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COMPUTATION OF THE EVAPORATION FROM THE BALTIC SEA FROM THE FLUX OF WATER VAPOR IN THE ATMOSPHERE

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1. *Introduction*

Direct measurements of the evaporation (or evapotranspiration) from the earth's surface with its varying nature and water content are very difficult to carry out. Therefore the most reliable data on the mean evaporation from large regions of land are those computed indirectly from measurements of precipitation, storage and discharge. The evaporation E then is determined from the equation

$$E = P - S - D\,, \tag{1}$$

where P denotes the precipitation, S the storage of surface, soil and ground water and D the discharge or runoff from the area considered. For estimation of the evaporation from sea surfaces the above formula is less useful because of the difficulty in estimating the discharge or net water transport.

Eq. (1) represents the water budget of the uppermost layer of the earth. A similar formula can be applied to the atmosphere above, if the storage now represents the change of the total content of water in gaseous, liquid and solid form in the atmosphere above the surface considered, and the discharge represents the net outflux of water across the vertical boundary of the corresponding part of the atmosphere. In this case S and D can be computed from meteorological observations just as in

Eq. (1) they were determined from hydrological measurements. If S_a and D_a are defined for the atmosphere in this way the evaporation is determined by an equation of the form

$$E = P + S_a + D_a \,. \tag{2}$$

Both equations are exactly valid if P, S and D (or S_a and D_a) can be measured.

As already mentioned Eq. (1) is not applicable on sea and ocean regions because of the difficulty in evaluating S and D. In principle, however, Eq. (2) in this case can be used just as well over sea regions as over land areas if sufficiently good meteorological observations are available for the determination of precipitation, change of water content in the atmosphere and net outflux of water from the atmospheric region considered.

The evaporation from a sea surface can also with satisfactory accuracy be computed by the aid of simple empirical formulae, which relate the evaporation to the difference in water vapor pressure at the interface between the water and the atmosphere and to the wind velocity. In its simplest form this relation may be expressed by

$$E = k(e_w - e_a)\, V \,, \tag{3}$$

where k is a constant, e_w and e_a denote the vapor pressure of the water surface and the air, respectively, and V is the wind velocity close to the surface. More complicated formulae have been derived, but for practical purposes the above simple formula gives just as satisfactory results as the more complicated expressions.

Different values for the constant k have been used. If the vapoı pressure difference, $e_w - e_a$, is expressed in millibars, the wind velocity, V, in m sec^{-1} and the evaporation, E, in mm/day the value of k, according to different authors, varies between 0.10 and 0.15. In his extensive study on the energy exchange between sea and atmosphere, Jacobs (1951) used a value of 0.143 for k. More recently Petterssen et al. (1962) considered 0.130 and Budyko (1956) 0.135 as the most likely value. For the Baltic Sea Eq. (3) has been used for computations of the evaporation by Simojoki (1948), Brogmus (1952) and Hankimo (1964). In these computations Simojoki assumed the constant k to be 0.110, Brogmus used values of 0.110—0.118 and Hankimo a value of 0.114.

The correct value of k is still somewhat questionable. Its empirical value depends largely on whether the evaporation is computed from climatological data or with aid of synoptic data. LAEVASTU (1965) has recently pointed out that synoptic data generally give a somewhat higher value for the evaporation than if the same value of k is used on climatological data. Also, it should be pointed out that k depends on the height where e_a and V are measured. Further on, it is doubtful whether the very simple empirical Eq. (3) may be successfully used under all conditions or whether more complicated formulae would give more exact results.

In the present investigation an attempt is made to compute the evaporation from the Baltic Sea by using the water budget of the atmosphere above the area considered. As already mentioned, this method is in principle exact. In reality, however, the usefulness of the method depends on the possibility to determine the water transport in the atmosphere with sufficient accuracy from the relatively few aerological stations available. Because of the limitations in this respect the method may be successfully applied on extended time periods, such as months or seasons, whereas for shorter time periods the computational errors often tend to become considerable.

2. *Formulae for computation of evaporation*

If the water budget of the atmosphere is used for computation of the evaporation, the evaporation per unit area and time, according to Eq. (2), is equal to the sum of precipitation, the change of water content of the atmosphere and the net outflux of water from the same part of the atmosphere. Since the water content and the outflux of water include water in gaseous, liquid and solid form the computation of the evaporation presupposes measurement of the total water content in the atmosphere. However, the regular aerological measurements do not give any information on the water content of the clouds. Generally this part is quite small in comparison with the content of water vapor. It can be assumed that, by neglecting the amount of water in the clouds, in most cases quite satisfactory values of the evaporation may be achieved, especially if the computation is extended over relatively large areas. This is equivalent to the assumption that all water vapor condensed immediately disappears from the atmosphere in form of observed precipitation. This assumption is naturally not always even approximately valid; the implication of this will be discussed in connection with the general interpretation of the results.

