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Volume VIII Number 1

THE WATER BALANCE

by

C. W. Thornthwaite

and

J. R. Mather

CENTERTON, NEW JERSEY

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PREFACE

During my early teaching period in the University of Oklahoma I became aware of the importance of the moisture factor in climate and came to appreciate that we could not gauge the wetness or dryness of a climate by considering the precipitation alone, but would need information on evaporation as well. In my climatic classification of 1931 I recognized the important role of moisture in climate and employed an empirical formula to obtain a moisture index.

Later, in the Soil Conservation Service, where I had the task of organizing climatic research leading to soil and water conservation we were again faced with the necessity to measure evaporation as reliably as precipitation. Between 1937 and 1942 some of my associates and I directed our labors toward the development of means to determine the evaporation from a natural surface. We explored various possibilities and adopted the so-called vapor transport method which involves measurement of the gradient of water vapor in the air layer near the ground and determination of the intensity of turbulent diffusion in the layer. The program involved perfection of turbulence theory and the improvement of instruments but was interrupted when the staff became involved in the war effort.

When I was faced with the problem of determining water needs for irrigation in the Republic of Mexico, while serving as consultant to the Comisión Nacional de Irrigación during World War II, I came to realize that we could not determine the amount by which precipitation fails to supply the water needs of crops unless one knows what these water needs are. This most important climatic element was defined as the amount of water which will be lost from a moist soil surface completely covered with vegetation. I called it potential evapotranspiration.

The concept of potential evapotranspiration first presented in 1944 was originally employed to provide information on the water needs of vegetation and in the solution of the hydrologic problems of stream flow and ground water storage. It was computed by means of an empirical formula. This new formula became the foundation of my climatic classification of 1948, which provides a more rational interpretation of the moisture factor. Now we could work out the water balance of an area, and could gain new insight into its site qualities by determining potential and actual evapotranspiration and water surplus and deficit. Scientists in all parts of the world became interested.

At the same time we sought various independent means of determining potential evapotranspiration for the purpose of verifying and if necessary modifying the formula. In 1945 I designed an instrument which we called an evapotranspirometer, which the Mexican Irrigation Commission put into service the following year. Since 1946 batteries of evapotranspirometers have been put into service in many places all over the world. There is now a considerable literature on evapotranspirometry.

From studies that my associates and I have conducted at the Laboratory of Climatology since 1946 we have gained new insight into the physics of evaporation and into the water relations of climates. We have reached a clearer understanding of the various problems which are involved in the application of the potential evapotranspiration concept to the determination of the water balance. These studies have been reported in a series of papers. Other papers seek to explain and refine the original procedures and use them in the solution of different research problems. These papers have been in great demand by scientists everywhere and many of them are now out-of-print.

We have felt for a long time that it would be desirable to collect together in one publication the pertinent papers dealing with the concept of potential evapotranspiration and the water balance. These papers should be edited so as to eliminate repetition and obsolete material. They would be issued as a single report which could serve to sum up the results of the past decade of work on evapotranspiration at the Laboratory of Climatology. This report would serve two purposes at one time. It would make available to those who have requested publications, the important material which has appeared in previous papers, now out-of-print, without the necissity of full reprinting of each article. It would also give researchers everywhere an opportunity to become acquainted with the latest developments in this important phase of climatology.

This is the task which Dr. Mather and I have undertaken in preparing the present report. There has been no effort to make it an original report. Rather we have used material already published in the Laboratory of Climatology, modifying it only so as to bring it up to date and to maintain continuity. In that way we have tried to show the development of the whole concept and at the same time provide a comprehensive survey of the whole field in relatively brief compass.

At the beginning of each chapter there is a detailed bibliography indicating from which previous articles the material in the chapter has been adapted. At the end of the report there is a comprehensive bibliography of other articles dealing with the concept of potential evapotranspiration and the water balance. This bibliography is by no means complete but it should give an idea of the wide-spread interest that has been aroused in recent years. It should also, of course, provide source material for those desiring additional information.

Full instructions for evaluating the water balance and samples of the tables and nomograms which are used to simplify the making of the computations are included as Appendix I of this report.

C. W. Thornthwaite

Centerton, New Jersey
January 31, 1956

ABSTRACT

In any adequate program of research development, or operation it is necessary to understand the climatic influences involved. In most instances, the primary climatic influence is the moisture relationship. This report attempts to explain the development and evaluation of the water balance and to show its usefulness in understanding these moisture relationships.

In order to be able to utilize the water balance it is first necessary to understand the concept of potential evapotranspiration. Thus, the first part of the report includes a discussion of the problems and limitations which are involved in any attempt to measure evapotranspiration experimentally or to compute it from available climatic data. Having shown that instruments such as the Piche, Class A water pan, or lysimeters do not provide adequate measures of potential evapotranspiration there is an effort to list the necessary conditions under which this important climatic factor can be measured directly. The report then describes the modifications which have been developed to provide more realistic results from the use of the water balance bookkeeping procedure.

The second part of the report is devoted to a discussion of the application of the various factors of the water balance in different lines of research. Checks on the validity of the results which are possible have shown that quite reasonable values of soil moisture storage, moisture deficit, and moisture surplus have resulted from the use of the procedure. Examples of its use in the definition of drought and the scheduling of supplemental irrigation, in the determination of stream runoff and fresh water accession to coastal estuaries, in the computation of soil moisture content and the ability of men to move over unpaved surfaces, and in the achieving of moisture indices which can be used in studies of classification and distribution indicate only a few of the many fields in which knowledge of the water balance can be used with worthwhile results.

The report concludes with a bibliography of articles dealing with potential evapotranspiration and the water balance which have been published by scientists in all parts of the world. An appendix provides the detailed instructions, tables, and nomograms which are necessary to permit the easy evaluation of the new water balance on either a daily or monthly basis.

INTRODUCTION*

The principal elements of climate that are observed are temperature, pressure, humidity, wind, solar radiation, and precipitation. The first four of these elements are qualities of the atmosphere, but the last two relate rather to the earth's surface, one constituting the source of soil-temperature and the other the source of soil-moisture. Radiation and precipitation are climatic factors of major importance, because climate has to do with more than the state of the atmosphere. Climate deals, in addition, with the conditions of temperature and moisture of the soil and with the interactions between soil-surface and atmosphere. Soil-temperature and soil-moisture are climatic elements.

The sum of the climatic elements that have been under observation does not equal climate. One element conspicuously missing from the list is evaporation. The combined evaporation from the soil surface and transpiration from plants, called "evapotranspiration", represents the transport of water from the earth back to the atmosphere, the reverse of precipitation. The rain gage measures precipitation within acceptable limits of accuracy. We know reasonably well how rainfall varies from one place to another over the inhabited parts of the earth and also how it varies through the year and from one year to another. On the other hand, few actual observations of the water movement from the earth to the atmosphere have been obtained and consequently we know next to nothing about the distribution of evapotranspiration in space or time.

Part of the moisture that is stored in the soil during rains evaporates directly back to the air from the soil surface and part is available to plants. Most of the water that enters a plant through its roots is later transpired from its leaves and stems. The amount of water in the root zone of the soil available to plants, to recharge the soil, and to supply streams and lakes depends in large measure on the relation between precipitation, which adds water, and evapotranspiration, which removes it. Since precipitation and evapotranspiration are due to different things, they are not often the same either in amount or in distribution through the year. In some places more rain falls month after month than the vegetation can use. The surplus moves through the ground and over it to form streams and rivers and flows back to the sea. In others, month after month, there is less water in the soil than the vegetation could use if it were available. There is no excess of rainfall and no runoff, except locally where the soil cannot absorb all the water as it falls. Consequently, there are no permanent rivers and there is no drainage to the ocean. In still other areas the rainfall is deficient in one season and excessive in another, so that a period of drought is followed by one with runoff. The march of precipitation through the year never coincides exactly with the changing demands for water in any part of the world.

Let us suppose that the rainfall and the water needed by plants at a place in southeastern United States vary through the year as shown in figure 1. Only a little more than a half-inch of water is needed in each of the winter months. The need rises rapidly during the spring and reaches a high point of nearly 7 inches in July. It falls rapidly during the autumn months. Between 3 and 4 inches of rain falls in each of the six winter and spring months. The largest amounts of rainfall, over 5 inches, come in July and August. In the driest months, October and November, only a little more than 2 inches of rain falls.

* The material in this chapter has been adapted from "Climate and Moisture Conservation," by C. W. Thornthwaite, Annals Ass. Amer. Geogr., Vol. 37, No. 2, 1947, pp. 87-100; "The Moisture Factor in Climate," by C. W. Thornthwaite, Trans. Amer. Geophys. Union, Vol. 27, No. 1, 1946, pp. 41-48; and "Modification of Rural Microclimates," by C. W. Thornthwaite, Background Paper No. 27, prepared for the Wenner-Gren Foundation International Symposium, "Man's Role in Changing the Face of the Earth," Princeton, June 16-22, 1955, 35 pp.

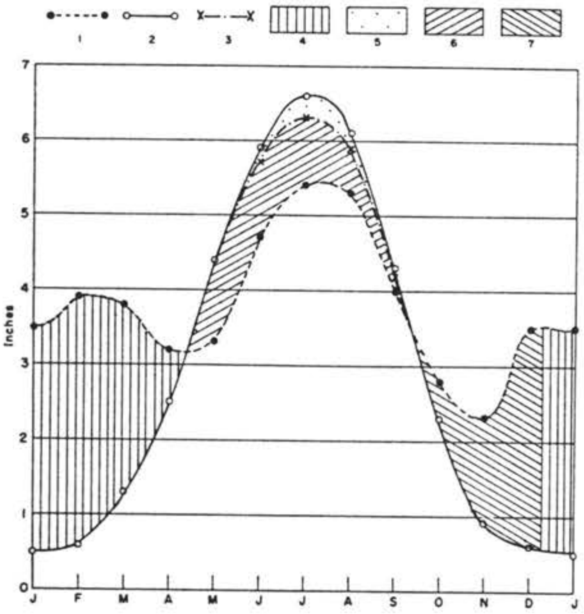

Figure 1

Relation of precipitation (1) to potential evapotranspiration (2) and actual evapotranspiration (3) at Camden, South Carolina showing periods of water surplus (4), water deficit (5), soil moisture utilization (6), and soil moisture recharge (7).

In this example, precipitation and water need do not coincide. There is too much precipitation in winter and too little in summer. In mid-autumn, water need falls below precipitation. For a while, the surplus rainfall replaces soil moisture that had been used up previously. From then on the surplus water raises ground-water levels and produces surface and subsurface runoff. But it is of little benefit to plants. In spring, both transpiration and evaporation increase rapidly and soon water need surpasses precipitation. The excess demands for water are satisfied in part by the current precipitation and from the stored soil moisture reserves. That part of the total water demands which is not met in this manner is known as the moisture deficit. As the soil dries and a smaller proportion of the water demands are satisfied through precipitation and the utilization of stored soil moisture, deficit increases and the plants begin to suffer severely from the lack of moisture. The actual loss of water from the soil and plants, the actual evapotranspiration, is equal to the water needs or potential evapotranspiration during those periods when precipitation exceeds water need. However, when precipitation is less than the demands for water the actual evapotranspiration is less than the potential and a moisture deficit equal to the differences between these two quantities exists.

When the seasonal course of precipitation is compared with the course of water need as in figure 1 one obtains information on many aspects of the water relations at a place. First, it is possible to determine the actual evapotranspiration or the actual loss of water from the plant and soil surfaces which, in almost all cases, is different from the potential water need. It is extremely difficult to measure actual evapotranspiration in practice because of the dependence of this quantity on such factors as soil type and method of land cultivation, type of plant cover, and moisture condition of the soil profile. Second, the difference between the potential and the actual water need provides a measure of the moisture deficit of a place, the amount by which the available moisture fails to satisfy the demand for water. Third, during those periods when the water need is greater than the precipitation there is a water demand which is met in part by utilization of stored soil moisture. Thus, it is possible to determine from the water balance the amount of moisture stored in the ground at any time either on a daily or monthly basis. Fourth, when precipitation is in excess of water need, the excess moisture will first be used to recharge the soil. When this has occurred, the remaining excess water will be water surplus and lost from the soil either by surface or subsurface runoff. This is the water that will ultimately find its way back to the streams and rivers or be used to recharge the ground water table. Fifth, if certain assumptions are made concerning the detention and melt of snow it is possible to determine from the water balance the daily or monthly accumulation of snow and its rate of release as water in the spring of the year.

The information, thus, provided by the determination of the water balance is of utmost utility in many different fields of research. For instance, knowledge of the moisture deficit is basic to any understanding of the economic feasibility of irrigation for it provides information on the total volume of water needed at any time and gives a definitive measure of drought. When compared with the moisture surplus in other seasons it makes clear whether there is enough water present during the year to permit irrigation. At the same time determination of the changes in soil moisture storage on a daily basis gives information on the state of the moisture in the ground at any time for use in the scheduling of the time and amount of supplemental irrigation to avoid the occurrence of any drought which will limit agricultural yields.

Information on the water surplus, the amount by which the precipitation exceeds the water needs when the soil is at field capacity is fundamental in any hydrologic studies which deal with the recharge of the ground water table or with the runoff of water in streams and rivers. By definition, this water surplus is that water which does not remain in the surface soil layers but is available for deep percolation to the water table and overland or subsurface flow to the water courses. Thus, information on water surplus climatically determined from the water balance provides a knowledge of stream flow which could only otherwise be obtained from extensive stream gaging installations and data on flow to the ground water table which would require detailed well records.

In oceanographic work, the water surplus is important in determining the amount of fresh water flow from the land to the ocean. This is particularly significant in the case of bays, estuaries and seas which are almost entirely landlocked, for the volume of fresh water runoff to the water body will be important in determining the salinity, density and other characteristics of the water.

More fundamental than knowledge of the flow into restricted basins is the picture of the large scale flux of moisture between land and sea. The water balance provides information on the detention of moisture on and within the land areas of the earth through the year. The seasonal fluctuations in the amount of land storage is reflected in the world-wide change in ocean levels through the year, a subject of vital importance not only to oceanographers but also to workers in other fields who are concerned with problems of large scale moisture flux.

It is during the period of water surplus that most of the problems with respect to the movement of men and vehicles over unpaved surfaces arise. On all but the most sandy soils, an increase in soil moisture content to or above the field capacity will result in a loss of bearing capacity and shearing strength and an increase in stickiness. This results in a lowered ability of men and vehicles to move over ground surfaces. On sands, of course, the reverse is true and as the moisture content increases up to the point where quicksand conditions exist the tractionability, or the ability of men and vehicles to traverse unpaved surfaces, improves.

The water balance will result in information not only on the periods of moisture surplus and deficit but will permit the magnitude of these quantities to be compared with one another and with the water need in order to provide climatic indices which can be used in classification and correlation studies. Precipitation and the climatically determined water need are truly active factors in climate and as such they must serve as the basis for any rational classification of the climates of the earth. At the same time a comparison of moisture surplus or deficit with the water need provides indices of humidity and aridity which can be correlated with the distribution of vegetation.

Probably more important than the use of the water balance in any one of these many lines of research is the fact that with the ability to secure information on all phases of the moisture relationships of an area from a knowledge of the climatic conditions comes the ability to determine long years of record of these moisture parameters. This is significant for it not only makes available data for periods when actual measurements were not made but also permits the accumulation of a record which can be utilized in statistical studies. Many problems such as the determination of probabilities of the amount of surplus water available for stream flow or of the occurrence of soil moisture conditions which would affect the movement of men or vehicles over unpaved surfaces or of drought can only be solved in this manner at present since measured values of the necessary parameters are not available for a long enough period of time. Such statistical studies are necessary in any attempt to understand the economic feasibility of any program of action as well as an aid in forecasting the outcome of the program.

The three important causal factors in climate which, if influenced by man, would result in some climatic modification are solar radiation, the general circulation, and surface features. Any realistic appraisal of the ways in which man might influence the first two of these shows that such influence must be transitory and of small effect. Most of the changes man can produce on the surface features, too, result in only local or temporary climatic changes. There is, however, one significant area - the water economy - where man can significantly influence climates over large areas and on a more permanent basis.

Almost every change of environmental conditions which man can make results in some change in the water economy or water budget at the earth's surface. In the regions where precipitation is continuously excessive and where drainage is the principal modification which man makes in the hydrologic regime, there is little possibility of bringing about any significant change in climate. Similarly, where precipitation is continually deficient, irrigation is man's means for changing nature. However, water is in such short supply that only a small fraction of the area can be irrigated. In the oases which result, there are important local microclimatic changes, but no changes that have any widespread influence.

There are, however, large areas of the world which have excessive precipitation during one season of the year and a lack of water during another. It is these areas where there are relatively abundant supplies of water available and where really significant climatic changes can be produced through the use of perhaps only a small amount of irrigation at just the proper time and in the correct amounts. These areas which have a small need for water now and which may either be producing crops with poor yields or no crops at all have potentialities for future great development not present in the desert areas with great water deficiencies or in areas with water surplus where drainage is necessary.

It is clear that many of man's activities can influence in one way or another the soil moisture relations. Many of these activities have resulted in a decrease in the amount of water which a soil can hold and hence tend to make it even more susceptible to droughts than it might otherwise be. For instance, the cultivation of soil greatly disturbs the soil structure and, when wheeled vehicles move over moist soil, some rearrangement of the soil particles occurs resulting in a compaction of the soil. This effect is to reduce the permeability of the soil, to lower its ability to hold water and to make it extremely difficult for water or plant roots to penetrate the compacted layer.

For example, the water available to plants in the upper layer of a Sassafras silt loam soil in a cultivated field in southern New Jersey has been found to be 2.17 inches per foot depth. The sod-covered soil along the border of this field holds 3.15 inches of water per foot. Thus, the loss of water holding capacity due to cultivation in this field is nearly an inch of water per foot of soil.

Light sandy soils, of course, are able to hold less water than heavy silt or clay soils. To a considerable extent plants compensate for the lower water holding capacity of sandy soils by deeper root penetration and more rapid root development. There is no similar compensation, however, when the water holding capacity of a soil is reduced by misuse. Misuse destroys the soil structure and reduces air capacity which inhibits root development. Thus, as water holding capacity diminishes, aeration does also, and the root zone becomes shallower.

The effects of vegetation changes, by grazing, burning, cropping, substitution of species and clearing, on the amount of water which enters or is retained in the soil and hence on the water economy of an area, are also noteworthy. The elimination of transpiration by stopping plant growth will always result in making more water available for soil moisture storage or for runoff. Whether the water enters the soil or runs off over the surface producing harmful erosion depends on the type of surface cover that remains. Uncontrolled burning of vegetation, of course, destroys not only the aerial parts of the plants but also much of the surface organic material and hence will lead to more runoff and less soil storage of the increased water supply. Grazing or over-grazing will result in two changes in the environmental conditions which will influence the microclimate. First, the removal of vegetation by the grazing animal will reduce transpiration and result in more of the precipitation being made available for soil storage or runoff. Second, however, the compaction of the soil by the hoofs of the animals will reduce the capacity of the soil to absorb water and hence make it less able to store water. The additional water made available by reduced transpiration will mostly run off with the possibility of erosion damage.

Since the water budget is such an important area of human influence it is desirable to consider how it can be determined from available data and how the information can then be used. Because knowledge of the water balance provides information on the periods of moisture surplus and deficit, and permits the determination of the amount of moisture stored in the soil or lost through runoff at any time the moisture budget becomes the basic tool in any evaluation of the water problems of a place or of an area.

PART I

THE CONCEPT OF POTENTIAL EVAPOTRANSPIRATION AND THE WATER BALANCE

CHAPTER 1. POTENTIAL EVAPOTRANSPIRATION*

The exchange of heat and of moisture between the earth's surface and the atmosphere are both processes of paramount importance to climate. Much study has been devoted to the problems of the heat budget and a vast body of knowledge regarding it has been accumulated. On the other hand, little is known of the moisture budget of the earth. At present, a number of maps of uncertain reliability show the evaporation over the oceans as computed by the use of various formulae. There are detailed maps of precipitation covering certain of the land areas and general maps of the whole earth. Maps of surface runoff have been derived from stream gaging records for certain of the land areas. For the areas where the distribution of precipitation and runoff are both known it has been possible to determine the evapotranspiration as a difference. In this indirect manner some knowledge of the magnitude and distribution of evapotranspiration has been obtained. These few moisture determinations and incomplete maps are of limited value in evaluating the moisture budget over extensive portions of the earth's surface.

For a number of years we have devoted a large share of our attention to an analysis of the moisture factor in climate. The novel feature of Thornthwaite's 1931 climatic classification was the recognition of the major importance of moisture regions. It employs an empirical formula to obtain a moisture index. Between 1937 and 1942 work was directed toward the development of means for determining the evaporation from any natural surface by the so-called vapor transport method which involved measurement of the gradient of water vapor in the air layer near the ground and determination of the intensity of turbulent diffusion in the layer. At the time of its presentation the method proved to be impractical because no instruments were available to make the precise measurements needed and because turbulence theory was insufficiently developed. In the intervening years considerable progress has been made in both instrumentation and theory; several improved instruments have been designed and developed, and recently significant advances in the theory of atmospheric turbulence have occurred. Even so, it is not likely that for some time it will be possible to measure the evaporation from the earth's surface with the same accuracy and ease that precipitation is measured.

Recognizing by 1943 that results from the vapor transport method would not be forthcoming for some time, it was necessary to return to the empirical methods of the 1931 classification and to re-examine the moisture index on which the classification was based in order to achieve practical solutions to such problems as the water requirements for irrigation. One can not determine the amount by which precipitation fails to supply water needs for crops without knowing what these water needs are.

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* The material in this chapter has been adapted from "The Water Balance in Arid and Semiarid Climates," by C. W. Thornthwaite, Desert Research, Proceedings International Symposium, Jerusalem, May 7-14, 1952, pp. 112-135; "Climatic Classification in Forestry," by C. W. Thornthwaite and F. Kenneth Hare, Unasylva, Vol. 9, No. 2, June 1955, pp. 50-59; "The Role of Evapotranspiration in Climate," by C. W. Thornthwaite and J. R. Mather, Archiv für Meteorologie, Geophysik und Bioclimatologie, Serie B, Band III, 1951, pp. 16-39; "A Re-examination of the Concept and Measurement of Potential Evapotranspiration," by C. W. Thornthwaite, in The Measurement of Potential Evapotranspiration, Mather (Ed.), Publications in Climatology, Vol. VII, No. 1, Seabrook, 1954, pp. 200-209; and "Investigation of Thornthwaite's Evapotranspiration Formula and Procedure," by J. R. Mather in Estimating Soil Tractionability from Climatic Data, Publications in Climatology, Vol. VII, No. 3, Centerton, 1954, pp. 379-384.

In order to make a map showing the distribution of water deficiency (the amount by which precipitation fails to supply sufficient water) it was first necessary to make a map of water need. This most important climatic element was 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 the vegetation. It was called potential evapotranspiration. It was computed by means of an empirical formula involving mean monthly temperature and relative humidity. In refining this formula to achieve more satisfactory results and to utilize only climatic data which are generally available it was possible to eliminate all factors but mean temperature and average length of day. That satisfactory results could be obtained without the use of wind, humidity, or solar radiation seems to be due to the fact that all of these important influences on evaporation including temperature vary together.

This new formula for determining potential evapotranspiration became the foundation of Thornthwaite's climatic classification of 1948. On the basis of the potential evapotranspiration concept advanced in that paper a more rational interpretation of the moisture factor in climate became available. Not only could the actual evapotranspiration and water deficit be determined but also the water surplus and runoff could be evaluated. This procedure has been described in a number of papers and is now well known, having served as the basis for studies made all over the world.

To justify the cardinal role applied to potential evapotranspiration, it will be necessary at this stage to digress upon the nature of evaporation over a land surface, since great confusion exists in the literature on this point.

Evaporation - the change in state of water from liquid to vapor - represents an important mass transfer from ground to atmosphere, the reverse of precipitation in the hydrologic cycle. But it is also an important agency of energy transfer, since vast amounts of heat are required to bring the evaporation about, and are then transferred to the air with the vapor as latent heat. Thus natural evaporation is much more than the reverse of rainfall; it is also a reverse flow to the downward stream of radiation from sun and atmosphere that warms the soil surface. A single parameter, measured evaporation or computed potential, hence at one time gives us some picture of two of the principal exchanges between earth and atmosphere.

Natural evaporation from the land whether evaporation from ponds, lakes and rivers, intercepted rainfall from the leaves and stems of plants, direct evaporation from the soil surface, or transpiration from green plant tissues, can proceed only when the vapor pressure of the ambient air is less than the vapor pressure at the evaporating surface, and can continue only while there is an external source of energy. The measurement of evaporation presents many difficulties.

It has been found that measured water surface evaporation or potential evapotranspiration varies inversely with the area of the evaporating surface with larger variations in dry climates and smaller ones in moist climates.

When a psychrometer is used to determine the humidity of the air a thermometer bulb is moistened and becomes an evaporating surface. The water evaporating from the wet bulb thermometer cools the bulb. The surface area of the bulb is small and the amount of water vaporized is very small. Heat flows into the water film on the bulb from the warmer surrounding air, and the evaporation process will reach equlibrium at a rate and at a wet bulb temperature such that the energy appropriated from the air is just sufficient to maintain the evaporation. Solar radiation contributes almost no energy to this process. The water from the wet bulb moistens the air but the amount is so minute that the effect on the moisture content of the air is completely negligible.

The Piche evaporimeter is also small having an evaporating surface of approximately 13 sq cm. The evaporation from the surface of a Piche evaporimeter is likewise incompetent to raise the humidity in the air to any significant degree. Here, most of the energy used in evaporation comes from the air.

The Weather Bureau Class A evaporation pan is 4 feet in diameter and on a summer day in a dry situation may evaporate 2 gallons of water. Solar radiation contributes an important share of the energy for evaporation; the amount depending on the turbidity of the water and on the albedo of the pan which varies greatly with type, age, and condition of the material used. Additional energy for evaporation is available from the air. The amount of water evaporated from the pan will do little to modify the moisture content of the air; but immediately over the water surface the humidity is raised, the moisture gradient reduced, and the evaporation impeded. The extent of this influence depends on the rate at which fresh air passes across the evaporating surface from outside.

If the area of the evaporating surface is large, the influence of the moisture condition of the air passing over the evaporating surface becomes small and solar radiation is the primary source of energy for evaporation. The influence of the evaporation on the atmospheric humidity is very important because this influence cannot fail to have a reciprocal effect on the evaporation. In moist air the temperature of the evaporating surface will rise to a point above the dew point of the air such that the evaporation will just use the energy that is available. Similarly, in dry air, rapid evaporation will lower the temperature of the evaporating surface until the evaporation is in accord with the available energy. The size of the water body or moist land area necessary to insure potential evapotranspiration is difficult to determine. The size of the area under the high moisture conditions has to be large enough so that the evapotranspiration from the area is not affected by external factors such as the advection of moist or dry air masses and their modification by local conditions. Water pans which vary from one another only in size are influenced differently by such conditions and, thus, result in different values of evaporation. Most land evaporation pans are certainly too small to be uninfluenced by external conditions and, in some cases, even soil pans surrounded by swamps and marshes appear to be too small. Expressed schematically the relation of the various pans to the larger, moist areas is shown in figure 1.1.

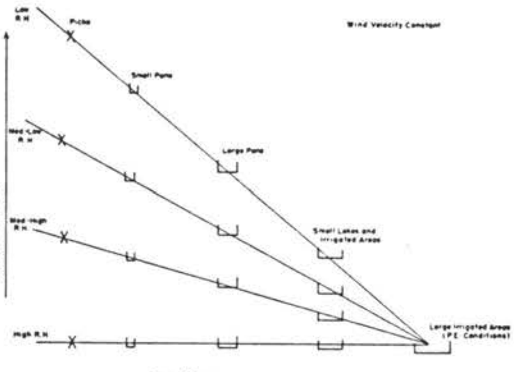

Figure 1.1

This diagram shows that the value of potential evapotranspiration is fairly independent of the size of the measuring pan when the humidity is high but, as the air becomes drier, the size of the pan greatly influences the rate of evaporation or evapotranspiration from it. Because of this limitation, pans or other instruments like the Piche evaporimeter do not give satisfactory measures of evapotranspiration except possibly under conditions of high humidity.

Because of the climatic significance of potential evapotranspiration, considerable effort has been directed toward the development of an instrument to measure this quantity. Soil tanks, covered with the same kind of vegetation that surrounds them and having a water supply fully adequate to the needs of the vegetation, have been used to measure evapotranspiration water losses. A simple, inexpensive vegetation-covered soil tank or evapotranspirometer came into use in 1945. It consists essentially of three parts; a field tank, a water supply and overflow apparatus, and a mechanism to control the water level in the field tank. This last part can be eliminated if daily sprinkling from above is used. The difference between the water added to the field tank by precipitation or supplied by aerial or subsurface watering and that lost from it to the overflow tank is the potential evapotranspiration. A number of articles on the installation and operation of evapotranspirometers have appeared.

In order to eliminate the moisture advection term from consideration, the evapotranspirometer tanks must be within a field which is planted to the same vegetation as is on the tanks and which receives the same watering treatment as the tanks themselves. The size of the buffer area needed will depend on the climate; in a moist climate such as Ireland, a square 50 meters on a side should be sufficient, but in the desert, probably a square 400 meters on a side would not be too large. The need for reliable observations in arid areas is very great but these are the areas where it is most difficult to establish and to maintain satisfactory conditions.

Since the rate of evapotranspiration is dependent on the moisture content of the soil, it becomes necessary to assure that the surface soil does not dry below "field capacity", because otherwise the ratio of evaporation heat loss to the total heat loss could not remain a constant.

Failure to maintain ideal conditions in an evapotranspirometer installation can result in either too much or too little water loss. If the soil in the tanks is allowed to become dry, the evapotranspiration rate will be less than the potential. If, on the other hand, the soil outside the tanks dries out, the evapotranspiration rate will be higher than the potential because of the advection of dry air. The condition of the vegetation is most important. If the vegetation in the tanks is higher than that outside, the water loss will be excessive. Even two or three centimeters will make a serious difference. If the vegetation in the tanks stands up prominently above the surroundings, the observations of water use are probably worthless and should be discarded. Tall growing crops will inevitably give excessive values of water use if they are grown in association with lower plants. For that reason, it is almost impossible to compare different kinds of vegetation in different adjacent tanks at the same time.

It will be difficult to maintain ideal conditions in an evapotranspirometer installation. Consequently, careful records of the conditions that actually prevail at all times should be made. Photographic notes probably best reveal the characteristics of the vegetation and should be made frequently.

From the foregoing discussion it should be clear that the water loss from ordinary evaporation pans or soil tanks can be very different from the true potential evapotranspiration. It should also be clear that there is no relation between potential evapotranspiration and expressions relating to the "evaporating power of the air" such as relative humidity or saturation deficit; they are not synonymous. The moisture content of the air is strongly influenced by the evaporation regime. Where the soil is dry and there is little evaporation, relative humidity is low and saturation deficit high. But if the conditions required for potential evapotranspiration are introduced, evaporation increases and more moisture is in the air. Thus, atmospheric moisture is not a conservative property of the air. Consequently, it is not possible to determine potential evapotranspiration by considering either relative humidity or saturation deficit. Similarly, pan evaporation is strongly influenced by the moisture content of the air, and so it is not possible to determine potential evapotranspiration from pan evaporation. The same criticism may be leveled at the evapotranspirometers unless they are operated under the proper conditions.

In dry climates or during periods of drought, pan evaporation is always higher than potential evapotranspiration. Any formula for determining potential evapotranspiration that contains a humidity term will also give excessive values in dry areas or dry periods. This discrepancy has resulted in the habitual use of too much water in irrigation, sometimes in grossly excessive amounts; wasting precious water and damaging valuable land at the same time.

Since 1946 batteries of evapotranspirometers have been operating in various places; two are in Nigeria within 8 degrees of the Equator and another on the Artic Circle in Canada. In all, records from more than twenty different evapotranspirometer installations have been collected by the Laboratory of Climatology which serves as a clearing house for the data. There is already a considerable literature on evapotranspirometry. From this increasing inventory of evaporation measurements, certain conclusions are permissible. Evapotranspiration clearly depends on:

(a) the external supply of energy to the evaporating surface, principally by solar radiation;
(b) the capacity of the air to remove the vapor, i.e., on wind speed, turbulent structure and the decrease of vapor concentration with height;
(c) the nature of the vegetation, especially as regards its capacity to reflect incident radiation, the extent to which it fully occupies the soil, and the depth of its root system;
(d) the nature of the soil, especially the amount of available water in the root zone.

Of these four, the meteorological controls (a) and (b) rank larger than the biotic and edaphic controls (c) and (d), a claim that will surprise many foresters. Actually (a) and (b) are closely related, since to a large extent the turbulent structure of the lower atmosphere depends on the same radiative and mass exchange processes that contribute the energy needed for evaporation. Similarly (c) and (d) are closely related.

There is no doubt that (a) is the master item in this list. Natural evaporation requires the latent heat of vaporization, and in practice this comes mainly from incoming solar radiation. Some of the radiation is reflected back from the surface, the percentage lost being known as the albedo. Some is used to heat the soil. Some goes to create convection within the air. The rest is used for evapotranspiration.

A word is necessary about interception, which is of considerable importance in forests. From the point of view of the climatologist it matters little whether the water evaporating from a plant comes from the soil via the root system, or is merely intercepted rain. Both processes require the same quantity of energy, and both constitute evapotranspiration. It is true, of course, that the intercepted water will not have figured in the physiology of the plant; but, supposing that soil moisture is not actually at the wilting point, the energy consumed in evaporating the intercepted water would otherwise have been used to evaporate transpired water. In other words, intercepted rainfall is not lost to the plant; it must afford relief to the drain on soil moisture.

Thus, to sum it up, we can fairly claim that evapotranspiration is the key process in the exchanges between earth and atmosphere.

Since it is not possible to obtain measurements of potential evapotranspiration everywhere, an alternative method of determining this quantity is necessary. In developing a formula which uses climatic factors for computing potential evapotranspiration a conservative climatic parameter must be used; one that will be relatively unaffected by the introduction of conditions of potential evapotranspiration. Atmospheric moisture is very sensitive to an increase of soil moisture necessary for potential evapotranspiration and is thus unsuitable. Maximum and minimum temperature are also both affected; maximum temperature is not as high over moist soil and minimum temperature is not as low. Thus diurnal range of temperature is not a conservative property either but would exhibit a reduction if the soil became moist. Since maximum and minimum daily temperature are affected in the opposite direction by changes in soil moisture, the mean is only slightly affected; mean daily temperature is one of the most conservative climatic elements. Temperature can serve as an index to potential evapotranspiration because there is a fixed relation between the net radiation used for heating and that used for evaporation when conditions exist to achieve the potential rate. However, since a higher proportion of the net radiation is spent on heating in arid areas, there is even with temperature a lack of conservatism which will result in some error in computing potential evapotranspiration.

The problem of developing a formula for potential evapotranspiration remains unsolved. The foregoing should help to explain, however, why there is a close relation between air temperature and potential evapotranspiration and why atmospheric humidity terms should be avoided.

