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AN APPROACH TOWARD A RATIONAL CLASSIFICATION OF CLIMATE

C. W. THORNTHWAITE

[With separate map, Pl. I, facing p. 94]

THE direction that the modern study of climate has taken has been dictated largely by the development of meteorological instruments, the establishment of meteorological observatories, and the collection of weather data. The catalogue of climatic elements consists of those that are customarily measured and usually includes temperature, precipitation, atmospheric humidity and pressure, and wind velocity. Increasingly, climatic studies have tended to become statistical analyses of the observations of individual elements. Because of this, climatology has been regarded in some quarters as nothing more than statistical meteorology.

THE ROLE OF EVAPORATION AND TRANSPIRATION

But 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 gauge 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, no instrument has yet been perfected to measure the water movement from the earth to the atmosphere, and consequently we know next to nothing about the distribution of evapotranspiration in space or time.

We cannot tell whether a climate is moist or dry by knowing the precipitation alone. We must know whether precipitation is greater or less than the water needed for evaporation and transpiration. Precipitation and evapotranspiration are equally important climatic factors. Since precipitation and evapotranspiration are due to different meteorological causes, 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 evaporates or than the vegetation uses. The surplus moves through the ground and over it to form

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➤ DR. THORNTHWAITE, consulting climatologist, has spent the growing seasons of 1946 and 1947 in experiments "fitting crops to weather" (see p. 4).

streams and rivers and flows back to the sea. In other places, month after month, there is less water in the soil than the vegetation would use if it were available. There is no excess of precipitation and no runoff, except locally where the soil is impervious and cannot absorb the rain on the rare occasions when 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 almost never coincides with the changing demands for water.

Where precipitation is in excess of water need, the climate is moist. Where the water deficiency is large in comparison with the need, the climate is dry. Where precipitation and water need are equal or nearly equal, the climate is neither humid nor arid.

POTENTIAL EVAPOTRANSPIRATION AS A CLIMATIC FACTOR

The vegetation of the desert is sparse and uses little water because water is deficient. If more water were available, the vegetation would be less sparse and would use more water. There is a distinction, then, between the amount of water that actually transpires and evaporates and that which would transpire and evaporate if it were available. When water supply increases, as in a desert irrigation project, evapotranspiration rises to a maximum that depends only on the climate. This we may call "potential evapotranspiration," as distinct from actual evapotranspiration.

We know very little about either actual evapotranspiration or potential evapotranspiration. We shall be able to measure actual evapotranspiration as soon as existing methods are perfected. But to determine potential evapotranspiration is very difficult. Since it does not represent actual transfer of water to the atmosphere but rather the transfer that would be possible under ideal conditions of soil moisture and vegetation, it usually cannot be measured directly but must be determined experimentally. Like actual evapotranspiration, potential evapotranspiration is clearly a climatic element of great importance. By comparing it with precipitation we can obtain a rational definition of the moisture factor.

Precipitation is a strictly physical process, which meteorologists have investigated in much detail. Evapotranspiration is likewise a physical process, yet since it is subject to biological control, it must be studied by methods unfamiliar to the meteorologist. Information concerning evapotranspiration has come chiefly not from the meteorologist but from the biologist. For this reason it is necessary to make use of the literature and apply the methods of plant physiology. Nevertheless, evapotranspiration represents the return flow of water to the atmosphere and is thus an important meteorological process.

The only method so far developed that measures the actual evapotranspiration from a field or any other natural surface without disturbing the vegetation cover in any way is the so-called "vapor transfer" method. Water vapor when it enters the atmosphere from the ground or from plants is carried upward by the moving air. It is carried upward in small eddies or bodies of air that are replaced by drier eddies from above. Although we cannot see water vapor, we can measure it in the air. We find that when evaporation is taking place the amount of moisture is greatest in the air near the ground and decreases with distance above it. If we determine the rate at which the air near the ground is mixing with that above it and at the same time measure the difference in water-vapor content at the two levels, we can determine both the rate and the amount of evapotranspiration. Furthermore, we can determine equally well the amount of water condensed as dew.[1]

This method is not easy either to understand or to use. It is hard to use because it requires physical measurements more precise than are usually made. Furthermore, the coefficient of turbulent transfer of air is not a constant. It varies from time to time and from place to place. It even varies with height at a given time and place. In spite of these difficulties the method can be perfected and will answer many important questions for climatology and biology.[2]

Scientists have tried in various ways to determine the amount of water used by plants. One of the earliest attempts was to remove leaves or branches from a plant, let them dry for a brief time, and weigh them to see how much water they had lost. Another method is to place plants in sealed containers and measure the moisture that accumulates in the confined air. Experimenters have grown thousands of individual plants in pots, weighing them periodically to determine the evapotranspiration losses. These methods are highly artificial, and generalization from them sometimes gives fantastic

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[1] See the note by John Leighly: New Occasions and New Duties for Climatology, *Geogr. Rev.*, Vol. 29, 1939, pp. 682–683.

[2] C. W. Thornthwaite and Benjamin Holzman: Measurement of Evaporation from Land and Water Surfaces, *U. S. Dept. of Agric. Tech. Bull. No. 817*, 1942; C. W. Thornthwaite: The Measurement of Evaporation and Transpiration from Natural Surfaces, *Proc. Hydrology Conference, State College, Pa., June 30–July 2, 1941* (*Pennsylvania State College School of Engineering Tech. Bull. No. 27*), 1942, pp. 185–197; *idem:* Atmospheric Turbulence and the Measurement of Evaporation, *Proc. Second Hydraulics Conference, June 1–4, 1942* (*Univ. of Iowa Studies in Engineering, Bull. 27*), Iowa City, 1943, pp. 280–288.

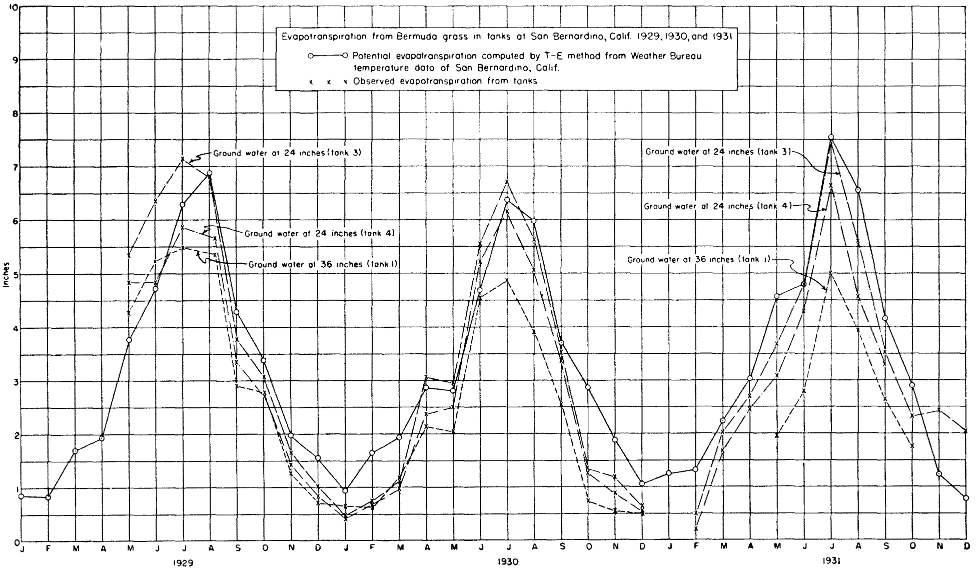

FIG. 1

results. For example, in a German study transpiration from an oak woodland was computed as being more than eight times the actual rainfall.[3]

There are other, less artificial methods of determining both water use and water need. In some irrigated areas rainfall, water applied by irrigation, and water outflow are all measured. The fraction of the water applied that does not run off is the evapotranspiration. Irrigation engineers have determined the evapotranspiration from plants growing in sunken tanks filled to ground level with soil in which water tables are maintained at different predetermined depths beneath the soil surface.[4] Lee found that the annual use of water by salt grass at Independence, Calif., ranged from 13.4 inches with the water table at 5 feet below the ground surface to 48.8 inches with it at 1.5 feet below. Debler reported a range in annual use of water by salt grass at Los Griegos, N. Mex., from 10.1 inches with the water table at 37 inches to 48.4 inches with it at 5 inches. Young and Blaney observed a range at Santa Ana, Calif., from 13.4 inches with free water 4 feet below the surface to 42.8 inches with it 1 foot below. Comparable figures for Bermuda grass at San Bernardino, Calif., were 28.19 inches with the water

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[3] B. G. Ivanov: Evaporation under Natural Conditions: Methods of Study and Results Attained (Hydrometeorological Publications), Moscow, 1939. (Translated from the Russian by Headquarters, Army Air Forces Weather Division.)

[4] A. A. Young and H. F. Blaney: Use of Water by Native Vegetation, *California Dept. of Public Works, Division of Water Resources Bull. No. 50,* 1942.

table at 3 feet and 34.37 inches with it at 2 feet. The observations from San Bernardino are shown in Figure 1.

Lowry and Johnson[5] published data on annual water use in 12 irrigated valleys in the western United States and one in the Dominican Republic. Since water is supplied by irrigation, water use in these valleys approximates potential evapotranspiration. It ranges from 18 inches in a mountain valley in Colorado to 58 inches in the Barahona district in the Dominican Republic.

Although the various methods of determining potential evapotranspiration have many faults and the determinations are scattered and few, we get from them an idea of how much water is transpired and evaporated and how much would be if it were available. We find that the rate of evapotranspiration depends on four things: climate, soil-moisture supply, plant cover, and land management. Of these the first two prove to be by far the most important.

Some scientists have believed that transpiration serves no useful purpose for the plant. We now understand that transpiration effectively prevents the plant surfaces that are exposed to sunlight from being overheated. Most plants require sunlight for growth. The energy of the sun combines water and carbon dioxide in the leaves into food, which is later carried to all parts of the plant and used in growth. This process, which is called "photosynthesis," is most efficient when the leaf temperatures are between 85° and 90° F. But a leaf exposed to direct sunlight would quickly become much hotter if the energy of the sun were not disposed of in some way. The surface of dry ground may reach a temperature of 200° F.; temperatures higher than 160° F. have been measured one-fourth of an inch below the ground surface. The plant is admirably designed to dissipate heat, the leaves being like the fins of a radiator, and some of the excess heat is conducted into the adjacent air and carried away in turbulence bodies. In this way the air is heated. But some of the excess heat energy is utilized in transpiration, to change water from a liquid into a vapor. Most of the heat of evaporation must come from the plant. Thus, the greater the intensity of sunshine, the greater will be the tendency to overheating, and the larger will be the transpiration of a plant exposed to it, if water is available for the process. Transpiration is a heat regulator, preventing temperature excesses in both plant and air. Dew formation at night is the reverse of this process and tends to prevent low temperature extremes, since the heat released goes

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[5] R. L. Lowry, Jr., and A. F. Johnson: Consumptive Use of Water for Agriculture, *Trans. Amer. Soc. of Civil Engineers*, Vol. 107, 1942, pp. 1243–1266 (Paper No. 2158); discussion, pp. 1267–1302.

mainly to the plant. Both transpiration and growth are related to temperature in the same way.

Atmospheric elements whose influence on transpiration has been studied include solar radiation, air temperature, wind, and atmospheric humidity. These factors are all interrelated. Although solar radiation is the basic factor, there seems to be a closer parallelism between air temperature and transpiration. The temperature of the transpiring part is most closely related to the rate of transpiration.

Transpiration and growth are both affected in the same way by variations in soil moisture. Both increase with increase of available water in the root zone of the soil, to an optimum. Above the optimum both are less, presumably because of poor aeration of the soil, which results in a lack of oxygen to supply the roots and an excess of carbon dioxide.[6] On the other hand, as water in the soil increases above the optimum for growth, direct evaporation from the soil surface also continues to increase.

We do not yet know how much we may increase or decrease transpiration by varying the type of plants or by modifying the plant cover. Since transpiration regulates leaf temperature, and since most plants reach their optimum growth at about the same temperature, we probably cannot change it very much except by reducing the density of the plant cover and thus wasting a part of the solar energy. If all the vegetation is removed from a field, there will be no transpiration. But as long as the root zone of the soil is well supplied with water, the amount of water transpired from a completely covered area will depend more on the amount of solar energy received by the surface and the resultant temperature than on the kind of plants.

Since potential evapotranspiration is an important climatic element, we need to know its distribution over the earth and how it varies through the year and from one year to another. Actual determinations are so few that it would be impossible to make a map of any area by means of them. At the present time the only alternative is to discover a relation between potential evapotranspiration and other climatic factors for which there are abundant data.

TEMPERATURE AND GROWTH

Many studies have been made of temperature and growth. Some investigators have measured the elongation of shoots, stems, and roots grow-

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[6] A. J. Loustalot: Influence of Soil-Moisture Conditions on Apparent Photosynthesis and Transpiration of Pecan Leaves, *Journ. of Agric. Research,* Vol. 71, 1945, pp. 519–532; G. W. Schneider and N. F. Childers: Influence of Soil Moisture on Photosynthesis, Respiration and Transpiration of Apple Leaves, *Plant Physiology,* Vol. 16, 1941, pp. 565–583.

ing under various controlled temperatures. Others have determined increase in width of leaves and increase in dry weight of the plant. Rate of development of various insects under controlled temperature has been determined. There is always an optimum temperature for growth, a temperature at which the growth rate is highest. At lower or higher temperatures growth is slower. The temperature at which growth is most rapid seems to vary somewhat with the material under study and with the length of exposure but is always near 30° C. (86° F.). In his investigation of maize seedlings Lehenbauer[7] obtained most rapid growth for a 3-hour period at 29° C. and for 9–12 hours at 32° C. He says that for periods of 3 hours or longer the greatest rates of growth occur within the temperature range 29° to 32°. Wadley[8] found that the rate of development of the green bug was most rapid at 30° C. Similarly, there are minimum and maximum temperatures beyond which growth does not occur. These limits vary likewise, but the minimum is near 0° C. and the maximum somewhere above 40° C.

About half a century ago Van't Hoff[9] propounded the principle that the velocity of a chemical reaction doubles or trebles with each rise in temperature of 10° C. This is an exponential law of the form

$$v = ca^t, \tag{1}$$

in which a has the value 1.0718 when velocity, v, doubles with a 10° rise in temperature, and the value 1.1161 when v trebles. The Van't Hoff law has been applied by biologists to physiological processes. The customary procedure is to determine temperature coefficients from actual growth measurements. The temperature coefficient is the quotient of two growth rates that are separated from each other by a 10° interval of temperature, and by the Van't Hoff rule it is expected to range between 2 and 3. In reality, the coefficient varies much more widely. It exceeds 10.0 in the temperature range between 0° C. and 10° C. and falls steadily to values less than 1.0, above the optimum temperature at which growth and temperature are inversely related.

Lehenbauer[10] published the following table of temperature coefficients for 12-hour growth rates of shoots of maize seedlings, for various 10° ranges of temperature:

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[7] P. A. Lehenbauer: Growth of Maize Seedlings in Relation to Temperature, *Physiological Researches*, Vol. 1, 1914, pp. 247–288.

[8] F. M. Wadley: Development-Temperature Correlation in the Green Bug, Toxoptera Graminum, *Journ. of Agric. Research*, Vol. 53, 1936, pp. 259–266.

[9] J. H. van't Hoff: Études de dynamique chimique, Amsterdam, 1884.

[10] *Op. cit.*, p. 281.

Temperature, °C.	12–22	13–23	15–25	18–28	20–30	21–31	22–32	25–35	32–42	33–43
Growth rate, 0.01 mm.	9–59	10–64	20–75	28–98	45–108	53–109	59–111	75–86	111–11	101–6
Coefficient	9.56	6.40	3.75	3.50	2.40	2.06	1.88	1.15	0.09	0.06

The temperature coefficient must have values of ∞ at the temperature minimum, 1.0 at the temperature optimum, and 0 at the temperature maximum. Only in the temperature range 20° to 30°, where the temperature coefficient is between 2 and 3, is the Van't Hoff law valid. Since the temperature coefficient drops continuously, the coefficient a in equation (1) must also drop. The Van't Hoff $a = 1.0718$ is valid at only one point on the curve and is no more significant than any other value of a. The fact that a drops steadily and falls below 1.00 when the optimum temperature is reached is far more significant than any average value of a (or of the temperature coefficient).

There is clearly some growth-inhibiting factor that is directly proportionate to temperature. It is only because of the accident that the temperatures which plants ordinarily experience under natural conditions are below the optimum, that growth is popularly assumed to vary directly with temperature. At temperatures above 30° C. the growth-inhibiting factor becomes greater than the growth-stimulating factor and the growth rate falls with rising temperature. Since the exponential equation $v = ca^t$ requires a continuously increasing growth rate with rising temperature, it does not truly express the relation between temperature and growth.

