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POTENTIAL EVAPORATION AND RIVER BASIN EVAPORATION

By Fred I. Morton[1]

INTRODUCTION

Although the return of water to the atmosphere from water, soil, and vegetation surfaces is one of the most important aspects of the hydrologic cycle, hydrologists and engineers have performed surprisingly little research on the subject. Undoubtedly, this is because of the emphasis that has been placed on the development of statisitical and runoff distribution techniques for the solution of practical problems. However, considerable research on evaporation and transpiration has been performed by workers in other fields such as meteorology, botany, geography, agriculture, and forestry. Although this work has been extremely varied in purpose, scope, method, and results, it has advanced the state of knowledge to the point where it has some application in the field of hydrology. This paper has been prepared to show that it is now (1965) possible to formulate a working hypothesis for estimating the evaporation from the water, soil, and vegetation surfaces of river basins and to derive supporting evidence from published climatological and hydrological records.

It has been customary to differentiate between the evaporation from water surfaces and the evaporation from land and vegetation surfaces by defining the latter as evapotranspiration. As the only significant difference is the control imposed on the quantity of water available for evaporation by soil moisture and vegetative processes, the distinction is unwarranted. Thus, the mass transfer and energy balance approaches to estimating evaporation from water surfaces are applicable also to estimating evaporation from moist soil and vegetation surfaces. This has led to the development of the concept of potential evaporation.

As usually defined, potential evaporation is the evaporation that would occur if there were an adequate moisture supply at all times. This definition does not specify the area of adequate moisture supply and thus fails to allow for

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the effect of evaporation on the energy available for evaporation, as reflected in climatological observations. Potential evaporation, as defined herein, is the evaporation from a moist surface so small that the evaporation has no appreciable influence on the air passing over it. Under this definition, potential evaporation may be derived from regional climatological observations and, because these observations reflect the regional weather, they are influenced significantly by the regional evaporation. Therefore, potential evaporation is a function of the regional evaporation. However, because regional evaporation is dependent on soil moisture and vegetative conditions in addition to the factors affecting potential evaporation, the two quantities are equal only when the soil and vegetation surfaces of the region are covered with moisture.

Both the mass transfer and energy balance approaches to the computation of evaporation require observations of the evaporating surface temperature. This seriously limits their utility because in the case of actual evaporation, it is difficult to make accurate measurements of the temperature of the evaporating surface and, in the case of potential evaporation, the evaporating surface might not even exist. However, it has been shown by H. L. Penman[2] and J. Ferguson[3] that it is possible to eliminate the need for surface temperature measurements under certain conditions by combining the mass transfer and energy balance equations into what is defined herein as the mass and energy transfer approach. The techniques used by Penman and Ferguson to accomplish this purpose appear to be different but are based on similar assumptions and provide similar results. As both methods are based on accepted physical principles, they should provide adequate estimates of potential evaporation.

The relationship between potential evaporation and the evaporation from water, soil, and vegetation surfaces of a region has always posed serious problems. Many workers have assumed a proportionality, varying from season to season. Such an assumption does not inspire confidence because it is known that the regional evaporation is dependent not only on the available energy, as reflected in the concept of potential evaporation, but also on such factors as soil moisture, conditions and vegetative processes. A new approach to the solution of the problem has been suggested by R. J. Bouchet.[4] From a consideration of the energy balance, he postulated that the potential evaporation is a measure of the quantity of energy remaining from the absorbed insolation after energy has been used for regional evaporation. Thus, a restriction in the regional evaporation, caused by a soil moisture deficiency or some vegetative process, results in increased temperatures, increased wind velocities, and decreased humidities, which are reflected in increased potential evaporation. With certain assumptions and approximations of relatively minor importance, the hypothesis leads to the conclusion that the regional evaporation is equal to the absorbed insolation less the potential evaporation.

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2 Penman, H. L., "Natural Evaporation from Open Water, Bare Soil and Grass," Proceedings, Royal Soc. of London, London, England, Series A, Vol. 193, No. 1032, 1948.

3 Bonython, C. W., "Evaporation Studies Using some South Australian Data," Transactions, Royal Soc. of South Australia, Adelaide, Australia, Vol. 73, Part 2, December, 1950.

4 Bouchet, R. J., "Evapotranspiration Réelle et Potentielle Signification Climatique," Publication No. 62, Internatl. Assn. of Scientific Hydrology, Genl. Assembly of Berkeley, held at the Univ. of California, Berkeley, Calif., in 1963, published in Gentbrugge, Belgium, 1963.

The foregoing hypothesis is attractive from the hydrologic and engineering viewpoint. It permits the use of climatological observations to estimate the end results of complex physical, chemical, and biological processes. The hypothesis appears to have a logical physical basis but, at the present time (1965), it lacks adequate experimental verification. As supporting evidence, Bouchet[4] presented annual evaporation data from regions of equatorial rain forest where the assumed equality of regional and potential evaporation permits both to be expressed as simple functions of the insolation. However, these data are not sufficiently numerous nor from regions with sufficient climatic diversity to provide adequate confirmation of the hypothesis.

The investigation described herein had two main objectives. The first of these was the formulation of a method for computing regional evaporation from climatological observations, based on a synthesis of the concepts of Ferguson[3] and Bouchet.[4] The formulation is presented in the sections on "Vapor Transfer," "Energy Balance," "Mass and Energy Transfer," and "Potential Evaporation and Regional Evaporation." The other main objective was a test of the method by (1) a comparison of computed potential evaporation, one of its principal components, with observed pan evaporation, and (2) a comparison of computed regional evaporation, its end product, with river basin evaporation derived from rainfall and runoff records. The comparisons were made with monthly data drawn objectively from regions of considerable geographic and climatic diversity within Canada. The tests are presented in the sections on "Pan Evaporation" and "River Basin Evaporation." The results of the investigation lead to the following conclusions: (1) Changes in regional evaporation resulting from changes in the availability of water may be detected by their effects on the potential evaporation; and (2) the supporting evidence justifies the use of the absorbed insolation less computed potential evaporation as a working hypothesis for estimating regional evaporation from climatological observations in studies of river basin yield or in studies of water requirements for large areas.

Notation.—The symbols adopted for use in this paper are defined where they first appear and are arranged alphabetically in the Appendix.

MASS TRANSFER

The evaporation from a moist surface may be estimated from a simple mass transfer equation of the type

$$E = \frac{f(u)}{p} (e_S - e_D) \quad \ldots\ldots\ldots\ldots\ldots\ldots (1)$$

in which E = evaporation, e_S = saturation vapor pressure at evaporating surface temperature, e_D = the saturation vapor pressure at air dew point temperature, p = standard atmospheric pressure at appropriate elevation, and f(u) = empiric wind function in evaporation units.

There is some doubt as to whether the standard atmospheric pressure, p, should appear directly in Eq. 1. The cogency of the doubt is difficult to assess because the effects are small and the experimental evidence is inconclusive.

A wide variety of empiric wind functions have been suggested for use in equations such as Eq. 1. This variety may be attributed to (a) the dependence of the wind function on the surface over which the wind passes, (b) the variation

TABLE 1.—LOCATION OF EVAPORATION PANS AND MONTHLY DISTRIBUTION OF OBSERVATIONS

Station	Latitude North	Longitude West	Elevation, in feet	Number of Observations, April, 1962 - Sept., 1964									
				Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Total
Truro A.	45° 22'	63° 16'	131			2	2	2	3	3	1		13
Knob Lake A.	54° 48'	66° 49'	1,681				2	3	3	1			9
Shipshaw	48° 26'	71° 12'	75			1	1	1	1	1			5
Ste Anne de Bellevue	45° 25'	73° 56'	90			1		1	1				3
Ormstown	45° 07'	74° 02'	135				1	1	1				3
Ottawa N.R.C.	45° 27'	75° 37'	320			1	1	1	1	1			5
Ottawa C.D.A.	45° 24'	75° 43'	260			2	3	3	3	3	2		16
Guelph O.A.C.	43° 32'	80° 15'	1,095			2	3	3	3	3	2		16
Fullarton	43° 23'	81° 12'	1,100							1			1
Harrow C.D.A.	42° 02'	82° 53'	625			2	2	3	2	3			12
Cameron Falls	49° 09'	88° 23'	750				1	1	1	2			5
Churchill A.	58° 45'	94° 04'	115					1	1				2
Resolute A.	74° 43'	94° 59'	209					1	1				2
Indian Bay	49° 37'	95° 12'	1,072			1	1	2	3	3	2		12
Gimli W.R.B.	50° 38'	97° 02'	725			2	2	2	3	3	2		14
Winnipeg Int. A.	49° 54'	97° 14'	786			2	3	3	3	3	1		15
Morden C.D.A.	49° 11'	98° 06'	992			2	2	2	2	2	1		11
Baldur	49° 19'	99° 20'	1,400			2	1	2	1	1			7
Riding Mountain Park	50° 42'	99° 41'	2,481			3	3	3	3	3			15
Estevan A.	49° 04'	103° 00'	1,884			1	1	3	2	2			9
Indian Head Forestry	50° 31'	103° 41'	1,919			2	3	3	3	3			14
Weyburn	49° 40'	103° 51'	1,860		2	3	3	3	3	2	1		17
Regina A.	50° 26'	104° 40'	1,884			2	2	2	2	3			11
Glen Moray	50° 46'	104° 55'	1,625				1	2	1	1			5
Holdfast	50° 27'	105° 20'	1,788			1	2	1	1	1			6
Imperial	51° 22'	105° 30'	1,700			2	3	1	1	1			8
Saskatoon S.R.C.	52° 08'	106° 38'	1,690			1	1	1	1	1			5
Outlook P.F.R.A.	51° 29'	107° 03'	1,774				1	2	2	2			7
Swift Current C.D.A. South	50° 16'	107° 44'	2,707			3	3	3	3	3			15
Swift Current C.D.A. S.R.L.	50° 17'	107° 45'	2,499			1	1	1	2	1			6
Vauxhall C.D.A.	50° 03'	112° 08'	2,555		2	2	3	3	3	3	2		18
Calgary A.	51° 06'	114° 01'	3,540						1	1			2
Mount Eisenhower	51° 15'	115° 57'	4,675				3	3	2	1			9
Beaverlodge C.D.A.	55° 11'	119° 22'	2,500			2	3	3	3	3			14
Summerland C.D.A.	49° 34'	119° 39'	1,491		2	2	2	1	1				8
Hudson Hope Portage Mtn.	56° 03'	122° 09'	2,260			2	3	3	3	3			14
McLeod Lake Compressor	55° 02'	123° 02'	2,300						1				1
Vancouver U.B.C.	49° 16'	123° 15'	305	2	3	3	3	3	3	3	2	2	24
Nanaimo Water Reservoir	49° 09'	123° 58'	375					1	1	1			3
Comox A.	49° 43'	124° 54'	75	1	2	2	2	3	3	2	1	1	17
Norman Wells A.	65° 17'	126° 48'	209				1	1	1				3
Fort Selkirk	62° 49'	137° 22'	1,490				1	1	1	1			4
Haines Junction	60° 45'	137° 35'	2,030				1	1	1				3
Total				3	11	52	71	80	81	71	17	3	389

of wind velocity with height of measurement and the difficulties of converting wind velocities at one height to wind velocities at another height and (c) the poor quality of evaporation, surface temperature, and dew point temperature data that must be used in deriving the function.