Ultimately the energy used in the process of evapotranspiration of water from the ground and plant surfaces is derived from the solar radiation reaching the earth. The heat budget equation at the earth's surface -- the main region of transformation of energy - may be approximately written as:

$$R = S + C + W$$

where R is the net radiation during any period of time or more precisely the sum of the solar and sky radiation less what is lost by reflection and long-wave back radiation, S is the energy stored in the earth's surface in the form of heat, C is the energy going into heating the air by conduction-convection, and W is the energy used in the phase transformations of water or snow to vapor. Thus, there are three possible sources of energy for evapotranspiration: solar radiation, heat that reaches the evaporating surface from the air, and heat that is stored in the evaporating body. The latent heat of vaporization is large, ranging from 574 cal/cc at 40°C to 596 cal/cc at 0°C. If the heat needed for the evaporation of a small film of water came from the water itself, the water surface would be cooled well below the dew point. Thus, with no external source of energy, the surface temperature would quickly drop to the dew point of the air and evaporation would cease.

Schmidt, Ångström, and many others have long pointed out the advantages as well as the limitations of the energy balance approach. Ångström, concluded that the most difficult question was that of determining the ratio between the heat carried away by convection and the heat used for evaporation. He determined the ratio that applied to the water surface of a Swedish lake but did not suggest any means of getting the ratios for land areas.

In 1953, while participating in an ambitious Air Force field expedition to O'Neill, Nebraska, the Laboratory of Climatology obtained a very extensive series of micrometeorological measurements from which it is possible to compute the various components of the net radiation for a number of days. The computations which are included in table 1.1 show that when the soil is very moist more than 80 percent of the net radiation is used in evaporation. When the soil is dry evaporation is greatly reduced and most of the net radiation is devoted to heating the air with very little remaining for evaporation. Between these extremes, the proportion of the net radiation that is spent on evaporation varies in a manner which has not yet been completely determined.

Table 1.1

Heat Used for Convection, Evaporation, and Storage in Soil and Soil Moisture Content, O'Neill, Nebraska, 1953

Date	Heat Used for Convection (C) (cal/cm hr)	Heat Stored in Soil (S) (cal/cm hr)	Heat Used for Evaporation (E) (cal/cm hr)	Total C+S+E (cal/cm hr)	$\frac{E}{C+S+E}$ %	Soil Moisture in 0-18" Profile (in.)
Aug. 13,14	56.3	29.7	377.2	463.2	81	1.65
18,19	59.1	-4.8	287.8	324.1	84	1.40
22	98.4	19.0	216.2	333.6	65	1.20
25	181.9	41.5	131.8	355.2	37	1.05
31	242.3	28.3	44.5	315.1	14	.75
Sept. 3,4	121.1	-47.5	136.5	210.1	65	1.20

In the United States the heat budget method of determining evaporation has been little used because of the difficulty in obtaining the many necessary observations. Certain simplifying assumptions made it possible to compute evaporation from a lake or other large free water surface but no way was found to determine the evapotranspiration from a land surface where the rate of evaporation is dependent on the amount of water in the soil. This difficulty has been overcome in the past by introduction of the concept of potential evapotranspiration.

During those periods when the soil moisture falls below field capacity it would be a simple matter to determine the evaporation from areas with varying amounts of soil moisture if the percentage of the net radiation utilized in the vaporization of water was proportional to the moisture in the soil. Except for the fact that heat from the soil and from the air are additional sources of energy for evapotranspiration it would thus be possible to determine the potential evapotranspiration directly from the net radiation.

If one assumes that the energy for evapotranspiration comes only from solar radiation and that there is no effect of stored heat or advection of energy, it is possible to compute the water loss from any surface when the solar radiation, the albedo of the surface, and the soil moisture conditions are known. Potential evapotranspiration differs for different vegetation species because of differences in albedo. As the albedo increases more of the incoming solar radiation is reflected back to the sky and less remains for heating and for evaporation. Ångström has given the following albedos for different surfaces: grass, .26; oak woodland, .175; pine forest, .14; moist sand, .09. We have found that the albedos of most common vegetables are very similar to that of grass.

As an example the maximum evapotranspiration from grass, oak woodland, pine forest, and moist sand has been computed for Riverside, California, and Miami, Florida, for a day in June when the total solar and sky radiation is at a maximum. The data are presented in table 1.2. Also included in the table are monthly values assuming that the daily rate is continued for a 30-day period.

Table 1.2

Maximum Daily and Monthly Evapotranspiration (mm)

	Grass		Oakwoods		Pine Forest		Moist Sand	
	Day	Month	Day	Month	Day	Month	Day	Month
Riverside, California	4.8	144	5.6	168	5.8	174	6.3	189
Miami, Florida	4.6	138	5.1	153	5.3	159	5.6	168

Actually these values are too large because of the assumption that no heat is stored in the ground and the use of the maximum possible solar and sky radiation. Thus, any measured values of evapotranspiration which are greater than these must be suspect and probably result from unnatural exposure or operating procedures.

These observations give the correct order of magnitude of maximum rates of potential evapotranspiration. Such rough computations provide a basis for evaluating the data obtained from the evapotranspirometers. If, for example, the water loss from grass covered tanks in south Florida is greatly in excess of the above figures for Miami (138 mm for June), it must be concluded that the installation is inadequate in some respect; perhaps the vegetation within and without the tank is not uniform or perhaps the soil outside the tank is not as moist as that inside.

In order to define potential evapotranspiration, it will be necessary to specify a number of conditions. First, the albedo of the evaporating surface must be a standard. Second, the rate of evapotranspiration must not be influenced by the advection of moist or dry air. And finally, the ratio of the energy utilized in evaporation to that heating the air must remain essentially constant.

Fresh turgid green leaves of grass such as are ordinarily used in an evapotranspirometer have an albedo of about .25. Other herbs and shrubs also have an albedo in the same range. After experiencing drought and becoming wilted these same plants have a higher albedo, up to .30. Moist bare soil has a much lower albedo; .09 for moist sand. Thus the net radiation and the evapotranspiration from a bare moist soil surface will be higher than from a vegetation covered surface. For a standard evapotranspirometer installation a solid cover of vegetation is necessary and the albedo should not vary much from .25.

The measurement of both potential and actual evapotranspiration is an extremely complex undertaking. The evapotranspirometer, when operated properly, i.e., when watered sufficiently so that there is no moisture deficiency and no appreciable moisture surplus in the soil in the tank, and when exposed homogeneously within a protective buffer area of the proper size to eliminate the effect of moisture advection, is an instrument which should give reasonably reliable values of potential evapotranspiration. Great care must be taken in the operation of the instrument and standardized soil, vegetation, cultivation, and watering practices must be maintained on the tanks in order to insure realistic observations which are comparable from one installation to another.

CHAPTER 2. THE EVALUATION OF THE WATER BALANCE*

The work which has been completed on the energy balance method of computing potential evapotranspiration has led to a much clearer understanding of the problems involved. However, considerable work must still be done in determining the exact contribution of each of the three sources of energy for evaporation under different conditions of climate, soil structure, and moisture. Because of these unknown factors, and because observations of net radiation are very few and cannot yet be computed directly, it is still necessary to refer to the empirical formula developed by Thornthwaite for determining the potential evapotranspiration at a point.

Average values of monthly and annual potential evapotranspiration have been computed from available climatic data for some 13000 stations around the world including about 3500 Weather Bureau stations in the United States. The average annual water need in the United States ranges from less than 18 inches in the high mountains of the West to more than 60 inches in three isolated areas in the deserts of Arizona and southern California (figure 2.1). It is less than 21 inches along the Canadian border of eastern United States and more than 48 inches in Florida and southern Texas.

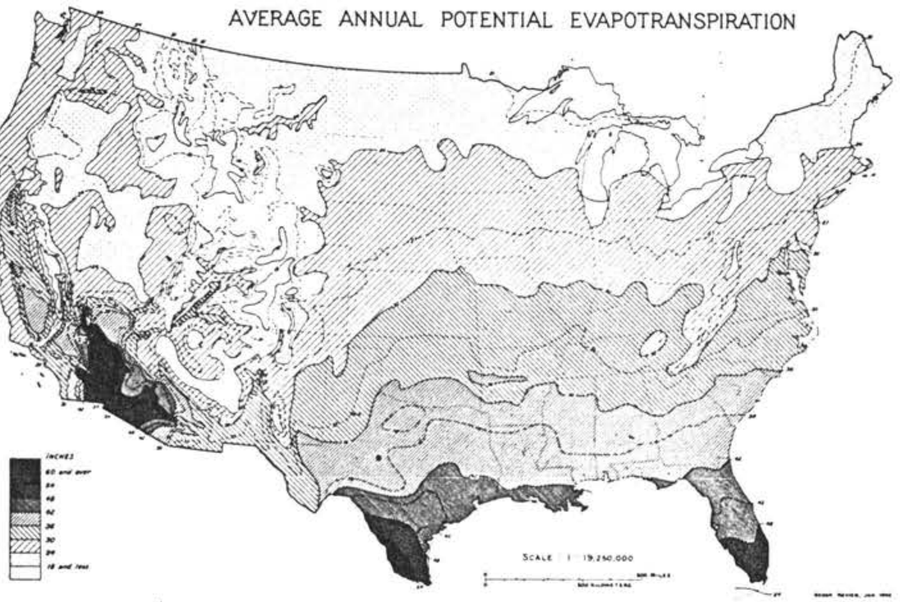

Figure 2.1

* The material in this chapter has been adapted from "The Computation of Soil Moisture," by C. W. Thornthwaite and J. R. Mather, in Estimating Soil Tractionability from Climatic Data, Publications in Climatology, Vol. VII, No. 3, Centerton, 1954, pp. 397-402; "Climatic Classification in Forestry," by C. W. Thornthwaite and F. Kenneth Hare, Unasylva, Vol. 9, No. 2, June 1955, pp. 50-59; "Modification of Rural Microclimates," by C. W. Thornthwaite, Background Paper No. 27, prepared for the Wenner-Gren Foundation International Symposium, "Man's Role in Changing the Face of the Earth," Princeton, June 16-22, 1955, 35 pp.; "The Water Balance in Arid and Semiarid Climates," by C. W. Thornthwaite, Desert Research, Proceedings International Symposium, Jerusalem, May 7-14, 1952, pp. 112-135; and "Climatology in Arid Zone Research," by C. W. Thornthwaite, Prepared for presentation at AAAS International Arid Lands Meetings, Albuquerque, April 26-May 4, 1955, 17 pp. (processed).

The march of potential evapotranspiration follows a uniform pattern through the year in most of the United States. It is negligible in the winter months as far south as the Gulf Coastal Plain and is only 2 inches a month in southern Florida. It rises to a maximum in July that ranges from 5 inches along the Canadian Border to 7 inches on the Gulf Coast. In some mountain areas and along the Pacific Coast it does not reach 5 inches in any month.

The march of precipitation is highly variable from one region to another. In much of the United States more than half the rain falls in the growing season. In the Pacific Coast states the distribution is reversed, with most of the rain falling in winter.

When the potential evapotranspiration is compared with the precipitation and allowance is made for the storage of water in the ground and its subsequent use, periods of moisture deficiency and excess are clearly revealed and an understanding of the relative moistness or aridity of a climate is obtained. In some stations precipitation is always more than the evapotranspiration so that the soil remains full of water and a water surplus $\underline{s}$ occurs. In other places, month after month, precipitation is less than potential evapotranspiration, there is not enough moisture for the vegetation to use and a moisture deficit $\underline{d}$ occurs. Stations with both wet and dry seasons, or with cold seasons of low water-need, normally show (1) a period of full storage, when precipitation exceeds water-need and a moisture surplus, $\underline{s}$, accumulates; (2) a drying season when stored soil moisture and precipitation are used in evapotranspiration, storage is steadily diminished, the actual evapotranspiration falls below the potential and a moisture deficiency, $\underline{d}$, occurs; and (3) a moistening season when precipitation again exceeds water-need and soil moisture is recharged. The values of $\underline{s}$ and $\underline{d}$ can be computed during the budgeting.

Although results obtained through the use of the water balance procedure developed in 1945 were quite satisfactory several of the assumptions used in the bookkeeping procedure were only approximations. Thus, recent work has been directed toward the improvement of these assumptions and the inclusion in the bookkeeping procedure of certain other factors which seem to be of great importance. This work has resulted in the revision of the bookkeeping procedure and the methods of computation so that it now provides more realistic values of the various moisture parameters from different soil types.

The moisture holding capacity of a soil depends on the depth of the soil layer considered, and the type and structure of soil. It can vary from just a few millimeters on a shallow sand to well over 400 mm on a deep well-aerated silt loam. The roots of plants compensate somewhat for the variable nature of soil for on sandy soils plants will be deep-rooted while on silts and clay the plants tend to be more shallow-rooted. Thus, the depth of water available to the roots of mature plants is not as variable as might be thought, at first. Of course, young plants or mature trees will have root systems which ramify through a markedly different depth of soil and so they will have available to them quite different amounts of moisture. Moderate or deeply rooting crops growing in humid areas utilize only about 100 mm of soil moisture between rains. The original bookkeeping procedure reflected this observation in the assumption of a 10 cm storage capacity of the soil.

As soil dries it becomes increasingly difficult for additional water to be lost by evaporation and transpiration. Thus, as the soil moisture content decreases so also does the rate of evapotranspiration. Studies have indicated that the rate of evapotranspiration is proportional to the amount of water remaining in the soil. For instance, when the soil moisture content is one-quarter of capacity the rate of evapotranspiration will be one-quarter of the potential rate.

From the above discussion, it can be seen that when the soil moisture content is near field capacity the rate of evapotranspiration will approximate the potential rate. Again the original bookkeeping procedure reflected this relation for it was assumed that as the 10 cm of available moisture was being lost by evapotranspiration the rate of evapotranspiration was equal to the potential value. After the 10 cm had been lost the evapotranspiration rate dropped to zero except after periods of rain when it was again equal to the potential rate.

Recent advances in our knowledge have permitted the preparation of tables which give the rate of evapotranspiration at any selected moisture content from soils with different values of total moisture holding capacity. Thus, it becomes a relatively simple matter to take account of the factors of variable moisture capacity and evapotranspiration rate in the water balance computations and to eliminate the objectionable features of the earlier assumptions. Actually, quite comparable values of actual evapotranspiration, water surplus and deficit are obtained using either the original 10 cm assumption or the new procedure of a variable storage capacity and evapotranspiration rate as would be expected from a consideration of the assumptions themselves. The new procedure is preferable, however, since it is more realistic than the older empirical one and depicts more exactly the processes going on in nature.

In computing the depth of water in a soil column, the gravitational water or moisture above field capacity and the capillary water or moisture below field capacity must be considered separately. At field capacity the soil contains no surplus of gravitational water and no deficit of capillary water. Thus, this value becomes an important and useful point in the computational process.

When the moisture content of the soil is at field capacity or above, the surplus water or water that is added by precipitation is lost slowly by downward percolation regardless of whether there is evapotranspiration or not. This gravitational water is only detained briefly, the period depending on the depth and permeability of the soil and the amount of gravitational water. It has been empirically found that in a 40 inch thickness of loan about 90 percent of the gravitational water in the soil on any given day is held over in the soil until the succeeding day. This percentage becomes smaller as the thickness of the soil layer decreases. Also the greater the amount of sand in the soil the smaller is the percentage of gravitational water held over from one day to the next. Thus, in computing the drying of a soil from an initial value of moisture above field capacity, it is necessary to determine separately both the loss of water by evapotranspiration and by gravitational flow.

There are many occasions such as in the preparation of general moisture maps or in the application of the results to strategic problems in tractionability when it is neither necessary nor desirable to make detailed daily computations of soil moisture or to use the specific hydropedologic data that would be required in determining the water balance on a daily basis. In these cases certain of the procedures employed in making the daily determinations can be generalized. For instance, in making monthly rather than daily computations it is possible to consider surface runoff and the percolation of gravitational water as one quantity, runoff. It is not necessary to employ any daily factor for the detention of gravitational water in the profile but merely a general monthly detention factor for all runoff, a factor which varies with the size of the watersheds and which for large watersheds is about 50 percent.

As an example of the results obtained from the water balance computation the monthly march of precipitation and of potential evapotranspiration have been compared for Berkeley, California and Seabrook, New Jersey (figure 2.2). The actual water balance computations for these stations are given in table 2.1.

In working out the actual steps involved in the water balance computations it is necessary to have the data of precipitation and potential evapotranspiration at the station in question plus the necessary tables which permit conversion of the information on potential water loss from a given depth of soil to values of actual water loss under conditions of a varying soil moisture content. Subtracting the potential evapotranspiration from the precipitation results in a series of positive and negative differences which represent potential additions and losses to the soil moisture storage. The negative values of the differences which indicate a potential loss of water from the soil now have to be converted into values of actual change in soil moisture storage due to the fact that as the soil dries water is lost from the soil at a rate somewhat less than the potential rate. At Berkeley, California, in June potential evapotranspiration is 79 mm greater than the precipitation. While this is the potential loss of water from the soil actually only 56 mm of moisture are removed from the soil. Evapotranspiration cannot go on at the potential rate for the soil moisture content does not remain at the optimum for evapotranspiration. Simple tables can be prepared which permit these computations to be carried out in a straightforward manner.

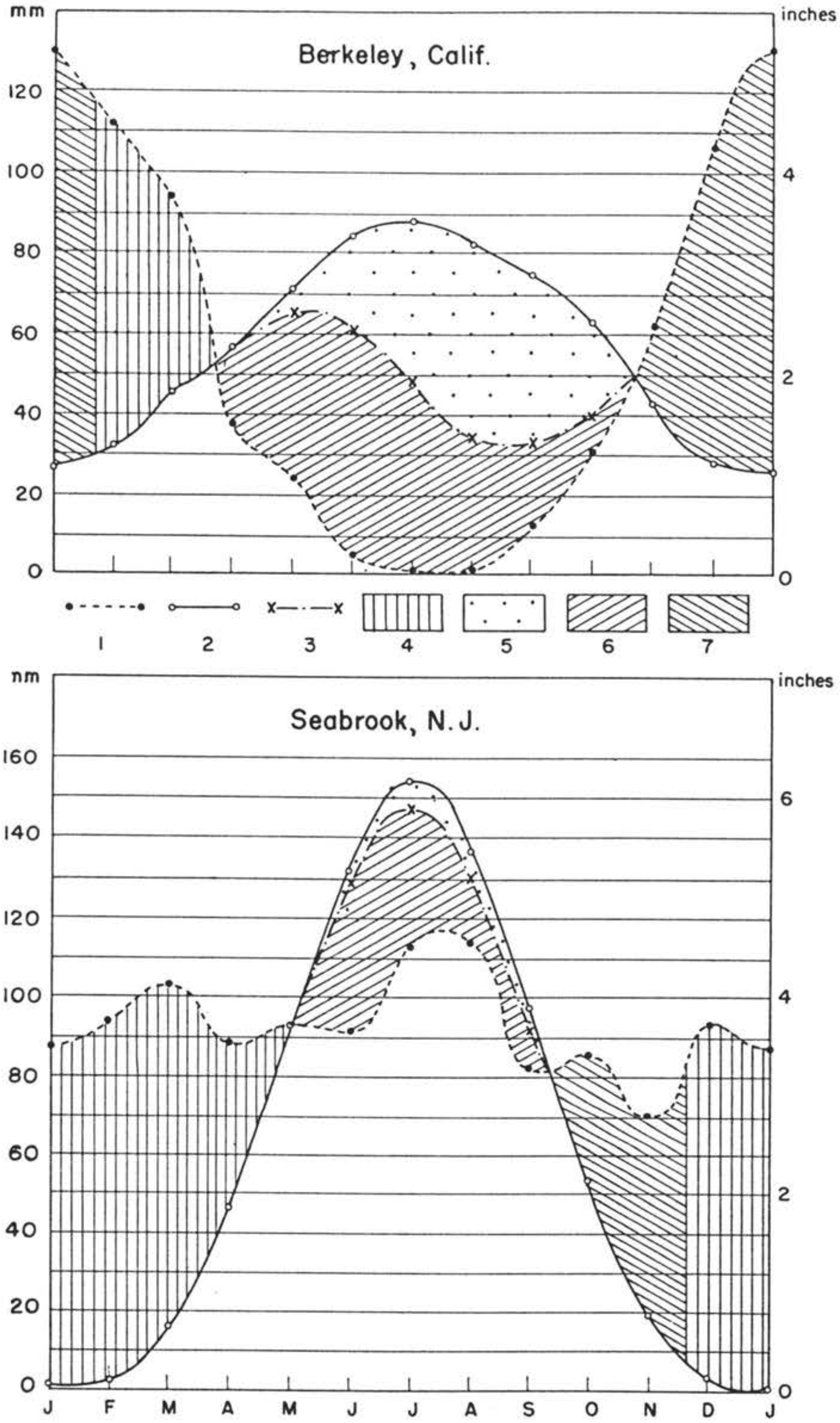

Figure 2.2

Average march of precipitation (1), potential evapotranspiration (2), and actual evapotranspiration (3) through the year at Berkeley, California and Seabrook, New Jersey. Diagram also shows periods of water surplus (4), water deficit (5), soil moisture utilization (6), and soil water recharge (7).

Table 2.1

Water Balance Computations for Berkeley, California and Seabrook, New Jersey.
300 mm depth of water stored in soil layer at field capacity.
(all values in mm)

Berkeley, California

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Potential Evapo.	26	32	45	56	71	84	88	82	75	63	43	28	693
Precipitation	130	112	94	37	24	5	1	1	13	31	62	106	616
Difference	104	80	49	-19	-47	-79	-87	-81	-62	-32	19	78	-77
Storage Change	104	22	0	-19	-41	-56	-47	-33	-19	-8	19	78	
Moisture Storage	278	300	300	281	240	184	137	104	85	77	96	174	
Actual Evapo.	26	32	45	56	65	61	48	34	32	39	43	28	509
Moisture Deficit	0	0	0	0	6	23	40	48	43	24	0	0	184
Moisture Surplus	0	58	49	0	0	0	0	0	0	0	0	0	107
Runoff*	0	29	39	19	10	5	3	1	1	0	0	0	107
Moisture Detention	278	329	339	300	250	189	140	105	86	77	96	174	

Seabrook, New Jersey

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Potential Evapo.	1	2	16	46	92	131	154	136	97	53	19	3	750
Precipitation	87	93	102	88	92	91	112	113	82	85	70	93	1108
Difference	86	91	86	42	0	-40	-42	-23	-15	32	51	90	358
Storage Change	0	0	0	0	0	-38	-35	-17	-10	32	51	17	
Moisture Storage	300	300	300	300	300	262	227	210	200	232	283	300	
Actual Evapo.	1	2	16	46	92	129	147	130	92	53	19	3	730
Moisture Deficit	0	0	0	0	0	2	7	6	5	0	0	0	20
Moisture Surplus	86	91	86	42	0	0	0	0	0	0	0	73	378
Runoff*	61	76	81	61	31	15	8	4	2	1	1	37	378
Moisture Detention	361	376	381	361	331	277	235	214	202	233	284	337	

* Assuming 50 percent of the water available for runoff in any month is held over until the following month.

When the precipitation is greater than the potential evapotranspiration the actual evapotranspiration equals the potential for at those times there is sufficient moisture in the soil so that evapotranspiration can proceed unhindered. When the precipitation is less than the potential evapotranspiration, the actual evapotranspiration equals the precipitation plus any moisture stored in the ground which is evaporated or transpired (the storage change). Moisture deficit and surplus follow simply from the bookkeeping calculations, the former being the difference between potential and actual evapotranspiration while the latter is the excess precipitation which occurs when the moisture holding capacity of the soil layer under consideration is full of water.

The moisture surplus is the water which is available for runoff into streams, rivers and lakes. On a moderate to large watershed there is considerable lag between the time the excess moisture falls as precipitation and the time it appears as runoff in streams. Thus it is necessary to consider the size of the watershed and the time period for which the water balance is being evaluated in determining the lag factor to be employed in the determination of runoff.

The moisture detention is the total of the moisture stored within the soil, the snow remaining on the soil surface, and the surplus water in the process of running off which has been detained temporarily. This value can rise considerably higher than the total water holding capacity of the soil for it includes both the so called gravitational water, the water above the field capacity of the soil and the moisture available for later surface or stream runoff. The water balance computations provide a good insight into the moisture relation of an area. For instance, at Seabrook, New Jersey, the potential evapotranspiration is negligibly small in winter but in early spring it begins a rapid rise which reaches the high point of the year of more than 150 mm in July. It falls rapidly during the autumn months. The corresponding precipitation is far more uniformly distributed through the year, being very close to 90 mm in nine of the twelve months. The rainiest months are July and August, each of which receive more than 110 mm; November, the driest month, has only 70 mm.

Rainfall and water need are not the same. At the time of maximum rainfall in July and August there is a water deficiency, whereas in November, when rainfall drops to the lowest value of the year, there is a water surplus. In early autumn when the water need is less than the precipitation the surplus rainfall first replaces soil moisture that had been used up previously and then raises ground water levels and produces surface and subsurface runoff. This surplus amounts to 378 mm at Seabrook. In spring, when the soil moisture is at field capacity actual and potential evapotranspiration are the same and all precipitation in excess of the potential evapotranspiration is realized as water surplus. When precipitation is less than potential evapotranspiration, the difference is made up in part from soil moisture storage; but as the soil becomes drier the part not made up is larger. This is the water deficit which amounts to 20 mm on the average at Seabrook, New Jersey.

In Berkeley, California, in a different climatic zone, nearly all of the rainfall comes in winter and there is almost no rain in summer. Here the winter water surplus is 107 mm and the summer water deficit is 184 mm.

A comparison of the water balance for Seabrook and Berkeley reveals some interesting facts. Both places have water surpluses and deficits during the year. The surplus at Seabrook is considerably greater than at Berkeley however. In addition, the net water balance shows an annual surplus of 358 mm at Seabrook and -77 mm at Berkeley. Thus at Seabrook and in other areas with similar water balances there is a large supply of readily available water which may be stored in the water table beneath the earth's surface - a supply which can be used for widespread irrigation and which will be replenished each year. On the other hand, at Berkeley water taken from the soil water reservoir for irrigation is not all replaced and there would be a year to year decline in the surface water table. Full irrigation of all land in such areas would not be possible. These two stations are illustrative of two different situations; in one area widespread climatic changes can be brought about through irrigation, but in the other, the influence of irrigations is of only local significance.

The use of actual values of potential evapotranspiration and precipitation for individual days reveals that certain errors are involved when individual monthly mean values are employed. It is not unusual for the rainfall to be distributed within a month in such a way that both deficiency and surplus occur in succession. Computations of the surplus and deficiency based on mean values conceal the true extent of these important quantities. An example of such an occurrence is shown in table 2.2. At Seabrook, New Jersey, in September, 1950, the total monthly rainfall was 253 mm while the monthly potential evapotranspiration was 81 mm. Precipitation exceeded water need by 172 mm. Since a total of 200 mm depth of water would fill the storage capacity of the soil layer under consideration and since 75 mm was stored in the soil at the beginning of the month, a total of 125 mm of water could go into storage during the month. The actual surplus computed on the basis of monthly figures then would be 47 mm and the deficiency 0. Based on daily computations of evapotranspiration and measured precipitation the water surplus for the month is seen to be 77 mm and the deficiency 19 mm. Thus a more accurate picture of the climate is revealed through the use of daily meteorological data.

When the water surplus or water deficiency is large the underestimation resulting from the use of means of potential evapotranspiration and precipitation is small and unimportant. We may take as an example a station in which the average precipitation of a month is greatly in excess of the evapotranspiration and the water surplus is large. The rainfall of the month varies from one year to another but if the lowest amount is still greater than the evapotranspiration it would make little difference by which method the water surplus is computed.

Table 2.2

Daily Soil Moisture Balance, September 1950, Seabrook, New Jersey.
Soil holds 200 mm of water at field capacity and has 75 mm at beginning of month. Ninety percent of available gravitational water held for percolation on each succeeding day.
(all values in mm)

Date	Adj. PE	Prec.	P-PE	Soil Moist. St.	Act. St. Ch.	Moist. Def.	Moist. Surp.	Avail. Grav. Water	Grav. Water St.	Soil Water Bal.
1	5		-5	73	2	3				73
2	5	1	-4	71	2	2				71
3	4	5	+1	72	+1	0				72
4	2		-2	71	1	1				71
5	3		-3	70	1	2				70
6	3		-3	69	1	2				69
7	3		-3	68	1	2				68
8	4		-4	66	2	2				66
9	4		-4	65	1	3				65
10	3		-3	64	1	2				64
11	3	169	+166	200	+136	0	30	30	27	227
12	2		-2	198	2	0		27	24	222
13	3	4	+1	199	+1	0		24	22	221
14	3	38	+35	200	+1	0	34	56	50	250
15	3		-3	197	3	0		50	45	242
16	3		-3	194	3	0		45	40	234
17	2		-1	193	1	0		40	36	229
18	2		-2	191	2	0		36	32	223
19	3	1	-2	190	2	0		32	29	219
20	3		-3	187	3	0		29	26	213
21	3	1	-2	185	2	0		26	23	208
22	2	29	+27	200	+15	0	12	35	32	232
23	2	3	+1	200	0	0	1	33	30	230
24	1	1	0	200	0	0		30	27	227
25	1		-1	199	1	0		27	24	223
26	1		-1	198	1	0		24	22	220
27	2		-2	196	2	0		22	20	216
28	2		-2	194	2	0		20	18	212
29	2		-2	192	2	0		18	16	208
30	3	1	-2	190	2	0		16	14	204
Total	81	253				19	77			

On the other hand, in subhumid and semiarid regions where both water deficiencies and surpluses can occur in a given month over the years the use of average annual values will give erroneous values for certain elements of the water budget. The Missouri Valley provides a good illustration. The Missouri river rises in the northern Rocky Mountains, flows across the semiarid Northern Great Plains, the subhumid prairies and enters the Mississippi near the dry margin of the humid climate. The water balance maps which have been prepared for this region include mean annual potential evapotranspiration or water need (figure 2.3), mean annual precipitation (figure 2.4), mean annual water deficiency (figure 2.5), and mean annual water surplus or runoff (figure 2.6). At the lower end of the valley, average annual water need is 32 inches and precipitation is 42 inches. Both water need and precipitation diminish to the north and west but precipitation drops off more rapidly so that in most of the valley precipitation does not equal water need. The water deficiency is less than 3 inches along the eastern margin but rises to more than 18 inches in some of the mountain valleys. The water surplus is more than 12 inches in the lowest parts of the valley and at high altitudes in the mountains but in most of the area it is less than 3 inches.

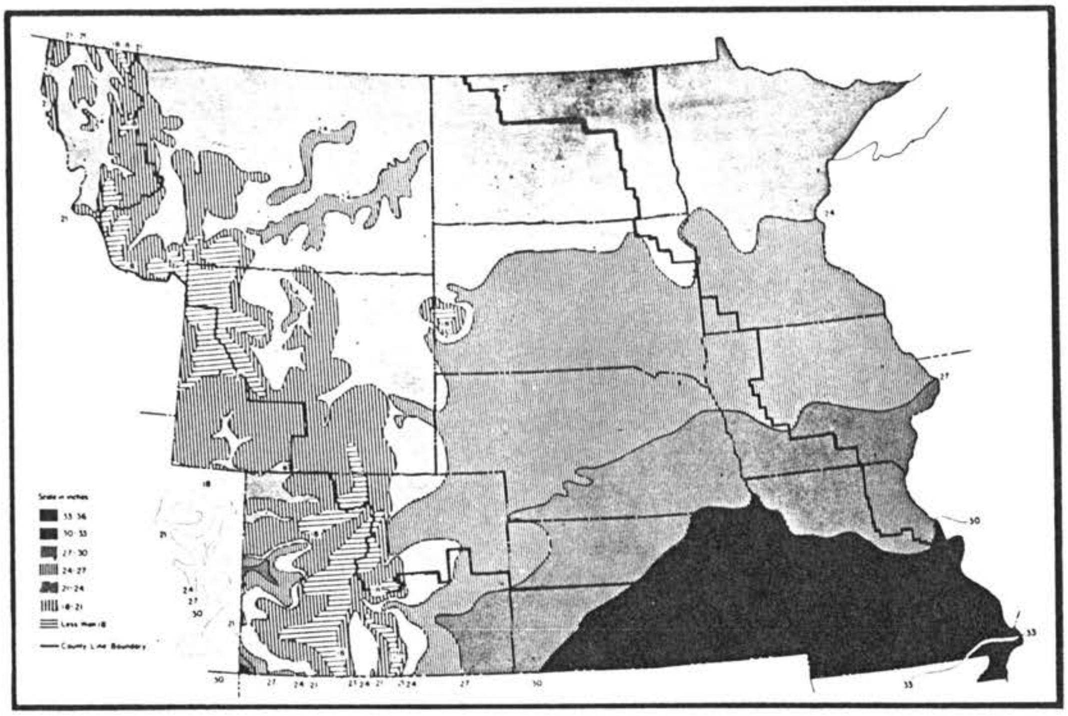

Figure 2.3 Average annual water need in the Missouri Valley, in inches.

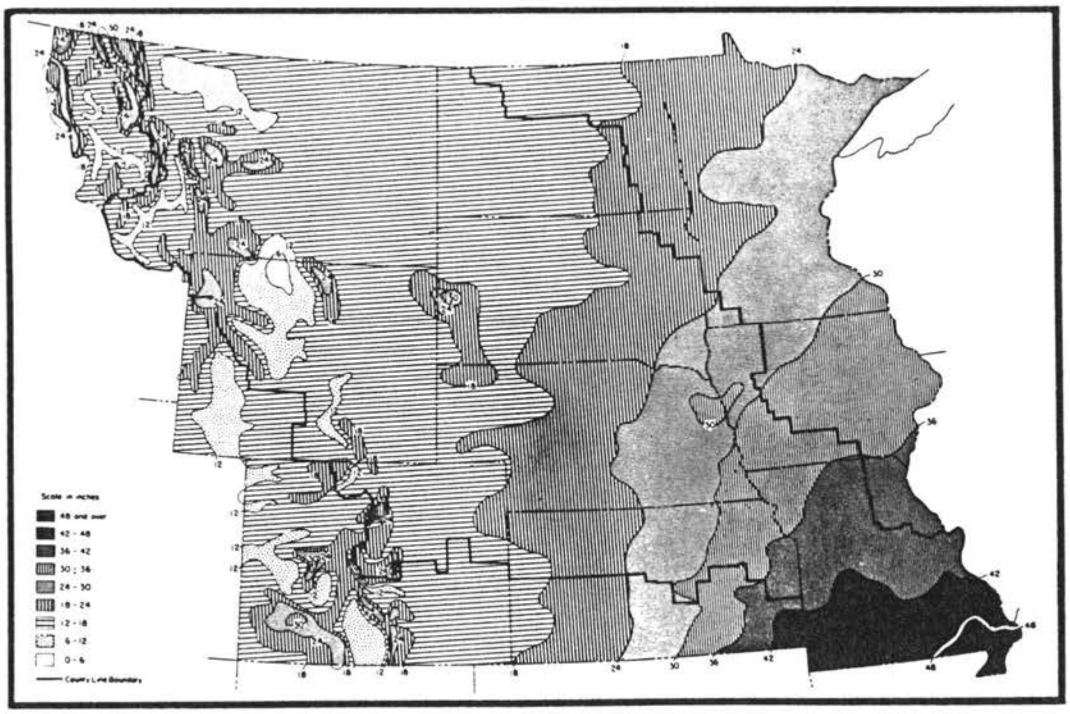

Figure 2.4 Average annual precipitation in the Missouri Valley, in inches.

