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JOURNAL OF MARINE RESEARCH

VOLUME III

1940

NUMBER 2

ON THE ANNUAL AND DIURNAL VARIATION OF THE EVAPORATION FROM THE OCEANS*

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The average annual evaporation of the oceans in different latitudes can be considered fairly well established (Mosby, 1936, Wüst, 1936), but so far little is known as to the character of the annual and the diurnal variations. Helland-Hansen (1930) has pointed out that evaporation will be facilitated if the sea surface is warmer than the air, whereas evaporation will decrease or cease if the sea surface is colder. He has therefore concluded that in middle latitudes the evaporation is at a maximum in fall and early winter, and at a minimum in summer, and, similarly, it may be concluded that the evaporation is at a maximum during the night and is somewhat smaller in the middle of the day. It will here be attempted to approach the problem on the basis of energy considerations which have been applied successfully in the study of the average evaporation from the sea (W. Schmidt, 1915), and of the evaporation from reservoirs and lakes (Ångström, 1920, Cummings and Richardson, 1927). A few specific examples will be discussed and a few conclusions will be drawn.

In this case, the energy equation states that the total amount of heat lost from a square unit of surface area in unit time, q_a , partly

* Contributions from the Scripps Institution of Oceanography, New Series, Number 90.

as heat used for evaporation, q_e , and partly as sensible heat given off to the atmosphere, q_c , must equal the difference between the surplus radiation (radiation from sun and sky minus effective back radiation from the sea surface) which is absorbed by the water, q_r , and the time change of the heat content of the water:

$$q_e + q_c = q_r - \int_0^{\infty} \frac{d}{dt} (c\rho\tau) dz \quad (1)$$

Here c , ρ , and τ mean specific heat, density and temperature, respectively. The amount of heat given off to the atmosphere by convection is generally assumed to be a small fraction of q_e , and $q_a = q_e + q_c$ represents, therefore, approximately the amount of heat used for evaporation.

The surplus of radiation can be found either by means of Kimball's data (1928) or Mosby's formula (1936), taking into account reflection at the surface and using Ångström's results (1920) as to the effective back radiation. Evaluation of the integral

$$\int_0^{\infty} \frac{d}{dt} (c\rho\tau) dz = c\rho \left[\int_0^{\infty} \frac{\partial\tau}{\partial t} dz + \int_0^{\infty} v_x \frac{\partial\tau}{\partial x} dz + \int_0^{\infty} v_y \frac{\partial\tau}{\partial y} dz + \int_0^{\infty} v_z \frac{\partial\tau}{\partial z} dz \right] \quad (2)$$

is generally more difficult, because it necessitates knowledge of the temperature changes within a given water mass. In certain instances approximately correct values can be determined.

ANNUAL VARIATION OF EVAPORATION

Helland-Hansen (1930) has pointed out that water masses of different origin may be carried into the area under examination, for which reason the observed temperature changes may be related to variations in the character of the water masses present and not to processes of heating or cooling. In order to overcome this difficulty, Helland-Hansen examined the "temperature anomalies" as defined by means of deviations from standard T-S curves, and on this basis he arrived at a series of curves showing the effect of heating in three areas of the eastern North Atlantic, the data being relatively complete for an area off the Bay of Biscay, with its center at 47° N. and 12° W. The data are from a region without very distinct currents, for which reason only the first term on the right-hand side of equation (2) has to be considered, and this can be computed from Helland-Hansen's data, introducing $c = 0.93$ and $\rho = 1.025$. Thus, the curve marked q_r in Figure 31A was obtained, showing the variation during the year of the

amount of heat involved in heating or cooling the water, and expressed in gm. cal./cm.² min. Kimball gives values of the total incoming radiation on the 21st of every month of the year at 48° N., 4° W., and the effective back radiation in this locality can be obtained by means of Ångström's table using the same values of cloudiness which were used by Kimball in his computation. Thus, the curve marked q_r

