ON THE ENERGY EXCHANGE BETWEEN SEA AND ATMOSPHERE*

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FOREWORD

The present paper is a preliminary report on an investigation of the exchange of heat and water vapor between the sea and atmosphere with particular reference to the regional and seasonal variations in these quantities over the North Atlantic and North Pacific oceans. The *mean annual* evaporation over the oceans according to latitude ranges can be considered fairly well established (Wüst, 1936; Mosby, 1936; McEwen, 1938), but there has been no extensive investigation of the seasonal and intra-latitudinal variations. Since a knowledge of evaporation rates is necessary before computations of the heat exchange between sea surface and atmosphere can be attempted, as a result of lack of these data, very little is known concerning the regional variations in the total energy exchange between sea and atmosphere.

The present study approaches the problem first on the basis of energy considerations such as have successfully been employed by Mosby (loc. cit.), and mean annual evaporation values have been computed for four selected areas in the North Atlantic and North Pacific oceans. The values so derived have then been compared with the evaporation within the same areas computed on the basis of more recent theoretical considerations of the processes of interchange of water vapor within the turbulent layer near the sea surface (Sverdrup, 1937; Montgomery, 1940), when using recently published data on humidities, sea surface temperatures and wind speeds over the oceans. However, the necessary climatic data over the oceans are of doubtful

* Contributions from the Scripps Institution, New Series, No. 158.

„Mass und Zahl behalten
ihre Bedeutung auch in der
modernen Geography."
(N. Krebs)

accuracy, particularly with reference to humidity, and a comparison of the two methods was necessary in order to determine the statistical adjustment to be applied to these data and further, to establish the validity of the latter method of computation.

By means of the final evaporation equation obtained, seasonal values of evaporation for each five-degree square in the North Atlantic and North Pacific have been computed. In addition, through the use of the regional and seasonal evaporation values so derived, and the Bowen formula (Bowen, 1926), which allows the computation of the ratio between evaporation and heat exchange, charts have been prepared for the same areas showing the total heat and energy exchange between sea surface and atmosphere for all seasons.

I. THE AMOUNT OF HEAT ABSORBED AT THE SEA SURFACE

It is obvious that, for the oceans as a whole, the energy loss from the sea surface during the course of the year must, for all practical purposes, equal the total incoming radiation absorbed at the surface. For small areas of the sea surface it is also evident that the evaporation rate expressed in grams or centimeters of depth may be determined entirely by energy rendered available for that purpose in accordance with the equation:

Notation

E = Evaporation in grams or centimeters.
E_1 = Mean annual evaporation computed by means of the energy equation.
E_2 = Mean annual evaporation computed by means of the theoretical evaporation equation of Sverdrup (uncorrected).
Q_e = Heat of evaporation.
Q_c = Heat exchanged between sea and atmosphere through convection.
Q_a = Total energy exchange between sea and atmosphere $(Q_e + Q_c)$.
Q_t = Total solar energy reaching the sea surface.
Q_b = Total back radiation at $\bar{C}$.
Q_v = Amount of heat added to or removed from the water mass through advective processes.
L_t = Latent heat of vaporization of water $= 594.9 - 0.51t$.
Q_s = Intensity of direct sun radiation upon a surface normal to the sun's rays.
S = Solar constant (1.932 gm. cal. cm^2/min.)
Q_0 = Total incoming radiation upon a horizontal surface with clear sky.

Q_{ab} = Total heat absorbed at the sea surface.
$eff\, Q_b$ = Effective back radiation with clear sky.
T = Turbidity factor (Linke).
m = Unit dry air mass at $p = 760$.
D = Ratio between direct and total radiation.
C = Percent of cloudiness.
r = Percent of incident solar heat reflected at sea surface.
R = Bowen ratio = Q_c/Q_e.
t_w = Temperature of sea surface.
t_a = Air temperature at height a.
e_w = Vapor pressure at sea surface (0.98 times e at t_w).
e_a = Vapor pressure at height a.
a = Height of observation.
H = Diffusion coefficient (0.235).
p = Pressure.
k_0 = Universal turbulence constant (von Kármán) = 0.4.
W = Friction velocity.
W_a = Velocity at height a.
z_0 = Roughness parameter (Rossby) = 0.6.
d = Thickness of laminar layer.
τ = Tangential stress of wind.
q = Specific humidity.
$\gamma_a = \dfrac{k_0}{\ln a/z_0}$
Γ = Evaporation coefficient (Montgomery).
c_p = Specific heat of air at constant pressure.
ρ = Density.
h = Solar altitude.
A = Eddy conductivity.
$\bar{k} = \bar{Q}_0/\bar{h}$. (Mosby—"Richtungskoeffizient").