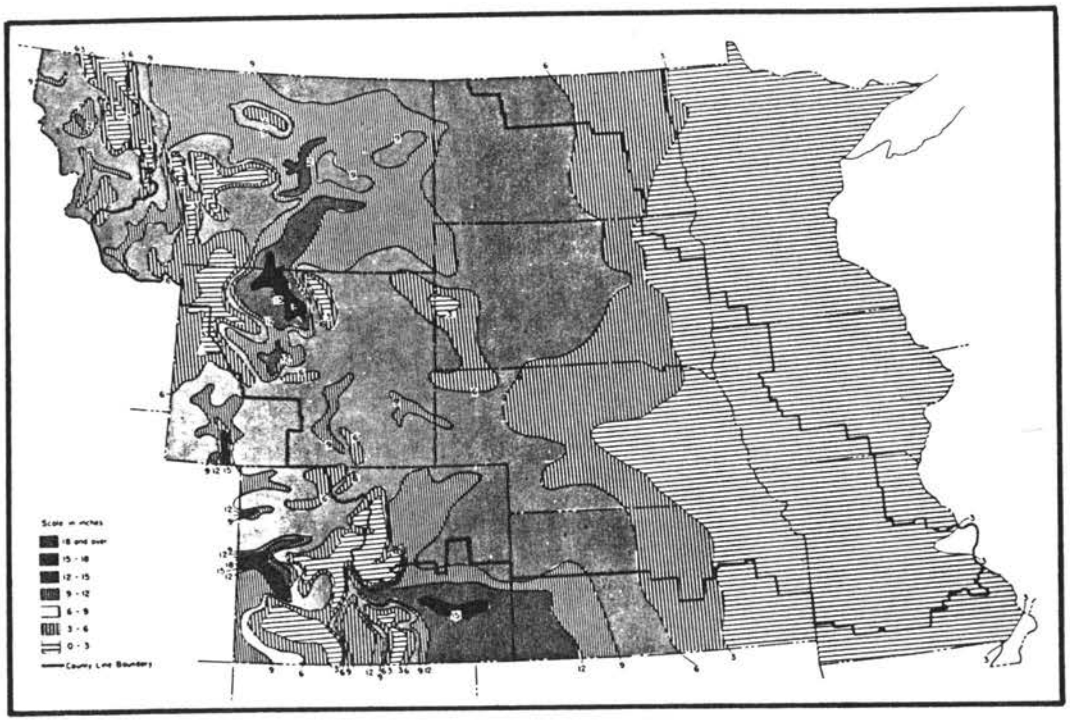

Figure 2.5 Average annual water deficiency in the Missouri Valley, in inches.

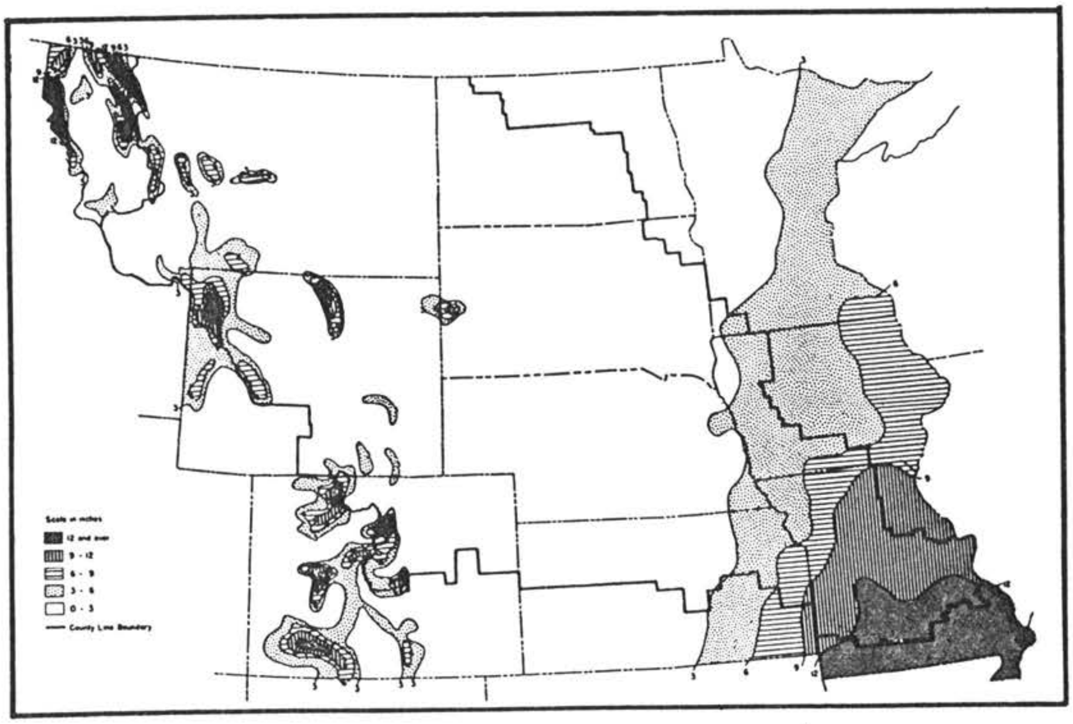

Figure 2.6 Average annual water surplus in the Missouri Valley, in inches.

In making the computations used in these maps long-time means of temperature and precipitation were used. For a number of selected stations in the Missouri Valley, the water balance for the individual years of the 25-year period, 1920 to 1944 has also been computed. Water deficiency and water surplus in the individual years at these stations are presented in figures 2.7 and 2.8. Both elements exhibit great variation from one year to another.

When surplus and deficiency are obtained by averaging the 25 individual yearly determinations the values are larger than when based on mean temperature and precipitation. The water surpluses and water deficiencies computed both ways for these various stations are compared in table 2.3.

Table 2.3

Comparison of Mean and Yearly Computations of Water Budget
(Data 1920-1944)

In Inches

Station	Average Annual Water Surplus		Average Annual Water Deficiency	
	Means	Average of 25 individual years	Means	Average of 25 individual years
Montana				
Havre	0	.25	9.61	12.43
Helena	0	.39	9.01	12.31
Miles City	0	.56	11.70	14.22
Wyoming				
Cheyenne	0	1.71	7.24	8.07
Sheridan	.63	2.17	8.27	10.57
Colorado				
Ft. Collins	0	1.13	8.43	11.74
Garnett	0	.03	13.19	13.43
Grand Junction	0	.12	20.04	20.49
Las Animas	0	.15	15.43	18.22
North Dakota				
Bismarck	0	.16	7.01	10.02
Devil's Lake	0	.83	4.14	5.97
Williston	0	.28	8.00	9.69
South Dakota				
Huron	.04	1.39	4.10	10.27
Pierre	0	.67	9.61	12.77
Rapid City	0	1.70	5.79	10.12
Nebraska				
Lincoln	1.34	2.51	1.93	8.31
North Platte	0	.72	7.68	11.93
Valentine	.16	1.41	5.47	9.97
Kansas				
Dodge City	0	1.15	9.13	12.23
Hayes	0	1.39	6.46	10.93
Manhattan	2.28	5.01	1.61	6.58
Minnesota				
St. Paul	3.19	5.25	.79	5.03
Worthington	3.15	4.45	0	4.24
Iowa				
Ames	4.37	7.17	.47	3.92
Charles City	6.61	8.96	0	3.27
Clarinda	4.49	6.44	.04	4.29
Missouri				
Columbia	9.49	12.35	.94	5.10
Kansas City	7.28	8.71	.59	6.30
Springfield	11.81	13.79	0	4.82

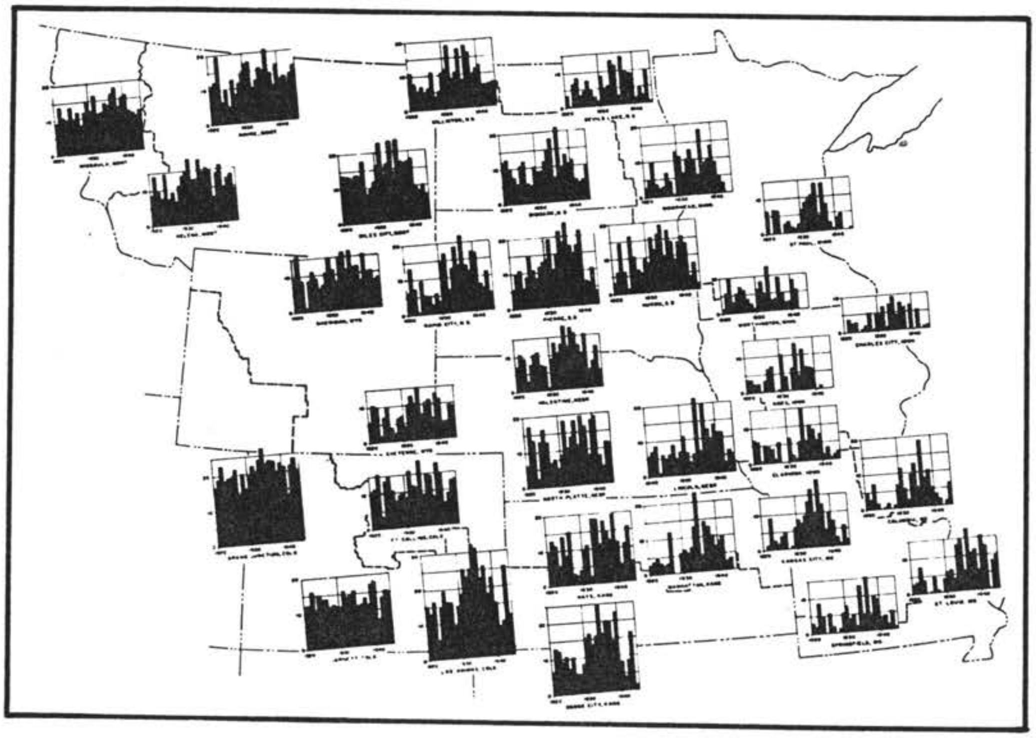

Figure 2.7 Water deficiency in the Missouri Valley, in inches.

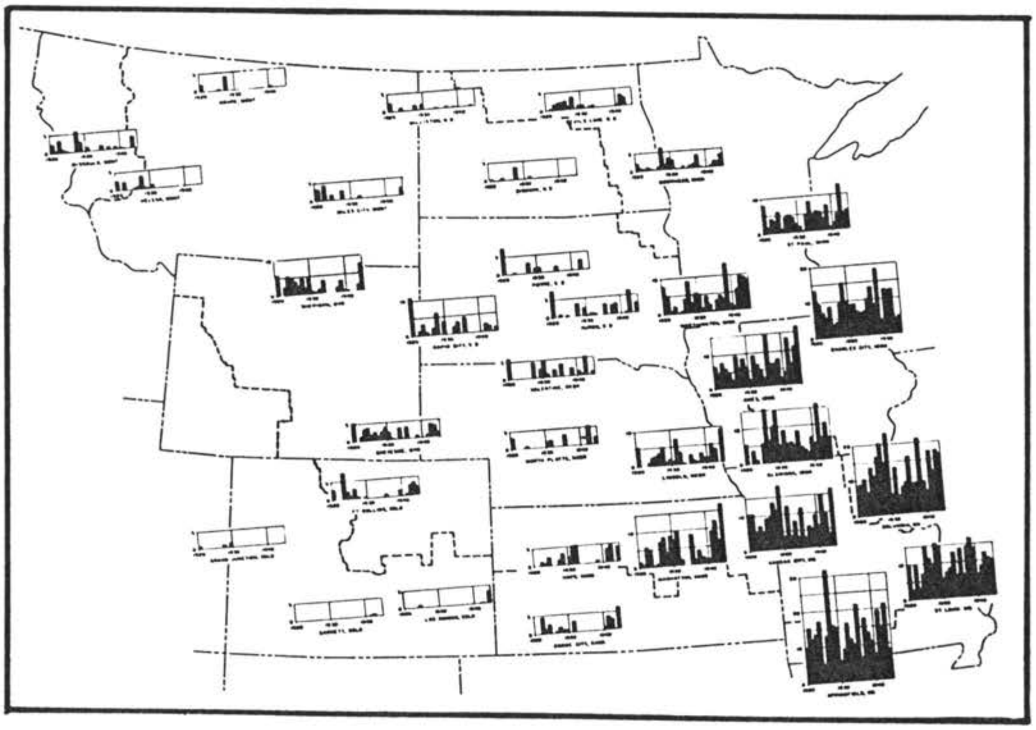

Figure 2.8 Water surplus in the Missouri Valley, in inches.

Use of the mean monthly data results in an underestimation of both water surplus and deficiency at all of the stations within the subhumid and semiarid parts of this region. Consequently, the maps of water surplus and water deficit based on average annual values (figures 2.5, 2.6) do not reveal fully the extent of the water problems. At Manhattan, Kansas, for example, mean values show that the annual surplus of water is 2.28 inches and the deficiency is 1.61 inches. One might conclude that since the surplus in some seasons is greater than the deficiency in others, more than enough water would be available to counteract the period of deficiency if suitable methods of storing the surplus water were developed. Actually, from computations based on data from the individual years of the 1920-1944 period, the average annual water surplus is found to be 5.01 inches but the annual water deficiency is 6.58 inches. On the basis of the data for individual years it is seen that insufficient water is available in the area to supply the irrigation needs. The actual amounts of available water surplus and deficiency at the other stations shown in table 2.3 also change considerably with the use of data for individual years.

The water balance in arid and semiarid regions is, as it is in even the humid regions, of critical importance in solving many of the complex moisture problems of a region. Because we are now able to set up such a balance through the use of the concept of potential evapotranspiration we can obtain a new understanding of problems which are fundamental to many fields of investigation.

PART II

APPLICATION OF THE FACTORS OF THE WATER BALANCE

CHAPTER 3. THE MOISTURE DEFICIT*

Every farmer knows that weather and climate are both important factors in crop production. They also realize that moisture deficiency in summer is the major deterrent to peak yields. Irrigation is the solution to this problem but scientific irrigation involves more than installing equipment and turning on the water.

If a farmer proposes to supply supplementary water to his crops he must have some practical means of determining how much water to use and when it is needed. Common practice among farmers is to watch the plants for signs of moisture deficiency as a basis for supplying water. This is, of course, not satisfactory for by the time the plants begin to show some signs of water need they are already suffering and the yield has been reduced correspondingly. Instead of watching the crop for indications of drought, some investigators have suggested watching the soil. One investigator has stated that the only known way to be sure that soil moisture is present in readily available form is by frequent examination of the subsoil by the use of a soil auger or similar tool. In recent years several devices have been developed to be installed permanently in the soil to give a continuous indication of the amount of moisture remaining. Among these devices are elements made of gypsum, fiber glass and nylon in which the electrical resistance varies with moisture. Many of these blocks have been used in some of the large irrigation enterprises to determine the time to apply water. Unfortunately, none of these devices gives the information that is needed.

The climatological approach is entirely different. The moisture in the soil is regarded as being a balance between what enters it as a result of precipitation and what leaves through evaporation and transpiration. An irrigation schedule is a natural outgrowth of this method of computing soil moisture. One can set up limits below which the soil moisture will not be allowed to fall for the particular crop and depth of root zone in question. Then by keeping daily account of how much water has been lost from the soil it is possible to know exactly when the predetermined level of soil moisture depletion is reached and to know just how much to irrigate to bring the moisture level back to a safe value. Shallow-rooted crops will have to be irrigated more frequently but with smaller amounts of water than will deeper-rooted pastures or orchards.

Figure 3.1 illustrates an irrigation schedule for three different types of crops growing at Seabrook, New Jersey in 1954. During the period illustrated, shallow-rooted crops had to be irrigated five times with a total of 3.75 inches of water while pastures and moderate-rooted crops were irrigated twice with a total of 3.00 inches of water. Deep-rooted orchards were irrigated once with a total of 2.25 inches of water. If irrigation is scheduled by keeping continuous account of the soil moisture, no great moisture deficiency can develop in the soil to limit growth and there will be no over-irrigation to damage both soil and crop and to result in a wasteful misuse of water.

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* The material in this chapter has been adapted from "The Water Budget and its Use in Irrigation," by C. W. Thornthwaite and J. R. Mather, in Water, Yearbook of Agriculture, 1955, pp. 346-358; "Agricultural Climatology, Weona, Arkansas," C. W. Thornthwaite Associates, Centerton 1955, 21 pp. (processed) (with permission of Mr. Jay Tschudy, Jr., President, Weona Farms Company); and "The Place of Supplemental Irrigation in Postwar Planning," by C. W. Thornthwaite, Publications in Climatology, Vol. VI, No. 2, Seabrook, N. J. 1953, pp. 11-29.

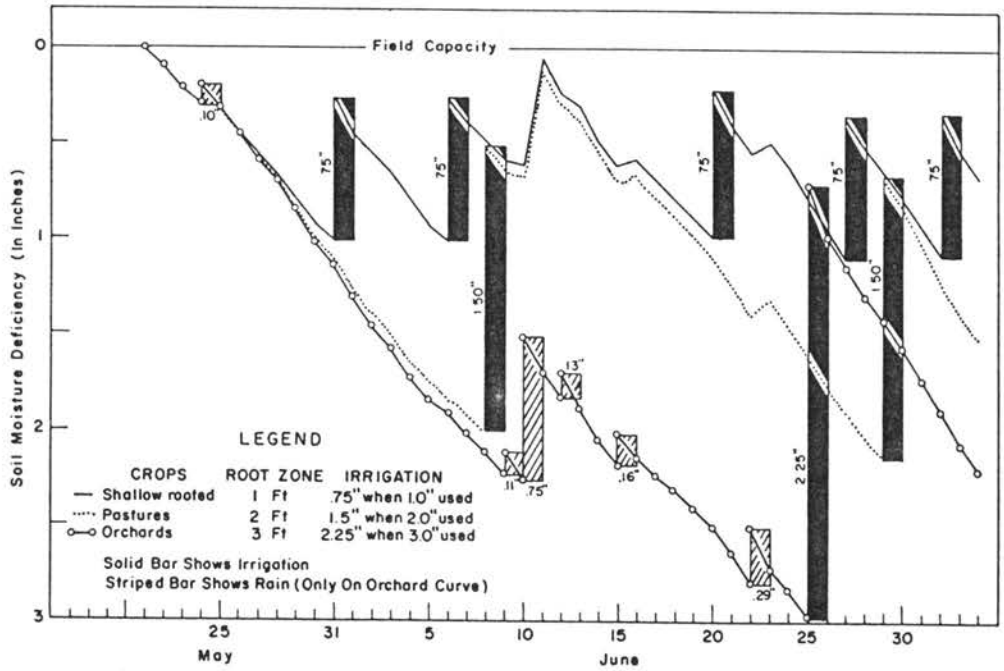

Figure 3.1 - Irrigation program for three different types of crops, Seabrook, New Jersey, 1954. Soil holds 3 inches of water per foot depth of soil at field capacity. Rain on May 21 brought soil moisture content to field capacity.

The purpose of irrigation is to counteract drought by making certain that the plants are not deprived of water at any time during the course of their development. Drought is the greatest natural hazard to agriculture. Everyone knows that drought makes farming difficult or impossible in the semiarid and arid regions of the earth. But few realize that even in the humid and subhumid regions, such as eastern United States and western Europe, droughts are frequent and severe.

There are four different kinds of drought. The first, which we may call permanent drought, is characteristic of the driest climates. The sparse vegetation is adapted to drought, and agriculture is impossible except by irrigating through the entire crop season. The second, or seasonal drought, is found in the climates that have well-defined rainy and dry seasons. The natural vegetation is made up of plants that produce seeds during the rainy season and then die, and of plants that remain alive but become dormant in the dry season. For successful agriculture, planting dates must be adjusted so that the crop develops during the rainy season. Otherwise, crops must be irrigated during the dry season. The third kind of drought results from the fact that rainfall is irregular and variable everywhere. These droughts depend upon the irregularity of rainfall and thus are not certain to occur in any definite season, but they are most probable at the time of maximum water need. We may call them contingent droughts. They may occur almost anywhere, even in the areas of seasonal drought, but are most characteristic of subhumid and humid climates. The fourth type of drought may be called invisible drought. It can occur in any area but is most common in the more moist climatic regions. Invisible drought can occur even during a period when there is rainfall on every day. When rains do not supply enough water to counteract the water loss by evapotranspiration the result is a borderline water deficiency which places a limitation on the growth of plants and reduces their yield in some cases to less than 50 percent of that possible. Invisible drought is not easy to recognize. Rainfall appears adequate and plants seem to be growing well yet because of a moisture deficiency they are unable to produce as luxuriantly as they would under optimum moisture conditions.

Many statistical studies of drought have been made but almost without exception they are mere tabulations of days receiving less than a specified amount of rain.

For example, in a probability analysis of drought in the United States published in 1942, a drought period was defined as one in which not more than 0.10 inch of precipitation occurred in any consecutive 48 hours. In 1946 the drought periods of six Georgia stations were tabulated, with drought being defined as "a period of 14 days or more in which there is not one-quarter of an inch of rainfall in any one 24-hour period." A suggestion was made in 1954 that the Weather Bureau punched cards be used to tabulate the number of days without rain during the growing seasons at various places in the last 20 years. These tabulations "would yield the basis for saying that in July, 3 years out of 5 will get one dry spell which will last 18 or more days". It was felt that such an analysis would enable agricultural engineers and others to tell farmers whether or not they should invest in irrigation equipment.

Actually such tabulations of number of days without rain do not give information about drought. It is evident that we cannot define drought as a shortage of rainfall alone because such a definition would fail to take into account the amount of water needed. Furthermore, the effect of a shortage of rainfall depends on whether the soil is moist or dry at the beginning of the period. Shantz explained that drought in its proper sense is related to soil moisture and that it begins when the available soil moisture is diminished so that the vegetation can no longer absorb water from the soil rapidly enough to replace that lost to the air by transpiration. Drought does not begin when rain ceases but rather only when plant roots can no longer obtain moisture in needed amounts.

To determine how severe drought may be in a place, we must compare water need with water supply in individual years. In that way we can determine how often water deficiencies of various amounts take place. As an example we may consider selected moisture data obtained during the 25 year period 1920 - 44, at four stations in agricultural areas of the United States: Hays, Kansas; Charles City, Iowa; Wooster, Ohio; and Auburn, Alabama (table 3.1).

Table 3.1

Comparative Moisture Data of Selected Stations
(Median values of 25 year period 1920 - 1944)

Station	Precipitation inches	Potential Evapo inches	Actual Evapo inches	Water Surplus inches	Water Deficiency inches
Hays, Kansas	22.13	31.26	21.02	0.00	11.30
Charles City, Iowa	30.94	25.91	23.19	8.90	2.72
Wooster, Ohio	35.63	26.46	23.82	11.26	2.91
Auburn, Alabama	49.84	39.45	33.07	18.43	6.73

In Hays, the least rainy station, the average rainfall is about 22 inches. In Auburn, the rainiest, it is about 50 inches. The average rainfall in Hays is about 10 inches less than the need; in Auburn it is 10 inches greater. In Auburn, however, much of the rainfall comes in winter, when it is not needed. It becomes surplus water and flows away. In the summer, however, water deficiency is large. Water deficiency in Hays also is large, ranging from nearly 20 inches to about 2 inches. In Hays there is not enough rainfall; in Auburn there is more than enough, but it is badly distributed through the year (figure 3.2).

In both Charles City and Wooster, with a better distribution of rainfall through the year, the water deficiencies are smaller, being less than 3 inches in half of the years of record. Drought intensity and frequency are a little smaller in Charles City than in Wooster. There is also a lower water surplus in Charles City - 8.90 inches, compared with 11.26 inches. Thus, of the four stations, Charles City most nearly approaches the ideal climate for agriculture, for its water supply most nearly coincides with water need.

ANNUAL WATER DEFICIENCY AT SELECTED STATIONS
DURING 25-YEAR PERIOD 1920-1944

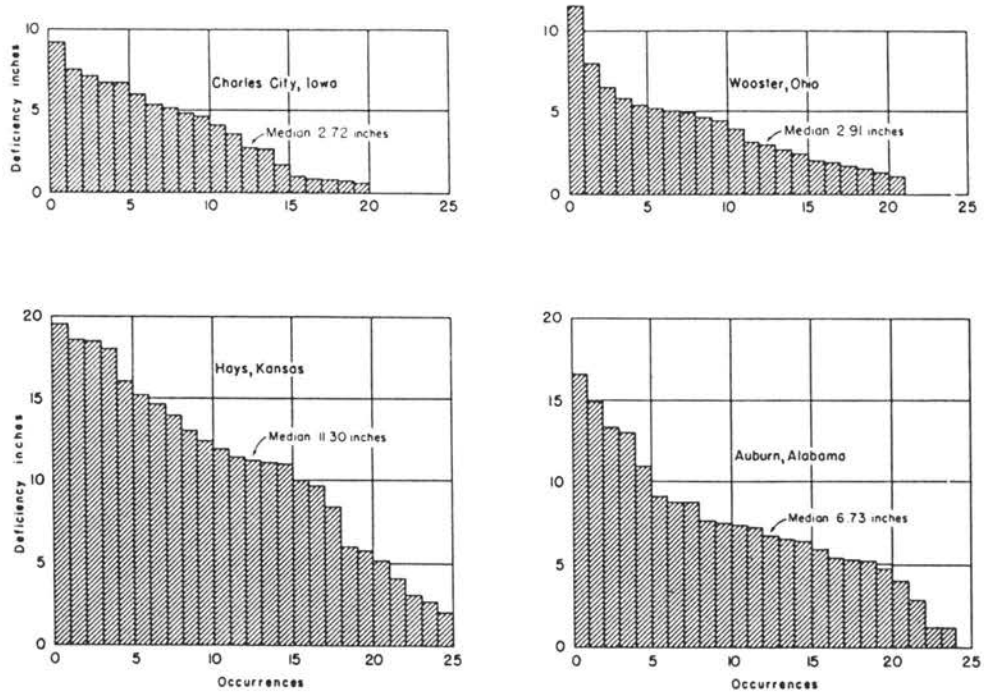

Figure 3.2

Differences in average yields in different localities are proportional to differences in drought incidence. The farmers of the East and Southeast get low returns from their work on the land partly because of high drought incidence resulting from the lack of coincidence between rainfall and water need. During much of the growing season the soil does not contain enough moisture, and in the non-growing season a large water surplus impoverishes the soil by leaching.

To farmers everywhere drought is a serious matter. Drought is hard to measure because we are not yet able to determine the water needs of plants very accurately. We do not know when to expect droughts or how intense they may be. Therefore, we cannot be sure which moisture-conservation measures may be best at a given time and place. Droughts deserve real study. Not until we have conquered drought by scientific irrigation will we achieve the maximum production from the soil.

A Case Study: An Irrigation Program for Weona Farms, Arkansas

From an agricultural point of view moisture and temperature are the two most important climatic parameters to consider. Any evaluation of the climatic environment of a place must of necessity be primarily concerned with the intensity and variability of these two factors.

Temperature by itself is not as useful in agricultural studies as is the potential evapotranspiration which may be considered to be the temperature expressed in terms of the water evaporated and transpired from a moist, vegetation-covered surface. Figure 3.3 shows the average course of precipitation, actual evapotranspiration, and potential evapotranspiration at Marked Tree, Arkansas, the nearest weather station to Weona Farms. Table 3.2 gives the actual monthly water balance computations for the same station.

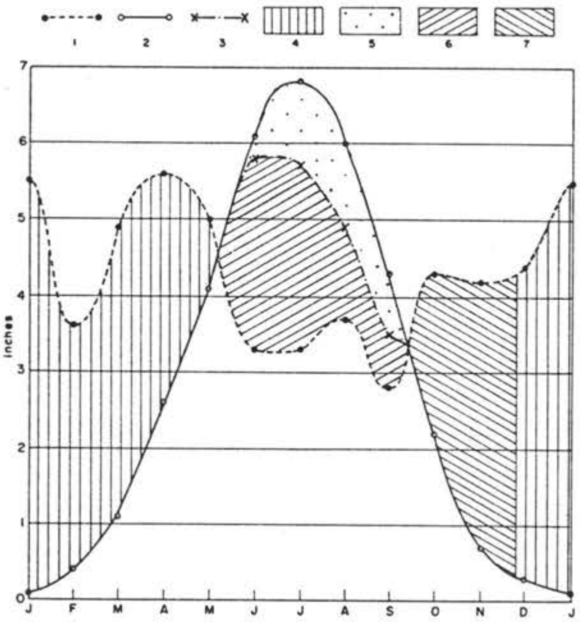

Figure 3.3

Average march of precipitation (1), potential evapotranspiration (2), and actual evapotranspiration (3) at Marked Tree, Arkansas. Diagram shows periods of water surplus (4), water deficit (5), soil moisture utilization (6), and soil moisture recharge (7).

Table 3.2

Average Water Balance, Marked Tree, Arkansas
(All values in inches)

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Potential Evapo.	0.1	0.4	1.1	2.6	4.1	6.1	6.8	6.0	4.3	2.2	0.7	0.3	34.7
Precipitation	5.5	3.6	4.9	5.6	5.0	3.3	3.3	3.7	2.8	4.3	4.2	4.4	50.6
Difference	5.4	3.2	3.8	3.0	0.9	-2.8	-3.5	-2.3	-1.5	2.1	3.5	4.1	15.9
Storage Change	0	0	0	0	0	-2.6	-2.5	-1.2	-0.7	2.2	3.5	1.4	
Moisture Storage	12.0	12.0	12.0	12.0	12.0	9.4	6.9	5.7	5.0	7.1	10.6	12.0	
Actual Evapo.	0.1	0.4	1.1	2.6	4.1	5.8	5.7	4.9	3.5	2.2	0.7	0.3	31.4
Moisture Deficit	0	0	0	0	0	0.3	1.1	1.1	0.8	0	0	0	3.3
Moisture Surplus	5.4	3.2	3.8	3.0	0.9	0	0	0	0	0	0	2.7	19.0

Variability of climatic factors from year to year is one of the most troublesome characteristics of climate. Although both temperature and precipitation vary, it is the precipitation which varies most widely and is most important in influencing the nature of the climate in mid-latitude regions. If one studies the 50-year record of monthly precipitation during the crop growing season (May-September) at Marked Tree, Arkansas, it can be seen that precipitation in any one month can vary from a trace (August, 1909) to a maximum of 10.8 inches (May, 1930). The annual precipitation has varied from over 77 inches to less than 30 inches. Fifty percent of the time the total rainfall for the five crop-growing months is less than 17.6 inches. In only one year in twenty is the summer rainfall greater than the average plant water need of 25.5 inches, a value which is determined from the average potential evapotranspiration for Marked Tree and given in table 3.2. Thus, once in every twenty years the total summer rain might be more than the plants could use, but even then, there may be some need for supplemental irrigation due to the fact that the rainfall might be unequally distributed through the season.

During the past fifty years only three times, in 1923, 1945, and 1951 did the rainfall from May through September exceed the average plant water need (figure 3.4). The year 1945 is an example of defective distribution of rainfall, with 8.89 inches in June, 3.33 in July, 0.44 in August, and 9.65 in September. When we compare these precipitation records with the average water need (potential evapotranspiration) for the corresponding months as given in table 3.2 we find a surplus of 2.8 inches in June, a deficit of 3.5 inches in July, a deficit of 5.6 inches in August, and a surplus of 5.4 inches in September.

MAY - SEPTEMBER PRECIPITATION
MARKED TREE, ARKANSAS

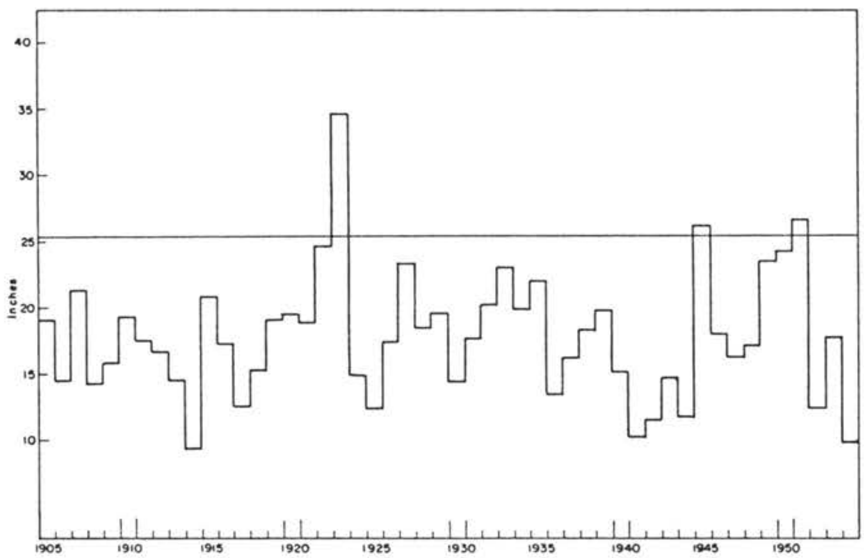

Figure 3.4

During two other years the value of total rainfall was less than 10 inches (1914, 1954) during the May to September period. Figure 3.4 shows clearly the possibility of great variation in the summer rainfall. Because of this, it is desirable to consider the comparison of precipitation and potential evapotranspiration for individual wet and dry years to bring out the incidence of drought as the average water balance cannot do, and to reveal the magnitude of the moisture problems in the Weona area. Figures 3.5 and 3.6 show the water balances for the years 1951 and 1954 respectively.

In 1951, the precipitation was over 3.6 inches in all months but March, May, and October. Because of this the water surplus was great, being over 24 inches for the year as compared with an average annual surplus of about 19 inches. In the summer there was no single period of soil moisture draw-down with accompanying water deficit as in the case of the average water balance (figure 3.3) or the balance for the year 1954 (figure 3.6). Instead three short periods of deficit of .4, .3, and .1 inch respectively separated by periods of soil moisture recharge occurred. The average water deficit is about 3.4 inches but in 1951 it amounted to only about .8 inch. Because the August precipitation in 1951 was considerably under the August water need there was still some need for supplemental irrigation as will be seen in later calculations of irrigation requirements (figure 3.10).

The water balance for 1954 (figure 3.6), a record dry year, is quite similar to the average water balance (figure 3.3) in shape although the extent of the water deficit is much greater - 11.5 inches compared with 3.4 inches - and the water surplus is less - 7.2 inches compared with 19 inches. In 1954 supplemental irrigation was needed from early June through September to supply the plant water needs adequately.

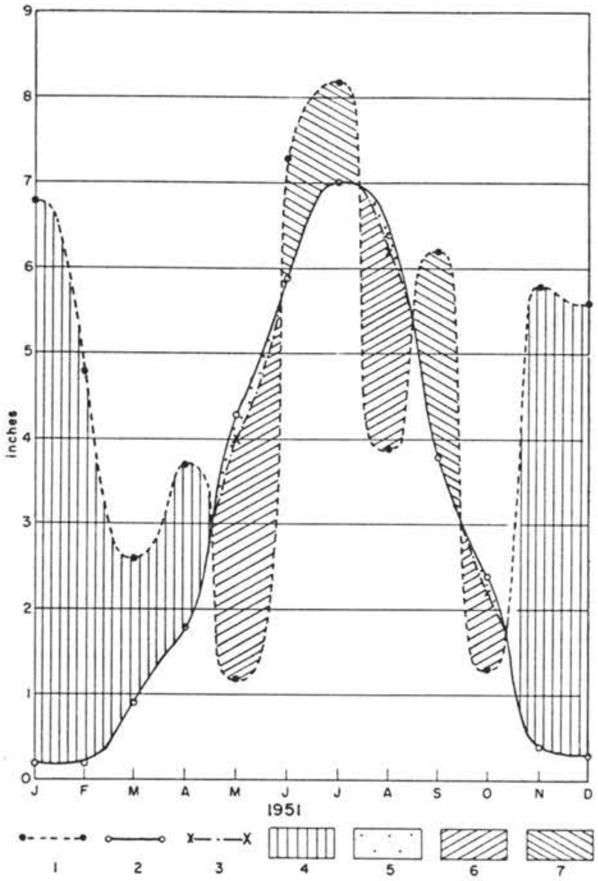

Figure 3.5

March of precipitation and water need through a moist year, Marked Tree, Arkansas, 1951.

