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Variability of Evapotranspiration in Illinois

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ABSTRACT

The average annual evapotranspiration for the state of Illinois is derived from the average precipitation and the average runoff from 43 stream gaging stations in and around the state. The lack of correlation between the evapotranspiration pattern and gross physiographic features is pointed out.

A comparison is made of three equations for calculating potential evapotranspiration. The reasons for selecting the Hamon equation are given. The Hamon equation is tested for applicability to the climate of Illinois by inserting it into a soil moisture accounting procedure and checking the results against measured amounts of soil moisture at three widely dispersed points in the state. The equation is used to calculate the evapotranspiration at 67 temperature-measuring stations in the state for an average year, a wet year, and a dry year. The water surplus or deficit and their spatial variability during these years is pointed out.

INTRODUCTION

Long-term averages of the amount of precipitation falling upon the state of Illinois show that an average of 99 billion gallons per day may be expected. By far the largest usage of this enormous amount of water is in the combined processes of evaporation and transpiration, which account for approximately 75 percent of the average daily precipitation. More than half of this percentage passes through the roots, stems, and leaves of growing plants; and it is possible that part of this water could be saved in the soils for future use if plants could be forced to reduce their rate of transpiration. It has been hypothesized by Roberts[1] and limited experiments have shown that the inclusion of fatty alcohols in the soils of growing test plants reduced the water required by the plants up to 40 percent.

Little is known of the variability in time and space of the processes of evaporation and transpiration since efforts at measuring evaporation have been widely spaced and transpiration is difficult to measure. The combination of the two processes, called evapotranspiration *(ET),* has been evaluated by Schicht and Walton[2] on three small watersheds in central Illinois. They found that annual evapotranspiration can be estimated within 2 or 3 inches by multiplying the mean annual potential evapotranspiration by a factor of 0.84 for years of normal or above normal precipitation. During dry years the factor might drop as low as 0.66.

This report establishes a basis for further study of the role of evapotranspiration in the water budget by delineating the spatial variability and by presenting a limited study of the temporal variability of evapotranspiration for the state. Because no technique has been devised whereby evapotranspiration may be measured direcdy, a procedure to determine the average evapotranspiration for the state from the average precipitation and average runoff from a number of watersheds in different parts of the state has been used. In addition, the role played by potential evapotranspiration as determined from the Hamon equation of 1961 has been evaluated and applied to estimating the water needs of the state in a dry year. Potential evapotranspiration has been defined as the water extracted from the supply of soil moisture by an extensive short green crop cover, completely shading the ground and never short of water.

EVAPOTRANSPIRATION

The classical means of calculating the average evapotranspiration from a stream basin has been to calculate the difference between measured streamflow and precipitation. In general, this measurement must involve records of precipitation and runoff which are continuous over many years. Basic in the calculation is the assumption that interbasin exhange of water does not occur and that the average basin storage is the same at the close of the period as it was at the beginning. Research reported by Russell[3] concludes that the static mean basin storage assumption is met for the 43 basins in and around the state of Illinois for which flow records are available for at least 25 years ending in 1960. Thirty-four of these basins are within the borders of the state. Table 1 lists the gaging stations, the basins, their length of record, and the average annual runoff. Figure 1 shows the distribution of the basins with respect to each other and to the state boundaries.

It is customary for the volume-flow for a stream basin to be plotted on a base map at the site of the gaging station. However, when the average equivalent depth of basin flow

Figure 1. Stream basins in and around Illinois for which at least 25 years of record were available in 1960

is plotted for comparison with other basins, the centroid of the basin is the appropriate plotting point. The point of plotting becomes important on large basins such as the Upper Fox in Illinois where a distance of 30 miles separates the centroid of the basin and the gaging station (Algonquin). These distances are most important where several basins are gaged at or near their entry into the same major stream. This may be illustrated by the Kankakee and Iroquois Rivers where they join and enter the Illinois River; plotting of the volume-flow at the gaging stations would result in a compression of the values into a small area near their entrance into the Illinois, whereas the centroid points are widely scattered.

To illustrate the difference that the plotting point makes, charts have been plotted using both the gaging sites (figure 2) and the centroids of the basins (figure 3). Lines of equal

Table 1. Long-Term Runoff for Selected Stream Basins in Illinois and Surrounding States

Gaging station	Basin	Mean annual runoff (inches)	Length of record (years)
Kane	Macoupin Creek	8.39	32
Vandalia	Upper Kaskaskia River	9.83	50
New Athens	Lower Kaskaskia River	9.94	36
Plumfield	Upper Big Muddy River	12.78	48
Murphysboro	Lower Big Muddy River	10.80	30
Geneseo	Green River	7.61	25
New Boston	Edwards River	7.80	26
Keithsburg	Pope Creek	7.76	26
Oquawka	Henderson Creek	8.19	26
Momence	Middle Kankakee River	7.18	45
Chebanse	Iroquois River	9.97	37
Wilmington	Lower Kankakee River	7.85	27
Dayton	Lower Fox River	8.27	45
Hanover	Apple River	8.87	26
Lowell	Vermilion River	8.28	29
Oakford	Lower Sangamon River	7.75	33
Monticello	Upper Sangamon River	9.63	50
Seville	Spoon River	8.29	46
Ripley	La Moine River	7.59	39
Forman	Cache River	17.34	36
Wilcox	Little Wabash River	10.84	46
Danville	Vermilion River	9.84	39
Wayne City	Skillet Fork	11.10	43
St. Marie	Embarras River	10.79	49
Algonquin	Upper Fox River	7.57	45
Freeport	Pecatonica River	9.00	46
Como	Rock River	8.94	46
Augusta, Iowa	Skunk River	6.85	46
Manchester, Iowa	Upper Maquoketa River	8.30	27
Durango, Iowa	Litde Maquoketa River	8.10	26
Maquoketa, Iowa	Lower Maquoketa River	8.17	47
New London, Missouri	Salt River	8.40	38
Troy, Missouri	Cuivre River	9.66	38
Eureka, Missouri	Meramec River	11.04	41
Zelma, Missouri	Castor River	16.48	39
Palmyra, Missouri	North River	7.82	25
Taylor, Missouri	South Fabius River	7.92	26
Canton, Missouri	Wyaconda River	7.56	28
Wayland, Missouri	Fox River	7.77	38
Princeton, Indiana	Patoka River	16.94	26
Bowhng Green, Indiana	Eel River	13.68	29
Monticello, Indiana	Tippecanoe River	11.67	29
Shelby, Indiana	Upper Kankakee River	11.82	38

depth of streamflow have been drawn. Figure 2 shows that a tight gradient is produced by the proximity of the gaging sites on the Kankakee River system. Figure 3, the chart of the centroid-plotted values, reveals that there is no tight gradient for that system.

The lines of equal depth of runoff cannot be taken to be anything other than the average runoff for the region through which they are drawn, not only because they are based upon long-term averages but also because the runoff from any point is a function of a number of factors that tend to cause the runoff from small areas to be irregular in space. These factors are differences is soil types, varying slopes, differences in infiltration from a number of causes, varying farming practices, and the type of growth on the land, whether forest, pasture, or tilled crops. The lines of equal runoff are justified only as a broad indication of the runoff to be expected from a given area of the state. With lines of equal precipitation the runoff may be used to derive lines of equal evapotranspiration.