What the growth-inhibiting factor is, is uncertain. It may reasonably be the operation of water deficiency in the plant tissue. The amount of water in a plant is a balance between that absorbed by the roots and that transpired from the leaves. As the transpiration rate rises with increasing temperature, it will presently exceed the rate of water absorption. Then a suction pressure, or tension, develops in the plant, acting as a brake on transpiration and on growth. The moisture balance in the plant is upset, transpiration is reduced, and growth is retarded. A more satisfactory equation of growth, therefore, is the following:

$$v = a \frac{bce^{ct}}{(e^{ct} + b)^2}, \tag{2}$$

in which a, b, and c are constants and e is the base of the Napierian system of logarithms. In this equation the numerator represents the growth-stimulating factor and the denominator the growth-inhibiting factor. The optimum temperature for growth is where numerator and denominator are equal.

When we fit the equation to a series of Lehenbauer's observations on the growth of maize seedlings, we have the following coefficients: $a = 1764.9$,

$b = 1118.8$, $c = .24$. The equation gives v as a percentage of the optimum growth rate. The Lehenbauer data and the growth curve derived from this equation are plotted in Figure 2. Some of the computed values of v are as follows:

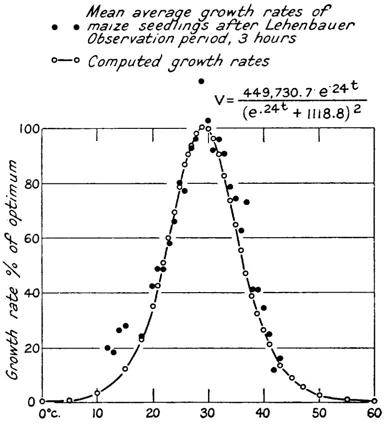


FIG. 2

°C.	v.
0	0.36
5	1.19
10	3.88
15	12.33
20	35.52
25	78.30
28	98.27
30	99.69
32	90.30
35	64.56
40	26.32
45	8.76
50	2.72
55	0.83
60	0.25

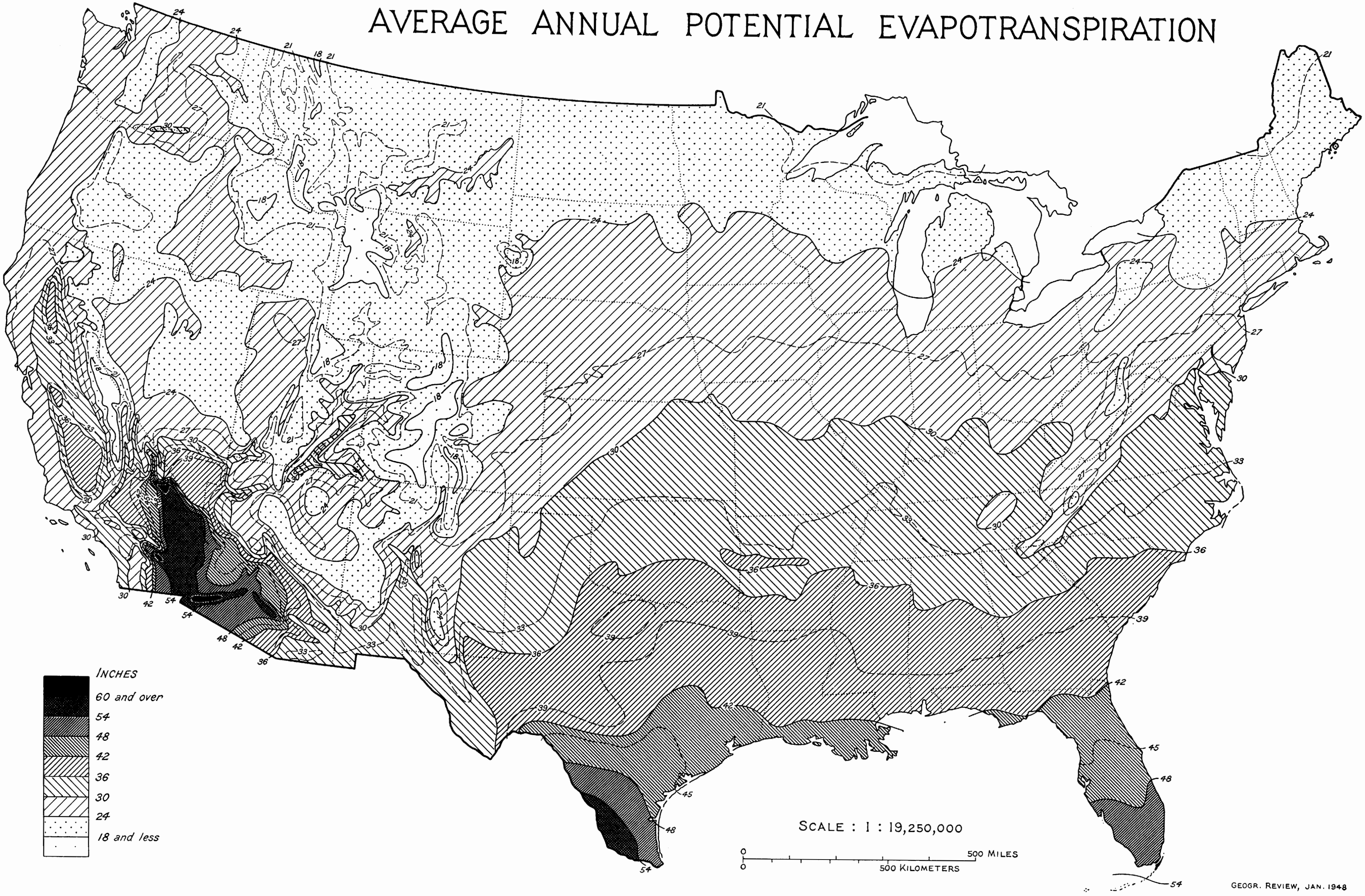
DETERMINATION OF POTENTIAL EVAPOTRANSPIRATION

Unfortunately, there are no comparable controlled experiments on the relation of transpiration to temperature. The most reliable measurements of transpiration and evaporation together are for long periods of time. When only monthly or annual totals of evapotranspiration are available, the temperature relation cannot be determined precisely.

The determinations have shown that potential evapotranspiration is high in the southern part of the United States and low in the northern part and that it varies greatly from winter to summer. From observations like those of Figure 1, it has been found that when adjustments are made for variation in day length, there is a close relation between mean monthly temperature and potential evapotranspiration. Study of all available data has resulted in a formula that permits the computation of potential evapotranspiration of a place if its latitude is known and if temperature records are available. The formula is given, and its use described, in the appendix.

Figure 3 shows the distribution of average annual potential evapotranspiration in the United States. It is based on normals of some 3500 Weather Bureau stations, revised to 1930. 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. It is less than 21 inches along the Canadian border of the eastern United States and more than 48 inches in Florida and southern

AVERAGE ANNUAL POTENTIAL EVAPOTRANSPIRATION



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Texas. Although potential evapotranspiration and precipitation are independent climatic elements, in arid regions potential evapotranspiration is increased because of the higher daytime temperatures due to the absence of clouds and rain and because of the small actual evapotranspiration. The high values in the Colorado and Gila Deserts and in the lower Rio Grande Valley are examples. In the arid section of the Columbia River Valley between Washington and Oregon the potential evapotranspiration is more than 30 inches, whereas it is only about 21 inches in the same latitude in the eastern United States.

The march of potential evapotranspiration through the year follows a uniform pattern in most of the country. 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. In most places the precipitation is less than the need during a part of the year. In times of excess rainfall water is stored in the soil. The part of this water that is within reach of roots is used before the plants begin to suffer; therefore drought does not begin immediately when rainfall drops below water need. The amount of water in the root zone available to plants varies with the soil structure and the distribution of roots. It is, accordingly, not a constant. However, except in areas of shallow soil the water storage capacity available to mature plants with fully developed root systems varies around a mean that is the equivalent of about 10 centimeters or 4 inches of rainfall. Curves comparing water need and precipitation at selected stations in the United States are given in Figure 4.

In Brevard, N. C. (Fig. 4-A), less than half an inch of water is needed in each of the winter months. The need rises rapidly during the spring, reaches a high point of more than 5 inches in July, and falls rapidly during the autumn. The total average annual water need is 28.50 inches. The precipitation ranges from less than 3½ inches in November to more than 6½ inches in July and is greater than the need in every month. The total average annual precipitation is 61.06 inches, more than twice the potential evapotranspiration. The surplus of 32.56 inches represents runoff. In Salisbury, N. Y. (Fig. 4-B), the average annual potential evapotranspiration is 21.81 inches and the precipitation is 48.39 inches, more than twice as much. In no month is the precipitation less than the need. Here the surplus that runs off amounts to 26.58 inches.

In Bar Harbor, Maine (Fig. 4-C), the precipitation falls to a minimum of less than 3½

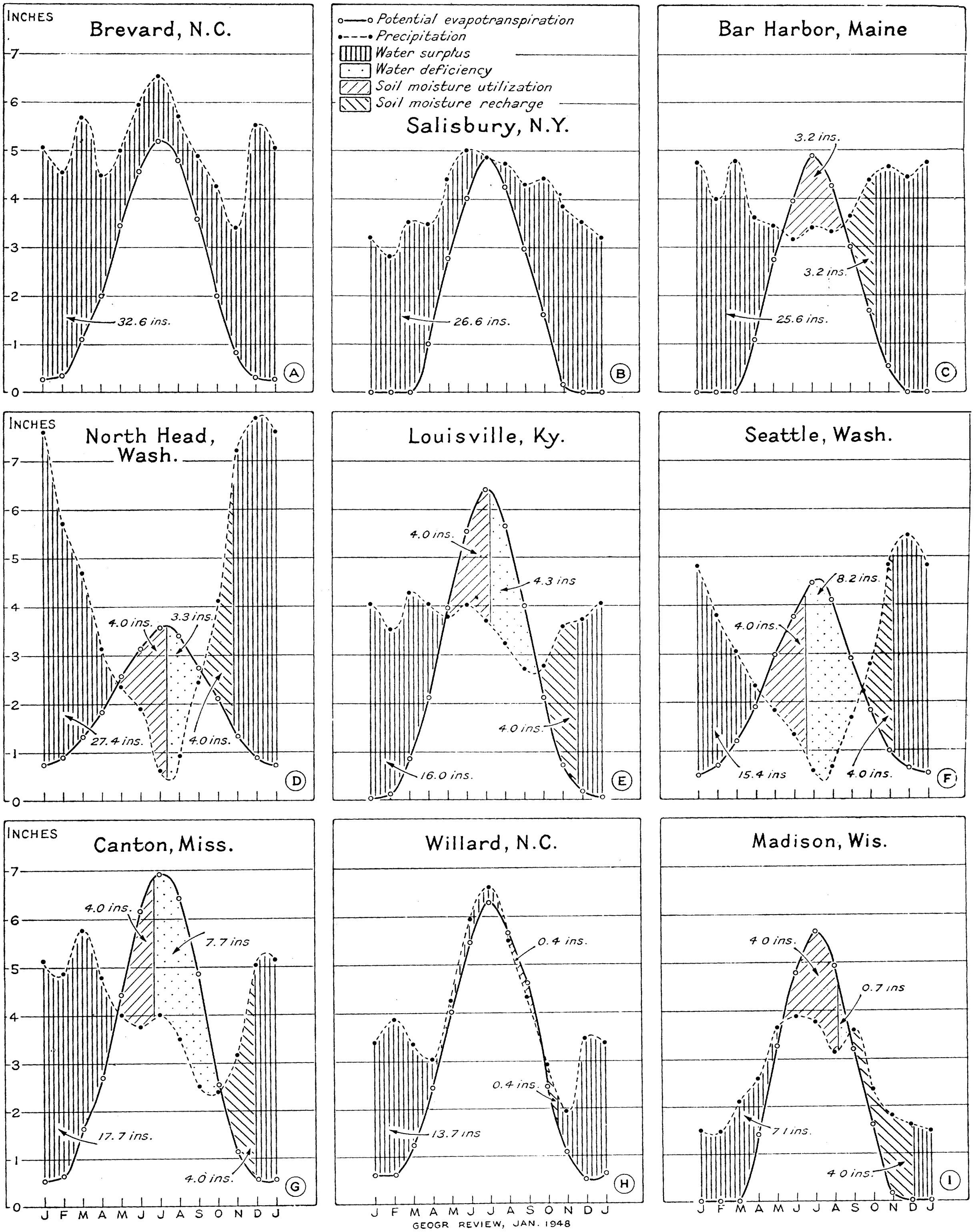

FIG. 4—March of precipitation and potential evapotranspiration at selected stations in the United States.

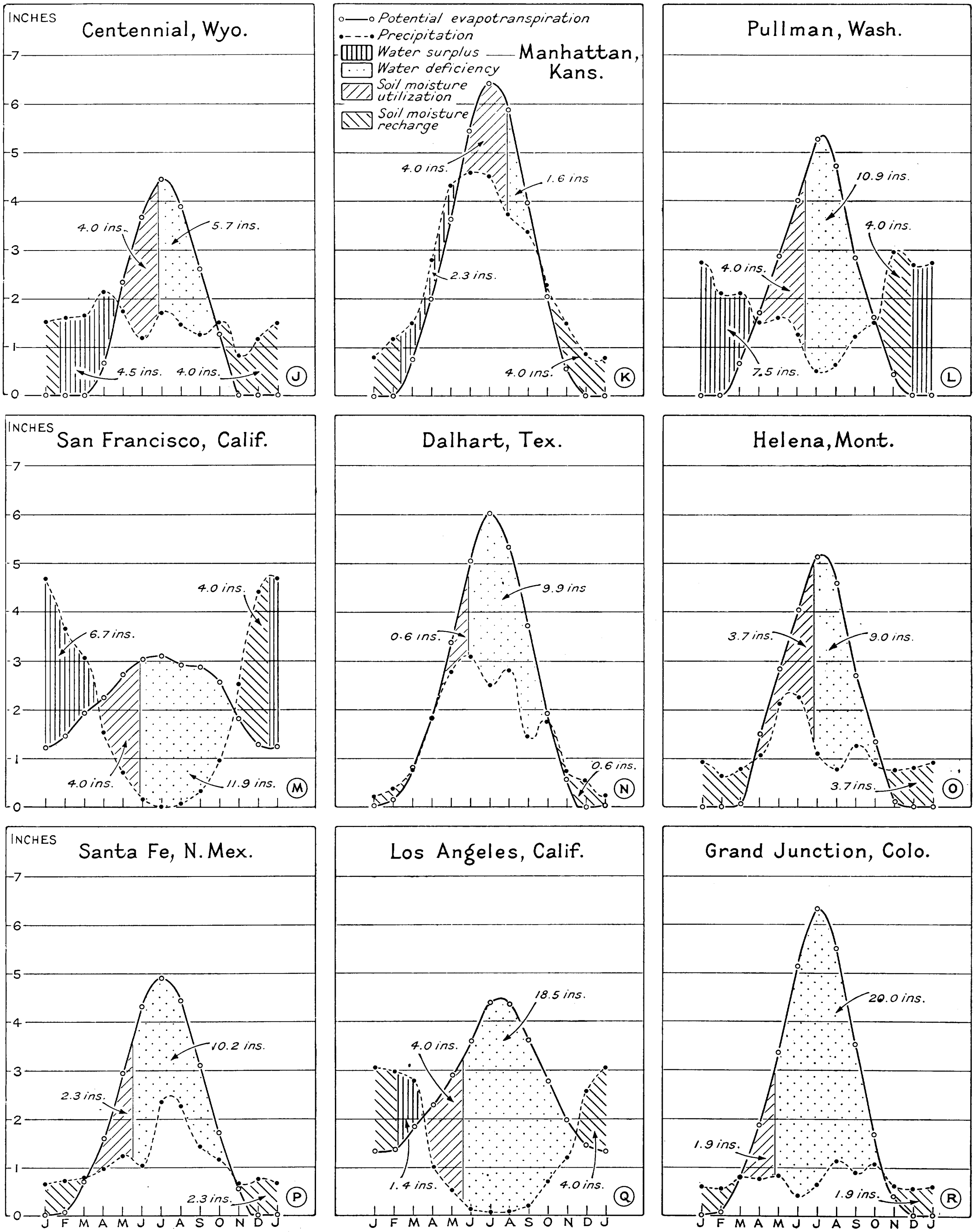

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FIG. 4 (cont'd.)

inches in each of the 3 summer months, at exactly the time when potential evapotranspiration rises above 4 inches. Current rainfall is less than need, but the difference is made up from water in storage in the soil. In North Head, Wash. (Fig. 4-D), the precipitation exhibits marked seasonal variation: it is more than 7 inches a month in November, December, and January and less than 1 inch in July and August. Although water need is not large in any month, the July maximum being only 3.6 inches, it is in excess of rainfall during 5 months, from May through September. There is too much rain in winter and too little in summer. In late September, as the days grow cooler and shorter, water need falls below precipitation. For a while the rainfall not immediately needed goes to replace stored soil moisture that has been used up. From then on, the surplus water is lost so far as plant growth is concerned. It raises ground-water levels and produces surface and subsurface runoff. But it is of no benefit to plants and adds nothing to atmospheric moisture. In spring, as the days lengthen and become warmer, plant growth accelerates, and both transpiration and evaporation increase rapidly. By midspring water need exceeds precipitation. Thereafter, until midsummer, the excess demands for water are satisfied from the soil-moisture reserves. When these reserves are exhausted, the vegetation must rely solely on current rainfall. And current rainfall is not enough. Water use falls below water need; the plants suffer, and growth is retarded.