The wind function used in the investigation was derived from monthly data for United States Weather Bureau Class A evaporation pans[5] for the period from April, 1962, to September, 1964, inclusive. All published values were used with the exception of those for which no water surface temperatures or wind speeds were available. Table 1 summarizes the location and altitude of the pans and the monthly distribution of the observations. The locations of the

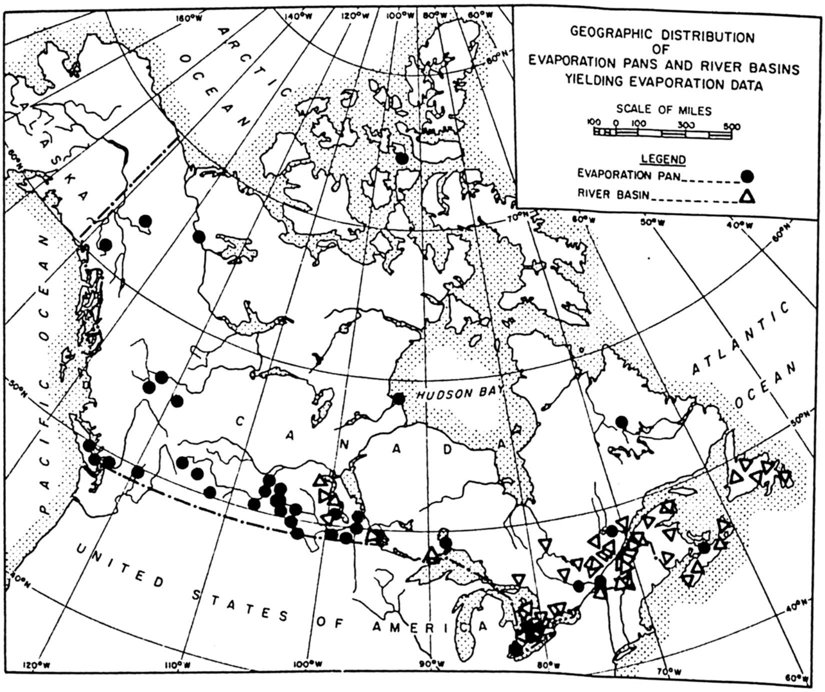

FIG. 1

pans are shown in Fig. 1. The individual monthly wind function values were computed by rearranging Eq. 1 and inserting the recorded evaporation, the appropriate standard atmospheric pressure, and the saturation vapor pressures at both the recorded water temperature and the dew point temperature recorded at the nearest psychrometric station.

Values of the computed wind function are shown plotted against the corresponding wind speeds in Fig. 2. Six outlying points have not been plotted. The scatter of the points partly explains the wide variety of wind function equations

5 "Monthly Records of Meteorological Observations in Canada," Meteorological Branch, Dept. of Transport, Toronto, Canada.

because many different, equally valid relationships could be drawn through the points. The line shown on Fig. 2 was chosen visually to combine the best fit with convenience of computation. It is represented by

$$f(u) = 500\left(0.6 + \frac{u}{100}\right) \qquad (2)$$

in which f(u) = wind function, in inches of water per month, and u = wind speed at pan rim, in miles per day.

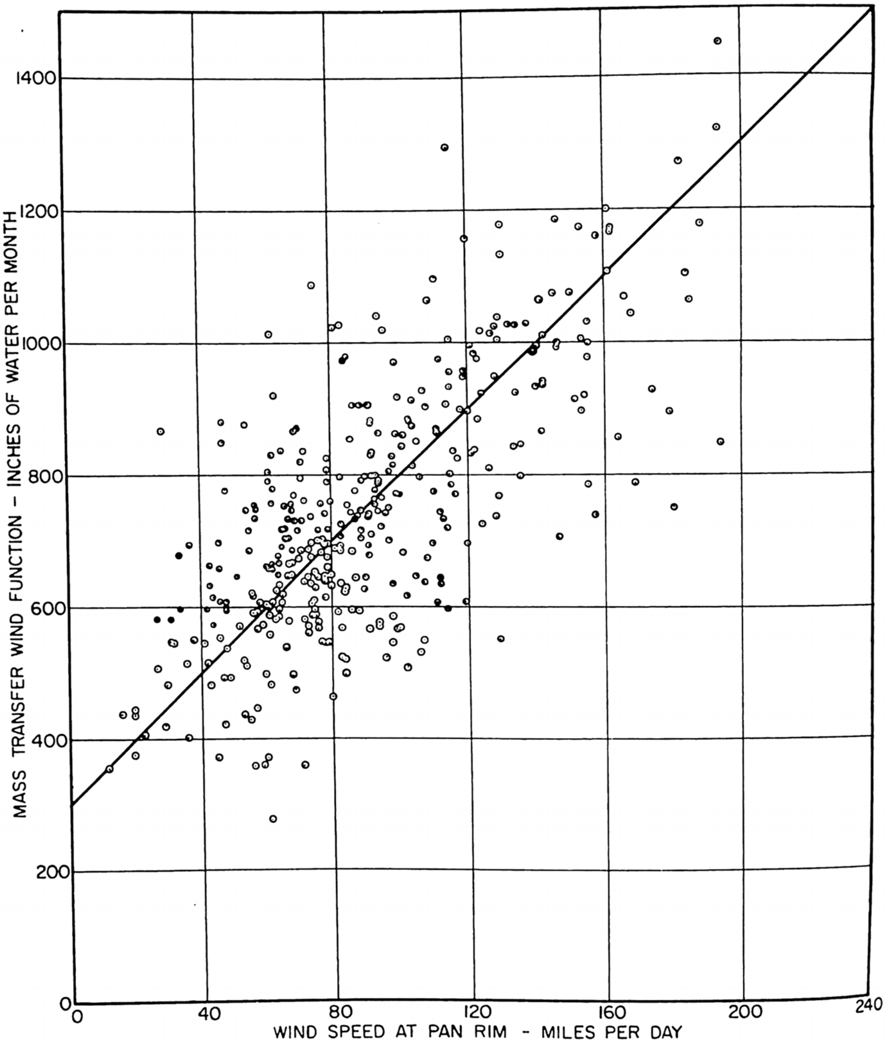

FIG. 2.—MASS TRANSFER WIND FUNCTION

The pan data used in the preparation of Fig. 2 provide some indication that variations in standard atmospheric pressure, i. e., elevation, affect evaporation to the extent shown in Eq. 1. However, the evidence is inconclusive.

Fig. 3 has been prepared to compare the recorded values of pan evaporation with the values computed by the mass transfer approach combining Eqs. 1 and 2. The data used in Fig. 3 include all evaporation data published with water

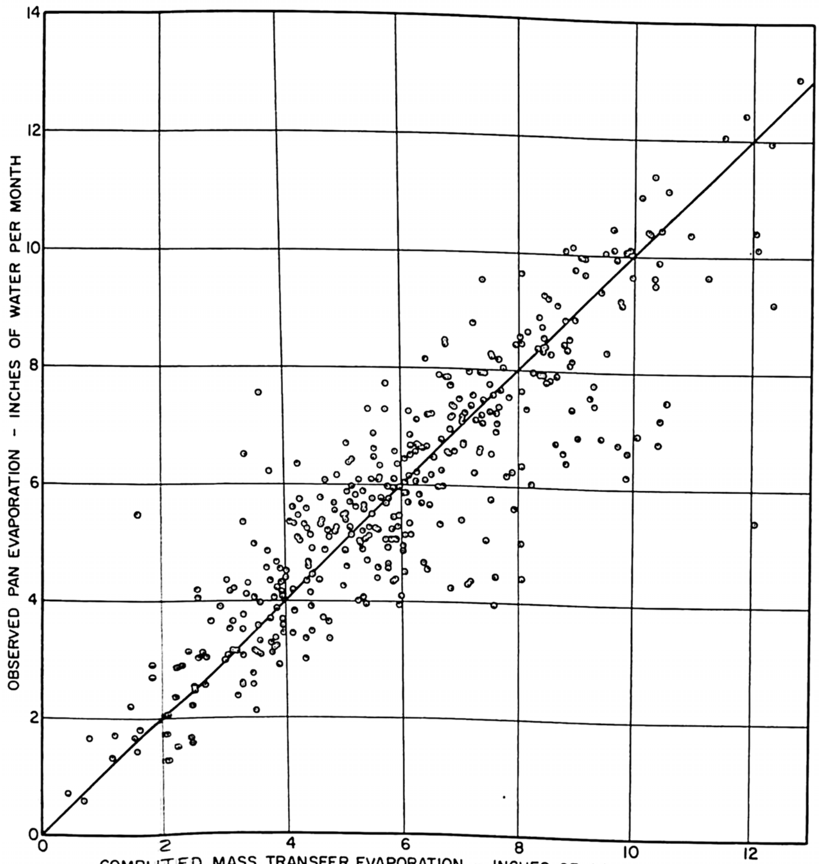

FIG. 3.—OBSERVED PAN EVAPORATION COMPARED WITH COMPUTED MASS TRANSFER EVAPORATION

surface temperatures and wind speeds for the period from April, 1962, to September, 1964, inclusive. The line on Fig. 3 is the line of equivalence. The comparison indicates that the mass transfer approach using observed surface water temperatures provides inaccurate results.

As Eq. 2 is based on wind speeds at the pan rim it is of interest to observe how these compare with the wind speeds published for climatological stations. Fig. 4 shows the monthly average wind speeds at hourly weather reporting stations throughout Canada plotted against the monthly average wind speeds at the pan rim wherever and whenever the values were recorded[5] for stations with reasonable geographic proximity during the period from April, 1962 to September, 1964, inclusive. Three outlying points have not been plotted. The relationship combining best fit and convenience may be expressed as u = 10 w, in which w = wind speed at the hourly weather reporting station, in

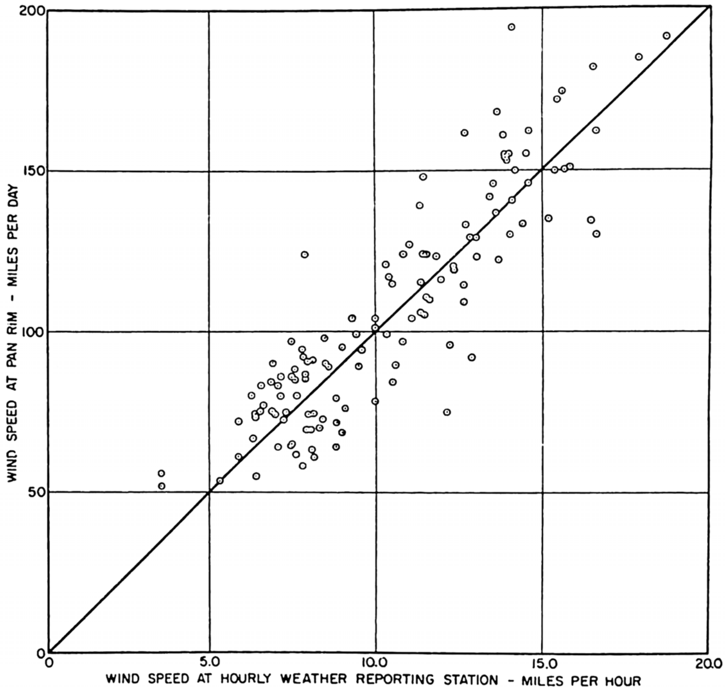

FIG. 4.—WIND SPEED AT PAN RIM COMPARED WITH WIND SPEED AT HOURLY WEATHER REPORTING STATION

miles per hour. Therefore, an adaptation of Eq. 2, for use with wind speeds recorded at hourly weather reporting stations is

$$f(u) = 500\left(0.6 + \frac{w}{10}\right) \quad (3)$$

ENERGY BALANCE

The energy balance for an evaporating surface is

$$E = (1-a)G - B - K \quad (4)$$

in which G = incident insolation, expressed in evaporation units, a = albedo of the surface, B = radiant heat transfer to the sky, in evaporation units, and K = sensible heat transfer to the air, in evaporation units.

Eq. 4 does not include terms for snow melt or heat storage below the evaporating surface. Therefore, results based on Eq. 4 do not apply to conditions where these factors are significant, such as evaporation for a short period of time, from deep lakes, or from a snow or ice covered surface.