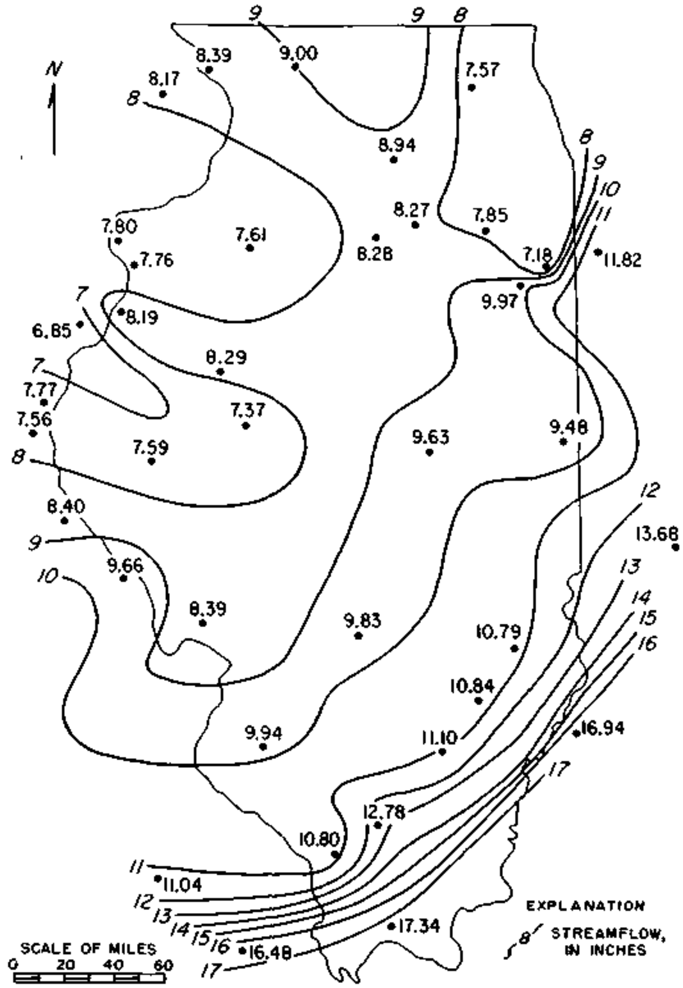

Figure 2. Streamflow with values plotted at gaging stations

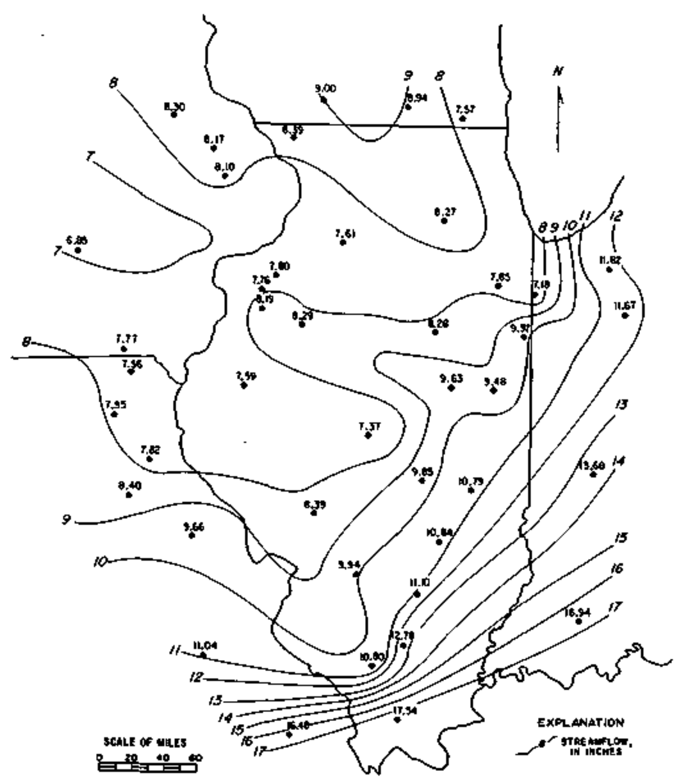

Figure 3. Streamflow with values plotted at centroids of basins

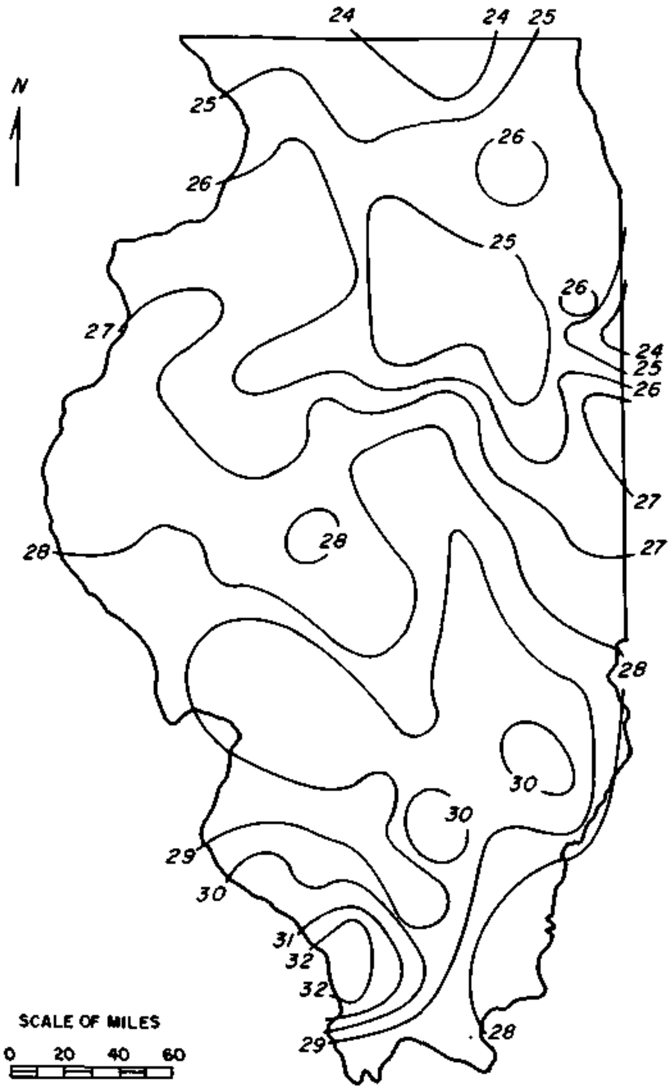

Figure 4. Evapotranspiration in inches determined with flow plotted at gaging stations

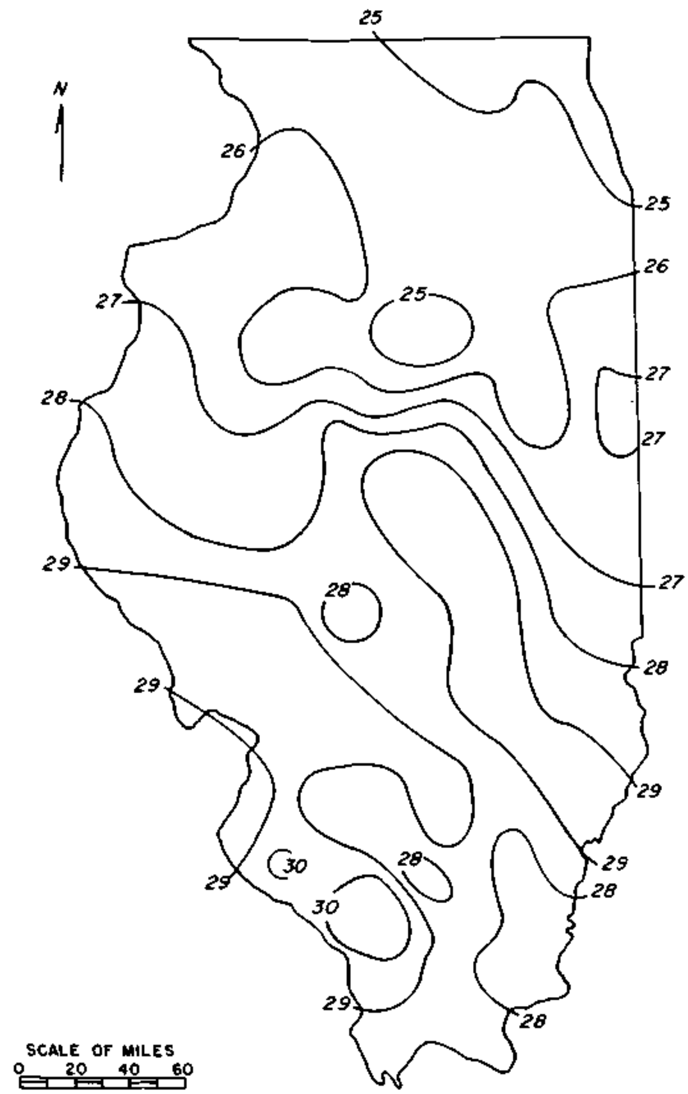

Figure 5. Evapotranspiration in inches determined with flow plotted at centroids of basins

Lines of equal evapotranspiration may be found from charts of isohyets and lines of equal runoff by the alignment of the two charts upon a light table and the drawing of lines through those points at which the runoff may be subtracted from the precipitation. The resultant charts of evapotranspiration for each of the two methods of plotting the streamflow data are shown as figures 4 and 5.