Per unit mass of air the net condensation of water vapor is equal to the decrease of specific humidity. For an individual mass unit the change of specific humidity can be written:

$$\frac{dq}{dt} = \frac{\partial q}{\partial t} + \mathbf{v} \cdot \nabla q + \omega \frac{\partial q}{\partial p} \tag{4}$$

where $\mathbf{v}$ denotes the horizontal wind vector and ω the vertical velocity in pressure coordinates ($\omega = dp/dt$). If Eq. (4) is multiplied by the density ϱ and integrated from the top ($p = 0$) to the bottom of the atmosphere ($p = p_0$) the equation can, by considering the hydrostatic equation $\partial p/\partial z = - g\varrho$, be transformed into:

$$\frac{1}{g}\int_0^{p_0} \frac{dq}{dt}\, dp = \frac{1}{g}\int_0^{p_0} \frac{\partial q}{\partial t}\, dp + \frac{1}{g}\int_0^{p_0} \nabla \cdot (q\ \mathbf{v})\, dp \tag{5}$$

where g is the gravity constant. The left-hand term in Eq. (5), however, is not only determined by the net condensation, but also by the transfer of moisture from the earth's surface in form of evaporation. If this quantity per unit area and time is denoted by E and correspondingly the precipitation by P, the water budget is expressed by

$$\frac{1}{g}\int_0^{p_0} \frac{dq}{dt}\, dp = E - P = \frac{1}{g}\int_0^{p_0} \frac{\partial q}{\partial t}\, dp + \frac{1}{g}\int_0^{p_0} \nabla \cdot (q\ \mathbf{v})\, dp \tag{6}$$

The above equation determines the difference between evaporation and precipitation as a function of the change of total water vapor content and the horizontal divergence of the water vapor flux. If the computation is extended over an area A of the earth's surface, Eq. (6) can be written in the form:

$$\int_A (E - P)\, dA = \frac{1}{g}\int_0^{p_0}\int_A \frac{\partial q}{\partial t}\, dA dp + \frac{1}{g}\int_0^{p_0}\int_A \nabla \cdot (q\ \mathbf{v})\, dA dp \tag{7}$$

Eq. (6) is useful if the flux of moisture in the atmosphere is well known as a function of space. In this case Eq. (7) gives the difference between the evaporation and precipitation integrated over an arbitrary area A. In most cases the mean difference, $\bar{E} - \bar{P}$, per unit time over the area A is of greatest interest. If we select an area limited by the vertical boundary between a number of aerological stations and L is the total length of the perimeter of the area Eq. (7) can, by using GAUSS' theorem, be transformed into

$$\bar{E} - \bar{P} = \frac{1}{g} \int_{0}^{p_0} \frac{\partial \bar{q}}{\partial t} dp + \frac{1}{gA} \int_{0}^{p_0} \oint (qv)_n \, dL dp \, . \tag{8}$$

In this equation $(qv)_n$ denotes the flux of water vapor per unit mass normal to the boundary, positive outwards, and the bars represent a mean areal value.

The above formulae for computing the difference between evaporation and precipitation were used by BENTON and ESTOQUE (1954) over the North American continent. They, however, replaced the ıeal flux of water vapor by the geostrophic flux. In spite of this approximation, which is not always permissible, they could show that the results compare favourably with computations of $E - P$ by the use of traditional hydrological methods. Other investigations with the same method have been made by BRADBURY (1957, 1958), MANABE (1957), NYBERG (1958), VÄISÄNEN (1962) and PALMÉN and HOLOPAINEN (1962). On a global scale the method has been used by STARR, PEIXOTO and LIVADAS (1958), and for studies of the atmospheric water budget in low latitudes by PALMÉN and VUORELA (1963), VUORELA and TUOMINEN (1964) and VAN DE BOOGAARD (1964).

The purpose of some of these investigations was primarily to compute the intensity of precipitation in selected synoptic situations where evaporation could be considered negligible in comparison with precipitation. Such cases are, for example, intense cyclones with strong moisture convergence. Generally the computed precipitation in such cases agrees rather well with the observed precipitation. The prime purpose of other investigations was to compute the mean evapotranspiration by assuming the precipitation to be known. Beacuse E generally is a relatively small quantity, satisfactory results may be achieved for synoptic cases only if all terms in Eq. (8) can be computed with a high degree of accuracy, which hardly is possible with aid of the available aerological data. However, if the computations are extended over sufficiently large areas and long time periods a considerable part of the errors of random nature are cancelled out.

