NO SOLUTION FOUND FOR FROST PENETRATION OR THAW EQUATION
C     DPOFST IS THE DEPTH TO CONSTANT TEMPERATURE.
      IF (SNOW.EQ.0.0) GO TO 135
      IF(DYMNTC-1.)136,134,134
  134 TPOFFT=4
      GO TO 115
  136 TPOFFT=9
      GO TO 110
  131 IF(MLTTIM-1)132,133,132
  133 MLTSM=MLTSM+1
      IF(MLTSM.GE.5)MLTWTR=0
      IF(MLTSM.GE.5)LITMOS=0.0
      GO TO 132
135   LNUM=0.0
      CNUM=0.0
      DAOFTW=DAOFTW+1
      KSUM=MAXPEN/RSUM
      MXPNSL=(MXPNSL+LITTER)*2.54
      IF(MXPNSL.LE.0.0)GO TO 1355
      IF(KSUM.LE.0.0)GO TO 1355
      FSTSUM=MXPNSL*MXPNSL*PENDNM/(KSUM*8.64E4)
      MXPNSL=MAXPEN-LITTER
      GO TO 1357
 1355 FSTSUM=0.0
 1357 FTSMMX=FSTSUM
      FSTSUM=0.0
      IF(DYMNTC.LE.0.0)GO TO 136
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      IF(THW.LE.0.0)THWSUM=0.0
      THWSUM=THWSUM+DYMNTC
      IF(MNTPCY.LT.0.0)GO TO 1358
      TWDNMK=0
      GO TO 1360
 1358 IF(DAOFFZ.NE.DAOFTW)GO TO 1359
      TWDNMK=0.0
      GO TO 1360
 1359 TWDNMK=(1.0*MNTPCY)/(2.0*(DAOFFZ-DAOFTW))
 1360 IF(LITTER.GT.0.0)GO TO 137
 1361 J=0
 1362 J=J+1
      JLAY=0
 1365 JLAY=JLAY+1
      IF(JLAY.LT.DPOFLY(J))GO TO 1367
      IF(J.LT.NOLAY)GO TO 1362
      PNCODE=9
      GO TO 115
 1367 SPCHET=(PCTMOS(J)*0.5+AVHTCY*(1.0-PCTMOS(J)))
      CDYSF1=(1-PROSTY(J))*CDYDS+PCTMOS(J)*CDYICE*KSOLIC
      CDYSF2=(PROSTY(J)-PCTMOS(J))*KSOLAR*CDYAIR
      CDFDNM=(PCTMOS(J)*KSOLIC+(PROSTY(J)-PCTMOS(J))*KSOLAR)
      CDYSF=(CDYSF1+CDYSF2)/(CDFDNM+(1-PROSTY(J)))
      DPTH2=DPTTSM
      DPTTSM=DPTTSM+1
      GO TO 138
  137 LNUM=LNUM+(PCTMOS(1)+LITMOS)*LITTER*80.0
      CNUM=CNUM+0.6*LITTER+LITMOS*LITTER
      DPTTSM=LITTER
      L=LNUM/DPTTSM
      C=CNUM/DPTTSM
      RSUM=RSUM+LITTER/0.6E-3
      THWNUM=(DPTTSM/RSUM)*THWSUM*8.64E4
      THWDNM=L+C*TWDNMK
      IF(THWDNM.LE.0.)THWDNM=0.001
      THW=0.394*SQRT(THWNUM/THWDNM)
      IF(THW.GT.LITTER)GO TO 1361
      GO TO 158
  138 LNUM=LNUM+PCTMOS(J)*80.0
      CNUM=CNUM+SPCHET
  139 L=LNUM/DPTTSM
      C=CNUM/DPTTSM
      IF (DAOFTW.EQ.1) GO TO 145
      IF (DAOFFZ.GE.DAOFTW) DAOFFZ=DAFZMX
  140 RSUM=RSUM+1./CDYSF
      KSUM=DPTTSM/RSUM
      THWNUM=KSUM*THWSUM*8.64E4
      IF(THWNUM.LE.0.0)GO TO 158
      THWDNM=L+C*TWDNMK
      THW=0.394*SQRT(THWNUM/THWDNM)
      IF (THW.LT.DPTTSM) GO TO 155
      IF(THW.GT.DPTTSM.AND.J.LT.NOLAY)GO TO 1365
      GO TO 150
  145 DAFZMX=DAOFFZ
      DAOFFZ=0
      GO TO 140
  150 PNCODE=20
      GO TO 115
  155 IF(THW.LT.DPTH2)GO TO 157
      GO TO 156
  157 DPTTSM=DPTH2+.25
      DPDIF=-.75
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      GO TO 1561
  156 DPDIF=THW-DPTTSM
 1561 DPTTSM=DPTTSM+DPDIF
      LNUM=LNUM+PCTMOS(J)*DPDIF*80.0
      CNUM=CNUM+SPCHET*DPDIF
      RSUM=RSUM+DPDIF/CDYSF
      KSUM=DPTTSM/RSUM
      L=LNUM/DPTTSM
      C=CNUM/DPTTSM
      THWNUM=KSUM*THWSUM*8.64E4
      THWDNM=L+C*TWDNMK
      WRITE(6,1580)
 1580 FORMAT(72H   LNUM     CNUM    DPTTSM    RSUM     KSUM    THWSUM   THWDNM=
     1 THWNUM   TWDNMK  )
      WRITE(6,1582)LNUM,CNUM,DPTTSM,RSUM,KSUM,THWSUM,THWDNM,THWNUM,
     1TWDNMK
 1582 FORMAT(3F8.4,F8.1,F8.6,F8.1,F8.2,2F8.4)
      THW=0.394*SQRT(THWNUM/THWDNM)
  158 ITHW=2
      IF (PEN.GT.MAXPEN) MAXPEN=PEN
      IF(THW.GT.THWMAX)THWMAX=THW
      IF(THW.GT.THWMAX)THSMMX=THWSUM
      IF (THW.GE.MAXPEN) GO TO 160
      TPOFFT=5
      IF(THW.LT.0.0)GO TO 165
      IF(AIRMIN.GT.32.0)GO TO 110
      AMNC=(AIRMIN-32.0)*0.5555
      DEPPRC=(1.+(AMNC/(2.0*(DYMNTC-AMNC))))*12.0*MNINFL
      IF(RUNOFF.GT.DEPPRC)GO TO 84
      IF((THW*(PROSTY(1)/100.0)-SFCSTG).LT.DEPPRC)GO TO 110
      MLTWTR=1
      GO TO 84
160   DAOFFZ=0
      DAFZMX=0
      TPOFFT=6
      MLTWTR=0
      MAXPEN=0.0
      LITMOS=0.0
      THW=0.0
      FREEZE=0
      NOINFL=0
      GO TO 115
  165 TPOFFT=9
      GO TO 110
      END
      SUBROUTINE EVPADJ(EPOT,SURFAC,MNSTG,SFCSTG,CRTPNT,SFCWLT,SFCPOT,
     1SFCCRT,POTSTG,PCP)
      REAL MNSTG
      DIF=EPOT-PCP
      IF(DIF)16,16,6
    6 EPOT=DIF
      DIF=0.0
C     EVAPORATION OF RAIN OF THE DAY OF RAIN CAN GO AT THE MAX RATE
C     REGARDLESS OF HOW DRY THE SOIL IS.