These two evapotranspiration charts reveal a more serious difference in the methods of plotting the streamflow data. Figure 4 has the same general features of higher and lower areas of evapotranspiration as figure 5. However, the magnitude of values of evapotranspiration are greater on figure 4 than on figure 5 by 2 inches per year. Figure 4 has a minimum value of 24 inches in the northern part of the state and a maximum of 32 inches in the southern part while figure 5 has values of 25 inches and 30 inches near the same locations. An outstanding difference is to be found in the Kankakee Basin where figure 4 has a value of 24 inches and figure 5 has a value of 26 inches. This area has been strikingly affected by the proximity of the stream gaging stations. It is believed that the centroid-plotted chart is the more reasonable of the two.

In general, the range of evapotranspiration in the state is from 25 inches along the northern border to 30 inches in the southern part of the state. However, the transition is not smooth from north to south. Figure 5 shows that there is an area of low evapotranspiration in the north-central portion of the state along with a relatively low area in the extreme southeastern portion. The long ridge of high evapotranspiration amounting to 29 inches from the southeast border to the center of the state is notable. The greatest evapotranspiration may be expected in the southwestern portion of the state.

The pattern of evapotranspiration over the state might be expected to be related to some identifiable feature of the physiography such as, slope, slope direction, and soil characteristics. However, the gross physiographic boundaries as listed by Leighton, Ekblaw, and Horberg[4] do not appear to be related to the evapotranspiration pattern. The patterns are illustrated in figure 6.

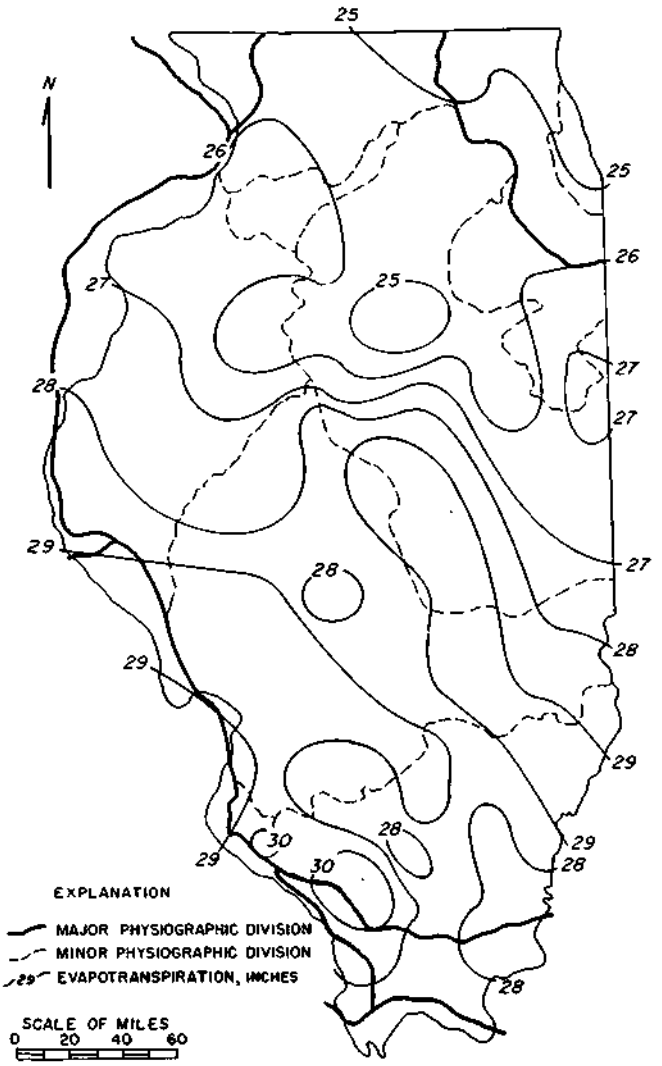

Figure 6. Physiographic divisions compared with evapotranspiration pattern

POTENTIAL EVAPOTRANSPIRATION EQUATION

Beginning with Dalton,[5] a number of equations based upon empirical data or semitheoretical reasoning have been proposed to estimate the amount of evapotranspiration. Dalton deduced that the amount of evaporation is proportional to the water vapor pressure deficiency of the air, which may be expressed as

$$e_s - e_a \propto \text{evaporation}$$

in which e_s is the saturation vapor pressure of the air and e_a is the actual vapor pressure of the air. This relationship when modified by the vapor pressure excess of the evaporating water is the basic assumption in most equations for determining evapotranspiration.

Instruments to measure the rate of movement of water vapor have not been developed except in exploratory experiments such as those of Thornthwaite[6] and Suomi.[7] Certainly, no systematic climatological survey has been made with instruments capable of measuring the flux of water vapor. Thus, those equations which have been most widely used have attempted to circumvent this instrumentation shbrtcoming through the assumption that evapotranspiration is directly related to the daily mean temperature and/or the solar radiation impinging upon the evaporation surface. The development of a theory that considers transpiration of plants is further complicated by the physiology of the plants, particularly the opening of the leaf stomata during the daylight hours in general but on occasion at other times.

The equation developed by Penman[8] in England has been generally accepted as the most reliable because it takes into account a greater number of the variables that could affect the amount of evapotranspiration. These variables include at least the daily wind travel, mean daily temperature, relative humidity, and cloudiness. At the same time, because it does require knowledge of many variables, the equation is cumbersome to calculate and may be unusable because some of the variables are not commonly measured.

An equation requiring a knowledge of fewer variables has been developed by Thornthwaite[9] and has been widely used in the United States. This equation requires only two variables, the daily mean temperature and the day length in units of 12 hours. Tables to facilitate the use of the equation also have been developed by Thornthwaite, since the calculation of potential evapotranspiration with the original equation is quite laborious.

Hamon[10] simplified the Thornthwaite equation to derive an equation which is forced to agree with the Thornthwaite equation through a coefficient. The Hamon equation is of the form

$$PE = 0.0055\, D^2\, Pt$$

in which PE is the daily potential evapotranspiration in inches, 0.0055 is the coefficient, D is the day length factor in units for a standard 12-hour day, and P_t is the saturation vapor density in grams per cubic meter at the daily mean temperature. Tables for D and P_t are given in the original paper by Hamon.[11]

Oddly enough, the Hamon equation can be used to estimate evapotranspiration at temperatures below the freezing point of water, whereas the Thornthwaite equation cannot. In the Thornthwaite equation, the mean temperatures in degrees centigrade appears in the numerator, so that temperatures below freezing result in negative values of *PE*. Temperature is not a direct factor in the Hamon equation, and since saturation vapor pressure density reaches zero at a temperature not reached in the earth's atmosphere, the Hamon equation always yields a small amount of *PE*.

Table 2 lists the annual potential evapotranspiration as calculated with the Penman, Thornthwaite, and Hamon equations from meteorological data provided by the Radiological Physics Division of the Argonne National Laboratories at Argonne, Illinois. Argonne is the only station in Illinois known to measure the variables required to compute the Penman equation as well as the Thornthwaite and Hamon equations.

The annual potential evapotranspiration calculated by the Penman equation is consistently larger than that calculated by the Hamon and Thornthwaite equations for the Argonne data. This also is illustrated by figure 7, which depicts *the* average monthly potential evapotranspiration

Table 2. Comparison of Annual Potential Evapotranspiration for Argonne, Illinois, from Different Equations

Year (May-April)	Annual potential evapotranspiration (*inches*) Hamon	Thornthwaite	Penman
1951	25.42	24.61	30.13
1952	26.49	23.87	32.95
1953	27.98	26.65	34.29
1954	27.19	25.97	33.11
1955	28.28	25.96	33.42
1956	27.40	26.30	33.58
1957	26.53	25.50	31.33
1958	25.35	24.47	30.80
1959	26.05	26.99	33.31

for the nine years of record from Argonne. Most striking in figure 7 is the large difference between the potential evapotranspiration from the Penman method and that from the other two methods during the spring and early summer months, an average difference of approximately 1 inch per month. The difference in the estimated potential evapotranspiration between Penman and Thornthwaite during the spring months generally has been recognized in the literature as a failure of the Thornthwaite equation.

Gerber and Decker[12] found that the Penman equation yielded results mat are highly correlated to evapotranspiration loss when the soil surface between row crops is wet, but that it failed to estimate evapotranspiration when the soil surface is dry. Pierce[13] found, in general, that the actual evapotranspiration exceeded the Thornthwaite potential evapotranspiration whenever soil moisture conditions were favorable.

