EVAPORATION OVER PARTS OF EUROPE

by

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Summary

Weather records from 31 meteorological stations in Europe are used to calculate the potential evaporation rate from neighbouring land surfaces and thence the estimated actual evaporation rate after correcting for shortage of summer rainfall. In northern Europe special consideration has had to be given to periods when snow covers the ground. Annual values range from about 20 cm/year in northern Norway to 60 cm/year in Italy and the Balkans: from west to east there is little change in total between Britain and Russia near Moscow, though the seasonal distribution differs markedly. A general check has been obtained from mean annual values of rainfall and run-off from 44 catchment areas—not so well distributed as the weather stations—and the agreement is usually within 10% in comparable environments.

* * *

1. Introduction

The treatment previously applied with some success to the British Isles (Penman 1950) has been extended to parts of Europe for which relevant weather and rainfall data could be obtained with reasonable ease. For many reasons the expected degree of precision is not as great as for the British Isles, but the attempt seemed worth while because (i) it gives, even if only approximately, a broad picture of the distribution of annual evaporation that is of some climatic interest; (ii) it gives a chance. of testing in more extreme climates the utility of the unavoidable empirical expressions based upon British experience; (iii) it demonstrates the importance of summer rain in determining how near actual evaporation will approach potential evaporation; (iv) it introduces a new physical aspect provided by the existence of continuous snow cover for several months in more northerly regions.

The basis of the paper is the same as that of its predecessor, namely the comparison of a theoretical mean annual evaporation, based on weather data, with an observed mean annual evaporation based on the difference between rainfall and run-off for catchment areas. The theoretical estimate is calculated from mean monthly values of air temperature, air humidty, wind speed, and duration of bright sunshine: to correct for deficiency of summer rainfall the amount has to be known; and a measure of snowfall has to be available for some regions.

2. Theoretical basis

Partly for completeness, and partly to make clear what new assumptions are made, a brief outline of basic theory follows. The estimated actual evaporation, E, is derived from an estimated potential evaporation, E_T, in turn derived from an estimated evaporation, E_O, from a hypothetical open water surface exposed to the prevailing weather. From consideration of energy balance, the heat budget as income is

$$\begin{aligned} H &= R_I - R_B \\ &= (1 - r)\, R_C - R_B, \end{aligned} \qquad (1)$$

where R_I is the net short-wave income, reduced by reflection from its incident value, R_C, and R_B is the net long-wave back radiation. The heat budget as expenditure is

$$H = E + K + M + S, \qquad (2)$$

where E is evaporation, K is sensible heat transfer to the air, M is energy used in melting of snow and S is sensible heat transfer to the soil. For the moment, M and S will be ignored. The value of H is determinable from standard weather data, and hence

$$H = E + K. \qquad (2a)$$

From consideration of turbulent transport, the evaporation rate is

$$E = f(u)\ (e_s - e_d), \qquad (3)$$

and the sensible heat transport is

$$K = \gamma f(u)\ (T_s - T_a). \qquad (4)$$

This assumes that $f(u)$ is the same for vapour transfer as for heat transfer. In equations (3) and (4) $f(u)$ is a function of wind speed at a known height, e_s and T_s are vapour pressure and temperature at the surface, and e_d and T_a are the vapour pressure and temperature in the air at screen height. It is convenient to define a new quantity, E_a, by

$$E_a = f(u)\ (e_a - e_d), \qquad (5)$$

where e_a is the saturation vapour pressure at mean air temperature; and also to introduce another parameter $\Delta = de/dT \doteq (e_s - e_a)/(T_s - T_a)$, i.e. the slope of the saturation vapour pressure curve near T_a.

Normally the surface temperature and vapour pressure are unknown and unknowable, but for open water they are uniquely related; and can be eliminated from (2a), (3), (4) and (5) to give

$$E_o = \frac{\Delta H + \gamma E_a}{\Delta + \gamma}. \qquad (6)$$

The only weather data needed to evaluate E_o to an adequate degree of approximation are mean air temperature, humidity, wind speed and duration of bright sunshine.