      IF(SURFAC.GT.0)GO TO 5
      IF(MNSTG.GT.CRTPNT)GO TO 10
      EPOT=EPOT*(MNSTG/POTSTG)
      GO TO 15
   10 DIF=MNSTG-CRTPNT
      IF(DIF.GT.EPOT)GO TO 15
      DDIF=EPOT-DIF
      EPOT=DIF+DDIF*(CRTPNT/POTSTG)
```

```
      GO TO 15
    5 IF(SFCSTG.GT.SFCCRT)GO TO 20
      EPOT=EPOT*(SFCSTG/SFCPOT)
      GO TO 15
   20 DIF=SFCSTG-SFCCRT
      IF(DIF.GT.EPOT)GO TO 15
      DDIF=EPOT-DIF
      EPOT=DIF+DDIF*(SFCCRT/SFCPOT)
   15 EPOT=EPOT+PCP
   16 RETURN
      END
      SUBROUTINE TRNSLS(LYSMO,LWRLOS,SFCLOS,N,NDLAY,LYRSMP,
     1POTSTG,SFCSTG,MNSTG,SFCPOT)
      DIMENSION PRTNFC(8)
      REAL LYSMO(N),LWRLOS,LYRSMP(N),MNSTG
      COMRAT=0.0
      NA=NDLAY
      DO 5 J=1,NA
      RATI=LYSMO(J)/LYRSMP(J)
      IF(RATI.LT.0.0)GO TO 3
      PRTNFC(J)=2.0*(N-J+1.)/((N+3.0)*N)
      IF(J.EQ.1)PRTNFC(J)=PRTNFC(J)*2.0
      GO TO 4
    3 PRTNFC(J)=0.0
    4 PRTNFC(J)=PRTNFC(J)*RATI
    5 COMRAT=COMRAT+PRTNFC(J)
      IF(COMRAT.LE.0.0)GO TO 7
      LOSI=SFCLOS/COMRAT
      DO 6 J=1,NA
    6 LYSMO(J)=LYSMO(J)-PRTNFC(J)*LOSI
      GO TO 8
    7 SFCLOS=0.0
    8 NB=NDLAY+1
      COMRAT=0.0
      DO 12 J=NB,N
      RATI=LYSMO(J)/LYRSMP(J)
      IF(RATI.LE.0.0)GO TO 9
      PRTNFC(J)=2.0*(N-J+1.)/((N+3.0)*N)
      GO TO 10
    9 PRTNFC(J)=0.0
   10 PRTNFC(J)=PRTNFC(J)*RATI
   12 COMRAT=COMRAT+PRTNFC(J)
      IF(COMRAT.LE.0.0)GO TO 14
      LOSI=LWRLOS/COMRAT
      MNSTG=0.0
      DO 13 J=NB,N
   13 LYSMO(J)=LYSMO(J)-PRTNFC(J)*LOSI
      GO TO 15
   14 LWRLOS=0.0
   15 DO 16 J=1,N
   16 MNSTG=MNSTG+LYSMO(J)
      RETURN
      END
      SUBROUTINE ESTVAP(PLNCO1,PLNCO2,DAY,R,DYMNTC,I,A)
      RATIO=PLNCO1/PLNCO2
      CORFAC=(DAY/12.)*(1/R)**2*RATIO*.54
      EVAPU=.2099/*((10*DYMNTC)/I)**A
      EPOT=EVAPU*CORFAC
      RETURN
      END
      SUBROUTINE LAYER(N,WP,FC,DPT,LYRSMP,FLDCPY,WLTPNT,TOTSTG,DCEN,
     1DTOT,NDLAY,ESFCLY,CRTPCT,CRTPNT,SFCWLT,DPTAV,SFCPOT,SFCCRT,
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     2POTSTG.PROSTY.FRACJ.ROOTBK)
      DIMENSION WP(N).FC(N).DPT(N).DCEN(N).CRTPCT(N).FRACJ(N)
      DIMENSION PROSTY(N)
      REAL LYRSMP(N)
      FLDCPY=0.0
      CRTPNT=0.0
      SFCPOT=0.0
      DTOT=0.0
      TOTSTG=0.0
      POTSTG=0.0
      WLTPNT=0.0
      DO 15 J=1.N
C     CONVERT POROSITY TO DECIMAL FRACTION
      PROSTY(J)=PROSTY(J)/100.0
      LYRSMP(J)=FC(J)-WP(J)
      CRTPNT=CRTPNT+CRTPCT(J)*DPT(J)/100.-WP(J)
      FLDCPY=FLDCPY+FC(J)
      WLTPNT=WLTPNT+WP(J)
      DTOT=DTOT+DPT(J)
      DCEN(J)=DTOT-(DPT(J)/2.0)
      POTSTG=POTSTG+LYRSMP(J)
      TOTSTG=TOTSTG+DPT(J)*PROSTY(J)
      IF(J.GT.NDLAY)GO TO 15
      ESFCLY=DTOT+DPT(J)
      SFCWLT=WLTPNT
      SFCCRT=CRTPNT
      SFCPOT=POTSTG
   15 CONTINUE
      DPTAV=DTOT/N
      IDTOT=DTOT
C     CONFRC IS A CONSTANT IN THE PARTITTIONING PROCESS.