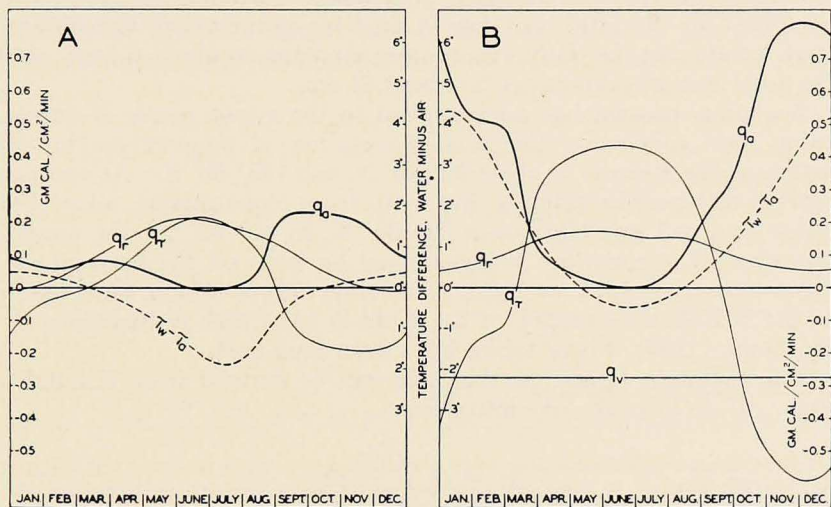


Figure 31. A. Annual variation of evaporation off the Bay of Biscay, in about 47° N. and 12° W. Curve q_r represents the surplus of radiation heat which is absorbed by the water, curve q_T represents the heat involved in heating and cooling the water. Curve q_a represents the heat given off to the atmosphere mainly by processes of evaporation. The curve $t_w - t_a$ represents the difference between the surface temperature of the water and the temperature of the air. B. Annual variation of evaporation within the region of the Kuroshio in about 32° 40' N. and 135° 50' E. The designation of the curves is the same as in figure 32 A. In addition, q_v represents the loss of heat of the water due to change of temperature in the direction of flow.

is found showing the variations through the year of the radiation surplus, also expressed in gm. cal./cm.²min. Assuming that the values of q_r are applicable to the same region as those of q_T , the amount of heat given off by evaporation and convection is found by subtraction of the two curves and is represented by curve q_a in Figure 31A.

The curve q_a represents roughly the annual variation of evaporation, and this computation leads therefore to the result that the evaporation is great from the middle of September to the middle of December, is small from January to May, and ceases in June and July. It is of particular interest to observe that a secondary minimum of evaporation appears to occur in February and that apparently the

increase of radiation surplus in early spring is accompanied by an increase in evaporation leading to a secondary maximum in March.

In the figure, the curve marked $\tau_w - \tau_a$ represents the temperature difference between the water and the air as derived from the charts for seasons in the Atlas of Climatological Charts (1939). The agreement between this curve and the curve q_a is good, as should be expected from Helland-Hansen's reasoning. Assuming that in this case 90% of the total heat loss is used for evaporation, one obtains that in late fall the daily evaporation amounts to about 0.5 cm. and the total annual evaporation to about 81 cm.

A similar method has been applied to the region south of Japan, using data concerning the annual variation of temperature in the waters of the Kuroshio in about $32^\circ 40' N.$ and $135^\circ 50' E.$ At stations located along a line running due south from Sionomisaki, about 100 miles long and extending from $33^\circ 30' N.$ to $31^\circ 50' N.,$ the annual variation of temperature was examined by Suda on the basis of observations from 1923 to 1927. The observations which were taken in the N.E.-flowing branch of the Kuroshio have been re-examined by Koenuma (1939), whose tables have been used here.

The radiation income in this area can be derived from Kimball's data. If, in this case, one introduces

$$q_\tau = c\rho \int_0^\infty \frac{\partial \tau}{\partial t} dz$$

one arrives at the result that in summer large amounts of heat are taken up from the atmosphere. This appears unreasonable and must be due to the fact that the other terms in (2) can not be neglected, because in this region a distinct current exists within which the temperature decreases in the direction of flow. It is not known how great this decrease is, nor whether it is independent of the seasons, but for the sake of simplicity it will be assumed that the term

$$q_v = c\rho \int_0^\infty v_x \frac{\partial \tau}{\partial x} dz$$

is constant throughout the year and so great that in summer q_a becomes zero and that other terms can be omitted. Figure 32B shows the curves q_r , q_τ , and q_v , and

$$q_a = q_r - q_\tau - q_v$$

which approximately represents the evaporation. In this case the evaporation is at a maximum in winter, and this result appears trust-

worthy, because in winter the cold and dry monsoon blows from the Asiatic Continent.