If the terms M and S are retained in equation (2a) the expression for E_o cannot be obtained uniquely. The difficulties can be adequately overcome, as follows.

3. The effect of heat storage

The stored heat goes through an annual cycle with a mid-February minimum and a mid-August maximum, and assuming a constant diffusivity (with time and depth) the amplitude of the variation is $\sqrt{2}\ \rho c\ \theta_o \sqrt{\varkappa T/4\pi}$, where ρc is the specific heat per unit volume, θ_o is the amplitude of the surface temperature, $\varkappa$ is the thermal diffusivity of soil, and T is the period (1 year) in seconds. Represensative values are : $\rho c = 0.50$, $\theta_0 = 10°C.$, and $\varkappa = 4 \times 10^{-3}$ cm²/sec., and in evaporation units the amplitude of the heat content is equivalent to 13.5 mm of evaporation. The effect is not very large and is primarily a phase shift delaying the time of maximum evaporation. It can be distributed sinusoidally throughout the year, but in aggregate it means that total estimated evaporation between mid-February and mid-August will be too big by about 27 mm and between mid-August and mid-February will be too small by the same amount.

For water it is difficult to get a corresponding estimate. The specific heat is now unity, but the heating is not confined to the surface and a turbulent diffusivity must be used. The effect may well be several times greater than for soil, and so winter minima for large stretches of deep water will be greater than calculated from equation (6), and the summer maxima will be less than calculated, and may be out of phase by one or more months. A quantitative demonstration of the effect appears in paragraph 9.

4. The effect of snow

The effect of snow is three-fold. First, it has a very high reflection coefficient, known to change from about 0.7 when fresh to about 0.18 when old (Angstrom, quoted from Sutton 1953), so that R_I has to be decreased in equation (1). Second, under snow the land is effectively an open water surface—there is no need for a correction for vegetation. Third, energy must be used to melt the snow, so reducing the amount available for evaporation.

To get an approximate solution of the snow problem the following assumptions are made:—

(i) In all months in which the mean air temperature is less than 0°C, all precipitation is snow having a uniform reflexion coefficient of $r = 0.63$ (a value chosen to be less than 0.7 and designed to ease the arithmetic because it makes $1 - r$ for snow = 0.4 times $1 - r$ for open water).

(ii) In the first spring month, with mean air temperature above 0°C, all precipitation is rain, the snow melts and during the melting period $r = 0.26$ (making $1 - r = 0.8$ times $1 - r$ for open water).

(iii) No melting occurs during the period $T_a < 0°C$.

(iv) During the melt period the surface temperature of the snow is 0°C, and hence evaporation during the melt period can be calculated from eq. (3).

The modified equation for the melt period is

$$E = \frac{\Delta (H - M) + \gamma E_a}{\Delta + \gamma} \qquad (7)$$

From (iv): $E = f(u)\,(e_o - e_d)$, where e_o is for 0°C. (3)

Hence, knowing H, E_a, $f(u)$ and e_d a value of M can be found, and knowing how much snow there is to melt (from (i)) the period of melt can be estimated: the remainder of the month will then be treated in the normal way. An example will illustrate. For Leningrad the mean precipitation from November to March ($T_a < 0°C$) is 144 mm. The energy needed to melt this would evaporate $\frac{80}{590} \times 144 = 19.5$ mm of water. For April, $e_d = 4.0$ mm Hg, $e_o = 4.6$ mm Hg, $f(u) = 0.70$, and hence $E_o = 0.70 \times 0.6 = 0.42$ mm/day. Also H (for $r = 0.26$) = 1.48, $E_o = 1.03$, $\Delta/\gamma = 1.8$; from which M = 1.8 mm/day in evaporation units. The total energy needed for melting (in the same units) is 19.5 mm: hence the melt period is 11 days, during which the evaporation is $11 \times 0.42 = 5$ mm. During the remaining 19 days of April equation (6) applies, and $E_o = 1.59$ mm/day.