$$\underline{E} = \frac{Q_t - rQ_t - Q_b - Q_c - Q_v}{L_t} \qquad \text{I-(1)}$$

where Q_t is the total heat reaching the sea surface, r is the percent reflected, Q_b the heat loss through back radiation to the sky, Q_c the heat exchanged between sea and atmosphere through convective processes, Q_v the quantity of heat added to or removed from the unit water column through advective processes and L_t is the latent heat of vaporization of water at temperature t.

The intensity of the direct sun radiation upon a surface normal to the sun's rays is:

$$Q_s = S_e^{-Ta_m m} \quad \text{g. cal./cm}^2 \text{ min.} \qquad \text{I-(2)}$$

where S is the solar constant (1.932 g. cal./cm min.), e the base of natural logarithms, T the turbidity factor after the definition of Linke (1922), m the unit air mass at $p = 760$ mm. and

$$a_m = 0.128 - 0.054 \log \text{m}. \qquad \text{I-(3)}$$

It is thus only necessary to know the sun's altitude h, the turbidity factor T and, in case of necessity, the air pressure p, in order to determine the intensity of solar radiation at any point on the earth's surface. This value times the sine of $h°$ gives, then, the intensity of the direct radiation on a horizontal surface. But in dealing with the total energy received at the sea surface, Q_0, it is also necessary to take into consideration the diffuse radiation from the sky as well as the direct radiation from the sun. From observations on the *"Maud"* expedition, Mosby (loc. cit.) obtains the following relationship between direct and total radiation:

$$D = 0.88 - \frac{2.4}{h} \qquad \text{I-(4)}$$

and for approximate values of T, obtains:

$$Q_0 = \frac{1}{D} Q_s \cdot \sin h \qquad \text{I-(5)}$$

It has been shown by Mosby, however, that the total incoming radiation on a horizontal surface at sea level can be related to the average solar altitude by the very convenient linear equation:

$$Q_0 = k\bar{h} \text{ g. cal./cm}^2 \text{ min.} \qquad \text{I-(6)}$$

The values of k vary somewhat with latitude within a range of 0.023 to 0.027 due to the varying water vapor in the atmosphere. The product $k\bar{h}$, then, gives the yearly average of the total incoming radiation with a clear sky in the stated units.

Ångstrom (1922), Kimball (1928) and others have determined an approximate relation between daily totals of solar radiation and the average cloudiness, the values for the ratio between radiation with clear sky and totally cloud-covered sky falling within the range 0.22 to 0.30. The differences are largely determined by the nature and thickness of the cloud cover and, perhaps, also by the percentage of cloudiness and solar altitude. Kimball presents 0.29 as the most likely average value. Using this ratio throughout he finds:

$$Q_t = Q_0 \, [0.29 + 0.71 \, (1 - \bar{C}/100)] \qquad \text{I-(7)}$$

where C is the percentage of sky covered by clouds. This allows the determination of the average incoming radiation for any latitude by means of the following formula:

$$Q_t = kh \, [1 - 0.71 \, (\bar{C}/100)] \qquad \text{I-(8)}$$

This formula, however, determines only that amount of heat incident upon the sea surface; a reduction is necessary to allow for the amount reflected back to the sky. According to Schmidt (1915), the percent reflected, r, varies according to latitude from a minimum of 3.3 at the equator to 8.0 at the poles. These data are given in tabular form by this author and also by Ångstrom (1920) and Mosby (1936). From this it follows that the total solar heat absorbed at the sea surface is:

$$Q_{ab} = Q_t \, (1 - r) \qquad \text{I-(9)}$$

Now, of the total heat absorbed by the water mass, some will be radiated back to space, the amount of which depends upon the absolute temperature of the water surface, the moisture content of the atmosphere and upon the nature and extent of cloud cover. Ångstrom (loc. cit.) has computed values for the back radiation at various temperatures and humidities and the data in the present calculations have been obtained from that source. As suggested, however, the values for the effective back radiation ($eff \; Q_b$) so obtained must be reduced to allow for the screening effect of any cloud cover present. According to Asklöff (1920), this correction factor amounts to $(1 - 0.0083 \, \bar{C})$. The final value must be further reduced (Ångstrom) by 6% to allow for internal reflection. Thus, the average back radiation is represented by the following formula:

$$Q_b = 0.94 \, [eff \; Q_b \, (1 - 0.0083 \, \bar{C})] \qquad \text{I-(10)}$$

The difference $Q_{ab} - Q_b$ represents, therefore, the amount of heat available in the sea for evaporation and for heating the air. It is necessary, as the first step in the computations, to disregard the advection factor. In order to determine the amount of heat used up in the evaporation process it is now only necessary to determine the proportion of energy given off as sensible heat to the atmosphere. All of the investigators of the problem referred to here have assumed a constant value for this proportion but it will be shown that the ratio Q_c/Q_e is highly variable, both with respect to season and with respect to the regional distribution.

II. THE EXCHANGE OF HEAT BETWEEN SEA AND ATMOSPHERE

Mosby (loc. cit.), concerning the exchange of sensible heat between sea surface and atmosphere, states:

„Um die mittlere Verdunstung bestimmen zu können, müssen wir nunmehr also noch diejenige Wärmemenge kennen, die durch turbulente Konvektion von der Meeresoberfläche zur Atmosphäre verfrachtet wird. Diese Wärmemenge hangt von vielen Verhältnissen ab, in erster Linie von dem Temperaturunterschied Wasser—Luft und von der Windstärke; es ist jedoch heute kaum möglich, sie gehau zu bestimmen."

In his determinations of evaporation from the oceans, Mosby assumes an arbitrary, but reasonable, value of 10% as the proportion of available energy given off as sensible heat to the atmosphere. He appears not to have been acquainted with the work of Bowen (1926), who has established the relation between evaporation and heat exchange at a water surface.

In considering the rates at which heat and water vapor are transported across the lower and upper surfaces of a volume of air above a body of water, Bowen has determined that the ratio, R, of the heat loss by conduction to that by evaporation can be obtained by means of the following formula:

$$R = 0.50\left(\frac{t_w - t_a}{e_w - e_a}\right)\frac{p}{760} \qquad \text{II-(1)}$$

where t_w and t_a are the temperatures in °C of the water and air, respectively, e_w and e_a are the vapor pressures in millimeters of the water surface and air and p is pressure (mm.). Bowen's derivation of the above formula is too lengthy to be reproduced here but the same equation can be derived quite simply as follows (Sverdrup et al, 1941):

Since the amount of water vapor, F, which is transported upward through the unit surface per second is

$$F = -A\frac{0.623}{p}\frac{de}{dz} \qquad \text{II-(2)}$$

where A is the eddy conductivity, then the amount of latent heat transported as water vapor is

$$Q_e = -L_t A\frac{0.623}{p}\frac{de}{dz} \qquad \text{II-(3)}$$

The amount of sensible heat transported away from the sea surface is

$$Q_c = -c_p A \frac{dt_a}{dz} \qquad \text{II-(4)}$$

where c_p is the specific heat of air at constant pressure. Therefore:

$$R = \frac{Q_c}{Q_e} = \frac{c_p}{L_t} \; \frac{p}{0.623} \; \frac{dt_a}{dz} \Big/ \frac{de}{dz} \qquad \text{II-(5)}$$

Setting $c_p = 0.237$ and $L_t = 585$, one obtains

$$R = 0.49 \; \frac{p}{760} \; \frac{dt_a}{dz} \Big/ \frac{de}{dz} \qquad \text{II-(6)}$$

Replacing the gradients dt_a/dz and de/dz by $(t_w - t_a)$ and $(e_w - e_a)$, one finds

$$R = 0.49 \; \frac{p}{760} \left(\frac{t_w - t_a}{e_w - e_a} \right) \qquad \text{II-(7)}$$

Since the present calculations require sea level pressures throughout, obviously the pressure term can be neglected, thus:

$$Q_c = RQ_e = \frac{RQ_a}{1 + R} \qquad \text{II-(8)}$$

$$Q_a - Q_c = \frac{Q_a}{1 + R} = EL_t \qquad \text{II-(9)}$$

and from (7) and (9)

$$E = \left[\frac{Q_a}{1 + 0.49 \left(\dfrac{t_w - t_a}{e_w - e_a} \right)} \right] \Big/ L_t \qquad \text{II-(10)}$$