3. *Region of computation*

For the computation of the evaporation the region of the Baltic Sea proper was selected. The equation used was Eq. (8). The area was limited by the polygon formed by the six aerological stations Stockholm, Copenhagen, Greifswald, Kaliningrad, Riga and Jokioinen. The positions of the stations are given in Fig. 1 which also gives the lengths of the sides of the polygon and the area enclosed by the sides. This area was 30.4×10^{10} m^2 large. Of the total area 77.2 per cent was covered by water and 22.8 per cent by land, if the numerous islands in the region were considered. It was necessary to include these relatively large land areas because no aerological stations exist on the Swedish coast of the Baltic Sea between Stockholm and Copenhagen and while it appeared, after testing, that the observations from the coastal station Liepaja on the eastern coast of the Baltic too often were missing.

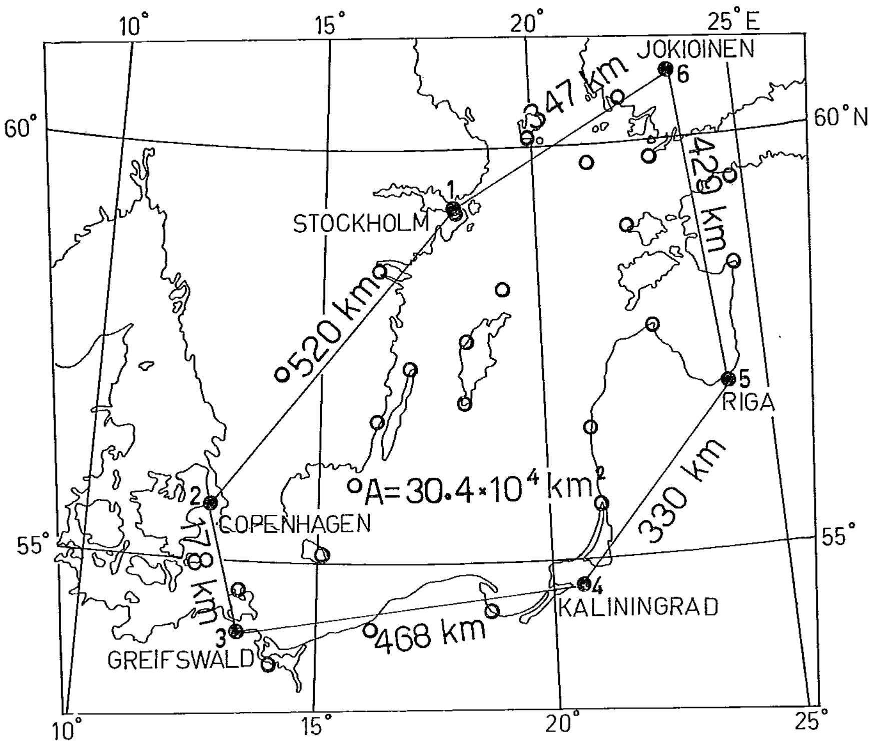

Fig. 1. Area used for the computation. The six aerological stations are marked by black circles and the stations with precipitation observations by open circles.

The relatively large proportion of land area, where the evaporation generally is quite different from the corresponding evaporation from the sea, makes it necessary to apply some approximate corrections on the computed values in order to determine the mean evaporation from the sea, which was the main purpose of the investigation. The method used for this correction will be presented later in connection with the discussion of the results.

4. *Method of the numerical computations*

Eq. (8) was applied on the aerological observations generally made twice a day (00 and 12 G.M.T.). The computation was carried out for the time period beginning on October 1, 1961 and ending on September 30, 1962. Theoretically this would mean the use of 730 complete series of observations. However, due to missing observations from one or several stations 23.7 per cent of the synoptic times had to be rejected. How this influenced the results is difficult to judge. It is quite possible that, to some extent, synoptic times with especially strong winds, and presumably strong evaporation, are relatively more numerous among the rejected cases than are situations with weaker winds. Hence it is possible that a slight underestimate of the mean evaporation appears in the average values for the months and the whole period.

For the evaluation of the two right-hand terms in Eq. (8) it was necessary to compute the change of the mean water vapor content in the atmosphere above the total area and the net outflux of water vapor across the vertical boundary of the polygon fixed by the six aerological stations. The last term represents the net outflux of water vapor at the fixed synoptic time. For the same time, however, the mean change of vapor content in the whole region could not be computed from observations. The same is also the case with the mean precipitation, $\bar{P}$, over the area. These two terms had therefore to be approximated from observations over time periods between the synoptic times. From this approximation considerable errors may occur. However, it may be expected that the influence of these errors on the mean values for longer time periods is relatively unimportant.

