SOIL MOISTURE REGIME AND SOIL WATER BALANCE UNDER UPLAND HARDWOOD FOREST DURING A PERIOD OF SUBNORMAL PRECIPITATION

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Richard L. Harlan

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Donald P. White
Major professor

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ABSTRACT

SOIL MOISTURE REGIME AND SOIL WATER BUDGET UNDER UPLAND HARDWOOD FOREST DURING A PERIOD OF SUBNORMAL PRECIPITATION

by Richard L. Harlan

The soil-moisture regime in a nine-foot profile under an old growth sugar maple-beech stand in south-central Michigan, was determined by the neutron-scattering technique during a $4\frac{1}{2}$ year period of subnormal precipitation. The predominant soil type within the study area is Hillsdale sandy loam, a Gray-Brown Podzolic soil developed under well-drained conditions on calcareous glacial till and under a humid continental-type climate. The area is a 15-acre woodlot that has been preserved in a relatively undisturbed condition. No past or present evidence of fire, grazing, or cutting, other than the removal of a few dead or down trees, is present.

The annual cyclic pattern of soil-moisture accretion and depletion is similar for each year, but varies in magnitude of fluctuations dependent upon the quantity and distribution of precipitation. Precipitation during the study period was 75 percent of normal. This deficiency is reflected in record low or near-record low groundwater levels during the period of study.

Infiltration of soil moisture advances behind a "wetting front", the moisture content of which is at or above "field capacity". Subsequent to the cessation of infiltration, redistribution of soil moisture within the profile results in almost uniform wetting in the upper five feet of the profile. Recharge occurs at a lesser rate below five feet even though the overlying horizons in the profile are not maintained at "field capacity".

Soil-moisture depletion begins with the breaking of vegetative dormancy in the spring and continues into the fall of the year. Infiltration of precipitation during the growing season normally does not penetrate to depths greater than three feet. Depletion occurs at all depths within the measured nine-foot profile but not at equal rates. The most readily available water is depleted at the greatest rates regardless of depth in the profile. The depletion rate decreases with both decrease in moisture content and with increase in soil depth.

Changes in total soil-moisture storage are partitioned into recharge, evapotranspiration, and deep seepage losses. The total rate of depletion is the sum of the loss rate due to evapotranspiration and the loss rate due to deep seepage below the depth of sampling. The deep seepage component is estimated from the total soil-moisture depletion curve and from assumed evapotranspiration rates. Based upon a mean growing season evapotranspiration rate of 0.07 inches/day, deep seepage losses account for an additional 0.5 inches of precipitation available as ground-water recharge during the growing season.

The mean annual evapotranspiration loss for the 1962 through 1965 water years was 22.27 inches or 96 percent of the mean annual precipitation. Deep seepage losses account for the remaining 4 percent.

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By

Richard L. Harlan

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VITA

Richard L. Harlan

Candidate for the Degree of
Doctor of Philosophy

Final Examination: January 27, 1967

Guidance Committee:

Dr. A.E. Erickson
Dr. W.J. Hinze
Dr. G. Schneider
Dr. D.P. White, Chairman
Dr. J.W. Wright

Dissertation: Soil-moisture regime and soil-water balance under upland hardwood forest during a period of subnormal precipitation.

Outline of Studies:
Major subject: Forest Hydrology
Minor subjects: Soil Science, Geology

Biographical Items:
Born July 7, 1939. Denver, Colorado

Undergraduate Studies:
Grinnell College, 1957-1958
University of Colorado, 1958-1961
B.A. Geology, 1961

Graduate Studies:
Michigan State University, 1961-1964
M.S. Forestry, 1964

Michigan State University, 1964-1967
Ph.D. Forestry, 1967

Experience:
Graduate Research Assistant, Department of Forestry, Michigan State University, 1964-1966; Research Officer, Forest Research Branch, Canada Department of Forestry and Rural Development, Calgary, Alberta, 1966 to date.

Member:

American Geophysical Union
Canadian Institute of Forestry
Sigma Xi
Society of American Foresters
Xi Sigma Pi

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CHAPTER I

INTRODUCTION

Many parts of the United States are facing water problems due to increasing requirements for large supplies of high quality water by an expanding population and rapid industrial growth. In order to solve these problems, more basic information is needed about surface and subsurface water -- its availability, quantity, quality, and use. In addition, quantitative data are needed on the influence of different vegetation types upon the total water economy of water-producing areas. Numerous soil-vegetation-water-climatic relationship investigations have been made, but this work has been concerned primarily with the influence of water supply upon plant growth and development. In hydrology, however, the principal focus is on the influence of the plant upon the water supply.

The water yield from surface and subsurface sources from a given land area is equal to the total precipitation less the so-called "natural water losses". These losses include both water returned directly to the atmosphere by evaporation and that taken up by plant life and subsequently transpired to the atmosphere. Evapotranspiration, the total water losses from water, vegetation, and soil surfaces and transpired from vegetation, makes the initial utilization upon precipitation. In effect, it regulates the amount of precipitation, after loss from surface runoff, available to the soil-moisture reservoir and as ground-water recharge.

With the increasing water requirements for domestic, agricultural, and industrial uses, hydrologists must study the unsaturated soil-water zone which is the dominant link between ground and surface water. The unsaturated zone must be defined in terms of those hydraulic characteristics that regulate fluid movement through it and the influence of vegetation upon the quantity of water available for man's needs.

As rainfall or snowmelt infiltrates into the soil, it is stored within the soil profile in order to replenish the soil moisture deficiency created within the root zone by evapotranspiration losses. Once this moisture deficiency has been satisfied, any surplus moisture moves downward beyond the depth of rooting and ultimately reaches the ground-water table.

The volume of surface runoff and ground-water recharge is only a portion of the total precipitation; it is the residual after the processes that lead to evaporation and transpiration have taken their toll. This toll, "the natural water loss", is, for the most part, not directly measureable and is therefore, traditionally determined by the application of the principle of conservation of mass to the hydrologic equation--that equation balancing the hydrologic budget. The hydrologic equation is:

$$\text{Precipitation} + \begin{matrix}\text{Surface and}\\ \text{subsurface}\\ \text{inflow}\end{matrix} = \text{Evapotranspiration} + \begin{matrix}\text{Surface}\\ \text{Runoff}\end{matrix} + \begin{matrix}\text{Change in}\\ \text{soil-mositure}\\ \text{storage}\end{matrix} + \begin{matrix}\text{Change in}\\ \text{ground-water}\\ \text{storage}\end{matrix}$$

This represents the balance between inflow of water, outflow, and change in soil-moisture and ground-water storage.

Theoretically all of these quantities may be measured directly or calculated except evapotranspiration, which can be computed (DeWiest, 1965).

However, in such glaciated regions as Southern Michigan, where there is a mosaic of vegetation, geologic, and hydrologic conditions, determination of the influence of a single vegetation cover type, occupying only a minor portion of a hydrologic unit, becomes obscure. Under these conditions, the influence of one cover type on the regional water balance must be studied from an alternate approach, either from the study of the soil-moisture regime, fluctuations in ground-water storage, or by empirical or theoretical methods of determining the water balance.

The first of these approaches was used in this study in determining the soil-water balance for an upland hardwood forest during a $4\frac{1}{2}$ year period of subnormal precipitation. Few soil-moisture studies reported in the forestry literature have gone beyond the descriptive phases and have made any attempt to estimate evapotranspiration or to calculate the soil-water budget. Furthermore, little attention has been given to the examination of the applicability of existing soil-moisture theory, developed primarily under agricultural and disturbed soil profiles, to forest-soil conditions.

Knowledge of the factors regulating soil-moisture retention and movement are necessary in order to treat the unsaturated zone as a dynamic part of the hydrologic cycle and to formulate a basis for the scientific management of vegetation for increased water yields.

CHAPTER II

PURPOSE AND SCOPE

The soil-moisture regime reflects the influence of vegetation upon water yield. During periods when the soil is wet, it functions principally as a stable and porous passageway to the ground-water reservoir for precipitation. When it is dry, however, it stores a great part or often all of the precipitation and allows only a small portion or none of the precipitation to reach the ground-water reservoir (Lull and Fletcher, 1962).

Past hydrologic studies have only begun to provide an insight into this complex physical and biological system and the intricacies of soil-moisture movement and mechanics of ground-water recharge. Adequate hydrologic data in these areas are requisite to the regulation, development, and assessment of the quantity and quality of available water within each hydrologic unit. Where man's needs for water results in the full development or major rearrangement of water distribution, the dynamic effects on water distribution within the unsaturated zone must be considered in the evaluation of the total water resource.

The basic approach in this study of the soil-moisture regime and soil-water balance under old-growth hardwoods is to define the primary components of the physical-biological system, to observe how the system functions, and to explain why they function in the observed manner. The objectives are, therefore:

1). to define the soil-vegetation-water interrelationships,

2). to quantitatively evaluate evapotranspiration and ground-water recharge, and

3). to evaluate and integrate the results in terms of previous investigations and existing soil-moisture theory.

CHAPTER III

REVIEW OF LITERATURE

Several classifications for describing the occurrence of fluids in porous media exist in the literature; the most widely accepted by hydrologists is that suggested by Meinzer (1923). Meinzer subdivided the porous material at or near the earth's surface into two general zones. The lowermost zone, the zone of saturation, is defined as "the zone in which the functional permeable rocks are saturated with water under pressure equal to or greater than atmospheric" (Amer. Geol. Inst., 1960). The uppermost zone, the zone of aeration, is defined as "the zone in which interstices of the functional permeable rocks are not filled (except temporarily) with water. The water is under pressure less than atmospheric" (Amer. Geol. Inst., 1960).

Within the zone of aeration three states of matter -- the solid soil matrix, the liquid-soil water, and the soil atmosphere -- exist simultaneously. Consequently, the occurrence and movement of water in the zone of aeration are more complex than saturated flow occurring within the zone of saturation (Remson and Randolph, 1962). The zone of aeration is subdivided into three subzones: the lowermost - the capillary fringe, the uppermost - the belt of soil water, and the central - the intermediate zone.

The capillary zone is defined as "a zone, in which the pressure is less than atmospheric, overlying the zone of saturation and containing capillary interstices some or all of which are filled with water that is continuous with the water in the zone of saturation but is held above that zone by capillarity acting against gravity." (Amer. Geol. Inst., 1960). The zone of capillarity or capillary fringe may extend upward into the root zone or to the ground surface. In both cases, large quantities of water may be lost to the atmosphere through the combined processes of evapotranspiration.

The uppermost zone, the belt of soil water, is defined as "that part of the lithosphere, immediately below the surface, from which water is discharged into the atmosphere in perceptible quantities by the action of plants or by soil evaporation" (Amer. Geol. Inst., 1960). The belt of soil water has been studied primarily from its relation to plant growth. The available moisture within the belt of soil water is seldom adequate at all times for growth of trees and other vegetation to continue at optimum rates consistent with climatic and other site conditions (Nash, 1962). During periods of inadequate precipitation, soil-moisture stresses develop and result in a reduction of the growth rate of vegetation. The moisture content within this zone is subject to wide seasonal variations due to depletion by evaporation and transpiration and accretion from precipitation.

The intermediate zone is "that part of a zone of aeration that lies between the belt of soil water and the capillary fringe" (Amer. Geol. Inst., 1960). It is invisioned as that segment of the zone of aeration in which water occurrence is not changed rapidly by changing conditions at either the land surface or at the water table. "Both the belt of soil water and capillary fringe are limited in thickness by local conditions such as the character of the vegetation and texture of rock or soil but the intermediate zone is not thus limited" (Meinzer, 1923).

Evapotranspiration.

Total evaporation, comprising evaporation from both the soil and plant surfaces as well as transpiration, from an area supporting vegetation is generally considered as one phenomena - evapotranspiration. In studying the water balance of a drainage basin, it is usually impractical to separate evaporation and transpiration. The meterological factors which affect evaporation also influence the transpiration process.

Although transpiration by vegetation is an evaporative process, it differs from other evaporation situations by reason of 1). the location of the evaporative surface above the ground; 2). the considerable distance between the source and the evaporative surfaces; and 3). the degree of control over the process by the living plant (Hoover, 1962). Transpiration is the most complex of all evaporative processes, for it is regulated by the meteorological situation, the conditions within the plant, and the water relations of the soil. When soil moisture is in short supply, the plant can decrease the amount of transpiration by closing the stomata. At the same time the plant must give off water in order to live, since essential food materials are carried from the soil by the sap stream (Geiger, 1965).

Evapotranspiration is also one of the most important factors in determining the water economy of an area. Since a certain amount of evaporation demands a definite amount of heat energy, it provides a link between the water budget and the heat or energy budget (Geiger, 1965). Evaporation is the most important of all factors in the energy budget after radiation. Energy is initially supplied for evapotranspiration by solar radiation with heat transfer both from the air to the evaporating surface and heat stored within the evaporating body itself (Thornthwaite and Mather, 1955).

According to Jeffrey (1964), evapotranspiration processes should be evaluated under a wide variety of conditions related to the atmospheric phase of the hydrologic cycle to reveal their quantitative importance in the mass transfer of water.

Gates (1962), in his examination of energy - transpiration relationships, noted that the rate of transpiration increases rapidly with increase in leaf temperature and leaf-air temperature differentials. The magnitude of differences varies with the amount of radiant energy absorbed by the leaf and the relative ambient air movement. Energy exchange, therefore, takes place between the leaf, atmosphere, soil, and surrounding foliage. The amount of energy absorbed by the foliage varies with the species and the stand conditions. The proportion of the total energy used in transpiration varies greatly with a number of factors including leaf orientation, stand structure, stand density, and available water (Jeffrey, 1964). Therefore, the soil-water content plays a significant role in natural processes that influence the energy balance at the earth's surface.

Potential Evapotranspiration.

In recent years, empirical and semiempirical methods for estimating the energy balance and "potential evapotranspiration" have received considerable attention, principally by Thornthwaite in the United States and Penman in England. Thornthwaite (1948) introduced the term "potential evapotranspiration" to express the combined effects of evaporation and transpiration occurring from plants and soil. The term, "potential evapotranspiration", is defined as the amount of water which will be lost from a surface completely covered with vegetation if there is sufficient water in the soil at all times for the use of vegetation (Baver, 1956). The best known empirical methods of determining potential evapotranspiration are those of Blaney and Criddle (1950), Penman (1948), and Thornthwaite (1948).

The concept of potential evapotranspiration implies a non-limiting water supply. While this condition rarely exists, particularly on a regional basis, the concept is useful as an index of climate. Knowledge of both precipitation and potential evapotranspiration is a key factor in ascertaining the natural water loss. The natural water loss from an area varies not only with potential evapotranspiration, but also with the availability of water.

Jeffrey (1964) points out that while it is true that the same climatic variables, radiation, humidity, wind, and temperature are involved in transpiration as in evaporation from a free water surface, these variables do not take into consideration physiological and morphological differences between plant species.

In England, Penman (1948) developed a method of using the energy balance approach for estimating evapotranspiration. This method treats evaporation from soils and crops as a purely physical process and requires knowledge of the following four meteorological parameters which are: a). duration of bright sunshine or net solar radiation, b). air temperature, c). air humidity, and d). wind speed.

The Thornthwaite climatic water balance provides a method to evaluate the moisture factor in climate. By means of a water balance bookkeeping procedure using the data of air temperature, precipitation, and location, it is possible to estimate, either on a monthly or a daily basis, the periods and amounts of water surplus, water deficit, soil-moisture storage, and runoff (Mather, 1961).

Potential evapotranspiration is controlled solely by available energy. When there is no shortage of water, actual and potential evapotranspiration should be equal; when there is a shortage of available moisture, actual evapotranspiration will be less than potential evapotranspiration (Mather, 1961).

The soil-water budget may be regarded as being a balance between what is added as a result of precipitation and what is lost through evaporation and transpiration. Therefore, soil moisture may be regarded as a bank account - precipitation adds to the account; evapotranspiration withdraws from it.

When the soil-moisture content is at or exceeds a critical moisture content, any water that is added to it by precipitation is lost by downward percolation from the soil. Movement of water through the soil-plant-atmosphere system occurs in response to an overall free energy gradient (Ehrenreich, 1962). Free energy gradients are determined by total pressure, temperature, concentration of water, and atmospheric pressure. Water migrates from regions of higher to lower potential energy as it moves within the soil, into the plant roots and through the plant to the leaves. The potential energy continuously decreases until the water reaches the leaves where evaporation occurs.

Water entering the soil surface is trapped temporarily in the porous mass of solid particles. Its immediate and subsequent disposal depends upon the characteristics of the soil, interactions between soil and water, and the influence of vegetation (Colman, 1953).

Soil-moisture availability.

The major factor distinguishing actual and potential evapotranspiration is soil-moisture availability. Considerable controversy exists in the literature as to whether water is equally available to plants over the range between permanent wilting point and "field capacity". "Field capacity" is defined as the "amount of water remaining in a well-drained soil when the velocity of downward flow into the unsaturated soil has become small" (Soil Sci. Soc. Amer., 1956). The processes of infiltration, drainage of water, diffusion of gases, conduction of heat, and movement of salines all depend upon the amount of water present in the soil.

Hoover (1962) notes that the removal of soil water by vegetation during the growing season rapidly depletes soil moisture. Transpiration removes water throughout the entire root zone at about the same rate, while evaporation depletes the surface layers most rapidly and removes water only very slowly from soil depths greater than two feet. Fraiser (1957) studied soil moisture on a variety of sites in Ontario and found that the surface soil layers dried out more rapidly than the deeper layers in the profile. In addition, Hoover, Olson, and Green (1953), working in a young loblolly pine plantation in the South Carolina Piedmont, observed that water was withdrawn from the zone in which it was most readily available, regardless of depth.

Gaiser (1952) found that the rate of moisture extraction from forest soils was such that all horizons of the profile simultaneously approach wilting point.

A five year investigation of soil-moisture fluctuations under forest cover, broomsedge field, bare soil, and on old-field sites on the Calhoun Experimental Forest near Union, South Carolina by Metz and Douglass (1959), indicate that forest types appear to deplete soil moisture to depths of at least 66 inches at relatively the same rate regardless of species.

As the soil dries, the rate of evapotranspiration decreases. According to Thornthwaite and Mather (1955), "when one-half of the (available) water is gone, the rate of evapotranspiration falls to one-half of the potential rate and plants begin to suffer from drought". Transpiration rates begin to be reduced at comparatively low soil-moisture tension but can continue at greatly reduced levels within the wilting range (Slatyer, 1957; West and Perkman, 1963).

Drought is related to soil-moisture content and begins when the available soil mositure is diminished so that vegetation can no longer absorb water from the soil rapidly enough to replace that lost to the atmosphere by transpiration. Soil-moisture deficits during the growing season, therefore, result in reduced transpiration rates. In many areas summer precipitation contributes little to streamflow, being largely utilized by evapotranspiration.

When soil moisture is high or maintained at optimum levels, land management practices and soil type or structure have little effect on the rate of evapotranspiration. When the root zone of the soil is well supplied with water, the amount used by the vegetation depends more on the amount of solar energy received by the surface and the resultant temperature than on the kind of vegetation growing in the area.

Penman (1963) summarized the results of Celjniker (1957) as follows:

a). When the soil-water content is high, a small decrease causes an increase in transpiration rate, a change associated with increase in soil and air temperature. Low transpiration values may be an indication of inadequate aeration limiting root activity.

b). There is then no major change in transpiration rate until the soil-water content falls to about three-quarters of the moisture equivalent and then there is a sharp fall in transpiration rate, attributed to a sudden decrease in the mobility of soil water.

c). Therefore, in drier soils, the transpiration rate is maintained at a low level.

Consequently, soil-water is not equally available throughout all ranges of soil-moisture contents encountered in the field. The availability decreases both as the soil-moisture content increases because of decreased aeration and as the moisture content approaches wilting point.

In addition, the amount of water available to the plant varies with the root distribution and depth of rooting as well as the mobility of water within the soil profile. Where roots reach a permanent groundwater table or into the capillary fringe, transpiration will take place at levels closer to the "maximum" throughout the growing season.

In deep well-drained soils, the depth of root penetration would be expected to be an important determinant of the annual quantity of transpiration. If on deep soils an adequately protective cover of shallow-rooted plants can be substituted for an existing cover of deep-rooted plants, one would expect transpiration to be reduced with a corresponding increase in usable water yield.

Measurement of Evapotranspiration.

No presently developed method of measuring actual evapotranspiration has proved to be completely satisfactory under all conditions encountered in the field. The best known technique for measuring actual water losses due to evapotranspiration is the use of a lysimeter. This technique has been widely used in experimentation on water utilization by agricultural crops and by grassland vegetation. However, this method is of limited applicability under mature forest conditions due primarily to the size requirements and economic limitations of hydrologic research. The main deficiencies of the lysimeter technique are in the edge effects and in the artificial conditions imposed by interruption of the soil column (Jeffrey, 1964).

Considerable interest has been shown in the determination of transpiration by riparian vegetation and evaporation from the stream surface and adjacent riparian areas. Theis and Conover (1959), Troxell (1936), White (1932), and most recently Reigner (1966), proposed systems for measuring the quantity of water involved in diurnal fluctuations of a water table due to evapotranspirational withdrawal by riparian vegetation. Urie (1965) developed a method of analysis for determining the water yield as net ground-water recharge and evapotranspiration from areas of homogeneous vegetation in glacial sands in the northwestern part of the Southern Peninsula of Michigan. Fluctuations in ground-water storage were evaluated by separating water-table fluctuations into recharge, seepage flow recession, and evapotranspiration.

All of the techniques for estimating the evapotranspiration losses from areas of riparian vegetation are restricted to areas in which the water table is close to the ground surface. In addition, the aerial extent and the hydrologic properties of the aquifer must be defined. Under certain conditions where the water table is below the depth of rooting, the quantity of recharge to the ground-water table may be determined and the evapotranspiration calculated as the difference between precipitation and the total of ground-water recharge and surface runoff.

Soil-moisture budgets have been used extensively within recent years - especially since the introduction of the neutron-scattering technique of soil-moisture measurement. Increments of soil moisture from precipitation are measured throughout the growing season, antecedant and subsequent soil moisture is determined and evapotranspiration calculated as a residual.

This method, however, fails to account for deep seepage of soil moisture and its contribution to the ground-water reservoir. Hewlett (1961) showed that the existence of a persistent tension gradient adjacent to streams may yield considerable quantities of water to stream flow although the soil-moisture content was below "field capacity". In addition, during periods of high soil-moisture contents, evapotranspiration is occurring but no reduction in soil-moisture content can be detected (Jeffrey, 1964). This method is not applicable where the roots of vegetation extend to the water table or to the capillary fringe.