(1) Precipitation; (2) Potential evapotranspiration; (3) Actual evapotranspiration; (4) Water surplus; (5) Water deficit; (6) Soil moisture utilization; (7) Soil moisture recharge.

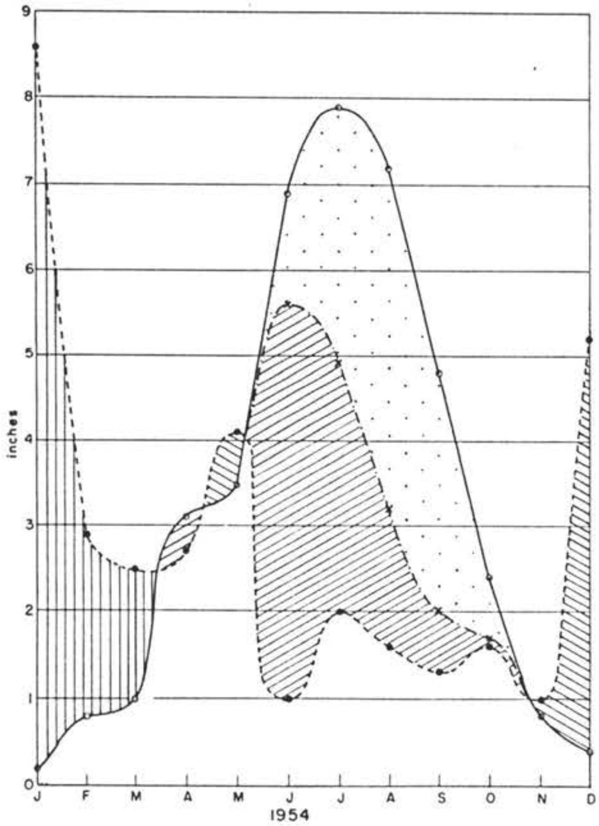

Figure 3.6

March of precipitation and water need through a dry year, Marked Tree, Arkansas, 1954.

Due to the poor distribution of precipitation within a period of a month these general water balances (figure 3.5, 3.6) cannot be used to reveal the exact need for supplemental irrigation which exists. It is only from daily computations of soil moisture content that the necessary information on irrigation requirements can be obtained. The various water balance graphs do show, however, that there is a marked change in the moisture conditions of the region around Weona, Arkansas, from one year to the next. Even in the so-called wet years, a poor distribution of rainfall results in some periods of moisture deficit and a need for supplemental irrigation. In the dry years, irrigation becomes a necessity if even fair yields are to be obtained.

In order to determine the most practical program of irrigation for Weona Farms the day-to-day record of precipitation and temperature at Marked Tree, the nearest weather station, has been investigated for the seven year period 1948-1954. From such a study it has been possible to determine the daily variation in soil moisture for the whole period of record and to determine the feasibility of irrigation, the amount of the irrigation application, and to learn something of the nature of the problems which will be involved in any program of supplemental irrigation.

Figures 3.7-3.13 show the variation in computed soil moisture deficit through each of the past seven years at Marked Tree. In every year the soil is full of water in the spring but as temperatures increase in May and June, resulting in greater evaporation and transpiration of the soil moisture, the soils begin to dry. Depending on the distribution of summer rainfall this drying of the soil continues and increases until about September when again the precipitation in most years begins to exceed the plant water use and the soil moisture deficit is reduced. From 1948 to 1951 summer rains were reasonably well distributed and the soil never had a moisture deficit greater than 5 inches. In two of the years it was never greater than about 3.5 inches. The last three years of the record have been much different with poorly distributed summer rainfall or almost none at all. In those years there was a rapid drying of the soil in May and June giving moisture deficiencies of about 6 inches early in the year. Unirrigated land largely remained with great moisture deficits for the rest of the crop-growing season. Whereas in 1948-1951 the October moisture deficits ranged generally between 0-2 inches, the October deficits during 1952-1954 were between 4-6 inches due to the lack of sufficient rainfall.

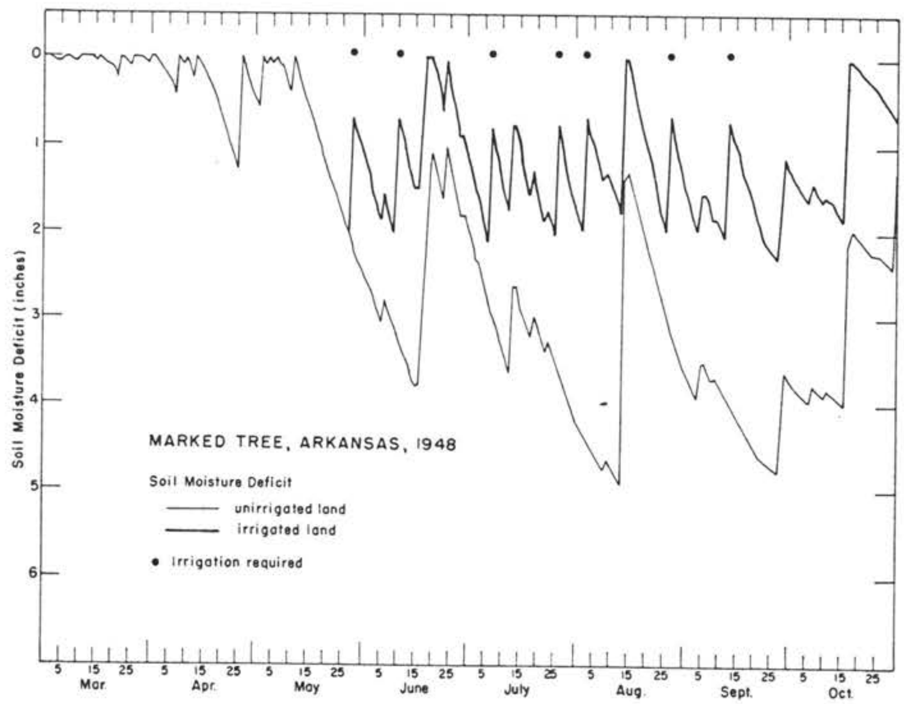

Figure 3.7

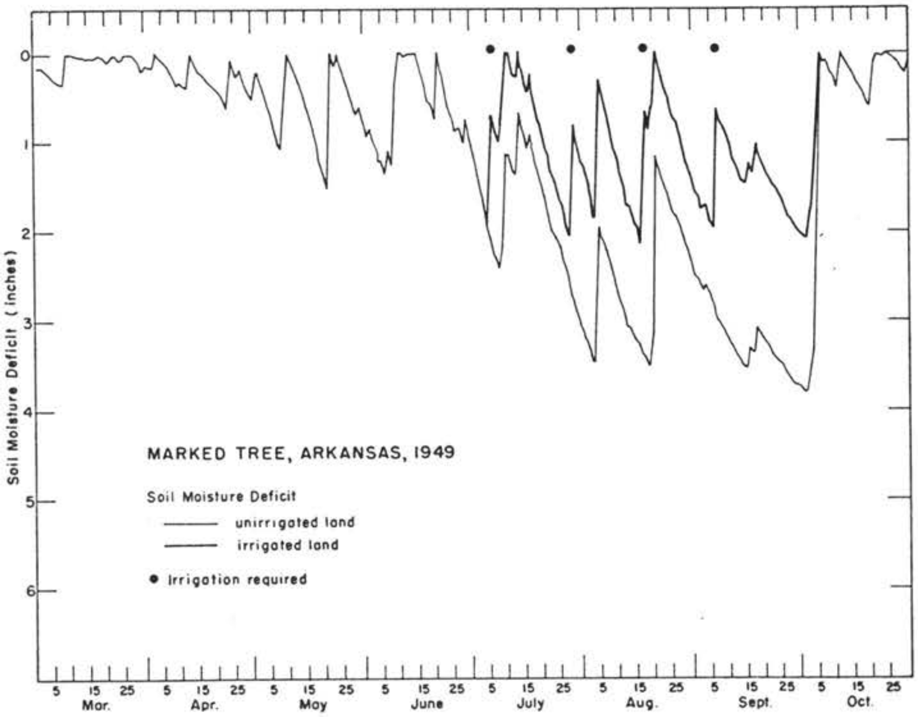

Figure 3.8

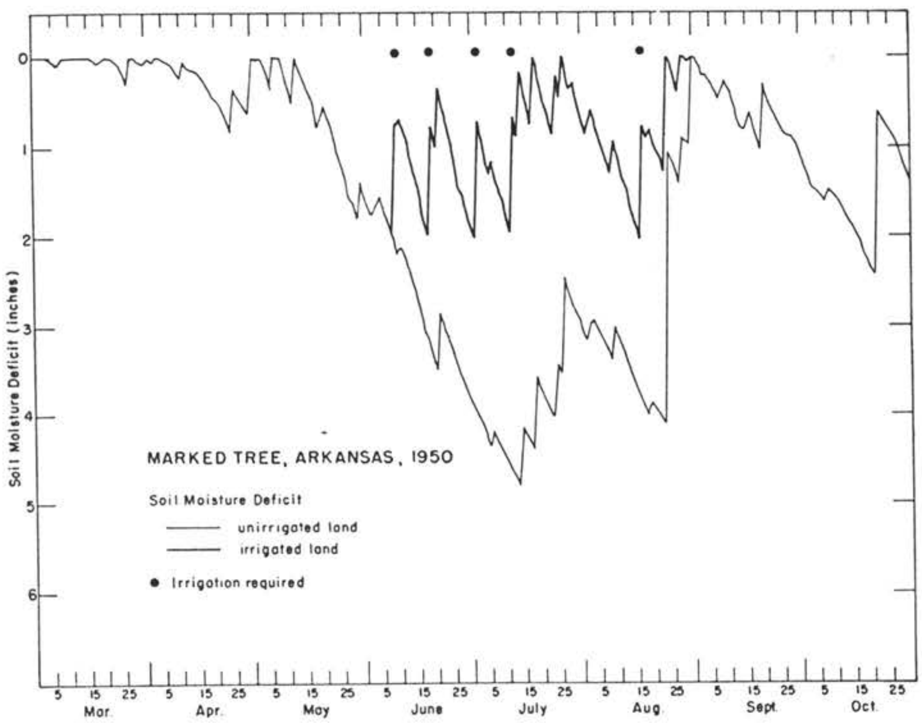

Figure 3.9

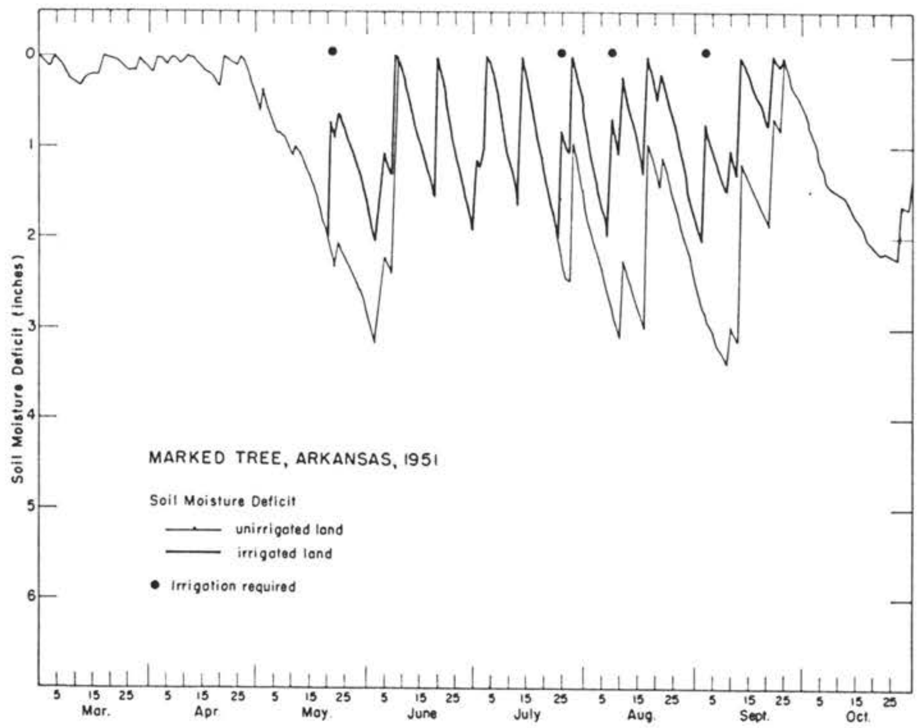

Figure 3.10

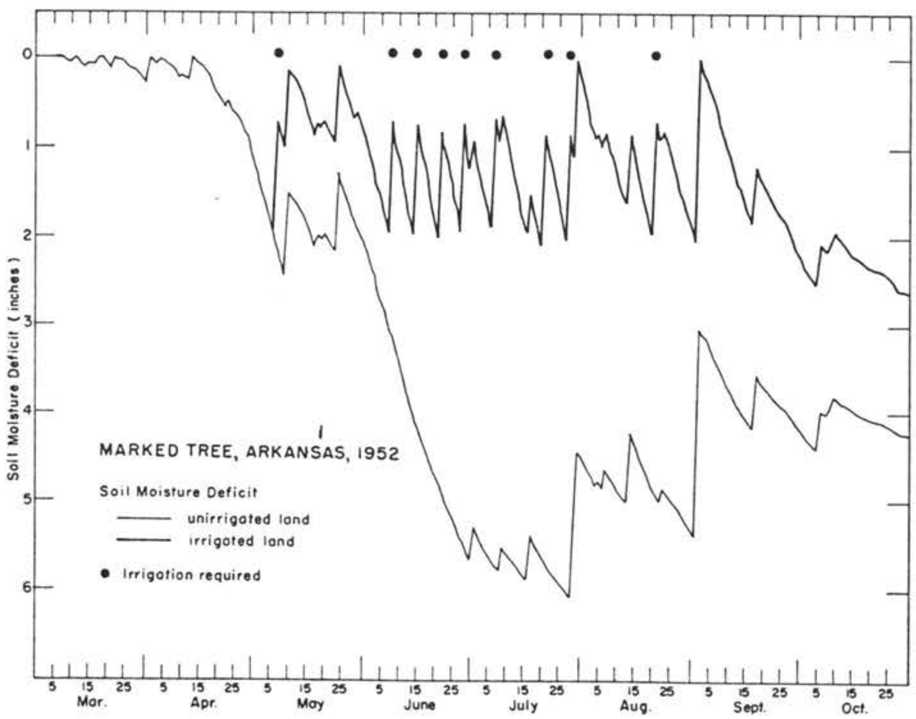

Figure 3.11

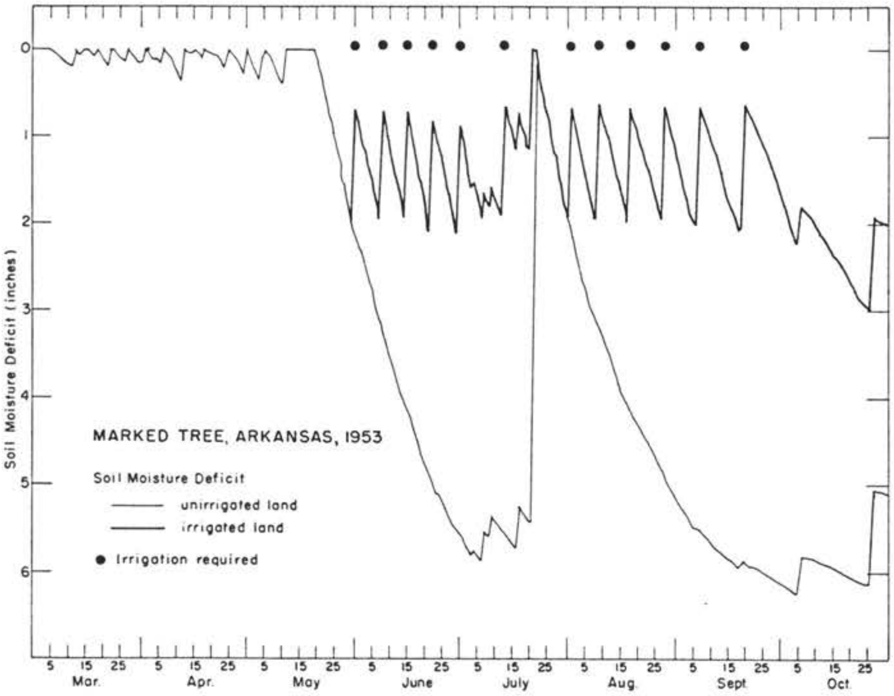

Figure 3.12

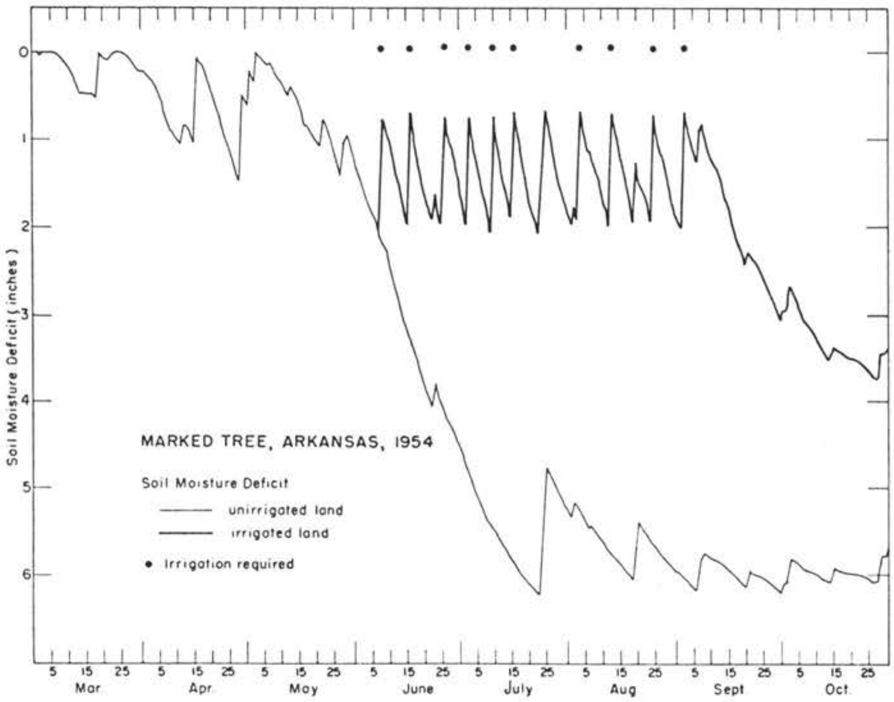

Figure 3.13

From studies of cotton irrigation which are available it appears that peak yields for this plant are achieved if the plant roots develop and grow in moist but not completely wet soils. Thus, it is desirable to let the soil dry out to a certain degree between irrigation applications. Also because of the possibility of precipitation in each of the summer months in northeastern Arkansas it is not desirable to bring the soil moisture in the whole profile back to field capacity at each irrigation. It is necessary to bring the soil profile, through irrigation, only back to 85 to 90 percent of field capacity and to allow a certain safety factor for rains which might occur shortly after irrigations. This will minimize the possibility of leaching of the soil due to excess water during the periods of rains.

From an analysis of the climatic conditions in the Weona area, it appears that it is most feasible to schedule irrigation whenever the soil moisture deficit drops to 2.0 inches and to apply 1.5 inches of water in each irrigation. This will return the soil moisture deficit to not less than .5 and provide a reasonable safety factor for unexpected rains. This irrigation schedule has been indicated on the graphs of soil moisture (figures 3.7-3.13) to show something of the nature of the problems involved. The times of irrigation have been marked on each diagram as well as the day-to-day variations in the soil moisture on the irrigated areas. Table 3.3 sums up the results of the seven years of hypothetical irrigation at Marked Tree.

Table 3.3

Irrigation Summary, Marked Tree, Arkansas, 1948-1954

	No. of Irrigation Applications	Water Applied (inches)	Water Applied (gal/acre)	No. of Days May-Sept. Soil Moisture Deficit on Irrigated Land Over .50 in.	1.00 in.	1.50 in.
1948	7	10.5	285,117	128	97	56
1949	4	6.0	162,924	111	70	30
1950	5	7.5	203,655	103	53	25
1951	4	6.0	162,924	113	70	24
1952	9	13.5	366,579	137	91	49
1953	12	18.0	488,772	129	92	52
1954	10	15.0	407,310	139	105	62

The table shows that irrigation was needed in each of the seven years and ranged from four 1.5 inch applications in the wet years of 1949 and 1951 due to a poor distribution of the rainfall, to twelve applications in the dry year of 1953. This resulted in a total water need of from 162,000 to 489,000 gallons of water per acre during the crop-growing season. The last three columns of table 3.3 give a breakdown of the soil moisture conditions during each of the years and show how well the soil moisture conditions were maintained between .5 and 2.0 inches of water deficit. In four of the seven years the soil moisture deficit was over 1.5 inches on almost 1/3 of the days during May to September. This would seem like a large number of days and might indicate that the soil was allowed to become and remain too dry for optimum yields. This is not the case, however, as is partially shown in table 3.4. This table gives the amount of water that would be lost from irrigated and unirrigated soil at Marked Tree during each of the seven years. That any water would be lost during the dry summer period results from the fact that occasional heavy storms provide more water than the soil needs at a given time. With irrigation, smaller storms occurring after irrigation might saturate the soil and cause runoff. The object in irrigation is to keep this loss of water as low as possible yet still maintain moist soil conditions. Table 3.4 shows that some water was lost from the irrigated soils in all years except 1954. In 1950 and 1951 even the unirrigated soils lost water due to excessive precipitation. The irrigated land lost larger amounts of water, of course, in those years due to the irrigation.

Table 3.4

Water Surplus (inches) from Time of First Irrigation until September 30 on Irrigated and Unirrigated Land Marked Tree, Arkansas, 1948-1954

	Unirrigated Land	Irrigated Land
1948	0	3.04
1949	0	2.13
1950	.22	3.68
1951	3.64	7.62
1952	0	.90
1953	0	4.10
1954	0	0

When table 3.4 is compared with the last three columns of table 3.3 it is apparent that the compromise irrigation program suggested will provide the most satisfactory solution to the irrigation problem at Weona. If the land were allowed to dry further before irrigation, or were irrigated less at each time to reduce the amount of water surplus during times of rain, there would be longer periods of water deficiency and a corresponding reduction in crop yields. If, on the other hand, more irrigation was applied to keep the soil moist at all times, the amount of water surplus would be larger and there would be more leaching of the soil and more waste of the water and soil resources. Thus, the program as outlined appears most feasible from a climatic viewpoint at the present time.

IT MUST BE POINTED OUT THAT THE PROGRAM OF IRRIGATION SUGGESTED ABOVE, 1.5 INCHES EVERY TIME THE MOISTURE DEFICIT REACHES 2.0 INCHES, HAS BEEN DETERMINED ON THE BASIS OF THE CLIMATIC CONDITIONS ALONE AND MAY HAVE TO BE MODIFIED ON THE BASIS OF ECONOMIC PRACTICABILITY. From the viewpoint of the operation of the irrigation system itself it may prove more desirable to irrigate between 2.0 and 2.5 inches depth of water every time a moisture deficit of 3.0 inches occurs. This will, of course, result in longer periods with the soil moisture deficit below 2.0 inches. This, in turn, will reduce crop yields somewhat but the reduction in yield may be more than offset by the savings resulting from a more practical irrigation program. It is felt that cotton yields, for instance, under such an irrigation program would be less than they would be under the irrigation program suggested from climatic considerations alone although they would still be considerably higher than under no irrigation or a haphazard program of watering. THE FINAL IRRIGATION PLAN MUST BE A COMPROMISE BETWEEN THE CLIMATIC AND ECONOMIC FACTORS WHICH ARE INVOLVED IN IRRIGATION SCHEDULING.

The program of irrigation, of course, must be changed if different crops are used. With shallow-rooted crops such as beans, peas, spinach, and the like, it would be more desirable to irrigate when the soil moisture deficiency was no more than 1.0 or 1.2 inches and to irrigate about .75 inch at each application. More frequent but smaller applications are necessary to prevent undue moisture deficits in the shallower plant root zones. Corn and grains would generally follow the irrigation program for cotton while orchards and pastures might not need irrigation until about 2.5 or 3.0 inches of deficit had occurred. With these deep-rooted crops each irrigation application might be as large as 2.0 to 2.5 inches.

It is possible for the individual irrigator to keep his own soil moisture balance and to schedule irrigation on his own farm provided he is willing to keep the necessary records. Forms, tables, and instructions for making the soil moisture calculations can be made available for any specific area but use of them involves considerable work and attention to detail. Thus, for most farming enterprises it is not feasible for the individual operator to maintain his own soil moisture balance sheet.

Through the use of the water balance bookkeeping procedure it has become possible to determine with good reliability the day-to-day variations in soil moisture content. This knowledge, has, in turn, led to a method of scheduling supplemental irrigation. This method enables a farmer to determine the time and amount of each irrigation application. With a rigorous control of the soil moisture content now possible a new era of scientific irrigation may become a reality.

The Future Direction of Irrigation

In 1949 the irrigated acreage in New Jersey stood at 28,117, less than 3 percent of the land in harvested crops in that year. The possibility of bringing the remaining New Jersey farm land under irrigation offers a great challenge. But there is a much greater challenge for scientific irrigation in southern New Jersey. There is an opportunity through irrigation to reclaim and develop the extensive empty areas which are unsettled and which now produce nothing.

The southern counties of New Jersey contain more than two and a half million acres of land, but only four hundred thousand acres were in harvested crops in 1949 of which only thirteen thousand were irrigated. At least one and a half million acres are considered non-arable waste land. The soils of this area, which is a part of the Atlantic coastal plain, are generally sandy and incapable of storing much water for use of crops during periods of summer drought. Since the soils dry out so completely in summer the vegetation is ravaged by fire. It is made up of stunted and fire-damaged pine and scrub oak. Large sections are known as the barrens and are considered worthless.

We have recently had some experience in irrigating this kind of land and have learned that it possesses great hidden potentialities. Almost everywhere in southern New Jersey ground water is readily available for irrigation within a few feet of the ground surface. Therefore, it is ideally situated to make profitable use of scientific irrigation. Although this vast area is non-arable by ordinary standards, when water is applied to it by proper irrigation and its mineral deficiencies are rectified by correct fertilization, it can be highly productive. If this waste area, that now produces nothing, were reclaimed and planted to potatoes it could easily produce five hundred million bushels a year; more than has ever yet been produced in the whole United States. This is only an illustration of the potentialities of this neglected region.

This huge job of reclamation in southern New Jersey can be done more cheaply and with less risk and uncertainty and would be economically more defensible than many of the large reclamation works of the West. It is not a job for the Federal Bureau of Reclamation, however, but for private interests. Since the task is expensive, most of the development will probably be done by large farming corporations. The insurance companies might take note of this opportunity for profitable investment of large sums of money.

What has been said of the possibilities of changing the climate in southern New Jersey by counteracting drought through supplemental irrigation applies almost equally to the entire Atlantic and Gulf coastal plains and the Mississippi delta from Long Island to Louisiana and Arkansas. In this vast area consisting of approximately 200 million acres, soil moisture becomes deficient during the summer but ground water is abundant within a few feet of the surface. Over the earth's surface there are many other areas as large as our own coastal plain where the climate exhibits the defect of a large water surplus in one season and a large deficiency in another and where irrigation provides the means to correct this defect. Through changes in the water balance which man brings about, sometimes deliberately and sometimes inadvertently, he exercises his greatest influence on climate. Through destruction of the natural ground cover and cultivation of the soil he reduces the water holding capacity of the soil and increases the incidence and severity of drought. In the regions of seasonal drought where there are alternating wet and dry periods, scientific irrigation can eliminate drought. This is where man is able to remedy a defect in climate and in so doing will do more to increase the food supplies of the world than he has ever done before in any other way.

CHAPTER 4. THE MOISTURE SURPLUS*

Since the moisture surplus as computed from the bookkeeping procedure represents water that is available for stream flow it is possible to obtain estimates of this latter parameter from climatic data alone. Thus, the procedure enables one to determine stream flow in areas where no stream gage records exist. If such records are available the information can be used as a check on the validity of the bookkeeping approach. Of course, all of the moisture available for runoff in a month will not be lost in that month for there will be some detention of moisture on the watershed past the end of the month. Studies have shown that on a moderate to large watershed only about 50 percent of the available surplus water will run off in a given month, the rest being retained for later runoff.

A number of illustrations of the results which are obtained from use of the water balance to determine stream flow can be cited. Figure 4.1 is a map of average annual surplus water over eastern United States as computed from the water balance procedure. Figure 4.2 is a map of the same region showing the average annual measured runoff from gaged streams. The first map was prepared entirely from climatic data while the second was drawn from the many measurements of stream flow made by the U.S. Geological Survey. These maps are sufficiently alike that one could be substituted for the other without loss.

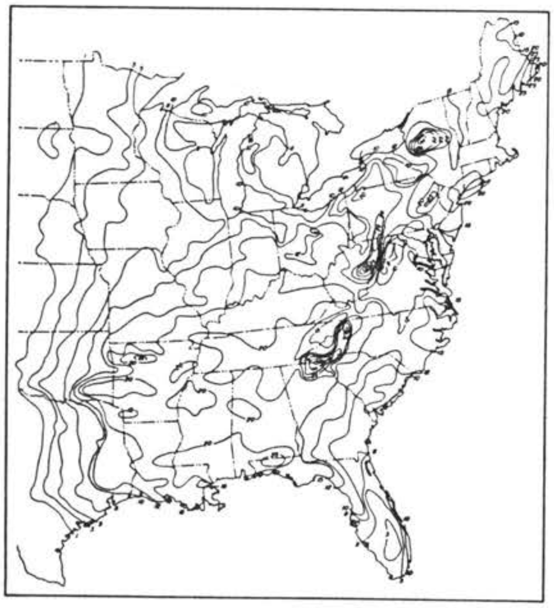

Figure 4.1

Figure 4.2

Figure 4.1 shows the average annual computed water surplus in Eastern United States while Figure 4.2, from National Resources Board Report of December 1, 1934, shows the average annual measured water runoff from gaged streams. (All values in inches.)

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* The material in this chapter has been adapted from "The Water Balance in Arid and Semiarid Climates," by C. W. Thornthwaite, Desert Research, Proceedings International Symposium, Jerusalem, May 8-14, 1952, pp. 114-135; Discussion by C. W. Thornthwaite of "An Accounting of the Daily Accretion, Depletion, and Storage of Soil-Water as Determined by Weighing Monolith Lysimeters," by Harrold, L. L. and Dreibelbis, F. R., Trans. Amer. Geophys. Union, Vol. 26, No. 2, 1945, pp. 283-297; "Report of the Committee on Transpiration and Evaporation, 1943-44," by H. B. Wilm, C. W. Thornthwaite et al, Trans. Amer. Geophys. Union, Part V, 1945, pp. 683-693; and "The Water Balance of the Lake Maracaibo Basin During 1946-53," by D. B. Carter, C. W. Thornthwaite Associates, Centerton, 1955, 39 pp. (processed). (Used with permission of the Creole Petroleum Corp.)

This agreement between measured and computed values of runoff can again be shown by reference to an arid region such as Arizona in southwestern United States. Figure 4.3 shows a map of water yield or runoff for Arizona. Although the author does not indicate how this map was obtained the inference is given in the accompanying publication that stream gaging records were used. Figure 4.4 shows water surplus for the same area computed from average monthly values of temperature and precipitation. The agreement clearly indicates the utility of the method of computing stream runoff from climatic data.

There are three watersheds near Coshocton, Ohio that are sufficiently large and that possess physical characteristics such that no significant amounts of water flow from them unmeasured. They are Killbuck Creek above Killbuck (466 square miles), Licking River above Toboso (672 square miles), and Wills Creek above Bird Run (730 square miles). Using the records of temperature and precipitation of single Weather Bureau stations centrally located in each of these watersheds, the monthly evapotranspiration and runoff have been computed for a number of years. To the extent that the precipitation and temperature registered at the station are not the same as for the watershed, the computed runoff should be expected to differ from the measured runoff. Monthly and annual computed values of runoff are, however, in good agreement with measured values (figures 4.5, 4.6) and the average values given in table 4.1, based on the series of years investigated, are almost identical.

Table 4.1

Computed and Observed Runoff from Tributary Watersheds of the Muskingum Drainage Basin.

Watershed	Weather Bureau Station	Computed Yearly Runoff (in.)	Observed Yearly Runoff (in.)
Killbuck Creek	Wooster	10.94	10.24
Licking River	Newark	12.94	12.80
Wills Creek	Cambridge	12.95	12.41

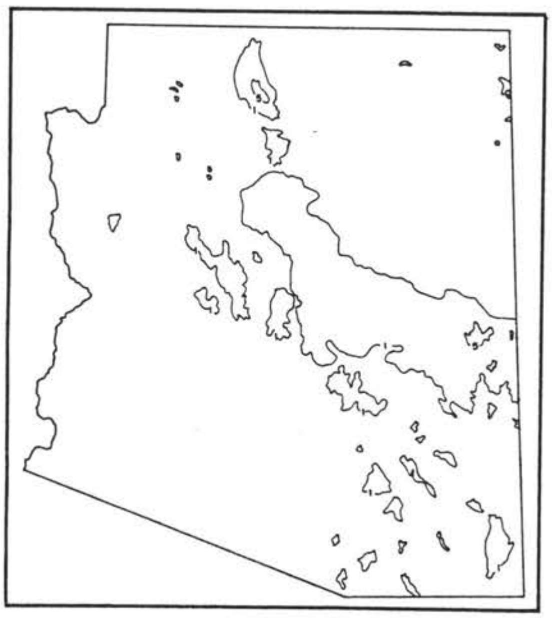


Figure 4.3 - Average annual water yield in Arizona, in inches.

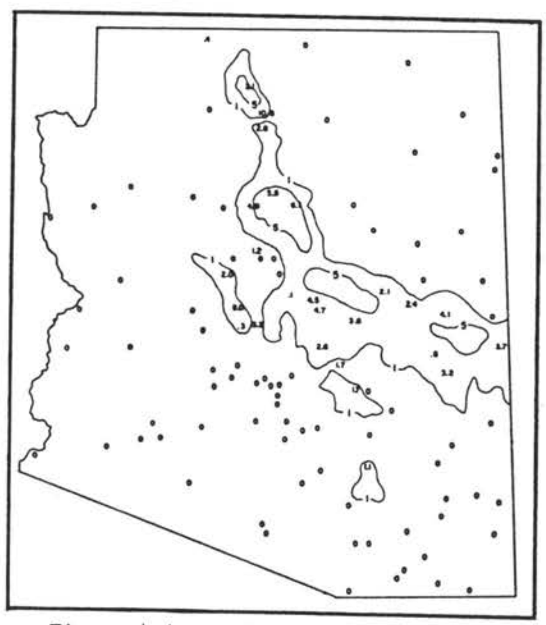


Figure 4.4 - Average annual water surplus in Arizona, in inches.