In Louisville, Ky. (Fig. 4-E), rainfall is fairly uniform through the year—between 3.5 and 4 inches a month except in autumn, when it falls below 3 inches. Since potential evapotranspiration is not uniform, however, there is a surplus of water in winter and a deficiency in summer, just as in North Head (Fig. 4-D) and Seattle (Fig. 4-F).

Although annual precipitation in Canton, Miss. (Fig. 4-G), is nearly 50 inches and summer rainfall averages 3.75 inches, the summer potential evapotranspiration is so large that there is a water deficiency of nearly 8.0 inches, almost the same as in Seattle.

In some places monthly precipitation and water need are nearly equal. For example, in Manhattan, Kans. (Fig. 4-K), winter rainfall is only a little greater than potential evapotranspiration and summer rainfall is only a little less. Water surplus and water deficiency are both small. In Madison, Wis. (Fig. 4-I), there is a maximum of rainfall in summer, but even so rainfall does not equal need; there is a small water deficiency in summer and a considerable water surplus in winter and spring. In Willard, N. C. (Fig. 4-H), potential evapotranspiration and precipitation are nearly the same in all the growing-season months. Rainfall is much greater than need in winter, however, and there is a water surplus of nearly 14 inches. In Dalhart, Tex. (Fig. 4-N), on the other hand, rainfall is equal to water need in winter but falls far short in summer, and there is a water deficiency of 10 inches.

In Grand Junction, Colo. (Fig. 4-R), rainfall is slight throughout the year. Water supply satisfies less than one-third of water need, and deficiency is very large. The climate is arid.

The curves of San Francisco, Calif. (Fig. 4-M), and Centennial, Wyo. (Fig. 4-J), show that in both places precipitation fails about equally to provide the water needed in June, July, and August. Water deficiencies, in inches, for June, July, and August are as follows: San Francisco 2.87, 3.07, 2.87; Centennial 2.48, 2.75, 2.44. But there is a vast difference in the relative aridity of the two places. Whereas July rainfall in Centennial provides about 1.5 inches of a needed 4.5 inches, in San Francisco there is no rain at all in July. In one place the water deficiency of July amounts to 62 per cent of the need, in the other to 100 per cent. The aridity of a place in a given period of time depends not on the numerical amount of the water deficiency but rather on the relation of this deficiency to the water need.

TABLE I—COMPARATIVE MOISTURE DATA OF SEATTLE, WASH., AND MANHATTAN, KANS.

In centimeters

ITEM	J	F	M	A	M	J	J	A	S	O	N	D	Y
						Seattle, Wash.							
Potential evap.	1.3	1.8	3.1	4.9	7.6	9.6	11.4	10.5	7.4	4.7	2.5	1.6	66.4
Precipitation	12.3	9.7	7.8	6.0	4.7	3.4	1.5	1.7	4.3	7.1	12.3	13.9	84.7
Storage change	0	0	0	0	−2.9	−6.2	−0.9	0	0	2.4	7.6	0	
Storage	10.0	10.0	10.0	10.0	7.1	0.9	0	0	0	2.4	10.0	10.0	
Actual evap.	1.3	1.8	3.1	4.9	7.6	9.6	2.4	1.7	4.3	4.7	2.5	1.6	45.5
Water deficiency	0	0	0	0	0	0	9.0	8.8	3.1	0	0	0	20.9
Water surplus	11.0	7.9	4.7	1.1	0	0	0	0	0	0	2.2	12.3	39.2
Runoff *	8.9	8.4	6.5	3.8	1.9	1.0	0.5	0.2	0.1	0.1	1.1	6.7	39.2
Moisture ratio	8.47	4.38	1.52	0.22	−.38	−.65	−.87	−.84	−.42	0.54	3.92	7.68	
						Manhattan, Kans.							
Potential evap.	0.0	0.0	1.9	5.1	9.3	13.9	16.4	15.0	10.1	5.2	1.4	0.0	78.3
Precipitation	2.0	3.0	3.8	7.1	11.1	11.7	11.5	9.5	8.6	5.8	3.8	2.2	80.1
Storage change	2.0	2.8	0	0	0	−2.2	−4.9	−2.9	0	0.6	2.4	2.2	
Storage	7.2	10.0	10.0	10.0	10.0	7.8	2.9	0	0	0.6	3.0	5.2	
Actual evap.	0.0	0.0	1.9	5.1	9.3	13.9	16.4	12.4	8.6	5.2	1.4	0.0	74.2
Water deficiency	0.0	0.0	0.0	0.0	0.0	0.0	0.0	2.6	1.5	0.0	0.0	0.0	4.1
Water surplus	0.0	0.2	1.9	2.0	1.7	0.0	0.0	0.0	0.0	0.0	0.0	0.0	5.8
Runoff *	0.0	0.1	1.0	1.5	1.6	0.8	0.4	0.2	0.1	0.1	0.0	0.0	5.8
Moisture ratio	∞	∞	1.00	0.39	0.19	−.16	−.30	−.37	−.15	0.11	1.71	∞	

*Assuming that 50 per cent of the water available for runoff in any month is held over until the following month. In watersheds of less than 100 square miles the percentage is probably smaller.

A moisture ratio that expresses the relative humidity or aridity of a month may be obtained by dividing the difference between precipitation and potential evapotranspiration by potential evapotranspiration,

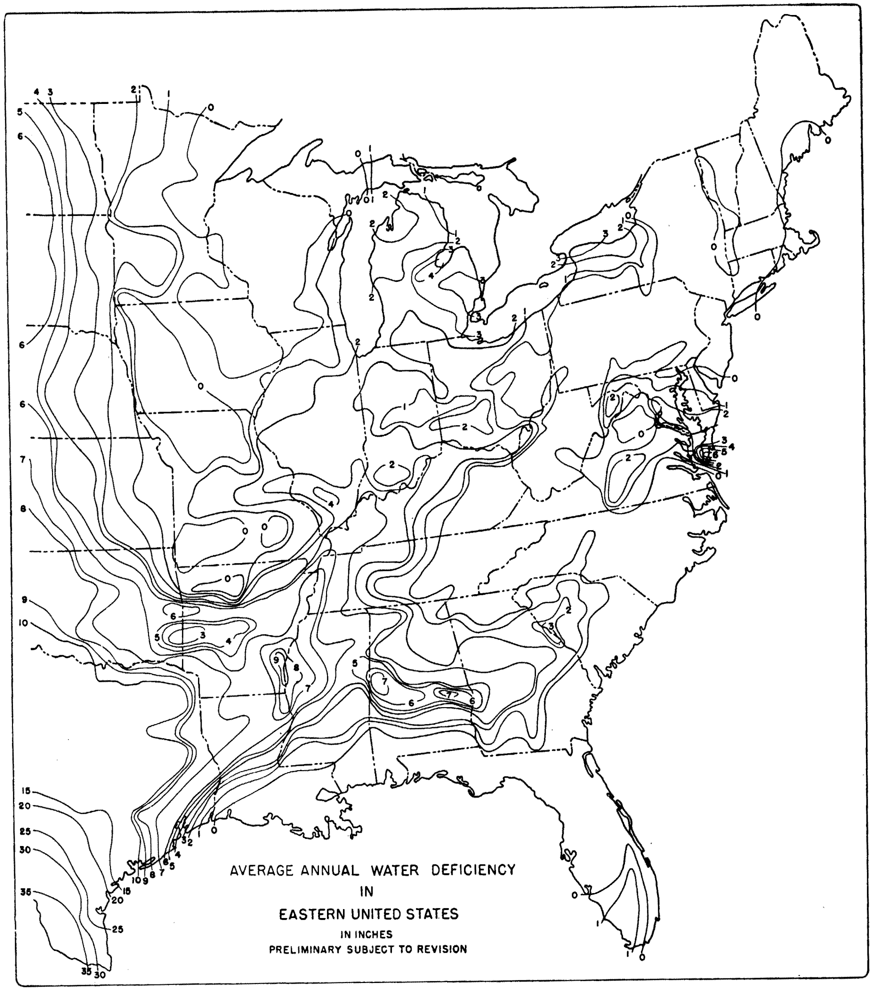
$$\frac{p - e}{e} \text{ or } \frac{p}{e} - 1. \tag{3}$$

Positive values of the ratio mean that the precipitation is excessive, negative values that it is deficient. A ratio of zero means that water supply is equal to water need.

WATER SURPLUS AND WATER DEFICIENCY

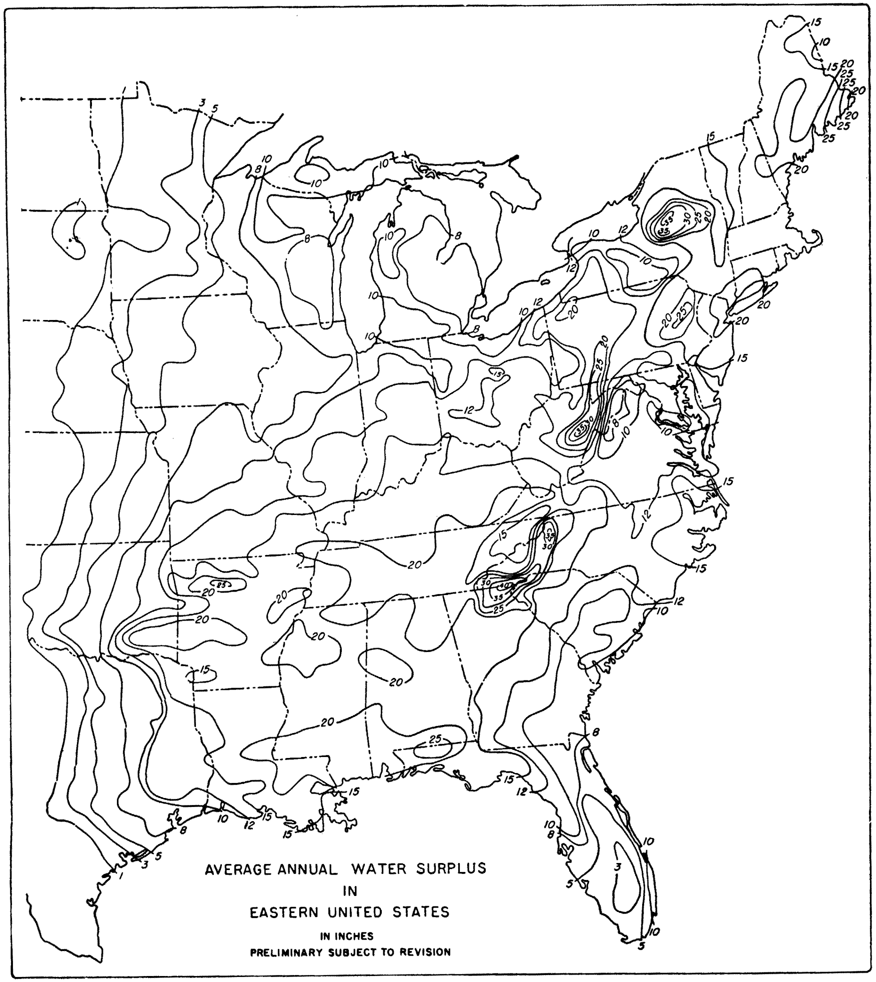
Numerical values of water surplus and water deficiency may be obtained as a simple bookkeeping procedure, precipitation being treated as income and potential evapotranspiration as outgo, and soil moisture as a reserve that may be drawn upon as long as it lasts. Sample computations for Seattle, Wash., and Manhattan, Kans., are given in Table I.

The distribution of average annual water deficiency in the eastern United States is shown in Figure 5. Deficiencies rise sharply to the west in the Great Plains States and are in excess of 10 inches in most of Texas. Deficiencies



AVERAGE ANNUAL WATER DEFICIENCY
IN
EASTERN UNITED STATES
IN INCHES
PRELIMINARY SUBJECT TO REVISION

FIG. 5



AVERAGE ANNUAL WATER SURPLUS
IN
EASTERN UNITED STATES
IN INCHES
PRELIMINARY SUBJECT TO REVISION

FIG. 6

Table II—Comparative Data on Irrigated Valleys in the Western United States

VALLEY	WEATHER STATION	OBSERVED WATER USE *cm.*	POTENTIAL EVAPOTRANS. *cm.*	DIFF. *cm.*	DIFF. %
New Fork	Pinedale, Wyo.	46.6	45.4	−1.2	2.6
Michigan and Illinois	Walden, Colo.	45.7	46.9	1.2	2.6
Southwest Area, San Luis	San Luis, Colo.	53.0	51.8	−1.2	2.4
West Tule Lake	Tule Lake, Calif.	67.7	57.9	−9.8	14.5
Garland Div., Shoshone	Lovell, Wyo.	61.6	59.3	−2.3	3.7
North Platte	Torrington, Wyo.	60.4	60.7	0.3	0.5
Mason Creek and Boise	Caldwell, Idaho	66.1	66.5	0.4	0.6
Uncompahgre	Delta, Colo.	68.3	67.5	−0.8	1.2
Mesilla	Agr. College, N. M.	83.2	84.0	0.8	0.9
Greenfields Div., Sun River	Great Falls, Mont.	60.3	60.2	−0.1	0.2
Pecos River near Carlsbad	Carlsbad, N. M.	89.6	91.7	2.1	2.3
Lower Rio Grande	Del Rio, Tex.	115.8	117.0	1.2	1.0
Barahona, Dominican Repub.	Barahona	147.3	146.3	−1.0	0.7

are surprisingly large in various areas of high precipitation in the Southern States—more than 6 inches in the Alabama Black Belt and in the lower Mississippi Valley.

The distribution of average annual water surplus in the eastern United States is shown in Figure 6. It is more than 10 inches in most of the East and Southeast but falls rapidly westward and is less than 1 inch in the Great Plains States. In various centers in the Appalachians the water surplus is more than 35 inches.

Reliability of Computed Potential Evapotranspiration

We can test the accuracy of the computed values of potential evapotranspiration in a number of ways. In Table II above the observations of annual water use in irrigated valleys published by Lowry and Johnson[11] are compared with computations of annual potential evapotranspiration. Although the computations refer to points of weather observation and the observations to irrigation projects differing in area, in only one place, West Tule Lake, Calif., is there a significant difference between computed and observed potential evapotranspiration.

There are no comparable direct observations of potential evapotranspiration in the eastern United States. There are, however, measurements of runoff from a great many streams for many years, which may be compared with computed values of water surplus. Observed mean annual runoff in 124 minor watersheds in the Tennessee Valley compared with water surplus

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[11] *Op. cit.*, pp. 1253 and 1301.

computed from data of temperature and precipitation at Weather Bureau stations[12] shows a close correspondence, especially noteworthy because the data of the Tennessee Valley Authority refer to areas ranging in size from a few square miles to nearly two thousand and are for a uniform period from 1920 to 1942, whereas the computations were made from averages of temperature and precipitation for periods of various lengths ending in 1930 and refer to the points of weather observation. Computed water surplus and measured runoff are equally close together in Ontario and Mexico, where potential evapotranspiration is respectively much smaller and much larger. The National Resources Board in its report of December 1, 1934, gives a map of average annual runoff in the United States, which may be compared with the map of water surplus, Figure 6. The correspondence is generally close, though there are discrepancies, some of which are due to admitted errors in the original runoff map.

From these and other tests it appears that the computed values of potential evapotranspiration are of the right order of magnitude throughout most of the United States. They are, nevertheless, only approximate. More exact determinations must await further study and the development of a more rational equation. Whether or not the formula can be used without modification to determine potential evapotranspiration in equatorial and polar regions is uncertain. This question also requires further study.

ESSENTIALS OF A CLIMATIC CLASSIFICATION

When the various climatic elements, such as temperature and precipitation, are plotted on maps, the values grade regularly except where gradients steepen on mountain slopes or along seacoasts. Nothing distinguishes any one value in the series from the others; hence divisions of the scales are arbitrary and frequently owe their origin to arithmetical or cartographical convenience. Divisions appearing on maps, selected perhaps to conform to a particular color scheme, gradually assume the role of climatic regions; for example, the 10-inch isohyet as the limit of agriculture.

Köppen made a significant advance in climatic classification when he undertook, first to identify the climatic regions and locate their boundaries from studies of the distribution of vegetation, and then to select numerical climatic values for the boundaries. Penck, De Martonne, and Thornthwaite employed the same method but used data of hydrology and soils to supple-

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[12] See Figure 3 (p. 691) in H. G. Wilm, C. W. Thornthwaite, and others: Report of the Committee on Transpiration and Evaporation, 1943–44, *Amer. Geophys. Union Trans. of 1944*, Part 5, Washington, 1945, pp. 683–693.

ment those of vegetation in locating the boundaries of climatic regions. Although this method represents a step forward in that the climatic boundaries are not arbitrary, it is still empirical.