The incident insolation, G, is observed at some climatological stations. Because the areal density of such observations is usually inadequate for general hydrologic purposes, it is common practice to relate the incident insolation to the extra-atmospheric insolation and the ratio of possible sunshine duration, i. e., the ratio of the duration of bright sunshine to the maximum possible duration of bright sunshine. The extra-atmospheric insolation is shown as a simple function of latitude and season for the latitudes of Canada in Table 2. Table 2 has been derived from the Smithsonian Meteorological

TABLE 2.—EXTRA-ATMOSPHERIC INSOLATION IN INCHES OF WATER EVAPORATED PER MONTH

Month	Latitude			
	40° N	50° N	60° N	70° N
January	7.4	4.6	1.7	--
February	9.2	6.6	3.8	1.4
March	13.6	11.2	8.4	5.5
April	16.2	15.0	13.0	11.0
May	19.4	18.8	17.8	17.0
June	19.8	19.7	19.4	19.9
July	19.9	19.6	19.0	19.1
August	18.0	17.0	15.5	13.9
September	14.4	12.4	10.2	7.5
October	11.4	8.6	5.9	2.9
November	8.0	5.2	2.5	0.2
December	6.8	3.8	1.1	--

Tables[6] and is based on a solar constant of 1.94 cal per sq cm per min, and a heat of vaporization of water of 590 cal per g.

Fig. 5 has been prepared to show the ratio of incident insolation to extra-atmospheric insolation as a function of the ratio of possible sunshine duration. The plotted points were derived from all published data[5] at six widely scattered climatological stations for the months May to October, inclusive of 1961, 1962, and 1963, and the months, May to September, inclusive, of 1964. The line drawn through the data visually is represented by

$$\frac{G}{G_O} = 1.17\left(0.18 + 0.55\,\frac{S}{S_O}\right) \quad \ldots\ldots\ldots\ldots\ldots (5)$$

in which G_O = the extra-atmospheric insolation, and S/S_O = ratio of possible sunshine duration (the ratio of the duration of bright sunshine to the maximum

6 "Smithsonian Meteorological Tables, Sixth Revised Edition," Smithsonian Institution, Washington, D. C., 1958, Table 132.

possible duration of bright sunshine). Eq. 5 has been expressed in the foregoing manner to show that the constants of 0.18 and 0.55 in the brackets are the same as those used by Penman.[2]

The absorbed insolation, (1-a) G, requires an estimate of the albedo, a, of the evaporating surface. Table 3 shows the seasonal and latitudinal variation of the albedo of a water surface in the latitudes of Canada as given by M. I.

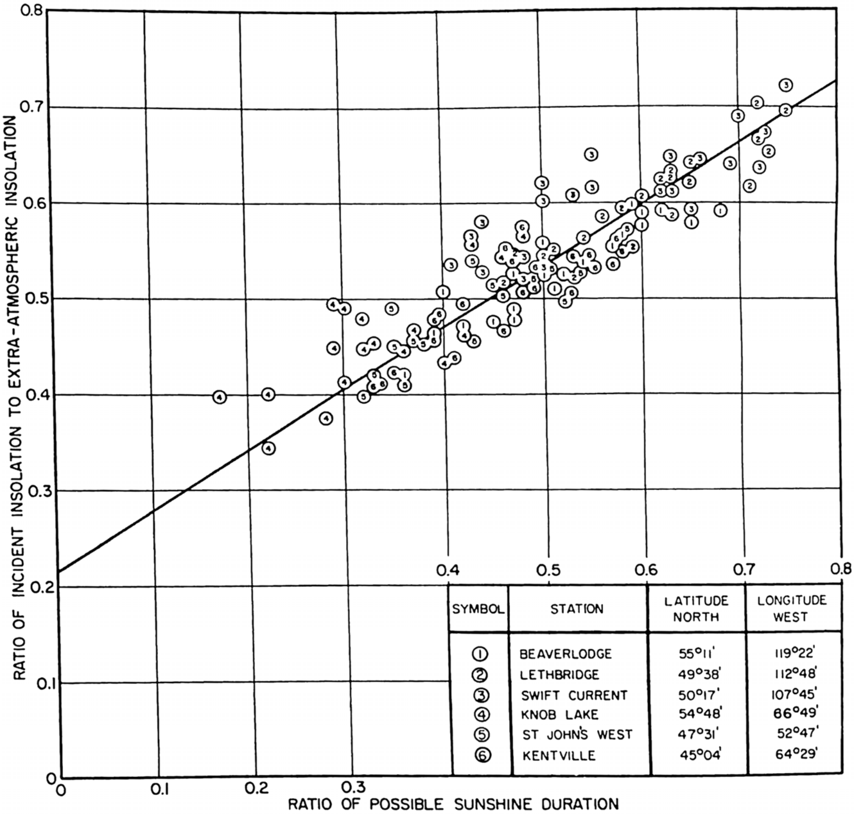

FIG. 5.—INCIDENT INSOLATION AS A FUNCTION OF EXTRA-ATMOSPHERIC INSOLATION AND SUNSHINE DURATION

Budyko.[7] The albedo for the lake, soil and vegetation surfaces of a region also has seasonal variations. Because quantitative estimates of such variations are scarce and unreliable, the albedo for a region during the snow-free months is assumed to be constant at 0.15.

The radiant heat transfer to the sky is not measured directly. It is equal to the long wave radiation from the evaporating surface less the long wave

7 Budyko, M. I., "The Heat Balance of the Earth's Surface," translated by N. A. Stepanova, U. S. Dept. of Commerce, Washington, D. C., 1958.

radiation returned to the surface from the atmosphere. The former can be estimated from Stefan's law and the latter can be measured. However, the instrumentation required for such measurements is expensive and complex so that it is usually necessary to estimate the atmospheric radiation from empirical relationships.

The customary method of estimating radiant heat transfer to the sky is to assume that the evaporating surface and the atmosphere radiate at air temperature--the former with an emissivity near unity and the latter with an empirically derived emissivity dependent on water vapor content and cloud cover. This approach has the following disadvantages:

1. The radiation from the evaporating surface and the radiation from the atmosphere are large quantities and the difference between them is small. Therefore, small errors in the assumed emissivity of the evaporating surface

TABLE 3.—ALBEDO OF WATER SURFACE

Month	Latitude			
	40° N	50° N	60° N	70° N
January	0.11	0.16	0.20	--
February	0.09	0.12	0.16	0.23
March	0.08	0.09	0.11	0.16
April	0.07	0.07	0.08	0.11
May	0.06	0.07	0.08	0.09
June	0.06	0.06	0.07	0.09
July	0.06	0.07	0.08	0.09
August	0.06	0.07	0.09	0.10
September	0.07	0.08	0.10	0.13
October	0.08	0.11	0.14	0.15
November	0.11	0.14	0.19	--
December	0.12	0.16	0.21	--

or the empirically derived emissivity of the atmosphere can lead to large errors in the computed radiant heat transfer to the sky.

2. It is assumed that the evaporating surface temperature is equal to the air temperature. The difference between the two temperatures is variable and can be significant in the computation of radiation.

3. The equations are cumbersome and difficult to compute. It may be concluded from the foregoing considerations that the customary method of computing radiant heat transfer to the sky produces a value of doubtful accuracy with considerable labor. As an alternative, it may be assumed that the evaporating surface radiates at the surface temperature with an emissivity near unity and that the atmosphere radiates at some equivalent emitting temperature with an emissivity near unity. According to this assumption,

$$B = \sigma \, (T_S + T_C)^4 - \sigma \, (T_E + T_C)^4 \quad \ldots\ldots\ldots\ldots (6)$$

in which σ = Stefan's constant, T_S = evaporating surface temperature, T_E = equivalent atmospheric emitting temperature, and T_C = absolute temperature adjustment (459.4°F).

In the temperature range experienced on evaporating surfaces, the variation of long wave radiation with temperature is almost linear. Therefore, Eq. 6 may be simplified to

$$B = \beta (T_S - T_E) \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (7)$$

in which β = radiant heat transfer coefficient approximately equal to 0.125 in. of water evaporated per month per degree Fahrenheit.

It is known that the equivalent atmospheric emitting temperature, T_E, should vary with the dew point temperature, the air temperature, and the extent of cloud. A simple but reasonable assumption, used in this investigation, is that T_E is equal to the air dew point temperature, T_D, or the average minimum air temperature, T_L, whichever is greater. Under this assumption, the atmosphere would radiate at the dew point temperature in calm, humid weather, and at the average minimum air temperature in windy, arid weather. Although unverified, the use of Eq. 7 with this assumption is justified on the grounds that it is simple, logical, permits the variation of surface termperature to be taken into account, and provides results that are sufficiently accurate for the purposes of this investigation.

An equation similar to Eq. 7 was used by H. Johnsson,[8] in his studies of the heat balance of Lake Klammingen, and by G. P. Williams,[9] in his studies of heat transfer coefficients for the water surface of McKay Lake. However, in both investigations, the equivalent atmospheric emitting temperature was assumed to be equal to the air temperature over the lake.

The sensible heat transfer from the evaporating surface to the air is estimated from

$$K = f(u)\ \gamma\ (T_S - T_A) \quad \ldots\ldots\ldots\ldots\ldots\ldots \quad (8)$$

in which γ = sensible heat transfer coefficient approximately equal to 0.00034 per degree Fahrenheit, and T_A = average daily air temperature.

It is usually assumed that the wind function for sensible heat transfer is equal to the wind function for transfer. Therefore, f(u) may be derived from Eq. 2 or Eq. 3.

MASS AND ENERGY TRANSFER

Both the mass transfer and energy balance approaches to the computation of evaporation require observations of the evaporating surface temperature. This seriously limits their utility because it is difficult to observe surface temperatures with any degree of accuracy. However, it has been shown by Penman[2] that it is possible to combine the mass transfer and energy balance approaches and thereby eliminate the need for surface temperature measurements. The result is

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8 Johnsson, H., "Termisk-Hydrologiska Studier i sjön Klämmingen," (Thermic-Hydrological Studies of Lake Klammingen), Geographiska Annaler, Stockholm, Sweden, Vol. 28, 1946.

9 Williams, G. P., "Heat Transfer Coefficients for Natural Water Surfaces," Publication No. 62, Internatl. Assn. for Scientific Hydrology, Genl. Assembly of Berkeley, held at the Univ. of Calif., Berkeley, Calif., in 1963, published in Gentbrugge, Belgium, 1963.

$$E = \frac{(1-a)G - B + E_O \dfrac{\gamma p}{\Delta}}{1 + \dfrac{\gamma p}{\Delta}} \qquad (9)$$

in which $E_O = f(u)/p\,(e_A - e_D)$, $\Delta = (e_S - e_A)/(T_S - T_A)$, and e_A = the saturation vapor pressure at average daily air temperature.

Eq. 9 may be modified to eliminate the need for surface temperature measurements in any relationship for radiant heat transfer and generalized for application to evaporation pans. However, it does have the following disadvantages:

1. The term Δ cannot be evaluated correctly because the evaporating surface temperature is unknown. It is customary to assume that Δ is equal to the slope of the saturation vapor pressure curve at the average daily air temperature. This assumption may lead to error if there is a significant difference between the surface and air temperatures.
2. With the use of such artificial terms as E_O and Δ, it is difficult to perceive the physical significance of variations in the climatological observations used in solving the equation.
3. The solution of the equation is cumbersome and complicated.

The first disadvantage may be eliminated and the last two disadvantages mitigated to some extent by using another approach to solve the problem. This approach has been credited to J. Ferguson by Bonython.[3] The mathematical development of the approach, as modified to incorporate Eq. 7 and generalized for application to evaporation pans, is presented below.

Eq. 1, for mass tranfer, and Eq. 4, for energy balance, may be combined in a manner applicable to both natural water surfaces and pans; the result is

$$\frac{f(u)}{p}(e_S - e_D) = k(1-a)G - m\,B - n\,K \qquad (10)$$

in which k = ratio of the insolation absorbing area to the evaporating area, m = ratio of the radiant heat transfer area to the evaporating area, and n = ratio of the sensible heat transfer area to the evaporating area.