In November, December, and January an average of .750 gm. cal./cm.² min. is given off to the atmosphere. Of this amount three-fourths or more is given off as latent heat of water vapor corresponding to an evaporation of about 1.4 cm. per day. On a similar basis the total evaporation in one year would amount to 266 cm. which is more than twice Wüst's average value in that latitude, 116 cm. per year. The difference indicates that the evaporation is greatest near the west-

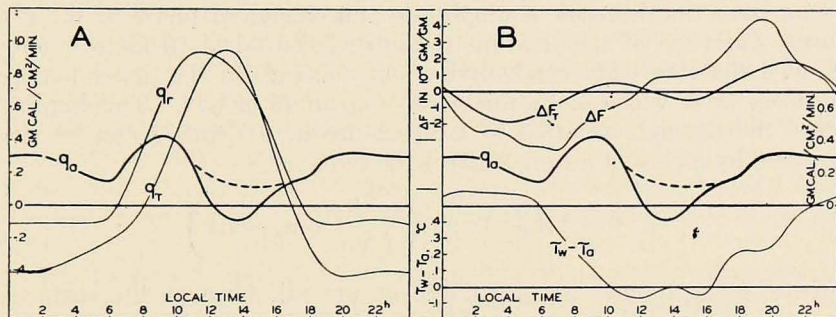


Figure 32. A. Diurnal variation of evaporation in the tropics, according to observations on four "Meteor" stations. The designation of the curves is the same as in figure 32A. B. The curve q_e represents the diurnal variation of heat given off to the atmosphere at four "Meteor" stations, the curves $\tau_w - \tau_a$ and ΔF represent the difference between water-surface temperature and air temperature, and the diurnal variation of the specific humidity of the air on the same four days. The curve ΔF_T represents the diurnal variation of the specific humidity of the air on 18 selected days in the tropics, according to Reger.

ern border of the North Pacific Ocean, as may be expected when the character of the currents and the atmospheric circulation is considered.

The temperature difference, water minus air, which is also shown in the figure as derived from the Atlas of Climatological Charts of the Oceans, varies as the computed evaporation and in winter is greater near Japan than anywhere else in the Pacific Ocean.

In the part of the Kuroshio under consideration, one may perhaps assume that the velocity between the surface and 200 meters is 50 cm./sec. With this value of v one finds that the value of q_v which was introduced corresponds to a decrease of the temperature of 0.2° C. in the upper 200 meters on a distance of about 400 km. It is impossible to verify this conclusion because the change in temperature in the direction of flow of the Kuroshio appears to be mainly caused by admixture of colder water. The estimated cooling due to the

amount of heat given off to the atmosphere appears, however, to be conservative.

DIURNAL VARIATION OF EVAPORATION IN THE TROPICS

In order to examine the diurnal variation of evaporation by means of the same method it is necessary to know the diurnal variation of temperature at different depths. From the tropics the observations at four of the "Meteor" anchor stations give the necessary information. These data have been examined by Defant (1932) who, from the decrease of amplitude and change of phase of the diurnal variation of temperature at the surface and at 50 meters, concluded that the eddy conductivity was constant in the upper homogeneous layer which had a thickness of about 70 meters. This implies that the diurnal variation of temperature at a depth z can be expressed by means of a sum of harmonic terms,

$$\tau = \Sigma A_n e^{-\kappa_n z} \sin \left(n \frac{2\pi}{T} t + \alpha_n - \kappa_n z \right)$$

where A_n represents the amplitude of the n th term at the surface, α_n , the phase of that term at the surface, T , the length of the day in hours, minutes, or seconds depending upon the units used for measuring the time, and where

$$\kappa_n = \sqrt{\frac{n\pi\rho}{\eta T}}$$

Here, η is the eddy conductivity. From change of phase and reduction of amplitude Defant found $\eta = 322$ gm./cm. sec. The amount of heat used in unit time for local heating and cooling is, therefore,

$$q_\tau = c_\rho \int_0^\infty \frac{\partial t}{\partial t} dz = \Sigma A_n \frac{n2\pi}{T} \frac{c_\rho}{\kappa_n 2\sqrt{2}} \sin \left(n \frac{2\pi}{T} t + \alpha_n + \frac{\pi}{4} \right)$$

In this case it may be assumed that q_v can be neglected.