In order to achieve a rational quantitative classification of climate, definite and distinctive break points must be discovered in the climatic series themselves. No such break points exist in the data either of precipitation or of potential evapotranspiration. Both run in continuous series from very low values to very large ones. But when they are taken together, there are some distinctive points, and we have the beginnings of a rational classification.

The primary climatic factors relate to moisture and heat. We are interested in whether a climate is moist or dry and warm or cold. We also need to know whether there is seasonal variation—whether the climate is moist in one season and dry in another.

The Moisture Factor

We cannot tell whether a climate is moist or dry by knowing precipitation alone; we must know whether precipitation is greater or less than potential evapotranspiration. Ignorance of the magnitude of water need has led to the development of a number of moisture indices, which attempt to evaluate the effectiveness of precipitation. It is recognized that water utilization or water loss, that is, evaporation and transpiration, must be related to precipitation somehow. Accordingly, a few of the indices make use of computed or measured values of evaporation usually from free water surfaces, lakes and reservoirs, or from evaporation pans.

In 1905 Transeau[13] produced an index of precipitation effectiveness, by using the quotient of total annual measured precipitation and annual evaporation as computed by Russell[14] for a year for about 150 stations, and made a map of the precipitation-evaporation ratio in the eastern United States. In 1933 Isozaki[15] computed ratios for 99 stations in Japan, using average annual precipitation and annual evaporation from pans, and made a moisture map of the Japanese Empire. Trumble,[16] making use of measurements of evaporation from pans in South Australia, derived an empirical equation relating

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[13] E. N. Transeau: Forest Centers of Eastern America, *Amer. Naturalist,* Vol. 39, 1905, pp. 875–889.

[14] Thomas Russell: Depth of Evaporation in the United States, *Monthly Weather Rev.,* Vol. 16, 1888, pp. 235–239.

[15] Masaru Isozaki: Thornthwaite's New Classification of Climate and Its Application to the Climate of Japan, *Journ. of Geogr.,* Tokyo Geographical Society, Vol. 45, 1933, pp. 234–245 (in Japanese).

[16] H. C. Trumble: The Climatic Control of Agriculture in South Australia, *Trans. and Proc. Royal Soc. of South Australia,* Vol. 61, 1937, pp. 41–62; *idem:* Climatic Factors in Relation to the Agricultural Regions of Southern Australia, *Trans. Royal Soc. of South Australia,* Vol. 63, 1939, pp. 36–43.

monthly evaporation to vapor-pressure deficit, which he computed from temperature values. He determined empirically that water need of a month exceeds water supply when computed evaporation is more than three times precipitation. His map of southern Australia shows the number of months of the year when precipitation is less than water need. Wilson[17] produced a moisture index by dividing evaporation from a Livingston porous cup atmometer by annual precipitation and made a moisture map of Ohio.

Numerous investigators, recognizing that temperature is the major control of evaporation, have substituted temperature for evaporation in their moisture indices. The Lang rain factor, $I = P/T$, indicates that the effectiveness varies directly with precipitation and inversely with temperature. De Martonne's index of aridity, $I = P/(T + 10)$, is a slight refinement of Lang's. Köppen's three formulae for delimiting the dry climates, $I = 8P/(5T + 120)$, $I = 2P/(T + 33)$, and $I = P/(T + 7)$, presented in 1918, 1923, and 1928 respectively, are similar to those of Lang and De Martonne. All use annual values of precipitation and temperature given in the metric system.

In 1931 Thornthwaite[18] utilized evaporation data from the continental United States and derived an empirical equation by which the precipitation-evaporation ratio could be determined from monthly values of precipitation and temperature. He used this moisture index in making a climatic map of North America and a later one of the earth.[19]

In 1936 Ångström[20] suggested a modification of De Martonne's index of aridity. He found that the index of aridity was proportionate to duration of precipitation, which was in turn directly proportionate to amount of precipitation and inversely proportionate to an exponential function of temperature. His humidity coefficient is $I = P/(1.07^t)$, in which the denominator of the fraction doubles with each rise of 10° C. in temperature, in accordance with Van't Hoff's law. Ångström published maps of northwestern Europe showing the humidity coefficient in January and July. In 1939 Church and Gueffroy[21] used the Ångström formula to make similar maps of the United States. In 1946 Setzer[22] utilized Van't Hoff's law in the

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[17] J. D. Wilson and J. R. Savage: An Evaporation Survey of Ohio, *Ohio Agric. Exper. Sta. Bull. 564,* 1936.

[18] C. W. Thornthwaite: The Climates of North America According to a New Classification, *Geogr. Rev.,* Vol. 21, 1931, pp. 633–655.

[19] *Idem:* The Climates of the Earth, *Geogr. Rev.,* Vol. 23, 1933, pp. 433–440.

[20] Anders Ångström: A Coefficient of Humidity of General Applicability, *Geografiska Annaler,* Vol. 18, 1936, pp. 245–254.

[21] P. E. Church and E. M. Gueffroy: A New Coefficient of Humidity and Its Application to the United States, *Geogr. Rev.,* Vol. 29, 1939, pp. 665–667.

[22] José Setzer: A New Formula for Precipitation Effectiveness, *Geogr. Rev.,* Vol. 36, 1946, pp. 247–263.

development of a moisture index. Although developed independently, Setzer's index is identical with Ångström's. Setzer published a map of the moisture index in the state of São Paulo, Brazil.

Ångström[23] has ably explained what the investigator wishes to accomplish with moisture indices:

> Especially with geographical, geological, or biological problems in view, [one has] tried to express the humidity or aridity of a climate through some kind of coefficient which is expected to be a measure on conditions of humidity not in the simple physical sense, but involving an indication of the climatic tendency to a surplus or deficiency of water with consequences upon the structure of the ground, the existence of rivers, floods, lakes or deserts etc. It is evident that the humidity of the soil is hereby a prime factor, and as this element is more seldom directly measured it seems justified to try to replace it by some kind of coefficient deduced from the current meteorological data.

A MOISTURE INDEX

It is now apparent that the actual evaporation and transpiration from the soil is not what must be compared with precipitation in order to obtain a moisture index, but, rather, the potential evapotranspiration. 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 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. Expressed as percentages these two indices are:

$$I_h = \frac{100s}{n} \quad \text{and} \quad I_a = \frac{100d}{n}, \tag{4}$$

where I_h and I_a are indices of humidity and aridity respectively, s is water surplus, d is water deficiency, and n is water need. The ultimate in the scale of aridity is where there is no precipitation and the water deficiency is consequently equal to the water need, making I_a equal to 100 per cent. As precipitation and potential evapotranspiration are independent of each other, the index of humidity does not reach a limit where water surplus equals water need, that is, where precipitation is twice potential evapotranspiration, but continues above 100 per cent.

Since water surplus and water deficiency occur at different seasons in

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[23] *Op. cit.*, p. 245.

most places, both must enter into a moisture index, the one affecting it positively, the other negatively. Although a water surplus in one season cannot prevent a deficiency in another except as moisture may be stored in the soil, to a certain extent one may compensate for the other. Water surplus means seasonal additions to subsoil moisture and ground water. Deeply rooted perennials may make partial use of subsoil moisture and thus minimize the effect of drought. Transpiration proceeds, but at reduced rates. For this reason, a surplus of only 6 inches in one season will counteract a deficiency of 10 inches in another. Thus in an over-all moisture index the humidity index has more weight than the aridity index: the latter has only six-tenths the value of the former. The moisture index is:

$$I_m = I_h - .6I_a \text{ or } I_m = \frac{100s - 60d}{n}. \tag{5}$$

Moist climates have positive values of I_m; dry climates have negative values. Figure 7 shows how climatic types are separated in terms of the moisture index, I_m, and makes clear how they are related to water surplus and water deficiency. The various climatic types together with their limits are as follows:

Climatic Type		Moisture Index
A	Perhumid	100 and above
B4	Humid	80 to 100
B3	Humid	60 to 80
B2	Humid	40 to 60
B1	Humid	20 to 40
C2	Moist subhumid	0 to 20
C1	Dry subhumid	–20 to 0
D	Semiarid	–40 to –20
E	Arid	–60 to –40

The index values –60, 0, and 100 are entirely rational limits of moisture regions. That the others are also may be seen in the nomograms of Figure 7. These definitions of moisture regions appear to be in final form. Further work is needed to improve the means for determining potential evapotranspiration, moisture surplus, and moisture deficiency. This may lead to revision of the location of the moisture regions but will not change the definition of them.

These climatic types are the same, and have the same meaning, as those proposed in an earlier climatic classification.[24] However, whereas the limits in the previous classification were determined empirically by study of vegetation, soils, drainage patterns, and so on, these limits are rational and

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[24] Thornthwaite, The Climates of North America (*op. cit.*).

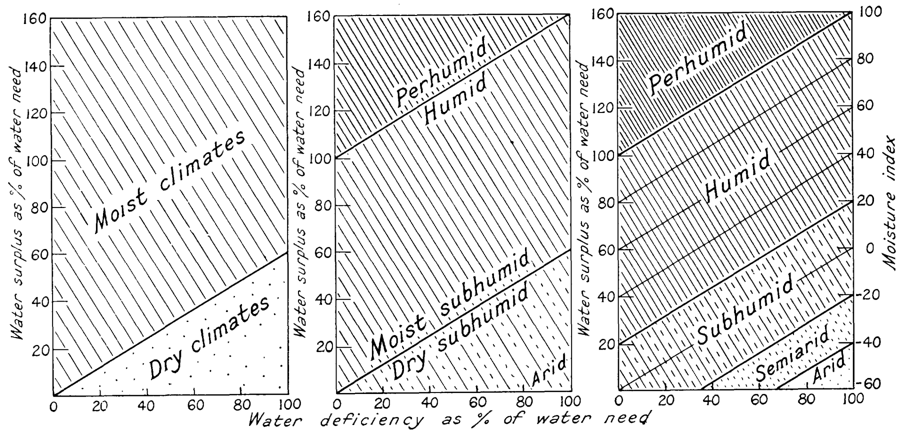

FIG. 7—Delineation of moisture regions on basis of water surplus and water deficiency.

are established solely in terms of the relation between potential evapotranspiration and precipitation. It is possible to convert the present moisture index, I, to the old P–E index by means of the following equation:

$$P\text{–}E = .8I + 48. \qquad (6)$$

The distribution of the moisture regions of the United States is shown in Plate I A. 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, since it separates regions of prevailing water surplus from those of water deficiency. Although dry climates predominate in the 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 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 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

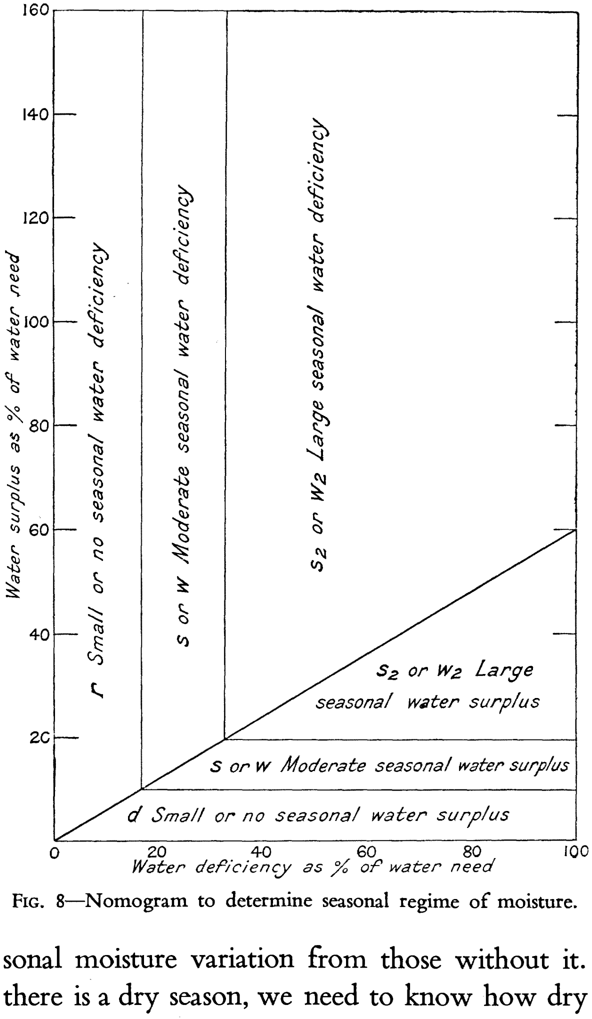

FIG. 8—Nomogram to determine seasonal regime of moisture.

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. Most of the San Joaquin Valley of California is arid.

SEASONAL VARIATION OF EFFECTIVE MOISTURE

It is important to know whether a place is continuously wet or continuously dry, or whether it is wet in one season and dry in another. The moisture index can indicate how humid or how arid a climate is, but it cannot at the same time distinguish climates with seasonal moisture variation from those without it. In the moist climates, if there is a dry season, we need to know how dry it is; in the dry climates, how wet the wet season is if one occurs. In moist climates water deficiency may be large, moderate, small, or nonexistent. In dry climates the same is true of water surplus (see Fig. 8).

By definition, a large water deficiency or surplus is an amount sufficient to make the climate one grade drier or moister than it would be otherwise. For example, in San Francisco the water surplus of 6.68 inches (24.6 per cent of the water need) makes the climate less dry, changing it from semiarid (D) to dry subhumid (C_1). Los Angeles, with a somewhat larger water deficiency and a smaller winter surplus, remains semiarid (D). A moderate water surplus or deficiency will, by definition, change the climate one-half grade.

The symbols s and s_2 indicate respectively moderate and large seasonal variation of moisture, with the drier season occurring in summer; w and w_2 are used similarly where the drier season occurs in winter. Where water need and precipitation march through the year approximately parallel to each other, there will be little or no seasonal variation of moisture. Here the symbols r and d are used, the first to designate the areas with little or no water deficiency in moist climates, and the second little or no water surplus in dry climates. The climatic subdivisions are defined in terms of the humidity and aridity indices as follows:

Moist Climates (A, B, C2)		Aridity Index
r	little or no water deficiency	0 – 16.7
s	moderate summer water deficiency	16.7 – 33.3
w	moderate winter water deficiency	16.7 – 33.3
s2	large summer water deficiency	33.3+
w2	large winter water deficiency	33.3+

Dry Climates (C1, D, E)		Humidity Index
d	little or no water surplus	0 – 10
s	moderate winter water surplus	10 – 20
w	moderate summer water surplus	10 – 20
s2	large winter water surplus	20+
w2	large summer water surplus	20+

The symbols s, s_2, w, and w_2 have the same meaning in both moist and dry climates in spite of the fact that they are defined differently. They refer to the season when rainfall is most deficient.

The winter dry, w and w_2, types do not occur in the United States. In fact, these types are much less widely distributed over the earth than would be deduced from a study of the seasonal march of precipitation. Water need is naturally larger in summer than in winter; and in many areas of maximum summer rainfall, water need and precipitation are essentially together through the year, and there is neither summer water surplus nor winter water deficiency. For example, in Nanking, China, the average precipitation of the three summer months is 55.0 centimeters and of the winter months only 12.0; but since water need varies even more widely between winter and summer, the climate is not of the w type. The water surplus occurs in late winter and spring, not in summer.

In Figure 9 representative stations are grouped according to whether the seasonal range of effective moisture is small, moderate, or large. Those in which precipitation most nearly parallels potential evapotranspiration and both water surplus and water deficiency are small are Ames, Iowa (C_2r), Alexandria, Minn. (C_2r), and Grafton, N. Dak. (C_1d).

The distribution in the United States of the climatic subtypes based on

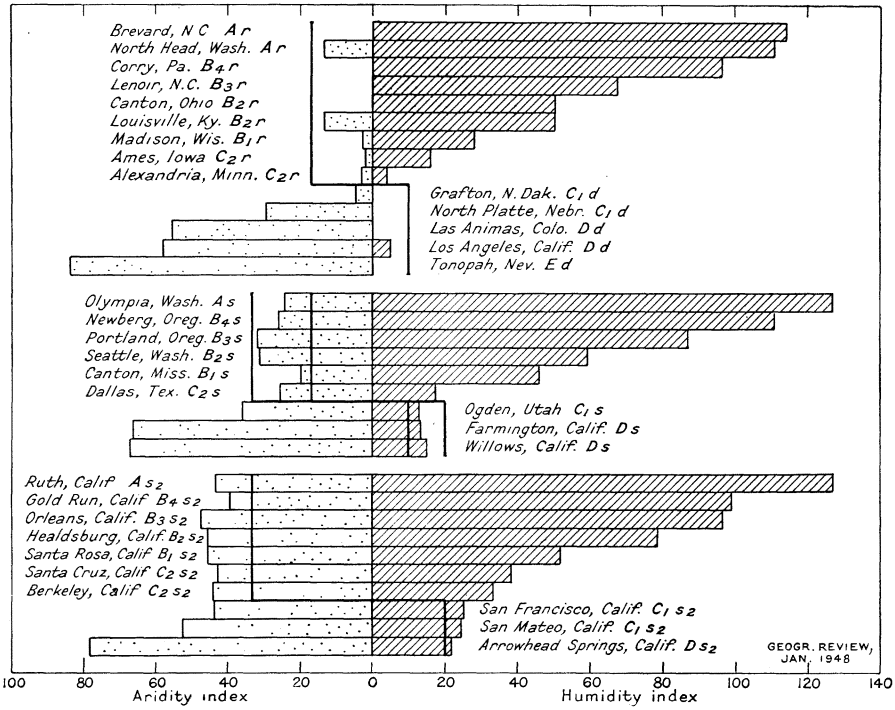

FIG. 9—Seasonal regime of moisture at selected stations.

seasonal variation of effective moisture is shown in Plate I B. Types r and d, representing small seasonal variation, are most extensive. Types s and s_2 prevail in the Pacific-coast region, where precipitation comes mainly in winter; s_2 does not occur except where the march of precipitation is opposite to that of water need and for this reason has its most extensive development in California. Type s may occur as well in regions of prevailing summer rain if summer water need sufficiently exceeds it. There are several areas of s climate in the southern United States and a few isolated stations in the Northeast. Canton, Miss. (B_1s), and Dallas, Tex. (C_2s), are examples. A large s area extends from eastern Texas and central Oklahoma eastward into Arkansas and Louisiana. A smaller area occupies the lower Mississippi Valley, and another crosses Alabama from Georgia to Mississippi along the well-known Black Belt.