The radiant heat transfer from the evaporating surface is to the sky and may be estimated from Eq. 7. However, the radiant heat transfer from the nonevaporating surface is with the surrounding objects and these are assumed to be radiating at the average daily air temperature. Therefore, the total radiant heat transfer may be estimated from

$$m\,B = \beta\,(T_S - T_E) + (m-1)\,\beta\,(T_S - T_A) \qquad (11)$$

The solution of Eqs. 8, 10, and 11 is

$$e_S + p\left(\frac{m\,\beta}{f(u)} + n\,\gamma\right)T_S = \frac{p\,k\,(1-a)G}{f(u)} + \frac{p\,\beta\,T_E}{f(u)} + \frac{p(m-1)\,\beta\,T_A}{f(u)} + p\,n\,\gamma\,T_A + e_D \qquad (12)$$

The five terms on the right side of Eq. 12 may be derived from climatological observations. This is not possible for the terms on the left side which depend on T_S. However, because e_S is a single valued function of T_S, e_S is a definable function of the left side of Eq. 12 and the heat transfer term $p[(m\ \beta/f(u)) + n\ \gamma]$. This functional relationship has been used in conjunction with Eq. 12 to prepare Fig. 6, in which, the ordinate is e_S, the saturation vapor pressure at the evaporating surface temperature, the abscissa is either the left side or the right side of Eq. 12, and each curve is for a different value of the heat transfer term. Because the abscissa is equal to the left side of Eq. 12, both the abscissa and ordinate may be derived from saturation vapor pressure tables for specific values of the heat transfer term and evaporating surface temperature. Therefore, the preparation of Fig. 6 is relatively simple. Because the abscissa is also equal to the right side of Eq. 12, both the abscissa and heat transfer term may be computed from climatological observations.

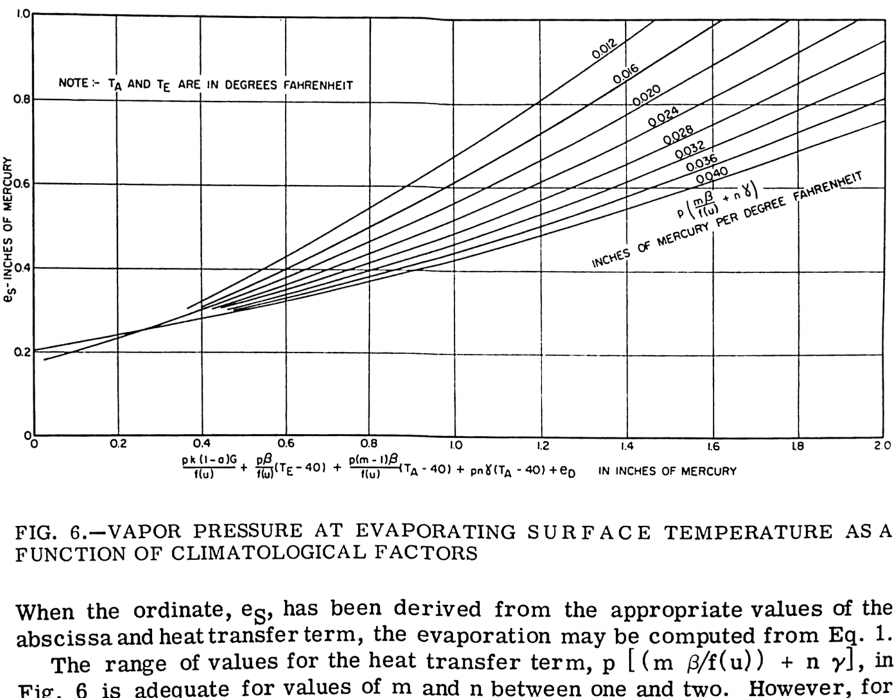

FIG. 6.—VAPOR PRESSURE AT EVAPORATING SURFACE TEMPERATURE AS A FUNCTION OF CLIMATOLOGICAL FACTORS

When the ordinate, e_S, has been derived from the appropriate values of the abscissa and heat transfer term, the evaporation may be computed from Eq. 1.

The range of values for the heat transfer term, $p[(m\ \beta/f(u)) + n\ \gamma]$, in Fig. 6 is adequate for values of m and n between one and two. However, for the normal case of evaporation from a natural water surface, where m and n are both equal to one, only those values of the heat transfer term less than 0.02 in. of mercury per degree Fahrenheit are required.

The effects on the computed evaporation of variations or errors in the climatological observations may be assessed from consideration of Eq. 1, the abscissa of Fig. 6, and the effect of the heat transfer term in Fig. 6.

1. According to Eq. 1, an increase in the air dew point temperature, e_D, results in a decrease in evaporation. However, this decrease is largely counteracted by the resultant increase in the saturation vapor pressure at the evaporating surface temperature, e_S, as indicated by the abscissa of Fig. 6.

2. In accordance with Eq. 1, an increase in the wind speed, u, results in an increase in evaporation. However, this increase is partly counteracted by a net decrease in e_S caused by a decrease indicated by the abscissa of Fig. 6, and an increase indicated by the effect of the heat transfer term in Fig. 6.

3. An increase in air temperature, T_A, or an increase in absorbed insolation, $(1-a)G$, results in an increase in e_S as indicated by the abscissa of Fig. 6. As there are no counteracting tendencies in either Eq. 1 or in the effect of the heat transfer term in Fig. 6, such an increase would result in a direct increase in evaporation.

From the considerations outlined above it is apparent that the importance of the climatological factors used in the mass and energy transfer approach varies inversely with the difficulty of measuring or estimating them accurately. Thus, the average daily air temperature and absorbed insolation, the two most important factors, can be measured or estimated easily, whereas the air dew point temperature, the least important of the factors, is difficult to measure. This characteristic is one of the principal advantages of the mass and energy transfer approach because areal coverage of observations usually varies with facility of measurement.

POTENTIAL EVAPORATION AND REGIONAL EVAPORATION

Fig. 6 and Eq. 1 are applicable to the computation of evaporation from a moist surface if the climatological observations reflect the weather directly over the surface. When, as is usual, the climatological observations reflect the regional weather, the computed evaporation is equal to the actual evaporation only when the moist area is so small that the evaporation has no appreciable influence on the air passing over it, or so large that it covers the surface of the region. Therefore, the use of Fig. 6 and Eq. 1 is usually limited to (1) the computation of potential evaporation which, by definition, meets the former requirement, and (2) the computation of evaporation from a region when the soil and vegetation surfaces are moist.

When the availability of water to the surfaces of a region is limited by soil moisture conditions or vegetative processes, Eq. 1 for vapor transfer is not applicable, although Eq. 4 for the energy balance remains valid. Bouchet[4] presented an analysis of the regional energy balance that indicated that variations in the regional evaporation caused by variations in the availability of water are reflected in the potential evaporation and therefore may be computed from climatological observations. A modified version of this analysis is presented below.

As the soil and vegetation surfaces of a region change from a humid condition to a less humid condition, the energy balances for both the region as a whole and for some small moist surface in the region change according to

$$E_R - E_{RH} = (1-a)\ (G - G_H) - (B_R - B_{RH}) - (K_R - K_{RH}) \quad . . (13)$$

and

$$E_P - E_{PH} = (1-a)\ (G - G_H) - (B_P - B_{PH}) - (K_P - K_{PH}) \quad . . (14)$$

in which the subscript R refers to the surface of the region as a whole, i. e., to regional evaporation, the subscript P refers to some small moist surface in the region, i. e., to potential evaporation, and the subscript H refers to a condition of regional humidity. It should be noted that the existence of a small moist area is not essential to the analysis as it would be too small to have any appreciable affect on the air passing over it.

The change in regional energy balance shown in Eq. 13 changes the energy content of the air over the region. This change is equal to the change in the sensible heat transfer from the surface of the region to the air plus any change in the quantity of energy brought into the region by large air mass movements. It is reasonable to assume that as the air passes over the small moist surface, the change in energy content per unit of area causes an equal and opposite change in the sensible heat transfer from the small moist surface to the air. This assumption is expressed mathematically as

$$K_P - K_{PH} = -(K_R - K_{RH}) - (W - W_H) \quad \text{. . . . . (15)}$$

in which W = energy brought into a region by large air mass movements, in evaporation units.

When the soil and vegetation surfaces of a region are moist, the regional evaporation is equal to the evaporation from a small moist surface in the region, i. e., $E_{RH} = E_{PH}$. Combining this equality, and the solution of Eqs. 13, 14, and 15 yields

$$E_R + E_P = 2\,E_{PH} + (1\text{-}a)\;(G\text{-}G_H) + (W\text{-}W_H) + \left[(1\text{-}a)G\text{-}B_R\text{-}B_P\right] - \left[(1\text{-}a)\,G_H\text{-}B_{RH}\text{-}B_{PH}\right] \quad \text{. . . . . (16)}$$

The quantity $2\,E_{PH}$ is independent of changing regional moisture conditions and the net effect of all the other terms on the right side of Eq. 16 is small. Therefore, the maximum value of the potential evaporation occurs when the region is completely arid and the regional evaporation is zero. That is,

$$E_{PI} = 2\,E_{PH} + (1\text{-}a)\;(G_I\text{-}G_H) + (W_I\text{-}W_H) + \left[(1\text{-}a)\;G_I\text{-}B_{RI}\text{-}B_{PI}\right] - \left[(1\text{-}a)\,G_H\text{-}B_{RH}\text{-}B_{PH}\right] \quad \text{. . . . . (17)}$$

in which the subscript I refers to conditions of regional aridity.

Under arid conditions, the sensible heat transfer would be from the air to the small moist surface, thereby compensating for the radiant heat transfer from the small moist surface to the sky. Therefore, it is reasonable to assume that the maximum evaporation from the small moist surface, i. e., the maximum potential evaporation, is equal to the gross energy supply to the region, or

$$E_{PI} = (1\text{-}a)\,G_I + W_I \quad \text{. . . . . (18)}$$

Combining Eqs. 16, 17, and 18 yields

$$E_R = (1\text{-}a)\,G - E_P + W - \left[(1\text{-}a)\,G_I\text{-}B_{RI}\text{-}B_{PI}\right] + \left[(1\text{-}a)G\text{-}B_R\text{-}B_P\right] \quad \text{. . (19)}$$

In order to use Eq. 19 it is necessary to make the following assumptions:

1. The energy brought into a region by large air mass movements is negligible. The quantity is small compared to the insolation, and the successive

additions and subtractions of energy tend to average out over a period of time. Therefore, any error resulting from the assumption decreases as the length of time considered increases.

2. The absorbed insolation less the radiant heat transfer to the sky from both the regional surface and the small moist surface, i. e., $(1\text{-}a)\,G\text{-}B_R\text{-}B_P$, remains constant under changing regional moisture conditions. A decrease in regional evaporation caused by a decrease in regional moisture results in (a) an increase in the surface temperature of the region, (b) a decrease in the surface temperature of the small moist surface, (c) an increase in the average daily air temperature, (d) a decrease in atmospheric humidity, and (e) a decrease in cloud quantity. The first two changes tend to stabilize the sum of the radiation from the surface of the region and the radiation from the small moist surface whereas the last three changes tend to stabilize the sum of the radiation from the atmosphere and the radiation from the sun.

With these two assumptions Eq. 19 is simplified and becomes

$$E_R = (1\text{-}a)\,G - E_P \quad \ldots\ldots\ldots\ldots\ldots (20)$$

If assumption 1 is incorrect, the errors resulting from the use of Eq. 20 would tend to be seasonal in maritime climates and random in continental climates. If assumption 2 is incorrect, the errors should have some functional relationship with the regional evaporation computed from Eq. 20.