Since no observations of humidity were available from the region inside the polygon the mean humidity in the region had to be determined as the weighted mean value along the boundary. Hence $\bar{q}$ was defined as

$$\bar{q} = \frac{1}{2L} \sum_{j=1}^{6} q_j (L_j + L_{j+1}) \tag{9}$$

where L is the length of the perimeter and $L_7 = L_1$ in order to ensure cyclic continuity at the boundary. This approximation implies that the mean value inside the polygon is assumed to be equal to the mean value at the boundary. This assumption could be quite erroneous in some synoptic situations, but most of the errors are obviously eliminated in the average values of several consecutive synoptic times. For the change of water vapor content hence the following formula was used:

$$\frac{1}{g} \int_0^{p_0} \frac{\partial \bar{q}}{\partial t} dp = \frac{1}{2Lg} \int_0^{p_0} \frac{\partial}{\partial t} \left\{ \sum_{j=1}^{6} q_j (L_j + L_{j+1}) \right\} dp \tag{10}$$

To compute the flux of water vapor across the boundary of the region the following approximation was used. The angle between the meridian and an arbitrary side of the polygon is denoted by β_j and the reported wind direction by α_j. Considering that the whole area is relatively small, the angle β_j was defined as the angle at the midpoint of the corresponding side. Further on it was assumed that the vapor flux across the boundary between two neighbor stations varied linearily. This means that the vapor flux was defined as the arithmetic mean of the flux at two neighbor stations. Under these simplifying conditions the second right-hand term in Eq. (8) can be written in the form

$$\frac{1}{gA} \int_0^{p_0} \oint (qv)_n dL dp = \frac{1}{2gA} \int_0^{p_0} \left\{ \sum_{j=1}^{6} L_j \left[q_j v_j \sin(\alpha_j - \beta_j) + \right. \right.$$

$$\left. \left. + q_{j+1} v_{j+1} \sin(\alpha_{j+1} - \beta_j) \right] \right\} dp \tag{11}$$

An attempt was also made to compute the water vapor flux by using geostrophic winds instead of the observed winds. The normal component of the mean geostrophic wind, $\bar{v}_{ng}$, across an arbitrary side of the polygon is determined by

$$\bar{v}_{ng} = \frac{g}{L_j f} (Z_j - Z_{j+1}) \tag{12}$$

where Z_j and Z_{j+1} are the heights of an arbitrary isobaric surface at two aerological stations, L_j the distance between them and f the average value of the Coriolis parameter. Hence the expression for the water vapor flux through the boundary of the whole region is given by

$$\frac{1}{g}\int_0^{p_0}\oint (qv)_{ng} dL dp = \frac{1}{2f}\int_0^{p_0}\left\{\sum_{j=1}^{6}(q_j + q_{j+1})(Z_j - Z_{j+1})\right\} dp \qquad (13)$$

if q is assumed to vary linearily at the boundary sides.

For the evaluations of the integrals in Eqs. (10), (11) and (13) the following assumptions were made. Since the specific humidity quite rapidly decreases upward in the atmosphere, satisfactory results can be achieved if the integration is performed up to a level where the water vapor content and the flux of water vapor are of negligible importance. As the upper boundary we therefore used 400 mb instead of 0 mb. The errors introduced by this assumption are generally small in comparison with other errors following from the approximations and from inaccuracy in the aerological data. Further it was assumed that the quantities to be integrated vary linearly between the standard isobaric surfaces p_0, 1000, 850, 700, 500 and 400 mb reported from the stations. This assumption can in some cases result in more serious errors than the limitation of the integration to the lower atmosphere. However, it was necessary because of the form in which the data are available.

5. *Correction for precipitation*

The computation gives, according to Eq. (8), the difference between the mean evaporation and precipitation over the whole area marked in Fig. 1 for the given synoptic times. To determine the corresponding mean evaporation the results had to be corrected for the mean precipitation at the same time. Since the amount of precipitation generally is reported for the 12 hour periods ending at 6 and 18 hours G.M.T., with the times of the aerological observations in the middle of these time periods, the evaluated values of $\bar{E} - \bar{P}$ were considered representative of the whole 12 hour periods.

For computation of the mean precipitation 27 stations were used. These are not evenly distributed over the area, as may be seen from Fig. 1. In some regions it was necessary to use some stations outside the polygon in order to secure a satisfactory distribution of stations. The existence of several stations on islands in the central parts of the Baltic Sea is very fortunate for the estimation of the actually observed mean precipitation. The mean precipitation was then simply defined as the arithmetical mean of the reported precipitation on the 27 stations.