      CONFRC=DTOT-ROOTBK
      JRT=DTOT-ROOTBK+1
      B=JRT-2
      DO 14 J=JRT.IDTOT
   14 CONFRC=CONFRC+J-B
      CONFRC=CONFRC+(ROOTBK+1.)
      JRT=ROOTBK
      L=0
      M=0
      DO 16 J=1.N
      L=M+1
      M=L+DPT(J)-1
      FRCSUM=0.0
      IF(M.GE.JRT)GO TO 25
      DO 17 K=L.M
      AK=K
      ANL=ROOTBK+2.0-AK
      FRCSUM=FRCSUM+ANL/CONFRC
      IF(K.EQ.1)FRCSUM=FRCSUM*2.0
   17 CONTINUE
      GO TO 16
   25 IF(L.GT.JRT)GO TO 28
      DO 26 K=L.JRT
      AK=K
      ANL=ROOTBK+2.0-AK
   26 FRCSUM=FRCSUM+ANL/CONFRC
      B=M-JRT
      FRCSUM=FRCSUM+B/CONFRC
      GO TO 16
   28 FRCSUM=(M-L+1)/CONFRC
   16 FRACJ(J)=FRCSUM
```

```
      RETURN
      END
      SUBROUTINE SMP(N.MNSTG.LYRSMP.LYSMO.EVAP.PCP.SFCSTG.DA.POTSTG.
     1AOFSMC.DCEN.DTOT.DAMXT.FIRST.SURFAC.NDLAY.DPTAV.DPT.SFCPOT.MLTWTR.
     2SNOMLT.RUNOFF.FRACJ)
      DIMENSION DCEN(N).DPT(N).FRACJ(N).PRTNFC(8)
      INTEGER DA.DAMXT.FIRST.SURFAC.TINDX.TINDX2
      REAL LYRSMP(N).LYSMO(N).MNSTG.LOSI.LADIF
      RUNOFF=0.0
      PCPLOS=0.0
      IF(FIRST.EQ.0)TINDX=0
      IF(FIRST.EQ.0)TINDX2=0
      NDLAY1=N
      IF(SURFAC.EQ.1)GO TO 8
      N=NDLAY-1
      GO TO 9
    8 N=NDLAY
    9 IF(DAMXT.LT.32.0)GO TO 4
      IF(TINDX.GT.0)GO TO 6
   25 IF(MLTWTR.GT.0)GO TO 18
      IF(MLTWTR.EQ.0.AND.TINDX2.GT.0)GO TO 19
    5 PCP1=PCP
      IF(DAMXT.LT.32..AND.SNOMLT.LT.0.001)GO TO 95
      PCP=PCP+SNOMLT
      IF(DAMXT.LT.32.)PCP=SNOMLT
      SNOMLT=0.0
      EVAP1=EVAP
      IF(PCP.LT.EVAP)GO TO 20
      IF(PCP.EQ.0.0)GO TO 90
      IF(PCP.EQ.EVAP)GO TO 90
      IF(LYSMO(1).EQ.LYRSMP(1).AND.PCP1.GT..1)GO TO 1
      PCP=PCP-EVAP
      MNSTG=MNSTG+PCP
      GO TO 10
    1 IF(LYSMO(2).EQ.LYRSMP(2))GO TO 2
      PCP=PCP*.6
      PCPLOS=PCP1+SNOMLT-PCP
      GO TO 7
    2 PCP=PCP*.4
      PCPLOS=PCP1+SNOMLT-PCP
    7 IF(MNSTG+PCPLOS.GT.POTSTG)GO TO 15
      WRITE(6.60)PCPLOS
   60 FORMAT(68H THE UPPER LAYER(S) IS(ARE) AT FIELD CAPACITY AND IT IS
     1LIKELY THAT .F6.3.21H INCHES HAVE RUN OFF.)
      GO TO 10
    4 IF(TINDX.GT.0)GO TO 25
      DLTFC=.12*LYRSMP(1)
      TINDX=1
      LYRSMP(1)=LYRSMP(1)+DLTFC
      POTSTG=POTSTG+DLTFC
      SFCPOT=SFCPOT+DLTFC
      GO TO 25
   18 IF(TINDX2.GT.0)GO TO 5
      TINDX2=1
      POTSTG=POTSTG-LYRSMP(1)
      SFCPOT=SFCPOT-LYRSMP(1)
      LYRSMP(1)=LYRSMP(1)*3.0
      POTSTG=POTSTG+LYRSMP(1)
      SFCPOT=SFCPOT+LYRSMP(1)
      GO TO 5
   19 POTSTG=POTSTG-LYRSMP(1)
      SFCPOT=SFCPOT-LYRSMP(1)
```

```
      LYRSMP(1)=LYRSMP(1)/3.0
      POTSTG=POTSTG+LYRSMP(1)
      SFCPOT=SFCPOT+LYRSMP(1)
      TINDX2=0
      GO TO 5
    6 LYRSMP(1)=LYRSMP(1)-DLTFC
      POTSTG=POTSTG-DLTFC
      TINDX=0
      GO TO 5
   10 IF(MNSTG.GE.POTSTG)GO TO 15
      JLOOP=0
      ADEDH2=PCP
      NLAY=1
   21 IF(LYSMO(NLAY).GE.LYRSMP(NLAY))GO TO 14
      LADIF=LYRSMP(NLAY)-LYSMO(NLAY)
      IF(LADIF.LT.ADEDH2) GO TO 13
      LYSMO(NLAY)=LYSMO(NLAY)+ADEDH2
   16 IF(NLAY.GE.N)GO TO 90
      JLOOP=JLOOP+1
      IF(JLOOP.GT.10)STOP
      RATIO1=LYSMO(NLAY)/LYRSMP(NLAY)
      RATIO2=LYSMO(NLAY+1)/LYRSMP(NLAY+1)
      RATDIF=RATIO1-RATIO2
      IF (RATDIF.GT.0.2) GO TO 12
      IF(NLAY.GT.N-2)GO TO 90
      RATIO3=LYSMO(NLAY+2)/LYRSMP(NLAY+2)
      RATDIF=RATIO2-RATIO3
      IF(RATDIF.GT.0.2)GO TO 51
      GO TO 90
   51 TRNSLY=RATDIF/2.0
      LYSMO(NLAY+1)=LYSMO(NLAY+1)-TRNSLY
      LYSMO(NLAY+2)=LYSMO(NLAY+2)+TRNSLY
      GO TO 16
   12 IF(NLAY.EQ.1.AND.JLOOP.EQ.1)RATDIF=RATDIF*ADEDH2
      TRNSLY=RATDIF/2.0
      LYSMO(NLAY)=LYSMO(NLAY)-TRNSLY
      LYSMO(NLAY+1)=LYSMO(NLAY+1)+TRNSLY
      GO TO 16
   13 ADEDH2=ADEDH2-LADIF
      LYSMO(NLAY)=LYSMO(NLAY)+LADIF
      LADIF=LYRSMP(NLAY+1)-LYSMO(NLAY+1)
      IF (N-NLAY.LT.2) GO TO 80
      IF(LADIF.LT.ADEDH2) GO TO 23
      IF (N-NLAY.LT.3) GO TO 80
      LYSMO(NLAY+1)=LYSMO(NLAY+1)+ADEDH2
   26 RATIO2=LYSMO(NLAY+1)/LYRSMP(NLAY+1)
      JLOOP=JLOOP+1
      IF(JLOOP.GT.10)STOP
      RATIO3=LYSMO(NLAY+2)/LYRSMP(NLAY+2)
      RATDIF=RATIO2-RATIO3