In Eq. 20, the potential evaporation, E_P, is the evaporation from a small moist surface such as an evaporation pan or an irrigated lysimeter. To provide an estimate of the potential evaporation of the region, the pan or lysimeter should be designed to prevent radiant and sensible heat transfer through the walls and bottom, and the evaporation measurements should be adjusted for the difference between the albedo of the moist surface and the albedo of the region. If no pan or lysimeter records are available, the potential evaporation may be computed from climatological records using Fig. 6 and Eq. 1.

PAN EVAPORATION

In the investigation presented herein, the potential evaporation was computed from climatological observations using Fig. 6 and Eq. 1. Therefore, it was considered necessary to make a systematic test of the procedure using potential evaporation data from different climatic regions. Because the evaporation from a United States Weather Bureau Class A land pan has no appreciable influence on the air passing over it, it may be classified as a special case of potential evaporation. Therefore, to make the test, monthly values of Class A pan evaporation,[5] for the period from April, 1962, to September, 1964, inclusive (see Table 1 and Fig. 1), were compared with the values of potential evaporation derived from Fig. 6 and Eq. 1. The pan evaporation records used in the test included all published data, with the exception of those for which no surface temperature or wind speed records were available. Therefore, the results are comparable with those computed by the mass transfer approach and plotted in Fig. 3.

The evaporation from a Class A pan is a special case of potential evaporation because the exposed metal surface of the pan absorbs insolation, transfers radiant heat to and from the surrounding objects, and transfers sensible heat to and from the air. Provision has been made for the incorporation of these effects into Fig. 6 through the use of the area ratios k, m, and n.

The ratios of metal area in contact with the air to the water surface area for a Class A pan are shown in Table 4. The pan wall above the water line and the pan bottom between the wooden supports are assumed to be only 40% effective for sensible heat transfer because the former does not have the full temperature gradient acting on it and the latter is not fully exposed to the wind. Under this assumption, the area of sensible heat transfer is double the area of the evaporating surface and the value of the ratio n is two.

Because of the lack of any reliable information on the subject, it is assumed that the effective area of radiant heat transfer is equal to the area of sensible heat transfer. Under this assumption, the area of radiant heat transfer is double the evaporating surface area and the value of ratio m is two.

M. A. Kohler et al.[10] found that the best agreement with the Lake Hefner pan data was achieved when the heat transfer term was equal to 0.000871 p in. of mercury per degree Fahrenheit. This value does not allow for the variation of radiant heat transfer with surface water temperature. The heat transfer term used in this study, $p\left[(2\beta/f(u)) + 2\gamma\right]$, is equal to 0.000871 p when the

TABLE 4

	Ratio to Water Surface Area	
	Actual	Effective
Pan wall below water line	0.63	0.63
Pan wall above water line	0.42	0.17
Pan bottom between wooden supports	0.50	0.20
Total	1.55	1.00

wind speed at the pan rim is 200 miles per day and is 14% greater when the wind speed has a more usual value of 100 miles per day.

When m = 2 and n = 2, a value of k = 1.08 gives the best agreement between observed pan evaporation and computed potential evaporation. Therefore it is assumed that the insolation absorbing area of a Class A pan is 8% greater than the evaporating area.

The adaptation of the mass and energy transfer approach to pan evaporation is dependent, to a large extent, on a priori reasoning because of the complexity of the adaptation and the relative insensitivity of evaporation to changes in m and n.

The selection of data from climatological records[5] and the methods of processing them for use in Fig. 6 and Eq. 1 are outlined below.

1. The incident insolation was computed from Eq. 5 using the ratio of possible sunshine duration at the nearest station where observations were

10 Kohler, M. A., Nordenson, T. J., and Fox, W. E., "Evaporation from Pans and Lakes," Research Paper No. 38, Weather Bur., U. S. Dept. of Commerce, Washington, D. C., May, 1955.

made. The absorbed insolation was based on the values of albedo for water tabulated in the section on the energy balance.

2. The average daily air temperature, T_A, and the average minimum air temperature, T_L, were taken from records for the evaporation station, or when these did not exist, from records for the nearest recording station.

3. The wind speed, u, was taken from records for the evaporation station and the wind function, f(u), was computed from Eq. 2.

4. The air dew point temperature, T_D, was taken from records of the nearest hygrometric station.

5. The standard atmospheric pressure, p, was derived from standard pressure tables using the elevation of the evaporation station.

The observed pan evaporation is shown plotted against the computed potential evaporation together with the line of equivalence in Fig. 7. The observations are the same as those used for the mass transfer approach in Fig. 3. From a comparison it may be concluded that the mass and energy transfer

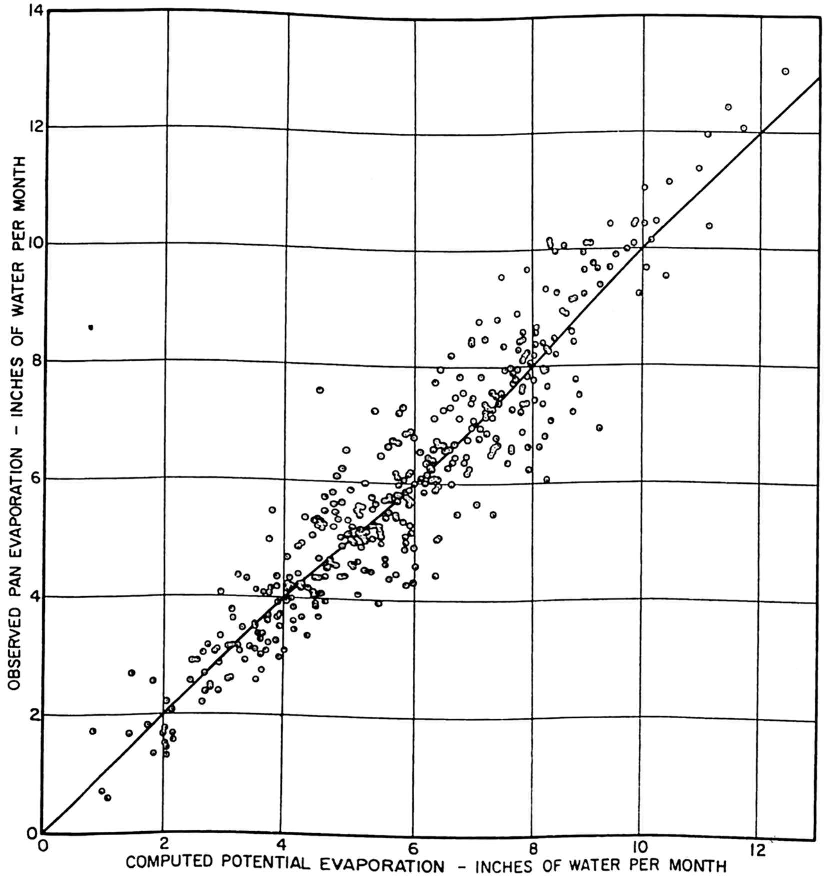

FIG. 7.—OBSERVED PAN EVAPORATION COMPARED WITH COMPUTED POTENTIAL EVAPORATION

approach used in Fig. 7 provides a better estimate of pan evaporation than the mass transfer approach used in Fig. 3.

An analysis of the differences between observed pan evaporation and computed potential evaporation indicates that there may be a slight seasonal pattern with the computed values being too low in the spring and too high in the fall. This is the only apparent nonrandom effect and appears to be the result of increase in albedo as the season advances. Such an increase in albedo could be caused by algal growth in the water.

In assessing the results of the mass and energy approach to potential evaporation, it should be noted that the test observations shown in Fig. 7 include all data published in conjunction with surface temperature and wind speed observations, with no attempt made to eliminate inaccurate observations. When consideration is given to the errors inherent in the measurement of pan evaporation and of the climatological data used in computing potential evaporation, the degree of agreement between the observed and potential evaporation is encouraging. Therefore, it may be concluded that Fig. 6 and Eq. 1 provide an adequate estimate of potential evaporation.

RIVER BASIN EVAPORATION

The relationship between regional and potential evaporation expressed in Eq. 20 was tested with monthly values of regional evaporation derived from river flow and rainfall records. The selection of data was based on certain river flow characteristics indicating that the evaporation during the month was equal to the rainfall less the runoff. These characteristics were sought in all available flow records[11,12,13] of medium-sized, unregulated rivers located in the central and eastern regions of Canada during the snow-free months for the period beginning October 1, 1949. The geographical, seasonal, and period-of-record limitations to the search outlined in the preceding sentence are as follows:

1. A medium-sized river has a drainage area between 80 sq miles and 3,500 sq miles.

2. An unregulated river has no relatively large, systematically operated storage reservoirs in the drainage basin. A river with relatively small reservoirs that are irregularly operated for logging or recreation purposes is considered to be unregulated.

3. The western edge of the central and eastern regions of Canada follows the Red River north to the Assiniboine, the Assiniboine River northwest to the 102nd meridian (the Manitoba-Saskatchewan border), and then follows the 102nd meridian north. It was considered that river basins to the west of this line would yield too little data to be worth searching.

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11 "Surface Water Data for Atlantic Drainage," Water Resources Branch, Dept. of Northern Affairs and Natl. Resources, Ottawa, Canada (published biennially).

12 "Surface Water Data for St. Lawrence and Southern Hudson Bay Drainage," Water Resources Branch, Dept. of Northern Affairs and Natl. Resources, Ottawa, Canada (published biennially prior to September 30, 1957 and annually thereafter).

13 "Surface Water Data for Arctic and Western Hudson Bay Drainage," Water Resources Branch, Dept. of Northern Affairs and Natl. Resources, Ottawa, Canada (published biennially prior to September 30, 1957 and annually thereafter).

4. The snow-free months are from June to October, inclusive, in forested or barren areas and from May to October, inclusive, in partly cultivated areas. In Ontario, to the south and west of the Trent canal system, the snow-free period extends from April to October, inclusive, provided that there is no snowmelt flood during the first part of April. A snowmelt flood is arbitrarily defined as a flood causing the average flow of the first ten days of April to exceed the average flow of the first day of April by at least 50%.

5. The period of record searched ended September 30, 1962, for all rivers tributary to the Atlantic Ocean and James Bay and ended September 30, 1961, for all rivers tributary to the western shore of Hudson Bay.

Monthly evaporation is equal to the difference between rainfall and runoff only if the stored water and the soil moisture in the river basin are the same at the beginning and end of the month. Because systematic records of these quantities are not available, the equality must be assessed from certain characteristics of flow records. To provide the objectivity that is essential for a test of this type, the selection of data was based on rigid criteria formulated in advance as a means of recognizing the flow characteristics signifying equality of stored water and soil moisture. Such criteria represent a compromise between hydrologic considerations applicable to a wide range of geographic conditions and the desirability of having a large number of data. The hydrologic considerations are as follows:

1. Equality of stored water at the beginning and end of the month should occur if the flows at the beginning and end of the month are on a recession curve and are equal.

2. Equality of soil moisture at the beginning and end of the month can be recognized from flow records only if the soil is primed. Such recognition is based on the occurrence of either a gradual increase in flow, denoting storage recharge, or a rapid increase in flow, denoting surface runoff, during the days immediately preceding the beginning and end of the month.

The foregoing considerations are quite general and difficult to interpret with objectivity. Furthermore, if strictly interpreted, they would yield little data. The following numerical selection criteria were developed as a compromise between these considerations and the necessity for an adequate quantity of data:

1. During the month and the following month, the daily flows of the first days should not be more than 0.10 cu ft per sec per sq mile higher or 0.50 cu ft per sec per sq mile lower than the daily flows of the preceding days. Furthermore, if the daily flow on the second day of either or both of the two months exceeds the average flow of the two preceding days, the daily flow for the first day of that month must be equal to or less than the daily flow of the preceding day.