5. Working equations

During a period of nine years in which the equations have been used it has very rarely proved possible to get a direct check on their accuracy. Circumstantial evidence has suggested that estimates of E_o and E_T (see below) may be slightly too big, and the recent report from Lake Hefner (U. S. Geological Survey 1952) has given an opportunity for re-assessing some of the empirical expressions used in the formulae. This re-assessment cannot be completed until the actual data from Lake Hefner are available, and at present only the formula will be given without explanation of changes. The net effect is that 'new' values of E_0 calculated for this paper are about 5 to 10% less than 'old' values

Eq. (5) Old $E_a = 0.35\,(1 + u_2/100)\,(e_a - e_d)$ mm/day

New $E_a = 0.35\,(0.5 + u_2/100)\,(e_a - e_d)$ mm/day

Eq. (1) Old $H = 0.95\,R_A\,(0.18 + 0.55\,n/N) - \sigma T^4\,(0.56 - 0.09\sqrt{e_d})\,(0.10 + 0.90\,n/N)$

New $H = 0.93\,R_A\,(0.20 + 0.48\,n/N) - 0.97\sigma T^4\,(0.47 - 0.077\sqrt{e_d})\,(0.20 + 0.80\,n/N)$

As an example. For Angers in June:—

Old $E_a = 4.65$; new $E_a = 3.87$ mm/day.

Old $H = 5.98$; new $H = 5.76$ mm/day.

Old $E_0 = 5.50$; new $E_0 = 5.11$ mm/day.

For the British Isles, where E_a is always much less than H in summer, the change in E_0 is less marked than this.

6. Potential transpiration

In principle, the value of the potential evaporation rate from a land surface can now be calculated from weather data without introducing a hypothetical open water surface (Penman 1953), but extra data are needed and can only be obtained in special circumstances. For the present purpose it is necessary to rely on the empirical conversion factors based on British experience, in the form

$$E_T = f \times E_0,$$

where the factor f has the values:—

May-August,	$f = 0.8$
Mar.-April. Sept.-Oct.,	$f = 0.7$
Nov.-February	$f = 0.6$.

It is known that one of the terms in f is a day-length factor which will lead to bigger summer values and smaller winter values in more northerly latitudes; and lead to a smaller annual range in more southerly latitudes. As it happens, however, small uncertainties in f are unimportant in most of the data for Europe because the correction for deficiency of summer rainfall has a much more marked effect in reducing estimates of actual evauoration.

7. Actual transpiration

One essential condition for evaporation is that there should be water to be evaporated: a potential rate of 1,000 mm a year has little meaning for a desert. As winter evaporation is usually small, the important rain is the summer rain, and where this is inadequate to meet the potential summer transpiration then the actual evaporation is less than the potential value E_T. In the early days of a dry period the vegetation can draw on stored water in the soil, but there is an upper limit to the amount that can be so withdrawn, a limit depending chiefly upon rooting depth of the vegetation and to some extent on soil type. For the present survey, British experience with grass will be extrapolated to the whole of Europe and the correction from potential transpiration to actual transpiration is made on the followging basis, working in months. Until $\Sigma(E_T - R)$ exceeds 75 mm then $E = E_T$: then begins a slow decline in which typical values are: — $\Sigma(E_T - R) = 95$, $\Sigma(E - R) = 94$; $\Sigma(E_T - R) = 125$, $\Sigma(E - R) = 109$; $\Sigma(E_T - R) = 165$, $\Sigma(E - R) = 114$; $\Sigma(E_T - R) = 225$, $\Sigma(E - R) = 120$. At this stage it is arbitrarily assumed that vegetation will no longer be transpiring and whatever higher value of $\Sigma(E_T - R)$ is reached, there will be no further increase in $\Sigma(E - R)$. This assumption has been used for the following places: — Angers, Athens, Lisbon, Odessa, Perpignan Rome. The most extreme places considered were Gursew and Madrid, but here the annual rainfall is so small that this sets an upper limit to possible evaporation.