      IF (RATDIF.GT.0.3) GO TO 22
      IF(N-NLAY.GE.3)GO TO 43
      GO TO 90
   43 NLAY=NLAY+1
      GO TO 26
   22 TRNSLY=RATDIF/2.0
      LYSMO(NLAY+1)=LYSMO(NLAY+1)-TRNSLY
      LYSMO(NLAY+2)=LYSMO(NLAY+2)+TRNSLY
      GO TO 26
   80 LYSMO(NLAY+1)=LYSMO(NLAY+1)+ADEDH2
      GO TO 90
   23 ADEDH2=ADEDH2-LADIF
```

```
+1)=LYSMO(NLAY+1)+LADIF
+2

STG-POTSTG+PCPLOS
STG
CPOT
.N
LYRSMP(J)
.LE.0.0)GO TO 90
+1) DA,RUNOFF
+ ON DAY ,I3,25H OF THIS MONTH THERE WAS ,F7.2,24H FOR RUN
CHARGE.)
DEDH2+LYSMO(NLAY)-LYRSMP(NLAY)
AY)=LYRSMP(NLAY)
Y+1
EQ.N)GO TO 89

=LYSMO(N)+ADEDH2

VAP-PCP
.LE.0.0)GO TO 90
.1)GO TO 34
.0
=1,N
SMO(J)/LYRSMP(J)
.LT.0.0)GO TO 37
J)=FRACJ(J)*RATI
5
(J)=0.0
=COMRAT+PRTNFC(J)
TRLOS/COMRAT
J=1,N
J)=LYSMO(J)-PRTNFC(J)*LOSI
40
1)=LYSMO(1)-WTRLOS
MO(1).LT.0.0)LYSMO(1)=0.0
P1
Y1
VAP1
=0.0
J=1,N
=ADFSMC+LYSMO(J)
GT.NDLAY)GO TO 42
=ADFSMC
DIF
=ADFSMC
N
```

GLOSSARY OF TERMS IN MAIN PROGRAM

A	The uncorrected estimated evaporation computed based on the Thornthwaite equation. It is only used when the Solar Radiation is missing.
ABDOSN	The value for the albedo when show is on the ground.
ACTLAY	The thickness in inches of the active layer of water replenishment or loss, generally the layer with sufficient roots for transpiration to be significant.
ADDPTH	Increase in snow depth in inches resulting from snow fall.
ADPT	The average dew point temperature (°F).
AIRMAX	The maximum air temperature for the day (°F).
AIRMIN	The minimum air temperature for the period(°F).
ALBDO	The albedo.
ALBDOS	The albedo representing summer conditions.
ALBDOW	The albedo representing winter conditions without snow.
ANGRAD	A function statement to convert angles in degrees to angles in radians.
ANMENC	The annual mean air temperature in degrees Celcius.
AOESMC	Actual (measured) or estimated soil moisture condition - the value in inches for the total profile.
APRTFC	A function to estimate the influence of humidity when the mean dewpoint is recorded near or above the mean air temperature.
ARMNYD	The minimum air temperature yesterday(°F).
ARMXYD	The maximum air temperature yesterday(°F).
ASPECT	The azimuthal direction of the representative slope for the basin, in degrees.
ASPRAD	The azimuthal direction of the representative slope in radians.
AVHTCY	The average heat capacity of the soil.
AVSCDY	The average thermal conductivity of an air dried soil determined by a measurement. If not available the program uses an average conductivity for dry soil.
CC	Cover code. Ground bare Harvested crop with stubble 1 Hardwoods without leaves 2 Hardwoods with leaves 3 Pasture 4 Open crops such as corn 5 Closed crops such as wheat 6 Conifers 7

CCODE	Word descriptions of cover code used for headings.
CCODE(J)	Members of code array, i.e., bare ground, etc.
CORLAT	Corrected latitude of level surface corresponding to natural slope of real surface.
COSZEQ	The angle of incidence of a direct ray from the sun when the sun is crossing the equator at solar noon.
CRTPCT	A threshold value for the available water below which the rate of evaporation is significantly reduced below the maximum possible rate, and is proportional to the amount of water remaining above the permanent wilting point. Above this threshold value the soil is assumed to lose water by evapotranspiration at the maximum rate. This value is given in percent soil moisture by volume for each layer.
CRTPNT	The threshold value of the available water left in the total profile given in inches above which actual evapotranspiration is the same as that from a lake and below which is reduced proportionately to the amount of available water remaining in the profile.
DA	The day of the month.
DAOFFZ	The length of the current freezing period in days.
DAOFTW	Number of consecutive days of thawing.
DAOFYR	The number of days so far this year.
DASINC	The number of days since pasture or meadow has been harvested as hay.
DAY	The length of a day for a given surface slope and time of year.