Because it had undergone only limited testing, and because of its greater simplicity in the number of variables required and in the amount of calculation, the Hamon equation was selected in this investigation for testing against values of evapotranspiration as indicated by gravimetric measurements of soil moisture loss. Such an equation would be very useful in accounting for soil moisture loss if it yielded reasonably accurate estimates of potential evapotranspiration.

TEST OF HAMON EQUATION FOR DAILY SOIL MOISTURE CALCULATIONS

The Hamon equation, as well as most others, cannot give a direct estimate of the actual evapotranspiration that has occurred; instead, it provides an approximation of the potential evapotranspiration from knowledge of the mean daily temperature and tabular values of the day length normalized to a 12-hour day. It can be assumed that when soil moisture is not limiting to crop growth, the potential evapotranspiration is the same as the actual evapotranspiration, if possible variations in transpiration efficiency of various field crops can be neglected.

The Agronomy Department of the University of Illinois measured soil moisture from samples that were collected within the state of Illinois between 1956 and 1960 at three locations, Urbana, Shabbona, and Carbondale (unpublished data). Sampling for each year was started at the end of the frozen soil period in the spring and was continued at

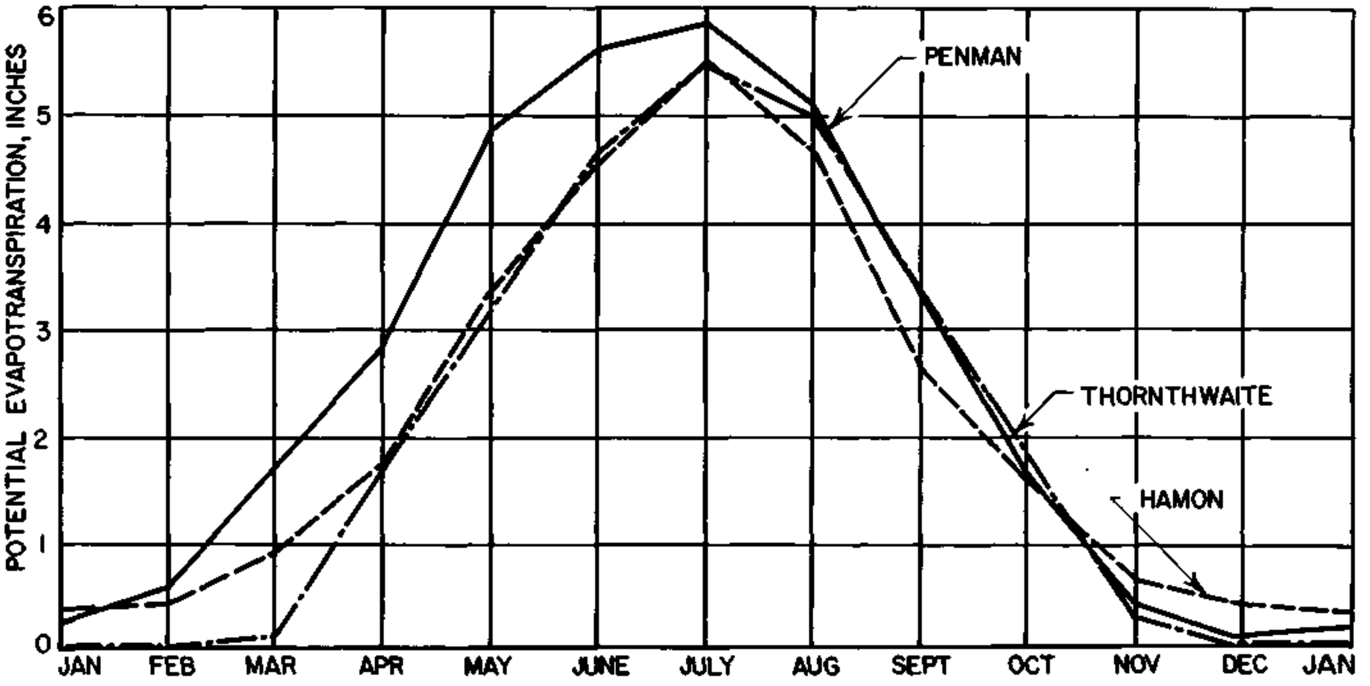

Figure 7. Average monthly potential evapotranspiration in inches for nine years at Argonne, Illinois, by three methods

intervals no oftener than twice each month until the end of the growing season. Two of the sampling sites, at Urbana and Shabbona, were beneath long established blue grass sod, whereas the Carbondale site was beneath a rye grass sod. A raingage was located within a few feet of each sampling site to provide a record of daily precipitation amounts.

Test cores for the gravimetric samples were made to depths of 60 inches at each of the blue grass sites, but to only 45 inches at the rye grass site at Carbondale. Tests were made to verify that the two types of grasses were capable of using moisture to these respective depths, despite the shorter rye root systems (20 inches for the blue grass at Urbana, and 6 inches for the rye grass at Carbondale). The gravimetrically determined soil moisture values were assumed to be correct although it is known that the method is subject to error due to disturbance of the soil and to naturally occurring spatial variations in soil characteristics.

The daily potential evapotranspiration for each of the three sites was computed with the Hamon equation from the average daily temperature and the appropriate day length factor. The daily potential evapotranspiration amounts were then subtracted successively from the measured soil moisture values, as shown in figures 8, 9, and 10. Daily rainfall when it occurred, was added to the soil moisture up to the limit of field capacity, the excess being assumed lost by water drainage, either surface or subsurface flow. This oversimplification of a complex process was necessary since there was no way of estimating the division of water absorbed by the soil and the amount drained by surface and subsurface flow.

Each measured amount of soil moisture was used as the starting point for a continuous accounting of soil moisture loss or gain until the next moisture sample was obtained. Uninterrupted calculations through an entire season were not used for two reasons: 1) if the first sample in the spring was not representative, all accounting based upon it would be in error; and 2) soil moisture recovery occurring after the plant wilting point had been reached would be lost.

However, uninterrupted soil moisture accounting for the full growing season from a spring gravimetric measurement might be feasible. This is shown by the data comparisons made for 1957 and for 1959 from Urbana, shown in figure 11. The 1957 graph seems to indicate that the accounting procedure, simple as it is, gives a reasonable approximation of the true soil moisture at the end of the season and, presumably, at any date between the beginning and ending dates. However, 1957, unlike 1959, was a year when soil moisture did not reach the wilting point for blue grass at Urbana. The calculations for 1959 show that the sod would have reached a zero moisture content on August 8 if it had been possible for evapotranspiration to continue at the potential rate, and that the soil moisture would have stayed at or near zero until September 23, when heavy rains occurred. The gravimetric tests made from August through September do not indicate zero soil moisture; rather, they indicate that the grass quit using soil moisture when