2. The average flow for the two days about the beginning of the month should be within 0.20 cu ft per sec per sq mile of the average flow for the two days about the end of the month.

3. The daily flows for each of the two days about the beginning of the month and each of the two days about the end of the month must be at least 0.50 cu ft per sec per sq mile.

TABLE 5.—EVAPORATION FROM RIVER BASINS

Reference No.	River	Year	Month	Data Derived from Observations: Temperatures of the Air, in degrees Fahrenheit: Average daily, T_A	Average minimum, T_L	Dew point, T_D	Wind speed, w, in miles per hour	Ratio of possible sunshine duration, $\frac{S}{S_O}$	Rainfall on river basin, in inches	Runoff from river basin, in inches	Computed Data, in inches: Evaporation from river basin	Absorbed insolation, 0.85 G	Potential evaporation, E_P	Absorbed insolation less potential evaporation, $0.85G - E_P$
(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)
1	Lewaseechjeech	1956	July	59.5	51.4	50.4	9.4	0.37	3.30	1.94	1.36	7.50	5.43	2.07
2	Indian	1961	July	59.6	49.6	51.9	9.6	0.41	4.10	1.34	2.76	7.94	5.46	2.48
3	Indian	1958	June	50.8	42.4	42.2	12.2	0.33	3.50	1.70	1.80	7.06	4.77	2.29
4	Rocky	1959	Oct.	44.7	36.8	39.4	15.9	0.31	2.70	1.71	0.99	3.24	2.30	0.94
5	Rocky	1956	Sept.	52.8	46.1	48.6	13.2	0.36	6.80	5.04	1.76	4.85	3.26	1.59
6	Rocky	1955	July	57.4	49.1	51.6	13.8	0.39	3.60	1.80	1.80	7.70	5.38	2.32
7	Piper's Hole	1959	Sept.	52.3	43.8	43.8	13.4	0.45	3.30	1.82	1.48	5.44	4.04	1.40
8	Bay du Nord	1959	Sept.	52.3	43.8	43.8	13.4	0.45	3.10	1.18	1.92	5.44	4.04	1.40
9	Bay du Nord	1958	July	59.1	51.2	50.2	12.7	0.37	4.25	2.10	2.15	7.50	5.79	1.71
10	Bay du Nord	1956	June	53.2	43.4	44.5	12.4	0.28	5.40	3.40	2.00	6.55	4.69	1.86
11	Bay du Nord	1952	June	55.0	46.2	48.2	11.2	0.34	4.50	2.16	2.34	7.20	4.81	2.39
12	N.E. Margaree	1959	Sept.	56.6	47.2	51.9	13.6	0.49	4.25	2.88	1.37	5.86	4.18	1.68
13	St. Mary's	1955	Sept.	54.1	41.9	50.2	6.9	0.53	3.49	0.68	2.81	6.23	3.66	2.57
14	St. Mary's	1953	Sept.	57.8	47.8	52.8	7.9	0.51	3.44	1.43	2.01	6.10	3.94	2.16
15	Medway	1953	Oct.	48.2	39.8	43.8	8.1	0.45	3.90	1.66	2.24	4.30	2.44	1.86
16	Medway	1956	Oct.	47.9	37.7	39.7	8.9	0.56	2.16	0.88	1.28	4.89	3.08	1.81
17	Roseway	1954	June	57.5	49.6	52.8	10.1	0.45	4.15	1.50	2.65	8.42	5.50	2.92
18	Shogomoc	1961	Aug.	65.1	52.0	55.3	8.1	0.58	2.25	0.42	1.83	8.63	6.35	2.28
19	Nepisiguit	1954	July	62.1	52.5	55.6	9.3	0.43	4.28	1.59	2.69	8.18	5.71	2.47
20	Nepisiguit	1961	Oct.	44.9	34.3	40.9	8.6	0.38	3.45	2.49	0.96	3.59	1.97	1.62
21	Nepisiguit	1959	July	68.8	53.3	58.3	10.5	0.56	2.11	0.88	1.23	9.56	7.60	1.96
22	St. Francis	1957	Oct.	44.1	33.6	36.8	10.2	0.30	2.50	0.94	1.56	3.18	2.01	1.17
23	York	1961	July	60.7	50.6	52.9	9.7	0.45	4.20	2.26	1.94	8.37	5.94	2.43
24	York	1955	Sept.	48.7	39.3	42.0	13.6	0.41	3.88	1.38	2.50	5.10	3.39	1.71
25	Dartmouth	1960	July	61.0	51.4	52.6	9.2	0.44	3.90	0.68	3.22	8.30	5.85	2.45
26	Dartmouth	1956	Oct.	43.6	35.1	33.6	12.6	0.49	1.60	0.94	0.66	3.97	2.89	1.08
27	Dartmouth	1951	Aug.	57.4	52.4	54.2	9.6	0.43	3.58	1.19	2.39	7.08	4.48	2.60
28	Rimouski	1953	July	62.6	51.8	53.4	11.6	0.52	3.70	1.06	2.64	8.74	6.65	2.09
29	Trois-Pistoles	1957	Oct.	43.3	34.0	36.9	10.2	0.30	2.84	1.06	1.78	3.12	1.95	1.17
30	Ouelle	1957	Oct.	42.3	33.4	36.4	10.2	0.33	2.62	1.10	1.52	3.37	1.86	1.51
31	Du Sud	1961	Sept.	59.9	50.8	53.0	9.4	0.48	4.43	2.02	2.41	5.78	4.18	1.60
32	Du Sud	1957	Oct.	43.9	34.8	36.8	9.4	0.31	3.02	1.91	1.11	3.31	1.99	1.32
33	Du Sud	1951	July	64.0	54.4	57.6	9.2	0.49	6.17	2.98	3.19	8.82	6.22	2.60
34	Etchemin	1960	June	61.0	50.5	50.0	9.9	0.58	3.84	1.78	2.06	9.76	7.09	2.67
35	Etchemin	1956	July	61.7	52.3	53.0	8.5	0.46	3.97	2.17	1.80	8.46	5.94	2.52
36	Etchemin	1951	Oct.	46.4	35.4	38.8	8.9	0.36	2.62	0.91	1.71	3.61	2.31	1.30
37	Beaurivage	1950	July	65.1	54.9	56.1	8.8	0.56	3.73	0.75	2.98	9.58	6.94	2.64
38	Becancour	1954	July	60.6	50.4	55.4	7.6	0.52	4.17	1.02	3.15	9.13	5.85	3.28
39	Becancour	1950	June	59.8	49.8	52.8	10.6	0.54	5.02	1.84	3.18	9.43	6.51	2.92
40	Watopeka	1961	July	66.0	54.2	57.6	7.8	0.48	4.30	1.03	3.27	8.69	6.24	2.45
41	Watopeka	1951	Oct.	45.9	35.7	39.5	9.4	0.42	2.04	0.78	1.26	4.01	2.42	1.59
42	Salmon	1954	June	60.5	50.6	52.4	7.2	0.38	6.40	3.53	2.87	7.65	5.15	2.50
43	Salmon	1952	June	61.4	50.6	52.2	9.9	0.56	5.44	3.36	2.08	9.60	6.86	2.74
44	Eaton	1952	June	61.2	50.6	52.2	9.9	0.56	5.06	2.50	2.56	9.60	6.84	2.76
45	Coaticook	1961	Aug.	62.8	52.1	56.2	8.1	0.49	4.32	0.99	3.33	7.85	5.36	2.49
46	Hall's Stream	1961	Aug.	62.8	52.1	56.2	8.1	0.49	4.20	0.75	3.45	7.85	5.36	2.49
47	Chateauguay	1954	June	63.4	54.8	57.0	8.0	0.42	4.93	1.49	3.44	8.11	5.61	2.50
48	Escoumains	1955	Aug.	61.3	50.6	54.4	8.8	0.45	2.80	0.51	2.29	7.26	5.10	2.16
49	Escoumains	1953	July	61.8	51.4	54.1	9.6	0.52	4.80	2.06	2.74	9.13	6.42	2.71
50	North Ste Anne	1952	July	67.0	52.6	59.6	8.5	0.54	4.05	2.16	1.89	9.35	6.72	2.63
51	North Ste Anne	1950	Sept.	49.6	36.6	44.6	9.2	0.47	2.74	0.95	1.79	5.66	3.33	2.33
52	Ste Anne	1950	Sept.	50.0	39.5	44.8	9.2	0.47	2.84	1.00	1.84	5.66	3.38	2.28
53	Batiscan	1961	July	65.7	55.4	55.3	10.3	0.41	4.00	2.48	1.52	7.95	6.29	1.66
54	Batiscan	1957	Oct.	42.2	32.2	35.9	9.5	0.33	3.58	2.16	1.42	3.38	1.92	1.46
55	Batiscan	1951	Oct.	44.1	34.3	38.9	9.6	0.34	2.80	1.85	0.95	3.42	1.97	1.45
56	Croche	1951	Sept.	53.2	42.0	46.6	8.8	0.42	4.09	1.67	2.42	5.24	3.44	1.80
57	L'Assomption	1961	July	64.2	58.2	54.8	8.3	0.46	3.98	1.92	2.06	8.47	6.17	2.30

TABLE 5.—*CONTINUED*

(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)
58	L'Assomption	1957	Oct.	43.0	30.4	35.6	8.8	0.32	2.66	1.46	1.20	3.41	2.00	1.41
59	L'Assomption	1952	July	69.1	58.0	61.1	9.0	0.58	4.20	1.18	3.02	9.77	7.24	2.53
60	L'Assomption	1961	July	64.7	52.5	55.3	8.2	0.45	3.61	1.59	2.02	8.38	6.12	2.26
61	L'Assomption	1958	July	63.4	51.7	55.2	8.5	0.48	6.05	2.91	3.14	8.70	6.17	2.53
62	Ouareau	1961	Sept.	59.1	46.1	51.8	9.4	0.44	2.20	0.83	1.37	5.46	3.98	1.48
63	Ouareau	1957	Oct.	44.0	29.9	35.6	8.8	0.32	2.58	1.27	1.31	3.40	2.12	1.28
64	Rouge	1957	Oct.	42.1	32.0	35.6	8.8	0.32	3.30	2.50	0.80	3.37	1.89	1.48
65	Coulonge	1961	Sept.	59.7	49.6	54.8	7.4	0.49	4.02	1.14	2.88	5.86	3.86	2.00
66	Kinojevis	1962	Aug.	60.4	48.8	52.7	6.8	0.56	4.58	0.85	3.73	8.30	5.41	2.89
67	Kinojevis	1949	Oct.	44.9	33.7	38.0	11.6	0.39	1.71	0.81	0.90	3.53	2.31	1.22
68	Ganaraska	1957	Oct.	47.5	37.8	40.6	10.4	0.42	2.76	0.85	1.91	4.17	2.68	1.49
69	Duffin	1950	Aug.	63.7	51.3	55.6	8.4	0.49	3.71	0.61	3.10	7.89	5.62	2.27
70	Credit	1954	April	43.4	33.2	36.2	10.4	0.38	4.30	2.01	2.29	6.11	3.42	2.69
71	Speed	1955	May	56.4	46.4	45.7	9.1	0.54	3.76	1.07	2.69	9.11	6.15	2.96
72	Conestogo	1955	May	56.4	46.4	45.7	9.1	0.53	3.88	0.59	3.29	9.01	6.10	2.91
73	Nith	1961	April	38.4	31.4	33.0	13.4	0.18	3.24	1.11	2.13	4.36	2.44	1.92
74	Big	1957	Oct.	48.0	37.5	40.8	9.8	0.49	2.30	0.72	1.58	4.68	2.95	1.73
75	Big Otter	1953	April	41.6	33.3	34.6	12.5	0.22	2.64	1.05	1.59	4.76	2.82	1.94
76	Thames	1953	April	41.8	33.4	34.5	12.5	0.26	2.65	1.05	1.60	5.10	2.98	2.12
77	North Thames	1954	April	44.7	34.6	38.6	12.5	0.36	4.80	3.04	1.76	5.93	3.47	2.46
78	Maitland	1961	April	38.4	31.2	32.5	13.4	0.16	2.69	1.60	1.09	4.21	2.45	1.76
79	Saugeen	1961	Oct.	51.0	41.4	44.1	12.0	0.43	1.71	0.44	1.27	4.22	3.06	1.16
80	Nottawasaga	1961	April	39.6	32.1	32.3	11.4	0.27	3.77	1.46	2.31	5.12	2.80	2.32
81	Aux Sable	1961	July	64.3	54.8	53.9	11.0	0.51	3.98	1.47	2.51	9.04	7.03	2.01
82	Shebandowan	1958	July	61.7	51.0	53.0	8.3	0.43	3.14	1.06	2.08	8.20	5.75	2.45
83	Whitemouth	1959	Sept.	53.3	43.2	45.9	11.4	0.44	3.01	0.81	2.20	5.23	3.68	1.55
84	Minnedosa	1960	May	53.5	41.4	35.8	12.6	0.64	3.12	0.54	2.58	9.93	7.61	2.32
85	Ochre	1959	May	48.3	37.7	33.8	15.1	0.48	3.05	1.05	2.00	8.30	6.11	2.19
86	Vermilion	1960	May	53.5	41.4	36.8	12.7	0.61	3.19	1.51	1.68	9.59	7.38	2.21
87	Woody	1954	June	56.5	46.6	45.2	10.8	0.40	4.90	2.46	2.44	7.83	5.79	2.04