DCEN	The depth of the center of each individual soil layer.
DENS	Snow density.
DENSIT	Average soil density in grams/cm^3.
DFSTAV	Average defusivity of the soil. Used in computing the depth to generally stable temperatures.
DLTAH(I)	Layer thickness in inches, an array.
DLTAVP	Difference between the maximum and minimum seasonal vapor pressures (in. of Hg).
DP	The dew point that is read in (°F).
DPOFST	Depth of stable temperatures. The point in the ground at which daily and seasonal temperatures cease to cause measurable change. This value is estimated in inches.

DPTAV	The average of the layer thicknesses in inches.
DPVPRS	The saturation vapor pressure in inches of Hg. dewpoint vapor pressure.
DTOT	The total depth of the active layer in inches.
DYMNTC	The daily mean temperature in °F.
DYMNTF	The daily mean temperature in °C.
EPOT	The potential evaporation. The evaporation computed based on meteorological conditions. The evaporation from a shallow lake.
ESFCLY	The estimated thickness in inches of the soil surface layer that is subject to direct evaporation.
EVAP	Measured pan evaporation in inches.
EVPDIF	The difference between measured pan evaporation and computed lake evaporation in inches.
EVPPOT	A remembering label for EPOT that the initial value is retained while EPOT is allowed to be adjusted.
FALDAY	An index. The number of days since vegetation started into the change to dormancy.
FALFRC	The fraction of the period which has passed during the fall transition from growing to dormant conditions.
FALL	A day of the year which has been observed or estimated on which the beginning of the change to dormancy begins.
FALPRD	The length of the period during which transpiration goes from nearly maximum to the winter level.
FC(I)	The field capacity for each soil layer in inches of water.
FIRST	An index which is zero for the first day and 1 thereafter.
FLDCPY	The field capacity of the total profile in inches of water.
FRACJ(I)	The fraction of the evapotranspiration loss that would come from a given layer.
FREEZE	An index which indicates that some frost has occurred.
FREZTP	Mean of two daily mean temperatures to allow for heat from a previous warm day to be lossed before frost is allowed to form.
FROST	An index which indicates that the basin is now in the dormant season and albedos, and soil moisture lost will be computed accordingly.
FRZFST	An index which when the mean daily dew point is high allows precipitation to be recorded before the effect of freezing is considered; but when the dew point is low, freezing occurs before precipitation is considered. The thickness of the snow blanket affects the depth of frost.

FSTSUM	Frost index.
GRASMN	If grass is a significant cover crop on the watershed and the soil moisture becomes depleted to a certain threshold value, the grass will die or go dormant. That value is GRASMN.
HC	An index which notes a crop harvest, a change in season, or any other unusual occurrence which will affect transpiration.
I	Constant used in the Thornthwaite evaporation estimation equation.
IPRCC	A memory variable used so that when a type of ground cover is returning to its summer condition from a state of dormancy, it may go from a single dormant class to the one from which it came. Example, Dormant to meadow, or dormant to cultivated crop.
IPRINT	An index which prints out the daily numerical values before the frost message.
IRTN	An index used to facilitate printing a message that transpiration is ceasing and that evaporation will occur only from the surface soil layers.
ITHW	An index which shows when set equal to 1 that thawing occurred on the previous day and that if freezing occurs today that stalactite frost is likely.
ITRNST	An index used to indicate that the basin is in transition from summer to winter conditions or visa versa.
J	Subscript.
J1	A subscript used to limit soil moisture values of layers to five without limiting the model to five layers.
J2	Subscript used in writing permeability condition.
J3	Similar to J2
K	Subscript
LABEL	An array of words giving the four cardinal directions for indicating the aspect direction of the watershed.
LAT	The latitude which is read in to the nearest thousandth of a degree.
LATRAD	The latitude changed to radians.
LGTMMN	The long term monthly average temperatures in °F.
LITDEX	An index which changes value of litter from initial value in the fall when the grass for instance is standing at near full natural height to a minimum value after it has been matted down by snow, rainfall, or drying.

LITFRZ	Effective thickness of vegetation and litter after the soil has frozen to a depth of 1 inch.
LITLOS	The apparent decrease in the depth of the litter in inches after grasses and low vegetation are crushed or matted down.
LITLOW	The effective value of the depth of vegetation in inches after it has been matted down.
LITMAX	The effective value of the depth of vegetation and organic material in inches in the fall before it is matted by the elements.
LITMOS	The moisture in % moisture by volume trapped in the litter above a frozen soil.
LITSNO	The effective value of the vegetative litter layer during and after being compressed by at least 2 inches of snow.
LITTER	The depth of inches of litter on the watershed.
LWRLOS	The fractional loss from the lower layers of the profile that lose water almost exclusively by transpiration. This is used during the period of transition from summer to winter and winter to summer.
LYRSMP(I)	The amount of soil moisture between field capacity and the wilting point for each layer in inches of water.
LYSMO(I)	The initial value of the soil moisture in each layer. After the model has begun this becomes the current value of soil moisture in each layer. The value is given in inches of water.
MAXPEN	Maximum frost penetration in inches.
MLTWTR	An index. It is 0 when normal conditions exist but when thawing begins but does not penetrate the frozen ground it equals 1. When it equals 1 the field capacity of the upper layer is allowed to increase such that the available water allowed before runoff occurs is tripled.
MNASD	The mean value of a term related to optical path length. The attenuation of the atmosphere is calculated for the atmosphere and water vapor. This term is mainly used only when solar radiation is not available and the Thornthwaite equation is being adjusted for seasonal changes.
MNINFL	The infiltration rate in inches per hour that a watershed might have when it is near field capacity and under cold conditions. The minimum infiltration capacity of a porous frozen soil.
MNSCAT	A variable to compute the amount of radiation that is scattered as opposed to that that comes directly. It is used in making adjustments when solar radiation is not directly measured.

MNSTG	The amount of available water in inches in the profile at any given time.
MNTPCY	The mean air temperature in °C, yesterday.
MNVPRS	The climatological long term mean atmospheric vapor pressure. Used with the seasonal variation in vapor pressure to correct solar radiation when it is not measured. It is entered in name of Hg.
MO	The month.
MXPNSL	The maximum frost penetration in inches into the soil alone. It does not include penetration in snow and litter.