The existence of the s type in the southeastern United States is surprising. Summer rainfall is fairly large, and it is not generally realized that water need is larger. Summer drought is serious in much of the eastern United States, even in the r type climates. In the s type summer drought is chronic. Supplementary irrigation becomes an important need.

AN INDEX OF THERMAL EFFICIENCY

Potential evapotranspiration is an index of thermal efficiency. It possesses the virtue of being an expression of day length as well as of temperature. It is not merely a growth index but expresses growth in terms of the water that is needed for growth. Given in the same units as precipitation it relates thermal efficiency to precipitation effectiveness.

In equatorial regions, where mean monthly temperature does not vary appreciably through the year, a mean annual temperature of 23° C. (73.4° F.) is a reasonable boundary between megathermal and mesothermal climates. Away from the equator, where there is a seasonal variation of temperature, mean annual temperature at the boundary is lower, 21.5° C. (70.7° F.), because the reduced growth and water need of winter are more than offset by an accelerated growth and an increased water need of summer. Where the mean temperature of every month is 23°, and if the station is directly on the equator, so that there is no variation in day length, the potential evapotranspiration is 114.0 centimeters (44.88 inches). This may be taken as the index that separates megathermal and mesothermal climates. Other boundaries in the series are in a descending geometric progression (p. 82).

Climatic regions based on temperature efficiency are analogous to those derived from the moisture index and are designated by similar symbols. Microthermal (C′) and mesothermal (B′) climates are subdivided like the subhumid (C) and humid (B) climates.[25] The distribution of the thermal regions in the United States is shown in Plate I C. Mesothermal (B′) climates are most extensive. Microthermal (C′) climates occur along most of the Canadian border and extend southward along the Rocky Mountains in a continuous belt to southern New Mexico. There are three areas of megathermal (A′) climate: southern Florida, southern Texas, and the arid parts of southern California and southwestern Arizona. Only small isolated areas of tundra (D′) occur, in the high Sierras of California, the northern Rockies of Wyoming and Montana, and the central Rockies of Colorado. There are no instrumental records of frost (E′) climate in the United States, and it does not appear on the map.

The megathermal (A′) climates of the United States are different from those of equatorial regions. Summer days are longer and hotter, and winter days are shorter and colder. Temperature and day length work together

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[25] Although the symbols are the same as those used in the original classification, the climatic limits differ considerably from the old ones. Consequently, the symbols have different meanings. For example, in the original scheme tundra was E′ and frost climate F′. Mesothermal climates are more extensive than previously shown.

to bring about seasonal variations in potential evapotranspiration or temperature efficiency. The difference is illustrated in Figure 10-A. Whereas at Belém, Brazil, the range is only from 10.3 to 13.1 centimeters, at Barahona, Dominican Republic, it is from 8.2 to 16.2 centimeters, and at Miami, Fla., from 5.2 to 17.8 centimeters. At the mesothermal border the mean temperature of the coldest month is as low as 50° F. in some stations. Freezing temperatures may occur almost annually, whereas in equatorial stations frosts are entirely unknown. For this reason certain plants popularly associated with tropical climates do not grow in the megathermal climates of the United States. However, plant growth continues, although at reduced rates, throughout the winter.

TE Index		Climatic Type	
cm.	*in.*		
		E'	Frost
14.2	5.61		
		D'	Tundra
28.5	11.22		
		C'_1	
42.7	16.83		Microthermal
		C'_2	
57.0	22.44		
		B'_1	
71.2	28.05		
		B'_2	
85.5	33.66		Mesothermal
		B'_3	
99.7	39.27		
		B'_4	
114.0	44.88		
		A'	Megathermal

The mesothermal (B′) climates occupy a wide range of latitudes in the eastern and central United States. Summer temperatures fall off slowly from south to north: the mean July temperature in the lower Rio Grande Valley of Texas is 85° F., and it is above 70° F. in the Yellowstone Valley of Montana. Day length, on the other hand, increases from south to north—about 14 hours in southern Texas and 16 hours in northern Montana. Consequently, average July potential evapotranspiration decreases slowly: it is about 18 centimeters in southern Texas and 15 centimeters in Montana. The fact that winter temperatures and day length both decrease from south to north has less influence because potential evapotranspiration is small even on the southern margin of the climatic zone. However, the growing season becomes shorter, and the northern edge of the zone, where annual potential evapotranspiration is only 50 per cent of the maximum for the climatic type, is reached near the Canadian border. The four subdivisions of the

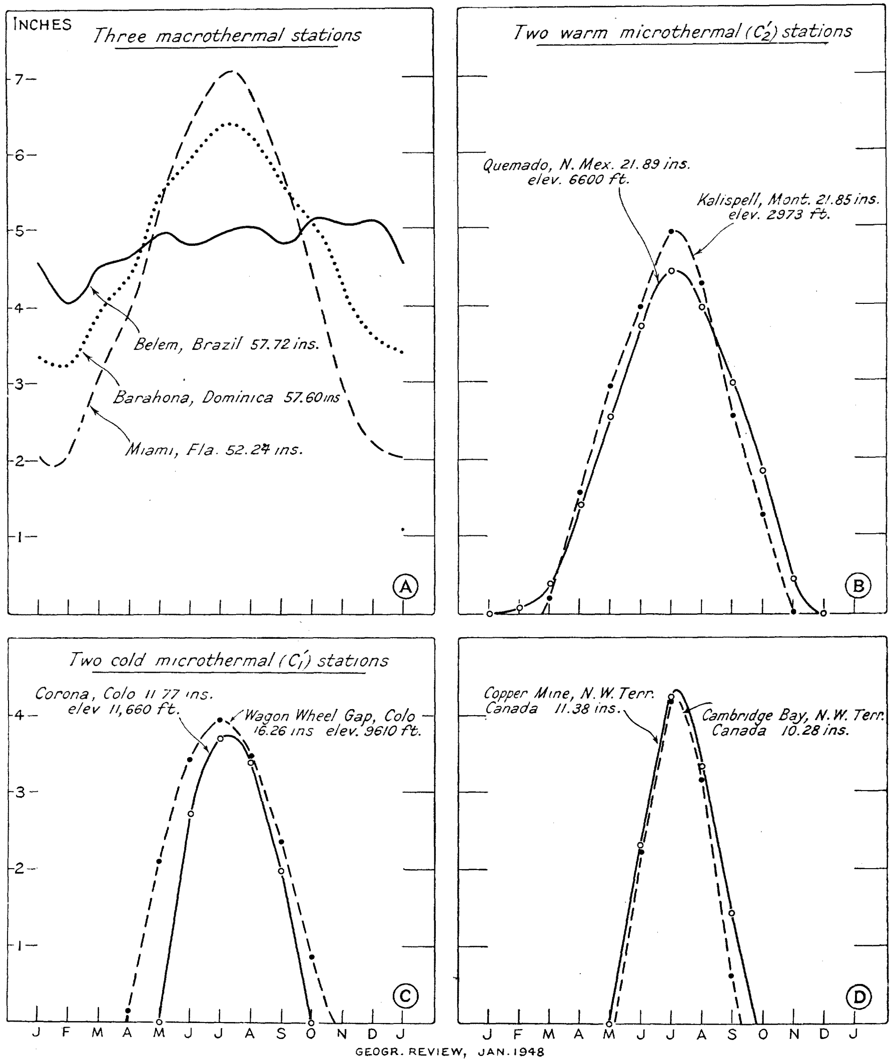

FIG. 10—March of potential evaporation. For *macrothermal* read *megathermal* (a preferable form, suggested by Mr. A. B. Hoen).

mesothermal climates appear to have real validity, but since no appropriate names for them have been found, they are identified only by subscripts 1 to 4 following symbol B'.

Microthermal (C') climates extend continuously from the Canadian border in Montana to within 200 miles of the Mexican border in New Mexico. They are at high elevations in the southern part of their range and descend to lower elevations in the north. Quemado, N. Mex., and Kalispell, Mont., have the same annual potential evapotranspiration and are both warm microthermal (C'_2); the first is 6600 feet above sea level, the second only 2973 feet. Figure 10–B shows the march of potential evapotranspiration in these two stations. Spring and summer temperatures are considerably lower in Kalispell. The growing season is shorter, but since summer days are longer, higher rates of water use and more rapid growth compensate for shorter time of plant activity.

The colder type of microthermal climates (C'_1) is not areally important in the United States.[26] The two most extensive areas are in Wyoming and Colorado, though smaller areas are found in several western states. Wagon Wheel Gap, Colo., is near the warm limit of the type, and Corona, Colo., is near the cold limit (Fig. 10-C). Although the temperatures of the warmest months are low (Corona mean July temperature, 48.8° F.; Wagon Wheel Gap, 56.2° F.), the potential evapotranspiration is surprisingly large, nearly 10 centimeters (4 inches). Annual precipitation is in excess of water need in both places, but summer rainfall is less than summer potential evapotranspiration. At Wagon Wheel Gap summer water deficiency amounts to 2.2 inches. Short growing season and low summer temperature do not prevent droughts.

No currently reporting Weather Bureau station in the United States has tundra (D') climate. However, it undoubtedly occurs in various isolated high areas, mostly in the West. The summit of Mt. Washington, in New Hampshire, has an annual water need of 12.9 inches, which places it on the tundra margin. Figure 10-D shows the march of potential evapotranspiration in two stations in the far north of Canada. Both are very near the microthermal-tundra boundary; Coppermine is just above the line, and Cambridge Bay just below. Although the mean temperature of July, the warmest month, is only about 50° F. in both places, the water need is astonishingly high, being above 10 centimeters (4 inches). Average annual potential evapotranspiration exceeds precipitation in both places, and summer drought is a characteristic feature of the climate. In southern and eastern Greenland, on the other hand, precipitation greatly exceeds evapotranspiration, and there is a large water surplus, a part of which nourishes the icecap, and a part of which runs off in summer.

In the climate of perpetual frost (E') temperatures need not remain continuously below the freezing point. Although frost effectively prevents vegetation growth, and thus transpiration, nevertheless some water passes into the atmosphere by evaporation or sublimation. At the outer limit of the frost climate potential evapotranspiration is 14.2 centimeters (5.61

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[26] This type is equivalent to taiga (D') in the previous classification.

inches). It is evident that even in the frost climate moisture differentiation is important, because if precipitation exceeds potential evapotranspiration, ice will accumulate and a glacier will develop. On the other hand, if precipitation is less than this critical value, ice accumulation is impossible. It is not certain that true frost (E′) climate exists anywhere in the United States. A snow line or mountain glacier does not necessarily indicate a frost climate.

SUMMER CONCENTRATION OF THERMAL EFFICIENCY

At the equator, where day length is the same throughout the year and where temperature is also uniform, seasonal variation of potential evapotranspiration will be small. With no variation no season can be called summer, and the potential evapotranspiration of any consecutive three months will constitute 25 per cent of the annual total. On the other hand, in the Polar Regions, where the growing season is entirely within the three summer months, the potential evapotranspiration of these months will constitute 100 per cent of the total. Between these extremes, as potential evapotranspiration falls from that characteristic of megathermal (A′) climates to that of frost (E′) climates, the part that is concentrated in summer gradually rises from 25 per cent to 100 per cent. This rise results from an increasing length of midsummer days, and an increase in the length of winter, with increase in latitude.

The astronomical motions of the earth tend to produce a certain fixed relation between the summer index of thermal efficiency and the annual index. This relation has been approximately determined from a series of stations in the interior low plains of North America. The summer concentration appears to be inversely proportionate to the logarithm of the annual index. The relation is described by the equation

$$s = 157.76 - 66.44 \log E, \qquad (7)$$

in which s is summer concentration in percentages and E is potential evapotranspiration in inches.

But there are a number of meteorological situations in which the relation is modified or altered and is, in consequence, abnormal. On small oceanic islands and seacoasts dominated by onshore winds in middle and high latitudes summer temperatures are less warm, and winter temperatures less cold, than elsewhere in these latitudes. The summer concentration of thermal efficiency is lower than it should be for the latitude. In mountains there is a more or less uniform reduction of temperature throughout the year. Annual and summer thermal-efficiency indices are both reduced, but not in the same proportion. Here, too, the summer concentration of thermal

Table III—Comparative Moisture Data of Selected Stations

In inches

Station	Water need	Summer need %	Precip-itation	Water surplus	Water def.	Surplus % of need	Def. % of need	Moisture index	Climatic type
A. Brevard, N. C.	28.50	51.1	61.06	32.56	0.00	114.2	0.	114.2	A $B'_2rb'_4$
B. Salisbury, N. Y.	21.80	67.2	48.39	26.58	0.00	121.8	0.	121.8	A $C'_2rb'_1$
C. Bar Harbor, Maine	22.32	59.1	47.09	25.55	0.00	114.5	0.	114.5	A $C'_2rb'_2$
D. North Head, Wash.	24.72	41.1	48.86	27.40	3.27	110.8	13.2	102.9	A B'_1ra'
E. Louisville, Ky.	31.77	55.5	43.47	15.95	4.28	50.2	13.5	42.2	$B_2B'_2rb'_3$
F. Seattle, Wash.	26.14	46.4	33.35	15.43	8.23	59.0	31.5	40.2	$B_2B'_1sa'$
G. Canton, Miss.	38.70	50.6	49.10	17.68	7.68	45.6	19.8	33.8	$B_1B'_3sb'_4$
H. Willard, N. C.	35.35	49.7	49.05	13.70	0.00	38.8	0.	38.8	$B_1B'_3rb'_4$
I. Madison, Wis.	25.12	61.4	31.54	7.09	0.67	28.2	2.7	26.6	$B_1B'_1rb'_2$
J. Centennial, Wyo.	18.86	63.7	17.68	4.49	5.67	23.8	30.1	5.7	$C_2C'_2sb'_1$
K. Manhattan, Kans.	30.83	57.8	31.50	2.28	1.61	7.4	5.2	4.3	$C_2B'_2rb'_2$
L. Pullman, Wash.	24.13	58.1	20.71	7.48	10.91	31.0	45.3	3.9	$C_2B'_1s_2b'_2$
M. San Francisco, Calif.	27.09	33.3	21.88	6.68	11.89	24.6	43.8	−1.7	$C_1B'_1s_2a'$
N. Dalhart, Tex.	29.02	56.9	19.09	0.00	9.92	0.	34.2	−20.5	D $B'_2db'_2$
O. Helena, Mont.	22.32	48.1	13.31	0.00	9.01	0.	40.4	−24.2	D $C'_2db'_4$
P. Santa Fe, N. Mex.	24.41	56.1	14.21	0.00	10.19	0.	41.7	−25.1	D $B'_1db'_3$
Q. Los Angeles, Calif.	32.01	38.7	14.96	1.42	18.47	4.4	57.7	−30.2	D B'_2da'
R. Grand Junction, Colo.	28.74	59.2	8.70	0.00	20.04	0.	69.9	−41.8	E $B'_2db'_2$
S. Fresno, Calif.	36.22	52.8	9.17	0.00	27.04	0.	74.7	−44.8	E $B'_3db'_3$
T. Tonopah, Nev.	25.43	58.7	4.84	0.00	21.38	0.	84.0	−50.4	E $B'_1db'_2$

efficiency is abnormally low. On the other hand, in some regions the normal seasonal march of temperature is exaggerated. Summer temperatures are increased by advection of hot tropical air, and winter temperatures are lowered by advection of cold polar air. In these regions the summer concentration is abnormally high.