III. COMPUTATIONS OF THE MEAN ANNUAL EVAPORATION FOR SELECTED AREAS IN THE NORTH ATLANTIC AND NORTH PACIFIC OCEANS BY MEANS OF ENERGY CONSIDERATIONS

In order to compare the values for evaporation obtained through the use of the energy equations given above with those determined by means of the more recent theoretical considerations of the interchange of water vapor and heat within the turbulent layer of the atmosphere, computations of the mean annual evaporation have been made for four selected areas in the North Atlantic and North Pacific oceans. Since, as a first step, it is necessary to disregard the advection factor, Q_v, the areas selected were those within which the dialatitudinal transport of ocean waters is at a minimum. For this reason, coastal areas have been eliminated from consideration. The two latitude ranges chosen were between 40° N. and 45° N. and between 20° N. and 25° N. In the North Atlantic, between 40° N. and 45° N., the evaporation was computed for the area bounded by meridians 20° W. and 50° W. and in the latitude range 20° N.–25° N., by the meridians 25° W. and 65° W. In the North Pacific, between 40° N. and 45° N., the computations have been made for the area bounded by meridians 140° W. and 160° E. and between latitudes 20° N.–25° N., by 130° W. and 140° E.

For the marine climatic data entering into the computations, the writer is indebted to the New Orleans and Washington, D. C., offices of the U. S. Weather Bureau for the loan of manuscript charts presenting recently compiled data on sea surface and air temperature, wet-bulb depression, wind speed and cloudiness over the oceans. Abridged forms of some of these charts appear in the *Atlas of Climatic Charts of the Oceans* (U. S. Weather Bureau, 1938).

The results of the computations are given in Table I. The average annual evaporation in the North Atlantic between latitudes 40° N. and 45° N. proves to be 84 cm/yr.; in the North Pacific, 68 cm/yr. Within the latitude range 20° N.–25° N., the evaporation is 168 cm/yr. in the Atlantic Ocean and 148 cm/yr. in the Pacific Ocean. These results would indicate that evaporation is greater at all latitudes in the North Atlantic than in the North Pacific. The value for the latitude range 40° N.–45° N. in the Atlantic is 16 cm/yr. greater than for the similar latitude range in the Pacific while it is 20 cm/yr. greater for the latitude range 20° N.–25° N. Analysis of the data in Table I indicates that the higher evaporation in the North Atlantic is largely due to the smaller amount of heat lost to the atmosphere rather than to any significant excess of energy received. The mean annual cloudiness within the latitude range 40° N.–45° N. is, however, somewhat higher in the North Pacific than in the North Atlantic which would account for a small proportion of the difference in evaporation at these higher latitudes.

TABLE I—Computations of Evaporation for

	Lat.	*Long.*	k	$\bar{h}°$	$\bar{C}$ (%)	Q_t/Q_o	Qt	r (%)	Q_{ab}
Atlantic	40°N–45°N	25°W–10°W	0.024	14.7	64	0.546	0.193	4.8	0.184
Ocean	20°N–25°N	25°W–65°W	0.023	18.2	46	0.673	0.282	3.8	0.271
Pacific	40°N–45°N	140°W–160°E	0.024	14.7	70	0.506	0.178	4.8	0.169
Ocean	20°N–25°N	130°W–140°E	0.023	18.2	46	0.675	0.283	3.8	0.272

IV. COMPUTATIONS OF THE ANNUAL EVAPORATION BY MEANS OF AVAILABLE DATA ON WIND SPEEDS, HUMIDITIES AND TEMPERATURES OVER THE OCEANS

As an outgrowth of the researches on turbulent mixing in the lower layers of the atmosphere, particularly by Taylor (1932), Prandtl (1935), von Kármán (1935), Rossby (1932; 1935) and Sverdrup (1936), it has been possible to prepare formulae showing evaporation as a function of moisture concentration and wind movement within the turbulent layer. Such formulae have been constructed by Sverdrup (1937) and Montgomery (1940) but since their discussions are presented in readily available publications, no attempt will be made in this paper to enter into the derivations of the equations.