To give an idea of the validity of the precipitation values used in our computation of the evaporation, a comparison between the total monthly precipitation for the whole period October 1, 1961—September 30, 1962 and the average monthly precipitation over the Baltic area after SIMOJOKI (1949) is presented in Table 1. Since our computation is confined to just one year whereas SIMOJOKI's data are computed for a period of 50 years (1886—1935) no complete agreement could be expected for singular months. Remarkable is, however, the close agreement between the total values of the annual precipitation.

Table 1. Monthly mean precipitation in the area A during the period October 1, 1961—September 30, 1962 compared with the average precipitation during the period 1886—1935 after SIMOJOKI.

	J	F	M	A	M	J	J	A	S	O	N	D	Year
Oct. 1, 1961—Sept. 30, 1962	55	36	20	35	55	38	51	84	68	40	44	43	569
1886—1935, SIMOJOKI	46	32	32	37	35	38	55	66	52	57	52	52	554
Difference	9	4	−12	−2	20	0	−4	18	16	−17	−8	−9	15

6. *Discussion of the results*

Owing to the different sources of error, the computed values of the evaporation for the synoptic times (or for the 12-hour periods) are of minor interest. As was already pointed out, the available data do not permit a satisfactory evaluation of the evaporation during very short time periods. Since a large part of the errors obviously have a more or less random character, it could be expected that at least the monthly mean values would be quite reliable. However, already average values for periods of the length 4—5 days seem to be satisfactorily correlated to the prevailing synoptic situations and to the evaporation to be expected during these periods from the empirical formula (3).

Since about 22.8 per cent of the whole area was land, this has to be considered in the interpretation of the results. To determine the mean evaporation from the sea we express the total mean evaporation, $\bar{E}$, as the sum of the evaporation from the sea area, $\bar{E}_s$, and from the land area, $\bar{E}_c$, by the equation

$$\bar{E} = 0.772\,\bar{E}_s + 0.228\,\bar{E}_c\,. \tag{14}$$

By solving this equation with respect to $\bar{E}_s$ we get

$$\bar{E}_s = \bar{E} + 0.295\,(\bar{E} - \bar{E}_c)\,. \tag{15}$$

If the above formula is applied on the mean monthly evaporation, it is necessary to know, at least approximately, the average monthly evaporation from the land areas included in the region. No completely satisfactory evaporation values are available for this purpose. It is well known that the evaporation from the Baltic Sea reaches its maximum intensity during late autumn or early winter, whereas the evaporation from the surrounding continental areas again shows a maximum in late spring and early summer. By using evaporation measurements from Sweden in the land areas bordering on the Baltic Sea it was possible to estimate the probable monthly values of $\bar{E}_c$ with some accuracy. Because the correction term in Eq. (15) contains the relatively small factor 0.295 the influence of errors in the monthly values of $\bar{E}_c$ is probably not too important. The following table presents the values used for $\bar{E}_c$ in our computation of $\bar{E}_s$:

Table 2. Monthly values of $\bar{E}_c$ (in mm) used in evaluating the final evaporation from the Baltic Sea

Month	$\bar{E}_c$	Month	$\bar{E}_c$	Month	$\bar{E}_c$	Month	$\bar{E}_c$
Jan.	5	April	38	July	60	Oct.	23
Febr.	7	May	73	Aug.	41	Nov.	13
March	15	June	76	Sept.	29	Dec.	7

These values correspond to an annual mean evaporation of 387 mm from the land areas bordering the Baltic Sea.

The computed monthly values of the evaporation from the whole area A (including the land areas) were smoothed by using the simple formula

$$\bar{E}_{ms} = 0.25\,(\bar{E}_{m-1} + 2\,\bar{E}_m + \bar{E}_{m+1}) \tag{16}$$

where $\bar{E}_{m-1}$, $\bar{E}_m$ and $\bar{E}_{m+1}$ are the mean values for three consecutive months and $\bar{E}_{ms}$ is the smoothed value for the month m. This smoothing was desirable while our computation was confined just to one year and the computed evaporation therefore hardly could be considered representative of the mean monthly evaporation. On these smoothed values then the correction of Eq. (15) was applied. In Table 3 the final values of the evaporation are presented. Here $\bar{E}_w$ represents the evaporation computed by use of observed winds and $\bar{E}_g$ the corresponding values of the evaporation computed by use of the geostrophic wind, both values corrected and smoothed according to formulae (15) and (16).

Table 3. Final monthly values of the evaporation from the Baltic Sea computed from the flux of water vapor in the atmosphere for the year October 1, 1961—September 30, 1962. The values are given in mm. $\bar{E}_w$ represents values computed from the observed flux, and $\bar{E}_g$ values computed from the geostrophic flux.