8. Calculation of evaporation

Most of the mean monthly values of duration of bright sunshine, air temperature, air humidity and wind speed have been supplied by the Climatology Branch of the British Meteorological Office. Where the data were complete, they were never averages over the same period of years for all four elements: where they were incomplete various devices have been used to make good deficiencies, as follows.

Sunshine (n/N). Values for another observatory have been used, if not too far away. For example, the Upsaala estimate of E is based on n/N as measured at Stockholm.

Humidity (e_d). For a few places, only four values were given. The other eight per year were obtained by interpolation. For a few others e_d was obtained from mean air temperature and mean relative humidity.

Wind speed (u_2). Some guesses were needed. Fortunately a large error in u_2 has only a small effect on the final value of E_0. Where values were assumed, all months were given the same value, a value chosen in the light of known windspeeds at stations with a similar geographical exposure.

Table 1 gives the calculated mean annual values of evaporation for 31 places, E_T being the potential transpiration, and E the estimated actual transpiration.

TABLE 1

Mean annual precipitation and evaporation (mm)

E_T = potential evaporation: E = estimated actual evaporation allowing for deficiency of summer rain. The last four columns indicate how inadequate data have been made good:—
n/N, Sunshine figures from named place; e_d, either interpolated, or from relative humidity;
u_2, assumed mean wind speed at 2 metres, the same for all months; Rain, total from named place.
Russia 86, 73 and 25 are meteorological stations that have not been identified by place-names that can be found in an ordinary atlas.

Place	Lat. N.	Long.	P	E_T	E	n/N	e_d	u_2	Rain
Abisko	68 21	18 49E	1050	236	236	—	—	—	Tromso
Upsaala	59 51	17 38E	536	305	303	Stockholm	int.	—	—
Leningrad	59 56	30 16E	613	377	367	—	—	—	—
Dorpat	58 23	26 43E	553	323	323	—	—	—	—
Memel	58 43	21 7E	670	444	415	—	—	—	—
Moscow	55 45	37 34E	595	403	402	Russia 73	—	—	—
Russia 86	55 9	30 28E	648	397	397	—	—	—	—
» 73	53 26	28 18E	609	416	416	—	—	—	—
» 25	52 16	29 48E	641	435	435	—	—	—	—
Glasgow	55 53	4 18W	945	377	377	—	—	—	—
Aldergrove	54 21	6 39W	807	363	363	—	—	—	—
Birmingham	52 29	1 56W	674	436	436	—	—	—	—
London	51 28	0 19W	606	503	470	—	—	—	—
Hannover	52 23	9 44E	725	416	416	—	—	—	Hamburg
De Bilt	52 6	5 11E	730	486	484	—	—	—	—
Vlissingen	51 24	3 32E	719	503	500	De Bilt	—	200 m.p.d.	—
Dresden	51 4	13 44E	670	533	533	Prague	—	—	—
Kaiserlautern	49 27	7 46E	637	473	464	Stuttgart	—	—	—
Wien	48 15	16 22E	660	552	542	—	—	—	—
Paris	48 51	2 20E	575	538	491	—	from R.H.	150 m.p.d.	—
Angers	47 28	0 43W	570	656	490	—	int.	—	—
Gursew	47 7	51 55E	158	786	(158)	Odessa	—	—	—
Odessa	46 29	30 46E	596	690	530	—	—	—	—
San Sebastian	43 19	1 59W	598	542	542	—	int.	—	—
Perpignan	42 42	2 53E	554	747	512	—	—	—	—
Belgrade	44 48	20 27E	620	661	574	Kalocsa	int.	—	—
Sofia	42 42	23 19E	640	684	618	—	from R.H.	150 m.p.d.	—
Lisbon	38 44	9 7W	732	962	518	—	from R .H.	200 m.p.d.	—
Madrid	40 24	3 43W	419	904	(419)	—	from R.H.	150 m.p.d.	—
Rome	41 54	12 28E	910	844	577	—	from R.H.	150 m.d.p.	—
Athens	37 59	23 45E	393	985	376	—	from R.H.	150 m.p.d.	—

The table also includes columns to show when deficiencies in data had to be made good.