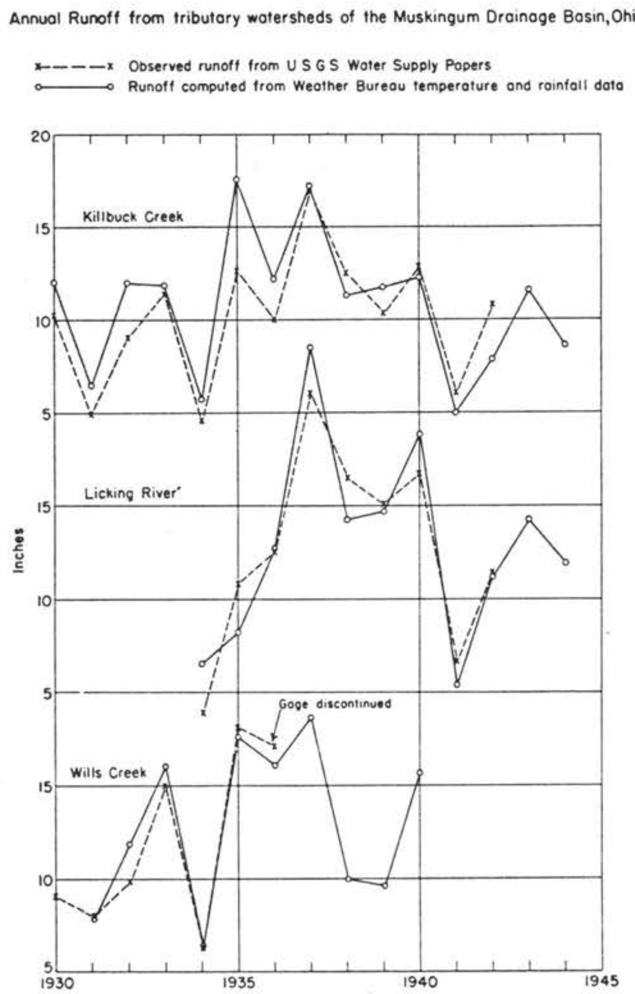


Figure 4.5

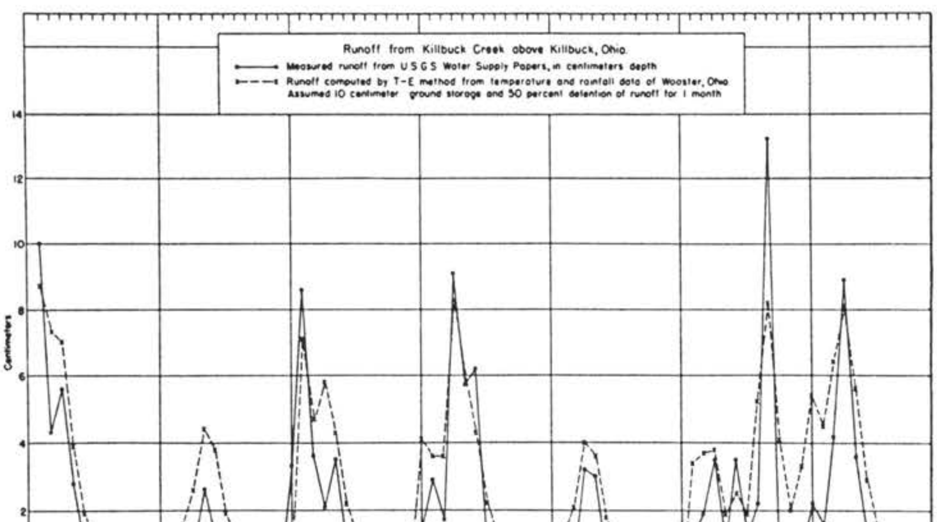


Figure 4.6

The greatest discrepancy between computed and observed runoff is for the year 1935 in Killbuck Creek, where computed runoff exceeds observed by 5.13 inches. But in that year the precipitation at Wooster was considerably more than that over the whole watershed. The annual precipitation at Wooster was 7.44 inches above normal and that at Ashland and Akron, near the upper end of the watershed, was respectively 2.12 inches and 2.21 inches below normal.

In 1943 a map of the measured stream runoff from 124 small watersheds in the Tennessee River Basin appeared. The close correspondence between the data on that map and the runoff as computed for the same area by means of the water balance bookkeeping procedure is excellent (figure 4.7), especially since the measured data cover a uniform period from 1920 to 1942 and the computed data result from averages of temperature and precipitation for various periods prior to 1930. Furthermore, the measured data of the Tennessee Valley Authority refer to areas ranging in size from a few square miles to nearly two thousand whereas the computations refer to points where the weather observations were made.

As a further check on this approach in the field of hydrology it is possible to compare the computed runoff with the actual observations for an individual watershed for a long series of years. This procedure involves the problem of getting true values of monthly precipitation for watersheds of considerable area. As an example

MEAN ANNUAL RUNOFF IN TENNESSEE RIVER BASIN

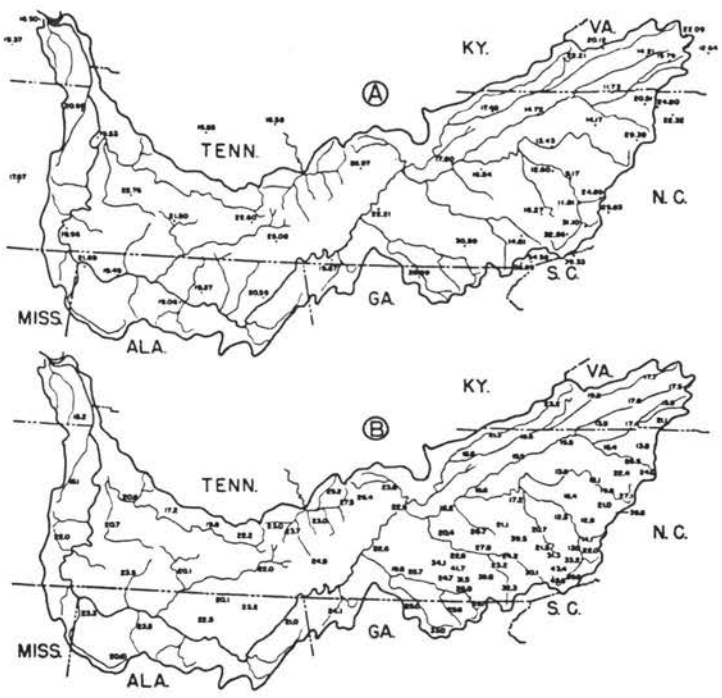


Figure 4.7

A - Water surplus (in inches) computed from temperature and precipitation data at Weather Bureau stations. Period of record variable ending in 1930. Figures refer to points of weather observations.

B - Observed runoff (in inches) adjusted to uniform period October 1930 through September 1942. From Tennessee Valley Authority, Hydraulic Data Division, Precipitation in Tennessee River Basin Annual 1943, page 16. Figures refer to watersheds.

the runoff from the Great Valley of Virginia has been computed from the averages of temperature and precipitation published by the Weather Bureau. This runoff has, in turn, been compared with the measured values of runoff from the James River Basin in Virginia. The two areas differ considerably but they have been used here to avoid the labor of assembling and computing the data for exactly comparable areas. Annual values of measured and computed runoff are given in table 4.2 while the seasonal course of the measured and computed values are given in figure 4.8. On the whole there is close agreement between the two sets of figures. Similar computations have been made for a number of other watersheds with equally acceptable results.

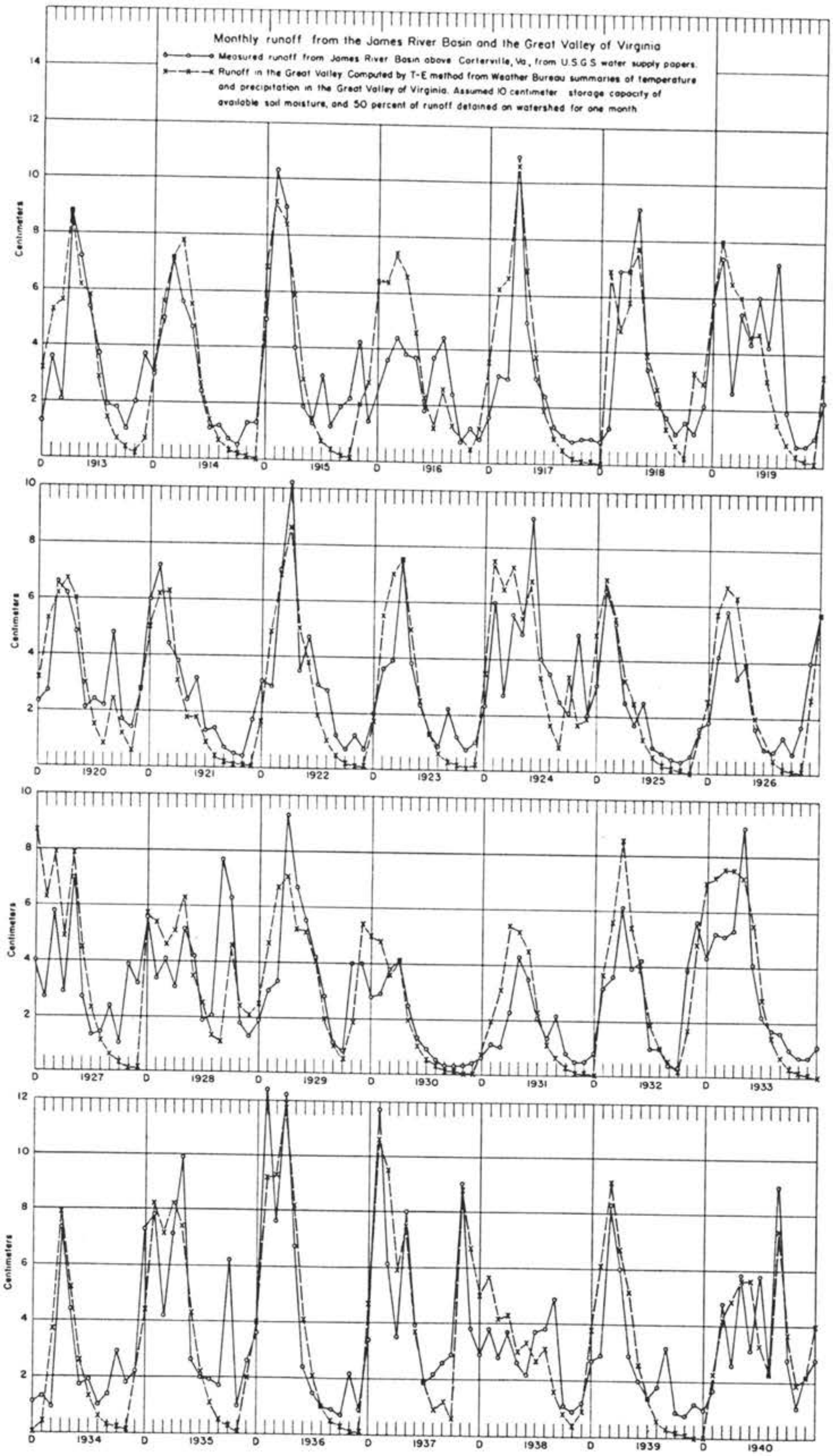
Table 4.2

Precipitation, Evapotranspiration and Runoff in the James River Basin and the Great Valley of Virginia (all values in inches)

Year	James River Basin[a] P	E	R	Great Valley of Virginia[b] P	E	R
1904	32.30	22.09	10.21	32.99	23.15	9.80
1905	44.01	29.26	14.75	41.93	26.85	13.43
1906	47.05	26.85	20.20	48.55	26.81	21.10
1907	44.44	24.79	19.65	45.11	24.09	20.87
1908	45.08	25.72	19.36	45.28	25.95	19.84
1909	34.20	17.83	16.37	38.89	24.60	16.14
1910	39.43	26.57	12.86	40.00	24.41	14.65
1911	41.39	28.60	12.79	44.85	26.22	17.76
1912	42.42	24.69	17.73	41.81	24.76	17.95
1913	45.01	27170	17.31	43.19	26.93	16.22
1914	36.59	22.45	14.14	38.78	22.24	15.12
1915	41.19	24.09	17.10	41.69	25.91	15.95
1916	37.92	25.28	12.64	39.84	15.87	15.12
1917	36.56	23.50	13.06	36.10	22.21	14.88
1918	45.30	28.35	16.95	47.28	26.77	18.19
1919	42.49	25.23	17.26	41.73	27.44	15.39
1920	45.15	27.89	17.26	40.99	23.98	16.34
1921	32.89	21.04	11.85	35.63	28.03	8.90
1922	41.67	25.90	15.77	41.06	27.36	13.66
1923	37.71	25.63	12.08	40.12	26.14	13.27
1924	47.30	27.52	19.78	44.30	23.47	20.28
1925	28.84	18.98	9.86	28.90	20.12	9.65
1926	39.67	26.48	13.19	41.14	23.86	14.92
1927	41.84	26.11	15.73	42.05	26.77	16.42
1928	42.94	26.04	16.90	40.16	25.12	16.30
1929	44.56	25.76	18.80	46.02	25.95	19.13
1930	21.16	14.16	7.00	23.70	18.58	6.73
1931	37.31	29.31	8.00	37.25	28.50	9.72
1932	44.21	29.37	14.84	45.23	24.41	17.40
1933	38.01	22.86	15.15	37.99	25.79	15.91
1934	44.43	41.10	13.33	41.34	27.28	11.34
1935	.....	.....	19.99	46.73	29.06	17.87
1936	.....	.....	20.44	45.59	25.00	20.28
1937	.....	.....	23.08	49.76	25.12	24.53
1938	.....	.....	13.08	39.37	26.58	13.27
1939	.....	.....	12.78	37.68	26.81	12.68
1940	.....	.....	17.31	44.49	23.94	18.66

[a] Data through 1934 from USGS Water-Supply Paper No. 772, p. 84; runoff for 1935 through 1940 from other Water-Supply Papers; E= P-R.

[b] E and R computed from temperature and precipitation data in USWB Climatological Data by Sections. Evapotranspiration plus runoff does not equal rainfall because of changes in soil-moisture and seepage-storage.


Monthly runoff from the James River Basin and the Great Valley of Virginia
Measured runoff from James River Basin above Cartersville, Va., from U.S.G.S. water supply papers.
Runoff in the Great Valley Computed by T-E method from Weather Bureau summaries of temperature and precipitation in the Great Valley of Virginia. Assumed 10 centimeter storage capacity of available soil moisture, and 50 percent of runoff detained on watershed for one month
Centimeters
D
1913
1914
1915
1916
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1918
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1920
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1938
1939
1940

Figure 4.8

The uses of information on stream runoff are many and varied. One of the most interesting occurs in the field of oceanography where it becomes possible through the use of the climatic bookkeeping procedure to determine the fresh water accessions from river flow and seepage to coastal estuaries. Often river flow is not gaged and seepage can only be roughly approximated. Previous estimates of such fresh water accessions were not detailed enough to be of help in problems concerned with the circulation of water in the estuary or the composition of the water itself.

Lake Maracaibo in Venezuela is a large coastal lake linked to the ocean by a narrow channel through which there is sometimes salt water gain from the ocean and at other times fresh water loss to the ocean. It is in a part of the world where no stream gaging stations exist, and where there is no possibility of getting such observations. Through use of the water balance bookkeeping procedure, however, it is possible to compute stream runoff on the basis of the climatic data at available weather stations. This has been done for the period 1946-1953 on a monthly basis in order to permit comparison directly with evidence available from other oceanographic research.

Over the water surfaces of the Maracaibo basin there is a loss of moisture by evaporation and transpiration; there is also a direct gain of moisture by the lake from precipitation onto its surface; and, there is an indirect gain of moisture by the lake of the runoff resulting from precipitation over the uplands. These are the components of the lake hydrography which constitute the fresh water accessions. They are balanced by a corresponding loss or gain through the straits at the mouth of the lake. The water balance is essentially an accounting of what happens to precipitation that falls over the lake and the watershed.

While the only significant source of fresh water for the basin is precipitation, its effect on the hydrography of the lake depends on whether the precipitation falls on the lake surface or onto an upland surface. In the former instance, the effect of precipitation on the level of the lake is direct and immediate while in the latter case there will certainly be some diminution of the moisture by evaporation and transpiration and also a delay in the arrival of the runoff at the lake. Accordingly, there are two dissimilar areas to be considered in the determination of the water balance of the Maracaibo basin. The lake and marsh area must be distinguished from the upland area.

The amounts of evapotranspiration and precipitation over the lake and marsh area and the amounts of upland runoff computed by means of the water balance during the course of this study cannot be measured easily. Adequate measurements of evaporation from the lake or of precipitation over the lake do not exist and there are no satisfactory measurements of stream flow and ground water discharge to compare with upland runoff. Likewise, actual measurements of water exchange between Lake Maracaibo and the Gulf of Venezuela which would provide a possible means of verification do not exist.

Water balances were calculated for eighty-three places in the Maracaibo basin from the available climatic data. Because of the large amount of data used, a single isolated extreme value had no extraordinary effect on the results. While the values of the water balances at every point are estimated, the methods of estimation are based on empirical relationships which have been verified as a result of world-wide studies. The procedure provides an approach to a problem which otherwise would be virtually insoluble.

On the basis of the water balance computations, a net monthly inflow of water from the Gulf of Venezuela to Lake Maracaibo was found to occur at least once in every year from 1946 to 1953 except during 1950. Inflows were usually sustained for three to four months duration; generally, March had the greatest inflow. In every year notable discharges from the lake occurred in October.

By making use of data from the climatological stations in the catchment area tributary to the lake and determining the potential evapotranspiration, the water surplus, and other elements of the water cycle month by month, it has been possible to work out the water economy of the lake and to show that at certain times there is an excess of water which flows into the ocean and in others a deficit which is made up by inflow from the ocean. The results of this study demonstrate clearly that the methods of climatology are indeed of real assistance in the solution of oceanographic problems.

The foregoing illustrations not only indicate the close agreement which can be found between measured and computed values of stream runoff but also show how it is possible to obtain from the method data on stream runoff from areas where no gaging stations exist and for long years of record. Since the method is straightforward to apply it should become a tool of increasing value to hydrologists once the utility of the approach becomes more familiar.

CHAPTER 5. THE MOISTURE STORAGE*

Since precipitation adds to the soil moisture content and evapotranspiration and runoff will subtract from it, it is possible to obtain at any time from knowledge of the balance between these two quantities the actual moisture content of the soil. This can be done on either a daily or longer term basis with the use of appropriate assumptions and procedures.

With the growing realization of the need for soil moisture information, there has been a rapid increase in the last few years in the number of instruments designed to measure soil moisture. In spite of improvements in instrument design and operation, soil moisture instruments do not give results which are very useful for either long- or short-range planning. Even if it were possible to make the necessary representative measurements of soil moisture, and in a few localities fairly good observations have been obtained, the length of the record is not great enough to permit their use in studies of the probability of occurrence of different levels of soil moisture at different times of the year. In time, it may be possible to obtain the necessary long-period records of soil moisture from certain areas in the United States and elsewhere but it will never be possible to have such records from all parts of the world. Thus such information must be obtained indirectly through the relation between soil moisture and other more readily available climatological observations using some type of water balance as described in earlier sections of this report.

To show the type of results which are obtained through such a computational procedure using only climatic data the day-to-day variations in soil moisture have been computed for a number of years at Coshocton, Ohio, and the results have been compared with the actual soil moisture determinations as published by the Soil Conservation Service (figures 5.1 - 5.3). For the type of soil at Coshocton it was found that the 40-inch profile contained a total of about 11.5 inches of water at field capacity. Accordingly the tables were used which give the rate of soil moisture depletion when the total water storage capacity in the soil is 11.5 inches. Ninety-three percent of the available gravitational water was detained in the soil each day. This percentage was determined as a result of observations over a long period of record on that particular soil depth and type.

At Coshocton during the winter there are a number of periods during which the air temperature is well below the freezing point and the ground is frozen to a certain depth. During these periods percolation of gravitational water would be greatly reduced or in some cases stopped entirely until thawing of the soil occurred. It was accepted in making these computations that percolation did not occur on days when the mean temperature of the air was below 30°F.

Considering the assumptions and approximations made in computing the soil moisture on the one hand and the methods of soil moisture determination employed on the other, agreement between the measured and computed values of soil moisture at Coshocton may be considered to be very satisfactory. It is clear that a number of problems still remain to be answered but the results indicate a significant advance over the earlier method of computing soil moisture. Although certain results concerning the relation between the intensity of rainfall and soil characteristics have been obtained, no attempt was made in the foregoing example to determine how much of the rainfall fell at rates above the infiltration capacity of the soil and hence was lost by surface runoff. Once this factor can be worked into the computing procedure some of the periods of disagreement between the measured and computed values of soil moisture such as during September and October, 1945, will probably be reduced or eliminated.

* The material in this chapter has been adapted from "Estimating Soil Tractionability from Climatic Data, Final Report," Publications in Climatology, Vol. VII, No. 3, Centerton, 1954, pp. 397-402; "Remarks on the Contribution of Climatology to Oceanography," by C. W. Thornthwaite, Oceanographic Convocation, Woods Hole, Mass. June 26, 1954, 6 pp. (processed); and "Estimating Soil Tractionability by Climatic Analysis," by C. W. Thornthwaite, Environmental Protection Section, Office of the Quartermaster General, Report No. 167, 1950, 54 pp.

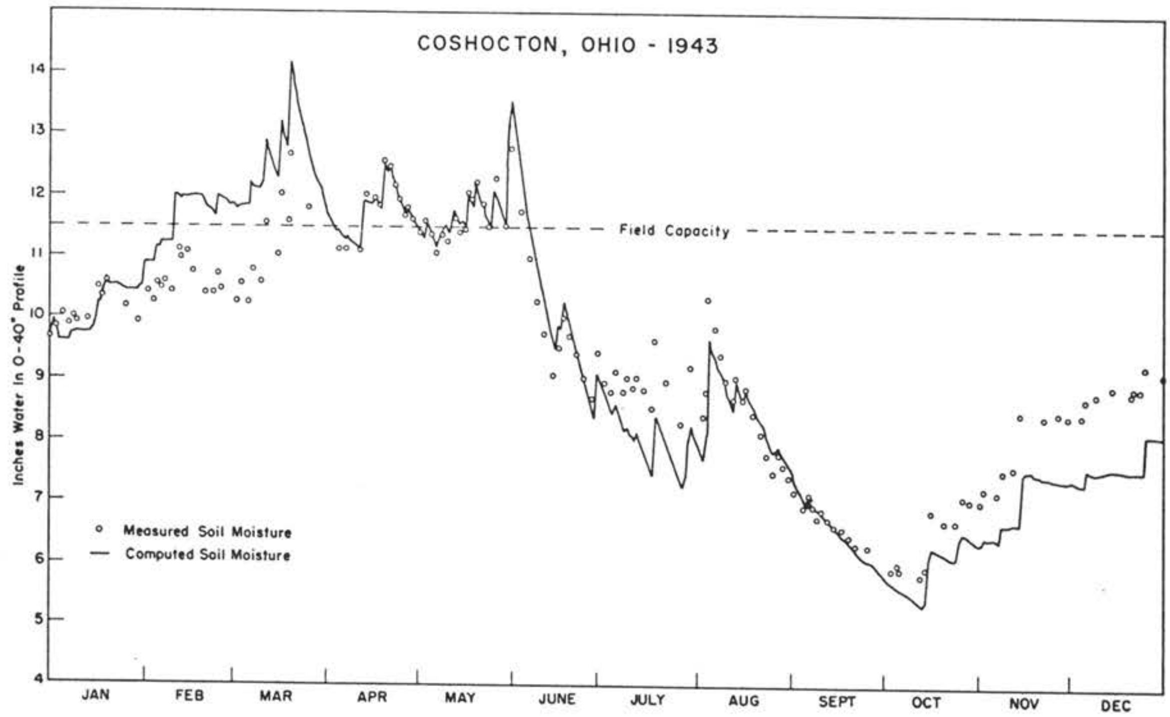


Figure 5.1

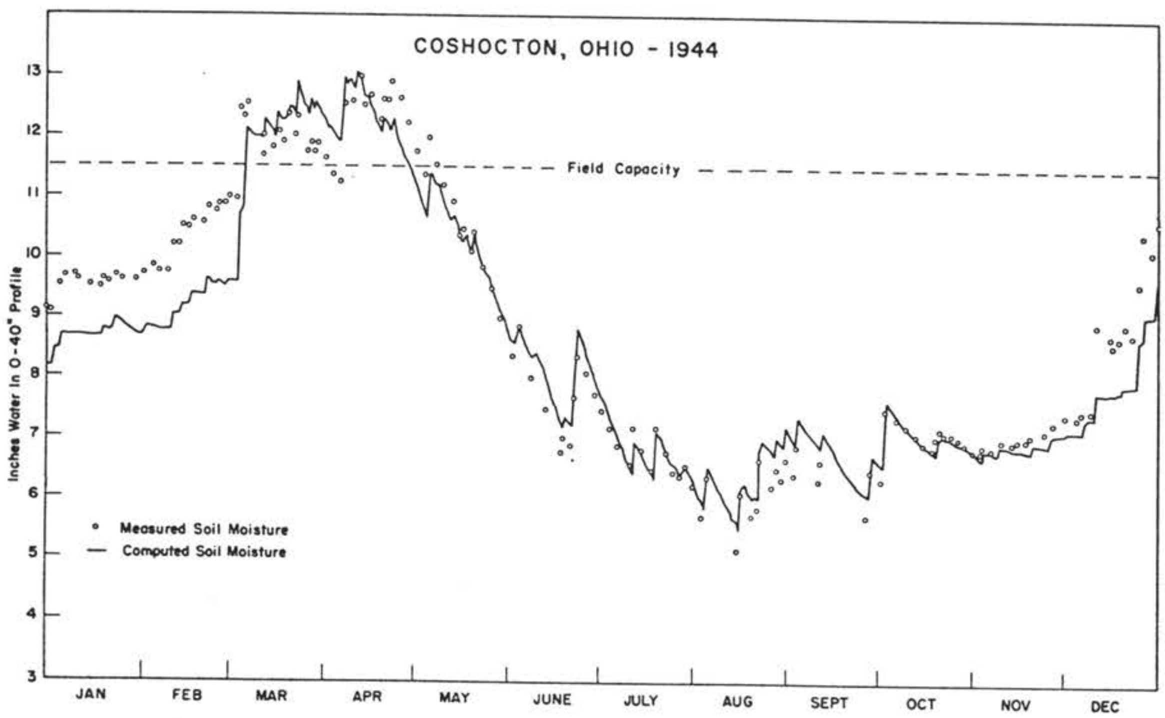


Figure 5.2

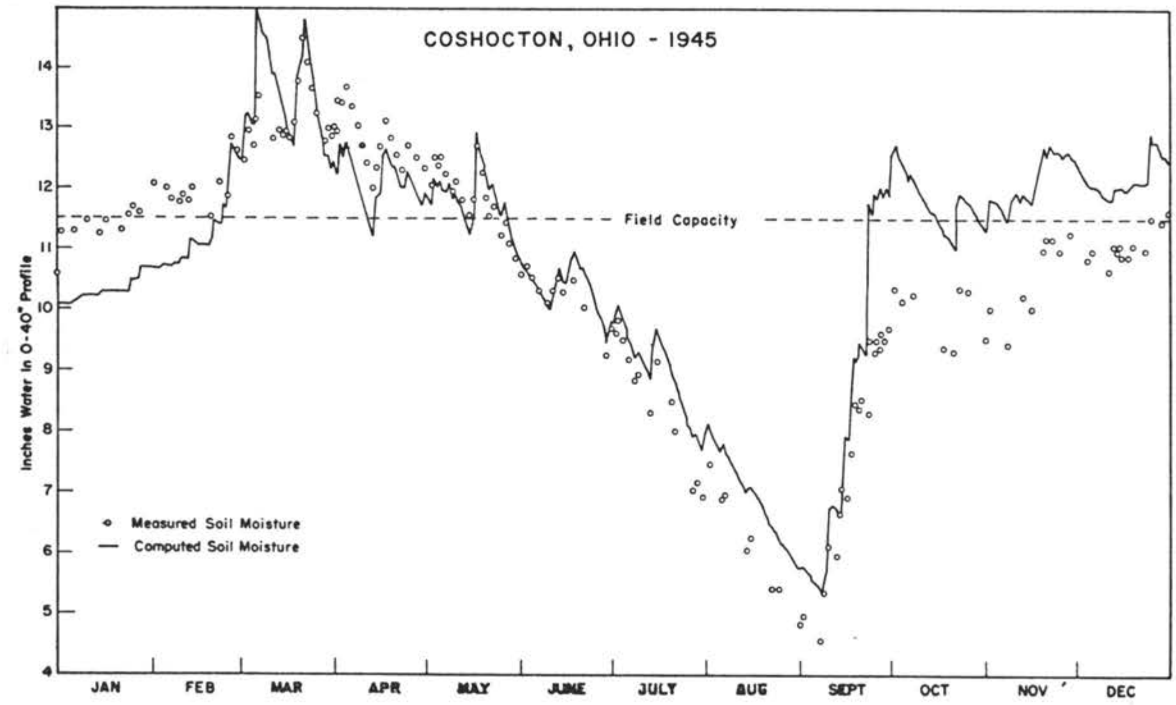


Figure 5.3

Another check of the validity of the computational procedure was obtained by comparing the measured and computed values of soil moisture at College Park, Maryland, in 1942. The results are shown in figure 5.4. Soil samples for moisture determinations were obtained by the Soil Conservation Service at 6-inch and 18-inch depths; the moisture computations are for the upper 24 inches of the soil. This two-foot layer contains 7 inches of water at field capacity. Only 80 percent of the available gravitational water was detained in the soil each day.

In making the computations for other localities with different types of soil and for root zones of different depths it is, of course, necessary to use the appropriate rate of soil moisture depletion determined for the particular amount of water held in the considered depth of soil at field capacity. Thus, it is possible to compute the trend of soil moisture in any type of soil and for any desired thickness of soil cover. Furthermore, the water regimes of individual layers of the soil profile can be determined separately from the climatic data. As an example, at College Park in 1942, if the measured value of soil moisture at the 6-inch depth is considered to be representative of the moisture content of the soil in the 0- to 12-inch depth and the measurement at 18 inches representative of the 12- to 24-inch layer, it is possible to compare the measured soil moisture in these two layers with that which would be computed for each layer using appropriate tables made up for the amount of the soil storage capacity in each case. This has been done and the results are presented in figure 5.4.

Similar computations for other places and other years support the conclusion that soil moisture can be determined with all needed precision from climatological data. It is apparent from the agreement found between measured and computed values that the climatologic approach will permit the accurate determination of the movement of water through soils and the amount of storage in any selected layer in the soil.

From the bookkeeping procedure it is possible to obtain values of the total amount of water temporarily in storage at the end of each month or day. This value, which includes not only the ground water storage but also the amount of water in streams, rivers and ponds, in transit in the soil capillaries or in the process of flowing over the land, is of inestimable value in other fields of research.

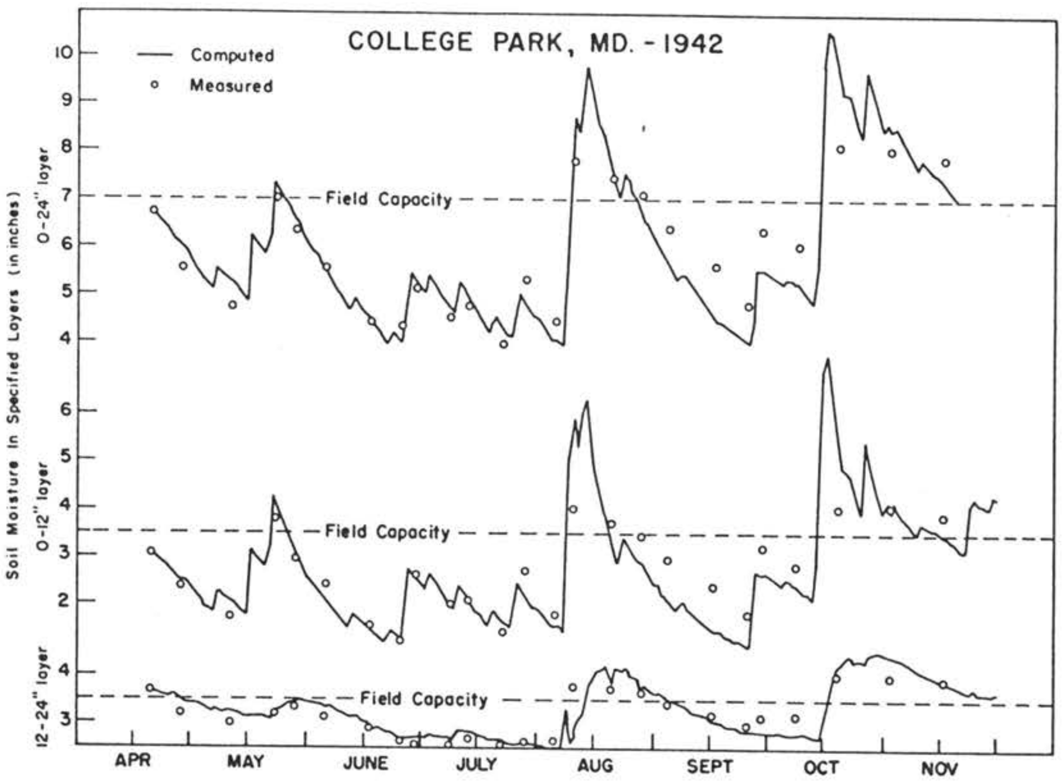


Figure 5.4

Consider first the total water detention at four stations in the United States (table 5.1).

Table 5.1

Total Temporary Water Storage at Selected Stations
(in mm)

	J	F	M	A	M	J	J	A	S	O	N	D
Seabrook	361	376	381	361	331	277	235	214	202	233	284	337
Berkeley	278	329	339	300	250	189	140	105	86	77	96	174
Montgomery	355	388	404	376	320	251	200	160	138	132	190	294
Albuquerque	19	19	18	16	12	8	6	4	3	3	3	11

In Seabrook, New Jersey, for example, at the end of March, 381 mm of water are temporarily trapped in the soil, in the ground water reservoir, and in the streams. At the end of September the water in temporary storage has fallen to 202 mm; the range between maximum and minimum values is 179mm. In Berkeley, the difference between the February maximum and the October minimum is 252 mm. In Montogomery, the range between March and October is 272 mm. In Albuquerque, on the other hand, the precipitation is never great enough in any month to produce a water surplus and consequently no water except a meager amount of soil moisture is held temporarily in storage on the land and the difference between the February maximum and the November minimum is only 16 mm.

These examples introduce an important oceanographic problem that has been a special concern for some time. By studying the records of tide gages from all over the world Munk has found that the oceans contain less water in March than in October. He seems to have lost 5×10^{19} grams of water. There are two possible places to look for this missing water; the atmosphere and the land. By extending the bookkeeping methods used on the individual stations, over the whole land area of the earth it becomes possible to determine how much water is in temporary storage on the land and in transit over it.

In the Northern Hemisphere the greatest supply of water is locked up in the land in spring and it reaches the minimum level in autumn. This is confirmed by the data of Seabrook, Berkeley, and Montgomery. Although there are local exceptions, it is generally true that there is a maximum of water locked up in the land in the form of snow, soil moisture, ground water and stream detention storage in March, and a minimum in October, just as Dr. Munk would like to have it.

But in the Southern Hemisphere the seasons are reversed and so are the months of maximum and minimum water storage in the land. However, the land area of the Southern Hemisphere is so much smaller than the Northern that the conditions in the Northern Hemisphere prevail. It would seem that the missing water has been found, but it remains to complete the inventory country by country to see if the answer that Dr. Munk requires is found.

The tractionability of the material over which vehicles move is determined by four characteristics of the material: bearing capacity, shearing strength, surface friction coefficient, and stickiness. Thus tractionability includes far more than "trafficability", a loose term which applies generally only to the first of these four characteristics.