Willardson and Pope (1963) proposed a system to correct for deep seepage by measuring seepage losses on unvegetated plots. In a series of experiments on furrow irrigation near Milford, Utah, soil-moisture measurements were taken to establish the quantity of water stored in a six-foot profile by an application of irrigation water. Measurements were made on the day before irrigation, on the first, second, and thirteenth days following irrigation. A moisture loss of 1.7 inches from the six-foot profile was noted during the first two days and the profile continued to lose moisture over the two week period of measurement. Analysis indicated that moisture loss continued for at least two weeks at an exponentially decreasing rate.

Similarly, Ogata and Richards (1957) of the U.S. Salinity Laboratory found that soil drained at an exponentially decreasing rate. The relationship is expressed by the equation

$$W = AT^B$$

in which W is the moisture content of the soil, A and B represent constants determined from field data and T is the time in days after irrigation.

Wilcox (1959; 1960) measured soil-moisture depletion on tarpaulin-covered plots and on nearby vegetated plots. As the result of his experiments, Wilcox concluded that evapotranspiration losses measured by soil-moisture depletion contains some unknown quantity of deep seepage or drainage. The quantity of deep seepage would vary on different soil types under the same vegetative cover.

In summary, little is known of the actual physical processes involved. However, soil moisture moves in an unsaturated condition under the influence of temperature and moisture gradients; therefore, it is reasonable to expect soil-water movement in response to gravity gradients and possible movement beyond the root zone or sampling depth (Willardson and Pope, 1963).

Swanson (1962) developed an approach using sap velocity rate as an index of transpiration. A heat pulse is introduced into a tree stem and the velocity of heat transfer recorded by means of probes inserted into the sapwood. This method is currently receiving much attention in the estimation of transpiration losses by different vegetation types in Arizona, Colorado, California and Canada.

Indirect methods based upon the energy concepts and turbulent diffusion of water vapor have been increasingly used in recent years. The heat budget approach to measuring evapotranspiration consists of measuring net radiation, soil-heat flux, air temperature, and vapor pressure gradients over the vegetative surface (Jeffrey, 1964). The energy approach to determination of the water budget is of limited applicability until more is learned about air dynamics and heat exchange under forest conditions.

Interception, Throughfall, and Stemflow.

Interception is a function of canopy characteristics, evaporation from foliage during rainfall, and the size of the storm. Each vegetative surface has a characteristic water storage capacity, which is dependent upon the nature and density of the cover. Once this storage capacity is satisfied, interception can only be increased by during-storm evaporation from the leaf surface and its replacement by incident precipitation (Jeffrey, 1964).

Helvey and Patric (1965) reviewed and summarized all the available references on rainfall interception in the eastern hardwood region and assessed the need for further research.

Interception is a two-part process consisting of 1). gross rainfall caught by the canopy and redistributed as throughfall, stemflow, or evaporation from the standing vegetation, and 2). that which is caught by the litter layer and evaporated without adding to moisture in the mineral soil (Helvey and Patric, 1965). Total interception loss, therefore, includes water retained on both the canopy and litter and subsequently returned to the atmosphere.

The quantity of litter interception is determined by the amount of litter on the ground and also by its drying rate, which is in turn determined by the local climate and moisture-holding capacity of the litter.

Helvey and Patric (1965) found a close agreement in the work of many investigators under a diversity of study conditions in the eastern hardwood regions. Throughfall, stemflow, and interception measurements were surprisingly similar over a wide range of canopy characteristics and predictable from gross rainfall records.

Although a certain quantity of the gross precipitation is lost by interception, the net effect is probably lessened due to reduction in the rate of evapotranspiration.

CHAPTER IV

INSTRUMENTATION AND DESCRIPTION OF STUDY AREA

The general study area lies within the Glaciated Central Ground-water region described by McGuinness (1963) and largely in the Eastern Lake Section of the Central Lowland physiographic province. The entire state of Michigan was glaciated during the Pleistocene epoch and consequently, the glacial geology of the region controls the distribution of water, soil development, and the surface and subsurface hydrologic characteristics.

Under these conditions, geologic factors are commonly as important as climatic factors in determining the magnitude and distribution of the natural water losses. These losses are the result of a combination of moisture availability and climatic factors. Moisture availability is controlled primarily by precipitation distribution and secondarily, by geologic factors. The permeability of the soil mantle initially determines how much precipitation will contribute to surface runoff and how much it will contribute to soil moisture and ground water (Crippen, 1965).

The study area occurs within a region in which agricultural and industrial activities prevail. Only 17% of the area in the southern half of the Southern Peninsula of Michigan is forested. Most of the forested land is in small, privately-owned tracts.

The study area, Toumey Woodlot, is a 15 acre tract located in the northwest corner of Ingham County in south-central Michigan (NE$\frac{1}{4}$, SE$\frac{1}{4}$, Sec.30, T4N, R1W. of the Michigan Meridian). The area is an old-growth sugar maple-beech stand that has been preserved in a nearly undisturbed condition. The only cutting that has taken place within the woodlot has been the occasional removal of dead or down trees for fuelwood. No past or present evidence of fire or grazing is evident.

In 1940, a 60 foot X 60 foot grid system was established within the study area to facilitate data collection in a long-term study of ecological conditions in a "relatively undisturbed sugar maple-beech stand". The grid numbering system, as shown in figure 1, is used for purposes of plot location and description.

Because of the small size of the woodlot, the microclimatic conditions within the woodlot are influenced by conditions in the surrounding farmlands and roadways. In order to minimize these effects, a 20-foot wide coniferous windbreak was planted in 1941-42 along the north edge of the woodlot facing open pasture. A 44-foot wide strip along the west side containing brush and pole-sized hardwoods was planted at the same time (Schneider, 1963). The woodlot was set aside at this time as an ecological reserve by Michigan State University for forestry and biological research.

No additional protection, however, was provided on the south side of the woodlot from outside influence from the cleared lands. Consequently, conditions within this portion of the woodlot may be more extreme and harsher than toward the center or northern part of the woodlot. Allowance for border effects was made, however, in the positioning of plots within the study area.

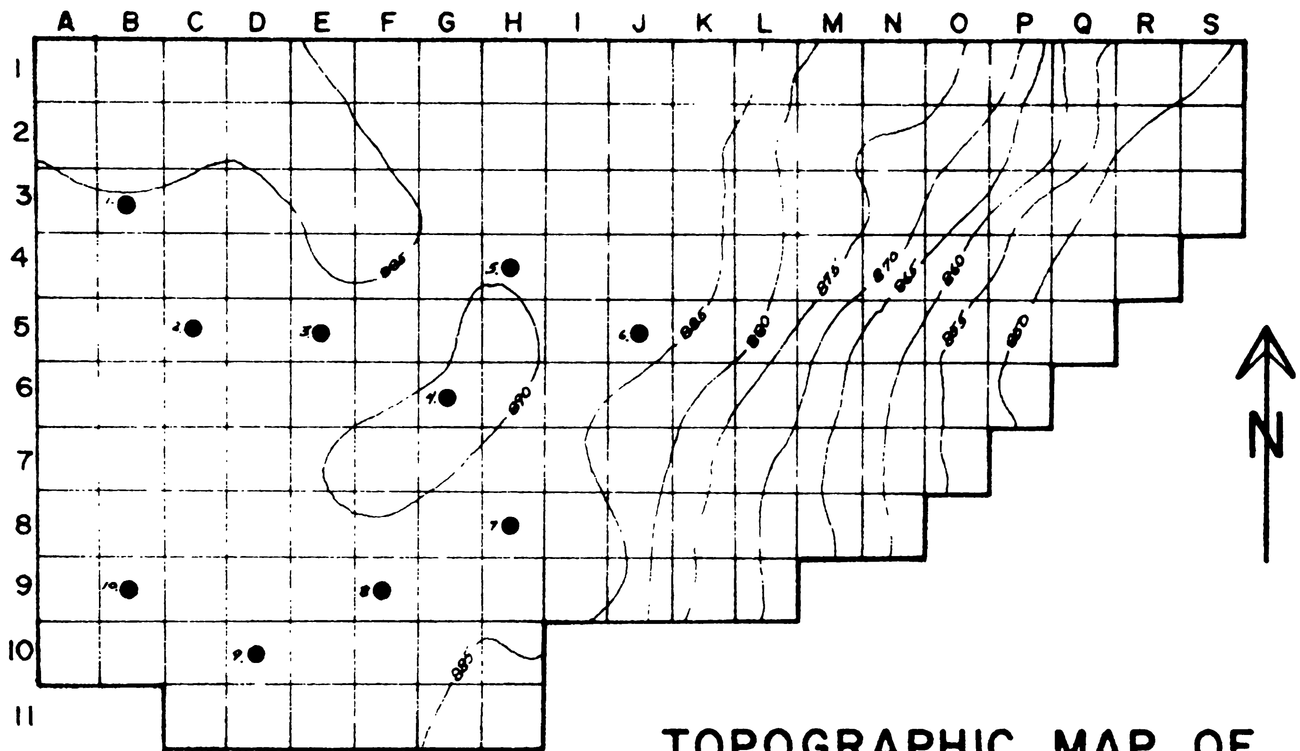
The ecological aspects of Toumey Woodlot are discussed by Schneider (1963, 1966).

Instrumentation.

Soil-moisture measurements by the neutron-scattering technique were begun in 1961 as part of an ecological study of Toumey Woodlot. Ten permanent sample points were established so as to sample the predominant soil type, Hillsdale sandy loam, occurring within the woodlot (figure 2). The locations of the permanent sample points are shown in figure 1.

At each soil-moisture sample point, a 10-foot, 1-3/4 inch inside diameter, steel conduit was driven into the soil for the entire 10 feet depth. Installation involved the use of a Veihmeyer tube, which was driven into the soil for the entire 10-foot depth, to probe the site for obstacles which might interfere with the movement of the access tubes into the hole.

Figure 1. Map of Toumey Woodlot showing grid system and location of the soil-moisture sample points. (Michigan State University, East Lansing)



LEGEND

● Soil-moisture access tube location

885 Contour

B-1 etc. Grid-system notation

TOPOGRAPHIC MAP OF TOUMEY WOODLOT

SCALE IN FEET

0 60 120

The access tubes were then driven into the soil, obtaining as tight a fit as possible (Schneider, 1963). This procedure is in accordance with recommendations of Horonjeff and Javete (1956). Horonjeff and Javete in a study of four different methods for installing access tubes for the measurement of soil moisture by the neutron-scattering technique found that the lowest error was observed as compared to gravimetric samples, when the access tubes were driven into the soil.

Soil-moisture measurements were taken at 1 foot intervals, beginning 1 foot below the ground surface and extending to 9 feet. It is assumed that little error resulted from the neutron loss to the atmosphere by beginning measurements 1 foot below the soil surface.

The statistical reliability of nuclear moisture measurements have been studied by many workers. The classical work by Jarrett (1946), which discusses statistical methods used in measurement of radioactivity, is used as the basis of many of the more recent studies (Merriam, 1960; Merriam and Knoerr, 1961).

The accuracy of the neutron soil-moisture meter is dependent upon: 1). the range of moisture contents at which the readings are made, and 2). the degree of confidence desired in interpreting the statistical accuracy (Nuclear-Chicago Corporation, n.d.).

According to graphs prepared by Merriam (1962), if an acceptable random counting error expressed as percent moisture by volume is 0.5 percent and the confidence level is 95 percent, a counting time of 0.60 minutes is required at 10 percent moisture by volume and 1.1 minutes for a moisture content of 20 percent by volume. Based upon these determinations for average moisture conditions, a one minute counting interval was used for the soil-moisture measurements.

The soil-moisture contents observed did not exceed 25 percent by volume for the period of study. Therefore, a random error equivalent to 0.10 inch of water per foot of soil was obtained from the graphs prepared by Merriam (1962).

The primary disadvantage of the neutron method is the depth of discrimination between soil layers without over-lapping measurements is about 12 to 36 inches, depending upon moisture conditions. The sphere of influence may be calculated by the formula,

$$\text{Sphere of influence} = \sqrt[3]{\frac{100}{\%\ \text{moisture by volume}}}$$

as presented by Douglass (1962b).

Geology and Topography.

The study area lies within a region typical of Southern Michigan, characterized by gently to roughly undulating glacial plains and moraines, numerous swampy depressions and small lakes

(Mich. Water Res. Comm., 1961). The major geomorphic features are represented by till plains, terminal moraines, outwash plains, glacial lake-beds, and broad alluvial valleys. The surface configuration represents an early phase in the geologic erosional cycle in that a complete drainage system has not been developed (Veatch _et._ _al._, 1941).

The surficial geology consists of a thick accumulation of glacial drift, approximately 70 to 150 feet thick in the general study area. The surficial material is quite variable both in the horizontal and vertical directions. It consists primarily of inter-bedded fluvial and morainic deposits, ranging in texture from coarse, water-deposited sands to unsorted sandy clay till. The glacial deposits are sufficiently thick to completely mask the direct influence of the underlying bedrock upon soil development.

Climate.

The climate is of the humid continental type, characterized by a fairly even distribution of precipitation throughout the year. Precipitation, generally associated with frontal movements or convectional currents, may occur in the form of rain, snow, sleet, frost, dew, or fog.

The mean annual precipitation for the Upper Grand River Basin, which includes the study area, for the 1900 through 1959 period is 31.13 inches. The maximum annual precipitation for this period was 38.97 inches occurring in 1905, and the minimum annual precipitation was 18.99 inches recorded in 1930 (Mich. Water Res. Comm., 1961). The mean annual snowfall is approximately 40 inches. Snowmelt makes a significant contribution to the total ground-water recharge. The Southern Michigan area generates approximately 12 inches of total runoff (McGuinness, 1941, 1963).

The mean annual potential evapotranspiration for the 1931-1953 period is approximately 650 mm or 25.6 inches (Messenger, 1962). The mean annual evaporation to be expected from shallow lakes and reservoirs is 23 inches (Kazman, 1965).

The Great Lakes have a stabilizing effect on temperature and becuase of the prevailing westerly winds, winters are milder and summers are cooler than interior states. The mean annual temperature for the 1931-1952 period was 48°F and for all previous years of record, 47.1°F (Mich. Water Res. Comm., 1961). The length of the growing season is approximately 150 days.

Soils.

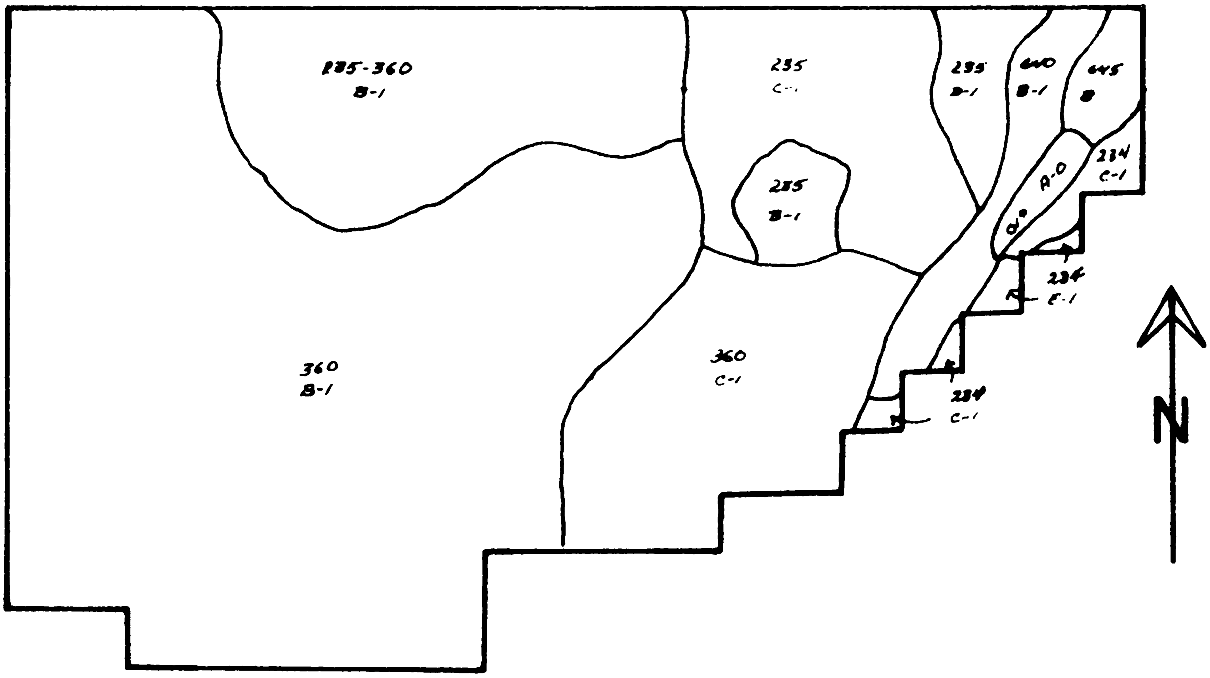
The prominent soil series that occur within the study area are the Hillsdale and the Spinks. Both series are well-drained Gray-Brown Podzolic soils developed in medium to fine-textured, calcareous glacial material of late Wisconsin age. The texture of the predominant soil series ranges from loamy sand to clay loam.

The Spinks soils are coarser textured than the Hillsdale series and commonly support a more xerophytic type vegetation. According to Veatch (1941), the native vegetation on the Hillsdale and Spinks soils is predominantly oak-hickory. However, the soils within Toumey Woodlot appear to have a higher moisture-holding capacity and therefore, the site is more favorable for growth of more mesophytic species such as sugar maple and beech.

The Hillsdale sandy loam is the major soil type in the study area, representing approximately two-thirds of the area. The Spinks loamy fine sand occupies approximately one-fourth of the study area; whereas Spinks loamy sand, Conover silt loam, Conover loam, and Carisle muck occupy only minor areas of the woodlot, primarily on the eastern part. The distribution of soil types within the study area is shown in figure 2. The study of the soil-moisture regime and soil-water balance was confined to the major soil type, the Hillsdale sandy loam.

The soil texture varies widely as is typical of glacial morainic material. However, the degree of variability remains within the limits of the sandy loam textural class for the Hillsdale series and within the limits of the loamy sand textural class for the Spinks series. For example, based upon six randomly located samples in the Hillsdale sandy loam, the range in percentage of sand in the C horizon ranges from 44 to 66 percent with a mean of 55 percent. In the same set of samples, the percentage of clay

Figure 2. Soil type map of Toumey Woodlot, (Michigan State University, East Lansing).



SOIL TYPE MAP OF TOUMEY WOODLOT

SCALE IN FEET

0 60 120

LEGEND

SOIL TYPE

010 Carlisle muck
234 Spinks loamy sand
235 Spinks loamy fine sand
360 Hillsdale sandy loam
640 Conover silt loam
645 Conover loam

SLOPE CLASS

A 0 - 2 %
B 2 - 6 %
C 6 - 12 %
D 12 - 18 %
E 18 - 25 %

EROSION CLASS

0 None or accumulation
1 Little or none

(after Schneider, 1963)

sized particles ranges from 11 to 22 percent with a mean of 18 percent. Although these soils are mapped as the same soil type, the differences in percentages of the soil fractions undoubtedly have a very strong influence on the productivity of the site and the soil-water relations.

The percentage of soil separates for an "average" soil profile for the Hillsdale sandy loam and the Spinks loamy sand as presented by Schneider (1963) are given in tables 1 and 2, respectively.

Vegetation.

Southern Michigan prior to the first white settlements in the 1830's, was covered by a dense hardwood forest. With the influx of settlers, however, much of the original forest was cleared by the early 1900's when a maximum area was under cultivation.

The composition of the presettlement vegetation was correlated with soil type. The beech-maple region of Michigan is characterized by the development on mesic sites of a climax in which sugar maple (Acer saccharum Marsh) and American beech (Fagus grandifolia Ehrh.) are the dominant trees in the canopy. This vegetation type is restricted to the southern Gray Brown podzolic region of Michigan (Schneider, 1966). As the soils become more sandy and consequently more subject to drought, the percentage of oak and hickory increases. The oak-hickory types are, therefore,

Table 1. Mechanical analysis of an "average" Hillsdale sandy loam profile in Toumey Woodlot (after Schneider, 1963).

Table 2. Mechanical analysis of an "average" Spinks loamy fine sand profile in Toumey Woodlot (after Schneider, 1963).

HILLSDALE SANDY LOAM

Horizon	Depth - Feet	% Sand	% Silt	% Clay	Texture
A_1	0.0-0.3	63.4	25.2	11.4	Sandy loam
A_2	0.3-1.3	62.1	25.8	12.1	Sandy loam
B_1	1.3-1.8	63.1	16.2	20.7	Sandy clay loam
B_2	1.8-4.0	65.3	10.5	24.2	Sandy clay loam
B_3	4.0-4.3	60.3	25.2	14.5	Sandy loam
C	4.3+	62.1	26.4	11.5	Sandy loam

SPINKS LOAMY FINE SAND

Horizon	Depth - Feet	% Sand	% Silt	% Clay	Texture
A_1	0.0-0.6	75.0	20.6	4.4	Loamy sand
A_2	0.6-1.7	76.1	19.4	4.5	Loamy sand
B_t	1.7-2.3	63.4	20.6	16.0	Sandy loam
Series of A_2- B_t	2.3-4.2	57.3	33.0	8.4	Sandy loam
C_{21}	4.2-4.5	60.3	28.1	11.6	Sandy loam
C_{22}	4.5-5.0	83.6	11.5	4.9	Loamy sand
C_{23}	5.0+	87.4	8.3	4.3	Loamy sand

associated with more xerophytic sites. The elm-ash-red maple-swamp white oak association with associated American basswood (Tilia american L.), shagbark hickory (Carya ovata (Mill.) K. Koch), and American sycamore (Platanus occidentalis L.) occur on the poorly drained mineral soils.

As part of a continuing ecological study, the "overstory composition", consisting of all woody stems of one-inch d.b.h. and larger, were measured and recorded by species and one-inch diameter class in the summers of 1940, 1950, and 1960 and again in the fall of 1965. Ecological conditions for the 1940 to 1960 period are reported by Schneider (1963, 1966).

The stand and stock table of the 13 major tree species in Toumey Woodlot is given in table 3 for the 1960 and 1965 inventories. The values given in the table are the mean values for the woodlot, expressed on a per acre basis.