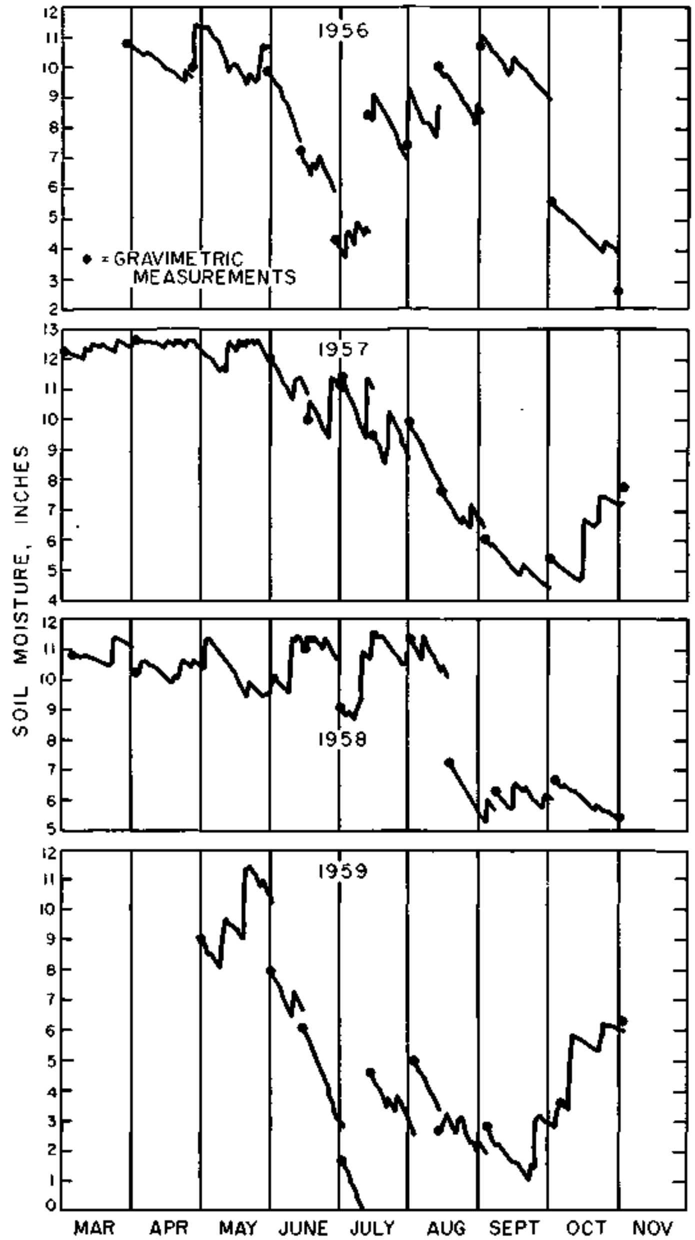

Figure 8. Comparison of measured and calculated values of soil moisture at Urbana

the available moisture was approximately 3 inches.

Figures 8, 9, and 10 show the daily accounting of soil moisture for the sampling sites at Urbana, Shabbona, and Carbondale, respectively. At each site there are certain time intervals during which the Hamon equation calculations seem to be well related to the soil moisture measurements. Intervals during which the two values do not appear to be related may be attributed to the unavoidable errors in the gravimetric method or to the failure of the grass to transpire at the potential rate.

Close correspondence between calculated and measured amounts is illustrated at Urbana (figure 8) for the periods of March 29 to April 27 and May 29 to June 18, 1956; August 1 to August 15, 1957; April 2 to June 16 and July

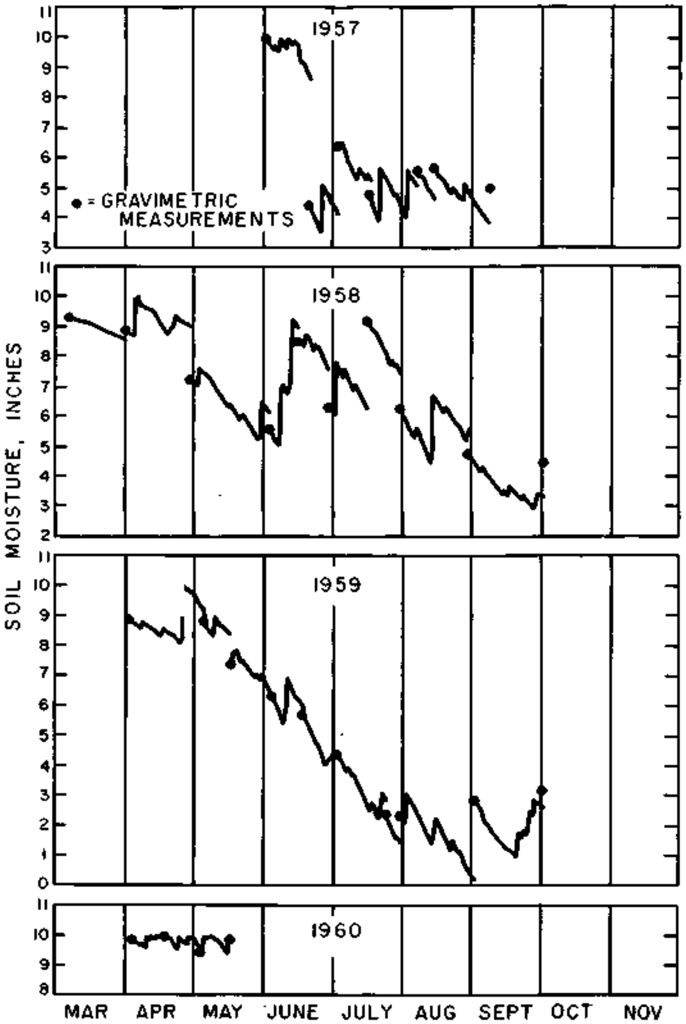

Figure 9. Comparison of measured and calculated values of soil moisture at Shabbona

1 to July 16, 1958; and September 4 to November 2, 1959. However, for Urbana in 1957 there is a period when the soil moisture was above field capacity, and comparison of the calculated soil moisture with the measured values, although good, should not be considered since both values are at or near an assumed limit. Also, on several occasions, notes in the 1959 soil moisture log book for Urbana mentioned that the character of the soil sample had obviously changed. In particular, the 1959 log included a notation that soil moisture had increased by 3.0 inches between July 2 and July 15 when there had been no rain, a feature also shown on figure 8.

In figure 9 for Shabbona in 1959 there is an interval from May 17 until July 31, if the measurement of July 24 is ignored, when the calculated soil moisture amounts fit very well. Only one day had a rainfall amount over 1.0 inch but a number of days had lesser amounts. During the latter part of July 1959 at Shabbona, soil moisture became inadequate to sustain the growth of the grass, and the grass began to wilt and ceased transpiring moisture. This is graphically illustrated in figure 9 where the daily accounting of soil moisture tends toward zero during August but the measured amount of September 2 shows that

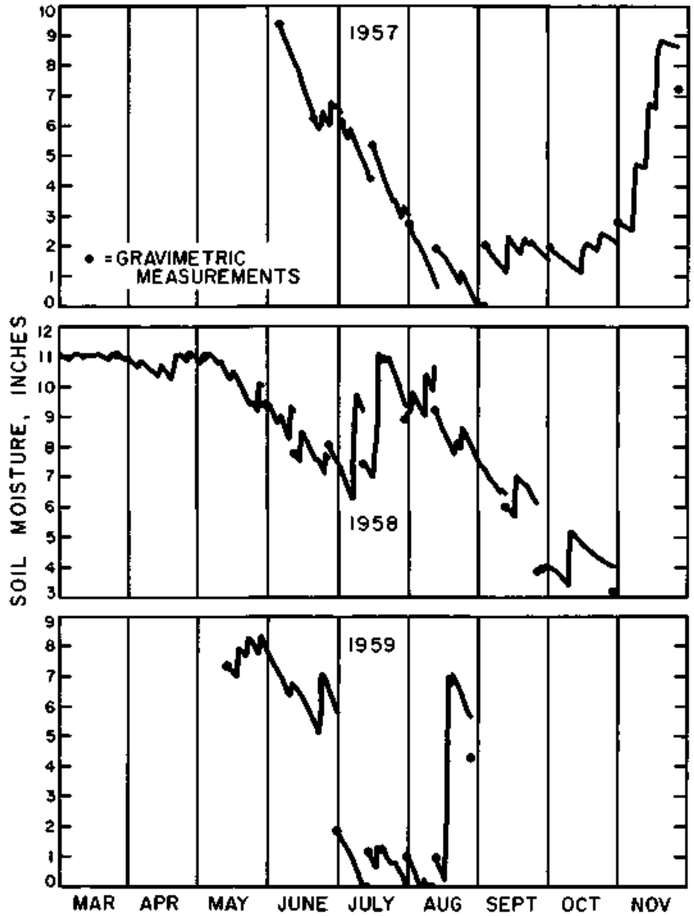

Figure 10. Comparison of measured and calculated values of soil moisture at Carbondale

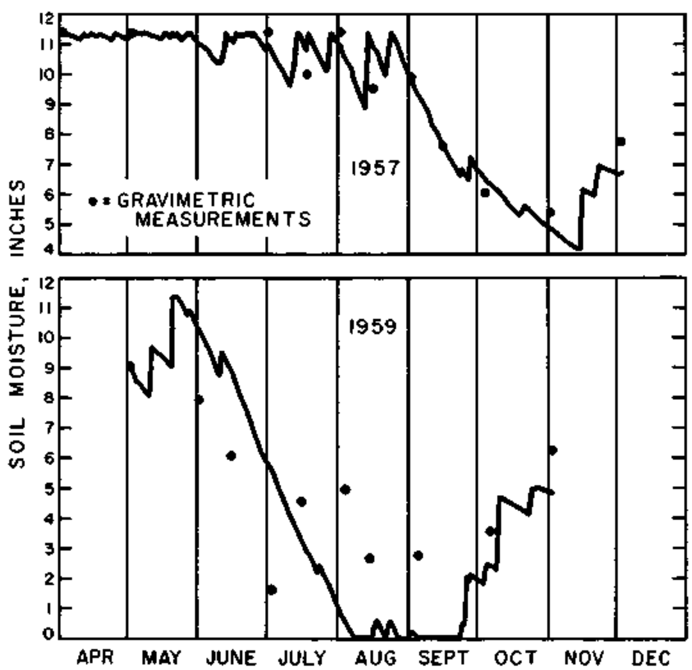

Figure 11. Comparison of seasonal calculations and measurements of soil moisture at Urbana

more than 2.5 inches of moisture remained. Similar examples may be found on figures 8 and 10.