POTENTIAL EVAPOTRANSPIRATION *Ins.*	*Cms.*	TEMPERATURE-EFFICIENCY TYPE	SUMMER CONCENTRATION *Percentage*	SUMMER CONCENTRATION *Type*
		A′		a′
44.88	114.0		48.0	
		B'_4		b'_4
39.27	99.7		51.9	
		B'_3		b'_3
33.66	85.5		56.3	
		B'_2		b'_2
28.05	71.2		61.6	
		B'_1		b'_1
22.44	57.0		68.0	
		C'_2		c'_2
16.83	42.7		76.3	
		C'_1		c'_1
11.22	28.5		88.0	
		D′		d′
5.61	14.2			
		E′		

The extent to which the summer-concentration percentage fails to meet the above requirements is a measure of its abnormality. San Francisco, for example, has a potential evapotranspiration of 27.09 inches and a summer concentration of 33.3 per cent. Thus summer concentration is that of full megathermal (a′) climate, though the temperature-efficiency type is actually only first mesothermal (B'_1). San Francisco is an example of a so-called "marine" climate.

Summer concentration of thermal efficiency in the United States is shown in Plate I D. The displacement of the various zones from the comparable ones of thermal efficiency (Pl. I C) is a measure of oceanity or continentality.

The abnormality of the summer-concentration percentage deserves special study, but it will be more profitable on a world basis than for the United States alone.

ELEMENTS OF THE CLASSIFICATION

Four symbols used together give a complete description of a climate. The necessary data for classifying the stations in Figure 4 are presented in Table III. Water need, in the first column, is, of course, potential evapotranspiration. The second column gives the percentage that summer potential evapotranspiration is of the annual total. The column labeled "Surplus as percentage of need" gives the humidity index, and that labeled "Deficiency as percentage of need" the index of aridity. The moisture index is obtained by subtracting six-tenths of the latter from the former.

The various subdivisions of mesothermal, microthermal, and humid climatic types do not have individual names but can be referred to only by symbol. Thus we can say first, second, third, or fourth mesothermal. Brevard, N. C. ($AB'_2rb'_4$), is perhumid, second mesothermal, with no season of water deficiency, and a temperature-efficiency regime normal to fourth mesothermal. San Francisco, Calif. ($C_1B'_1s_2a'$), is dry subhumid, first mesothermal, with large winter water surplus, and a temperature-efficiency regime normal to megathermal.

Superficially the present system is similar to its predecessor[27] in that the same factors are employed; namely a moisture factor, a heat factor, and the seasonal variation of the two. Actually, the two systems are fundamentally different. In the earlier classification, climatic types were identified, and boundaries were located, empirically, through study of the distribution of vegetation, soils, drainage features, and so on. In the present classification, climates are defined rationally, and boundaries are determined by the data.

The difference may be illustrated by the change in point of view respecting vegetation. The earlier study adopted Köppen's position that the plant is a meteorological instrument which integrates the various factors of climate and which, with experience, can be "read" like a thermometer or a rain gauge. In the present study, vegetation is regarded as a physical mechanism by means of which water is transported from the soil to the atmosphere; it is the machinery of evaporation as the cloud is the machinery of precipitation.

Climatic boundaries are determined rationally by comparing precipitation and evapotranspiration. The subdivisions of the older classification were justly criticized as being vegetation regions climatically determined. The present climatic regions are not open to this criticism, since they come from a study of the climatic data themselves and not from a study of vegetation.

This classification can be improved. A first step will be to develop better means of determining potential evapotranspiration. Additional observations are needed, particularly in the tropics and in high latitudes. With new data available, the present formula can be revised, or perhaps a new and more rational formula can be devised. A truly rational method of delimiting the temperature-efficiency regions has not yet been developed. A relation be-

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[27] Thornthwaite, The Climates of North America (*op. cit.*).

tween the heat factor and the moisture factor may exist that will provide the rational basis. So far it has not been discovered.

There is an encouraging prospect that this climatic classification which is developed independently of other geographical factors such as vegetation, soils, and land use, may provide the key to their geographical distribution.

Soil-forming processes are related to water surpluses and water deficiencies; so are hydrological regimes and drainage patterns. The problem of the origin of the prairies may come near solution in an analysis of annual frequency of water deficiency and water surplus. Furthermore, much can be learned regarding soil productivity and best land use through a study of magnitude and frequency of water deficiency. Finally, we have a better understanding than ever before of the qualities of a climate when we are able to compare the potential evapotranspiration through the year with the precipitation.

APPENDIX I

THE DETERMINATION OF POTENTIAL EVAPOTRANSPIRATION

The relation between mean monthly temperature and potential evapotranspiration adjusted to a standard month of 30 days, each having 12 hours of possible sunshine, in four selected areas is shown in Figure 11. In each area the relationship within the range of temperature involved is well expressed by an equation of the form

$$e = ct^a, \tag{8}$$

in which e is monthly evapotranspiration in centimeters and t is mean monthly temperature in °C. From these observations it is seen that there is no simple relationship between monthly evapotranspiration and monthly temperature. The coefficients c and a vary from one place to another. Thus an equation having coefficients derived from observations made in a warm climate does not yield correct values of potential evapotranspiration for an area having a cold climate, and vice versa. In Figure 12 the lines showing the relationship between temperature and evapotranspiration in several areas tend to converge where potential evapotranspiration is 13.5 centimeters and temperature is 26.5° C. At lower temperatures there is increasing divergence in potential evapotranspiration.

In a general equation constants c and a must be allowed to vary with a factor that is small in cold climates and large in hot climates. Mean annual temperature is not satisfactory because in some places it is affected by below-freezing temperatures. A special equation was developed for the purpose. A monthly index is obtained from the equation $i = (t/5)^{1.514}$. Summation of the 12 monthly values gives an appropriate heat index, I. While this index varies from 0 to 160, the exponent a in the above equation varies from 0 to 4.25. The relation between the two is closely approximated by the expression

$$a = 0.000000675I^3 - 0.0000771I^2 + 0.01792I + 0.49239. \tag{9}$$

The coefficient c in equation (8) above varies inversely with I. From these relations a general equation for potential evapotranspiration was obtained. It is

$$e = 1.6\ (10t/I)^{a}, \tag{10}$$

in which a has the value given in equation (9).

The formula gives unadjusted rates of potential evapotranspiration. Since the number of days in a month ranges from 28 to 31 (nearly 11 per cent) and the number of hours in the day between sunrise and sunset, when evapotranspiration principally takes place, varies with the season and with latitude, it becomes necessary to reduce or increase the unadjusted rates by a factor that varies with the month and with latitude.

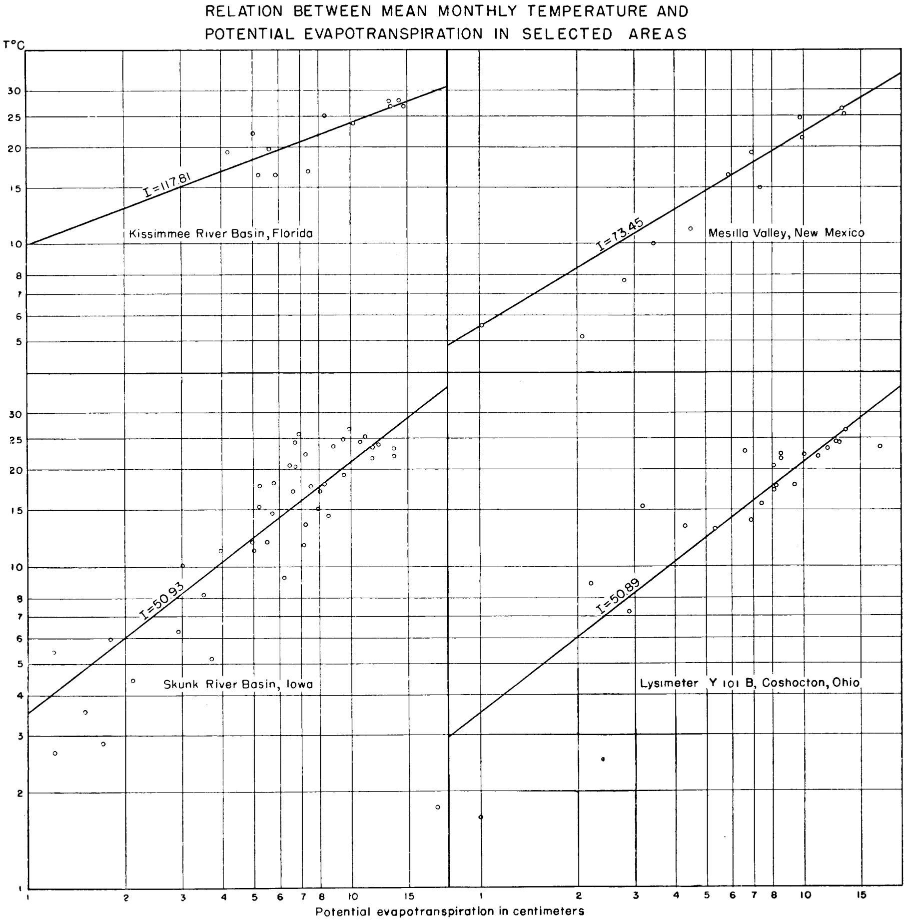
RELATION BETWEEN MEAN MONTHLY TEMPERATURE AND POTENTIAL EVAPOTRANSPIRATION IN SELECTED AREAS

FIG. 11

This mathematical development is far from satisfactory. It is empirical, and the general equation does not accord with the newly developed law of growth. Furthermore, the equation is completely lacking in mathematical elegance. It is very complicated and without nomograms and tables as computing aids would be quite unworkable. The chief obstacle at present to the development of a rational equation is the lack of understanding of why potential evapotranspiration corresponding to a given temperature is not the same everywhere. Not until this matter is understood will a rational method be possible.

In spite of the lack of a theoretical foundation, this method makes it possible to arrive at values of potential evapotranspiration that are approximately correct and that give an entirely new approach to the problems of climatic classification.

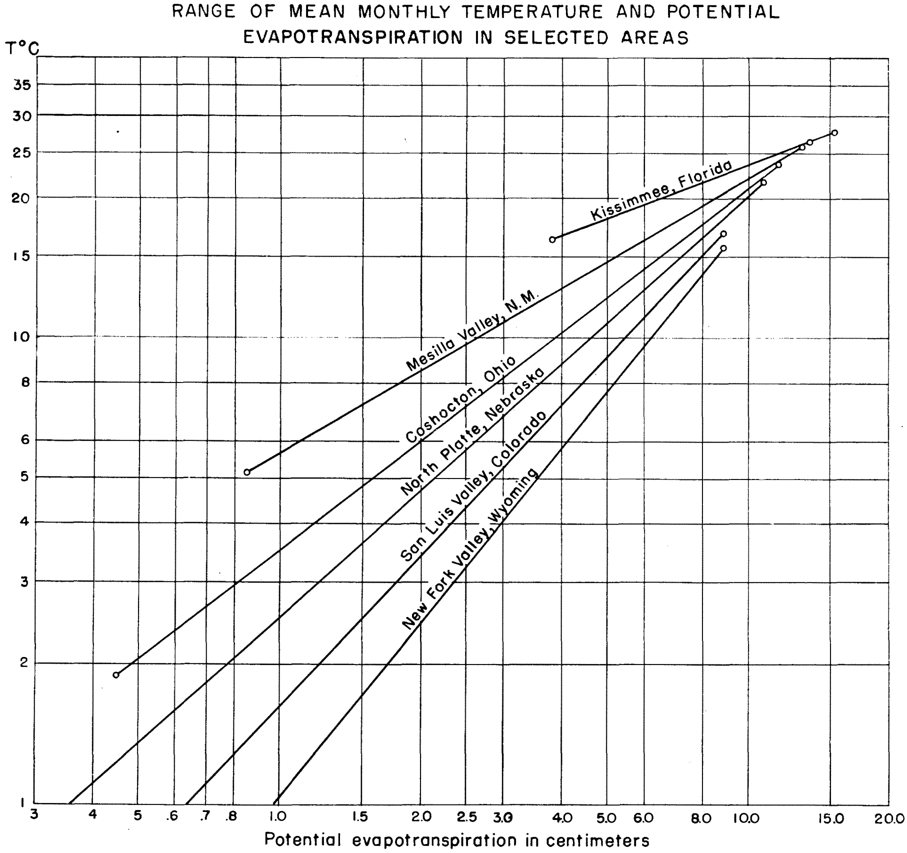

FIG. 12

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 and tables.

The first step is to obtain the heat index, *I*. Table IV gives monthly values of *i* corresponding to monthly mean temperatures. Summation of the 12 monthly values gives the index, *I*.

The next step is to determine unadjusted values of potential evapotranspiration from the nomogram in Figure 13.[28] Since there is a linear relation between the logarithm of temperature and the logarithm of unadjusted potential evapotranspiration, straight lines on the nomogram define the relationship. All lines pass through the point of convergence at

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[28] A table has been prepared for this step, but it is too long to publish here. It will be placed on file at the American Geographical Society.

TABLE IV

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									

$t = 26.5°$ C. and $PE = 13.5$ cm. The slope of the line is determined by the heat index of the station. For example, the heat index of Brevard, N. C., is 56.0, and the line ruled on the nomogram represents the relationship between potential evapotranspiration and temperature at that place. At a mean temperature of 10° C. (50° F.) the unadjusted potential evapotranspiration is 3.6 centimeters.[29] Knowing the index (I) of the station, one sets a straightedge in

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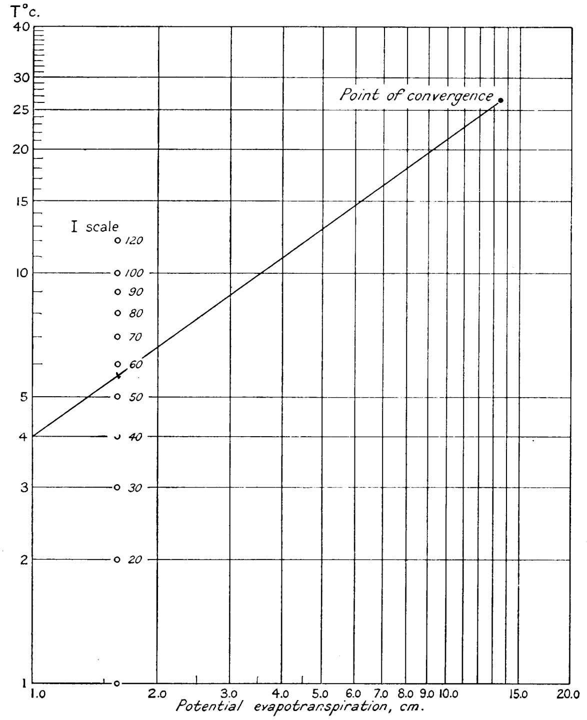
[29] For accurate computation it is desirable to make a nomogram on a larger scale on logarithm paper that is commercially available. Keuffel and Esser No. 359-112L (logarithmic, 2 x 3 cycles) is satisfactory. The point of convergence and the heat-index scale are easily added. A transparent rule may be used.

Table V—Mean Possible Duration of Sunlight in the Northern and Southern Hemispheres Expressed in Units of 30 Days of 12 Hours Each

N. Lat.	J	F	M	A	M	J	J	A	S	O	N	D
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

the appropriate position on the nomogram and reads potential evapotranspiration corresponding to the given mean temperature of the month. The nomogram is used only when tempera-



T.°C.	PE
26.5	13.50
27.0	13.95
27.5	14.37
28.0	14.78
28.5	15.17
29.0	15.54
29.5	15.89
30.0	16.21
30.5	16.52
31.0	16.80
31.5	17.07
32.0	17.31
32.5	17.53
33.0	17.72
33.5	17.90
34.0	18.05
34.5	18.18
35.0	18.29
35.5	18.37
36.0	18.43
36.5	18.47
37.0	18.49
37.5	18.50
38.0	18.50

FIG. 13—Nomogram for determining potential evapotranspiration and table for use at higher temperatures.

ture is 26.5° C. or less. (The accompanying table gives potential evapotranspiration corresponding to the higher temperatures). Twelve values are obtained for the 12 months. These are unadjusted values for months of 30 days of 12 hours each.