Sugar maple is the most important and abundant species in the study area. It comprises approximately 60 percent or 81 square feet of the 137 square feet of basal area on an "average" acre in Toumey Woodlot. American beech, the second most important species, makes up approximately 20 percent of the total basal area. In addition, lesser amounts of slippery elm (Ulmus rubra Mühl.), American basswood, American elm (Ulmus americana L.), white ash (Fraxinus americana L.), northern red oak (Quercus rubra L.), and black cherry (Prunus serotina Ehrh.) occur within the 15 acre study area. The relative abundance of the different species in

Table 3. Stand and stock table of 13 major tree species in Toumey woodlot (1960 and 1965). Data for 1960 from Schneider (1963).

Note: Sawtimber sized species include: sugar maple, American beech, basswood, American elm, black cherry, white and red oak, bitternut hickory, slippery elm, white ash, red maple, black maple, and hophornbeam.

d.b.h.	1960 TREES / ACRE	1960 B.A. / ACRE	1965[1)] TREES / ACRE	1965[1)] B.A. / ACRE
1	274.69	1.65	283.40	1.51
2	109.14	2.40	122.81	2.68
3	45.12	2.21	47.69	2.51
4	22.35	1.94	24.08	2.10
5	14.47	1.97	14.58	1.99
6	11.06	2.17	10.25	2.01
7	7.63	2.04	8.74	2.34
8	6.02	2.10	5.93	2.02
9	5.35	2.36	5.34	2.36
10	3.79	2.07	2.58	1.41
11	5.10	3.37	3.90	2.53
12	4.59	3.60	4.29	3.37
13	4.23	3.90	4.51	4.16
14	4.37	4.67	3.85	4.12
15	4.43	5.44	3.76	4.61
16	5.50	7.68	4.06	5.67
17	4.59	7.23	4.65	7.33
18	3.78	6.68	4.14	7.32
19	3.47	6.83	3.54	6.97
20	3.48	7.59	3.34	7.29
21	3.18	7.65	2.74	6.59
22	2.74	7.23	3.18	8.39
23	1.99	5.74	1.99	5.74
24	1.47	4.62	2.22	6.97
25	1.78	6.07	1.24	4.23
26	1.18	4.35	1.41	5.20
27	1.03	4.10	0.87	3.46
28	1.32	5.65	1.03	4.40
29	0.66	3.03	0.73	3.35
30	0.14	0.69	0.66	3.24
31	0.29	1.52	0.14	0.73
32	0.44	2.46	0.52	2.90
33	0.29	1.72	0.29	1.72
34	0.21	1.32	0.14	0.88
35	0.07	0.47	0.29	1.94
36	0.07	0.49	0.14	0.99
37	0.14	1.05	--	--
38	--	--	0.14	1.10
39	--	--	--	--
40	0.07	0.61	--	--
41	--	--	0.07	0.64
45	0.07	0.77	0.07	0.77
TOTAL	560.30	137.44	583.31	137.54

1) Personal communication, Schneider, G., May, 1966.

terms of basal area and number of stems on the "average" acre are given in table 4. The common names and scientific names of the tree species are listed in Appendix I.

The ground-cover vegetation is not considered for purposes of the study of the soil-moisture regime and soil-water balance. The influence of the ground-cover vegetation is restricted primarily to the surface few inches of the soil and therefore, for purposes of this study, is assumed to have a uniform influence over the entire study area.

The overstory vegetation, 1-inch d.b.h. and larger, are summarized for each of the ten 60 foot X 60 foot soil-moisture study plots in table 5. All values are converted to a "per acre" basis. The range in basal area is from 80.46 to 223.55 square feet per acre. The mean basal area per acre is 149.43 square feet which is somewhat higher than the overall mean of 137.54 for the whole woodlot.

On the study plots, sugar maple and American beech make up approximately 95 percent of the total number of stems 1-inch d.b.h. and larger. Reproduction is variable, ranging from light to abundant. There is no significant correlation between basal area and the amount of reproduction.

The mean stem diameter is 4.8 inches. The minimum stem diameter is 3.36 inches on plot 9 (D-10) and the maximum is 7.43 inches on plot 10 (B-9). There is no significant correlation

Table 4. Mean number of stems, basal area, and percent of total basal area of the major tree species on an "average" acre in Toumey Woodlot (1965).

SPECIES	STEMS / ACRE	B.A. / ACRE	% of TOTAL B.A.
Sugar maple	403.96	81.42	59.2
American beech	44.61	27.64	20.1
Basswood	9.92	7.75	5.6
Black cherry	1.83	0.93	0.7
Slippery elm	34.05	3.55	2.5
American elm	7.50	4.34	3.2
Northern red oak	5.89	5.77	4.2
White oak	0.43	0.39	0.3
White ash	36.40	3.82	2.8
Bitternut hickory	2.13	0.30	0.2
Red maple	0.07	0.14	0.1
Black maple	1.18	0.90	0.7
Hophornbeam	27.62	0.35	0.3
Miscellaneous	11.60	0.24	0.1
TOTAL	575.59	137.54	100.0

Table 5. Basal area, mean plot diameter, and percent species composition on soil-moisture study plots. (1965)

Plot No.	Grid Location	B.A. per Acre	Mean Dia. - Inches	Percent of Basal Area Sugar Maple	Amer. Beech	Misc.	Reproduction
1	B-3	142.01	5.24	88.	6.	6.	Light
2	C-5	80.46	3.68	43.	51.	6.	Abundant
3	E-5	109.64	6.09	22.	78.	--	Moderate
4	G-6	176.70	6.10	86.	14.	--	Moderate
5	H-4	157.00	3.56	67.	33.	--	Abundant
6	J-5	108.22	4.00	93.	7.	--	Moderate
7	H-8	202.45	3.43	85.	12.	3.	Light
8	F-9	214.12	5.15	76.	7.	17.	Moderate
9	D-10	80.63	3.36	91.	2.	7.	Light
10	B-9	223.55	7.43	60.	30.	10.	Abundant
Mean		149.48	4.80	71.1	24.0	4.9	

between position in the woodlot and basal area or mean stem diameter.

The mean basal area for the ten sample plots is not significantly different at the 95 percent level of confidence from the mean basal area as determined by a 100 percent sample of Toumey Woodlot. Therefore, the soil-moisture plots may be considered to be representative of the vegetation in Toumey Woodlot. In addition, the large variation in basal area on the soil moisture study plots (Standard deviation 53.87 sq. ft.) indicates a high degree of variability within the stand itself. There are no easily visible major differences between plots; therefore, further analysis to determine differences does not appear to be valid.

CHAPTER V

METHODOLOGY

Study of the soil-moisture, climatic, soil, and vegetation data are divided into three phases. The first phase involves an empirical description of the soil-moisture regime during the period of study. The cyclic pattern of soil-moisture recharge and depletion is repeated each year. Recharge begins in the fall and continues through the winter into the growing season when a general depletion of soil moisture occurs. The relative magnitude of soil-moisture recharge and depletion may be expected to vary from year to year depending upon the existing moisture and climatic conditions. Therefore, the actual quantity of water added to the soil-moisture reservoir, as well as the relative efficiency of precipitation in its contribution to soil-moisture storage, must be evaluated in terms of the pre-existing conditions.

For purposes of description, the mean soil-moisture profile determined from the ten sampling points is used. The mean profile for the vegetation - soil type is of interest rather than changes occurring in the soil-moisture regime at an individual point. In addition, emphasis will be placed upon differences or changes in soil-moisture contents rather than in terms of absolute values. By basing calculations primarily upon differences rather than absolute values, any errors inherent in calibration of the neutron probe will be minimized. Also, absolute values are of only minor importance in this type of study where we are primarily interested in evaluation of consumptive moisture losses by the vegetation type concerned.

The first phase of the study will deal also with time-phenomena variations and variations in magnitude of soil-moisture changes occurring at different depths within the soil profile. These time-depth relationships will be of primary interest in the second phase of the study which is concerned with the analysis of soil-moisture accretion and depletion patterns. Analysis will concern the evaluation of redistribution of soil moisture within the soil profile, as well as the existence of deep seepage.

Because of limited instrumentation, examination of the soil-moisture accretion patterns will be based primarily upon a purely empirical approach. Early soil-water investigations were concerned primarily with the water storage capacities of soil following infiltration. Such work did not consider the redistribution of soil water following infiltration or the physical explanation of the phenomenon (Biswas, Nielson, and Biggar, 1966). The actual advance of a wetting front within the soil profile is of minor importance in terms of recharge to the ground-water table except during periods of maximum soil-moisture storage.

Nevertheless, because of the redistribution of soil moisture, recharge to the ground-water system may occur even though the entire soil profile is not at "field capacity".

"Field capacity" for the mean nine-foot soil profile is estimated from the soil-moisture contents measured on dates following periods of recharge. Using an arbitrary criteria of "field capacity" as the moisture contents of a soil that has been allowed to drain freely for at least 48 hours, the measured soil-moisture contents for each day meeting this criteria are plotted as a function of depth in the profile. The "field capacity" is considered to be the sum of the maximum measured moisture contents for each depth.

The second phase of the study will, therefore, be concerned with the quantitative evaluation of soil-moisture movement, redistribution, and drainage from the soil profile.

The final phase of the study will deal with the determination of the soil-water budget in terms of actual evapotranspiration and the quantity of ground-water recharge. In this phase of the study, fluctuations in soil-moisture storage will be attributed to recharge, evapotranspiration, and deep-seepage components.

Evapotranspiration losses will be calculated by one of two methods, depending on whether soil-moisture accretion or depletion has occurred. When depletion has occurred,

evapotranspiration will be considered to be equal to the difference between any two consecutive soil-moisture values plus precipitation during that period. If accretion has occurred, evapotranspiration will be considered to equal the difference between precipitation and the amount of soil-moisture recharge. These procedures of calculation of evapotranspiration, however, ignore possible recharge to the ground-water system or loss below the depth of measurement due to deep seepage. Therefore, the values of evapotranspiration as calculated above must be corrected for deep-seepage losses.

This method of calculation of evapotranspiration losses results also in a certain error during the snow accumulation periods as precipitation is stored at the soil surface and does not immediately contribute to soil-moisture storage. However, an under-estimate of evapotranspiration losses during the snowmelt period may balance out this error.

To estimate deep-seepage losses, a linear relationship between the rate of evapotranspiration and soil-moisture content is assumed until the assumed evapotranspiration rate is equal to the total soil-moisture depletion rate. When the assumed evapotranspiration rate is equal to or greater than the total moisture depletion rate, the evapotranspiration rate will be assumed to equal the total soil-moisture depletion rate. Lull and Fletcher (1962) use a similar linear relationship between the rate of depletion and soil-moisture content in a study of the comparative influence of hardwoods, litter, and bare soil on the soil-moisture regimen. In this manner, Lull and Fletcher (1962) obtained a significant correlation (at the 1 percent level of significance) between the average daily rate of soil drying and soil-moisture content.

Based upon results presented by Patric, Douglass, and Hewlett (1965), Richards, Gardner, and Ogata (1956) and Nixon and Lawless (1960), it is assumed that the rate of soil-water drainage decreases exponentially with time and with decrease in soil-moisture content. This assumption will agree with results presented by Remson, Randolph, and Barksdale (1960) which indicate a slow but continual downward drainage under a negative hydraulic head.

The difference between the total soil-moisture depletion curves and the linear evapotranspiration rate curve is considered to be the correction value for deep-seepage loss. This method is only an approximation of deep-seepage losses, dependent upon the validity of the assumptions. Much work is needed in this area of soil physics and soil-vegetation relations to fully understand the processes involved and to more accurately define the components of the soil-water budget.

Data are presented and analyzed by water year (October 1 through September 30 of the following calendar year) rather than by calendar year. The beginning of the water year, October 1, more closely corresponds to the date of occurrence of the minimum total soil-moisture content and the beginning of the annual period of soil-moisture recharge. For this reason, the use of the water year serves as a natural break for both analysis and descriptive purposes.

CHAPTER VI

RESULTS

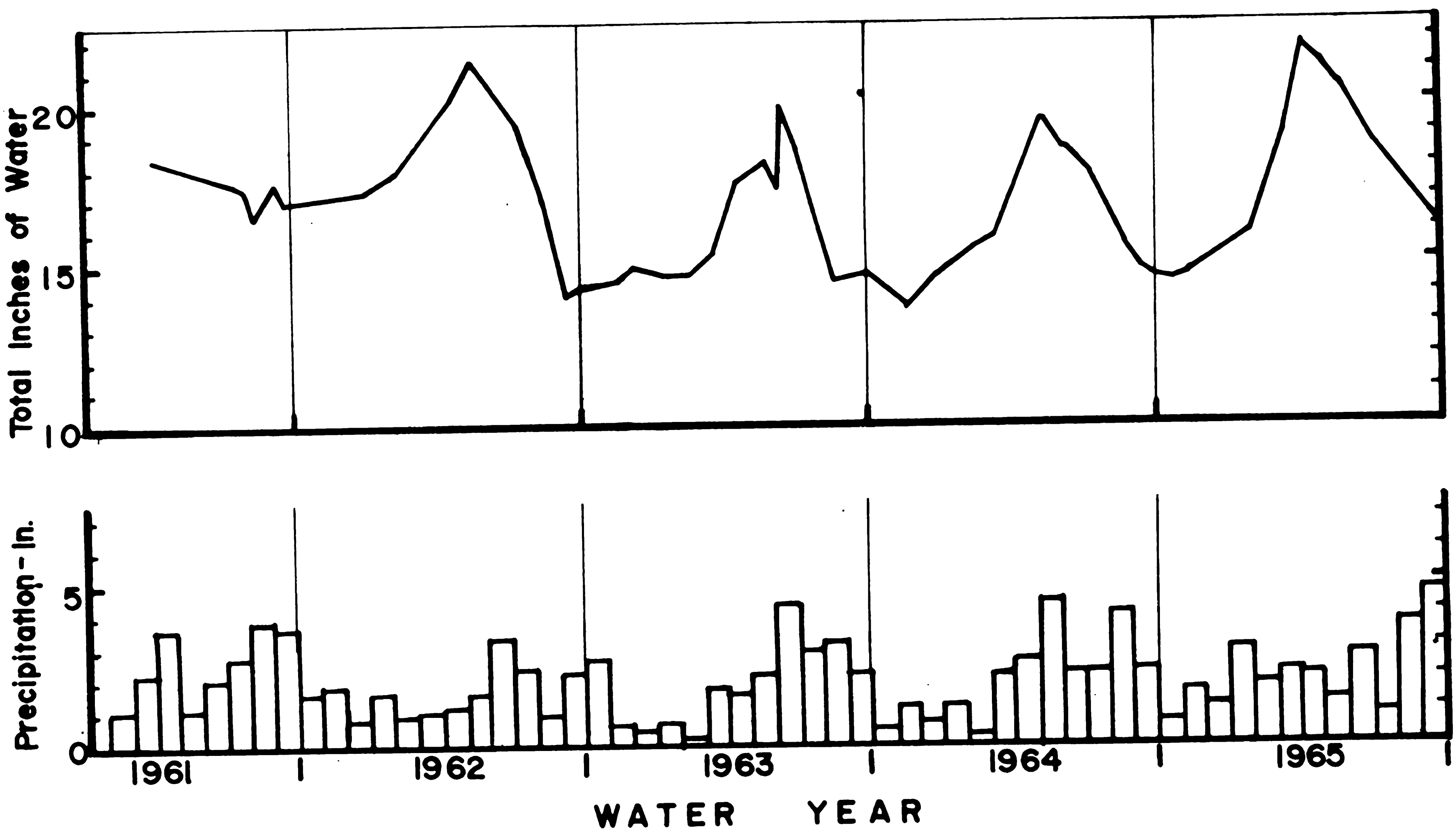
Results from the study of the soil-moisture regime under an old-growth hardwood stand for a $4\frac{1}{2}$-year period are presented and analyzed in two sections. The first is concerned with the annual cyclic patterns of soil-moisture accretion and depletion and their relationship to climatic variables. The second section is concerned with the analysis of soil-moisture accretion patterns following infiltration and soil-moisture depletion.

Soil moisture regime and its relation to climatic variables.

Examination of the composite soil-moisture regime for the April, 1961, through September, 1965, period (figure 3) shows an annual cyclic pattern of recharge and depletion. The profile represents the mean total soil-moisture content in a 9-foot profile for ten soil-moisture sampling points. Recharge begins in the fall of the year and continues through the spring snowmelt period when the maximum soil-moisture content is reached. With the resumption of vegetation growth in the spring of the year, soil moisture is continually depleted by evapotranspiration through the growing season into the fall when recharge again replenishes the soil-moisture reservoir.

Figure 3. Mean soil-moisture regime for a 9-foot profile under an old-growth hardwood stand (1961-1965 water years).

Note: The water year is defined as the period from October 1 through September 30 of the following calendar year.


Total Inches of Water
20
15
10
Precipitation - In.
5
0
1961
1962
1963
1964
1965
WATER YEAR

The annual cyclic pattern is similar from year to year but varies in detail dependent upon the distribution and amount of precipitation, the condition of the vegetation, and the moisture conditions of the soil. The mean annual precipitation for the 1961-65 water year period, recorded at the U.S. Weather Bureau Station on Michigan State University Horticulture Farm, one-half mile north of the study area, was 23.16 inches, or 77 percent of the 1941-60 mean annual precipitation. Seventy-five percent of all monthly precipitation values (table 6) were below the 1941-60 normal.

The accumulated precipitation deficit for the 1961-65 water years was 34.61 inches. A maximum accumulated deficit of 37.48 inches occurred during July, 1965. The mean annual deficit is 6.91 inches or 26.3 percent of the normal precipitation. The monthly precipitation deficits for the 1961-65 water years are presented in table 7.

Approximately 48 percent of the total precipitation deficit occurred during the November through April period. Precipitation deficits during this period have significant implications in terms of the quantity of water available as ground-water recharge. This is the period of lowest evapotranspiration losses. Theoretically, once the soil-moisture content is increased to "field capacity", the infiltration of any additional quantities of water into the profile will result in recharge to the ground-water system.

Table 6. Monthly precipitation: 1961-1965 water years (Horticulture Farm, Michigan State University, East Lansing).

	1941-60	1961	1962	1963	1964	1965
	Inches					
Oct.	2.59	1.07	1.68	2.69	0.42	0.65
Nov.	2.07	1.53	1.91	0.51	1.03	1.55
Dec.	1.64	0.11	0.88	0.40	0.63	1.11
Jan.	1.60	0.16	1.86	0.54	1.18	2.86
Feb.	1.37	1.14	0.90	0.20	0.33	1.70
Mar.	2.11	2.07	1.13	1.87	2.28	2.13
Apr.	2.71	3.55	1.23	1.65	2.53	2.05
May	3.93	1.04	1.82	2.30	4.49	1.34
Jun.	3.53	2.17	3.42	4.07	2.09	2.84
Jul.	2.97	2.71	2.41	2.91	2.23	0.79
Aug.	3.02	3.97	1.85	3.05	4.04	3.73
Sep.	2.52	3.69	2.22	2.11	2.19	4.78
TOTAL	30.06	23.21	21.31	22.30	23.44	25.53

Table 7. Monthly precipitation deviations from the 1941-1960 mean. 1961-1965 water years. (East Lansing, Michigan).

	1961	1962	1963	1964	1965	MEAN	% Months with Deficits
	----------------------	Inches	----------------------				
Oct.	-1.52	-0.91	+0.10	-2.17	-1.94	-1.29	80.
Nov.	-0.54	-0.16	-1.56	-1.04	-0.52	-0.76	100.
Dec.	-1.53	-0.76	-1.24	-1.01	-0.53	-1.01	100.
Jan.	-1.44	+0.26	-1.06	-0.42	+1.26	-0.28	60.
Feb.	-.023	-0.47	-1.17	-1.04	+0.33	-0.52	80.
Mar.	-0.04	-0.98	-0.24	+0.17	+0.02	-0.21	60.
Apr.	+0.84	-1.48	-1.06	-0.18	-0.66	-0.51	80.
May	-2.89	-2.11	-1.63	+0.56	-2.59	-1.73	80.
Jun.	-1.36	-0.11	+0.54	-1.44	-0.69	-0.61	80.
Jul.	-0.26	-0.56	-0.06	-0.74	-2.18	-0.76	100.
Aug.	+0.95	-1.17	+0.03	+1.02	+0.61	+0.29	20.
Sep.	+1.17	-0.30	-0.41	-0.33	+2.26	+0.48	60.
TOTAL	-6.85	-8.75	-7.76	-6.62	-4.63	-6.91	75.

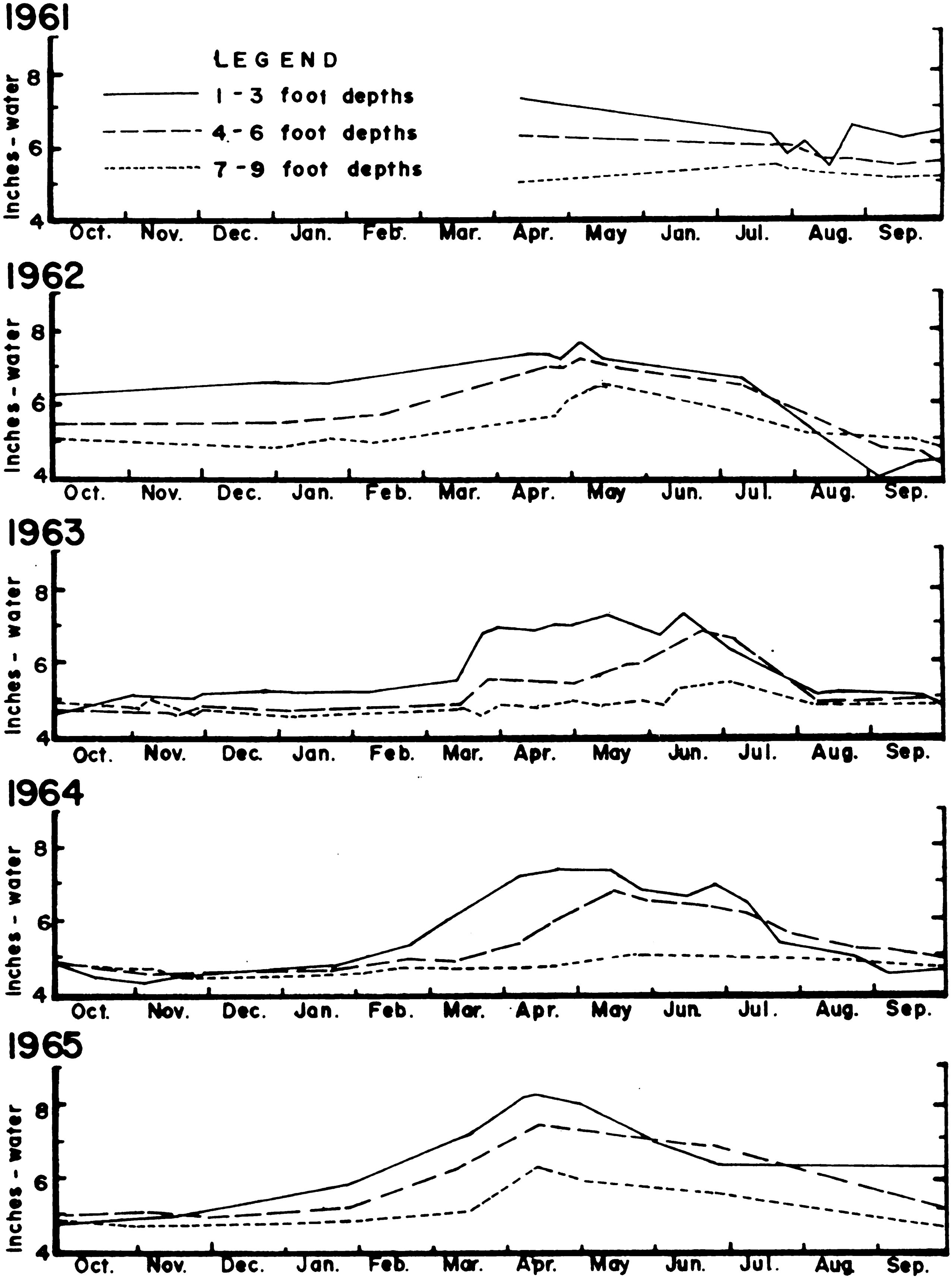
The deficiency of precipitation during the study period is reflected in record low or near-record low ground-water levels in South Central Michigan. Ground-water levels failed to recover during the winter and early spring recharge periods. According to McGuinness (1941), snowmelt normally contributes a major portion of the total ground-water recharge.