9. Evaporation from Lake IJssel

Complete direct checks on the computations cannot be obtained, but an incomplete check is possible for Lake IJssel. Volker (1951) gives values of mean monthly evaporation from the lake based on the behaviour of a nearby evaporation pan with a correction for the difference in lake and pan water temperatures. These values appear in the second column of table 2. The corresponding theoretical estimates for De Bilt and Vlissingen appear alongside and in yearly total both are equally good and satisfactory, but there is a marked phase difference between the two kinds of estimates (columns 2 and 5). This is attributed to neglect of the stored heat term S in the energy balance, and the final column of the table shows the result of correcting the mean for changes in stored heat. If δT is the increase in mean temperature in a month, and d is the depth of the water, then in evaporation units $S = \delta T . d/59$ mm. As T is not known, the mean of De Bilt and Vlissingen air temperatures has been used to give order of magnitude; and d has been taken as 450 cm. (The depth is usually given as 4 to 5 m.) The very good agreement between this corrected estimate and Volker's estimate must be regarded as fortuitous, but it does show the importance of changes in heat storage in monthly estimates, and their unimportance in the annual total.

TABLE 2

Evaporation from Lake IJssel (mm/month)

Month	Volker estimate	Present estimate De Bilt	Present estimate Vlissingen	mean	Corrected mean
Jan.	9	0	6	3	4
Feb.	8	15	15	15	10
Mar.	23	36	36	36	25
Apr.	49	70	67	68	56
May	82	101	101	101	83
June	100	120	114	117	99
July	109	112	115	114	109
Aug.	105	93	99	96	102
Sept.	81	58	64	61	79
Oct.	54	25	31	28	48
Nov.	20	7	11	9	23
Dec.	12	0	6	3	9
Year	653	637	665	651	647

10. Catchment data

Where precipitation and run-off are measured an estimate of annual evaporation for a catchment area can be obtained from the difference between long-term average annual precipitation and long-term average annual run-off. Except for a few values obtained from other sources, the catchment data have been supplied by Dr. S. Hénin of Versailles, France. In table 3 are some of the available data, and only one of the obvious omissions is deliberate—for Switzerland. Here values for four catchments range from 212 to 316 mm per annum, but they have been omitted because there is no entry in table 1 for a comparable climate. Other omissions are because of lack of data:— Portugal, Spain, Central Europe and the Balkans. A recent map by Reichel (1953) covers the whole of Europe but gives no values for individual catchments.

TABLE 3

Catchment data

River	Country	P (mm)	E (mm)	River	Country	P (mm)	E (mm)
Ijo	Finland	559	200	Seine	France	715	484
Karis	»	662	343	Durance	»	1250	545
Kolback-san (?)	Sweden	(650)	(350)	Selune	»	1000	464
Savean	»	(750)	336	Isle	»	870	492
Soran	»	(700)	371	Creuse	»	810	518
Dee	Scotland	1720	348	Truyère	»	1127	498
Shannon	Ireland	970	385	Saône	»	960	518
Nene	England	608	430	Isère	»	1450	400
Thames	»	755	490	Adour	»	930	548
Mean of 5	Denmark	732	349	Argens	»	904	599
Volkhov	U.S.S.R.	574	341	Havel	Germany	571	448
Daugara	»	600	347	Ems	»	729	454
Niemen	»	579	383	Elbe	»	692	500
Oka	»	558	401	Tanaro	Italy	985	515
Pripet	»	573	465	Panaro	»	1126	559
Don	»	418	330	Arno	»	1016	636
Dnieper	»	548	410	Tiber	»	967	663
Bug	»	497	433	Ofanto	»	632	475
Dniester	»	548	441	Bradano	»	676	464
Meuse	France	890	493	Silicy (Mean 4)	»	721	512
Vilaine	»	725	460	Sardinia (Mean 5)	»	860	549
Eure	»	600	500				