Periods when calculated moisture accounting corresponded well with measured soil moisture amounts at Carbondale are June 5 to August 1, 1957; June 12 to June 27, 1958; and July 11 to July 31, 1958. These are illustrated in figure 10. Measurements at Carbondale were infrequent in 1959 until plant wilting occurred, so that conclusions may not be drawn for that year. There is some indication on the 1958 graph for Carbondale that the assumption of total absorption of all precipitation when soil moisture is below field capacity is not valid. When heavy rains occurred between gravimetric measurements (June 27 to July 11) the calculation for soil moisture at the end of the period is substantially overestimated. Nevertheless, the use of water by the plants appears to occur at the potential rate, as shown by the very close agreement between calculation and measurement from June 5 to June 20, 1957, when no rain occurred but soil moisture levels were high.

The above examples indicate that, from March through November in Illinois, the Hamon equation may be used to account for soil moisture loss when moisture is adequate for grass growth.

MEAN ANNUAL POTENTIAL EVAPOTRANSPIRATION FOR ILLINOIS

The range of average annual potential evapotranspiration for the state of Illinois is delineated on the map in figure 12. The isolines on this map depict the *PE* calculated from the Hamon equation and based upon the 45-year normal temperatures by months for 82 Weather Bureau reporting stations.[14] (See figure 15 for locations of these stations.) Figure 13 shows the 45-year mean temperature for Illinois.

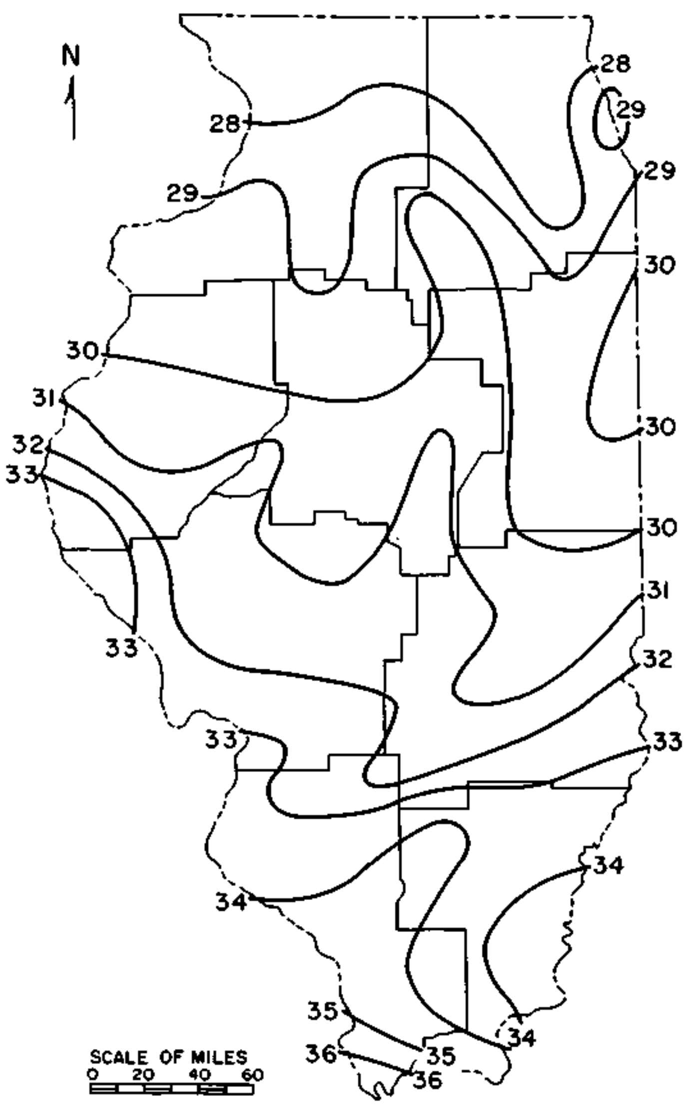

Figure 12. Mean annual potential evapotranspiration, inches (45-year)

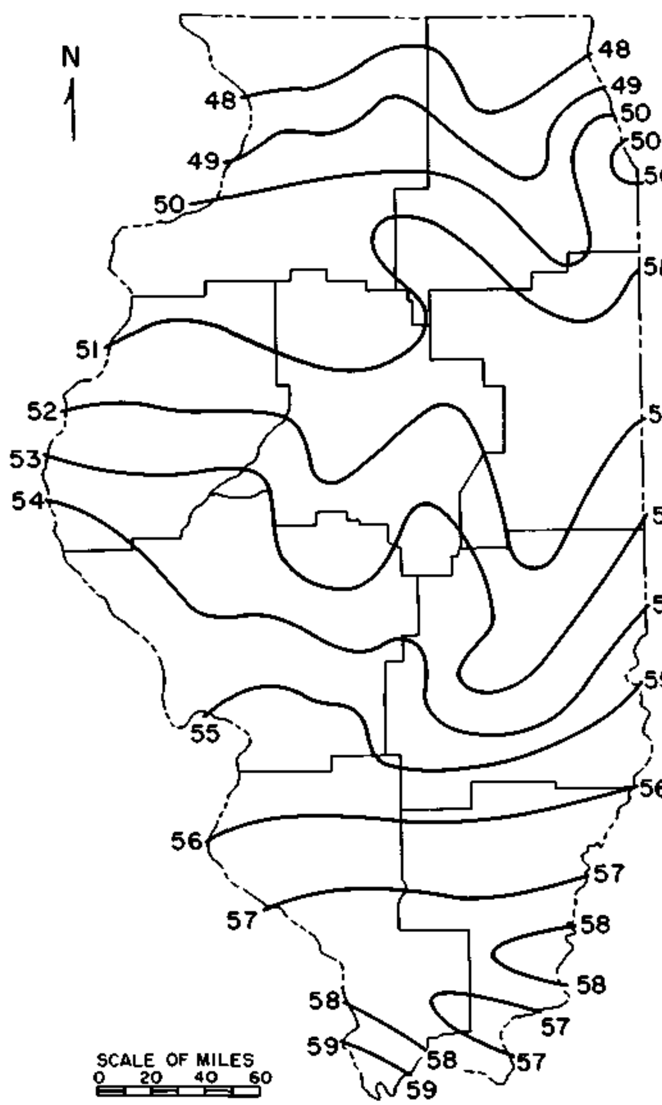

Figure 13. Mean annual temperature, °F (45-year)

In the calculations for figure 12, the monthly values of *PE* have been adjusted to the summation of daily average temperatures by the relationship

$$\Sigma PE_d = KPE_m$$

where K has a value of 1.04; and PE_d and PE_m are the daily and monthly average *PE* respectively. Since *PE* is not a linear function of temperature, as shown by figure 14, K is necessary to equate PE_m to ΣPE_d

Figure 12 shows that the northern part of the state has the lowest *PE* and the southern tip of the state has the highest *PE.* The heat island caused by the urban area of the city of Chicago is prominent as an area of increased *PE.*

As would be expected, there is a close correspondence between figures 12 and 13 although the correspondence is not exact. The pattern of the isolines of the two charts are very similar, but the gradients are different because of the nonlinearity of the PE-temperature relationship.