NDLAY	The number of the lowest layer for which direct evaporation has a significant effect.
NFLTRN	Excess water available for infiltration or runoff. In some subroutines the same variable is called RUNOFF. It is estimated in inches.
NFTNYD	Excess water from precipitation or snowmelt available to runoff or water table recharge from the previous day. The value is in inches.
NOINFL	An index to indicate that the ground is impervious.
NOLAY	The number of layers.
ORGCNT(2)	The organic content of a soil layer.
ORGMNI	The lower threshold for organic content of a layer below which when freezing occurs unless the soil is extremely dry, concrete frost most likely is the type that will occur.
ORGMN2	The upper threshold for organic content of a layer above which should freezing occur it will most likely be of the granular form. Between these two thresholds it may be difficult to predict the type of frost that will form.
OSNDPT	The previously estimated snow depth.
PCP	Precipitation entered in inches.
PCPW	Liquid precipitation to indicate when rain and show might be expected that precipitation is liquid.
PCPWEQ	Estimated precipitable water based on climatological records. Used in correcting solar radiation when not measured.
PEN	The depth of frost penetration in inches.
PENDNM	The denominator of the penetration equation.
PNCODE	The frost penetration code.

0....No frost in ground or litter.
1....There is frost in the soil or litter.
2....The soil is protected from frost by snow or litter.
3....or greater... There is no solution found for the freeze thaw equation or conditions do not make a clear cut estimation possible.

PI	The circular and sperical geometric constant, 3.1416....
PLNCO1	A plain coefficient. A coefficient used in correcting sloping surfaces to a flat horizontal plane and the seasonal movement of the sun.
PLNCO2	A plain coefficient to indicate solar radiation on a plain surface on the day of an equinox.
POTSTG	Potential storage in a soil profile between field capacity and the wilting points. It is the sum of the variables LYRSMP.
PROSTY(I)	The porosity of each layer.
QUAD	The circle of directions is broken into four quadrants: the northern one is from 45° West of North to 45° East of North and so forth around the circle. QUAD is the subscript to tabulate these quadrants.
R	The ratio of the seasonal distance from the sun to the Earth varies on a seasonal basis, to the mean distance.
ROOTBK	This is the depth in inches at which the transpiration process appears to begin to extract moisture from a depth increment in a relatively constant fashion rather than as a function of depth. It is postulated as a breakpoint in the root distribution as a function of depth.
RRF	The relative roughness factor relating the net gain or loss of snow on a watershed from or to its surroundings.
RSUM	The thermal resistance of the soil.
RUNOFF	The amount of moisture that is lost to infiltration of runoff after the soil profile is brought to field capacity. It is estimated in inches.
SACOR	The slope-aspect correction to latitude to give an equivalent horizontal surface subject to the same extra-terrestial solar radiation as the real sloping surface.
SARG	The phase adjustment used in describing the variation in the solar delination.
SCOND(I)	The words porous or impervious that describe the permiability of a frozen soil.
SFCCRT	The threshold value of soil moisture separating the maximum evapotranspiration rate from a reduced rate proportional to the remaining soil moisture left in the layer that is subject to direct evaporation. It is estimated in inches.

SFCLOS	The fraction of moisture loss from the surface layer subject to direct evaporation during the period of transition from winter to summer conditions and summer to winter conditions. It is estimated in inches.
SFCPOT	The maximum available water that can be stored in the surface layer that is subject to direct evaporation. Estimated in inches.
SFCSTG	The current soil moisture above the wilting point that is stored in the furface layer subject to direct evaporation. It is estimated in inches.
SFCWLT	The wilting point for the surface layer that is subject to direct evaporation. It is entered in inches.
SIGMA	A constant used in fitting the seasonal water vapor curve in the upper midwest.
SIGSQR	A term used for fitting the change in season precipitable water.
SLOPE	The representative slope of the watershed in degrees.
SLPRAD	The angle of slope converted to radians.
SNODPT	The estimated snow depth in inches based on an assumed density and the precipitation occurring below 32°C which has not been calculated to have melted.
SNOFAL	The observed increase in the snow depth in inches. Generally a result of precipitation.
SNOLOS	A decrease in snow depth in inches.
SNOMLT	Snowmelt water in inches.
SNOW	The amount of water equivalent in inches falling as precipitation when the mean air temperature was below 1°C.
SOILTP	The name of the soil type.
SPRDAY	An indexing register to keep track of the number of days since transition to summer transpiring conditions began.
SPRFRC	Spring fraction - the ratio of the number of days passed in the spring season to the number estimated for complete transition from a dormant to a fully active condition.
SPRING	The day of the year when the transition starts from dormancy to full transpiring operation.
SPRPRD	The estimated number of days for the transition from winter dormancy to summer transpiration.
SR	Solar radiation in langleys corrected for slope and aspect and later for albedo.

SRAD — Solar radiation raw data.

SRCOR — Is a solar radiation correction determined in DALNCR to adjust raw solar radiation data.

SUMGRS — The latest day of growing season that drouth dormant plants would be expected to be revived by rain and transpire again.

SURFAC — An index. It is zero if the vegetation is such that transpiration is taking water from deeper layers in the soil profile. It is 1 when either the cover or the season is such that water loss is by evaporation only and indicates that only the surface layers are active in moisture exchange processes.

THW — The depth of thaw in inches.

THWMAX — Maximum depth of thaw when thaw does not completely penetrate the frost layer.

THWSUM — Accumulated thaw index - parallel to frost index.

TMIXFC — Temperature indexing function. Since weather stations may tend to be affected by airports or urban areas where they are located, this function is to relate the station temperatures to those on the watershed.

TOTSTG — The number of inches of water that could be stored if the soil structure remained the same and all vertices were filled with water.

TPOFFT — A frost code. See the freeze-thaw routine for more detail.

UNCDAY — A day length that is not affected by the slope of the watershed.

WILIPT — (uncorrected day length)

WLTPNT — The permanent wilting point--the amount of water in inches at 15 bars of tension.

WIND — The measured wind movement in miles per day generally at 18" above the surface.

WP(I) — The inches of water retained in each layer at the permanent wilting point (15 bars tension).