Approximately 45 percent of the mean precipitation deficit occurred during the three month - May, June, and July - period; whereas 52 percent of the deficit occurred during the four month period - April through July. This period corresponds, in part, with the period of greatest potential growth, and therefore, it has its obvious implications in terms of growth reduction.

Infiltration of summer precipitation seldom penetrates to depths greater than three feet. It is stored within the soil profile in order to satisfy the soil-moisture deficit normally present during the growing season and is subsequently utilized by evapotranspiration. Recharge to the ground-water system is dependent therefore, upon precipitation during the dormant season and snowmelt. Only minor quantities of water are contributed to ground water during the growing season.

Variations in the soil-moisture regime for the period of study for the 1-3, 4-6, and 7-9 foot depths, are shown in figure 4. The relative magnitude of variations in soil-moisture content decreases with increase in depth. Moisture conditions below three

Figure 4. Soil-moisture patterns for the 1-3, 4-6, and 7-9 foot depths for the 1961-1965 water years (Toumey Woodlot).


1961
1962
1963
1964
1965
LEGEND
1 - 3 foot depths
4 - 6 foot depths
7 - 9 foot depths
Inches - water
4
6
8
Oct.
Nov.
Dec.
Jan.
Feb.
Mar.
Apr.
May
Jan.
Jun.
Jul.
Aug.
Sep.

feet are relatively insensitive to addition of increments of precipitation at the surface, except during the spring recharge period. Moisture conditions at these depths do not show as sharp fluctuations in moisture conditions as shown by the surface layers.

Recharge in the surface layers of the soil profile begins in the fall of the year, coincident with the start of the new water year. Even though recharge is occurring in the surface layers, soil-moisture depletion continues at depth due to one or a combination of several phenomena occurring simultaneously within the profile. Continued evapotranspiration withdrawal by the forest vegetation and downward seepage of soil water may contribute to the continued soil-moisture depletion from the lower depths in the soil profile during the dormant season. In addition, thermal and moisture gradients within the soil profile may result in the redistribution of soil water (Hoekstra, 1966).

During the dormant season, the surface soil is subjected to freezing conditions and a reduction in temperature of the surface soil layers results. According to Geiger (1965), in frozen soil the water vapor pressure gradient in the unfrozen soil below frozen surface layers is always directed upward and is independent of the diurnal temperature variation at the surface. Both water rising under capillarity and water vapor migrating through the soil pores come up against a frozen barrier. Water vapor condenses at this interface, resulting in a gradual increase in moisture content. This phenomena together with a corresponding decrease in bulk density upon freezing has been shown by Krumbach and White (1964) and Ferguson, Brown, and Dickey (1964). Therefore, we may expect a gradual increase in soil-moisture content near the soil surface as well as soil-water depletion at depth by moisture migration due to thermal gradients.

Taylor and Cavazza (1954) noted the soil-water migration by vapor diffusion or by a combination of vapor diffusion and capillarity can be appreciable when thermal gradients are presented. However, "there can be little doubt that the most complex and least understood area in the field of soil-water relationships is that of the effect of temperature gradients applied to moist soils" (Winterkorn, 1958). Instrumentation in this study, however, was not such that thermal gradients or differences in potential could be detected so that direction of soil-moisture movement within the profile could be determined.

The dates of occurrence of the minimum soil-moisture contents (table 8) for each water year show a definite time lag with increasing depth in the profile. Similarly, the time of occurrence of the maximum moisture contents for each water year shows a 2-4, and 4-6 week delay in the 4-6 and 7-9 foot depths,

Table 8. Dates of occurrence of maximum and minimum moisture contents measured at Toumey Woodlot during the 1962-1965 water years.

Depth	1962 Max.	1962 Min.	1963 Max.	1963 Min.	1964 Max.	1964 Min.	1965 Max.	1965 Min.
1	Apr.14	Sep. 7	Mar.25	Oct. 1	Apr.23	Oct.18	Apr.14	Oct.27
2	May 3	Sep. 7	Apr.23	Nov. 3	Apr.23	Nov. 8	Apr.14	Oct.27
3	May 3	Sep. 7	Jun.14	Oct. 1	May 15	Oct.18	Apr.14	Oct.27
4	May 3	Sep. 7	Jun.21	Nov. 3	May 15	Oct.18	May 3	Nov.10
5	May 3	Sep. 7	Jun.21	Nov. 3	May 15	Oct. 5	Apr.14	Nov.10
6	May 3	Sep.28	Jun.21	Nov.10	May 27	Mar.12	Apr.14	Dec. 2
7	May 3	Sep.21	Jun.14	Mar.11	May 27	Mar.12	Apr.14	Dec. 2
8	May 12	Sep.21	Jun.21	Mar.11	May 15	Nov. 8	Apr.14	Dec. 2
9	May 17	Sep.30	Jul. 5	Mar.11	May 27	Nov.27	Apr.14	Feb. 1

respectively, over that for the 1-3 foot depths.

Soil-moisture depletion throughout the growing season occurred at all depths within the soil profile, but not at equal rates. The rate of soil-moisture depletion was highest in the surface layers which was also closely associated with the highest moisture contents. In general, the rate of soil-moisture depletion decreased with both decrease in soil-moisture content and with increase in depth in the profile. The rate of soil-moisture depletion is greatest during June and July and gradually decreases in August and September.

The rate of soil-moisture depletion is correlated with the mean monthly temperature, but independent of total monthly pan evaporation (U.S. Weather Bureau Class A Evaporation pan). The mean monthly air temperatures for East Lansing are given in table 9 for the 1961-65 water years.

Analysis of mean monthly temperature and pan evaporation data indicate a significant (at the 95 percent level of confidence) correlation as shown in figure 5. Further analysis indicates that pan evaporation increases in a curvilinear manner with increase in mean monthly temperature from April to July. The maximum pan evaporation (7.43 inches of water) during July coincides with the maximum mean temperature (70.8^{o}F) for the 1961-65 water years.

Pan evaporation decreases from the maximum occurring in July at a much greater rate than it increased to the maximum

Table 9. Mean monthly air temperature: 1961-1965 water years (East Lansing, Michigan).

	1961 °F	1962 °F	1963 °F	1964 °F	1965 °F	Mean 61-65 °F
Oct.	51.1	53.7	53.6	54.1	48.2	52.1
Nov.	42.4	39.6	38.4	43.9	43.3	41.5
Dec.	23.9	27.3	24.1	21.0	26.8	24.6
Jan.	21.4	19.1	14.6	27.5	23.5	21.2
Feb.	29.8	20.8	17.4	26.6	24.6	23.3
Mar.	38.1	32.6	35.9	35.4	23.2	40.7
Apr.	42.7	46.6	48.7	48.9	43.9	46.1
May	50.5	64.1	55.6	62.5	63.6	59.2
Jun.	66.3	67.6	68.4	69.0	66.3	67.5
Jul.	71.5	69.4	71.3	72.6	69.4	70.8
Aug.	69.2	69.5	66.2	67.3	67.7	67.9
Sep.	67.4	60.6	60.4	62.6	63.6	62.9
Mean	47.9	47.6	46.2	49.3	47.0	48.2

evaporation, although the mean monthly temperature does not decrease at as great a rate. This situation gives a hysteresis-type loop to the relationship as shown in figure 6. These differences may be attributed in part to differences in wind velocity.

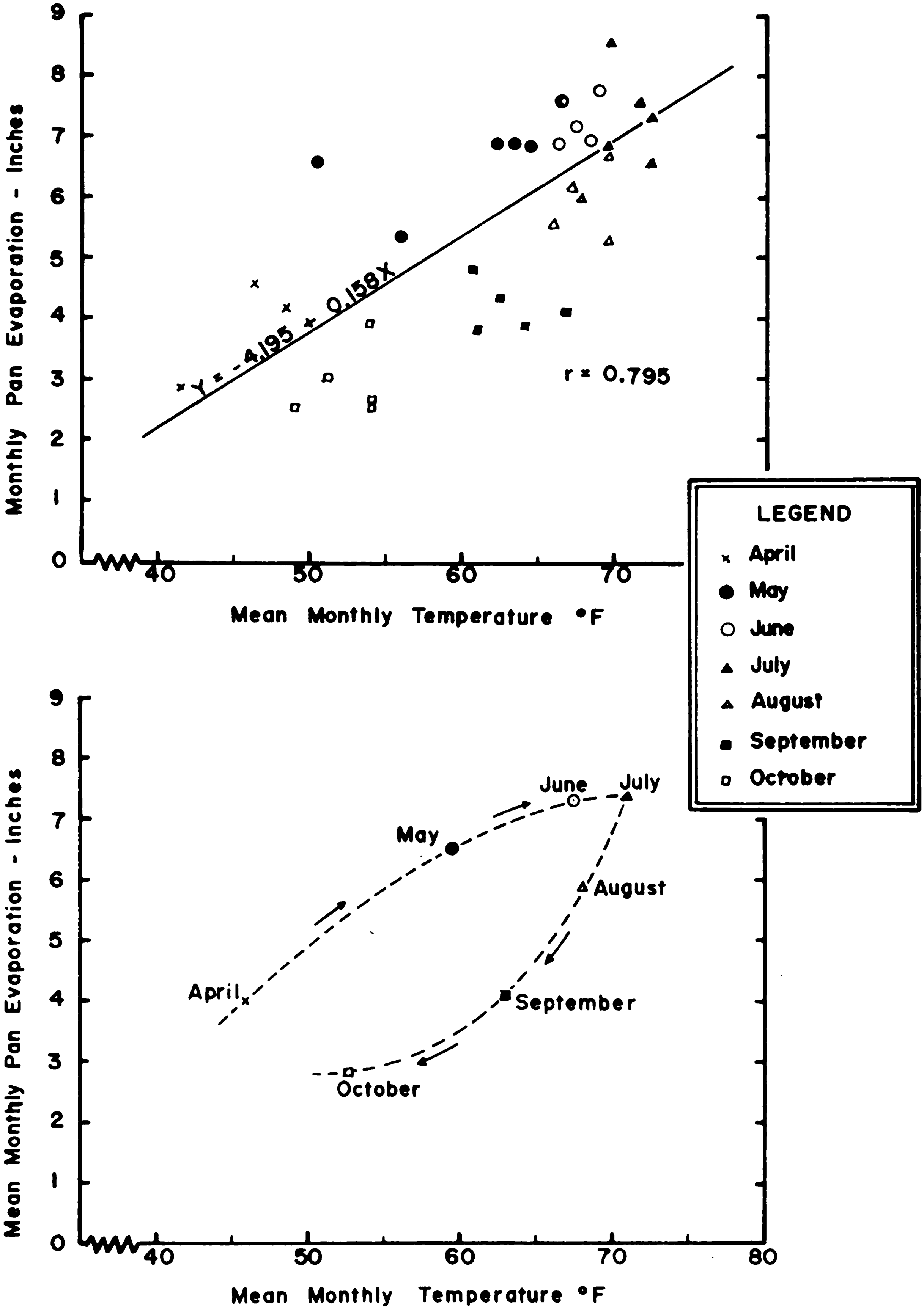
A serious weakness exists in the use of temperature as an expression to estimate evapotranspiration in that temperature is related to potential evapotranspiration and not necessarily to actual evapotranspiration. If an adequate supply of moisture were always available to satisfy the demands of potential evapotranspiration, the temperature variable would be closely related to actual evapotranspiration. This condition is met only during relatively short intervals following recharge of the soil-moisture storage capacity.

Soil-moisture accretion may occur in the lower depths of the profile simultaneously with depletion in the surface depths. Depletion in the surface horizons contribute to both evapotranspiration and to continual deep seepage, even though the moisture content is below "field capacity". Under these conditions, water drains from the upper soil layers and contributes to soil-moisture storage in the lower horizons or to ground water under a negative hydraulic gradient.

The soil-moisture gradients are normally directed downward - the greatest moisture contents normally occur at the surface

Figure 5. Linear relationship between the mean monthly temperature (°F) and monthly pan evaporation (Inches) - 1961-1965 water years. (East Lansing, Michigan).

Figure 6. Hysteresis-type relationship between mean monthly temperature (°F) and mean monthly pan evaporation (Inches) - 1961-1965 water years. (East Lansing, Michigan).


Monthly Pan Evaporation - Inches
Y = -4.195 + 0.158X
r = 0.795
Mean Monthly Temperature °F
LEGEND
April
May
June
July
August
September
October
Mean Monthly Pan Evaporation - Inches
June
July
May
August
April
September
October
Mean Monthly Temperature °F

and decrease with increasing depth. During the July-August period, soil-moisture gradients are directed both upward and downward from the 4-6 foot zone. Toward the end of the growing season, soil-moisture gradients within the nine-foot profile tend to be minimized. Similar moisture gradient patterns are shown (figure 4) during the 1962, 1963, 1964, and 1965 water years.

Movement of soil water may occur either from regions of higher to lower moisture contents or from regions of lower to higher moisture contents depending upon the soil-water energy conditions. Recharge to the ground water, for example, occurs against a moisture gradient under a potential energy gradient. Moisture gradients, therefore, are only one component of the total potential distribution in the soil-water-air system.

Analysis of variance of the soil-moisture contents on the date of occurrence of the maximum total soil-water contents (May 3, 1962), show highly significant differences between moisture contents at different depths within the soil profile and between permanent soil-moisture sampling plots. The moisture contents in the 1, 2, 3, 4, and 5-foot depths are all significantly greater (at the 95 percent level of confidence) than those in the 6, 7, 8, and 9-foot depths. However, the moisture contents within the 1-5 foot depths and the 6-9 foot depths at any given depth are not significant from the moisture content at any other depth within the same group.

Significant differences between soil-moisture plots may be expected due to differences in soil, vegetation, and topography. There are no significant correlations between maximum moisture contents and vegetation density, basal area, species composition, or soil texture.

Analysis of variance of the minimum observed total soil-moisture contents occurring on October 18, 1963, show there are significant differences between soil-moisture contents with depth, but no significant differences in soil-moisture contents between plots.

Correlation coefficients determined for the relationships between the maximum soil-moisture contents, minimum soil-moisture contents, and available water and basal area on the study plots are all non-significant at the 95 percent confidence level.

Pattern of soil-moisture accretion and depletion.

Development of soil-moisture theory has been based upon the application of Darcy's equation, derived originally for flow in a saturated medium, to unsaturated flow (Hewlett, 1961). In unsaturated flow, the larger pores are filled with air and therefore, soil-water movement is closely dependent upon the properties of air-water interfaces. Under unsaturated conditions, soil water exists as part of a multiphase system and can move within the belt of soil water and within the intermediate zone as liquid

water, as water vapor, and as adsorbed water. Moisture transfer in the adsorbed phase is generally negligible except under special conditions such as in a very dry soil with a high specific surface area.

Much of the hydraulic conductivity of soil is through pores of silt size material (0.05-0.002 mm) or larger. As a result, at the "critical moisture content" when pores of silt size and larger are dewatered, the rate of downward drainage becomes small.

Water applied to the soil surface moves downward behind a "wetting front" - the moisture content of which is between some critical level and saturation. For most soils, this critical moisture level corresponds to "field capacity" which is that moisture level at which the capillary conductivity becomes small. As the wetting front advances, it leaves the overlying soil at or near the critical moisture content.

Investigations by Bodman and Colman (1944) have shown that the downward movement of water advances behind a "wetting front" of sharp differential moisture contents. Water applied to the soil surface distributes itself uniformly throughout the soil only if the quantity applied is sufficient to bring the entire soil to the critical moisture level. Once the applied water wets a certain depth to this critical moisture level, practically all movement ceases or becomes very slow (Remson and Randolph, 1962). More water must then be applied to obtain further rapid penetration of the wetting front (Veihmeyer, 1939).

The wetting-front phenomena results because a certain degree of saturation is necessary before any given distribution of pore sizes can begin to transmit water rapidly. At lower moisture contents, water movement through these pores depend upon vapor diffusion which is a relatively slow process. Therefore, rapid transmission of soil water can occur only at moisture contents where the larger pores are sufficiently saturated for water to move through them as a liquid. Below these moisture contents, water movement in the vapor form through the larger pores and as a liquid through the smaller pores is very slow and the wetting front remains practically stable (Remson and Randolph, 1962).

Following cessation of infiltration, redistribution of soil water occurs in response to differences in potential within the profile. Continual movement of soil water tends to equalize the moisture contents within the profile.

Studies by Remson, Randolph, and Barksdale (1960) of the intermediate zone at Seabrook, New Jersey, suggest that downward gradients of hydraulic head produce slow but continuous rates of drainage even during the season of soil-moisture depletion. Hewlett (1961) concluded that there is some evidence

that slow drainage will result in gradients of soil-moisture stress on natural slopes and that persistence of negative hydraulic gradients can result in a steady accretion to ground water and streamflow from unsaturated soil.

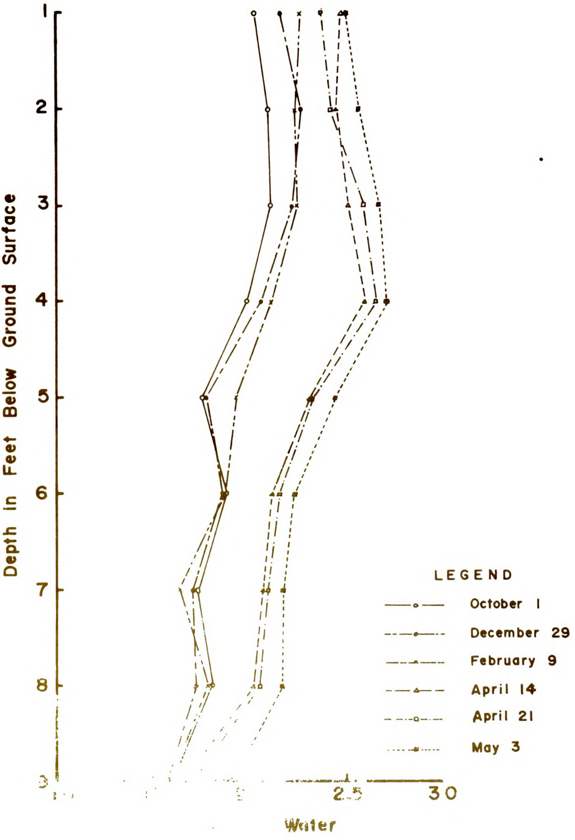
In addition to redistribution of soil-moisture within the profile, evapotranspiration draft upon the soil-moisture supply tends to simultaneously reduce the soil-moisture contents. The roots of vegetation withdraw water held within the soil matrix and release it into the atmosphere via the stems and foliage. The rate at which soil moisture is depleted is dependent upon the soil-moisture content, the depth under consideration, and the stage of vegetation development. Our present level of knowledge is inadequate, especially under forested conditions, to formulate a theoretical basis for separating soil-moisture changes into moisture redistribution and evapotranspiration loss components.

Furthermore, the neutron-scattering technique of soil-moisture measurement is not sensitive enough to distinguish fine differentials in soil-moisture contents occurring within a short depth interval. More refined methods, such as gamma attenuation, are needed to study the advance of the wetting front and subsequent redistribution of soil water within the profile. The neutron-scattering technique is, however, highly useful in determination of gross moisture contents within a spherical volume of approximately 12-inch minimum diameter. The general patterns of soil-moisture accretion and depletion can be studied by the use of this technique of soil-moisture measurement. For most hydrologic studies, the neutron method gives sufficient refinement of measurement to meet the study requirements.

Examination of the soil-moisture data for Toumey Woodlot for periods of recharge provide some indication as to the nature of the accretion patterns under upland hardwood forest vegetation. The accretion patterns, in general, represent the distribution of soil water within the profile on selected sampling dates during the October through April recharge period. The moisture patterns, however, do not show the existence of a definite wetting front but represent the moisture conditions within the soil profile after redistribution of soil water has, at least in part, occurred.

The soil-moisture accretion patterns for the 1962, 63, 64 and 65 water years are shown in figures 7 through 10. The figures represent the mean soil-moisture contents for ten sample plots on the dates indicated. The moisture distribution does not tell us about the actual advance of the wetting front or the physical processes involved. Nevertheless, it does reveal the soil-moisture distribution within the profile which is of major importance to the overall moisture availability for vegetation requirements and for ground-water recharge.