The bearing capacity of a surface is the load per unit area which the underlying materials can support without being crushed or without settling enough to impede movement. Concrete pavement, rock and dry soil have a high bearing capacity. Concrete can carry heavy loads without being crushed and practically without any permanent deformation or settling. Even the elastic deformation of a concrete pavement, that is the compression shrinkage which is fully restored when the load is removed, is so small that it can be observed only with precise strain gages. Freshly fallen snow on the other hand, has a very low crushing strength and is entirely inelastic. It can support only a small load per unit area and deforms permanently under small loads. Snowshoes and skis are used to distribute weight over a large area which, when multiplied by the low strength per unit area of the snow, will be sufficient to support the load. Heavy tracked vehicles can travel over soft ground because their weight is distributed over a large area.

Shearing strength is the resistance of the material, whether pavement, soil, snow, or ice, to tangential force, as exerted on it by wheels of power-driven vehicles or by the hoofs of animals and feet of men. In a power-driven vehicle, the propelling force is transmitted to the soil or pavement either through lugs or through friction between the tire and the surface. The lug or the tire of the power-driven vehicle, or the hoof or foot of a draft animal or of a man, engages a section of the material underlying the traction surface of some depth, width and length which depend on the material. The external force applied to this section which tends to shear it off at the bottom and on the sides is counteracted by its shearing strength which is due to cohesion and internal friction between the elementary particles of the material. When the shearing strength is exceeded, the section of the material is severed from its position, compressed somewhat and slipped out or ejected from under the foot, hoof or wheel. The action just described takes place in greater or lesser degree on all traction surfaces. It is pronounced on wet plastic soils and particularly apparent on dry, relatively deep sands which have a very low shearing strength although they possess appreciable bearing power when confined.

Surface friction is the resistance to relative motion of two bodies in contact. It is determined by the character of the surfaces of the bodies and by the pressure that holds them in contact. Surface friction varies directly with the load on the wheel, track, hoof or foot in contact with the traction surface. The coefficient of friction between two surfaces is the ratio of the force required to move one over the other to the total force pressing the two together.

Stickiness is a property of some soils and some kinds of snow causing it to stick to wheels, hoofs and feet and to "ball up", making movement of vehicles, animals and men quite difficult.

Soils vary widely in bearing power, shearing strength, surface friction, and stickiness, and thus in tractionability.

The one thing common to all soils is the effect of moisture content on these reactions. Not all soils respond alike or to the same extent or even in the same direction, but in all soils moisture content is the determining factor. Soil moisture is largely determined by precipitation which adds water and evapotranspiration which removes it. Since evapotranspiration is a function of temperature and length of day, the basic factors of tractionability common to all soils are climatic.

With respect to their response to moisture and therefore to tractionability, soils can be divided into two main groups, plastic and cohesive and nonplastic or cohesionless. Plastic soils can be molded when wet and on drying firmly retain the shape imparted to them by molding. Nonplastic soils will not retain a molded form on drying. If water is added when plastic soils are being molded, a point is reached at which the soil begins to stick to the fingers. Up to a certain point, the stickiness increases with an increase in moisture; it ceases at saturation when free water appears at or near the surface. Stickiness is a property of plastic soils only.

Bearing capacity and shearing strength of plastic soils are closely interdependent: both decrease with increase in soil moisture. Conversely, in nonplastic soils both bearing capacity and shearing strength increase with moisture content; in a dry state their shearing strength is very low while the bearing capacity remains fairly high, especially when material is confined.

The percentage of clay in a soil determines whether it is plastic or nonplastic. A washed screened sand from which all of the clay has been removed will not become muddy. When a soil mixture is largely sand with minute amounts of clay, the reaction is characteristically nonplastic; when about 15 percent of clay is present a soil begins to react as a plastic material. A clay content of 15 percent has been used as the dividing line between plastic and nonplastic soils.

Several moisture constants which may be related to tractionability have been established by usage. They are:

a. Wilting Coefficient - The moisture content below which water cannot be withdrawn from the soil by plants. At moisture contents below the wilting point, the shearing strength of nonplastic soils becomes very low and the bearing and shearing strengths of plastic soils are high.

b. Field Capacity - The moisture which a soil can retain against gravity, or the moisture content of a soil after the excess free water has drained out. It is also referred to as maximum capillary water and for many soils is the same or nearly the same as the "water equivalent". Tractionability of nonplastic soils is greatly improved at and beyond this point.

c. Upper Plastic Limit (or liquid limit) - The moisture content at which plastic soils flow when tapped. A close linear relationship exists between the upper plastic limit and the moisture equivalent or field capacity. At this moisture content plastic soils lose practically all of their bearing power and shearing strength.

d. Water Holding Capacity - The moisture content at which all pores in the soils are filled with water; also referred to as saturation and occurs after prolonged rains before the gravitational "free" water has time to drain out. At this moisture content, nonplastic soils behave much like plastic soils at the upper plastic limit and plastic soils lose their stickiness and become slippery.

Appreciating the value of the method of estimating soil moisture from climatic data and recognizing the need for its improvement and extension for the purposes of determining the ability of men and vehicles to move over unimproved surfaces, the Geophysics Research Division of the Air Force Cambridge Research Center in 1951 asked the Laboratory of Climatology to prepare a proposal for undertaking such work. The proposal was approved and work began in January, 1952, on a contract directed toward the study of "Research in Forecasting Soil Tractionability". A one year extension of that research contract was undertaken in January 1955 although it was clear that significant results would not be forthcoming in such a limited period of time.

The concept of forecasting as applied to soil moisture and tractionability has two aspects -- the tactical and the strategic. A typical tactical problem would be to forecast how soon after a soaking rain which resulted in poor tractionability would conditions improve sufficiently to permit certain specific operations to be carried out. Such a forecast would require, in turn, a reliable and rather specific forecast of amounts and intensities of rainfall or of rates and amounts of snowmelt, and possibly of rates of thawing of frozen ground, as well as of temperature over a period of several days. With our present knowledge such forecasts for several days in advance are inaccurate. But even if such forecasts were possible, the present methods of estimating soil moisture from climatic data and water properties of soils are as yet not adequate to insure the order of reliability that would be expected in the solution of specific tactical problems.

The strategic aspect is quite different. A typical strategic problem might be to determine the probability of good, poor, or satisfactory tractionability at a given period of time (season, month, or even week). Here it would be more practical to utilize past records rather than forecasts to determine values of tractionability for different periods during a number of past years and their probabilities of occurrence.

In order to adapt the work already completed on the water balance bookkeeping procedure to the needs of tractionability it is necessary to convert the information on the total detention of water in the whole soil profile into figures of moisture detention in the active layer for tractionability - about a 2-foot depth of soil. This problem was resolved by determining the distribution of moisture content in the upper two feet of a soil profile as a function of the total moisture content of a 4-foot profile, the depth used in evaluating the bulk of the data for other purposes. To do this involved a study of a large number of soil moisture profiles from different soil and vegetation environments and resulted in the selection of average soil moisture profiles which express the distribution of moisture vertically in the soil as a function of different total values of soil moisture content.

Figures 5.5 and 5.6 show typical moisture profiles in both cultivated and noncultivated areas. Under cultivated crops, with an appreciable amount of bare soil the soil moisture profile often shows a pronounced drying of the surface soil layers with very little reduction in moisture content in the deeper layers. On the other hand, under a field of well established vegetation which is not cultivated the moisture content of a much deeper layer of soil is reduced because of the action of the roots. Only in the deepest layer is the moisture content little affected by the evapotranspiration.

When these two types of moisture profiles are averaged, to obtain a hypothetical moisture profile which might best represent both extremes one obtains essentially a straight line relation between moisture content and depth. As the soil dries the profile becomes less and less steep for the drying extends over a greater and greater depth of soil.

Once the typical moisture profiles have been selected it is possible to relate the moisture in the depth for tractionability to the moisture content in the total profile used in the water balance bookkeeping procedure. It is also possible to select certain of the moisture profiles which seem to represent significant conditions and to use the values of total moisture content which these profiles indicate in defining the numerical boundaries which will serve in the analysis of the moisture data.

SOIL MOISTURE PROFILES UNDER CULTIVATED AND FORESTED CONDITIONS

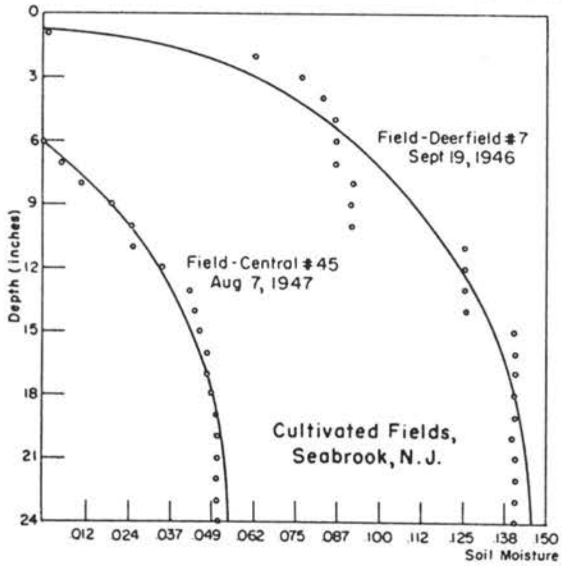


Figure 5.5

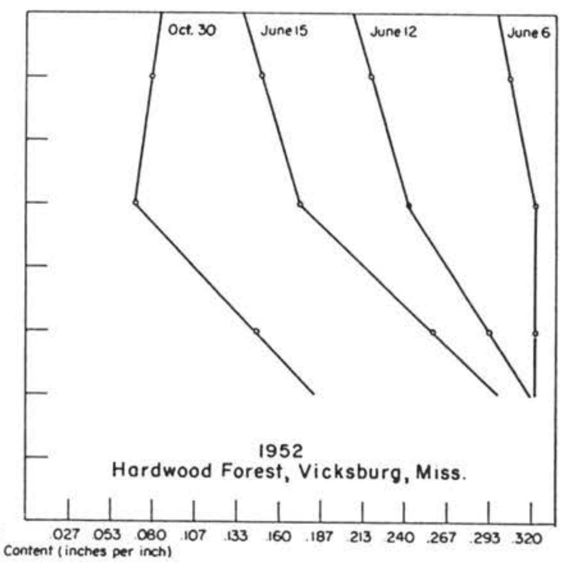


Figure 5.6

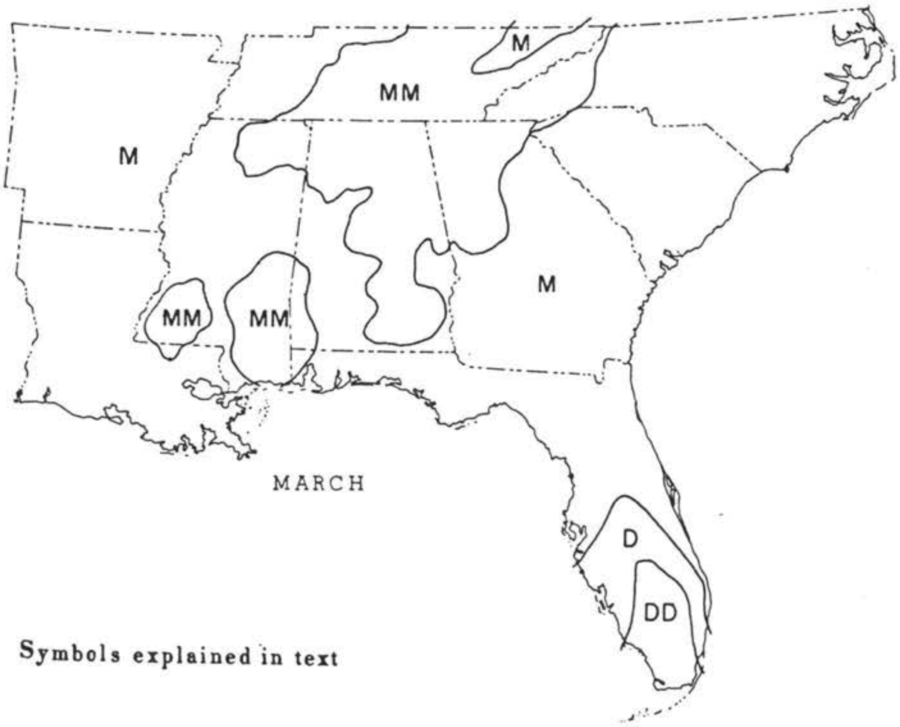
On non-frozen soils five mutually exclusive classes of ground conditions, based on the soil moisture content in the significant soil depth for tractionability, two feet, have been recognized and mapped. A number of additional classes of ground conditions exist on frozen soil, including such classes as deep snow, permafrost, ice caps, and glaciers, but a discussion of the types which occur under frozen conditions is beyond the scope of this report.

In appraising the significance of the several classes of ground conditions, the important influences of exposure and topography on soil moisture must always be kept in mind. As finally mapped, the moisture classes would refer to average conditions in an area. Steep slopes and southern exposures would be represented by the lower soil moisture included in the range for the class while the upper values within the range would apply to depressions with impaired drainage and to northern exposures.

The five classes of ground conditions which have been mapped over southeastern United States in figure 5.7 along with an interpretation of their significance in terms of tractionability are described below. The moisture percentages used are only approximate.

> DD (Very dry) - The average soil moisture content of the two foot depth for tractionability is less than 33 percent of the moisture content at field capacity. The moisture content at the soil surface is 0 while at the depth of two feet it can vary between 0 and 66 percent of field capacity. Within the range of soil moisture covered by DD the tractionability of plastic soils is excellent. All parts of the relief with plastic soils except swampy areas are sufficiently hard and firm to support vehicles and transport of all kinds. When soil moisture remains in this class for extended periods, heavy traffic may grind the surface to fine dust which may impede fast moving equipment. The tractionability of sandy soils and particularly of deep shifting sands is very poor except in seeps and depressions where some moisture may have accumulated.

AVERAGE GROUND MOISTURE CONDITIONS



Symbols explained in text

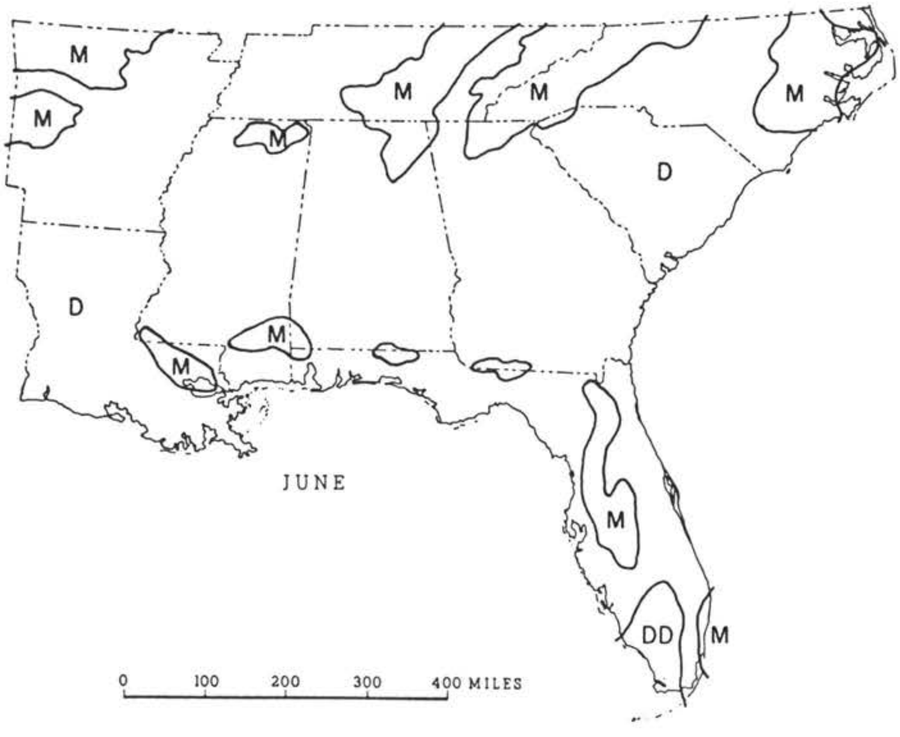


Figure 5.7

IN SOUTHEASTERN UNITED STATES

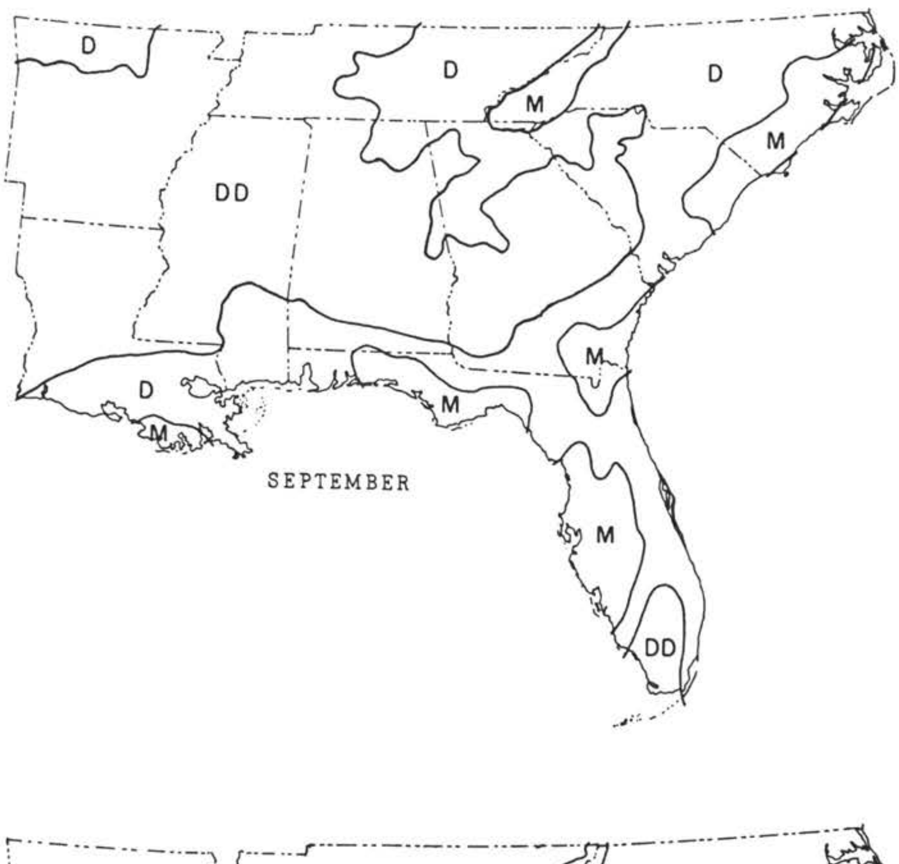


Figure 5.7 (cont.)

D (Dry) - The average moisture content of the two foot layer for tractionability is between 33 and 75 percent of the moisture content at field capacity. At the soil surface the moisture content ranges from 0 to 50 percent of field capacity while at the two foot depth it ranges from 66 percent of field capacity up to field capacity itself. Within this class of ground conditions the tractionability of plastic soils remains high and that of sand remains poor. Plastic soils tend to soften somewhat but not enough to impede movement of even the heaviest equipment. No appreciable improvement is noted in the tractionability of sandy soils except immediately after heavy showers. The surface of plastic soils becomes definitely slippery during and just after such showers.

M (Moist) - The moisture content of the two foot layer of soil for tractionability is between 75 and 115 percent of field capacity. At the soil surface the moisture content will vary from 50 percent of field capacity to field capacity while at a depth of two feet it will vary from field capacity to 133 percent of field capacity. Within the range of moisture represented by M conditions the tractionability of plastic soils deteriorates rapidly. The soil becomes softer and offers considerable difficulty for heavy wheeled vehicles. Track laying vehicles have no trouble except after heavy showers when the surface is muddy and slippery. Tractionability of sands and other non-plastic soils is good and continues to improve with additional moisture.

MM (Very moist) - The average soil moisture content in the top two feet ranges from 115 to 155 percent of field capacity. At the soil surface the moisture content ranges from field capacity to the liquid limit (about 150 percent of field capacity) while at a depth of two feet the moisture ranges from 133 to 167 percent of field capacity. MM conditions represent very poor tractionability on plastic soils. Adhesion of plastic soils to foreign objects becomes greater than the cohesion between soil particles. Traffic of practically every kind is extremely difficult on bare soil but light animal-drawn vehicles and men can travel over grassed areas without too much difficulty until the surface is cut up. The tractionability of sands and non-plastic sandy soils is excellent in this class of soil moisture.

W (Wet) - The average moisture content in the top two feet of soil ranges from 155 to maximum water holding capacity (about 200 percent of field capacity). At the surface of the soil the moisture content is between the liquid limit and the maximum water holding capacity while at a depth of two feet it ranges between 167 percent of field capacity and the maximum water holding capacity. W conditions represent a range of moisture content where free water begins to appear at or near the surface. At this point the stickiness of plastic soils is greatly diminished but at the same time the bearing strength and friction coefficient are so greatly reduced that the movement of power driven transport is nearly impossible. The tractionability of most sandy soils remains fair although this is the moisture range in which sands with certain grain size exhibit the characteristics of quicksand.

Over southeastern United States (figure 5.7) in winter MM conditions are found in a belt through eastern and central Tennessee and northern Georgia, Alabama and Mississippi. Isolated patches of MM conditions occur in southern Mississippi and Alabama. The rest of the southeastern part of the country experiences M conditions except for southern Florida where the low precipitation and high evapotranspiration still results in D and DD conditions.

In June the increased evapotranspiration results in the occurrence of D conditions over most of the southeast with the exception of isolated areas of M conditions in the mountains of North Carolina and eastern Tennessee and along the Gulf Coast. These conditions of dryness intensify through the summer so that by September a wide band of DD soil conditions are found through the mid-south. Only in the mountainous regions and at isolated places along the Gulf coast is there sufficient precipitation to result in M soil moisture conditions.

With the decrease in evapotranspiration in fall the soil moisture content increases and by December most of the mid-south area is once more M. A small band of D conditions is found through south-central Georgia and Alabama and again in southern Florida while very moist (MM) conditions exist in the mountains of Tennessee and North Carolina.

In addition to maps of the geographic variation of soil moisture values by months, it would be most desirable to determine the expectancies of different values of soil moisture through the year for a large number of places so that maps or tables of the geographic distribution of the expectancies of different levels of soil moisture might be prepared. The ideal presentation would be maps or tables giving expectancies for each week or even for each day of the year. This ideal is practically unattainable because the climatic records that are available for many parts of the world lack the necessary detail and are for too limited periods.

Using the present computing procedure and the most readily available data, it appears that to achieve maximum uniformity in the presentation of the data it would be possible to use only monthly weather conditions in the computation of soil moisture at a place.

A twenty year record of monthly values of soil moisture has been obtained from the climatic data at two stations in southeastern United States. Using the graphical method of determining probabilities which was developed by Hazen in his work on flood flows, it is possible to graph on arithmetic probability paper the twenty values of soil moisture content at the end of each month. From such probability graphs, the soil moisture contents which can be expected to be equalled or exceeded different percentages of the time at two different stations have been determined. The seasonal courses of different probabilities of soil moisture based on 12 separate determinations, one at the end of each month are given in figures 5.8 and 5.9. The distributions are only approximate since the determination of probabilities for each day of the year and the use of a longer period of record would result in certain slight changes in the patterns shown.

Plastic soils should be tractionable if the moisture content in the active layer for tractionability is 140 mm or less. To insure success of a operation it is desired that the chances of having a soil moisture content above this value are 1 in 10 or less. On figures 5.8 and 5.9 it is seen that the 10 percent probability line intersects the 140 mm soil moisture line on July 5 and November 28 for Marked Tree, Arkansas, and June 17 and December 26 for Louisville, Kentucky. During those periods it might be expected that the soil moisture content would be below the critical value more than 90 percent of the time. Of course, local heavy showers would bring the value of soil moisture in the top layer temporarily above the critical value but this effect would be transitory. It must be remembered that the probabilities shown on figures 5.8 and 5.9 have been determined from a limited past record and there is no assurance that future conditions will be exactly similar.

Data for a number of days or periods through the year presented in a form similar to figures 5.8 and 5.9 for all regions would be of inestimable value in planning military or civilian operations. They provide the only available source of data which can be used in any attempt to forecast future soil moisture conditions or to explain past situations in areas where measured values are not available.

The present work on tractionability has been directed toward a refining of some of the concepts and assumptions employed in the general water balance bookkeeping procedure in an effort to develop a method which would be of utility in the solution of both tactical and strategic military problems. The work has been concerned more with physical principles and with techniques and procedures than it has been with large-scale mapping or statistical summaries of results. It is felt that these last

EXPECTANCIES OF SOIL MOISTURE IN TWO FOOT DEPTH OF SOIL
MARKED TREE, ARKANSAS

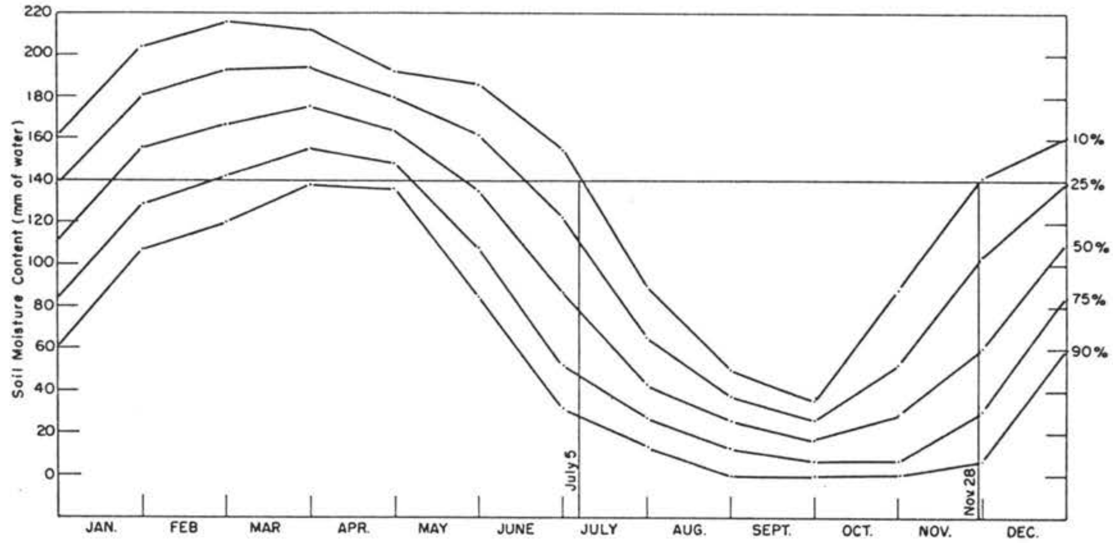


Figure 5.8

EXPECTANCIES OF SOIL MOISTURE IN TWO FOOT DEPTH OF SOIL
LOUISVILLE, KENTUCKY

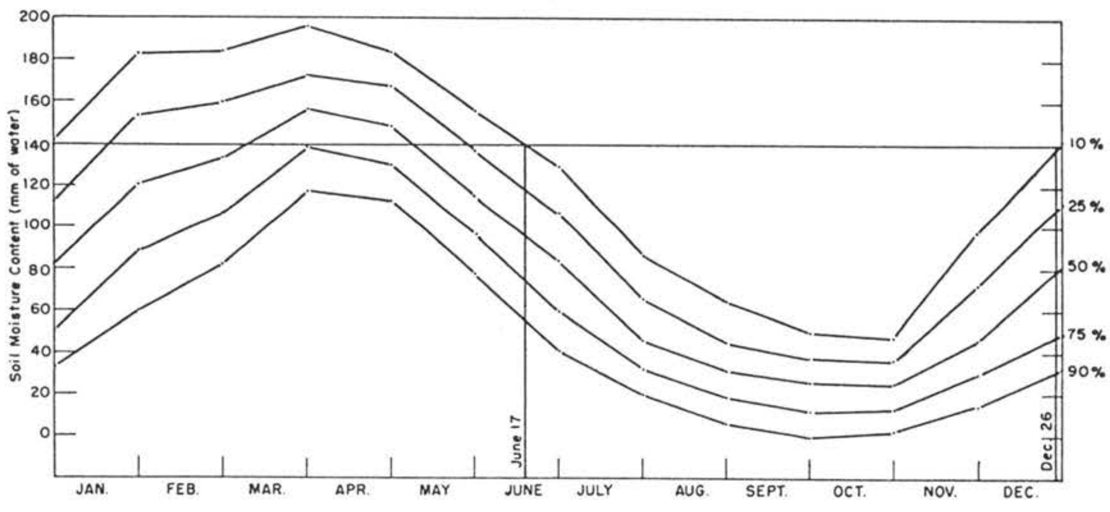


Figure 5.9

quantities can be achieved at any time once the underlying concepts have been fully understood and expressed. As a result of the work to date we are achieving at present a much more complete understanding of the utility of the whole water balance not only as it applies to tractionability problems but also as it can be utilized in other lines of research.

CHAPTER 6. MOISTURE INDICES*

Climates, like nearly all other variable quantities, can be classified. The difficulty of rationally classifying a natural complex such as soil or vegetation is clear; yet we know that classification is necessary, and we know that it gropes toward reality. It is likewise with climate. Faced with complex geographical variation of climate, we seek to recognize climatic types and their geographical equivalent, climatic regions. The purpose of a climatic classification is to provide a concise description of the various climatic types in terms of the truly active factors of climate, primarily those of moisture and heat.

The Asia-Pacific Forestry Commission, at its second session in Singapore, December 1 - 13, 1952, asked the World Meteorological Organization, through the Director of the Forestry Division of FAO, to recommend a standard scheme of climatic classification . . . "based on the relationships between climate and vegetation . . . . Such a classification would be extremely useful to forestry agencies of our member governments as a guide for the choice of exotic species for trial plantings."

In 1954 the 8th General Conference of UNESCO in Montevideo discussed a program for research in the "humid tropics". The proposed program for 1955 and 1956 included the following recommendation: "Maps delimiting the humid tropical areas and showing those areas which experience similar climatic conditions are considered to be a prerequisite for the work of the programme." This interest by different agencies within the United Nations organization is but the reflection of a growing awareness in all countries of the need for accurate and detailed climatic information.

It is doubtful whether any classification that now exists deserves to be adopted as "standard". It may be a long time before anything approaching a standard can be achieved.

The earliest attempts at climatic classification in the modern sense were made in the mid-nineteenth century, chiefly by natural historians and biologists. From the very beginning the initiative in such classification has rested with the biologists, and it has been from these students of the living cover of the earth that most of the actual systems of classification have come.

In 1866, A. Grisebach published the earliest significant world map of vegetation regions. Coming as it did shortly after the great plant physiologist, Alphonse de Candolle, had published his massive investigations into the factors influencing the distribution of plant species, Grisebach's attempt at the classification of world vegetation was a challenge to students of climate to look for correspondences between temperature or precipitation data and the world distribution of vegetation. At the same time, among biologists there was extensive investigation of the effect of climate on the phenology, growth and development of plant species. The monographs of Carl Linsser are the outstanding monument to this phase; his work, concerned with the effects of temperature on phenology and rainfall on vegetation, led him to divide the world into climatic zones, thereby making himself the first to attempt a true classification of climates on the basis of vegetation zones.

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* The material in this chapter has been adapted from "Climatic Classification in Forestry," by C. W. Thornthwaite and F. Kenneth Hare, Unasylva, Vol. 9, No. 2, June 1955, pp. 50-59; "Grassland Climates," by C. W. Thornthwaite, Publications in Climatology, Vol. V, No. 6, Seabrook, 1952, 15 pp.; "An Approach Toward a Rational Classification of Climate," by C. W. Thornthwaite, Geogr. Rev., Vol. 38, No. 1, 1948, pp. 55-94; "Climate in Relation to Crops," by C. W. Thornthwaite and J. R. Mather, Meteorological Monographs, Vol. 2, No. 8, 1954, pp. 1-10; and "Climates of Africa and India According to Thornthwaite's 1948 Classification," by D. B. Carter, Publications in Climatology, Vol. VII, No. 4, Seabrook, 1954, pp. 453-474.

Thus by 1975 the idea that climates might be classified according to the type of vegetation or physiological response they produced was well established. But it was left to a St. Petersburg trained biologist, Wladimir Köppen, to take this idea and elevate it into a primacy of place that it has never lost. Köppen's interest carried him into climatology, which he dominated for some sixty years, and his climatic classification continues in use even today among geographers.

In spite of the very wide currency it has achieved, the Köppen system is unsatisfactory. It is a very crude, blunt instrument; the regions it defines are large and unwieldy, and do not correspond to major divisions of the vegetation of the earth. The fact that a particular isopleth of mean air temperature happens to follow a soil or vegetation boundary is to a large extent fortuitous. Any effective system of classification must endeavor to answer the question - what are the real, active processes of climatic control? And how can we devise suitable parameters for these processes? These are the questions that the Thornthwaite climatic classification of 1948 attempts to answer.

A moisture index, which is obtained by comparing the water need at a place with the moisture surplus and deficit, is an essential part of the classification. Where precipitation is exactly the same as potential evapotranspiration all the time and water is available just as needed, there is neither water deficiency nor water excess, and the climate is neither moist nor dry. As water deficiency becomes larger with respect to potential evapotranspiration, the climate becomes more arid; as water surplus becomes larger, the climate becomes more humid. Where there is a water surplus and no water deficiency, the relation between water surplus and water need constitutes an index of humidity. Similarly, where there is a water deficiency and no surplus, the ratio between water deficiency and water need constitutes an index of aridity. Since water surplus and water deficiency occur at different seasons in most places, both must enter into a moisture index, the one affecting it positively, the other negatively.

The distribution of the moisture regions of the United States, based on this moisture index, is shown in figure 6.1. The moist climates of the East and the dry climates of the West are separated by moisture index 0, in a line that extends from western Minnesota southward to the Gulf of Mexico. This is a very important line,

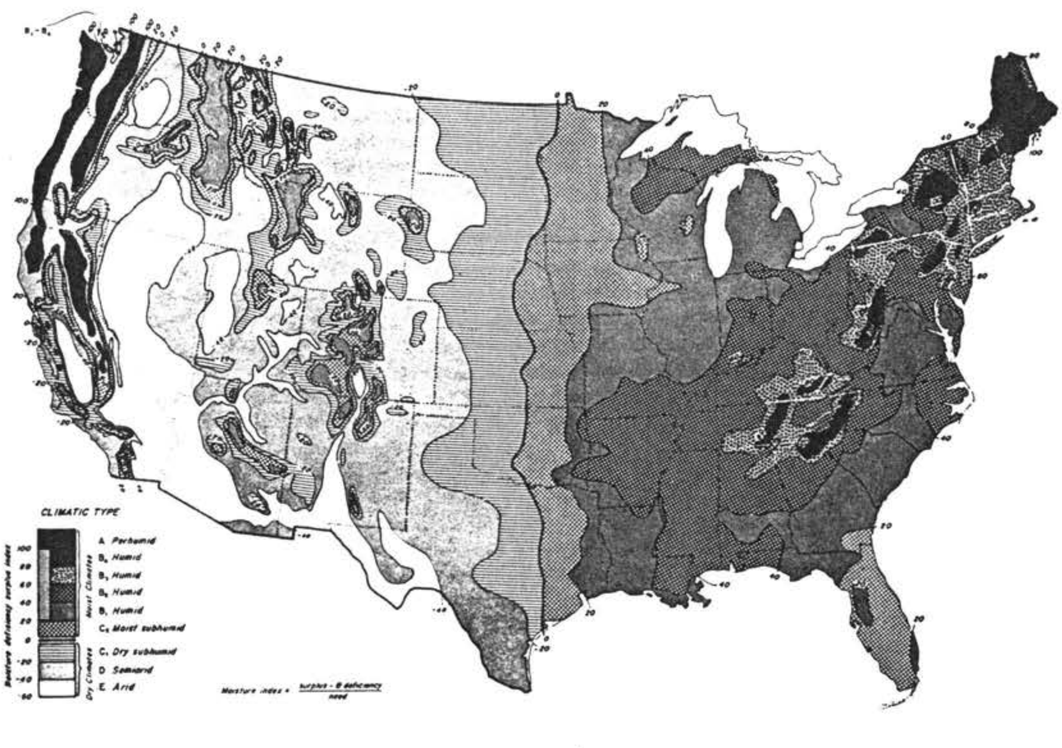

Figure 6.1

since it separates regions of prevailing water surplus from those of water deficiency. Although dry climates predominate in western United States, there is a belt of moist climates along the Pacific coast as far south as the San Francisco peninsula. Moist climates appear also as islands in the western mountains.