By means of the hourly values of sea-surface temperature as given by Defant the curve marked q_τ in Figure 32A has been computed, but it should be observed that the amplitude of the curve is subject to some uncertainty because it depends upon the accuracy of the value of η . Taking into account the location of the stations, the dates on which they were occupied, and the average hourly values of cloudiness, as stated in the tables of meteorological observations (Kuhlbrodt and Reger, 1938), the curve q_r has been computed by means of

Mosby's formula and the Ångström values of back radiation. The difference between these two curves, q_a , represents the amounts of heat given off in unit time to the atmosphere by convection or lost by evaporation. It may again be assumed that the heat used for evaporation is considerably greater than that given off by convection, and that the curve q_a , therefore, roughly represents the diurnal variation of the evaporation.

In the tropics the diurnal variation of evaporation appears to be somewhat similar in character to the annual variation in middle latitudes in the eastern North Atlantic. The maximum evaporation takes place during the early hours of the night, falls off during the latter part of the night towards a secondary minimum in the early morning, rises sharply when the sun appears above the horizon, but drops again to a low minimum in the early afternoon hours. Thus, a semi-diurnal period appears more pronounced than a diurnal, but it is possible that, owing to uncertainties as to the amplitudes of q_r and q_{τ} , the afternoon minimum, as indicated by the dashed part of the curve q_a , appears too low in this special case. The total evaporation in one day is about 0.5 cm.

It is possible to point out certain observations which substantiate the conclusion that in the tropics a semi-diurnal period prevails. In general, q_r can be expressed as the sum of a constant term and a series of harmonic terms,

$$q_r = \bar{q}_r + \Sigma b_n \sin \left(n \frac{2\pi}{T} t + \beta_n \right)$$

and, similarly, q_{τ} , can be expressed as a series of harmonic terms,

$$q_{\tau} = \Sigma a_n \sin \left(n \frac{2\pi}{T} t + \alpha_n + \frac{\pi}{4} \right)$$

Therefore,

$$q_a = \bar{q}_r + \Sigma c_n \sin \left(n \frac{2\pi}{T} t + \gamma_n \right)$$

where

$$c_n = \sqrt{a_n^2 + b_n^2 - 2a_n b_n \cos \left(\beta_n - \alpha_n - \frac{\pi}{4} \right)},$$

$$\text{tg } \gamma_n = \frac{a_n \sin \left(\alpha_n + \frac{\pi}{4} \right) - b_n \sin \beta_n}{a_n \cos \left(\alpha_n + \frac{\pi}{4} \right) - b_n \cos \beta_n}$$

We are here interested in the variation of the evaporation and can therefore disregard the constant term. It is evident that each one of the harmonic terms will attain its minimum value if $\alpha_n = \beta_n - \pi/4$. The surplus radiation q_r is, in general, symmetrical around noon and the phase angles of the first two harmonic terms have the values $\beta_1 = 270^\circ$ and $\beta_2 = 90^\circ$. In the tropics, the amplitude of the second harmonic term is relatively great, but in middle latitudes it nearly disappears. It is striking that harmonic analysis of a large number of observations of sea-surface temperatures in the tropics leads to the result that $\alpha_1 = 270^\circ - 45^\circ = 225^\circ$. Hann's analysis of the "Challenger" data gives 227.5, whereas Krümmel and Wegemann from other data obtain 223.8 and 225.8, respectively (Defant, 1932). This indicates that the diurnal variation of the evaporation is as small as possible. There exists, on the other hand, no agreement between the second terms. From the four days dealt with above, Defant's analysis gives $\alpha_2 = 19^\circ$, and the studies by Krümmel and Wegemann give $\alpha_2 = 26^\circ$ and 26.9° , respectively, instead of 45° . Therefore, it may be expected that the semi-diurnal term will be relatively larger. If we make the assumption that the amplitudes b_2 and a_2 are equal, we find that the phase angle of the semi-diurnal evaporation term is about 170° , meaning that the first minimum comes at 3.3^h and the second minimum at 15.3^h . In middle latitudes the diurnal terms dominate both in q_r and q_τ (Wegemann, 1920), and it appears therefore probable that the diurnal variation of the evaporation is more prominent there, and is mainly dependent upon the difference $a_1 - b_1$, because α_1 is close to 225° .