Month	J	F	M	A	M	J	J	A	S	O	N	D	Year
$\bar{E}_w$	67	45	40	22	6	16	35	45	43	44	74	91	528
$\bar{E}_g$	101	69	33	13	24	38	48	64	68	97	132	126	813

The table shows very clearly the annual course of the evaporation with a very pronounced maximum in late autumn and early winter and a pronounced minimum in late spring and early summer. Somewhat surprising is the quite systematic difference between $\bar{E}_g$ and $\bar{E}_w$, which for the whole year amounts to 285 mm. This great difference indicates that the use of the geostrophic moisture flux as an approximation for the real, from wind measurements computed flux, can lead to erroneous conclusions. To give an idea of the possible errors resulting from the use of geostrophic wind, we shall look at Eq. (7) which is equivalent to Eq. (8). The former equation can be written

$$\bar{E} = \bar{P} + \frac{1}{g}\int_0^{p_0} \frac{\partial \bar{q}}{\partial t}\,dp + \frac{1}{g}\int_0^{p_0} \overline{\mathbf{v}\cdot\nabla q}\,dp + \frac{1}{g}\int_0^{p_0} \overline{q\nabla\cdot\mathbf{v}}\,dp \qquad (17)$$

where the bars again indicate mean values over the whole area A.

The last term in the above equation represents the contribution of the wind divergence to the evaporation. This term vanishes approximately if the observed winds are replaced by geostrophic winds. It can easily be shown that the sign of the divergence term, on the average, is negative because of a negative correlation between q and $\nabla\cdot\mathbf{v}$. This negative correlation depends on the fact that the content of moisture in the atmosphere generally is larger in cases of low-level convergence (cyclonic situations) than in cases of low-level divergence (anticyclonic situations). Hence the neglect of the divergence term in Eq. (17) must, on the average, result in too high values for $\bar{E}$. We therefore reject the values of evaporation computed under the assumption of geostrophic water vapor flux and assume the values computed from the observed winds to be more reliable. This conclusion is in agreement with BRADBURY's results which indicate that the values of $P-E$ computed from actual wind data are in better agreement with observed values.

7. *Errors in the computation*

There are three principal sources of error in our computation of the evaporation. These are: 1) errors in the observational data used in the computation; 2) errors following from the approximations made in evaluating the terms in Eq. (8); and 3) errors due to the neglect of the flux and atmospheric content of water in liquid and solid form.

It is obvious that many errors of different types appear in the large amount of observations used in our computation. It was already mentioned that because of obvious errors in the observations on one or several stations or the total lack of data from one or several stations the computation could not be carried out for all synoptic times of the year. As the result of this elimination of many synoptic times, some systematic errors in the montly mean values may appear, especially if the weather situation has an influence on missing data. Probably more serious is the possible existence of systematic errors owing to the differences in the observation technique in the different countries surrounding the Baltic. Also local conditions on the six aerological stations may influence the results. It is, however, not possible to estimate the errors of this type.

The approximations used in evaluating the different terms in Eq. (8) may introduce considerable errors. The assumption of a linear variation of the water vapor flux between the neighboring stations is, of course, very dangerous and can in special synoptic situations result in large errors. Also these errors, however, are largely eliminated from the mean values of longer time periods. Similarly it can be assumed that errors in the approximation of $\frac{\partial \bar{q}}{\partial t}$ and $\bar{P}$ could be large for singular synoptic times, but obviously quite small in the mean values for extended periods. A quite important source of error in the evaluation of the water vapor flux follows from the use of only the standard isobaric levels and the assumption of linearity in the vertical distribution of the humidity flux especially because of the lack of additional levels between 1000 and 850 mb, where a considerable part of the flux divergence occurs. It is probable that this lack of additional data appreciably reduces the absolute values of the moisture flux divergence in some cases characterized by strong positive or negative divergence.

The errors in the computed values of the evaporation, resulting from the assumption that the content and flux of water in the atmosphere entirely depends on the distribution of specific humidity, may in some special cases be considerable. This assumption means that all the condensed water vapor immediately leaves the atmosphere in form of precipitation. In some cases the air enters the region with no or very few clouds, but again leaves the region with quite dense clouds formed over the sea. Such cases occur especially in outbreaks of cold air. In cases of this type Eq. (8) gives an underestimate of the evaporation. In the mean monthly values the influence of this error, however, probably is small but no estimate has been made.