Perhumid climates are not extensive in the United States. They occur along the coasts of Washington, Oregon, and northern California and on the western slopes of the Cascades and the Sierras. There is a single small island in the high Rockies of Colorado, and other small islands on high elevations in the Appalachians. A narrow belt along the Maine coast is perhumid. Humid climates are most extensive in the East, but they also occur adjacent to perhumid climates on the Pacific coast and on high areas elsewhere in the West. Subhumid climates are most extensive in the Middle West, wide belts of moist subhumid and dry subhumid extending from the Canadian border in Minnesota and North Dakota to the Gulf coast of Texas. Much of the Florida peninsula is moist subhumid. Smaller subhumid areas occur in the West, mostly as belts along the lower mountain slopes. The Great Plains and much of the intermountain region are semiarid. Arid climate is most extensive in the Southwest. There are, however, small arid areas in Washington, Oregon, Idaho, Montana, Wyoming, Utah, and Colorado.

Studies have shown that the humidity and aridity indices can be used to differentiate the climates of the grasslands and steppes from those of the forest and deserts. Expressed as percentages the two indices are:

$$I_h = \frac{100\ s}{n} \qquad \text{and} \qquad I_a = \frac{100\ d}{n}$$

where s is water surplus, d is water deficiency, and n is water need. These indices are represented graphically in figure 6.2 for a number of stations across the United States near the 41st parallel of latitude. The stations are from different vegetation zones and show how vegetation is influenced by water surplus and deficit. In the eastern part of the country, where the humidity index is 50 or more and the aridity index is practically 0, oak-chestnut forests comprise the natural vegetation complex. Further west through Ohio, Indiana, and Illinois, the humidity index falls to

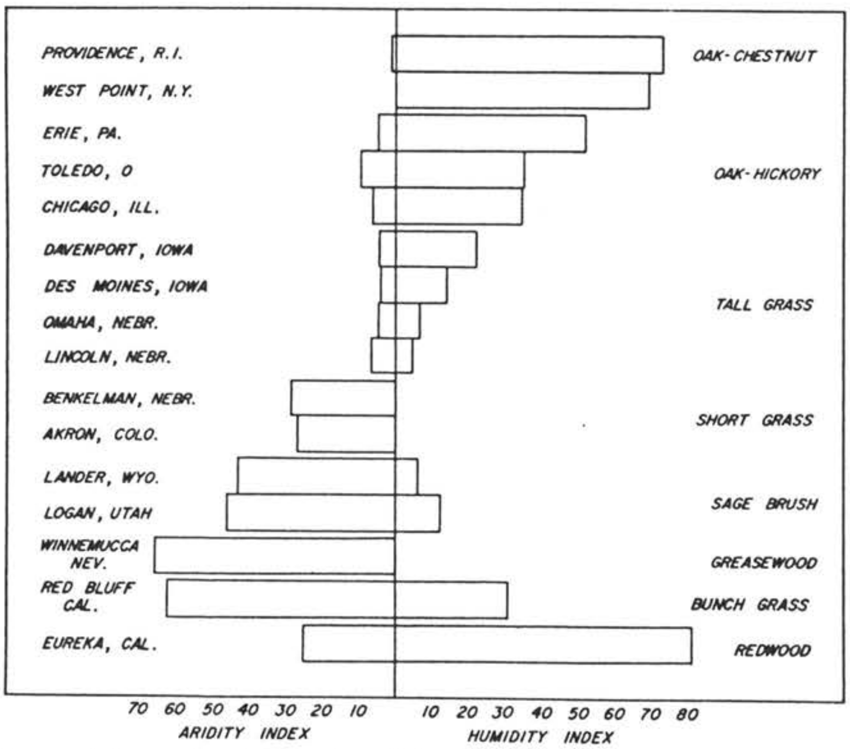

Figure 6.2 - Humidity and aridity index at selected stations near latitude 41°N across the United States.

30 to 40 while the aridity index rises to 5 to 10. Here the climax natural vegetation is oak-hickory. In Illinois, the vegetation is transitional between woodland and grassland. This is most clearly seen if one compares the river valley areas with the interfluves. On the higher, level interfluve areas, grasslands prevail. In the valleys, with water draining in from the nearby slopes and ridges in addition to the natural precipitation, soil moisture is increased and woodland vegetation prevails. Thus, topographic features, through their influence on soil moisture will play an important role in the determination of the natural vegetation.

As one moves farther westward through Iowa and Nebraska, the humidity index decreases from 20 to about 5 and the aridity index remains about 5 to 10. In this region, there is a good balance between water surplus and deficiency. Both are small and if one considers the annual distribution of the periods of surplus and deficit it is usually found that neither continues for more than a month or two at a time during the year. This is the region of the tall grass prairies. Again in the valleys where soil moisture is augmented by seepage flow or there is natural subsurface irrigation, woodland vegetation occurs. Certain drainage conditions in the loess soils and the sand hills of Nebraska may locally result in unexpected vegetation associations but here as elsewhere the vegetation conforms to the soil moisture pattern. Thus, soil and geologic conditions as well as topography can influence the vegetation through their effect on soil moisture.

The line of the 41st parallel next crosses the Great Plains. On a wide belt extending from Canada to Mexico east of the Rockies, the average water surplus is zero. Only in exceptionally rainy periods is there any water surplus to add to ground water and to produce surface runoff. With local sources of runoff virtually absent there are extensive areas without surface drainage features except for the through-flowing streams that originate in the mountains. These areas are flat, unbroken and treeless. Dust, which originates in the arid areas to the west, is caught by the grassy vegetation and accumulates to produce the flat surface. The surface is protected and preserved by the prairie sod. The flat surface in turn prevents the development of springs and seeps so that there is no local concentration of ground water to encourage the development of trees and shrubs. Thus, in the grasslands, vegetation and surface are mutually dependent upon each other.

In western Nebraska and Colorado, the humidity index becomes essentially 0 while the aridity index rises to 20 to 30. The decrease in soil moisture indicated by this shift in the indices results in a change from tall grass to short grass prairies. As the aridity index increases without a corresponding increase in the humidity index through Wyoming, Utah, and Nevada, the short grass prairies are replaced by a sagebrush and greasewood type of desert vegetation. Interestingly, in central California where a large aridity index exists in conjunction with a humidity index of about 30, the soil moisture regime permits the development of a bunch grass type of semi-desert vegetation. Here, with short periods of water surplus and long periods of deficit tall and short grass prairies do not occur. On the California coast, the humidity index again increases to almost four times the aridity index and a forest vegetation association results.

Figure 6.2 shows that in regions having humidity indices over 30 to 35 with no appreciable aridity index, resulting in high values of soil moisture in nearly all months of the year, forests predominate. Where water surplus and deficit are small and almost balance, tall grass prairies exist. With longer periods of low soil moisture, short grass replaces the tall grass prairies. In the regions of almost continuous deficit only the widely spaced desert vegetation of sagebrush and greasewood can survive the periods of low soil moisture content.

The evidence is unmistakable that there is a definite grassland climate. It is marked by a specific balance between moisture deficit and surplus which results in few periods during the year of either very dry or moist soil conditions. As a result of this environment grasses flourish. It is possible to explain the presence of grasslands on the basis of climate alone, without bringing in other agencies such as man or fire. Grasslands are a natural phenomenon, in equilibrium with their climatic environment. Thus, through climatic information it is possible to differentiate the grasslands in general and even the small scale variations in different grasslands within the broad group. This must lead to a more rational interpretation of the vegetation associations of the world.

The task of correlating the indices of the Thornthwaite classification with observed vegetation distribution and with soils over the earth has hardly begun. Eastern United States is made up almost entirely of humid climate and together with southeastern Canada, comprises one of the largest humid regions in the world. Within this region several distinct forest types have developed, in part, due to variations in the thermal index but largely because of variations in the moisture index. The humid climate exhibits a wide range in effective moisture between its subhumid and perhumid boundaries and it has accordingly been divided into four subregions. Within each of these subdivisions there is a northern and a southern zone. The forest types associated with these climatic regions are approximately as in table 6.1.

Table 6.1

Moisture Subdivisions of Humid Climate, Moisture Indices, and Characteristic Forest Types in Northern and Southern Zones in the United States

Climate	Moisture Index	Northern Zone	Southern Zone
B_4	80 - 99.9	Spruce - fir	(absent in eastern U.S.)
B_3	60 - 79.9	Birch - beech - maple - hemlock	Oak - chestnut
B_2	40 - 59.9	Beech - maple	Oak - pine
B_1	20 - 39.9	Oak - hickory	Oak - hickory

The moisture factor, and hence the moisture index, is of importance principally in temperate and tropical latitudes. In the economically valuable coniferous forests of the sub-artic, thermal control appears to outweigh the significance of moisture. In a recently published series of articles Hare has shown that the broad physiognomic subdivisions of the Boreal Forest are highly correlated with the distribution of annual PE. He suggests the divisions shown in table 6.2.

Table 6.2

Suggested Relation Between Potential Evapotranspiration and Vegetation in Boreal Forest.

Forest Sub-Zone	Annual PE (cm)	Vegetation
Tundra		Tundra
	31	
Forest-tundra		Tundra on interfluves; woodland in valleys.
	35	
Woodland		Predominantly *Cladonia*-rich open woodland; closed-crown forest in isolated groves.
	42	
Forest		Closed-crown forest occupying most of the mesic sites.
	52	
Mixed temperate forests		Forest dominated by non-Boreal species, typically deciduous broadleaves.

The attempts at correlation of climate and vegetation that are presented in the foregoing paragraphs are highly tentative. Actually, since the development of climax vegetation formations and of mature soil types is most closely related to the march of soil moisture it is felt that the present approach represents a step in the right direction. It is certain that no system of classification exists, that is more soundly based. In particular, in establishing climatic analogues for the guidance of programs of exotic plant introduction, the indices described above seem likely to be of considerable value.

Since it is clear that whether grasses and other herbaceous plants occupy an area alone or share it with trees and shrubs depends on the soil moisture regime it is necessary to go further than merely to obtain a useful and rational climatic classification. It is necessary to develop more fully the water balance approach, to determine the actual influence of soil moisture on climax vegetation formations and to make maps of these active controls of vegetation distribution. Thus, foresters should not hope for a complete solution to problems of plant distribution in a "standard" climatic classification but should seek information on the soil moisture regime, and understand the importance of periods of moisture deficiency and surplus in the formation of vegetation complexes. Maps of the distribution of these latter elements would be more important than conventional maps of climatic regions.

PART III

SUMMARY AND BIBLIOGRAPHY

CHAPTER 7. SUMMARY AND BIBLIOGRAPHY OF ARTICLES RELATING TO POTENTIAL EVAPOTRANSPIRATION AND THE WATER BALANCE.

It is becoming increasingly apparent that in any adequate program of research, development, or operation it is necessary to understand the climatic influences involved. In most instances, the primary climatic influence is the moisture relationship. The foregoing report has attempted to explain the development and evaluation of the water balance and to show its usefulness in understanding the moisture relationships which exist in the world around us.

Beginning with the realization that the return flow of moisture from the surface and from plants to the atmosphere is as important a climatic factor as the downward flow of moisture in the form of precipitation and that one cannot determine the moisture relationships of an area without comparing the precipitation with the potential evapotranspiration, it has been possible to develop a simple bookkeeping procedure permitting the daily or monthly determination of the various moisture factors which are influential in describing the moisture regime of an area. Through the continued use of the bookkeeping procedure over many years and as a result of the increase in our knowledge which has accrued from recent developments in meteorologic theory and instrumentation there has been a refinement of some of the original assumptions in order to include additional terms which make the procedure more reliable.

Checks on the validity of the results which are possible have shown that quite reasonable values of soil moisture storage, moisture deficit, and moisture surplus have resulted from the use of the procedure. Although the use of the water balance in other fields of research has not been extensive it has already been found to be a tool of considerable power and utility.

The present report on the water balance should be considered as provisional and introductory. It is to be expected that as more scientists become aware of this research tool and begin to investigate it in detail as they apply it to different research problems there will be considerable improvement in the bookkeeping procedure to make it more rational and yet simpler to apply. Continued work will reveal how it can be utilized in other fields of research to provide information which is unavailable in any other way. It has taken a number of years to bring the water balance to its present state of development and utility. Its promise of value in future research makes worthwhile the effort so far expended and insures that its future development will be rapid.

During the decade since the concept of potential evapotranspiration was first presented there have been an increasing number of papers by scientists in all fields and in all parts of the world that deal with this subject. Some have been analytical in nature, attempting to compare the results from this approach with the results which can be obtained from others. Some suggest modifications or improvements in the approach. The great majority, however, deal with the use of the concept as a research tool in an attempt to understand more clearly some of the many complex moisture problems in our environment.

With the thought that a listing of these various papers would be of some value to other researchers, the following preliminary bibliography of articles dealing with the concept of potential evapotranspiration and the water balance has been assembled. It is hoped that readers of this report will call attention to other references so that this list can be made more useful.

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APPENDIX I

INSTRUCTIONS FOR EVALUATING THE WATER BALANCE

A) MONTHLY WATER BALANCE

To carry out the computations at a given station, all that is needed are the data of mean monthly temperature, the latitude of the station, the monthly precipitation, and the tables and charts included with these notes. Sample water balance computations at three stations are included below. These examples will be discussed in detail in the following notes.

Seabrook, New Jersey
(All values except T and I in mm)

	J	F	M	A	M	J	J	A	S	O	N	D	Y
T°C	0.9	1.2	5.9	11.3	17.5	22.3	24.7	23.7	20.2	14.0	7.6	2.3	
I	.07	.12	1.29	3.44	6.66	9.62	11.23	10.55	8.28	4.75	1.89	.31	58.21
Unadj PE	1	2	16	41	75	106	122	115	93	55	23	4	
PE	1	2	16	46	92	131	154	136	97	53	19	3	750
P	87	93	102	88	92	91	112	113	82	85	70	93	1108
P-PE	86	91	86	42	0	-40	-42	-23	-15	32	51	90	358
Acc Pot WL					(0)	-40	-82	-105	-120				
ST*	300	300	300	300	300	262	227	210	200	232	283	300	
Δ ST						-38	-35	-17	-10	+32	+51	+17	
AE	1	2	16	46	92	129	147	130	92	53	19	3	730
D	0	0	0	0	0	2	7	6	5	0	0	0	20
S	86	91	86	42	0	0	0	0	0	0	0	73	378
RO	61	76	81	62	31	15	8	4	2	1	0	37	378
SMRO													
Tot RO	61	76	81	62	31	15	8	4	2	1	0	37	
DT	361	376	381	361	330	277	234	213	201	232	283	336	

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Bismarck, North Dakota
(All values except T and I in mm)

	J	F	M	A	M	J	J	A	S	O	N	D	Y
T°C	-13.4	-12.1	-4.3	5.6	12.5	17.6	21.0	19.6	14.5	7.2	-1.9	-9.6	
I				1.19	4.00	6.72	8.78	7.91	5.01	1.74			35.35
Unadj PE				26	61	88	106	98	71	34			
PE	0	0	0	30	79	116	141	120	74	32	0	0	592
P	11	11	23	39	59	85	57	46	31	24	14	14	414
P-PE	11	11	23	9	-20	-31	-84	-74	-43	-8	14	14	-178
Acc Pot WL				(-227)	-247	-278	-362	-436	-479	-487			
ST	97	108	131	140	131	118	89	69	60	58	72	86	
Δ ST	11	11	23	9	-9	-13	-29	-20	-9	-2	14	14	
AE	0	0	0	30	68	98	86	66	40	26	0	0	414
D	0	0	0	0	11	18	55	54	34	6	0	0	178
S	0	0	0	0	0	0	0	0	0	0	0	0	0
RO	0	0	0	0	0	0	0	0	0	0	0	0	0
SMRO	0	0	0	0	0	0	0	0	0	0	0	0	0
Tot RO	0	0	0	0	0	0	0	0	0	0	0	0	0
DT	97	108	131	140	131	118	89	69	60	58	72	86	

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* See discussion under General Comments, page 96.

Concord, New Hampshire
(All values except T and I in mm)

	J	F	M	A	M	J	J	A	S	O	N	D	Y
T°C	-6.0	-5.2	0.0	6.7	13.3	18.0	20.9	19.5	15.5	9.6	2.9	-3.4	
I				1.56	4.40	6.95	8.72	7.85	5.55	2.69	.44		38.16
Unadj PE				30	63	88	104	96	75	44	12		
PE	0	0	0	34	79	113	133	115	78	42	10	0	604
P	68	63	75	74	75	80	90	81	76	76	86	66	910
P-PE	68	63	75	40	-4	-33	-43	-34	-2	34	76	66	306
Acc Pot WL				(0)	-4	-37	-80	-114	-116				
ST	434	497	300	300	296	265	229	204	204	238	300	366	
Δ ST					-4	-31	-36	-25	0	+34	+62		
AE	0	0	0	34	79	111	126	106	76	42	10	0	584
D	0	0	0	0	0	2	7	9	2	0	0	0	20
S	0	0	75	40	0	0	0	0	0	0	14	0	129
RO	2	1	38	39	19	10	5	2	1	1	7	4	129
SMRO			20	89	44	22	11	6	3	1	1		197
Tot RO	2	1	58	128	63	32	16	8	4	2	8	4	326
DT	435	497	514	426	359	296	244	211	207	239	307	369	

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In order to determine potential evapotranspiration, mean monthly values of temperature must be available, and the latitude of the station must be known. Three steps are involved in the computation and all three are accomplished by means of a nomogram (figure I.1) and tables (tables I.1, I.2).

I: Heat Index

The first step is to obtain the heat index I. Table I.1 gives the monthly values of i corresponding to mean monthly temperatures. Summation of the 12 monthly values gives the index I. i is zero when the mean temperature is 0°C or less.

Example: Seabrook, N.J.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 1 T°C	0.9	1.2	5.9	11.3	17.5	22.3	24.7	23.7	20.2	14.0	7.6	2.3	
Line 2 I	.07	.12	1.29	3.44	6.66	9.62	11.23	10.55	8.28	4.75	1.89	.31	58.21

Unadj PE: Unadjusted Potential Evapotranspiration

The second step is to determine the unadjusted potential evapotranspiration (in mm) using the nomogram of figure I.1* by drawing the regression line from the point of convergence through the I value for the station. For temperatures over 26.5°C use table accompanying graph for all indices I. Potential evapotranspiration is zero at temperatures below 0°C.

Example: Seabrook, N.J.

	J	F	M	A	M	J	J	A	S	O	N	D
Line 1 T°C	0.9	1.2	5.9	11.3	17.5	22.3	24.7	23.7	20.2	14.0	7.6	2.3
Line 3 Unadj PE	1	2	16	41	75	106	122	115	93	55	23	4

Adj PE: Adjusted Potential Evapotranspiration

The third step is to adjust these values of potential evapotranspiration for month and day length by multiplying by the proper correction factors, sample values

* See discussion under General Comments, page 96.

of which are given in table I.2. For stations poleward of 50° use correction factor for 50°.

Example: Seabrook, N.J.

Seabrook, N.J. is located at latitude 40°N. Monthly correction factors for this latitude are:

J	F	M	A	M	J	J	A	S	O	N	D
.84	.83	1.03	1.11	1.24	1.25	1.27	1.18	1.04	.96	.83	.81

Multiply the unadjusted potential evapotranspiration for each month by the appropriate correction factor to obtain adjusted value of potential evapotranspiration (in mm).

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 3 Unadj PE	1	2	16	41	75	106	122	115	93	55	23	4	
Line 4 Adj PE	1	2	16	46	92	131	154	136	97	53	19	3	750

P: Precipitation

Enter monthly precipitation in mm on appropriate line.

P-PE: Precipitation Minus The Potential Evapotranspiration

To determine periods of moisture excess and deficiency it is necessary to obtain the difference between precipitation and potential evapotranspiration. A negative value of P-PE indicates the amount by which the precipitation fails to supply the potential water need of a vegetation covered area. A positive value of P-PE indicates the amount of excess water which is available during certain periods in the year for soil moisture recharge and runoff.

In the great majority of stations there is only one so called "wet" season and one "dry" season per year. Thus, there will be only one set of consecutive negative and one set of positive differences. At these stations, two possibilities exist. At some places the excess precipitation (positive P-PE) during the year may be greater than the potential water loss (negative P-PE) (see example Seabrook, N.J.), while in other places the reverse may be true (see example Bismarck, N.D.). This latter situation will occur in dry areas where the precipitation is not sufficient to bring the soil moisture up to its maximum value of water holding capacity at any time during the year. Here the water deficiency even at the end of a period of rain and moisture recharge is some value other than zero. At stations with positive totals the water deficiency at the end of the wet period is always zero.

Example: Seabrook, N.J.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 4 Adj PE	1	2	16	46	92	131	154	136	97	53	19	3	750
Line 5 P	87	93	102	88	92	91	112	113	82	85	70	93	1108
Line 6 P-PE	86	91	86	42	0	-40	-42	-23	-15	32	51	90	358

Example: Bismarck, N.D.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 4 Adj PE	0	0	0	30	79	116	141	120	74	32	0	0	592
Line 5 P	11	11	23	39	59	85	57	46	31	24	14	14	414
Line 6 P-PE	11	11	23	9	-20	-31	-84	-74	-43	-8	14	14	-178

Acc Pot WL: Accumulated Potential Water Loss

The negative values of P-PE, representing a potential deficiency of water are summed month by month as an aid in the computational steps which follow. At dry stations (annual total P-PE is negative) it is necessary to find a value of potential water deficiency with which to start accumulating the negative values of P-PE. This can be done easily with the use of a nomogram (figure I.2*) and a table of soil moisture retention (table I.3). Obtain the potential deficiency (the sum of the negative values of P-PE), find this value on the border of table I.3 and then locate in the body of the same table the value of actual retention which corresponds to this value of potential deficiency. Find this number on the vertical scale on the left side of the graph (figure I.2). Follow horizontally across on this line until it intersects the sloping line whose value equals the sum of the positive P-PE's. Read the value of potential deficiency with which to start accumulation of P-PE's on horizontal scale at bottom of graph directly under this intersection.

Example: Seabrook, N.J.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 6 P-PE	86	91	86	42	0	-40	-42	-23	-15	32	51	90	358
Line 7 Acc Pot WL					(0)	-40	-82	-105	-120				

Since the sum of all the P-PE values is positive the value of accumulated potential water loss with which to start accumulating the negative values of P-PE is 0.

Example: Bismarck, N.D.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 6 P-PE	11	11	23	9	-20	-31	-84	-74	-43	-8	14	14	-178
Line 7 Acc Pot WL				(-227)	-247	-278	-362	-436	-479	-487			

Sum of P-PE is negative so that it is necessary to use table I.3 and figure I.2 to obtain proper value with which to begin accumulating negative P-PE values. Sum the negative values of P-PE. The total for Bismarck is -260. Look up on table I.3 value of actual retention which corresponds to value of potential deficiency of 260. Locate this value, 125, on the left hand side of the nomogram (figure I.2) and go across until it intersects sloping line having the value of the sum of the positive P-PE's (82). On the horizontal scale below this intersection is the value of the potential deficiency with which to start the accumulation of the negative values of P-PE. In the case of Bismarck as shown on figure I.2 this value is -227 which is entered for April, the month just before the series of negative P-PE values.

ST: Storage

Table I.3 gives values of the soil moisture retained in the soil, the soil moisture storage, after a given amount of accumulated potential water loss has occurred. Look up each value of the accumulated potential water loss line in table I.3 and enter values of storage (retention) in appropriate places on storage line. After the soil moisture storage for each of the months with negative values of P-PE has been found from the table, the positive figures from line P-PE representing additions of moisture to the soil must be included. If accumulative adding brings the value of soil moisture storage over 300* this value is entered on the storage line until the next negative value of P-PE is reached since the soil cannot hold more water in storage (see example Seabrook, N.J.). If the temperature is below -1°C it is assumed that the precipitation falls as snow. Under those conditions total storage can go higher than 300 since the snow will remain on the surface. If the total storage is above the water holding capacity of the soil when the temperature again rises above -1°C, the storage reverts to 300 since the snow is considered to be

* See discussion under General Comments, page 96.

moisture surplus in the process of running off (see example Concord, N.H.). If the total storage is less than 300 at the time temperatures above -1°C again occur the snow melt is held in storage in the ground (see example Bismarck, N.D.).

Example: Seabrook, N.J.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 6 P-PE	86	91	86	42	0	-40	-42	-23	-15	32	51	90	358
Line 7 Acc Pot WL					(0)	-40	-82	-105	-120				
Line 8 ST	300	300	300	300	300	262	227	210	200	232	283	300	

Example: Concord, N.H.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 1 T°C	-6.0	-5.2	0.0	6.7	13.3	18.0	20.9	19.5	15.5	9.6	2.9	-3.4	
Line 6 P-PE	68	63	75	40	-4	-33	-43	-34	-2	34	76	66	306
Line 7 Acc Pot WL				(0)	-4	-37	-80	-114	-116				
Line 8 ST	434	497	300	300	296	265	229	204	204	238	300	366	

Example: Bismarck, N.D.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 1 T°C	-13.4	-12.1	-4.3	5.6	12.5	17.6	21.0	19.6	14.5	7.2	-1.9	-9.6	
Line 6 P-PE	11	11	23	9	-20	-31	-84	-74	-43	-8	14	14	-178
Line 7 Acc Pot WL				(-227)	-247	-278	-362	-436	-479	-487			
Line 8 ST	97	108	131	140	131	118	89	69	60	58	72	86	

<u>ΔST: Change in Soil Moisture</u>

As an aid in later computations it is desirable to obtain the difference in the amount of soil moisture storage from one month to the next. When the value in the storage line is 300 or over, it is assumed that there is no change in soil storage although there may be a change in above surface storage. This is not reflected in the values in the Δ ST line.

Example: Concord, N.H.

	J	F	M	A	M	J	J	A	S	O	N	D
Line 1 T°C	-6.0	-5.2	0.0	6.7	13.3	18.0	20.9	19.5	15.5	9.6	2.9	-3.4
Line 8 ST	434	497	300	300	296	265	229	204	204	238	300	366
Line 9 ΔST	0	0	0	0	-4	-31	-36	-25	0	+34	+62	0

<u>AE: Actual Evapotranspiration</u>

When the precipitation is greater than the potential evapotranspiration, the soil remains full of water and the actual evapotranspiration will equal the potential. When the precipitation drops below the potential evapotranspiration the soil begins to dry out and actual evapotranspiration becomes less than that potentially possible. In those months, the actual evapotranspiration equals the precipitation plus the amount of water drawn from the soil moisture storage (the ΔST, disregarding its sign).

Example: Seabrook, N.J.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 4 PE	1	2	16	46	92	131	154	136	97	53	19	3	750
Line 5 P	87	93	102	88	92	91	112	113	82	85	70	93	1108
Line 9 ΔST	0	0	0	0	0	-38	-35	-17	-10	+32	+51	+17	
Line 10 AE	1	2	16	46	92	129	147	130	92	53	19	3	730

D: Moisture Deficit

The amount by which the actual and potential evapotranspiration differ in any month is the moisture deficit for that month.

Example: Seabrook, N.J.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 4 PE	1	2	16	46	92	131	154	136	97	53	19	3	750
Line 10 AE	1	2	16	46	92	129	147	130	92	53	19	3	730
Line 11 D	0	0	0	0	0	2	7	6	5	0	0	0	20

S: Moisture Surplus

After the soil moisture storage reaches 300 any excess precipitation is counted as moisture surplus and is subject to runoff. If the temperature of the month is below -1°C so that the precipitation falls as snow there is no surplus since all of the precipitation is treated as storage. Only when the temperature rises above -1°C can a surplus again occur. The surplus water which results from the melting snow is considered separately (in line 14) from the surplus which results from rainfall (line 12).

Example: Concord, N.H.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 1 T	-6.0	-5.2	0.0	6.7	13.3	18.0	20.9	19.5	15.5	9.6	2.9	-3.4	
Line 6 P-PE	68	63	75	40	-4	-33	-43	-34	-2	34	76	66	
Line 8 ST	434	497	300	300	296	265	229	204	204	238	300	266	
Line 9 ΔST	0	0	0	0	-4	-31	-36	-25	0	+34	+62	0	
Line 12 S	0	0	75	40	0	0	0	0	0	0	14	0	129

RO: Water Runoff

Studies have shown that for large watersheds only about 50 percent of the surplus water which is available for runoff in any month actually does run off. The rest of the surplus water is detained on the watershed and made available for runoff during the next month. If periods shorter than a month are considered or the watershed is only a few square miles in area the detention of surplus water may differ from 50 percent. In the example for Concord, N.H. a factor of 50 percent is used.

Example: Concord, N.H.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 12 S	0	0	75	40	0	0	0	0	0	0	14	0	129
Line 13 RO	2	1	38	39	19	10	5	2	1	1	7	4	129

SMRO: Snow Melt Runoff

Empirical evidence now available indicates that there is a relation between the elevation of the watershed, and the runoff of the water made available from melting snow provided the soil moisture content is at field capacity. It has been found that the lower the elevation of the watershed, the more rapid is the runoff of the water from the melting snow. Of course, in the first month that the temperature rises above -1°C the snow melt runoff cannot equal 50 percent of the available snow for much of the heat must go into the process of melting the snow first. Until the temperature rises considerable above freezing, snow melt remains a relatively slow process. It must be emphasized that the empirical relations found relate to the runoff of the water from the melting snow and not to the rate of snow melt itself.

It has been found that in areas above 1600 m (5000 ft) if the soil moisture storage is 300mm approximately 10 percent of the water made available from the melting snow will run off during the first month with temperatures above -1°C, 25 percent of the remainder during the second month, and 50 percent of the remainder in each of the following months until it is all gone. In areas below 1600 m if the soil moisture storage is 300 mm, 10 percent of the water available from the melting snow will run off during the first month with temperatures above -1°C while 50 percent of the remainder will run off in succeeding months. If the soil moisture storage is less than 300 mm there is no runoff of snow melt since it is assumed to go into storage in the soil.

Example: Concord, N.H. Elevation 339 feet

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 1 T°C	-6.0	-5.0	0.0	6.7	13.3	18.0	20.9	19.5	15.5	9.6	2.9	-3.4	
Line 5 P	68	63	75	74	75	80	90	81	76	76	86	66	
Line 8 ST	434	497	300	300	296	265	229	204	204	238	300	366	
Line 14 SMRO			20	89	44	22	11	6	3	1	1		197

Snow 197

Example: Bismarck, N.D. Elevation 1670 feet

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 1 T°C	-13.4	-13.1	-4.3	5.6	12.5	17.6	21.0	19.6	14.5	7.2	-1.9	-9.6	
Line 5 P	11	11	23	39	59	85	57	46	31	24	14	14	
Line 8 ST	97	108	131	140	131	118	89	69	60	58	72	86	
Line 14 SMRO	0	0	0	0	0	0	0	0	0	0	0	0	0

Snow 73

Tot RO: Total Runoff

The total runoff from an area is the sum of the water surplus runoff and the snow melt runoff. Because of the lag introduced by large watersheds there can be appreciable runoff during periods when the evapotranspiration is more than the precipitation and a moisture deficit is occurring.

Example: Concord, N.H.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 13 RO	2	1	38	39	19	10	5	2	1	1	7	4	129
Line 14 SMRO			20	89	44	22	11	6	3	1	1		197
Line 15 Tot RO	2	1	58	128	63	32	16	8	4	2	8	4	326

DT: Total Moisture Detention

The moisture detention is the total of the water stored within the soil, the snow remaining on the soil surface and the surplus water in the process of running off which has been detained for a month.

Example: Concord, N.H.

	J	F	M	A	M	J	J	A	S	O	N	D	Y
Line 8 ST	434	497	300	300	296	265	229	204	204	238	300	366	
Line 12 S	0	0	75	40	0	0	0	0	0	0	14	0	129
Line 13 RO	2	1	38	39	19	10	5	2	1	1	7	4	129
Line 14 SMRO			20	89	44	22	11	6	3	1	1		197
Line 16 DT	435	497	514	426	359	296	244	211	207	239	307	369	

Snow 197

Note: For March, detention of 514 consists of 75-38=37 mm of water runoff detained, 197-20=177 mm of snow detained and 300 mm of storage. 300+37+177=514.

B) DAILY WATER BALANCE

The computation procedure for determining the daily variation in soil moisture storage is quite similar to that already discussed for the monthly water balance. A special form has been prepared which permits the computations to be carried out simply and directly. Data for one period at Seabrook, N.J. have been evaluated as an example.

The computation form, or soil moisture record, permits the evaluation of one month of data per sheet. The first three columns deal with the determination of adjusted potential evapotranspiration from data of mean air temperature. The manner of determining potential evapotranspiration is entirely similar to that described previously for the monthly computations. If the nomogram (figure I.1) is used to determine unadjusted potential evapotranspiration, one goes across horizontally using the mean temperature of the day instead of the monthly mean temperature. The value of unadjusted potential evapotranspiration read on the horizontal scale above the intersection of the temperature line with the I regression line gives the daily unadjusted evapotranspiration. The correction factor then used to convert the unadjusted value to an adjusted potential evapotranspiration depends only on the deviation of the actual day length from a standard 12 hour day. Before making the daily computations a table of such day length correction factors should be prepared. It is also quite desireable to prepare tables of the relation between temperature and unadjusted potential evapotranspiration for a given I value if the daily record for a station is to be evaluated. This permits more rapid computations than would be possible through use of the nomogram. Examples of the tables of daily unadjusted evapotranspiration and of the day length correction factor for one station are found in tablesI.4 and I.5.

As in the case of the monthly computations, in making the daily determinations it is necessary to obtain the difference between the precipitation and the potential evapotranspiration. When the potential evapotranspiration is greater than the precipitation, this value is negative and indicates a drying of the soil.

In order to carry out the calculations of soil moisture storage it is necessary to know the value of soil moisture content with which to begin. If the record for the current year is being evaluated it is possible to obtain this starting value by direct measurement of the actual and total moisture content of the soil layer under consideration. If, however, a past record of soil moisture is being investigated the starting value of soil moisture content can only be obtained by assuming a value of soil moisture storage of 300 after a period of rain during the moist season of the year and evaluating the daily record for a period of about a year prior to the actual date for which the daily calculations are wanted. During that time the assumed value of soil moisture storage will approach the true value and by the end of the year it should closely approximate the true value. Actual calculations can then begin with a realistic value of soil moisture storage.