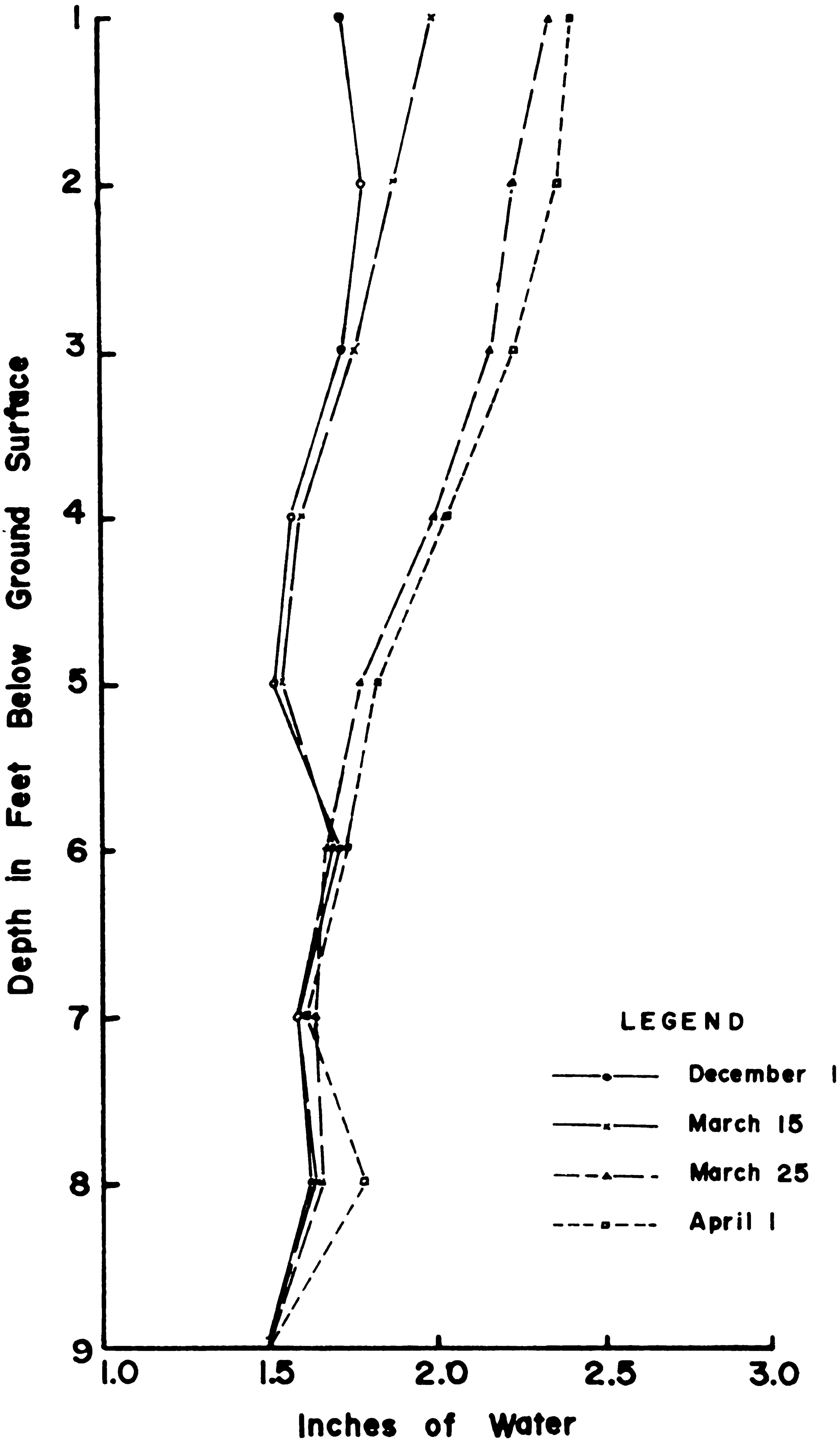
Figure 7. Soil-moisture accretion patterns for the 1962 water year. (Toumey Woodlot).



Depth in Feet Below Ground Surface
1
2
3
4
5
6
7
8
25
30
Water
LEGEND
October 1
December 29
February 9
April 14
April 21
May 3

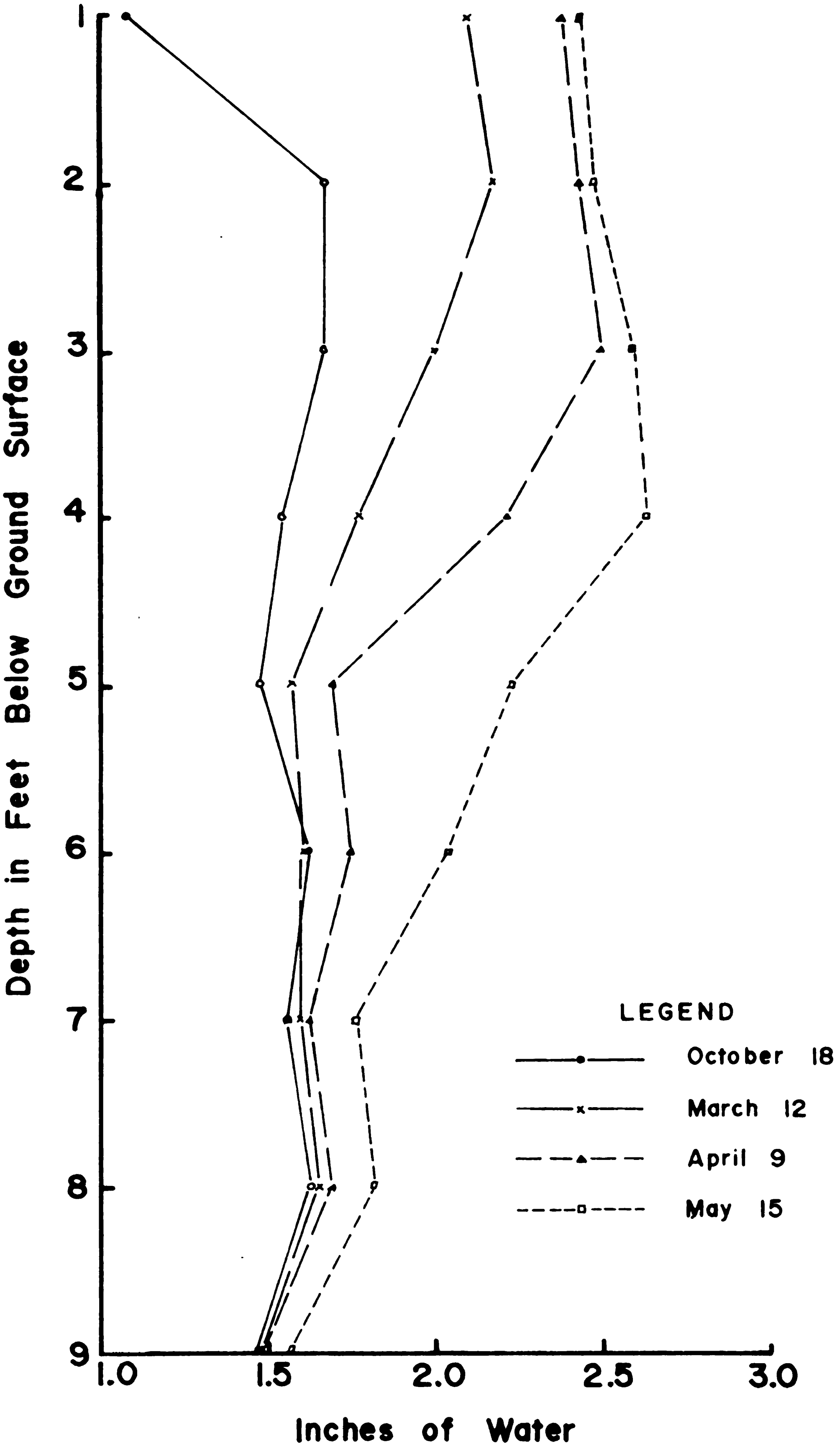
67

Figure 8. Soil-moisture accretion patterns for the 1963 water year. (Toumey Woodlot)



Depth in Feet Below Ground Surface
1
2
3
4
5
6
7
8
9
1.0
1.5
2.0
2.5
3.0
Inches of Water
LEGEND
December 1
March 15
March 25
April 1

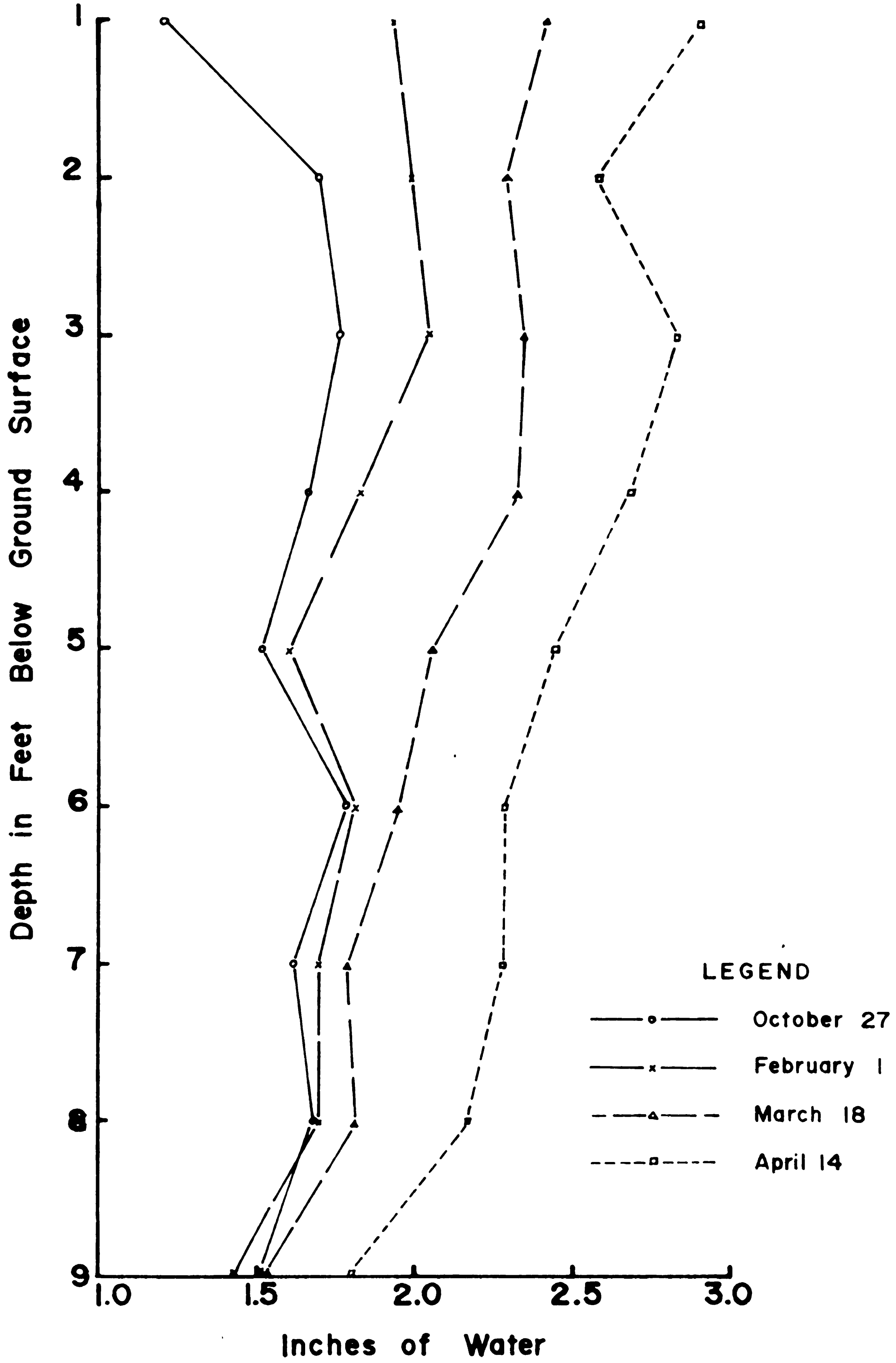
Figure 9. Soil-moisture accretion patterns for the 1964 water year. (Toumey Woodlot)



Depth in Feet Below Ground Surface
1
2
3
4
5
6
7
8
9
1.0
1.5
2.0
2.5
3.0
Inches of Water
LEGEND
October 18
March 12
April 9
May 15

6

Figure 10. Soil-moisture accretion patterns for the 1965 water year. (Toumey Woodlot)



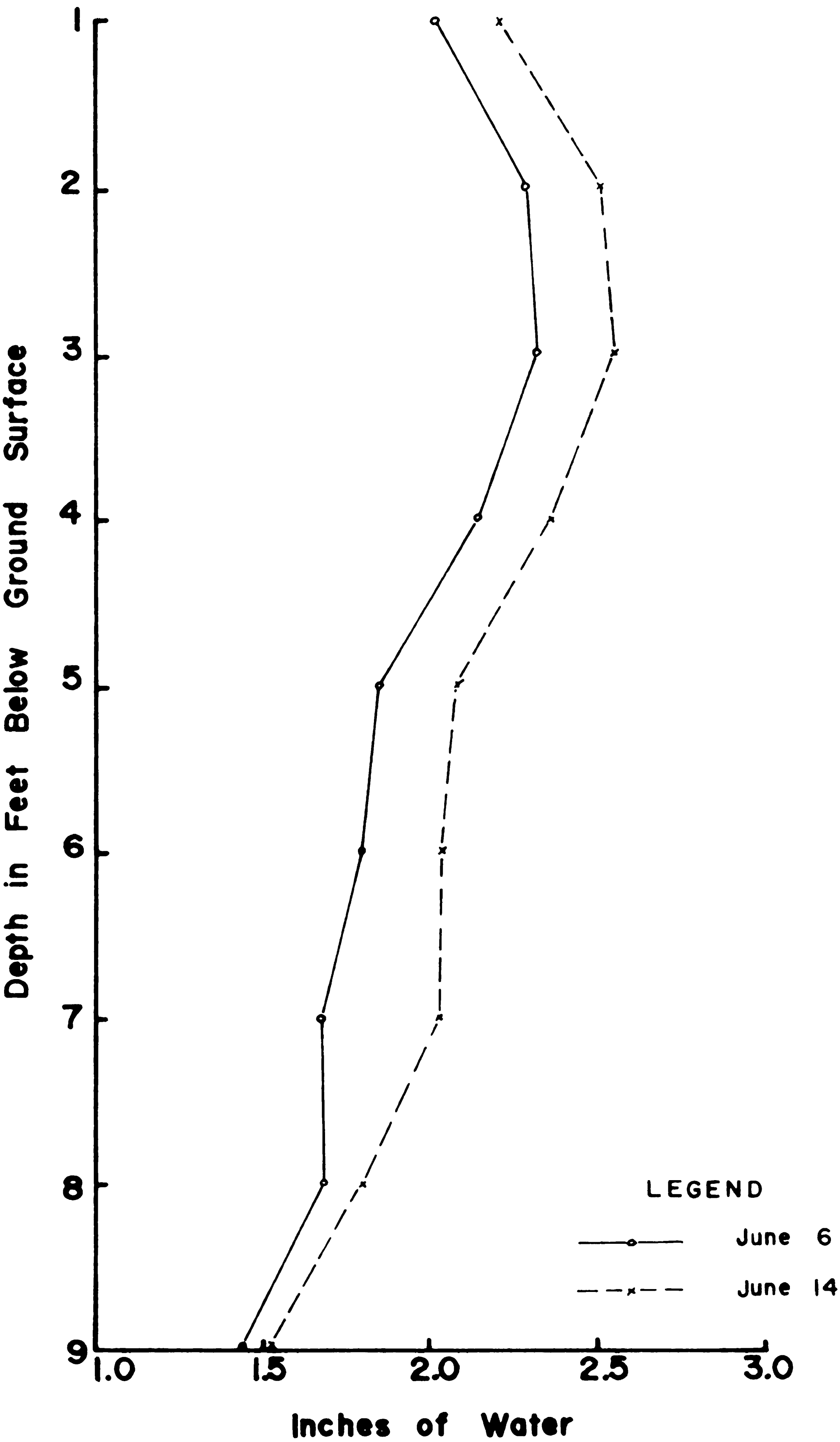
Depth in Feet Below Ground Surface
1
2
3
4
5
6
7
8
9
1.0
1.5
2.0
2.5
3.0
Inches of Water
LEGEND
October 27
February 1
March 18
April 14

Accretion to the soil-moisture reservoir appears to occur fairly uniformly throughout the upper 5 feet of the profile. The quantity of recharge decreases with increase in depth below the 5-foot level. This general pattern of accretion is fairly constant and without regard to the initial soil-moisture content or to "field capacity". This distribution pattern of soil-moisture recharge is due to redistribution of soil moisture following cessation of infiltration. Moisture accretion may occur at depth more rapidly than would normally be expected because of numerous root channels within the profile. Former root channels provide a porous and highly permeable passageway for the downward movement of water. This point has received little attention in the forestry literature. The existence of root channels may be of major significance in the soil-moisture regime in areas of native forest vegetation. The presence of root channels is of primary significance in its effect on the infiltration capacity of the surface soil.

A similar soil-moisture accretion pattern is shown in figure 11 for a period (June 6-14, 1963) during which 4.07 inches of precipitation was recorded. Recharge occurred throughout the 9-foot profile even though the soil-moisture content in the surface horizons is below "field capacity".

Under these conditions it may be assumed that some recharge below the depth of rooting or to the ground-water system

Figure 11. Soil-moisture distribution in the soil profile before and after a two week period during which 4.07 inches of precipitation was recorded. (Toumey Woodlot)


Depth in Feet Below Ground Surface
1
2
3
4
5
6
7
8
9
1.0
1.5
2.0
2.5
3.0
Inches of Water
LEGEND
June 6
June 14

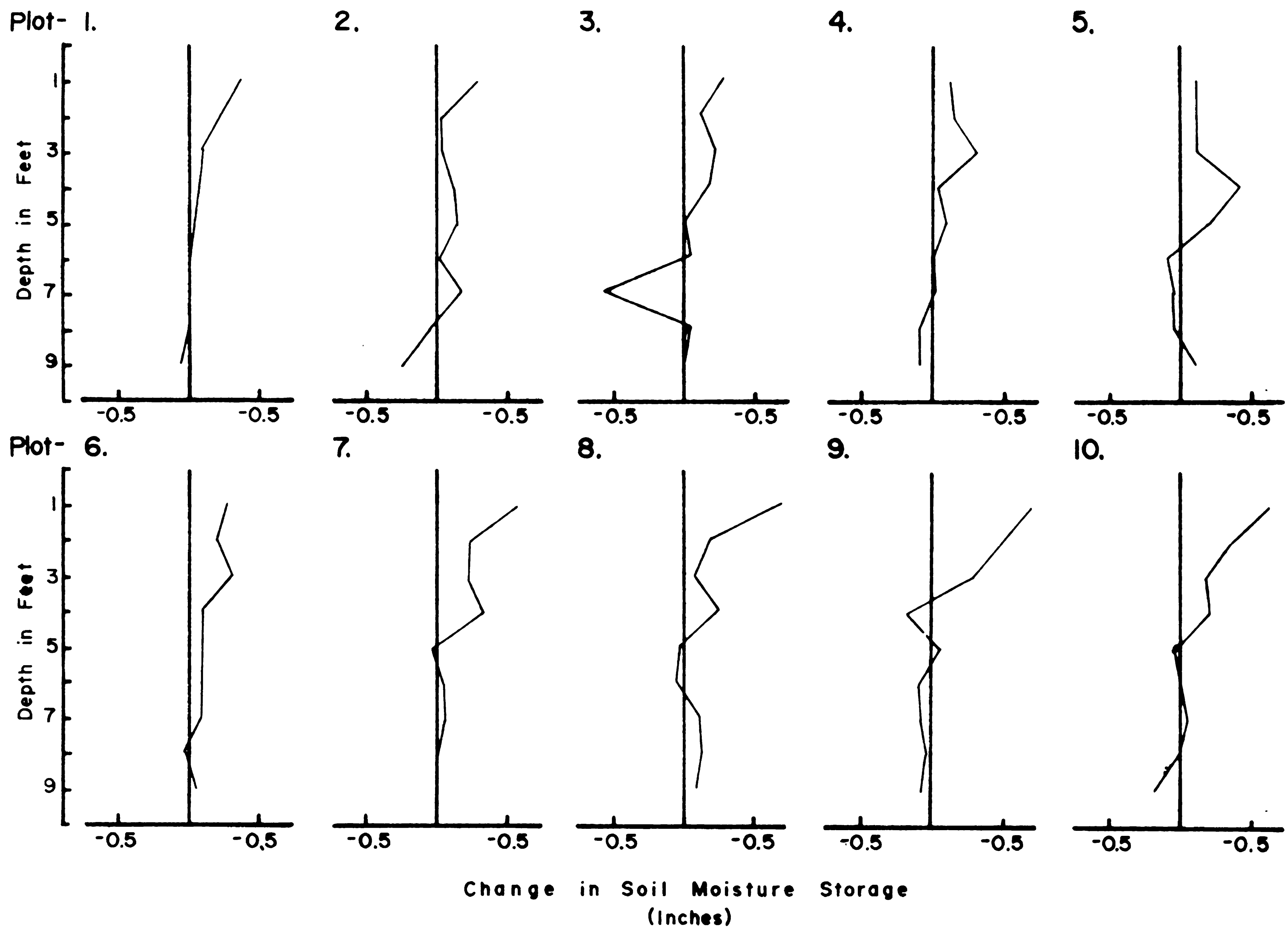
72"

does occur. Nevertheless, our knowledge of unsaturated flow and the nature of the soil matrix is limiting and therefore, separation of the total losses into evapotranspiration and deep-seepage components is difficult.

Soil-moisture depletion, or a net reduction in the total soil-water content, does not begin simultaneously at all depths within the profile, or at the same depth at different locations within Toumey Woodlot. But soil-moisture accretion and depletion may occur concurrently at different depths within the soil profile or at the same depth at different locations. This high degree of variability is illustrated in figures 12 and 13 for periods of general depletion and accretion, respectively.

Figure 12 illustrates the changes in total soil-moisture contents by one foot depth intervals for the June 21 through July 5, 1963 period. During this period, no precipitation was recorded; the gross precipitation during the June 6-12 period totaled 4.07 inches. Examination of the soil-moisture changes at various depths in the profile indicate that equilibrium conditions, with regard to soil-water redistribution, have not been reached. Soil-water movement from both the surface litter and underlying mineral matrix and from within the profile has occurred during the interval of measurement, even though no additional precipitation occurred during the period.