Finally, these values of potential evapotranspiration are adjusted for day and month length. Table V gives correction factors by which the unadjusted potential evapotranspiration of each month must be multiplied. The corrections must be the appropriate ones for the latitude of the station. The correction factor for 50° N. latitude will be used for all stations farther north. The same limitation applies in the Southern Hemisphere. Potential evapotranspiration is in centimeters. Summation of the 12 monthly values gives annual potential evapotranspiration in centimeters. The computations for Brevard, N. C., follow:

	J	F	M	A	M	J	J	A	S	O	N	D	Y
t.°C.	0.9	1.1	4.7	9.2	13.3	17.3	18.8	18.4	16.1	10.3	4.7	1.2	9.7
I	0.07	0.10	0.91	2.52	4.40	6.55	7.43	7.19	5.87	2.99	0.91	0.12	39.06
Unadj. PE	0.2	0.4	2.0	4.2	6.3	8.4	9.2	9.0	7.2	4.7	1.9	0.4	
Adj. PE	0.2	0.3	2.1	4.6	7.6	10.2	11.4	10.4	7.4	4.6	1.6	0.3	60.7

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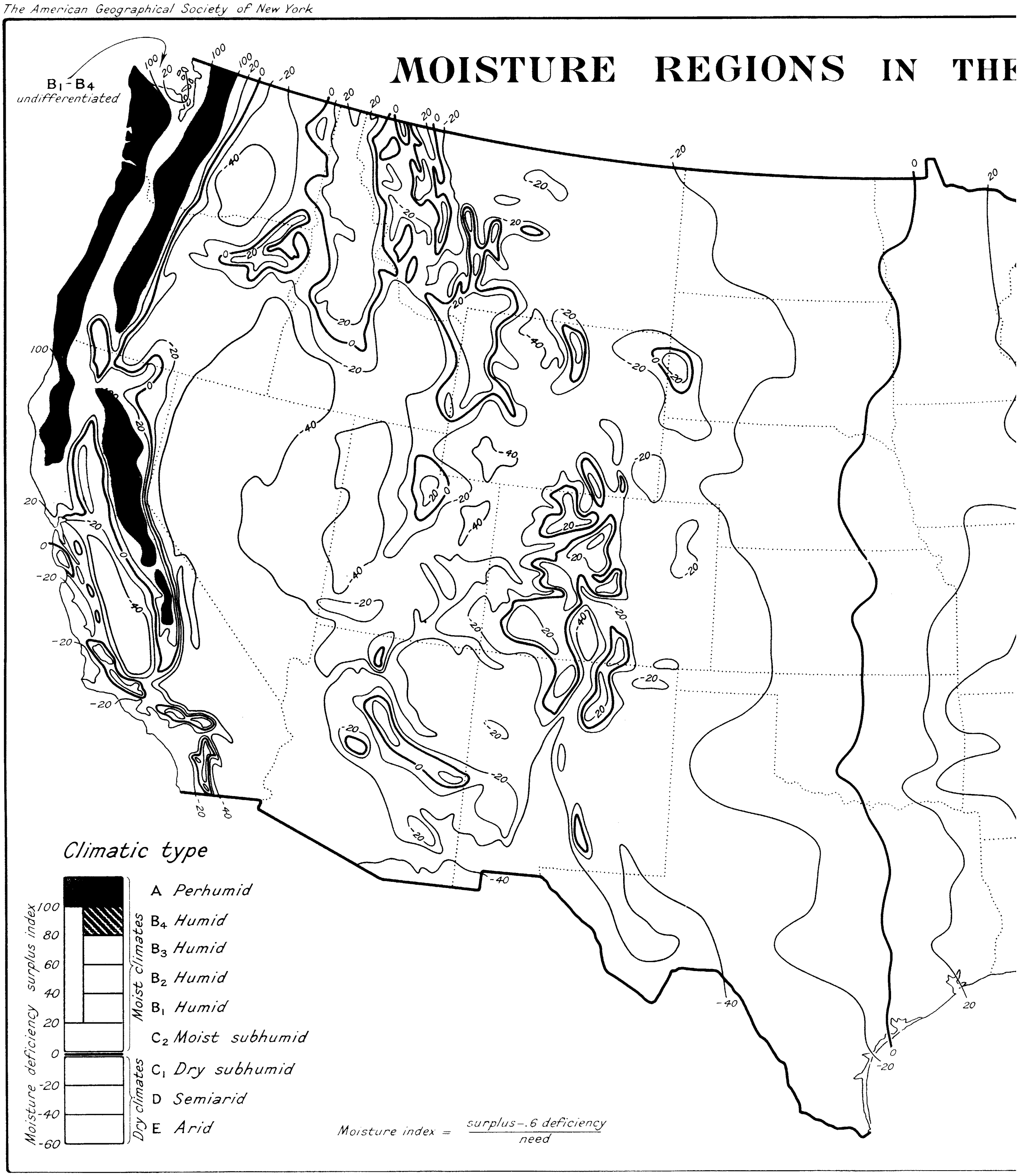
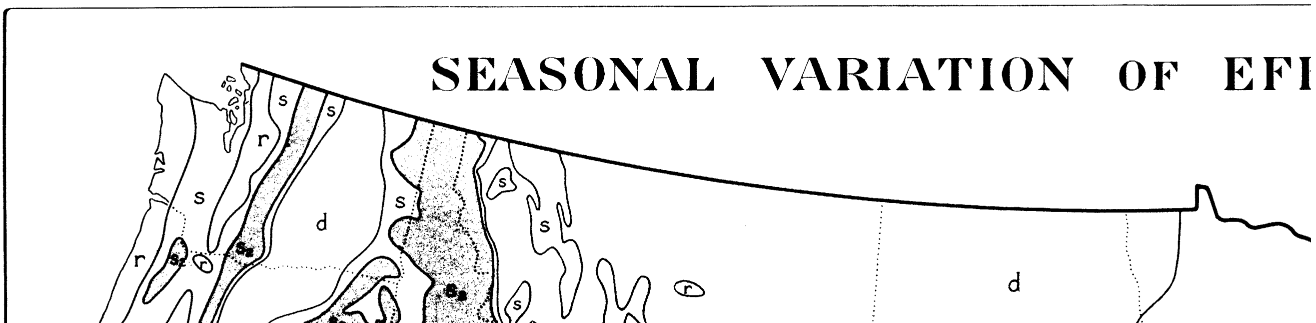
MOISTURE REGIONS IN THE

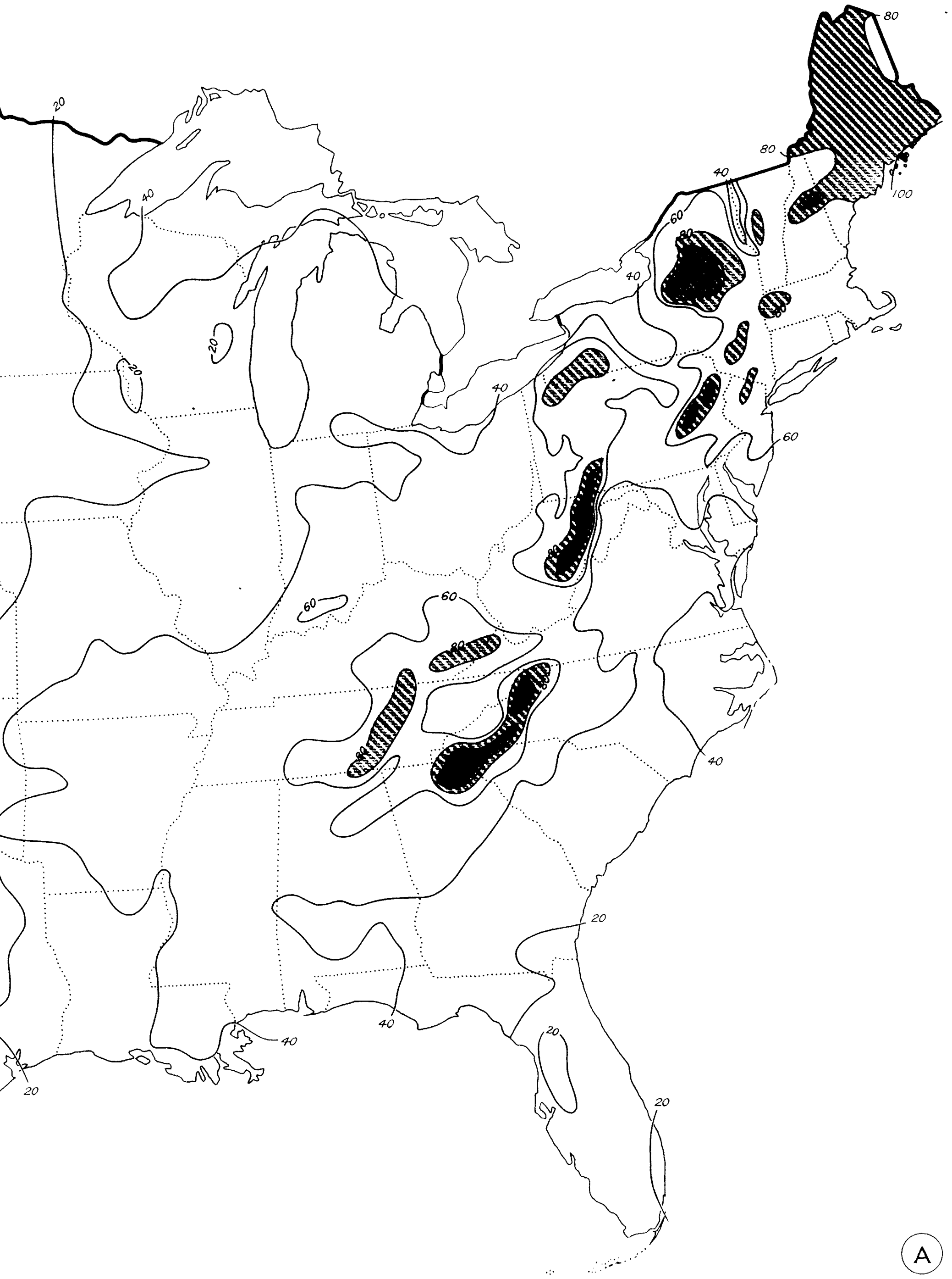


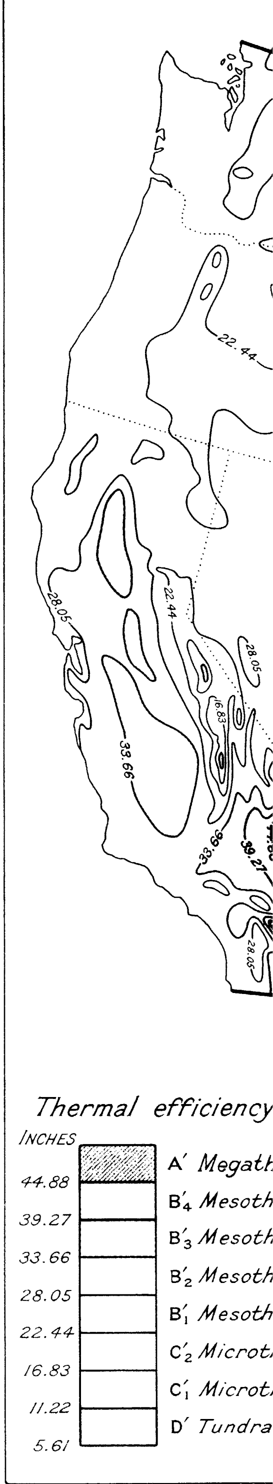
$$\text{Moisture index} = \frac{\text{surplus} - .6\ \text{deficiency}}{\text{need}}$$