WSNO — The watershed designation, generally a number.

YDP — Yesterday's mean dew point.

YR — The year.

GLOSSARY of Terms in SNOWR not found in MAIN

RAD1	Melt due to longwave radiation during warm part of the day.
RAD2	Melt due to longwave radiation during the cold part of the day.
ADPTC	Dew point temperature in degrees °C.
VAPRS	Vapor pressure of water in inches of Hg.
YSNOW	The amount of snow estimated present yesterday in inches of water equivalent.
CDHFDT	Estimated mean dewpoint during the cold half of the day.
CDHFTM	Mean air temperature during the cold half of the day.
CNDCMT	Condensation melt during the cold half of the day.
CNDMLT	Condensation melt
CNDWMT	Condensation melt during the warm half of the day.
DASNMT	Daily snow melt due to seasonal degree day factor.
DMINTC	Minimum air temperature in °C.
DPVPRS	Dewpoint vapor pressure.
DVMXTC	Daily maximum temperature °C.
EVSNMT	Snow melt due to the seasonal degree day factor occurring during the cold half of the day.
RADMLT	Snowmelt due to longwave radiation.
RANMLT	Snowmelt due to rain.
RANNIT	Snow melt due to rain during the cold half of the day.
SNODNS	Snow density
WMHFDT	Main dew point temperature during the warm half of the day.
WMHFTM	Mean air temperature during the warm half of the day.
YSNDPT	The estimate snow depth yesterday

Terms in DALNCR not found in the MAIN Program

ANG	The angle made by a line from the center of the earth to the rising or setting sun and a line from the center of the earth to the sun at solar noon.
DEC	The solar declination
ADEC	The declination in radians
COSZ	The average of the cosine of the angle the sun makes the horizon
RARG	A phase shift or correction to the calendar for the cyclic variation in the distance between the earth and the sun.
HAFDA	The time in hours between noon and sunrise or sunset.
SCTRD	A correction to the extraterrestrial radiation to consider scattering.
SRADW	Measured solar radiation
ABSRBD	A correction to the extraterrestrial radiation to consider absorption
PCPWTR	Estimate of precipitable water
UNCANG	Uncorrected value of ANG
UNCDAY	Uncorrected length of day.

Terms found in PRECPR not found in MAIN

DENS	Snow density
OSNDNS	Snow density prior to new estimate
YSNDPT	Snowdepth yesterday

Terms found in HAYCUT not found in MAIN

DIF	Difference in the water to be lost and that available for loss
HAYCTF	The equation for the return to normal evaporation after a harvest
LWRFRC	Percent of available water unavailable without transpiration
LWRWTR	Percent of water above critical cutoff point that is found in lower zone not subject to direct evaporation
SFCFRC	Ratio of SFCSTG/SFCPOT - that moisture currently stored to that at field capacity
SFCWTR	Amount of soil moisture in zone subject to direct evaporation,greater than critical moisture.

Terms found in EVAPR not found in MAIN

TA	Corrected or adjusted air temperature used in evaporation equation
TSAT	Adjusted dewpoint temperature in evaporation equation
AIRVP	A term related to the vapor pressure of the air
SATVP	A term related to the saturated vapor pressure.
ARGNUM	Argument of exponential used in evaporation equation. An intermediate result
FSTNUM	Intermediate results in evaporation equation
PRTNUM	Intermediate results in evaporation equation
WINDFC	Wind function from aerodynamic equation

Terms found in THNWTR not found in MAIN

TSUM	Summing variable to sum monthly mean temperatures
TEMNC(I)	Long term averages of monthly mean temperatures
ANMENC	Annual mean temperature in °C.

FREEZE THAW ROUTINE (FREZR) GLOSSARY

ACZONE	The region between the 0°C. isotherm and the depth of unchanging temperature
AMNC	Mean air temperature in °C.
BTMSMP	The soil moisture percent by volume of the lowest soil layer
C	The volumetric heat capacity of the profile undergoing freezing or thawing.
CDFDNM	A name given to the term in the denominator in the equation to compute the conductivity of frozen soil.
CDYAIR	The thermal conductivity of air
CDYDS	The thermal conductivity of dry soil
CDYHUM	The thermal conductivity of humus
CDYICE	The thermal conductivity of ice
CDYSF	The thermal conductivity of frozen soil
CDYSF1	A convenience variable used in calculating the conductivity of frozen soil. Just an intermediate step
CDYSF2	Same or similar to the above
CDYWTR	Thermal conductivity of water
CHTHUM	Volumetric heat capacity of humus
CNUM	A summing variable used to determine the volumetric heat capacity of the freezing or thawing profile
DAFYRY	The value of the day in the year yesterday used for testing whether days of thaw are consecutive
DAOFFZ	The number of days in the current freeze period
DAOFTW	The number of days in the current thaw period
DATWMX	The longest number of days during which there was a thaw that did not completely thaw the ground. This is to determine how far frozen ground has been previously thawed.
DCEN	The depth of the center point of each soil layer
DEPPRC	A variable which computes the time when percolation of excess water to lower layer might occur on a day when the maximum temperature is above freezing and the minimum is below freezing
DFSVTY	Soil defusivity
DPDIF	The difference in depth between penetration of freezing or thawing in a given layer and the bottom of that particular layer
DPOFLY	The thickness of each layer. An array

DPTH2	A lower (shallower) limit to frost penetration used in iterating frost computation
DPTLFT	Variable used in computing composite porosity to determine soil conductivity to use in thawing from below
DPTTSM	A summing variable to compute the total depth
DUMMY	The depth reduction in frost due to heating from below
FTSMMX	The maximum value that the FSTSUM has reached to this point in the winter season.
FSTPRM	A recomputed value of FSTSUM (the frost index) to account for the insulating layer of snow whenever snow is melted or compacted
FSTSUM	The freezing index. The cumulative sum of average temperatures less than 0°C.
J30	A subscript
K	A subscript
KSOHUM	The ratio of the thermal conductivities (TC) of soil mineral to humus
KSOLAR	The ratio of the TC of soil mineral to air
KSOLIC	The ratio of the TC of soil mineral to ice
KSUM	A summing variable used to compute the mean thermal conductivity of the profile undergoing freezing or thawing.