11. Comparison of theoretical and water-balance estimates

The values in tables 1 and 3 have been plotted on a map of Europe (Fig. 1) and the agreement between the two kinds of estimates is usually good. Without detailed discussion of possible errors in the theoretical basis, in the meteorological measurements or in the river gauging it should be sufficient to state that agreement within 10% can be regarded as satisfactory. In places there are bigger discrepancies, some of which could arise as follows. Usually the meteorological observatory is on a drier, sunnier site than the nearby catchment area. If both have an adequate summer rainfall then the catchment evaporation will be less than the station evaporation because it gets less sunshine: this may be true near Odessa (E = 53 cm) in comparison with catchments to the northwest (E = 41 to 44 cm). If summer rainfall is very deficient then increased rain in the catchment may more than compensate for decreased sunshine. This type of behaviour is suspected near Lisbon, Madrid and Athens. From Reichel's map the values of rainfall *minus* runoff at these three places appear to be 60, 40 and 50 cm; Fig. 1 gives 52, 42 and 38 cm. For the Wien, Belgrade and Sofia areas Reichel indicates values of rainfall *minus* runoff of about 50 cm/year for all three: Fig. 1 gives 54, 57 and 62 cm/year calculated from meteorological data.

The agreement is sufficiently good to suggest that the method can be used to estimate mean annual evaporation for areas where river data are not available. It may be inferred that if the annual total is correct, then the monthly values may be reliable too.

12. Annual cycle of evaporation

In table 4 appear monthly values of estimated potential transpiration E_T, and estimated actual evaporation (E%) as a fraction of the total annual estimated actual evaporation (ΣE). These are for three contrasting climates which all yield