When additions of moisture to the soil bring the storage up to 300 that value is entered on the soil moisture record sheet under storage. As the soil dries it is

DAILY SOIL MOISTURE BALANCE, SEABROOK, NEW JERSEY, MAY-JUNE 1953

Water holding capacity of soil 300 mm. Soil moisture content at start 295 mm. Ninety percent of available gravitational water on any day held for later percolation. (All values except T in mm)

Date	Mean T°F	Unadj PE	Adj PE	P	P-PE	Soil Moist ST	Act ST Change	Moist Def	Moist Sur	Avail Grav Water	Grav Water ST	Soil Moist Bal
May						295						
30	60	2.1	3	1	-2	293	-2	0				293
31	54	1.5	2	18	+16	300	+7	0	9	9	8	308
June												
1	53	1.4	2	11	+9	300	0	0	9	17	15	315
2	55	1.6	2	1	-1	299	-1	0		15	14	313
3	64	2.6	3		-3	296	-3	0		14	13	309
4	66	2.8	3		-3	293	-3	0		13	12	305
5	74	3.8	5		-5	288	-5	0		12	11	299
6	77	4.1	5	15	+10	298	+10	0		11	10	308
7	78	4.3	5	1	-4	294	-4	0		10	9	303
8	73	3.7	5		-5	289	-5	0		9	8	297
9	72	3.5	4		-4	285	-4	0		8	7	292
10	75	3.9	5		-5	280	-5	0		7	7	287
11	68	3.0	4		-4	277	-3	1		7	6	283
12	63	2.5	3		-3	274	-3	0		6	5	279
13	66	2.8	4	1	-3	271	-3	0		5	5	276
14	62	2.3	3	14	+11	282	+11	0		5	4	286
15	57	1.8	2		-2	280	-2	0		4	4	284
16	64	2.6	3		-3	278	-2	1		4	4	282
17	68	3.0	4		-4	274	-4	0		4	3	277
18	66	2.8	4		-4	270	-4	0		3	3	274
19	70	3.3	4		-4	267	-3	1		3	3	270
20	76	4.0	5		-5	262	-5	0		3	2	264
21	79	4.4	6		-6	257	-5	1		2	2	259
22	80	4.5	6		-6	252	-5	1		2	2	254
23	77	4.1	5		-5	248	-4	1		2	2	250
24	72	3.5	4		-4	244	-4	0		2	2	246
25	68	3.0	4		-4	241	-3	1		2	1	242
26	73	3.7	4		-4	238	-3	1		1	1	239
27	81	4.7	6		-6	233	-5	1		1	1	234
28	80	4.5	6		-6	228	-5	1		1	1	229
29	79	4.4	6	2	-4	225	-3	1		1	1	226
30	78	4.3	5	1	-4	222	-3	1		1	1	223

necessary to convert the values of potential change in storage given in the P-PE column into values of actual changes in storage for with drying the actual evapotranspiration will be less than the potential. A table has been prepared which gives the actual soil moisture retention or storage for given values of P-PE. This is the same table which was used for a similar purpose in making the monthly computations (table I.3). Since the values of P-PE are not accumulated as in the case of the monthly computations it is necessary to accumulate them as the work is carried out by finding the value of soil moisture storage in the body of table I.3 and then counting ahead a number equal to the value of P-PE to obtain the new value of soil moisture storage.

Moisture is not stored in the soil above the value of 300 but is considered to be surplus water in transit from the area. Of course, at temperatures below -1°C the soil is considered to be frozen and no percolation of water through the soil and out of the area occurs.

The moisture deficit is the difference between the potential water loss, P-PE, and the actual change in storage or actual water loss.

The moisture surplus on any given day is the excess of precipitation over potential evapotranspiration after the soil moisture storage reaches 300.

The sum of the gravitational water storage from the previous day and the moisture surplus gives the total available gravitational water on any given day. It has been found that only a certain percentage of the total quantity of surplus water will run off on any day. In the bookkeeping procedure the remaining percentage will be held as gravitational water storage and made available for runoff on the following day. Again only a percentage of this total will actually be lost by runoff on the second day. The percentage of surplus water lost each day depends on the soil type and structure and the depth of the soil layer under consideration. It has been found to be near 90 percent for a deep loam soil.

The soil moisture balance is the sum of the soil moisture storage and the gravitational water storage on any day. It can have a value well above 300 due to the inclusion of gravitational water in the accounting process.

C) GENERAL COMMENTS

In the foregoing discussions of the monthly and the daily moisture balances it must be pointed out that only the case in which the water holding capacity of the soil is 300 mm is considered. Soils vary in their ability to hold water and plants will vary in the depth of their root systems so that in actual practice, the amount of water available to a plant when the soil is full of water can vary from less than 50 mm to over 400 mm. In order to work out the water balance for areas with moisture holding capacities which are different from 300 it is necessary to have the appropriate tables and nomograms, prepared for the moisture holding capacity in question, and to substitute the new value of moisture holding capacity in place of the value 300 in these instructions. More detailed instructions which are in the process of being prepared will include tables for use with other values of moisture holding capacity.

In the foregoing instructions only stations with one consecutive group of positive and negative differences of P-PE have been described. There are some stations which have two or more groups of positive and negative P-PE's. These stations are more difficult to evaluate because of the problems involved in determining the proper value with which to start accumulating the potential water loss. Special forms are available which permit the data at these stations to be evaluated with comparative ease. Detailed instructions on the handling of these more difficult stations are being prepared.

Figures I.1 and I.2 included with these notes are not particularly suitable for actual use since it has been necessary to omit many of the background guide lines on the paper for clarity. It is, thus, desirable to prepare figures I.1 and I.2 on larger scale graph paper to permit them to be read more exactly if actual data from a station are to be evaluated.

Table I.1

MONTHLY VALUES OF i CORRESPONDING TO MONTHLY MEAN TEMPERATURES

T°C	.0	.1	.2	.3	.4	.5	.6	.7	.8	.9
0			.01	.01	.02	.03	.04	.05	.06	.07
1	.09	.10	.12	.13	.15	.16	.18	.20	.21	.23
2	.25	.27	.29	.31	.33	.35	.37	.39	.42	.44
3	.46	.48	.51	.53	.56	.58	.61	.63	.66	.69
4	.71	.74	.77	.80	.82	.85	.88	.91	.94	.97
5	1.00	1.03	1.06	1.09	1.12	1.16	1.19	1.22	1.25	1.29
6	1.32	1.35	1.39	1.42	1.45	1.49	1.52	1.56	1.59	1.63
7	1.66	1.70	1.74	1.77	1.81	1.85	1.89	1.92	1.96	2.00
8	2.04	2.08	2.12	2.15	2.19	2.23	2.27	2.31	2.35	2.39
9	2.44	2.48	2.52	2.56	2.60	2.64	2.69	2.73	2.77	2.81
10	2.86	2.90	2.94	2.99	3.03	3.08	3.12	3.16	3.21	3.25
11	3.30	3.34	3.39	3.44	3.48	3.53	3.58	3.62	3.67	3.72
12	3.76	3.81	3.86	3.91	3.96	4.00	4.05	4.10	4.15	4.20
13	4.25	4.30	4.35	4.40	4.45	4.50	4.55	4.60	4.65	4.70
14	4.75	4.81	4.86	4.91	4.96	5.01	5.07	5.12	5.17	5.22
15	5.28	5.33	5.38	5.44	5.49	5.55	5.60	5.65	5.71	5.76
16	5.82	5.87	5.93	5.98	6.04	6.10	6.15	6.21	6.26	6.32
17	6.38	6.44	6.49	6.55	6.61	6.66	6.72	6.78	6.84	6.90
18	6.95	7.01	7.07	7.13	7.19	7.25	7.31	7.37	7.43	7.49
19	7.55	7.61	7.67	7.73	7.79	7.85	7.91	7.97	8.03	8.10
20	8.16	8.22	8.28	8.34	8.41	8.47	8.53	8.59	8.66	8.72
21	8.78	8.85	8.91	8.97	9.04	9.10	9.17	9.23	9.29	9.36
22	9.42	9.49	9.55	9.62	9.68	9.75	9.82	9.88	9.95	10.01
23	10.08	10.15	10.21	10.28	10.35	10.41	10.48	10.55	10.62	10.68
24	10.75	10.82	10.89	10.95	11.02	11.09	11.16	11.23	11.30	11.37
25	11.44	11.50	11.57	11.64	11.71	11.78	11.85	11.92	11.99	12.06
26	12.13	12.21	12.28	12.35	12.42	12.49	12.56	12.63	12.70	12.78
27	12.85	12.92	12.99	13.07	13.14	13.21	13.28	13.36	13.43	13.50
28	13.58	13.65	13.72	13.80	13.87	13.94	14.02	14.09	14.17	14.24
29	14.32	14.39	14.47	14.54	14.62	14.69	14.77	14.84	14.92	14.99
30	15.07	15.15	15.22	15.30	15.38	15.45	15.53	15.61	15.68	15.76
31	15.84	15.92	15.99	16.07	16.15	16.23	16.30	16.38	16.46	16.54
32	16.62	16.70	16.78	16.85	16.93	17.01	17.09	17.17	17.25	17.33
33	17.41	17.49	17.57	17.65	17.73	17.81	17.89	17.97	18.05	18.13
34	18.22	18.30	18.38	18.46	18.54	18.62	18.70	18.79	18.87	18.95
35	19.03	19.11	19.20	19.28	19.36	19.45	19.53	19.61	19.69	19.78
36	19.86	19.95	20.03	20.11	20.20	20.28	20.36	20.45	20.53	20.62
37	20.70	20.79	20.87	20.96	21.04	21.13	21.21	21.30	21.38	21.47
38	21.56	21.64	21.73	21.81	21.90	21.99	22.07	22.16	22.25	22.33
39	22.42	22.51	22.59	22.68	22.77	22.86	22.95	23.03	23.12	23.21
40	23.30									

Table I.2

MEAN POSSIBLE DURATION OF SUNLIGHT IN THE NORTHERN AND SOUTHERN HEMISPHERES

EXPRESSED IN UNITS OF 30 DAYS OF 12 HOURS EACH

	J	F	M	A	M	J	J	A	S	O	N	D
N. Lat.												
0	1.04	.94	1.04	1.01	1.04	1.01	1.04	1.04	1.01	1.04	1.01	1.04
5	1.02	.93	1.03	1.02	1.06	1.03	1.06	1.05	1.01	1.03	.99	1.02
10	1.00	.91	1.03	1.03	1.08	1.06	1.08	1.07	1.02	1.02	.98	.99
15	.97	.91	1.03	1.04	1.11	1.08	1.12	1.08	1.02	1.01	.95	.97
20	.95	.90	1.03	1.05	1.13	1.11	1.14	1.11	1.02	1.00	.93	.94
25	.93	.89	1.03	1.06	1.15	1.14	1.17	1.12	1.02	.99	.91	.91
26	.92	.88	1.03	1.06	1.15	1.15	1.17	1.12	1.02	.99	.91	.91
27	.92	.88	1.03	1.07	1.16	1.15	1.18	1.13	1.02	.99	.90	.90
28	.91	.88	1.03	1.07	1.16	1.16	1.18	1.13	1.02	.98	.90	.90
29	.91	.87	1.03	1.07	1.17	1.16	1.19	1.13	1.03	.98	.90	.89
30	.90	.87	1.03	1.08	1.18	1.17	1.20	1.14	1.03	.98	.89	.88
31	.90	.87	1.03	1.08	1.18	1.18	1.20	1.14	1.03	.98	.89	.88
32	.89	.86	1.03	1.08	1.19	1.19	1.21	1.15	1.03	.98	.88	.87
33	.88	.86	1.03	1.09	1.19	1.20	1.22	1.15	1.03	.97	.88	.86
34	.88	.85	1.03	1.09	1.20	1.20	1.22	1.16	1.03	.97	.87	.86
35	.87	.85	1.03	1.09	1.21	1.21	1.23	1.16	1.03	.97	.86	.85
36	.87	.85	1.03	1.10	1.21	1.22	1.24	1.16	1.03	.97	.86	.84
37	.86	.84	1.03	1.10	1.22	1.23	1.25	1.17	1.03	.97	.85	.83
38	.85	.84	1.03	1.10	1.23	1.24	1.25	1.17	1.04	.96	.84	.83
39	.85	.84	1.03	1.11	1.23	1.24	1.26	1.18	1.04	.96	.84	.82
40	.84	.83	1.03	1.11	1.24	1.25	1.27	1.18	1.04	.96	.83	.81
41	.83	.83	1.03	1.11	1.25	1.26	1.27	1.19	1.04	.96	.82	.80
42	.82	.83	1.03	1.12	1.26	1.27	1.28	1.19	1.04	.95	.82	.79
43	.81	.82	1.02	1.12	1.26	1.28	1.29	1.20	1.04	.95	.81	.77
44	.81	.82	1.02	1.13	1.27	1.29	1.30	1.20	1.04	.95	.80	.76
45	.80	.81	1.02	1.13	1.28	1.29	1.31	1.21	1.04	.94	.79	.75
46	.79	.81	1.02	1.13	1.29	1.31	1.32	1.22	1.04	.94	.79	.74
47	.77	.80	1.02	1.14	1.30	1.32	1.33	1.22	1.04	.93	.78	.73
48	.76	.80	1.02	1.14	1.31	1.33	1.34	1.23	1.05	.93	.77	.72
49	.75	.79	1.02	1.14	1.32	1.34	1.35	1.24	1.05	.93	.76	.71
50	.74	.78	1.02	1.15	1.33	1.36	1.37	1.25	1.06	.92	.76	.70
S. Lat.												
5	1.06	.95	1.04	1.00	1.02	.99	1.02	1.03	1.00	1.05	1.03	1.06
10	1.08	.97	1.05	.99	1.01	.96	1.00	1.01	1.00	1.06	1.05	1.10
15	1.12	.98	1.05	.98	.98	.94	.97	1.00	1.00	1.07	1.07	1.12
20	1.14	1.00	1.05	.97	.96	.91	.95	.99	1.00	1.08	1.09	1.15
25	1.17	1.01	1.05	.96	.94	.88	.93	.98	1.00	1.10	1.11	1.18
30	1.20	1.03	1.06	.95	.92	.85	.90	.96	1.00	1.12	1.14	1.21
35	1.23	1.04	1.06	.94	.89	.82	.87	.94	1.00	1.13	1.17	1.25
40	1.27	1.06	1.07	.93	.86	.78	.84	.92	1.00	1.15	1.20	1.29
42	1.28	1.07	1.07	.92	.85	.76	.82	.92	1.00	1.16	1.22	1.31
44	1.30	1.08	1.07	.92	.83	.74	.81	.91	.99	1.17	1.23	1.33
46	1.32	1.10	1.07	.91	.82	.72	.79	.90	.99	1.17	1.25	1.35
48	1.34	1.11	1.08	.90	.80	.70	.76	.89	.99	1.18	1.27	1.37
50	1.37	1.12	1.08	.89	.77	.67	.74	.88	.99	1.19	1.29	1.41

Table I.3

SOIL MOISTURE RETENTION TABLE - 300 mm

Soil moisture retained after different amounts of potential evapotranspiration have occurred. Water holding capacity of soil is 300 mm.

PE	0	1	2	3	4	5	6	7	8	9
					Water Retained in Soil					
0	300	299	298	297	296	295	294	293	292	291
10	290	289	288	287	286	285	284	283	282	281
20	280	279	278	278	277	276	275	274	273	272
30	271	270	269	268	268	267	266	265	264	263
40	262	261	260	260	259	258	257	256	255	254
50	254	253	252	251	250	249	248	248	247	246
60	245	244	244	243	242	241	240	240	239	238
70	237	236	236	235	234	233	232	232	231	230
80	229	228	228	227	226	225	225	224	223	222
90	222	221	220	219	219	218	217	216	215	215
100	214	214	213	212	212	211	210	209	209	208
110	207	207	206	205	204	204	203	202	202	201
120	200	200	199	198	198	197	196	196	195	194
130	194	193	192	192	191	191	190	189	189	188
140	187	187	186	186	185	184	184	183	182	182
150	181	181	180	179	179	178	178	177	176	176
160	175	175	174	173	173	172	172	171	171	170
170	170	169	168	168	167	167	166	166	165	164
180	164	163	163	162	162	161	160	160	159	159
190	158	158	157	157	156	156	155	155	154	154
200	153	153	152	152	151	151	150	150	149	149
210	148	148	147	147	146	146	145	145	144	144
220	143	143	142	142	141	141	140	140	139	139
230	138	138	138	137	137	136	136	135	135	134
240	134	133	133	132	132	132	131	131	130	130
250	130	129	128	128	128	127	127	126	126	126
260	125	125	124	124	124	123	123	122	122	121
270	121	121	120	120	119	119	119	118	118	117
280	117	117	116	116	115	115	115	114	114	114
290	113	113	112	112	112	111	111	110	110	110
300	109	109	109	108	108	108	107	107	106	106
310	106	105	105	105	104	104	104	103	103	103
320	102	102	102	101	101	101	100	100	100	99
330	99	98	98	98	98	97	97	97	96	96
340	96	95	95	95	94	94	94	93	93	93
350	92	92	92	92	91	91	91	90	90	90
360	89	89	89	88	88	88	88	87	87	87
370	86	86	86	86	85	85	85	84	84	84
380	84	83	83	83	82	82	82	82	81	81
390	81	80	80	80	80	80	79	79	79	78
400	78	78	78	77	77	77	77	76	76	76
410	76	75	75	75	74	74	74	74	74	73
420	73	73	72	72	72	72	72	71	71	71
430	71	70	70	70	70	70	69	69	69	68
440	68	68	68	68	67	67	67	67	66	66

Table I.3 (continued)

PE	0	1	2	3	4	5	6	7	8	9
450	66	66	66	65	65	65	65	64	64	64
460	64	64	63	63	63	63	63	62	62	62
470	62	62	61	61	61	61	61	60	60	60
480	60	60	59	59	59	59	59	58	58	58
490	58	58	57	57	57	57	57	56	56	56
500	56	56	55	55	55	55	55	54	54	54
510	54	54	54	53	53	53	53	53	52	52
520	52	52	52	52	51	51	51	51	51	50
530	50	50	50	50	50	50	49	49	49	49
540	49	49	48	48	48	48	48	48	47	47
550	47	47	47	47	46	46	46	46	46	46
560	46	45	45	45	45	45	45	44	44	44
570	44	44	44	44	44	43	43	43	43	43
580	43	42	42	42	42	42	42	42	42	41
590	41	41	41	41	41	41	40	40	40	40
600	40	40	40	39	39	39	39	39	39	39
610	38	38	38	38	38	38	38	38	38	37
620	37	37	37	37	37	37	36	36	36	36
630	36	36	36	36	36	36	35	35	35	35
640	35	35	35	34	34	34	34	34	34	34
650	34	34	33	33	33	33	33	33	33	33
660	32	32	32	32	32	32	32	32	32	32
670	32	31	31	31	31	31	31	31	31	31
680	30	30	30	30	30	30	30	30	30	30
690	30	29	29	29	29	29	29	29	29	29
700	28	28	28	28	28	28	28	28	28	28
710	28	27	27	27	27	27	27	27	27	27
720	27	26	26	26	26	26	26	26	26	26
730	26	26	26	26	25	25	25	25	25	25
740	25	25	25	25	25	24	24	24	24	24
750	24	24	24	24	24	24	24	24	23	23
760	23	23	23	23	23	23	23	23	23	23
770	22	22	22	22	22	22	22	22	22	22
780	22	22	22	22	22	21	21	21	21	21
790	21	21	21	21	21	21	21	21	20	20
800	20	20	20	20	20	20	20	20	20	20
810	20	20	20	20	20	19	19	19	19	19
820	19	19	19	19	19	19	19	19	19	19
830	18	18	18	18	18	18	18	18	18	18
840	18	18	18	18	18	18	18	17	17	17
850	17	17	17	17	17	17	17	17	17	17
860	17	17	17	17	16	16	16	16	16	16
870	16	16	16	16	16	16	16	16	16	16
880	16	16	16	16	15	15	15	15	15	15
890	15	15	15	15	15	15	15	15	15	15
900	15	15	14	14	14	14	14	14	14	14
910	14	14	14	14	14	14	14	14	14	14
920	14	14	14	14	13	13	13	13	13	13
930	13	13	13	13	13	13	13	13	13	13
940	13	13	13	13	12	12	12	12	12	12

Table I.3 (continued)

PE	0	1	2	3	4	5	6	7	8	9
950	12	12	12	12	12	12	12	12	12	12
960	12	12	12	12	12	12	12	12	12	12
970	12	12	11	11	11	11	11	11	11	11
980	11	11	11	11	11	11	11	11	11	11
990	11	11	11	11	11	11	11	10	10	10
1000	10	10	10	10	10	10	10	10	10	10
1010	10	10	10	10	10	10	10	10	10	10
1020	10	10	10	10	10	10	10	10	10	10
1030	9	9	9	9	9	9	9	9	9	9
1040	9	9	9	9	9	9	9	9	9	9
1050	9	9	9	9	9	9	9	9	8	8
1060	8	8	8	8	8	8	8	8	8	8
1070	8	8	8	8	8	8	8	8	8	8
1080	8	8	8	8	8	8	8	8	8	8
1090	8	8	8	8	8	8	8	8	8	8

****

	0	5		0	5		0	5		0
1100	7	7	1200	5	5	1300	4	4	1400	3
1110	7	7	1210	5	5	1310	4	4	1410	3
1120	7	7	1220	5	5	1320	4	4	1420	2
1130	7	7	1230	5	5	1330	3	3	1430	2
1140	6	6	1240	5	5	1340	3	3	1440	2
1150	6	6	1250	4	4	1350	3	3	1450	2
1160	6	6	1260	4	4	1360	3	3	1460	2
1170	6	6	1270	4	4	1370	3	3	1470	2
1180	6	6	1280	4	4	1380	3	3	1480	2
1190	6	5	1290	4	4	1390	3	3	1490	2

****

	0		0		0		0
1500	2	1600	1	1700	1	1800	1
1550	2	1650	1	1750	1	1850	1
1900	0						

Table I.4

UNADJUSTED DAILY RATES OF POTENTIAL EVAPOTRANSPIRATION
FOR I OF 58.2 SEABROOK, NEW JERSEY

T°F	PE cm	T°F	PE cm	T°F	PE cm	T°F	PE cm	T°F	PE cm	T°F	PE cm	T°F	PE cm
30.0	.00	40.0	.04	50.0	.12	60.0	.21	70.0	.33	80.0	.45	90.0	.59
30.5	.00	40.5	.04	50.5	.12	60.5	.22	70.5	.33	80.5	.46	90.5	.60
31.0	.00	41.0	.04	51.0	.12	61.0	.22	71.0	.34	81.0	.47	91.0	.60
31.5	.00	41.5	.05	51.5	.13	61.5	.23	71.5	.34	81.5	.47	91.5	.61
32.0	.00	42.0	.05	52.0	.13	62.0	.23	72.0	.35	82.0	.48	92.0	.62
32.5	.00	42.5	.05	52.5	.14	62.5	.24	72.5	.36	82.5	.49	92.5	.63
33.0	.00	43.0	.06	53.0	.14	63.0	.25	73.0	.37	83.0	.50	93.0	.63
33.5	.00	43.5	.06	53.5	.15	63.5	.25	73.5	.37	83.5	.50	93.5	.64
34.0	.01	44.0	.07	54.0	.15	64.0	.26	74.0	.38	84.0	.51	94.0	.65
34.5	.01	44.5	.07	54.5	.16	64.5	.26	74.5	.38	84.5	.52	94.5	.66
35.0	.01	45.0	.07	55.0	.16	65.0	.27	75.0	.39	85.0	.52	95.0	.66
35.5	.01	45.5	.08	55.5	.17	65.5	.27	75.5	.39	85.5	.53	95.5	.67
36.0	.01	46.0	.08	56.0	.17	66.0	.28	76.0	.40	86.0	.54	96.0	.68
36.5	.02	46.5	.09	56.5	.18	66.5	.29	76.5	.41	86.5	.54	96.5	.69
37.0	.02	47.0	.09	57.0	.18	67.0	.29	77.0	.41	87.0	.55	97.0	.69
37.5	.02	47.5	.09	57.5	.19	67.5	.30	77.5	.42	87.5	.56	97.5	.70
38.0	.03	48.0	.10	58.0	.19	68.0	.30	78.0	.43	88.0	.56	98.0	.71
38.5	.03	48.5	.10	58.5	.20	68.5	.31	78.5	.43	88.5	.57	98.5	.71
39.0	.03	49.0	.11	59.0	.20	69.0	.32	79.0	.44	89.0	.58	99.0	.72
39.5	.03	49.5	.11	59.5	.21	69.5	.32	79.5	.45	89.5	.58	99.5	.73
40.0	.04	50.0	.12	60.0	.21	70.0	.33	80.0	.45	90.0	.59	100.0	.74

Table I.5

DURATION OF SUNLIGHT IN UNITS OF 12 HOURS LATITUDE 40° N.

Date:	J	F	M	A	M	J	J	A	S	O	N	D
1	.78	.84	.94	1.05	1.16	1.23	1.25	1.19	1.09	.98	.87	.80
2	.78	.85	.94	1.06	1.16	1.24	1.25	1.19	1.09	.98	.87	.79
3	.78	.85	.95	1.06	1.16	1.24	1.25	1.18	1.08	.98	.87	.79
4	.78	.85	.95	1.06	1.17	1.24	1.24	1.18	1.08	.97	.87	.79
5	.78	.86	.95	1.07	1.17	1.24	1.24	1.18	1.08	.97	.86	.79
6	.79	.86	.96	1.07	1.17	1.24	1.24	1.17	1.07	.97	.86	.79
7	.79	.86	.96	1.07	1.17	1.24	1.24	1.17	1.07	.96	.86	.79
8	.79	.87	.97	1.08	1.18	1.24	1.24	1.17	1.07	.96	.85	.79
9	.79	.87	.97	1.08	1.18	1.24	1.24	1.17	1.06	.95	.85	.78
10	.79	.87	.97	1.09	1.18	1.25	1.24	1.16	1.06	.95	.85	.78
11	.79	.87	.98	1.09	1.19	1.25	1.24	1.16	1.06	.95	.84	.78
12	.79	.88	.98	1.09	1.19	1.25	1.23	1.16	1.05	.94	.84	.78
13	.80	.88	.98	1.10	1.19	1.25	1.23	1.16	1.05	.94	.84	.78
14	.80	.88	.99	1.10	1.19	1.25	1.23	1.15	1.04	.94	.84	.78
15	.80	.89	.99	1.10	1.20	1.25	1.23	1.15	1.04	.93	.83	.78
16	.80	.89	.99	1.11	1.20	1.25	1.23	1.15	1.04	.93	.83	.78
17	.81	.89	1.00	1.11	1.20	1.25	1.22	1.14	1.03	.93	.83	.78
18	.81	.90	1.00	1.11	1.21	1.25	1.22	1.14	1.03	.92	.82	.78
19	.81	.90	1.01	1.12	1.21	1.25	1.22	1.13	1.03	.92	.82	.78
20	.81	.90	1.01	1.12	1.21	1.25	1.22	1.13	1.02	.92	.82	.78
21	.81	.91	1.01	1.12	1.21	1.25	1.22	1.13	1.02	.91	.82	.78
22	.82	.91	1.02	1.13	1.22	1.25	1.22	1.12	1.02	.91	.82	.78
23	.82	.91	1.02	1.13	1.22	1.25	1.21	1.12	1.01	.91	.81	.78
24	.82	.92	1.02	1.13	1.22	1.25	1.21	1.12	1.01	.90	.81	.78
25	.82	.92	1.03	1.14	1.22	1.25	1.21	1.12	1.01	.90	.81	.78
26	.83	.92	1.03	1.14	1.22	1.25	1.21	1.11	1.00	.89	.81	.78
27	.83	.92	1.03	1.14	1.22	1.25	1.20	1.11	1.00	.89	.81	.78
28	.83	.93	1.04	1.15	1.23	1.25	1.20	1.11	.99	.89	.80	.78
29	.83	.93	1.04	1.15	1.23	1.25	1.20	1.10	.99	.88	.80	.78
30	.84		1.05	1.15	1.23	1.25	1.19	1.10	.99	.88	.80	.78
31	.84		1.05		1.23		1.19	1.10		.88		.78

Table for Determining Monthly Potential Evapotranspiration at Temperatures Above 26.5°C
(For daily values divide by 30)

T°C	PE(mm)	T°C	PE(mm)	T°C	PE(mm)
26.5	135.0	30.5	165.2	34.5	181.8
27.0	139.5	31.0	168.0	35.0	182.9
27.5	143.7	31.5	170.7	35.6	183.7
28.0	147.8	32.0	173.1	36.0	184.3
28.5	151.7	32.5	175.3	36.5	184.7
29.0	156.4	33.0	177.2	37.0	184.9
29.5	158.9	33.5	179.0	37.5	185.0
30.0	162.1	34.0	180.5	38.0	185.0

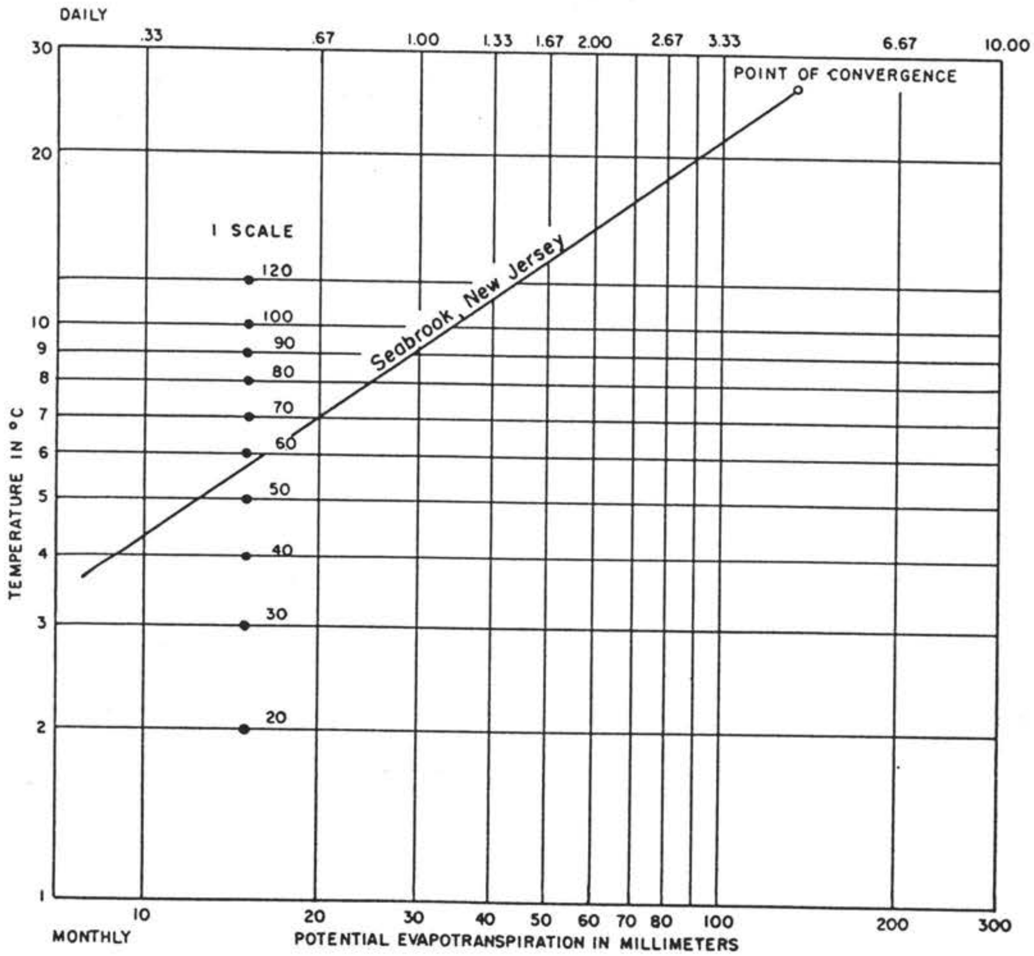

Figure I.1
Nomogram for determining potential evapotranspiration from mean temperature.

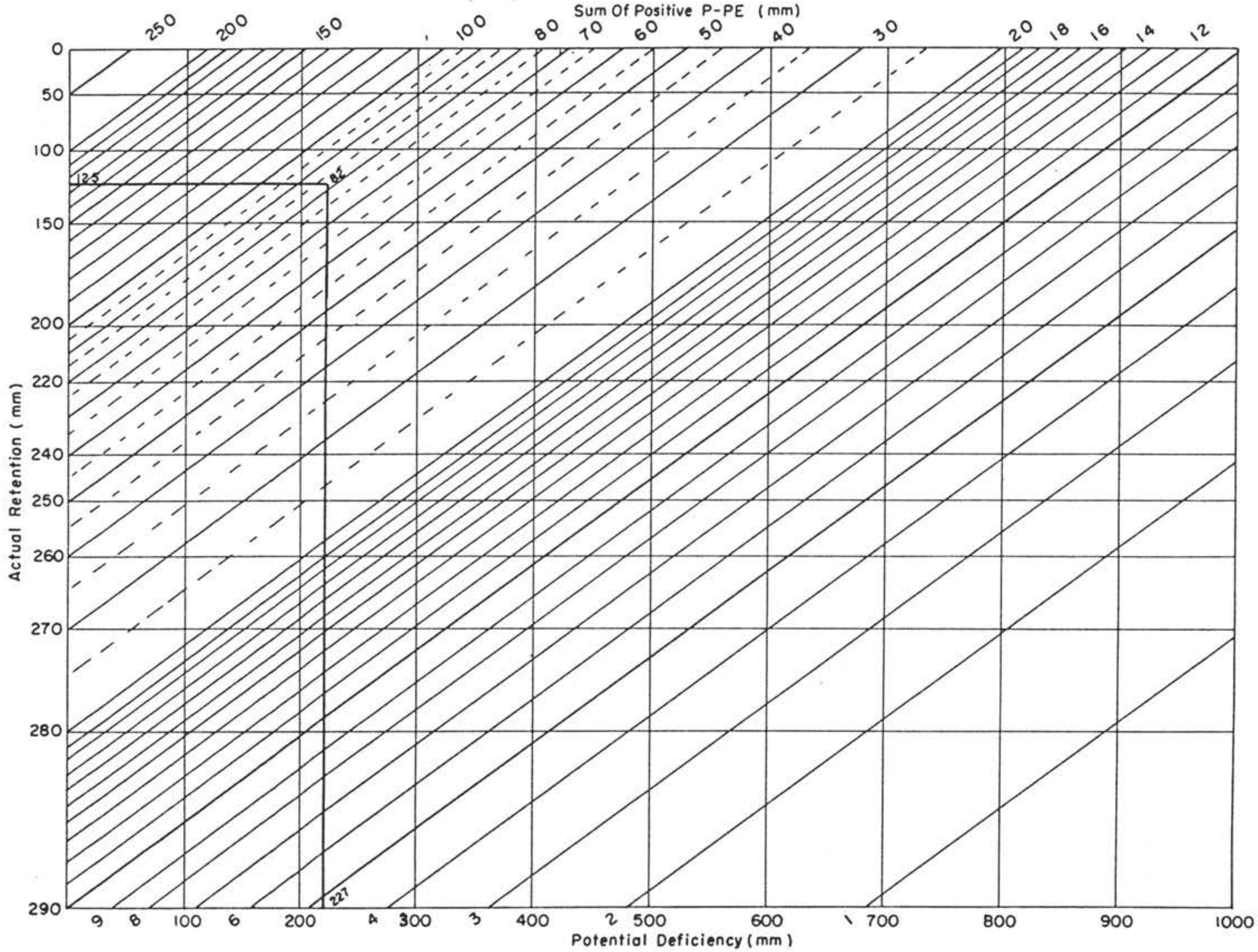

Figure I.2
Nomogram to permit determination of value of potential water loss with which to start accumulation of negative P-PE (line 7).