Figure 12. Pattern of soil-moisture changes for ten soil moisture sampling points for the June 21 through July 5, 1963 period. (Toumey Woodlot)



Plot- 1.
2.
3.
4.
5.
Plot- 6.
7.
8.
9.
10.
Depth in Feet
1
3
5
7
9
-0.5
-0.5
Change in Soil Moisture Storage
(Inches)

[illegible]

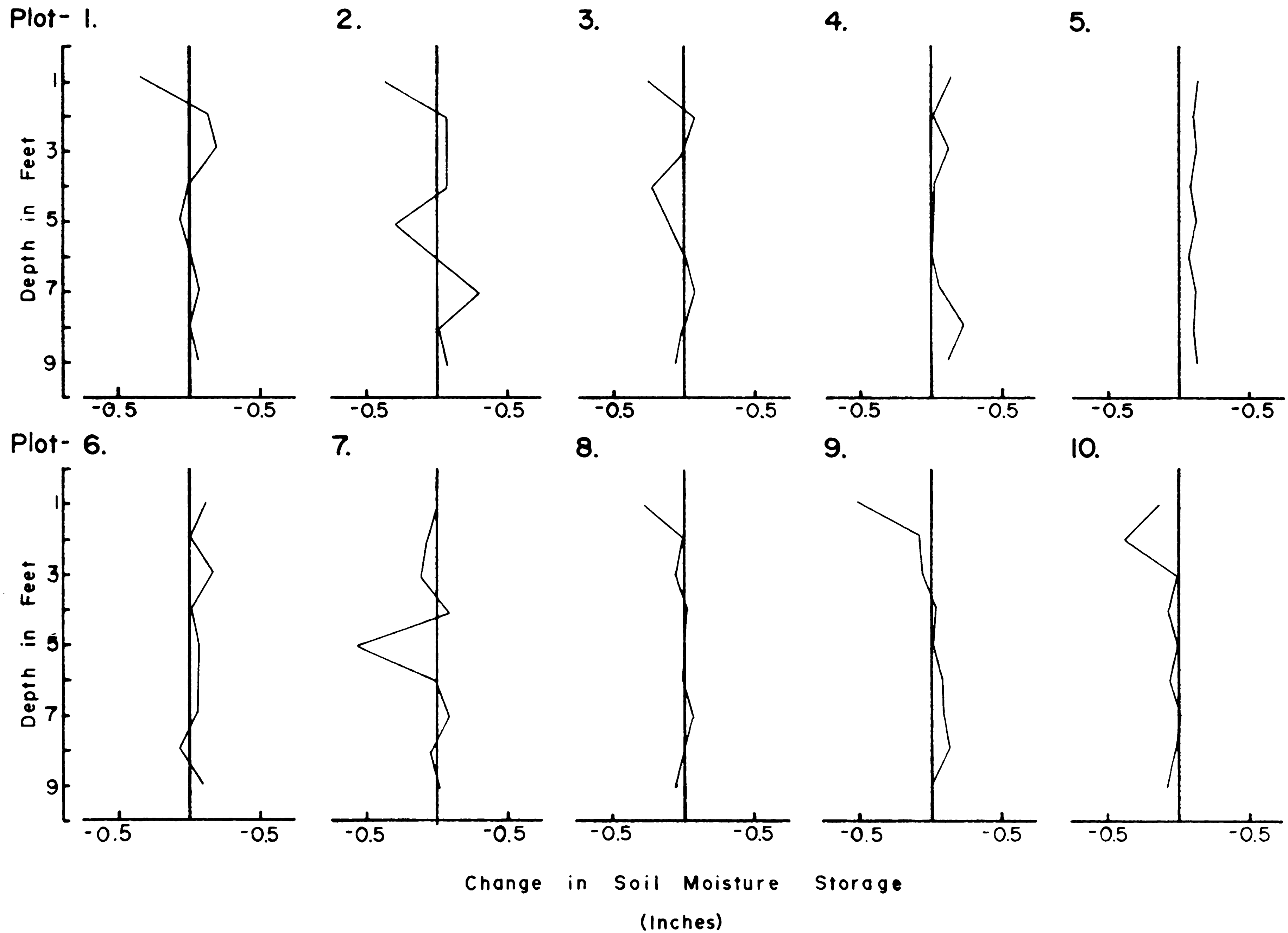
In general, figure 12 indicates that depletion of soil moisture has occurred in the upper 4 to 5 feet of the profile from both evapotranspiration draft and from the downward redistribution of soil water. Accretion has normally occurred at the lower depths in the profile. However, the quantity of soil water that has moved downward below the depth of sampling is unknown and must be estimated.

Similarly, the pattern of soil-moisture changes for the January 30 through February 9, 1962 period (figure 13) illustrates the high degree of variability between plots. There are no general trends common to all plots. The mean values for the ten soil-moisture sampling plots show a general increase in soil-moisture content in the surface layers and in the 4 to 5-foot depth. Depletion occurred in the 3-foot depth and in all depths below the 6-foot level.

Soil-moisture depletion normally occurs throughout the profile during the growing season, but not at equal rates at all depths. The differential rates of soil-moisture depletion within the profile will be discussed more fully later in the section.

Soil-moisture depletion curves can be used collectively to determine the relative rates with which moisture is withdrawn from the soil profile through moisture uptake by plant roots and lost below the depth of rooting by downward drainage. It is virtually

Figure 13. Pattern of soil-moisture changes for ten soil-moisture sampling points for the January 30 through February 9, 1962 period. (Toumey Woodlot)


Plot- 1.
2.
3.
4.
5.
Plot- 6.
7.
8.
9.
10.
Depth in Feet
1
3
5
7
9
-0.5
-0.5
Change in Soil Moisture Storage
(Inches)

impossible to separate evapotranspiration from drainage losses by soil-moisture measurements without the interruption of recharge to the ground-water system. For this reason, experimenters often neglect to account for possible loss of soil water by drainage through and from wet profiles (Patric, Douglass, and Hewlett, 1965).

The failure to separate evapotranspiration losses from deep seepage is not as serious a problem in studies on the relationships between available water and plant growth, as long as the occurrence of deep seepage is recognized. However, in hydrologic studies drainage from the soil profile may be a major portion of the "total water yield" from an area. It has been shown that drainage from the soil profile is an important factor in maintaining stream flow during dry periods in some areas of the country (Hewlett, 1961). In addition, the prediction of the results of vegetation manipulation for increased water yields must consider the possibility of increasing the quantity of soil-water drainage and consequently, increasing the total water yield.

The concept of field capacity as a constant as sometimes applied to agricultural soils is not a workable term in hydrology due to continued downward movement of soil water, even though the soil-moisture content is below this "value". The rate of soil-water drainage from the soil decreases exponentially with time from the maximum soil-moisture contents (Patric, Douglass, and

Hewlett, 1965; Richards, Gardner, and Ogata, 1956; Nixon and Lawless, 1960). Willardson and Pope (1963) observed a similar exponentially decreasing rate of soil-water drainage.

Therefore, deep seepage of soil water is of major importance when the soil-moisture content is at or near maximum; it would then decrease in importance as the soil-moisture content decreases. Gaiser (1952) found that the rate of soil-moisture depletion was such that all horizons of the profile simultaneously approach the wilting point. This general concept is supported by Hoover (1962).

Patric, Douglass, and Hewlett (1965) compared soil water absorption by forest vegetation from the top 20 feet of Southern Appalachian and South Carolina Piedmont soils. They found that early absorption from uniformly moist soil is related primarily to root concentration. However, latter absorption from drier soil tended toward uniform extraction even beyond 20 feet in depth. Water loss from the lower zone is both by local absorption by roots and by upward movement through the soil in response to potential gradients.

Absorption of soil water is related to the distribution of roots and available moisture in the profile. Absorption is increased while water is readily available until the potential evapotranspiration rate is reached.

Soil-moisture depletion curves were constructed for each 1-foot depth in the soil profile and for periods of depletion in time. From these it is evident that there is a definite decrease in depletion rate with increase in depth in the profile (e.g. the slope of the depletion curve decreases with increase in depth below the ground surface). This may be explained in part by several phenomena, which are: 1). continued drainage of soil water within the profile, 2). decrease in root concentration and soil-water uptake efficiency with increase in depth, and 3). the redistribution of soil water within the profile due to differences in potential energy. Lull and Fletcher (1962) similarly observed that the rate of soil-moisture loss tended to decrease with increasing profile depth.

The highest rates of soil-moisture drying or depletion are associated with the near surface depths. In these surface layers, evaporation plays an important role in soil drying as well as transpiration. Evaporation, by itself, decreases in importance with increase in depth in the profile. Transpiration, on the other hand, is able to utilize water from all depths in the profile by means of moisture uptake by the roots of vegetation. However, there is considerable question as to whether soil water at all depths within the profile is equally available for plant use.

The rates of soil-moisture depletion per foot depth during the growing season varied from 0.001 inches per day per foot to 0.030 inches per day per foot (figure 14). The lower rates of depletion are associated with the lower soil-moisture contents.

The total daily rate of soil-moisture for the 1962-65 water years was 0.075 inches of water during the June, July, and August growing period. This period coincides with the period of maximum soil-moisture depletion. The mean daily pan evaporation measured at the U.S. Weather Bureau Station at East Lansing, Michigan, was 0.219 inches of water or 292 percent of the daily evapotranspiration rate from the study area.

A composite depletion curve for the total profile is shown in figure 15. The depletion curve is based upon soil-moisture depletion rates for the 1962-65 water years. Deviations from the curve as shown are attributed to precipitation during the latter part of the depletion period. Nevertheless, there is a surprisingly close agreement between the slope of the depletion curve for the four water years in question. The depletion curves for each water year were shifted to pass through a common point at which the total water content was equal to 19 inches. Time in days at this common point was arbitrarily considered to be 0.

Composite depletion curves for the 1-foot depth, the 2-3, 4-5, 6-7, and 8-9 foot depths are shown in figures 16 and 17 a, b, c, d. It is assumed that the rate of depletion is linear with time

Figure 14. Semilogarithmic plot of the growing season soil-moisture depletion rate as a function of soil-moisture content for the 1962-1965 water years. (Toumey Woodlot)

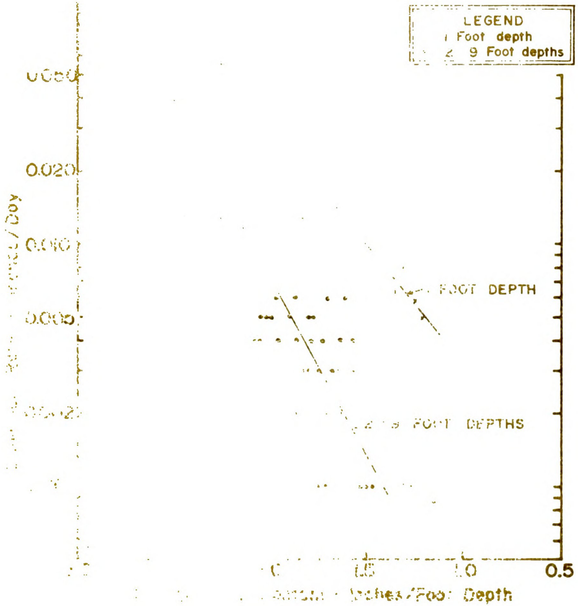

LEGEND
Foot depth
9 Foot depths
FOOT DEPTH
FOOT DEPTHS
0.020
0.005
0.5
Depth

Figure 15. Composite soil-moisture depletion curve for the nine-foot profile for the 1962-1965 growing seasons. (Toumey Woodlot)

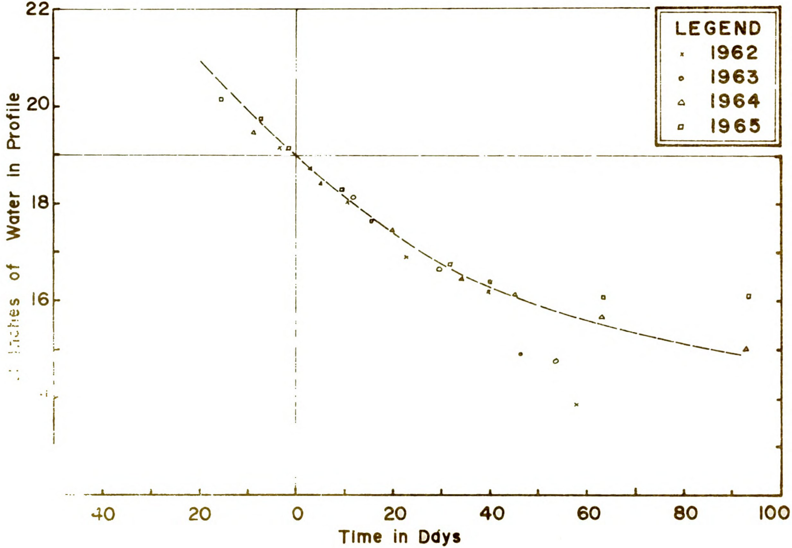
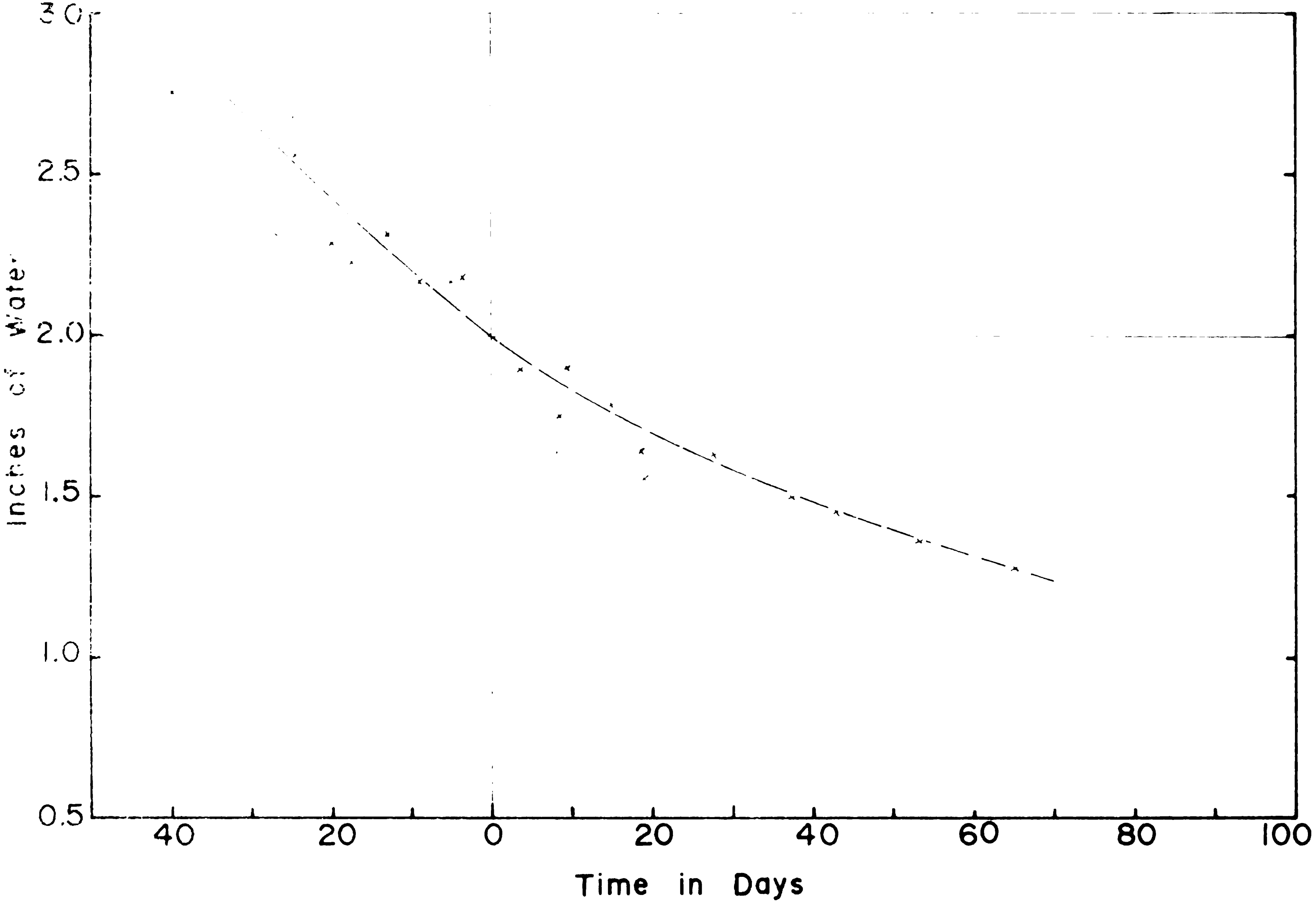

LEGEND
1962
1963
1964
1965
22
20
18
16
Inches of Water in Profile
40
20
0
20
40
60
80
100
Time in Ddys

Figure 16. Composite soil-moisture depletion curve for the 1-foot depth for the 1962-1965 growing seasons. (Toumey Woodlot)

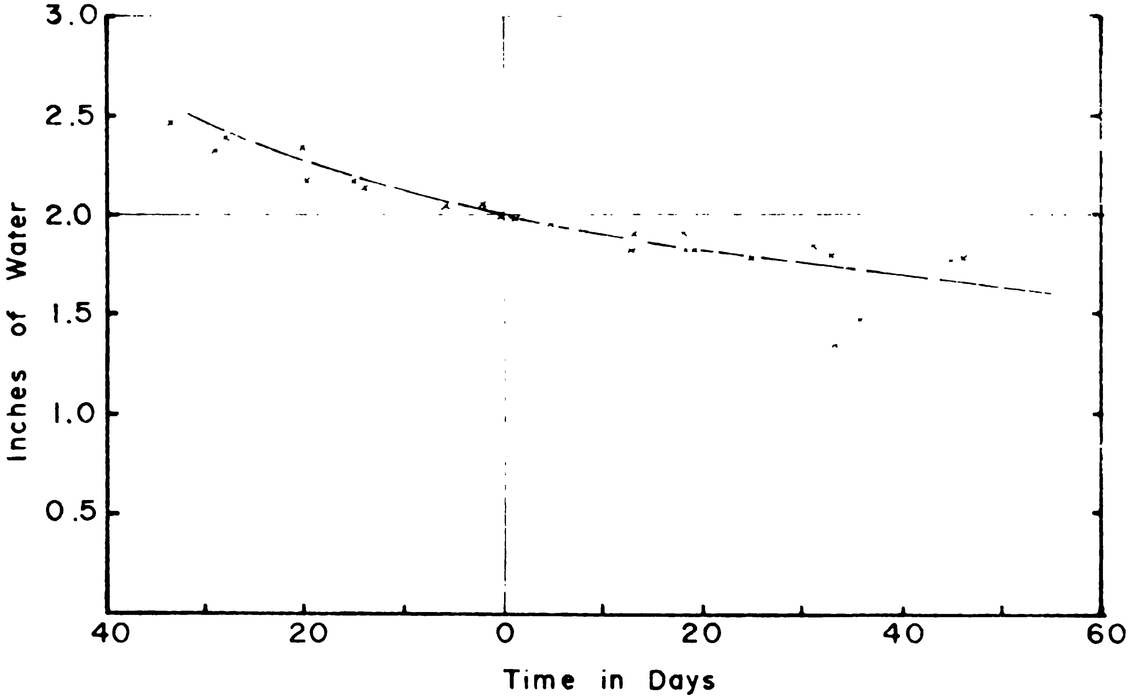


3.0
2.5
2.0
1.5
1.0
0.5
Inches of Water
40
20
0
20
40
60
80
100
Time in Days

Figure 17a. Soil-moisture depletion curve for the 2 and 3-foot depths for the 1962-1965 growing seasons. (Toumey Woodlot)

Figure 17b. Soil-moisture depletion curve for the 4 and 5-foot depths for the 1962-1965 growing seasons. (Toumey Woodlot)

2 & 3 FOOT DEPTHS



4 & 5 FOOT DEPTHS

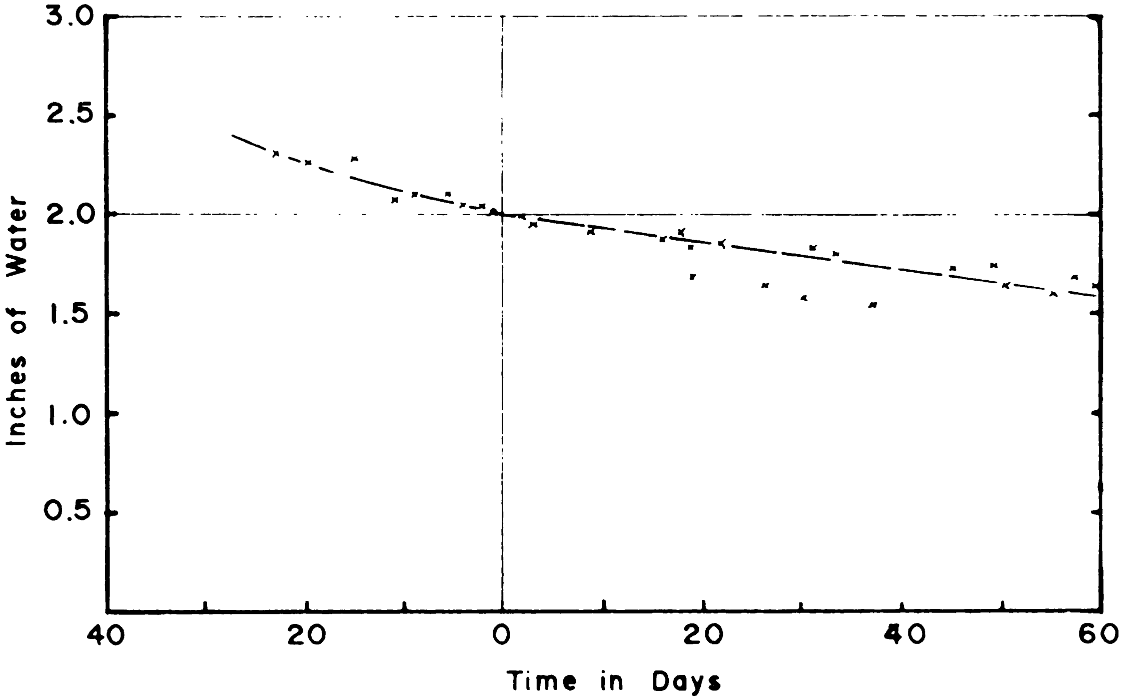
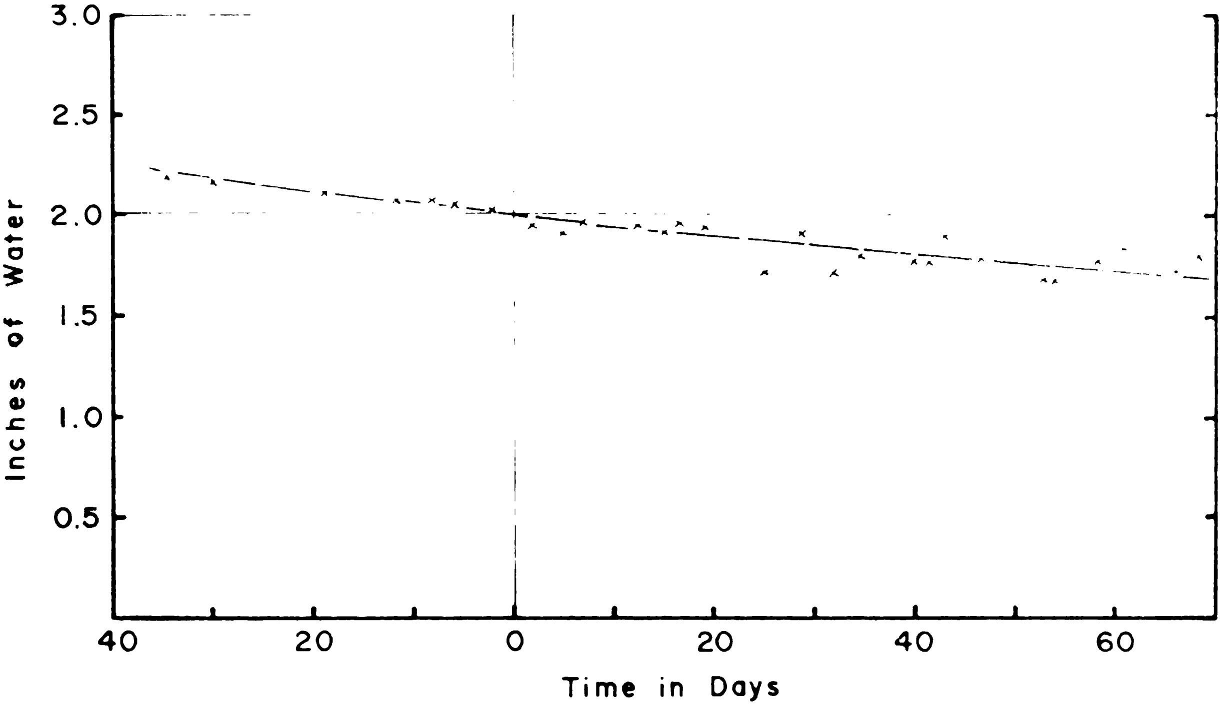