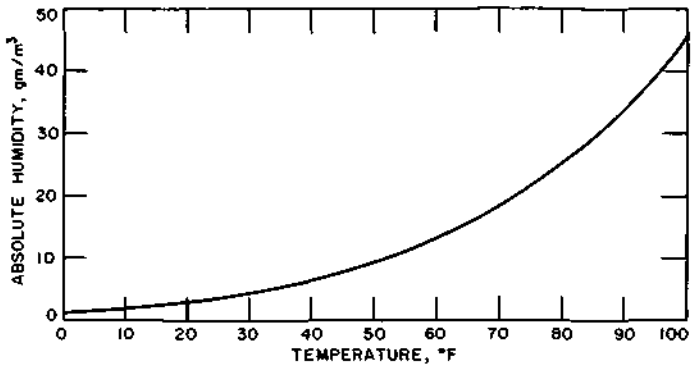

Figure 14. Variations of absolute humidity with temperature

MOISTURE VARIABILITY

The variability of moisture for evapotranspiration in the normal year, in selected wet and dry years, and within varying geographical areas of the state, was investigated. For the areal study, the divisions of the state established by the Crop Reporting Service of the U.S. Department of Agriculture were used. These nine crop-reporting sections, which follow county lines, are shown in figure 15 along with the stations for which potential evapotranspiration was calculated (figure 12).

The values of mean annual potential evapotranspiration given in figure 12 represent the maximum loss that could be expected for the normal year. The mean annual precipitation for Illinois, given in figure 16, was used with figure 12 to obtain the mean annual difference between precipitation and *PE,* as shown in figure 17. For convenience, this difference is here termed the *potential water surplus,* which represents the amount of moisture available in excess of maximum needs for evapotranspiration. When the amount of moisture was less than the maximum evapotranspiration needs, the difference is termed the *water deficit* and is expressed as a negative value.

For the normal year, then, figure 17 indicates that no area of Illinois has a shortage of water for evapotranspiration, and that considerable variability exists from place to place within the state. The largest potential surplus (11 inches) is in the Shawnee Hills area of southern Illinois; the lowest surplus value (3 inches) occurs in three locations, all in the northern half of the state.

Areal averages of potential water surplus for the normal year for the nine sections of the state are given in table 3. The variability by sections is not so great as that shown in figure 17, as would be expected since point values have greater variability than areal averages. The Southeast Section, which includes much of the Shawnee Hills area, has the highest average potential water surplus, and the Northeast and West Sections have the lowest.

Average potential water surplus for 1957, which was a relatively wet year, are given in table 3 for the nine state sections. For that year, the Southeast Section had a water surplus of nearly 26 inches, which was more than 17 inches above its normal-year value. The Southwest and East Southeast Sections were not far behind with surplus values of over 23 and 22 inches, respectively. The Northwest Section showed the lowest potential surplus, but this exceeded its normal-year surplus by 1.29 inches.

In 1956, subnormal rainfall prevailed over broad portions of Illinois, particularly in the northern half. The variability of potential water surpluses for 1956 is depicted in figure 18. Large negative values, ranging from -6 to -12 inches, appear in a band across north-central Illinois, in portions of the west, and in the lower southeastern corner. However, in the south-central part of the state two small areas show surpluses of 8 inches or more, which are near or slightly above the normals shown in figure 17.

Table 3. Normal and Abnormal Annual Water Surpluses

(Amounts in inches)

Section	Normal-year water surplus (45-year)	Selected wet year, 1957: Water surplus	Selected wet year, 1957: Amount above normal	Selected dry year, 1956: Water deficit or surplus	Selected dry year, 1956: Amount below normal
Northwest	5.51	6.80	1.29	-1.89	7.40
Northeast	4.27	10.57	6.30	-6.04	11.31
West	4.30	8.30	4.00	-4.64	8.94
Central	5.18	10.76	5.58	-4.54	9.72
East	5.39	14.24	8.85	-1.79	7.18
W. Southwest	5.96	11.28	5.32	-0.33	6.29
E. Southeast	8.50	22.36	13.86	2.79	6.71
Southwest	8.19	23.84	15.65	2.08	6.11
Southeast	8.66	25.90	17.24	1.28	7.38

Section averages for 1956, listed in table 3, indicate large negative values for the Northeast, West, and Central Sections. The Southwest, Southeast, and East Southeast Sections had some water surplus, but the amounts were considerably below their normal-year surpluses.

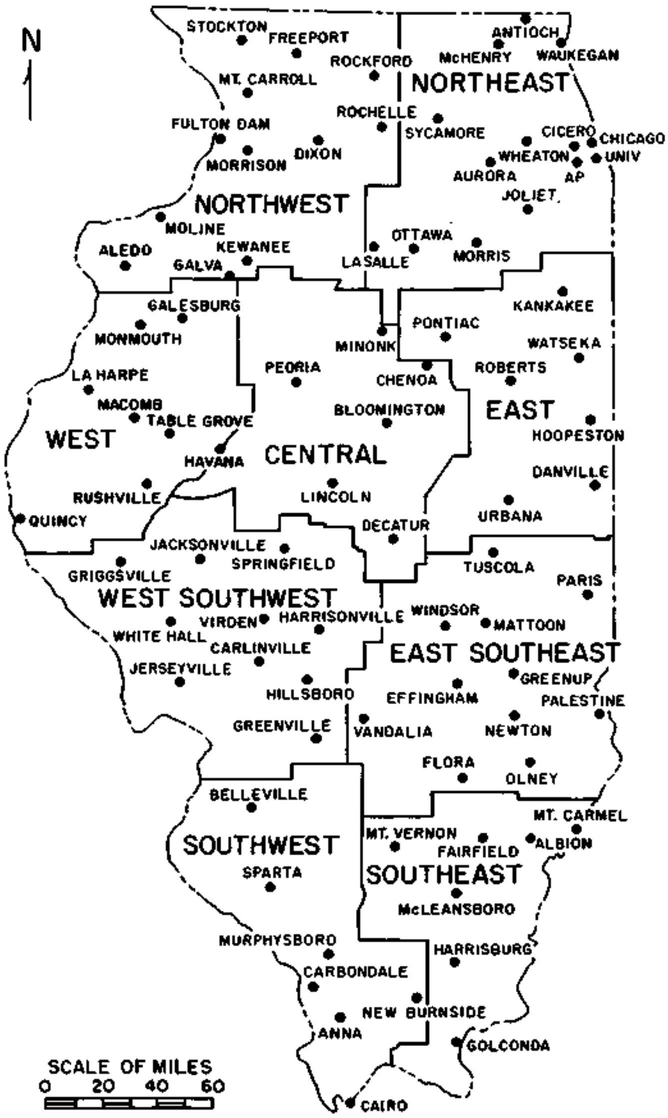

Figure 15. Crop-reporting sections in Illinois and locations of weather stations

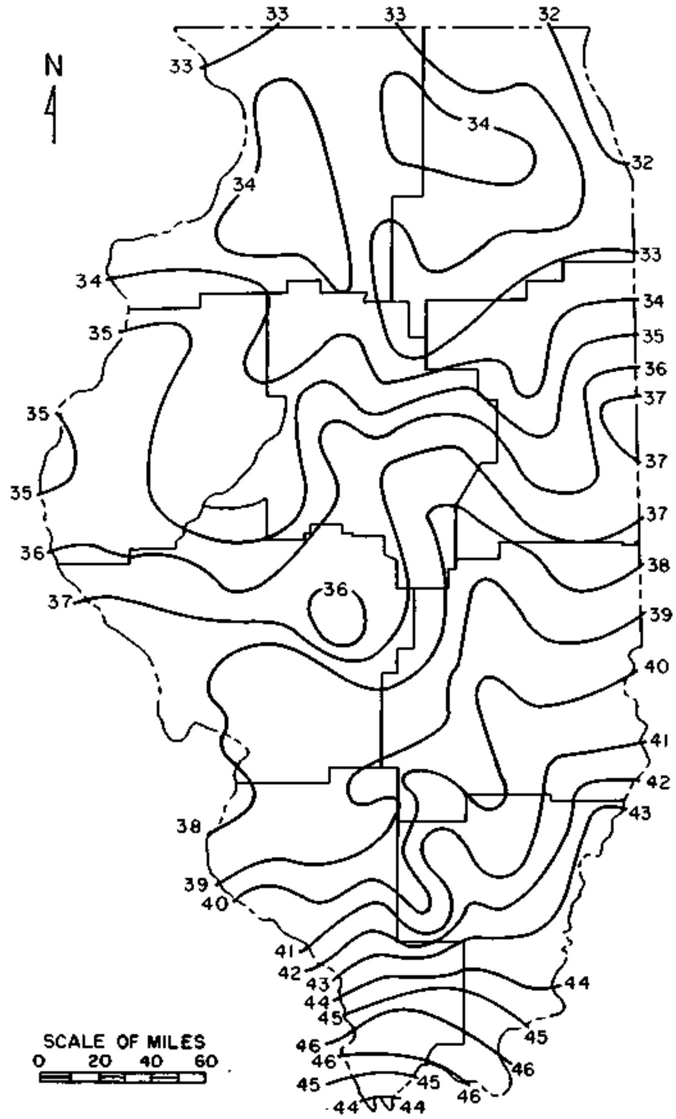

Figure 16. Mean annual precipitation, inches (1906-1950)