SEASONAL VARIATION OF EF

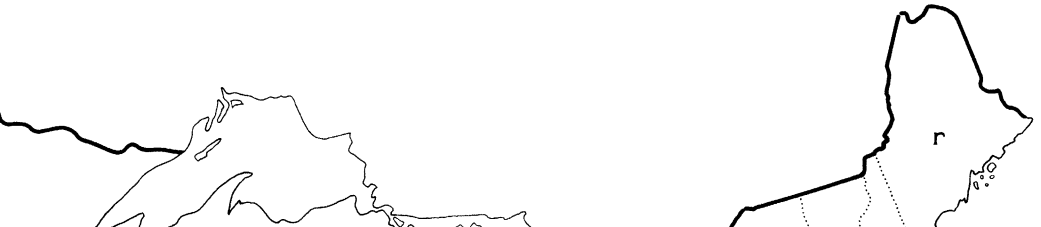


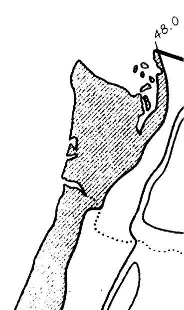
THE UNITED STATES



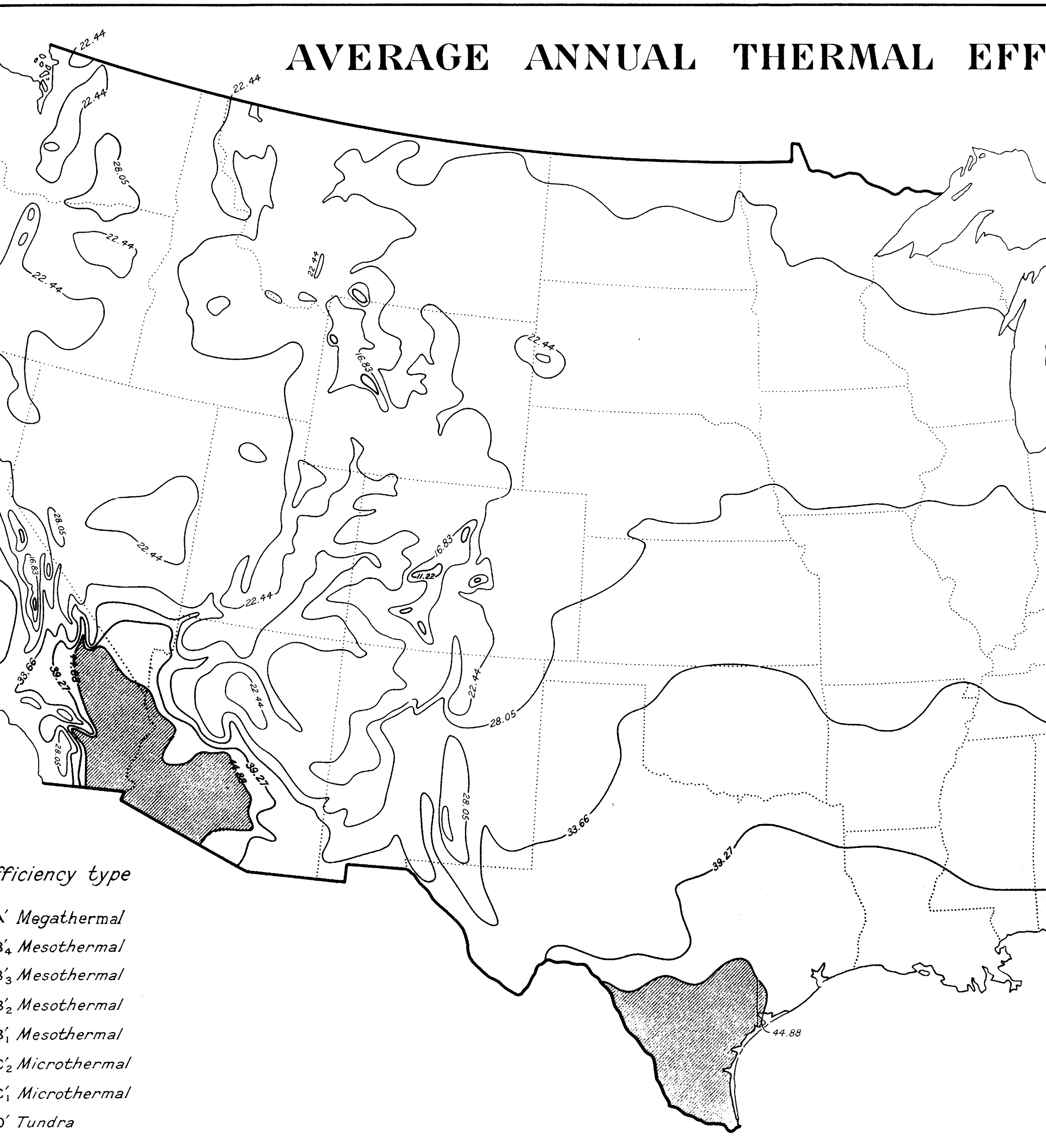


EFFECTIVE MOISTURE

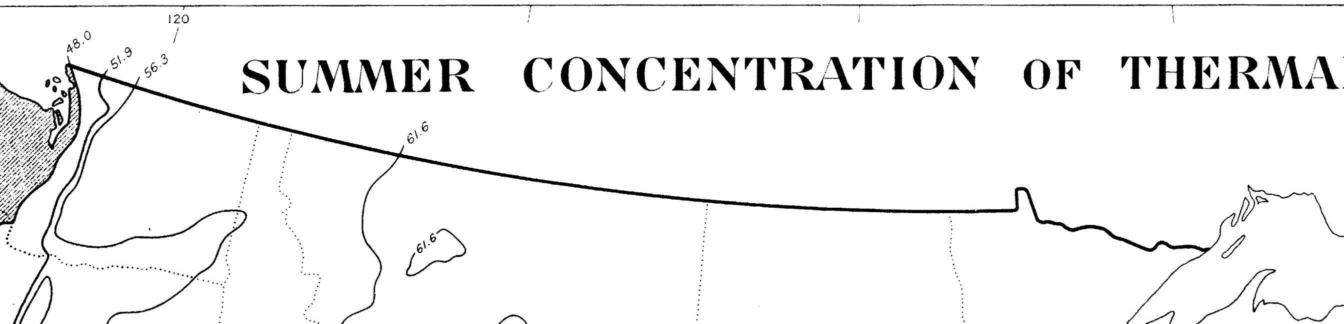


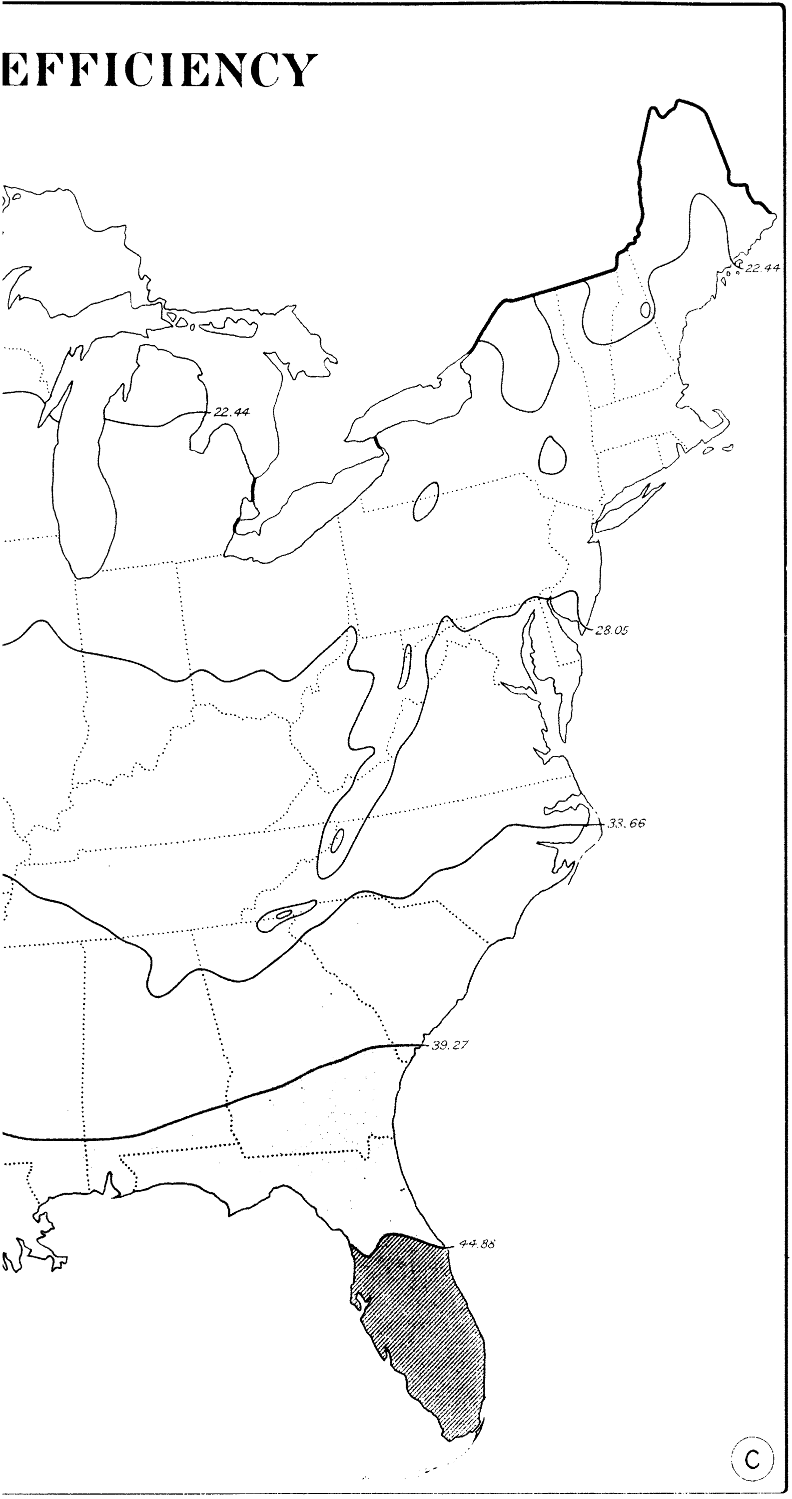


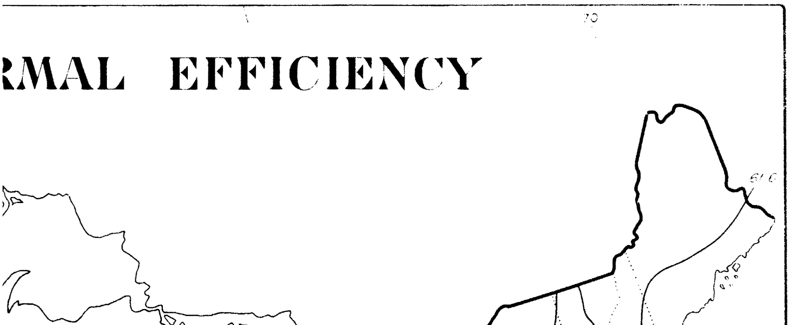
AVERAGE ANNUAL THERMAL EFF

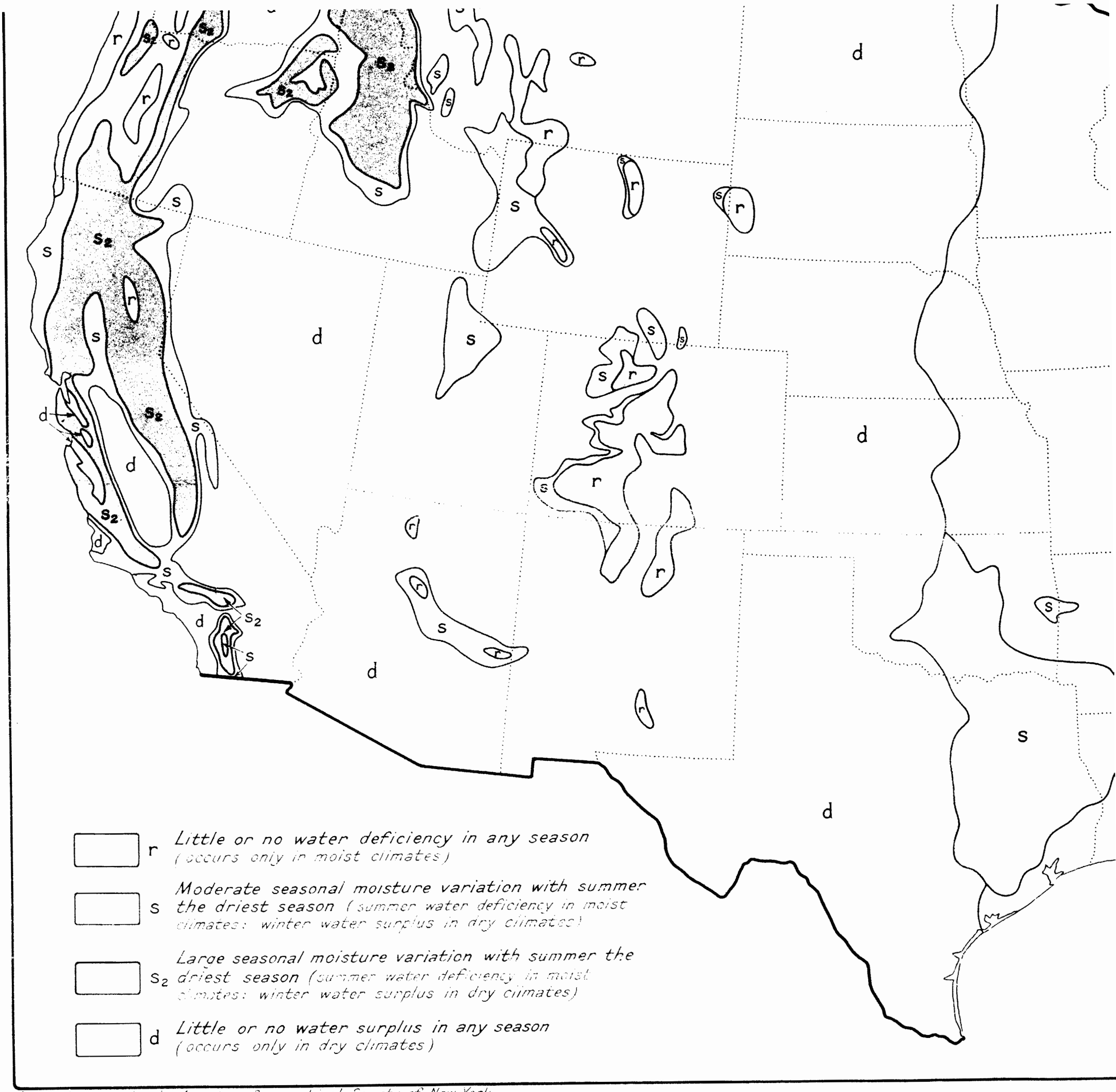


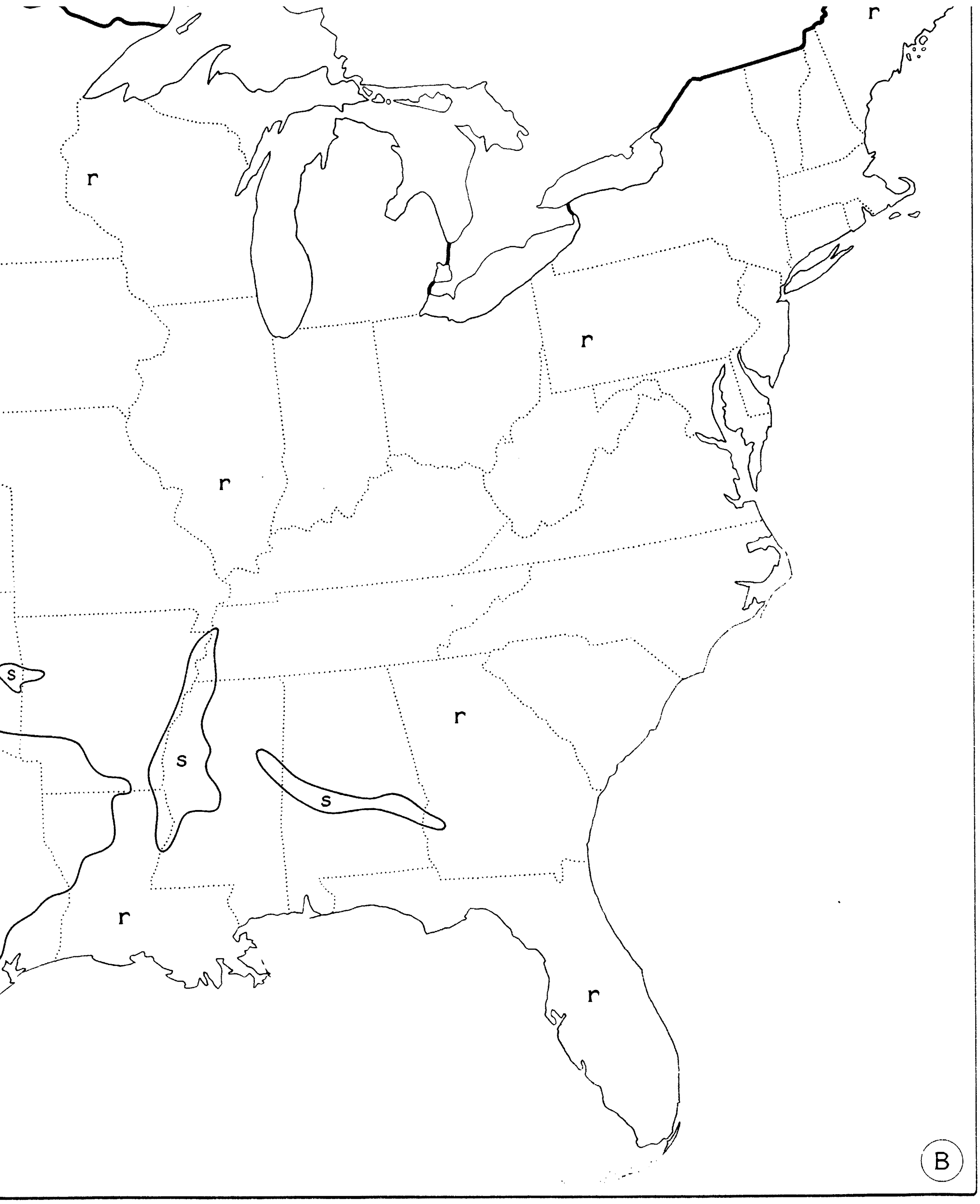
SUMMER CONCENTRATION OF THERMA

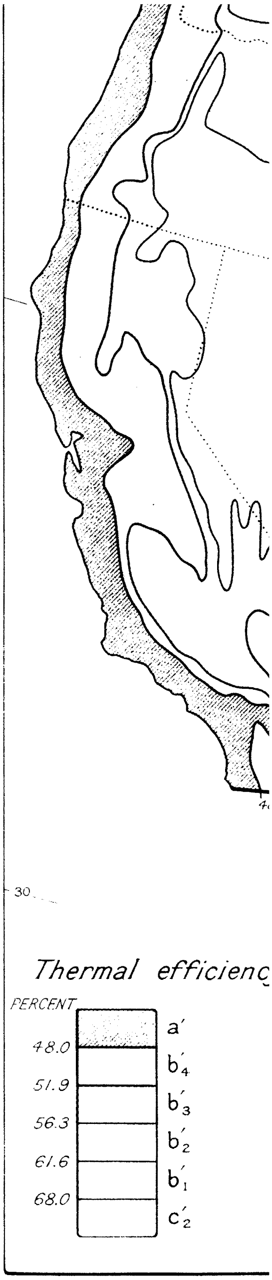



EFFICIENCY
22.44
22.44
28.05
33.66
39.27
44.88
C

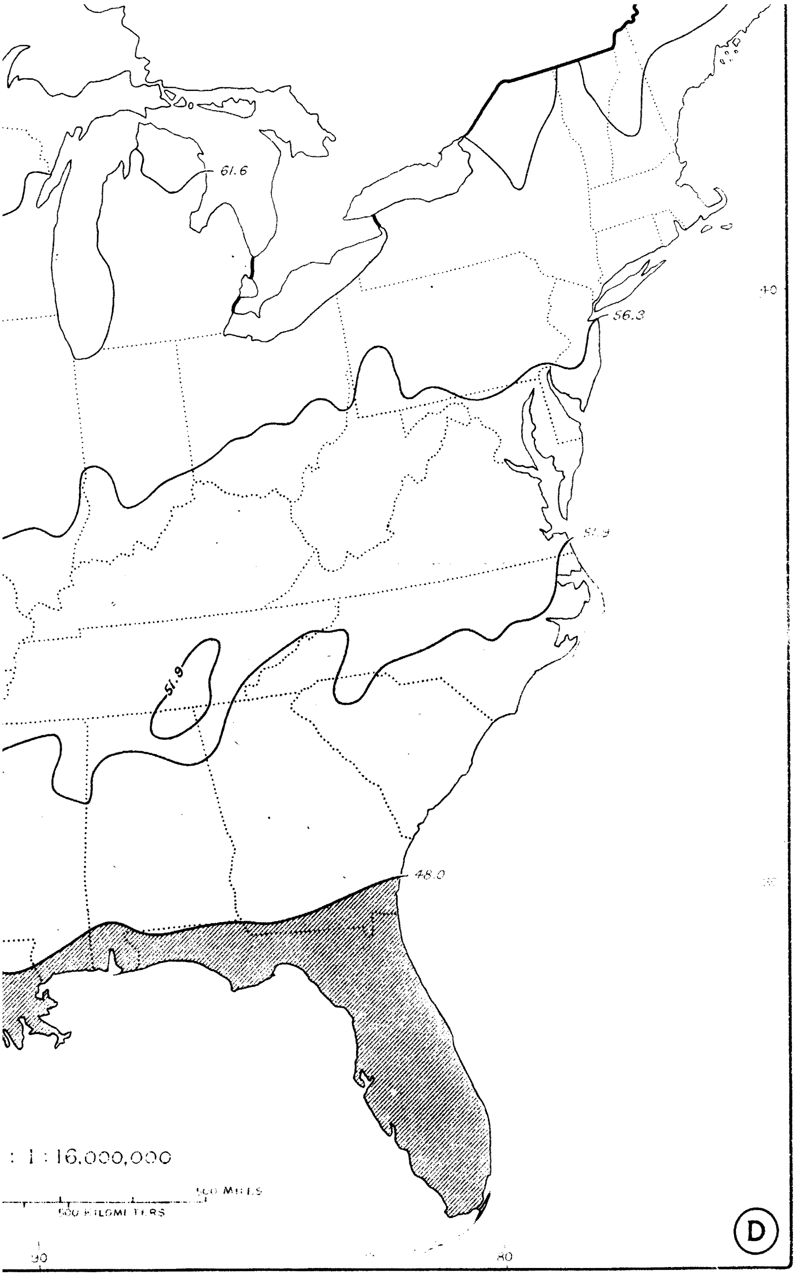

MAL EFFICIENCY


r Little or no water deficiency in any season (occurs only in moist climates)
s Moderate seasonal moisture variation with summer the driest season (summer water deficiency in moist climates: winter water surplus in dry climates)
s2 Large seasonal moisture variation with summer the driest season (summer water deficiency in moist climates: winter water surplus in dry climates)
d Little or no water surplus in any season (occurs only in dry climates)
r
s
s2
d


r
r
r
r
r
r
r
s
s
s
s
B


30
Thermal efficienc
PERCENT
a'
48.0
b'_4
51.9
b'_3
56.3
b'_2
61.6
b'_1
68.0
c'_2


61.6
61.6
68.0
68.0
48.0
48.0
51.9
fficiency type
a′
b′4
b′3
b′2
b′1
c′2
SCALE: 1:16,0
0
500
110
100
90



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