FACTORS WHICH INFLUENCE THE DIURNAL VARIATION OF EVAPORATION

The evaporation from a water surface must, on the one hand, be nearly proportional to the difference between the surplus radiation and the heat used for temperature change of the water, and on the other hand must equal the amount of water vapor which is transported upwards from the surface by processes of eddy diffusion. This upwards transport depends mainly upon the wind velocity, the difference between vapor pressure at the surface and in the air at a short distance from the surface, and the state of turbulence of the air (Sverdrup, 1937). Over the oceans the diurnal variations of the two former factors are relatively small, and it may therefore be assumed that the state of turbulence of the air is of dominant importance. It appears reasonable to assume that shortly after sunrise the turbulence will increase, for which reason the evaporation will also increase. In the tropics

the increase will not continue unimpeded, however, because towards noon the air temperature may become higher than that of the temperature of the sea surface, the stratification becomes stable and the turbulence will be reduced in the early hours of the afternoon. In the late afternoon when the air temperature has dropped below that of the surface, the turbulence will again increase and will reach a maximum some time after sunset. Such a diurnal variation of the turbulence would account for the double period of the evaporation in the central part of the tropics, but in the trade-wind areas and in middle latitudes the air temperature remains below that of the water, and in these circumstances turbulence does not decrease at noon. There, a single period of the evaporation should be expected, probably with minimum in the early forenoon when the temperature difference, water minus air, is smallest (Kuhlbrodt and Reger, 1938). It may be pointed out that over land the evaporation in low latitudes appears to have a single diurnal period with its minimum value at sunrise and its maximum value in the early afternoon. Such a period is in agreement with the fact that over land the surface is warmer than the air throughout the day, and the turbulence has its maximum in the early afternoon hours.

THE DIURNAL VARIATION OF EVAPORATION AND OF THE SPECIFIC HUMIDITY OF THE AIR IN THE TROPICS

In Figure 32B are shown the computed diurnal variations of the heat given off to the atmosphere, q_a , at the four "Meteor" stations, and curves representing the hourly deviations of the specific humidity from the average value, ΔF , and the difference between water and air temperature, $\tau_w - \tau_a$. The minimum of evaporation in the early afternoon hours appears, as was pointed out, to be associated with the fact that in these hours the water was colder than the air, for which reason the turbulence and, consequently, the upwards transport of water vapor was reduced.

The variation of the specific humidity of the air, ΔF , indicates, on the other hand, the presence of a semi-diurnal period which is slightly displaced in relation to the evaporation. These curves are all based on observations from the four days only, but in his discussion of the meteorological observations from the "Meteor," Reger points out that in the tropics the specific humidity shows a pronounced semi-diurnal variation with minima about 4^h and 15^h, and maxima at 10^h and 21^h. Reger's general curve for the tropics is also entered in the figure, and is marked ΔF_T .

If the results as to the evaporation which have been obtained by our example can be generalized it appears that there is a close correspondence between the diurnal variation of the evaporation and that of the specific humidity of the air, as may be expected from a consideration of the processes involved. Variations of the specific humidity in a cube of air at a short distance from the sea surface depend upon the difference between the supply of water vapor through the lower horizontal surface of the cube and the upwards transport of water vapor through the upper horizontal surface. Near the sea surface differences in transport through the vertical surfaces of the cube can be neglected, and if the upwards transport of water vapor associated with the turbulence of the air were constant, the specific humidity would not vary. If now the evaporation increases but the upwards transport remains unaltered the specific humidity must increase. Suppose next that the increase of the evaporation stops so that the supply of water vapor reaches a new but higher value than previously. Then the specific humidity will a little later become constant but will have attained a somewhat higher value. When the evaporation decreases similar considerations are applicable and one is therefore led to the conclusion that under these conditions the specific humidity must vary as the evaporation, but the minima and maxima of the specific humidity must come some time after the corresponding extreme of evaporation. Similarly, a decrease in the upwards transport must lead to an increase of the specific humidity if the evaporation remains constant. Present knowledge of variations of the evaporation and the state of turbulence is too scanty to permit a further examination of the question, but in the tropics the agreement between the variations of the evaporation and the specific humidity is so striking that it appears worth while pointing out. In the trade-wind area a single diurnal period of the specific humidity dominates and there the semi-diurnal period in the evaporation should be expected to be small.

SUMMARY

The annual variation of evaporation is examined by means of the energy equation in an area off the Bay of Biscay and in a part of the Kuroshio off southern Japan. In both areas maximum evaporation appears to take place in late fall or early winter, the evaporation in the Kuroshio area being nearly three times that off the Bay of Biscay. In both areas the evaporation ceases in summer, but off the Bay of Biscay a secondary maximum occurs in March.

The diurnal variation of evaporation in the tropics is examined in a similar manner. A semi-diurnal period is found which is explained as a result of the varying state of turbulence and which appears to throw light on the previously established semi-diurnal variation of the specific humidity of the air.

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