Comparison of the 1956 sectional averages with figure 18 indicates that the average potential water surplus for a particular section does not satisfactorily describe the surplus or deficiency that can occur over smaller areas. Sharp variances from large to small areas will occur as the result of the natural variability of rainfall in Illinois. One section as a whole may have a water surplus, while a small portion of that section experiences severe drought. An example of this in 1956 is the Southeast Section which averaged a potential water surplus of 1.28 inches while one area within the section had a deficiency exceeding 6 inches. As another example, the Northeast Section had an average value of -6.04 inches, but values within the section ranged from -12 to +2 inches.

It would be convenient to be able to say that the water deficit amounts for 1956 in figure 18 represent the actual amounts of water that must be applied through irrigation to supplement precipitation. However, even if pumping and transmission losses within the irrigation system are neglected, the actual amount of water necessary to provide for evapotranspiration at the potential or maximum rate would be greater than the amounts shown in figure 18, as the result of several variability characteristics of Illinois rainfall. The usefulness of actual precipitation is lessened, for instance, when it does not occur at the time it is needed or when a portion of it is lost to surface runoff during heavy rains. If a 10-percent loss of rainfall were attributed to such surface runoff, then a useful estimate of the irrigation necessary to replace soil water used in the evapotranspiration process might be obtained by calculating the loss due to potential evapotranspiration and subtracting 90 percent of the rainfall from it. As an example, the small area in north-central Illinois that shows a water deficit of -12 inches on figure 18 would require approximately 14 inches of irrigation water instead of 12 inches.

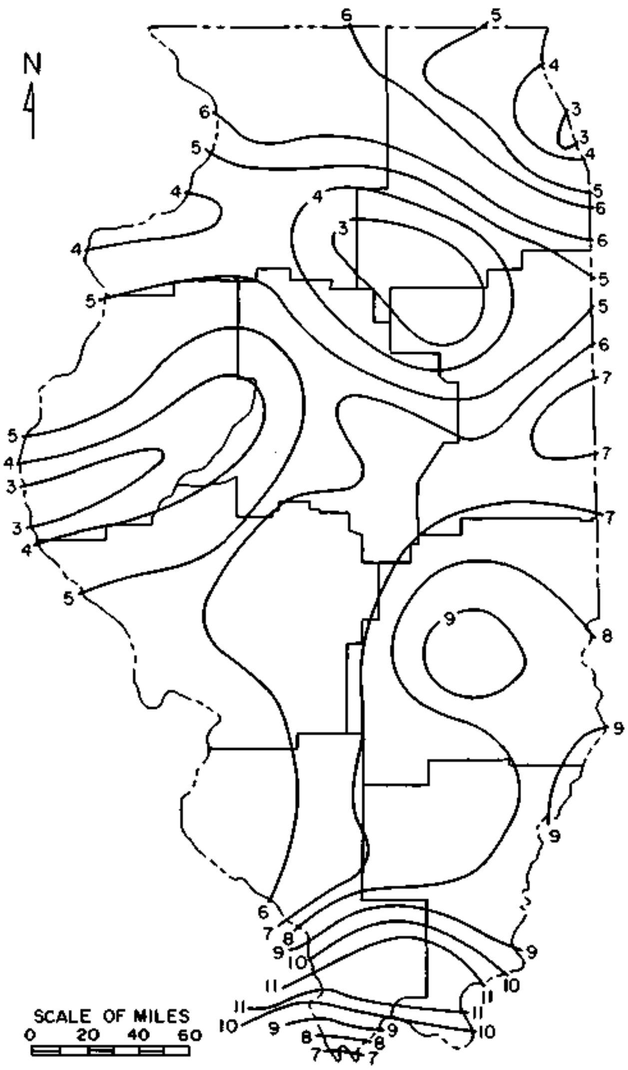

Figure 17. Mean annual difference between precipitation and potential evapotranspiration, or the potential water surplus, inches

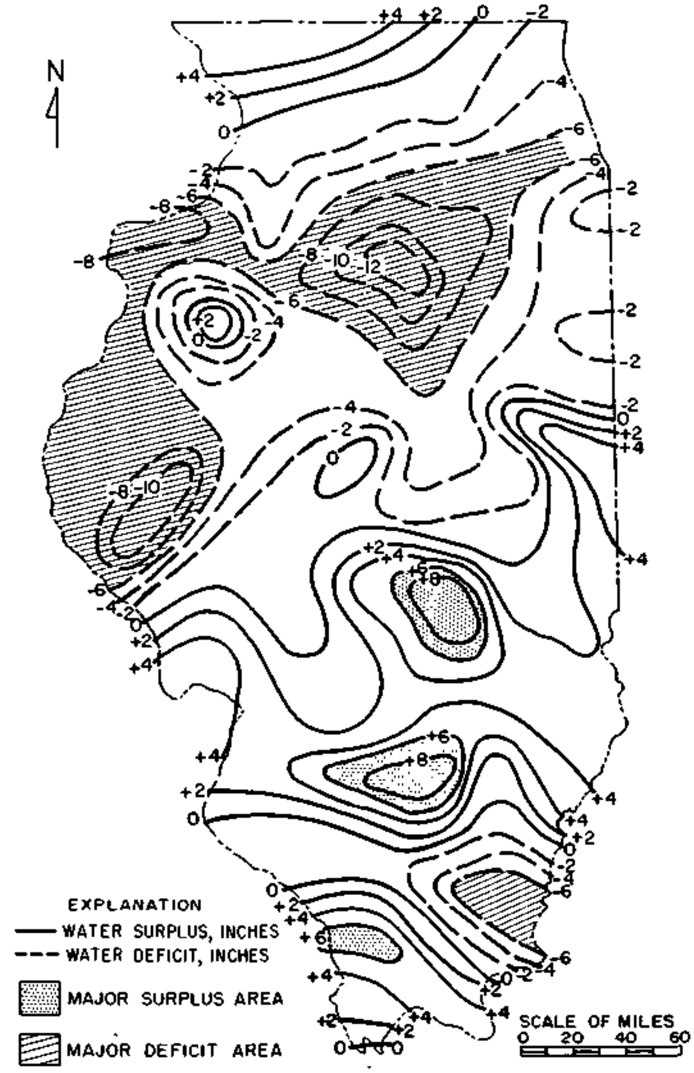

Figure 18. Potential water surplus in 1956 (difference between precipitation and potential evapotranspiration)

SUMMARY

The annual average evapotranspiration was found to vary from 25 inches in northern Illinois to 30 inches in southern Illinois. The greatest evapotranspiration should be expected in the area between Carbondale and DuQuoin. The lowest evapotranspiration should be expected in the northeast corner of the state and in the Minonk area. An area of lower than the area average evapotranspiration is found centered in the Shawneetown area whereas a ridge of above average evapotranspiration extends from Mt. Carmel to Lincoln forming a sharp gradient in the average evapotranspiration between Minonk and Lincoln.

The Hamon equation for calculating potential evapotranspiration was found to be able to determine the loss in soil moisture as compared with gravimetric measurements. The potential evapotranspiration was found to vary from a high of 36 inches in the Shawnee Hills to a low of 26 inches near the Wisconsin border for an average year. On the average it was found that no area of the state receives so little precipitation that evapotranspiration would exceed it. However, it was found that the Shawnee Hills should experience the largest water surplus whereas the northeast corner of the state should have the least water surplus. In 1956 the. Northeast Section experienced a water deficit of more than 6 inches. In 1957 the Southeast Section had a water surplus above potential requirements of almost 26 inches. However, whether a wet or dry year, the variability over small distances can amount to more than 100 percent of the Section average.

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