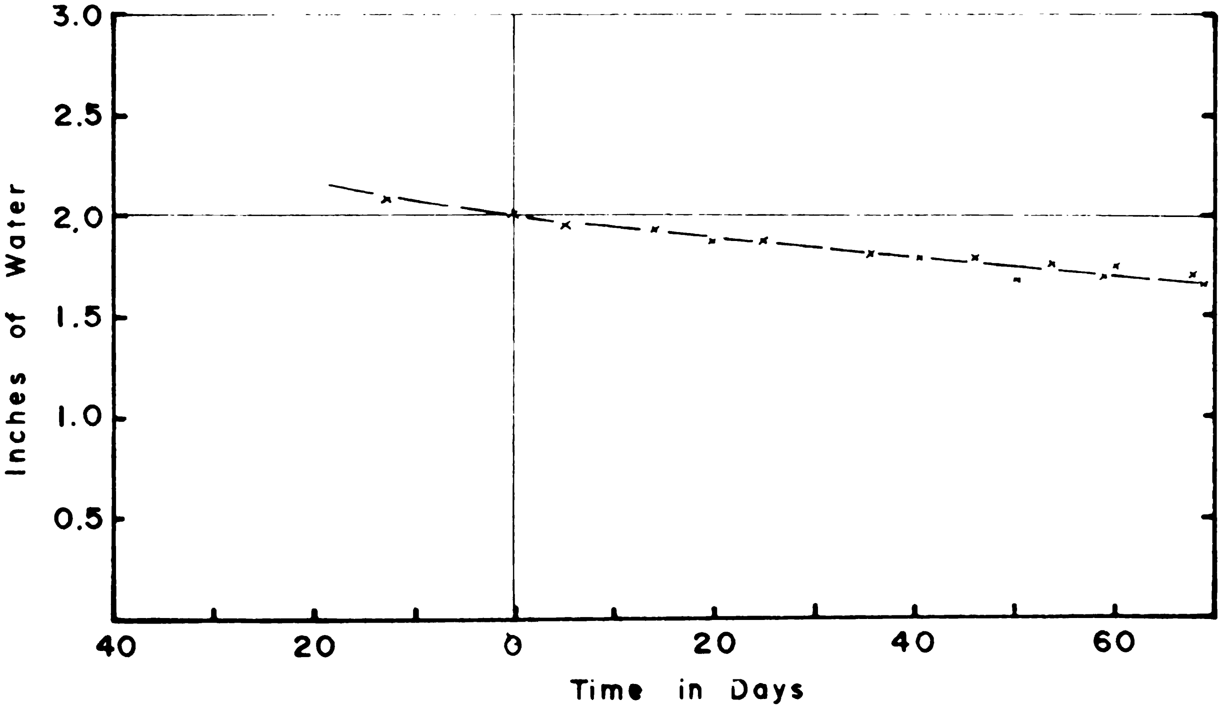
Figure 17c. Soil-moisture depletion curve for the 6 and 7-foot depths for 1962-1965 growing seasons. (Toumey Woodlot)

Figure 17d. Soil-moisture depletion curve for the 8 and 9-foot depths for 1962-1965 growing seasons (Toumey Woodlot)

6 & 7 FOOT DEPTHS



8 & 9 FOOT DEPTHS



over the central range of moisture contents encountered. The deviation of the depletion curves from the straight line indicated represents an unbalance between the inflow and outflow of soil water. The deviations from this straight line relationship are especially noteworthy in the surface depths of the profile. With increase in depth, there appears to be a better balance between the inflow and outflow of soil water. Depletion, therefore, in the lower depths more nearly represents evapotranspiration losses rather than losses due largely to internal drainage.

CHAPTER VII

SOIL WATER BUDGET

Few soil-moisture studies in the past, especially those reported in the forestry literature, have made any attempt to estimate the soil-water budget under different vegetation conditions. The few notable exceptions have been concerned with the soil-water budget under field crops or irrigated areas. This deficiency stems from the lack of a theoretical basis for separating soil-moisture depletion into evapotranspiration and deep-seepage components.

Subsequent to periods of soil-water recharge, water in the larger, non-capillary pores moves downward within the profile in response to gravity until capillary or upward forces prevent additional movement. As was noted previously, the rate of downward movement of soil water decreases exponentially with time. Therefore, it may be assumed that downward movement within the profile decreases during the periods of greatest evapotranspiration or depletion. Furthermore, the rate of drainage from the soil profile will be a function of the soil-moisture content. The first derivative of the total soil-moisture depletion curve (figure 15) will define the rate of total soil-moisture depletion as a function of moisture contents. This functional relationship for the 1962-65 growing seasons (figure 18) indicates a curvilinear relation between soil-moisture content and the growing season rate of daily total soil-moisture depletion. As the soil-moisture content decreases, the rate of total soil-moisture depletion rapidly decreases, whereas the daily rate of total soil-moisture depletion sharply increases with an increase in soil-moisture content. The exponentially decreasing rate of soil-water deep seepage may be expressed in the form:

$$W = AT^{B}$$

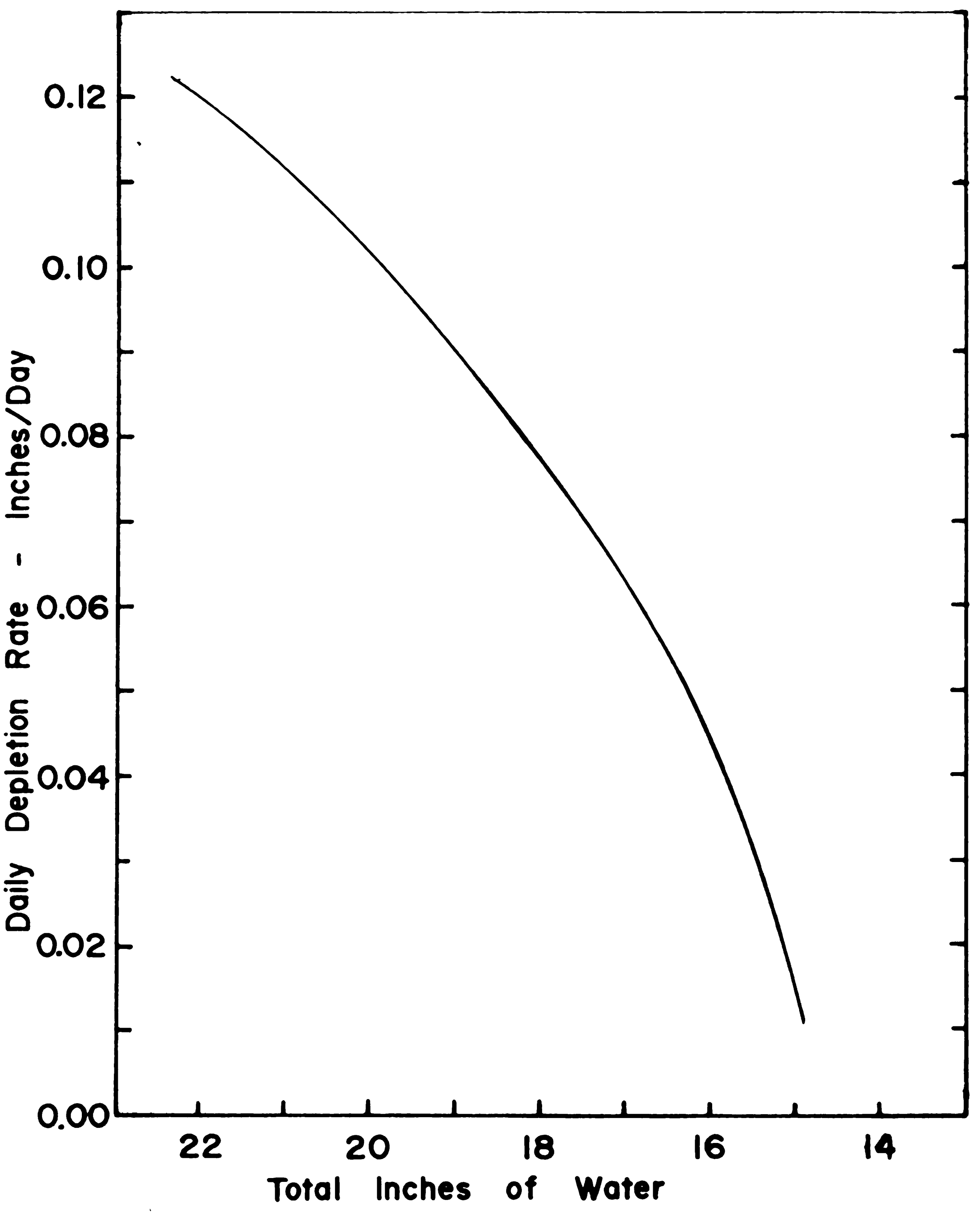
in which W is the moisture content of the soil, A and B represent constants for a particular soil, and T refers to time in days of drainage.

The total rate of soil-moisture depletion from the profile will consist of the sum of the loss rate due to deep seepage and the loss rate due to evapotranspiration. If the rate of evapotranspiration losses is assumed to be constant over a specified interval, the slope of the evapotranspiration curve will be constant.

The rate of deep seepage on any given day, therefore, will be the difference between the slope of the total soil-moisture depletion curve and the slope of the evapotranspiration rate curve. If the rate of evapotranspiration is assumed to be constant, the slope of the total depletion curve can be defined by the expression (Willardson and Pope, 1963):

$$\frac{d\,(\text{Total Depletion})}{dT} = C + AB\,(T)^{(B-1)}$$

Figure 18. Functional relation of daily rate of total soil-moisture depletion and total soil-moisture contents for the nine-foot profile: 1962-1965 growing season. (Toumey Woodlot)


Daily Depletion Rate - Inches/Day
0.12
0.10
0.08
0.06
0.04
0.02
0.00
22
20
18
16
14
Total Inches of Water

in which C is the assumed constant evapotranspiration rate. The remaining terms are as defined previously.

Evapotranspiration and deep-seepage losses must be interrelated and each influences the depletion rate of the other. The influence of one phenomena, however, upon the other is not known and therefore, calculations must be based upon simplifying assumptions.

The most common assumption is based on the classical concept of "field capacity". Under this concept, drainage from the profile becomes insignificant when the soil-moisture content is at or below "field capacity". Furthermore, it is assumed that recharge to the ground-water system will not occur unless the entire soil profile is at or above "field capacity". Total soil-moisture depletion is considered to be due solely to evapotranspiration losses when the soil-moisture contents are less than "field capacity".

The error induced by this simplifying assumption, however, is not too serious in investigations of soil-moisture availability and growth or survival. However, in hydrologic studies, the deep-seepage component of the total soil-moisture depletion may be a significant portion of the total water yield from an area, especially during periods of deficient precipitation.

The quantitative evaluation of the deep-seepage component of the total soil-water depletion becomes highly important under irrigated crops in relation to leaching of salts from the soil profile. Much of the work in the area of soil-water relations has been oriented in this direction.

The "field capacity" for the mean nine-foot profile in Toumey Woodlot is estimated to be approximately 21 inches of water. The estimate is based upon the examination of the soil profile following periods of soil-moisture recharge during the 1962, 63, 64, and 65 water years. The total soil-moisture contents equaled or exceeded "field capacity" only twice during the period of study. These periods were April, 1962 and April, 1965. Based upon the concept of "field capacity", recharge to the ground-water system occurred only twice during the period of study and only for short durations. The ground-water reservoir is not recharged until water is added at the water table; however, recharge at the water table is determined by the amount of recharge water that passes through the soil zone.

Although the rate of recharge may vary, the periods of larger amounts of recharge should be correlated with periods of heavier precipitation.

Two different approaches were explored for estimating the components of the soil-water budget for the 1962 through 1965 water years. In the first of these approaches (table 10), calculation of the deep-seepage component is based upon the assumption that soil-water movement below the depth of sampling occurs only

Table 10. Soil-water budgets for the 1962 through 1965 water years based upon the classical "field capacity" concept. (Toumey Woodlot)

	Water Year 1962	1963	1964	1965	Mean
	Inches				
Precipitation	21.31	22.30	23.44	25.53	23.15
Change in S.M. Storage	-2.79	+0.47	0.00	+1.42	-0.22
Total Water Losses	24.10	21.83	23.44	24.11	23.37
Estimated Deep Seepage	0.85	0.00	0.00	1.65	0.62
Evapotranspiration	23.25	21.83	23.44	22.46	22.75

when the entire soil profile is at or above "field capacity". The total soil-moisture storage did not reach "field capacity" during the 1963 and 1964 water years. Therefore, the evapotranspiration losses for these years would be equal to the total water loss for the respective year.

The change in soil-moisture storage (table 10) is the net difference between the total soil-moisture storage on October 1 and September 30 for each water year. These values represent the mean net difference for the 10 soil-moisture plots.

Loss of soil water below the depth of sampling is assumed to have occurred during the 1962 and 1965 water years as the total soil-moisture contents exceeded "field capacity" during the spring recharge periods. The deep-seepage component of the total soil-water budget for the 1962 and 1965 water years is estimated by examination of the soil-moisture and precipitation data for the periods when the soil-moisture contents were at or above "field capacity" (21 inches of water). The deep-seepage component is considered to consist of the total storage in the profile above "field capacity" plus the precipitation, corrected for interception losses, during the time interval the soil-moisture contents were at or above 21 inches. Interception losses were estimated by formulas presented by Helvey and Patric (1965). The difference between the total water losses and deep seepage is the evapotranspiration or consumptive use.

Evapotranspiration losses for the 1963 and 1964 water years utilized 98 percent and 100 percent respectively of the total annual precipitation.

The evapotranspiration losses calculated in this procedure include interception losses as well as soil-moisture depletion by evapotranspiration. Interception losses for the 1961 - 1965 period, as determined by formulas presented by Helvey and Patric (1965), represent 17.4 percent of the mean gross annual precipitation or 4.03 inches of water annually. Although the interception losses may be quantitatively significant, the overall effect may be somewhat lessened due to a temporary reduction in the rate of soil-moisture depletion and transpiration rate. For hydrologic purposes, however, interception, transpiration and evaporation may be considered as one quantity - evapotranspiration or consumptive use.

If we accept the hypothesis that continual downward movement of soil water occurs even though the moisture content is below "field capacity", the soil-water budget as calculated previously will over-estimate evapotranspiration losses. More significantly, the quantity of deep seepage and recharge to the ground-water system will be underestimated. Vegetation management and manipulation for increased water yields is aimed primarily at increasing this latter quantity. It is toward this goal that the

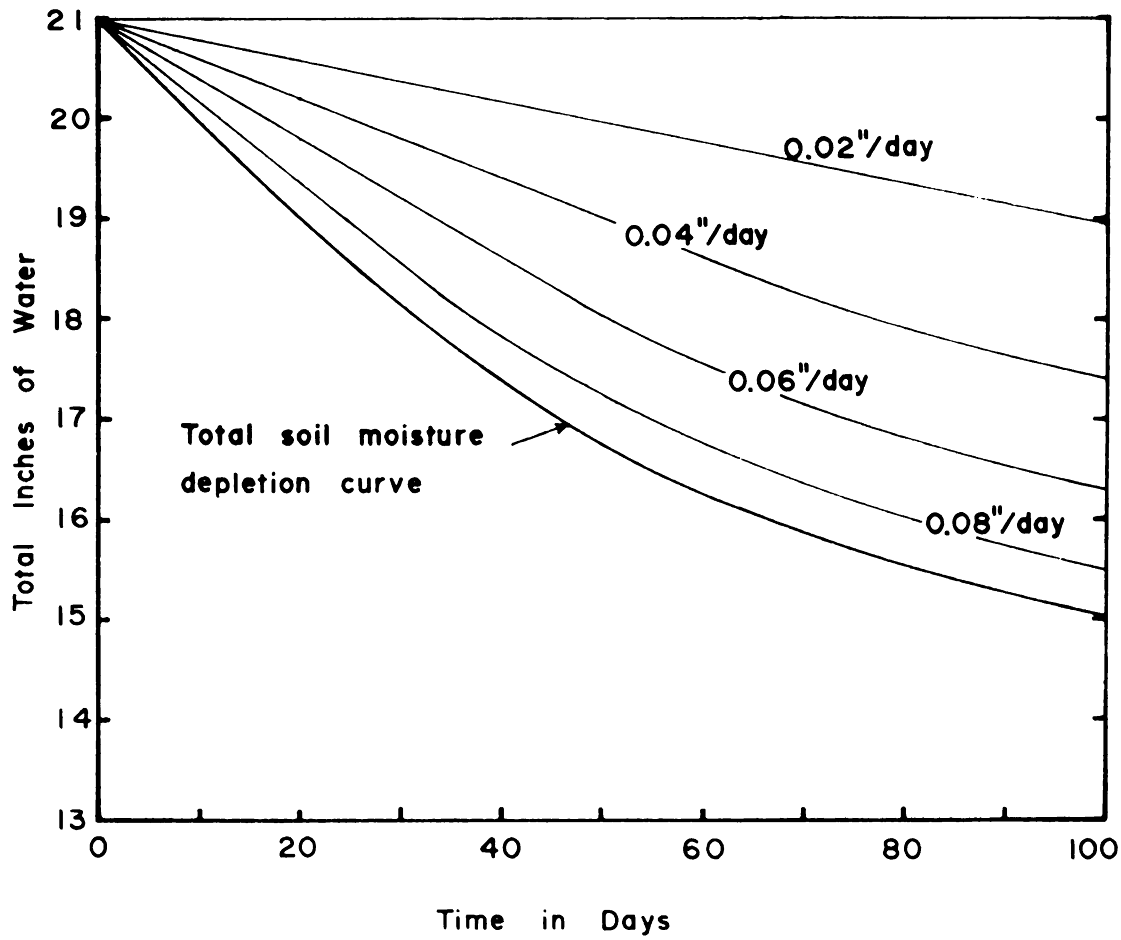
greatest potential benefits lie in watershed management.

The deep-seepage component of the total soil-water depletion curve can be estimated if the rate of evapotranspiration is known or can be approximated. If we assume that the rate of evapotranspiration is linear with time from the maximum moisture content to the moisture content at which the slopes of the total depletion curve and the evapotranspiration depletion curve are equal, the deep-seepage component may be determined by subtraction of the two curves. In the regions where the slope of the total depletion curve is less than that of the assumed evapotranspiration rate, the slope of the evapotranspiration curve will be considered to be identical to that of the total soil-moisture depletion curve. Under this latter condition, deep seepage is considered to be negligible.

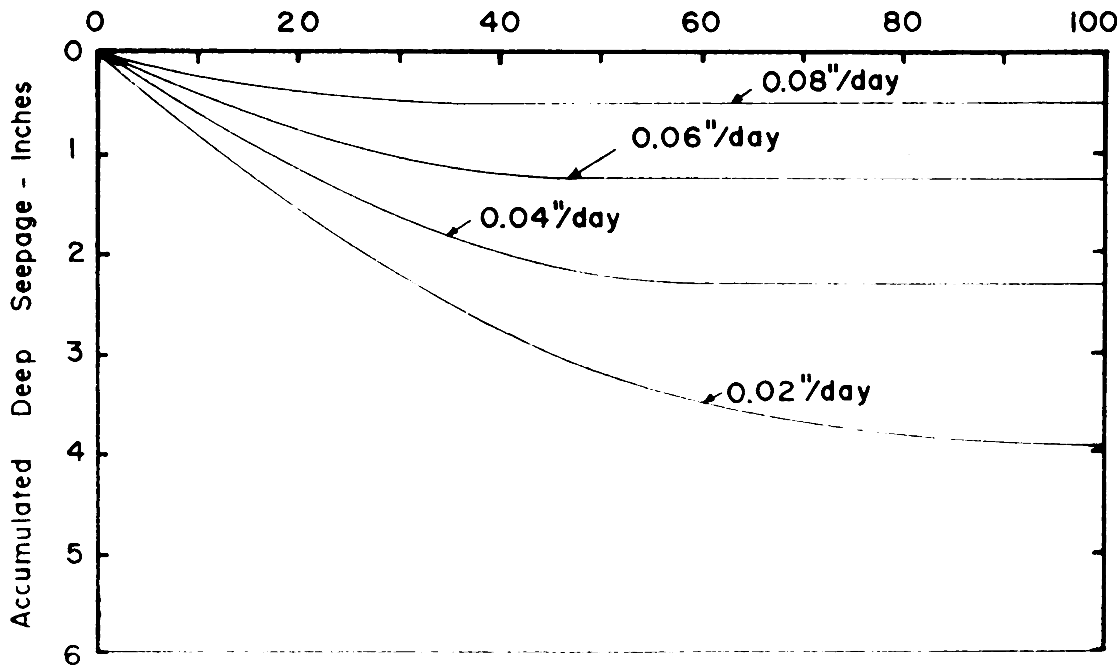
The accuracy of the approximation depends upon the validity of the assumptions of evapotranspiration rates in relation to the total depletion curve.