KWATAR	Ratio of the thermal conductivities of soil mineral to air
KWATSL	The ratio of thermal conductivities of water and soil
ITHW	A counter used to check on freezing following a day of thawing
JLAY	A counter to allow penetration to go inch by inch through soil layers
L	The composite latent heat of the profile undergoing freezing or thawing
LDEEP	Potential latent heat per unit change of state at bottom of frozen layer
LITK	Litter TC
LITLOS	An adjustment to the total frost penetration to account for compaction of litter
LITMOS	The moisture in the litter layer
LITTER	The depth of the littler layer

LNUM	A summing variable used in determining the composite latent heat of the profile undergoing freezing or thawing
LYSMO	The day by day soil moisture content of each given layer, in inches of water
MAXPEN	The maximum depth of frost penetration
MLTSM	Counter to return ground to porous condition after 5 days of thaw conditions
MLTTIM	A counter to check for days of consecutive melt
MLTWTR	An index which is set to 1 when thawing begins over frozen ground which allows the surface layer to collect more water than the normal field capacity and increases the likelihood that concrete frost will form if it has not already.
MNSTG	The available water in the total profile
MNTPCY	The mean air temperature yesterday in °C.
NDLAY	The number of layers of soil or the lowest layer of soil (numbered from the surface) subject to direct evaporation
NOINFL	An index indicating that the watershed has been frozen imperviously
PCTMST	Percent moisture (average) of the lower layers
PENDNM	An intermediate calculation - the denominator in the penetration equation
PENNUM	The numerator in the penetration equation
PNCODE	The frost penetration code: 1. Frost does exist in the soil; 2. Soil protected from frost by snow; 3. Indeterminent results
PNDNMK	Frost penetration numerator term involving the annual mean temperature
PNNMSM	The penetration equation numerator for a snow layer
PRSTAV	Porosity of lowest layer or average of lower layers
S1FDNM	Intermediate result in soil thermal conductivity computation
S1LDNM	Similar to S1FDNM
SLCYFT	An intermediate term in computing the composite conductivity of moist soil.
SLCYSD	An intermediate term in computing the composite conductivity of moist soil
SLFDNM	An intermediate term in computing the conductivity of the soil below the frozen layer
SOLCNY	The composite soil thermal conductivity

SOLCN1	An intermediate term used in computing the conductivity of the soil below the frozen layer
SOLCN2	Same definition as above
SOLDNM	The numerator term in computing the composite TC
SPCHET	Composite volumetric heat capacity
THSMMX	The maximum value to this point in a season for the variable THWSUM
THWDNM	An intermediate term - the numerator in the thaw equation
THWMAX	The maximum depth to this point in the season which has thawed providing that a complete thaw has not already taken place. This term only has a value if the ground below it is still frozen.
THWNUM	An intermediate term in the thaw equation-numerator
THWSUM	A thaw index. Cumulative degree days above 0°C. for mean air temperatures
TPOFFT	A code: 1. Generally a quick freeze-granular frost expected. 2. Overmoist ground-needle ice and possible heaving expected. 3. Concrete frost expected with impervious soil. 4. Snow on ground is melting and some thawing of soil from below is expected. 5. Some thawing has taken place but the ground is still frozen down to a certain depth. 6. The ground is now free of frost 7. The ground is freezing but conditions make the type of frost indeterminent 8. There is something wrong in the computation.
TWDNMK	An intermediate term. The denominator in the thaw equation
WTPROS	Porosity of a layer weighted by the thickness of its layer
YTPOFT	Type of frost determined previous day.

Terms found in EVPADJ not found in MAIN

DIF	Potential evaporation less precipitation
DDIF	Intermediate difference

Terms found in TRNSLS not found in MAIN

N	Defined as NOLAY IN MAIN
NA	Shortened variable name for NDLAY
NB	Top layer not affected by direct evaporation
LOSI	The basic loss increment computed for use in partitioning loss as a function of depths and soil moisture present
RATI	The ratio of available water currently present in a layer to that present at field capacity
COMRAT	The total number of loss increments into which the loss must be divided on a given day.
PRTNFC	Depth partitioning function for evaporation

Terms found in ESTVAP not found in MAIN

EVAPU	Uncorrected estimate of evaporation based on the Thorthwaite equation
RATIO	Ratio of solar radiation on a plane surface on an equinox to that at any other times of year

Terms found in LAYER not found in MAIN

AK	Real (floating point) form of subscript K
ANL	Is a counter to sum the potential-evaporation increments of each inch of soil in each layer of soil
B	Counter used in determining the total loss increments in the region of soil where the loss is constant and independent of depth
CONFRC	Total potential loss units
DPT	Same as DLTAH in MAIN
DPTAV	Average thickness of layers
FRCSUM	Potential loss units in each layer
JKLM	Subscripts

JRT	A counter used in dividinglosses proportional to depth to a certain point and independent of depth below that point
ROOTBK	The point at which the loss changes from being proportional to depth to relatively independent of depth

Terms found in SMP not found in MAIN

ADEDH2	Water ad
DLTFC	Increment to field capacity when ground is frozen
DPT	DLTAH in MAIN
DAMXT	Maximum air temperature
COMRAT	Total potential loss units
EVAP1	Temporary name for EVAP to maintain entry point value
LOSI	Ratio of water to be loss by evaporation to the number of potential loss units
JLOOP	Counter to count number of passes through soil moisture of passes through soil moisture adjustment procedure. Used to prevent infinite looping
LADIF	Difference between the soil moisture in a layer at field capacity and currently present
N	Same as NOLAY in MAIN
NLAY	Layer subscript
NDLAY1	Temporary name for N or NOLAY
PCP1	Temporary name for PCP
PCPLOS	Amount of precipitation and/or snowmelt estimated to go to runoff and lost to infiltration due to the high intensity of supply.
RATDIF	The difference in the ratio of adjacent layers of soil moisture present to the potential soil moisture at field capacity
RATI	Ratio of soil moisture present in a layer to potential moisture present at field capacity
RATIO1	Ratio of soil moisture in a specific layer (i) to the potential soil moisture in the layer
RATIO2	Ratio of soil moistures in the (i + 1) layer
RATIO3	Ratio of soil moistures in the (i + 2) layer

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