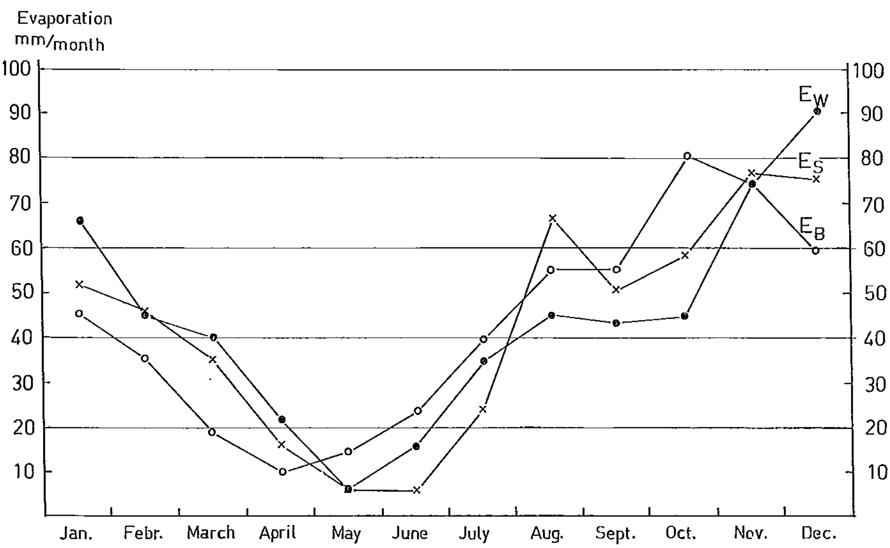

Fig. 2. Monthly values of the computed evaporation from the Baltic Sea proper for the time period from October 1, 1961 to September 30, 1962, E_W, compared with the monthly mean evaporation after SIMOJOKI, E_S, and BROGMUS, E_B.

8. *Comparison with some previous estimates of the evaporation from the Baltic Sea*

All previous computations of the evaporation in the Baltic area have been performed with the aid of empirical formulae of the type (3). As was pointed out in the Introduction, quite reliable results may be achieved in this way for sea regions. A comparison of the results presented here and those obtained from some of the previous computations is therefore of considerable interest.

For the lighthouse Bogskär in the northern Baltic Sea proper SIMOJOKI (1948) computed the mean monthly evaporation for a period of 13 years. Similar computations were made by BROGMUS (1952) for the central region of the Baltic (»Gotland Sea»). In Table 4 the computed values of the monthly mean evaporation after SIMOJOKI and BROGMUS are presented and compared with the $\bar{E}_w$-values from Table 3. The same values are presented graphically in Fig. 2.

Table 4. Comparison of the monthly evaporation (in mm) from the Baltic Sea according to SIMOJOKI, BROGMUS and the present investigation.

Month	SIMOJOKI	BROGMUS	$\bar{E}_w$	0.5 (S+B)$-\bar{E}_w$
January	52	45	67	−19
February	46	36	45	− 4
March	35	19	40	−13
April	16	10	22	− 9
May	6	15	6	4
June	6	24	16	− 1
July	24	40	35	− 3
August	67	55	45	16
September	50	55	43	9
October	58	81	44	26
November	77	74	74	2
December	75	60	91	−23
Year	512	514	528	−15

The agreement between the different values is surprisingly good. In judging the values of the last column, it should be observed that the values computed by us are strongly influenced by the fact that they are confined to one year only. For instance, the relatively low value of the evaporation in October 1961 and the high evaporation in December the same year in comparison with the values after SIMOJOKI and BROGMUS are easily explained by the special conditions prevailing during these months. October 1961 was characterized by unusually high air temperature which suppressed the evaporation, whereas December was characterized by several strong outbreaks of cold air with the water temperature still being high because of the preceding warm autumn.

More recently HANKIMO (1964) has published some new values of the evaporation from the Baltic Sea proper covering a part of the same time period as that used by us. His computations were carried out by the use of Eq. (3) for the months December, 1961—May, 1962. In the following table his results are compared with our results for the same months.

Table 5. Comparison between evaporation computed by HANKIMO and the corresponding values in the present investigation.

Month	Dec.	Jan.	Febr.	March	April	May	Dec.—May
Hankimo	103	50	62	39	13	9	276
$\overline{E}_w$	91	67	45	40	22	6	271
Difference	12	−17	17	−1	−9	3	5

For the whole period the agreement is very good. The somewhat large differences for some of the months can partly be explained by the fact that we had to exclude some days from our computation because of missing observations. On the whole, however, the strong decrease of evaporation from its maximum value in December to its minimum value in May is similar in both series.

9. *Weather associated with very strong evaporation*

Since the evaporation from a sea surface is approximately proportional to the difference $e_w - e_a$ and the wind velocity V, the most intense evaporation occurs in weather situations characterized by strong outbreaks of cold air, especially if the sea temperature is relatively high compared with the air temperature. Synoptic situations of this type are quite common over the Baltic Sea in autumn and early winter.

As an example of such a weather situation we select the period December 18—22, 1961. The computed values of the evaporation during this period are presented in Table 6.