4. The last five days of the month and the preceding month should include a daily flow at least 0.20 cu ft per sec per sq mile higher than the minimum daily flow of the five days immediately preceding the day on which it occurs. However, the foregoing requirement may be ignored for either the month or preceding month if the daily flow of the last day of the month is at least 0.10 cu ft per sec per sq mile higher than the minimum daily flow of the last five days of the month.

5. The tolerance in applying the foregoing criteria is 1 cu ft per sec or 0.005 cu ft per sec per sq mile, whichever is greater.

There is little doubt that selection based on the foregoing criteria will include data for which the basic requirement of equality of stored water and soil moisture at the beginning and end of the month does not apply, and will exclude data for which the basic requirement does apply. However, the criteria do provide an objective, physically sound method of selecting data which should generally meet the basic requirement.

By searching all river flow records[11,12,13] falling within the geographical, seasonal, and period-of-record limitations stated previously, it was found that 89 monthly flows met the criteria. Two of these flows occurred during months when the rainfall measurement coverage for the river basins was not only inadequate but almost nonexistent. The remaining 87 monthly flows, expressed in terms of basin yield, are shown together with the river name, the year, the month, and other relevant data in Table 5. The geographical

TABLE 6.—RIVER

River	Province	Reference No. in Table 5	Gage Location		Drainage area, in square miles
			Latitude N	Longitude W	
(1)	(2)	(3)	(4)	(5)	(6)
Lewaseechjeech	Newfoundland	1	48°37'	57°56'	180
Indian	Newfoundland	2-3	49°31'	56°07'	376
Rocky	Newfoundland	4-6	47°13'	53°34'	110
Piper's Hole	Newfoundland	7	47°57'	54°17'	300
Bay du Nord	Newfoundland	8-11	47°45'	55°26'	454
N. E. Margaree	Nova Scotia	12	46°22'	60°59'	142
St. Mary's	Nova Scotia	13-14	45°10'	61°59'	523
Medway	Nova Scotia	15	44°25'	65°03'	132
Medway	Nova Scotia	16	44°10'	64°40'	535
Roseway	Nova Scotia	17	43°50'	65°22'	191
Shogomoc	New Brunswick	18	45°57'	67°20'	91
Nepisiguit	New Brunswick	19	47°24'	65°48'	712
Nepisiguit	New Brunswick	20-21	47°30'	65°41'	807
St. Francis	New Brunswick	22	47°12'	68°57'	520
York	Quebec	23-24	58°50'	64°38'	389
Dartmouth	Quebec	25-27	48°56'	64°38'	288
Rimouski	Quebec	28	48°25'	68°33'	800
Trois-Pistoles	Quebec	29	48°05'	69°11'	387
Ouelle	Quebec	30	47°25'	69°57'	310
Du Sud	Quebec	31-33	46°50'	70°45'	311
Etchemin	Quebec	34-36	46°38'	71°03'	443
Beaurivage	Quebec	37	46°39'	71°17'	274
Becancour	Quebec	38-39	46°22'	71°37'	546
Watopeka	Quebec	40-41	45°35'	71°59'	127
Salmon	Quebec	42-43	45°37'	71°24'	329
Eaton	Quebec	44	45°28'	71°39'	250
Coaticook	Quebec	45	45°17'	71°54'	199
Halls Stream	Quebec	46	45°03'	71 30'	85
Chateauguay	Quebec	47	45°17'	73°48'	948
Escoumains	Quebec	48-49	48°20'	69°26'	354
North Ste Anne	Quebec	50-51	46°55'	71°52'	294
Ste Anne	Quebec	52	46°42'	72°05'	686
Batiscan	Quebec	53-55	46°33'	72°25'	1,750
Croche	Quebec	56	47 36'	72°44'	787
L'Assomption	Quebec	57-59	46°16'	73°46'	211
L'Assomption	Quebec	60-61	46°01'	73°26'	504
Ouareau	Quebec	62-63	46°02'	73°43'	492
Rouge	Quebec	64	46°21'	74°47'	997
Coulonge	Quebec	65	45°52'	76°41'	2,000
Kinojevis	Quebec	66-67	48°27'	78°22'	647
Ganaraska	Ontario	68	43°59'	78°18'	94
Duffin	Ontario	69	43°51'	79°04'	110
Credit	Ontario	70	43°33'	79°39'	320
Speed	Ontario	71	43°32'	80°16'	229
Conestogo	Ontario	72	43°31'	80°31'	317
Nith	Ontario	73	43°11'	80°27'	398
Big	Ontario	74	42°41'	80°32'	228
Big Otter	Ontario	75	42°41'	80°48'	269
Thames	Ontario	76	42°58'	81°13'	519
North Thames	Ontario	77	43°15'	81°09'	416
Maitland	Ontario	78	43°53'	81°20'	628
Saugeen	Ontario	79	44°27'	81°20'	1,570
Nottawasaga	Ontario	80	44°15'	79°49'	456
Aux Sable	Ontario	81	46°13'	82°04'	524
Shebandowan	Ontario	82	48°33'	89°41'	1,080
Whitemouth	Manitoba	83	49°56'	95°88'	1,450
Minnedosa	Manitoba	84	50°01'	100°13'	1,490
Ochre	Manitoba	85	51°04'	99°47'	202
Vermilion	Manitoba	86	51°11'	100°01'	261
Woody	Manitoba	87	52°16'	101°08'	821

BASIN CHARACTERISTICS

Physiographic Region				Bedrock		Soil Mantle		Relief		Surface Storage		Vegetation		
Appalachian	Great Lakes-St. Lawrence Lowland	Canadian Shield	Manitoba Cuesta	Igneous	Sedimentary	Shallow	Deep	Gentle	Rough	Low	High	Forest	Cultivated	Moss Covered Barren
(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)	(15)	(16)	(17)	(18)	(19)	(20)	(21)
X				X		X			X		X	X		X
X					X	X			X	X		X		
X					X	X			X	X		X		X
X					X	X		X			X	X		X
X					X	X			X		X	X		X
X				X		X			X	X		X		
X					X	X			X		X	X	X	
X				X		X			X		X	X		
X				X	X	X			X		X	X		
X				X	X	X			X		X	X		
X				X		X			X		X	X		
X				X		X			X	X		X		
X				X		X			X	X		X		
X					X	X			X		X	X		
X					X	X			X	X		X		
X					X	X			X	X		X	X	
X					X	X		X		X		X	X	
X					X		X	X		X		X	X	
X					X	X		X			X	X	X	
X					X	X			X	X		X	X	
X	X				X		X		X	X		X	X	
X	X				X		X	X		X		X	X	
X	X				X		X	X		X		X		
X	X				X		X	X		X		X	X	
X					X	X			X	X		X	X	
X					X	X			X	X		X	X	
X					X	X			X	X		X	X	
X					X	X			X	X		X	X	
	X	X			X		X	X		X		X		
		X		X		X			X		X	X		
		X		X		X			X	X		X		
		X		X		X			X	X		X		
		X		X		X			X		X	X		
		X		X		X			X		X	X		
		X		X	X	X			X		X	X	X	
		X		X	X	X			X		X	X		
		X		X		X			X		X	X		
		X		X		X			X		X	X		
		X		X		X			X		X	X		
		X		X		X		X			X	X		
	X				X		X	X		X		X	X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
	X				X		X	X		X			X	
		X		X		X			X		X	X		
		X		X		X		X			X	X		
		X		X		X		X			X	X	X	
			X		X		X	X			X	X	X	
			X		X		X		X	X		X	X	
			X		X		X		X	X		X	X	
			X		X		X		X	X		X	X	

distribution of the river basins is shown in Fig. 1 and brief descriptions of the salient characteristics of the river basins are provided in Table 6.

In Table 5, the evaporation from river basins is the difference between rainfall and runoff and is compared with the difference between absorbed insolation and potential evaporation. The data used in computing these quantities are also shown in Table 5. The selection of these data from climatological records[5] and the methods used in processing them are as follows:

1. The rainfall on the river basin was derived from isohyetal maps prepared from published rainfall records.
2. The ratio of possible sunshine duration for the river basin, S/S_O, was derived from the records of one, two, or three of the sunshine measuring stations in the vicinity of the river basin.
3. The average daily air temperature, T_A, and the average minimum air temperature, T_L, were derived from records at one, two, or three climatological stations in the vicinity of the river basin.
4. The air dew point temperature, T_D, was derived from the records of one or two of the nearest hygrometric stations in the vicinity of the river basin. The average dew point computed from station records was adjusted for the average difference in elevation between the air temperature measurement stations and the hygrometric stations. The adjustment was equal to a reduction of 1° F per 1,000 ft increase in elevation.
5. The wind speed, w, was derived from the records of one, two, or three hourly weather reporting stations in the vicinity of the river basin.
6. The absorbed insolation, 0.85 G, was derived from the incident insolation assuming an albedo of 0.15. The incident insolation was computed from Eq. 5.
7. The potential evaporation was derived from Fig. 6 and Eq. 1. In this derivation the wind function, f(u), was computed from Eq. 3, the atmospheric pressure was assumed to be constant at 28.5 in. of mercury (standard atmospheric pressure at an elevation of 1,200 ft above sea level), and the ratios of insolation absorbing area, radiant heat transfer area, and sensible heat transfer area to evaporating area, k, m, and n, were assumed to be unity.

Normally, the published data were adequate for objective estimates of the foregoing quantities. However, in some cases the data were so sparse or so variable that it was necessary to give weight in a subjective manner to such factors as topographic and normal climatic conditions. Such subjectivity was of most significance in its effect on the selection of isohyetal configuration and sunshine measuring stations.

Fig. 8 has been prepared to provide a graphical comparison between the river basin evaporation and the absorbed insolation less potential evaporation. The numbers inscribed in the plotted points are the reference numbers of Table 5 and the solid line is the line of equivalence. As there is no systematic functional divergence between the plotted points and the line of equivalence, it may be concluded that variations in the quantity $(1-a)G-B_R-B_P$ under different regional moisture conditions are either so small that they may be ignored, as assumed in the derivation of Eq. 20, or are compensated for by other erroneous assumptions.