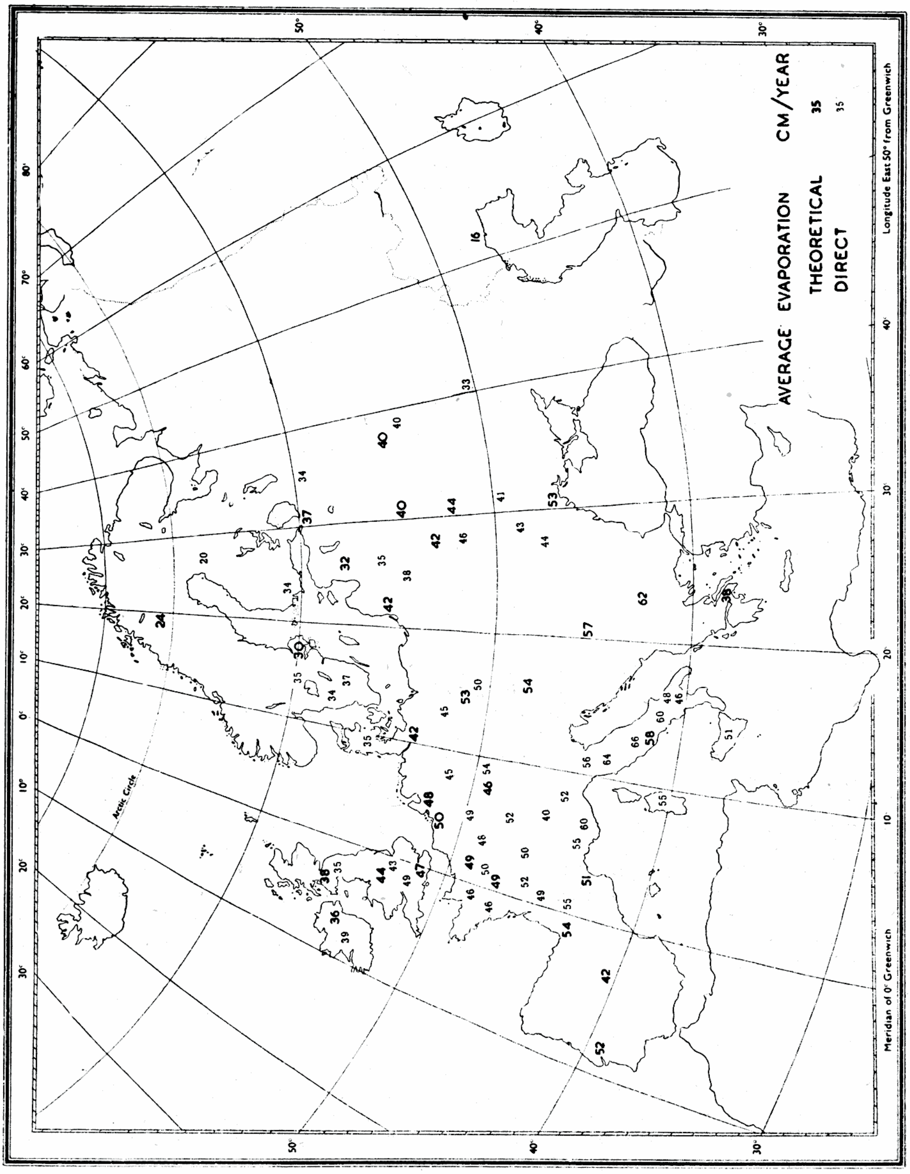

Fig. 1

a value of $\Sigma E = 42$ cm per annum. Birmingham and Memel, with almost identical annual rainfall and potential transpiration, show a marked contrast in seasonal distribution of potential transpiration, Memel having a higher summer maximum and a lower winter minimum. In June and July (on average) Memel rain is inadequate to maintain maximum transpiration. At Madrid the limiting effect of inadequate summer rainfall reaches an extreme: there are five months in which $E < E_T$, and in the last four of them $E = R$, i.e. the only evaporation that takes place is the re-evaporation of the month's rain.

TABLE 4
Annual cycle of evaporation

Month	Birmingham		Memel		Madrid	
	E_T	E%	E_T	E%	E_T	E%
Jan.	2	0	—6	—1	13	3
Feb.	10	2	—2	—0	26	6
Mar.	26	6	5	1	51	12
April	43	10	28	7	77	18
May	68	16	80	19	113	22
June	80	18	104	25	145	7
July	82	19	105	19	172	3
Aug.	67	15	80	19	150	3
Sept.	38	9	39	9	85	8
Oct.	16	4	15	4	45	11
Nov.	4	1	2	0	17	4
Dec.	0	0	—6	—1	10	2
ΣE_T (mm)	436		444		904	
ΣE (mm)		436		415		419
ΣP (mm)	674		670		419	

13. Conclusion

This survey is essentially exploratory. It has shown that physical principles can be used to estimate the order of magnitude of evaporation from natural land surfaces on the basis of simple weather data. In extreme cases two corrections are needed. First, in high latitudes, a correction for snow: second, in most inland and all low latitudes, a correction for insufficiency of summer rainfall. Further work on this problem would increase detail in two ways. First, with complete weather data for other places (and better data for some of those already studied) a more complete distribution of estimates could be made. Second, concentration on particular river basins could give month by month estimates of evaporation over a period of years and so lead to a picture of the hydrology of the basin similar to that produced for the Stour and the Thames (Penman, 1951). Such a detailed survey would provide a very critical test of the generalizations and assumptions made in this paper, but the evidence presented here suggests that the attempt could be made with a fair prospect of success.

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