Table 6. Evaporation (in mm/12 h) during a period of a strong outbreak of cold air over the Baltic Sea.

Time	Dec. 18	Dec. 19		Dec. 20		Dec. 21		Dec. 22	Mean
	12h	00h	12h	00h	12h	00h	12h	00h	
Evapor.	6.5	4.5	5.6	0.7	7.8	5.7	5.0	2.9	4.8

The mean evaporation over the whole area A in Fig. 1 was during this period of 4 days 4.8 mm/12 h, corresponding to about 12 mm/day if the computed values are corrected for the land areas according to Eq. (15). This mean evaporation from the sea region corresponds to a flux of latent heat of 0.5 cal cm^{-2} min^{-1}. It shows the strong influence of the Baltic Sea as a heat source for the surrounding coastal regions in the cold season, especially if the simultaneous flux of sensible heat, which is almost equally strong (Hankimo, 1964) is considered.

Acknowledgements

The authors wish to express their gratitude to the late Dr. Aimo Väisänen for his valuable help in preparing the aerological data for the computations. We also wish to thank the Finnish Cable Works Ltd for the use of the electronic computer Elliott 803, and the Head of the Department of Meteorology, University of Helsinki, Professor L. A. Vuorela for his helpfulness in many respect.

Financial support for this study was received from the National Research Council for Sciences and from the Sohlberg Foundation of the Finnish Society of Sciences.

REFERENCES

BENTON, G. S. and ESTOQUE, M. A., 1954: Water-vapor transfer over the North American continent. *J. Meteor.*, **11**, 462—477.

BRADBURY, D. L., 1957: Moisture analysis and water budget in three different types of storms. *Ibid*, **14**, 559—565.

—»— 1958: Moisture analysis and water budget in some nondeveloping cases. *Geophysica*, **6**, 179—187.

BROGMUS, W., 1952: Eine Revision des Wasserhaushaltes der Ostsee. *Kieler Meeres-forschunden*, Band IX, 1.

BUDYKO, M. I., 1956: *The heat balance of the earth's surface.* Hydrometeorological Publishing House, Leningrad. Translated by N. A. Stepanova, U. S. Weather Bureau, 1958.

HANKIMO, J., 1964: Some computations of the energy exchange between the sea and the atmosphere in the Baltic area. *Finnish Meteorological Office Contributions*, No. 57.

JACOBS, W. C., 1951: The energy exchange between sea and atmosphere and some of its consequences. *Bull. Scripps Inst. Oceanogr.* **6**, 27—122.

LAEVASTU, T., 1965: Daily heat exhange in the North Pacific, its relations to weather and its oceanographic consequenses. *Soc. Sc. Fennica, Comm. Phys.-Math.* **31**, (in press).

MANABE, S., 1957: On the modification of air-mass over the Japan Sea when the outburst of cold air predominates. *J. Meteor. Soc. Japan*, Ser. II, **35**, 311—326.

NYBERG, A., 1958: A determination of some monthly values of evapotranspiration in Finland and in a region in the Northeastern Atlantic. *Geophysica*, **6**, 377—387.

PALMÉN, E. and HOLOPAINEN, E. O., 1962: Divergence, vertical velocity and conversion between potential and kinetic energy in an extratropical disturbance. *Geophysica* **8**, 89—113.

PALMÉN, E. and VUORELA, L. A., 1963: On the mean circulations in the Northern Hemisphere during the winter season. *Quart. J. Roy. Meteor. Soc.* **89**, 131—138.

PETTERSSEN, S., BRADBURY, D. L. and PEDERSEN, K., 1962: The Norwegian cyclone model in relation to heat and cold sources. *Geofys. Publ.* **24**, 243—280.

SIMOJOKI, H., 1948: On the evaporation from the Northern Baltic. *Geophysica*, **3**, 123—126.

SIMOJOKI, H., 1949: Niederschlag und Verdunstung auf dem Baltischen Meer. *Fennia* **71**, No. 1.

STARR, V. P., PEIXOTO, J. P. and LIVADAS, G. C., 1958: On the meridional flux of water vapor in the Northern Hemisphere. *Geofis. pur. appl.* **39**, 174–185.

VAN DE BOOGAARD, H. M. E., 1964: A preliminary investigation of the daily meridional transfer of atmospheric water vapour between the Equator and 40°N. *Tellus*, **16**, 43–54.

VUORELA, L. A. and TUOMINEN, I., 1964: On the mean zonal and meridional circulations and the flux of moisture in the Northern Hemisphere during the summer season. *Pure Appl. Geophys.* **57**, 167–180.

VÄISÄNEN, A., 1962: A computation of the evaporation over Finland during a rainless period based on the divergence of the water-vapor flux. *Geophysica*, **8**, 159–165.