The accumulated deep-seepage component of the soil-water budget for assumed evapotranspiration rates of 0.02 inches/day, 0.04 inches/day, 0.06 inches/day, and 0.08 inches/day are shown in figure 19. The values of accumulated deep seepage range from 7 percent to 63 percent of the total seasonal soil-moisture depletion. Based upon the mean daily soil-moisture depletion rate

Figure 19. Deep seepage components of the total soil-moisture depletion curve for assumed growing season daily evapotranspiration rates and a maximum moisture content of 21 inches. (Toumey Woodlot)



21
20
19
18
17
16
15
14
13
Total Inches of Water
0.02"/day
0.04"/day
0.06"/day
0.08"/day
Total soil moisture depletion curve
0
20
40
60
80
100
Time in Days



0
20
40
60
80
100
0
1
2
3
4
5
6
Accumulated Deep Seepage - Inches
0.08"/day
0.06"/day
0.04"/day
0.02"/day

of 0.07 inches/day, approximately 0.7 inches of water moves downward below the depth of sampling during the first 100 days of the growing season. The accumulated value of deep seepage serves as a correction term which subtracted from the estimated evapotranspiration losses will give a more realistic value of consumptive use.

A similar type of relationship may be established for separating deep seepage and evapotranspiration losses during the dormant season. This relationship, however, will be applicable over short intervals between periods of recharge. The soil-moisture recharge and depletion patterns will be influenced by thermal gradients and soil freezing during the winter months, as well as the redistribution of soil water within the profile.

The component of deep seepage during the dormant season is considered to be small in comparison to that during the recharge period and the beginning of the growing season because of the relatively low total moisture contents. Soil-moisture recharge during the late fall and winter months normally is minor and restricted to the surface horizons. The major portion of the total seasonal recharge occurs during the later March through early May period. Recharge to the lower depths in the profile continues into the depletion period. Based upon these observations, the total quantity of deep seepage during the dormant season is small and therefore, is not determined as a separate quantity.

The soil-water budgets for the 1962 through 1965 water years based upon an assumed mean evapotranspiration rate of 0.07 inches/day are presented in table 11. The deep-seepage component is estimated from figure 19. The starting point was adjusted to correspond to the maximum measured moisture content for each water year. The deep-seepage component as determined is approximately 0.5 inches greater on the average than that determined under the classical "field capacity" concept.

The components of the soil-water budgets are estimates and their verifications must be based upon more refined techniques of measurement and a complete understanding of soil water - transpiration rate relationships.

Table 11. The soil-water budgets for the 1962 through 1965 water years based upon an assumed mean evapotranspiration rate of 0.07 inches/day. (Toumey Woodlot)

	Water Year 1962	1963	1964	1965	Mean
	Inches				
Precipitation	21.31	22.30	23.44	25.53	23.15
Change in S.M. Storage	-2.79	+0.47	0.00	+1.42	-0.22
Total Water Losses	24.10	21.83	23.44	24.11	23.37
Estimated Deep Seepage	1.55	0.20	0.30	2.35	1.10
Evapotranspiration	22.55	21.63	23.14	21.76	22.27

CHAPTER VIII

SUMMARY AND CONCLUSIONS

The soil-moisture regime in a nine-foot profile under a "realtively undisturbed" old-growth sugar-maple-beech stand was determined by the neutron-scattering technique during a $4\frac{1}{2}$ year period of subnormal precipitation. Precipitation was approximately 75 percent of the 1941-60 mean and soil-moisture deficiencies within the entire soil profile were noted. More significantly, the quantity of water available as ground-water recharge after losses to evaporation and transpiration was limited. This deficiency of precipitation during the 1961-65 period is reflected in the persistence of record low or near record-low ground-water levels in the south-central Michigan area during the study period.

The annual cyclic pattern of soil-moisture accretion and depletion in the total soil profile was similar from year to year but varied in magnitude due to differences in the distribution and amount of precipitation. Soil-moisture recharge began in the fall of the year, coincident with the start of the water year (October 1) and continued through the snow-melt period into May. Depletion began with the breaking of vegetative dormancy and the appearance of foliage in the spring of the year. Soil-moisture storage was continuously depleted throughout the growing season until the fall period when recharge again occurred.

Precipitation during the growing season normally did not penetrate to depths greater than three feet. Only minor periods of soil-moisture recharge were observed during the growing season. Based upon observations of soil-moisture recharge and depletion, precipitation during the growing season was utilized completely by evapotranspiration.

Soil-moisture depletion curves for the 1962, 1963, 1964, and 1965 growing seasons were similar for all years. The rate of soil-moisture depletion tended to decrease with both increase in depth in the soil profile and with decrease in the soil-moisture content.

The total soil-moisture depletion rate at any given time during the growing season is the sum of the evapotranspiration rate and the rate of deep seepage. The rate of deep seepage decreases exponentially with time and, therefore, decreases in importance with decrease in soil-moisture storage. Based upon assumed linear rates of evapotranspiration with time of 0.02 inches/day, 0.04 inches/day, 0.06 inches/day, and 0.08 inches/day, soil-moisture losses from the sampling zone due to deep seepage accounted for 63, 37, 17, and 7 percent, respectively, of the total seasonal soil-moisture depletion. Based upon a mean total soil-moisture depletion rate of 0.07 inches/day, deep seepage accounts for approximately 13 percent of the total seasonal soil-moisture depletion.

The existence of deep seepage of soil water below the depth of sampling is supported by examination of the soil-moisture accretion patterns. Recharge to the soil profile occurred behind a wetting front, the moisture content of which was at or near "field capacity". Following the cessation of the advance of the wetting front or infiltration, redistribution of soil moisture within the profile occurred in response to differences in total potential. By the redistribution of soil water within the profile, recharge occurred at all depths within the sampling zone even though the overlying horizons in the profile were not maintained at "field capacity". Recharge during the dormant season, after the redistribution of soil water, occurred almost uniformly in the upper five-feet of the profile. Recharge occurred at the lower depths of sampling but at a decreasing rate. Soil-moisture accretion patterns tended toward a uniform moisture distribution throughout the profile. This equilibrium condition, however, is probably never obtained under field conditions due to continuous changes in water content, temperature, and total potential.

The soil-moisture gradients within the soil profile are directed downward in the profile during the fall and winter recharge periods. Soil water tends to move downward in the profile under the moisture gradients imposed upon the soil-water system. Furthermore, the reduction of the temperature and the presence of frost in the surface soil during the winter months results in an upward migration of water vapor, tending to off-set the downward movement under moisture gradients.

The rate of soil-moisture depletion is greatest in the surface layers of the profile and decreases with depth in the profile. Soil-moisture gradients in the upper three feet of the profile are reversed toward the middle of the growing season. Under these conditions, soil water tends to move both upward and downward from the three to five foot depths.

Soil-moisture depletion occurs at all depths within the nine-foot profile but not at equal rates. The most readily available water appears to be depleted most rapidly regardless of the position in the profile. At all depths within the soil profile, the moisture content simultaneously approaches the "wilting point". This is supported by the minimization of the soil-moisture gradients within the soil profile toward the end of the growing season.

Estimated evapotranspiration losses as calculated under the assumption that no recharge occurs to the ground-water table unless the entire soil profile is at or above "field capacity", accounted for a mean annual loss of 22.75 inches or 98 percent of the mean annual precipitation. This estimate of evapotranspiration losses is an over-estimate because the calculation procedure does not account for deep-seepage losses below the depth of sampling.

Based upon a mean daily evapotranspiration rate of 0.07

inches/day, the partitioning of the total soil-moisture depletion curve for the growing season into deep-seepage and evapotranspiration components accounts for an additional 0.5 inches of water available as ground-water recharge during the growing season.

Deep-seepage losses below the depth of sampling are considered to be negligible during the dormant season because of the relatively low soil-moisture contents and depletion rates (less than 0.01 inches/day) measured during this period.

Soil-moisture recharge occurs primarily in the late March to early May period. Recharge during the early part of the dormant season is normally small and confined primarily to the surface horizons. Recharge to the lower depths in the profile during the spring recharge period normally lags four to six weeks behind recharge to the surface horizons. The maximum moisture storage in the lower horizons is not reached until after the beginning of the seasonal soil-moisture depletion period. Therefore, continued deep seepage following recharge in the spring of the year is included in the partitioning of the total soil-moisture depletion into deep-seepage and evapotranspiration components.

Evaluation of the results from this study and those from previous investigations reported in the literature, point up several common inadequacies in our basic understanding of vegetation-soil-water interrelationships. These inadequacies are primarily in the areas of the mechanics of soil-moisture movement and retention, ground-water recharge, and soil water-vegetation relationships.

The formation of a theoretical basis for evaluating the component parts of the hydrologic cycle and the soil-water budget is in its infancy. Intensive investigations under laboratory and field conditions are needed to fully understand soil-water behavior and to formulate a sound basis for partitioning changes in soil-moisture storage into recharge, evapotranspiration, and deep-seepage losses.

Furthermore, the relationships between soil-moisture availability, vegetation growth, and evapotranspiration are inadequately understood. Research in these areas is not only basic to forestry and hydrology but also to soil science, agriculture, and to intelligent management of our forested and agricultural lands to best meet the needs of the present and future generation.

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APPENDICES

APPENDIX I

Common and botanical names of tree species mentioned in text.

Common Name	Scientific Name
Ash, white	*Fraxinus americana* L.
Basswood, American	*Tilia americana* L.
Beech, American	*Fagus grandifolia* Ehrh.
Cherry, black	*Prunus serotina* Ehrh.
Elm, American	*Ulmus americana* L.
Elm, slippery	*Ulmus rubra* Mühl.
Hickory, bitternut	*Carya cordiformis* (Wangenh.) K. Koch
Hickory, shagbark	*Carya ovata* (Mill.) K. Koch
Hophornbeam, eastern	*Ostrya virginiana* (Mill.) K. Koch
Maple, black	*Acer nigrum* Michx. f.
Maple, red	*Acer rubrum* L.
Maple, sugar	*Acer saccharum* Marsh.
Oak, northern red	*Quercus rubra* L.
Oak, swamp white	*Quercus bicolor* Willd.
Oak, white	*Quercus alba* L.
Sycamore, American	*Platanus occidentallis* L.

APPENDIX II

Table 1. Mean measured soil-water contents expressed in inches of water per foot depth for the 1961 water year. (Toumey Woodlot, Michigan State University Campus, East Lansing).

SAMPLING DATE

DEPTH	Apr. 8	Jul. 21	Jul. 28	Aug. 4	Aug. 8	Aug. 18	Aug. 25	Sep. 8	Sep. 15	Sep. 30
1	2.34	1.91	1.73	1.93	1.79	1.58	2.24	2.15	1.99	2.02
2	2.47	2.06	1.97	2.08	2.00	1.93	2.10	2.15	2.07	2.10
3	2.47	2.21	2.09	2.08	2.05	1.98	2.09	2.08	2.08	2.11
4	2.26	2.05	2.11	2.04	2.00	1.94	2.03	2.04	1.97	1.99
5	1.99	1.93	1.85	1.85	1.81	1.74	1.76	1.80	1.75	1.75
6	1.93	2.02	1.97	1.94	1.94	1.96	1.87	1.91	1.88	1.88
7	1.69	1.86	1.80	1.80	1.80	1.75	1.75	1.78	1.72	1.72
8	1.79	1.94	1.92	1.89	1.88	1.89	1.84	1.96	1.82	1.80
9	1.53	1.77	1.68	1.69	1.68	1.65	1.61	1.63	1.61	1.60
Total	18.47	17.75	17.12	17.30	16.95	16.42	17.29	17.50	16.89	16.97

Table 2. Mean measured soil-water contents expressed in inches of water per foot depth for the 1962 water year. (Toumey Woodlot, Michigan State University Campus, East Lansing).

Depth	Oct. 1	Dec.29	Jan.20	Feb. 9	Apr.14	Apr.21	Apr.26	May 3
1	2.02	2.17	2.11	2.27	2.48	2.38	2.34	2.50
2	2.10	2.26	2.24	2.24	2.45	2.43	2.46	2.56
3	2.11	2.23	2.29	2.24	2.51	2.60	2.58	2.66
4	1.99	2.06	2.11	2.11	2.60	2.66	2.62	2.71
5	1.75	1.76	1.82	1.91	2.32	2.32	2.35	2.44
6	1.88	1.87	1.88	1.87	2.12	2.15	2.17	2.23
7	1.72	1.70	1.73	1.64	2.08	2.09	2.17	2.17
8	1.80	1.72	1.83	1.79	2.02	2.05	2.09	2.17
9	1.60	1.57	1.60	1.58	1.65	1.70	1.77	1.90
Total	16.97	17.34	17.61	17.65	20.23	20.38	20.56	21.34
Depth	May 12	May 17	July 7	Aug. 3	Sept. 7	Sept.21	Sept.28	Sept.30
1	2.28	2.23	2.16	1.64	0.93	1.21	1.23	1.23
2	2.46	2.45	2.33	1.98	1.48	1.68	1.70	1.70
3	2.60	2.57	2.39	2.05	1.33	1.67	1.66	1.66
4	2.59	2.59	2.30	1.97	1.52	1.57	1.55	1.55
5	2.35	2.32	2.08	1.86	1.61	1.46	1.45	1.45
6	2.20	2.12	2.11	1.96	1.79	1.68	1.70	1.70
7	2.24	2.19	1.94	1.78	1.73	1.62	1.62	1.62
8	2.20	2.21	2.08	1.94	1.83	1.69	1.72	1.71
9	1.99	2.00	1.86	1.80	1.64	1.56	1.57	1.56
Total	20.91	20.68	19.25	16.98	13.86	14.14	14.07	14.08

Table 3. Mean measured soil-water contents expressed in inches of water per foot depth for the 1963 water year. (Toumey Woodlot, Michigan State University Campus, East Lansing).

Depth	Oct. 1	Nov. 3	Nov.10	Nov.26	Dec. 1	Dec.21	Jan. 4
1	1.23	1.76	1.69	1.66	1.69	1.86	1.70
2	1.70	1.67	1.74	1.72	1.75	1.76	1.84
3	1.66	1.67	1.72	1.63	1.70	1.66	1.66
4	1.55	1.48	1.62	1.52	1.55	1.52	1.55
5	1.45	1.43	1.54	1.43	1.50	1.50	1.48
6	1.70	1.70	1.62	1.67	1.70	1.67	1.64
7	1.62	1.55	1.67	1.55	1.58	1.50	1.55
8	1.71	1.66	1.84	1.62	1.64	1.60	1.61
9	1.56	1.46	1.52	1.44	1.50	1.45	1.46
Total	14.18	14.38	14.96	14.24	14.61	14.52	14.49
Depth	Jan.11	Feb. 8	Mar.8	Mar.11	Mar.15	Mar.25	Apr. 1
1	1.69	1.80	2.01	1.87	1.97	2.41	2.38
2	1.82	1.78	1.78	1.87	1.85	2.21	2.34
3	1.70	1.70	1.78	1.70	1.73	2.15	2.21
4	1.55	1.55	1.61	1.55	1.58	1.99	2.02
5	1.50	1.50	1.55	1.48	1.52	1.76	1.81
6	1.66	1.66	1.68	1.63	1.68	1.66	1.70
7	1.55	1.55	1.62	1.50	1.61	1.56	1.60
8	1.62	1.62	1.64	1.50	1.63	1.54	1.78
9	1.46	1.46	1.49	1.30	1.49	1.36	1.49
Total	14.55	14.62	15.16	14.40	15.06	16.64	17.33

Table 3 (continued)

Depth	Apr. 15	Apr. 23	May 2	May 13	May 20	May 27	June 6
1	2.22	2.33	2.27	2.32	2.29	2.16	2.02
2	2.34	2.39	2.38	2.46	2.42	2.38	2.30
3	2.22	2.24	2.28	2.46	2.41	2.41	2.33
4	2.00	1.98	2.02	2.17	2.20	2.21	2.15
5	1.81	1.78	1.79	1.87	1.87	1.90	1.85
6	1.73	1.72	1.74	1.78	1.78	1.82	1.80
7	1.61	1.60	1.73	1.66	1.63	1.66	1.67
8	1.62	1.61	1.64	1.66	1.63	1.67	1.68
9	1.48	1.46	1.51	1.43	1.52	1.51	1.43
Total	17.03	17.11	17.36	17.81	17.75	17.72	17.23
Depth	June 14	June 21	July 5	Aug. 9	Aug. 16	Sept. 25	Sept. 30
1	2.22	2.16	1.75	1.45	1.51	1.50	1.35
2	2.51	2.38	2.16	1.81	1.78	1.78	1.73
3	2.56	2.52	2.34	1.76	1.74	1.81	1.76
4	2.38	2.50	2.29	1.66	1.63	1.73	1.66
5	2.08	2.11	2.05	1.58	1.54	1.55	1.52
6	2.03	2.08	2.07	1.73	1.72	1.80	1.75
7	2.02	1.93	1.95	1.68	1.69	1.68	1.64
8	1.80	1.86	1.86	1.68	1.66	1.76	1.72
9	1.51	1.60	1.63	1.50	1.49	1.56	1.52
Total	19.11	19.14	18.10	14.85	14.76	15.17	14.65

Table 4. Mean measured soil-water contents expressed in inches of water per foot depth for the 1964 water year. (Toumey Woodlot, Michigan State University Campus, East Lansing).

Depth	Oct. 1	Oct. 5	Oct. 18	Nov. 8	Nov. 27	Jan. 23	Feb. 25	Mar. 12	Apr. 9	Apr. 23	May 15
1	1.35	1.21	1.09	1.13	1.20	1.50	1.79	2.10	2.39	2.50	2.44
2	1.73	1.68	1.67	1.64	1.66	1.73	1.93	2.17	2.44	2.50	2.47
3	1.76	1.72	1.67	1.67	1.69	1.73	1.84	2.00	2.50	2.56	2.59
4	1.66	1.58	1.54	1.54	1.62	1.62	1.64	1.78	2.21	2.46	2.63
5	1.52	1.48	1.48	1.51	1.48	1.63	1.51	1.56	1.69	1.91	2.23
6	1.75	1.70	1.63	1.68	1.67	1.68	2.02	1.61	1.75	1.81	2.02
7	1.64	1.61	1.56	1.62	1.60	1.56	1.61	1.60	1.62	1.65	1.76
8	1.72	1.67	1.63	1.62	1.62	1.63	1.65	1.66	1.69	1.69	1.82
9	1.52	1.48	1.48	1.48	1.45	1.46	1.49	1.48	1.50	1.51	1.56
Total	14.65	14.13	13.75	13.89	13.99	14.54	15.48	15.96	17.79	18.59	19.52

Depth	May 27	Jun 18	Jun 26	Jul.10	Jul.24	Aug. 5	Aug.23	Sep. 8	Sep.20	Sep.30
1	2.09	2.26	2.24	2.17	1.64	1.55	1.50	1.37	1.28	1.26
2	2.34	2.17	2.36	2.15	2.00	1.92	1.85	1.80	1.74	1.72
3	2.53	2.36	2.41	2.17	2.04	1.98	1.91	1.82	1.78	1.77
4	2.46	2.34	2.33	2.15	2.06	2.00	1.93	1.81	1.73	1.71
5	2.17	2.08	2.05	2.02	1.86	1.84	1.75	1.66	1.58	1.56
6	2.05	2.04	1.99	1.96	1.94	1.91	1.86	1.85	1.81	1.80
7	1.85	1.82	1.82	1.75	1.74	1.73	1.70	1.66	1.63	1.63
8	1.75	1.78	1.74	1.74	1.74	1.73	1.72	1.70	1.69	1.69
9	1.61	1.57	1.55	1.56	1.56	1.55	1.54	1.52	1.51	1.51
Total	18.85	18.42	18.49	17.67	16.58	16.21	15.76	15.19	14.75	14.65

Table 5. Mean measured soil-water contents expressed in inches of water per foot depth for the 1965 water year. (Toumey Woodlot, Michigan State University Campus, East Lansing).

Depth	Oct. 1	Oct. 27	Nov. 10	Dec. 2	Feb. 1	Mar. 18	Apr. 14	May 3	May 19
1	1.26	1.20	1.32	1.49	1.93	2.41	2.90	2.75	2.54
2	1.72	1.69	1.73	1.75	1.99	2.29	2.57	2.56	2.38
3	1.77	1.75	1.75	1.79	2.04	2.34	2.82	2.66	2.50
4	1.71	1.66	1.65	1.65	1.81	2.33	2.68	2.69	2.58
5	1.56	1.51	1.51	1.51	1.60	2.05	2.44	2.40	2.33
6	1.80	1.78	1.78	1.75	1.80	1.94	2.28	2.22	2.16
7	1.63	1.61	1.60	1.60	1.69	1.78	2.27	2.18	2.16
8	1.69	1.69	1.68	1.66	1.68	1.81	2.16	2.08	2.06
9	1.51	1.50	1.48	1.45	1.44	1.50	1.76	1.69	1.75
Total	14.65	14.39	14.50	14.65	15.98	18.45	21.88	21.23	20.46

Depth	June 5	June 17	July 1	July 15	Aug. 2	Aug. 10	Sept. 1	Sept.24
1	2.55	2.22	1.89	1.84	1.79	1.63	1.47	2.29
2	2.44	2.30	2.15	2.13	1.98	1.97	1.86	2.02
3	2.48	2.35	2.23	2.20	1.99	1.92	1.86	1.80
4	2.54	2.40	2.26	2.00	1.89	1.73	1.74	1.71
5	2.26	2.16	2.07	2.03	1.89	1.76	1.63	1.56
6	2.13	2.09	2.05	2.04	2.00	1.93	1.89	1.82
7	2.06	2.00	1.94	1.89	1.87	1.80	1.75	1.70
8	2.01	1.99	1.97	1.92	1.87	1.84	1.82	1.73
9	1.69	1.73	1.77	1.70	1.62	1.57	1.50	1.46
Total	20.16	19.24	18.33	17.75	16.90	15.95	15.52	16.09