An analysis of the deviations between the plotted points and the line of equivalence in Fig. 8 indicates that there is a seasonal pattern, with the river basin evaporation tending to be higher than the absorbed insolation less potential evaporation during August and September, and lower during April, May, and October. This may be the result of appreciable quantities of energy supplied with a definite seasonal pattern by large air mass movements. However, the most probable explanation is that there is a seasonal variation of albedo caused by color changes in the soil and vegetation surfaces and the occurrence of occasional snow storms during April, May, and October.

The agreement between river basin evaporation and absorbed insolation less potential evaporation is exceptionally good in view of the wide geographic

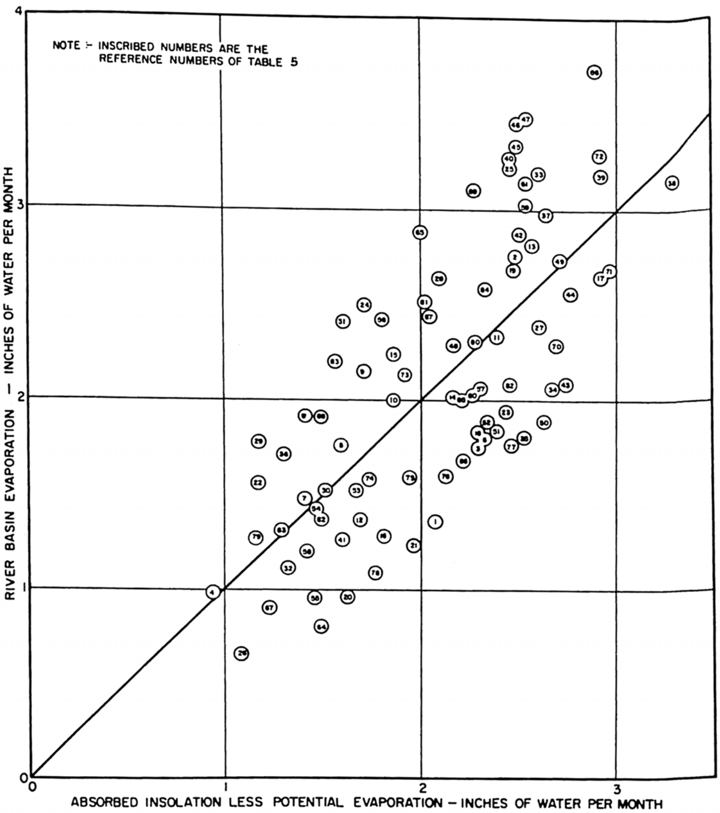

FIG. 8.—RIVER BASIN EVAPORATION COMPARED WITH ABSORBED INSOLATION LESS POTENTIAL EVAPORATION

range of the data, the inadequacy of the selection criteria, and the errors inherent in measuring or estimating the factors used in deriving the three quantities. The degree of agreement justifies the use of Eq. 20, in conjunction with Eq. 1 and Fig. 6, as a working hypothesis for the computation of regional evaporation from climatological observations.

CONCLUSIONS

According to Eq. 20, the sum of the regional and potential evaporation is equal to the absorbed insolation. The two controlling factors are the absorbed insolation and the availability of water. When the soil and vegetation surfaces of the region are moist, the regional evaporation is equal to the potential evaporation and both are equal to one half of the absorbed insolation. When soil moisture conditions or vegetative processes limit the availability of water, the regional evaporation decreases and the energy thus freed is transferred to the atmosphere as increased turbulence and heat, where it is reflected in an equivalent increase in potential evaporation. When no water is available, the regional evaporation becomes zero and the potential evaporation reaches its maximum value, which is equal to the absorbed insolation

It is apparent from the foregoing considerations that the potential evaporation is equal to the quantity of energy remaining from the absorbed insolation after energy has been used for regional evaporation. Furthermore, it is apparent that the difference between the potential evaporation and one half of the absorbed insolation is a measure of the quantity of water that is not evaporated because of soil moisture conditions or vegetative processes. This permits the regional evaporation, a result of complex physical, chemical, and biological processes, to be estimated by its effect on climatological observations. However, it must be emphasized that potential evaporation is an effect of, rather than a cause of, regional evaporation. Thus, increases in potential evaporation under conditions of constant gross energy supply, as reflected in increased air temperatures, increased wind speeds, and decreased dew point temperatures, do not cause increased regional evaporation, as is usually assumed; they are, in fact, caused by decreased regional evaporation.

Because the potential evaporation varies from 50% to 100% of the absorbed insolation, there is a tendency for the evaporation from pans and irrigated lysimeters to scatter about a line with a slope of 70% to 75% when plotted against the absorbed insolation. Damagnez et al.[14] have presented potential evaporation data from Tunisia illustrating this tendency. As a large shallow lake is a region with a moist surface, the evaporation from such a lake should be equal to 50% of the absorbed insolation, or 65% to 70% of the potential evaporation. This is near the accepted percentage ratio of lake evaporation to pan evaporation and provides a possible explanation for it.

The results of this investigation leave little doubt that changes in regional evaporation resulting from changes in the availability of water are reflected in the potential evaporation. Therefore, the foregoing examination of the implications of variations in the availability of water is, in its essentials, correct. Furthermore, the supporting evidence justifies the use of Eq. 20, in conjunction with Eq. 1 and Fig. 6, as a working hypothesis for estimating regional evaporation from climatological records in studies of river basin yield or in studies of water requirements for large areas. However, the quality and limited climatic diversity of the data do not permit a critical appraisal of the empirical relationships used in Eq. 1 and Fig. 6 and the

14 Damagnez, J., Riou, Ch., DeVillele, O., and ElAmmami, S., "Estimation et Mesure de l'Evapotranspiration Potentielle en Tunisie," Publication No. 62, Internatl. Assn. for Scientific Hydrology, Genl. Assembly of Berkeley, held at the Univ. of California, Berkeley, Calif., in 1963, published in Gentbrugge, Belgium, 1963.

assumptions used in the formulation of Eq. 20. The relationships requiring further refinement and the assumptions requiring further verification are, (1) the empiric wind function, (2) the equation for radiant heat transfer to the sky, (3) the equation for sensible heat transfer to the atmosphere, (4) the assumption that changes in the energy content of the air per unit of area result in equal and opposite changes in the sensible heat transfer from a small moist surface to the air, as expressed in Eq. 15, (5) the assumption that the maximum potential evaporation is equal to the gross energy supply to the region, as expressed in Eq. 18, (6) the assumption that the energy brought into the region by large air mass movements may be neglected, as expressed in the transition from Eq. 19 to Eq. 20, (7) the assumption that the absorbed insolation less the radiant heat transfer to the sky from both the regional surface and a small moist surface in the region, i. e., $(1-a)G-B_R-B_P$, remains constant under changing regional moisture conditions, as expressed in the transition from Eq. 19 to Eq. 20. Refinements to the three empirical relationships depend on progress in instrumentation. However, verification of the four assumptions may proceed independently.

Further verification of Eq. 20 requires accurate estimates of potential evaporation and regional evaporation. The potential evaporation may be computed from climatological observations, in a manner similar to that described herein, although the accuracy of these computations will be uncertain until better estimates of radiant heat transfer, sensible heat transfer, and the wind function are available. These sources of uncertainty may be eliminated if a small moist surface, such as a pan or lysimeter, designed to provide good accuracy and to prevent radiant and sensible heat transfer through the wall and bottom, is used to measure potential evaporation. Regional evaporation may be derived from river basin rainfall and runoff data, in a manner similar to that presented herein, or from the water budgets of large swamps or shallow lakes. When runoff records are too short to provide adequate regional evaporation data, they may provide a check of Eq. 20 when compared with runoff that has been synthesized from the rainfall and the computed regional evaporation (the absorbed insolation less potential evaporation) using a method similar to that proposed by J. C. I. Dooge.[15,16]

Verification techniques may vary under different climatic conditions. Arid regions with regional evaporation equal to zero permit an evaluation of the assumption that the maximum potential evaporation equals the gross energy supply to the region, i. e., the absorbed insolation. In a continuously humid region, such as would exist in large swamps and shallow lakes, or might exist in large bogs and the equatorial rain forests, the validity of Eq. 20 requires that both the potential evaporation and the regional evaporation be equal to one half the absorbed insolation. Thus, values of either potential or regional evaporation may be used to check the validity of Eq. 20 and provide a sensitive indication of the variability of the quantity $(1-a)\ G-B_R-B_P$ under changing regional moisture conditions. In regions that are neither humid nor arid, such as those that yielded data for this investigation, the verification of Eq. 20

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15 Dooge, J. C. I., "The Routing of Groundwater Recharge through Typical Elements of Linear Storage," Publication No. 52, Internatl. Assn. of Scientific Hydrology, Genl. Assembly of Helsinki, held in Helsinki, Finland, in 1960, published in Gentbrugge, Belgium, 1960.

16 Dooge, J. C. I., "Indirect Estimation of Low Flows," Proceedings, Institution of Civ. Engrs., Ireland, Dublin, Ireland, March, 1961.

requires values of both potential and regional evaporation. Although this requirement complicates the verification process, it is likely that the bulk of supporting evidence will continue to be found in such regions.

The type of research mentioned previously stresses further verification. However, considerable work is required to extend the scope of this investigation. The conclusions presented herein do not apply to conditions where heat storage or snow melt are of significance, such as evaporation for a short period of time, from deep lakes or from ice and snow covered surfaces. The modification of the relationship between regional and potential evaporation for application to such conditions should give rise to many interesting and complex problems.

APPENDIX.—NOTATION

The following symbols have been adopted for use in this paper:

a = albedo of surface;
B = radiant heat transfer to sky in evaporation units;
B_P = B from small moist surface;
B_{PH} = B_P under conditions of regional humidity;
B_{PI} = B_P under conditions of regional aridity;
B_R = B from region;
B_{RH} = B_R under conditions of regional humidity;
B_{RI} = B_R under conditions of regional aridity;
E = evaporation;
E_O = $f(u)/p\,(e_A - e_D)$;
E_P = potential evaporation;
E_{PH} = E_P under conditions of regional humidity;
E_{PI} = E_P under conditions of regional aridity;
E_R = regional evaporation;
E_{RH} = E_R under conditions of regional humidity;
E_{RI} = E_R under conditions of regional aridity;
e_A = saturation vapor pressure at average daily temperature of air;
e_D = saturation vapor pressure at air dew point temperature;
e_S = saturation vapor pressure at evaporating surface temperature;
$f(u)$ = empiric wind function in evaporation units;
G = incident insolation in evaporation units;
G_H = G under conditions of regional humidity;
G_I = G under conditions of regional aridity;
G_O = extra-atmospheric insolation in evaporation units;
K = sensible heat transfer to air in evaporation units;
K_P = K from small moist surface;
K_{PH} = K_P under conditions of regional humidity;
K_R = K from region;
K_{RH} = K_R under conditions of regional humidity;
k = ratio of insolation absorbing area to evaporating area;
m = ratio of radiant heat transfer area to evaporating area;

n = ratio of sensible heat transfer area to evaporating area;
p = standard atmospheric pressure at appropriate elevation;
S/S_O = ratio of possible sunshine duration (ratio of the duration of bright sunshine to maximum possible duration of bright sunshine);
T_A = average daily air temperature;
T_C = absolute temperature adjustment = 459.4°F;
T_D = air dew point temperature;
T_E = equivalent atmospheric emitting temperature;
T_L = average minimum air temperature;
T_S = evaporating surface temperature;
u = wind speed at pan rim, in miles per day;
W = energy brought into region by large air mass movements, in evaporation units;
W_H = W under conditions of regional humidity;
W_I = W under conditions of regional aridity;
w = wind speed at hourly weather reporting station, in miles per hour;
β = radiant heat transfer coefficient = 0.125 in. of water evaporated per month per degree Fahrenheit;
γ = sensible heat transfer coefficient = 0.00034 per degree Fahrenheit;
$\Delta = (e_S - e_A)/(T_S - T_A)$; and
σ = Stefan's constant.