However, the basic assumptions are that the transport of water vapor between sea surface and atmosphere takes place by ordinary diffusion through a shallow boundary layer and by eddy conductivity above this layer. It is also shown that for the unstable atmosphere, i. e., where the potential temperature decreases with height, the eddy conductivity, itself, is a linear function of height and of the roughness character of the sea surface. It is also shown that, under these conditions, the vapor pressure is a function of the logarithm of height. According to Sverdrup:

$$E_2 = \frac{H \dfrac{0.623}{p} \rho \; (e_u - e_a) W}{\dfrac{H}{K_0} \ln \dfrac{a + Z_0}{d + Z_0} + dw} \qquad \text{IV-(1)}$$

where H is the diffusion coefficient, p is the atmospheric pressure, ρ is density, a altitude of observation, k_0 is von Kármán's constant (0.4), W is the friction velocity where

$$W = \sqrt{\frac{\tau}{\rho}} = \frac{k_0}{\ln \dfrac{a + z_0}{z_0}} \; W_a = \frac{0.165}{\log \dfrac{a \pm z_0}{z_0}} \; W_a$$

Selected Areas by Means of the Energy Equations

$\bar{t}_w$ (°C)	$\bar{t}_a$ (°C)	$\bar{e}_a$ (mm)	$\bar{e}_w$ (mm)	$effQ_b$	Q_b	$Q_{ab}-Q_b$	R	Q_c	Q_e	$\bar{L}_t$	E_1 (cm/yr.)	
16.4	16.1	11.9	13.7	0.187	0.083	0.101	0.076	0.007	0.094	586.9	84.0	*Atlantic*
24.8	25.2	19.8	23.0	0.166	0.096	0.175	−0.056	−0.010	0.186	582.3	167.5	*Ocean*
11.1	10.4	8.5	9.8	0.186	0.074	0.095	0.249	0.019	0.076	589.8	67.7	*Pacific*
24.6	23.9	18.6	22.8	0.166	0.097	0.176	0.073	0.012	0.164	582.4	148.2	*Ocean*

W_a is the velocity at level a and z_0 is the roughness parameter (0.6,—Rosby) and

$$dW = \frac{dW_a\ k_0}{\ln \frac{a}{z_0}} = 4.12$$

According to Montgomery:

$$E_2 = \rho k_0\ \gamma_a\ \Gamma\ (q_w - q_a)\ W_a \qquad \text{IV-(2)}$$

where $\gamma_a = \dfrac{k_0}{\ln \frac{a}{z_0}}$ and Γ is his "evaporation coefficient." Both γ_a and Γ depend upon the hydrodynamic character of the sea surface but the dependence is such that the product $\gamma_a\ \Gamma$ is very nearly constant for either rough or smooth surface. If this is true, equation IV-(2) can be reduced to a formula of the type:

$$E_2 = K\ (e_w - e_a)\ W_a \qquad \text{IV-(3)}$$

Equation IV-(1) also reduces to a formula of this type.

Since $d = \dfrac{\ln \frac{a}{z_0}\, dW}{k_0\, W_a} = \dfrac{71.2}{W_a}$

when $z_0 = 0.6$ and $a = 600$

Then, at $W_a = 500$; $d = 0.14$

$W_a = 1500$; $d = 0.05$

Taking $H = 0.235$ and $p = 1000$, from equation IV-(1):

$$E_2 = \frac{0.235 \cdot 0.623 \cdot 1.2 \times 10^{-6} \cdot \frac{0.4}{6.9}\ (e_w - e_a)\ W_a}{0.6 \cdot 6.6 + 4.1}$$

$$= 1.25 \times 10^{-9}\ (e_w - e_a)\ W_a \qquad \text{IV-(4)}$$

where e is in millibars and W_a is in meters per second at a height of six meters.

The 24-hour evaporation in millimeters or decigrams is therefore given by:

$$E_2 = 0.108\ (e_w - e_a)\ W_a \qquad \text{IV-(5)}$$

Equation IV-(1) applies to the adiabatic atmosphere but under conditions of atmospheric stability, the computed values for E_2 will be somewhat too great. However, when considering evaporation over periods of a day or longer, no serious error is introduced as the amounts computed under stable conditions are small when compared with those computed under adiabatic conditions.

Through the use of equation IV-(5), computations of the mean annual evaporation have been made for the same four areas in the North Atlantic and North Pacific. The results are given in Table II. As

TABLE II—MEAN ANNUAL EVAPORATION FOR SELECTED AREAS COMPUTED BY MEANS OF EQUATION (5), E_2, COMPARED WITH THE MEAN ANNUAL EVAPORATION, E_1, COMPUTED BY MEANS OF THE ENERGY EQUATIONS

	Lat.	*Long.*	E_1 (*cm/yr.*)	E_2 (*cm/yr.*)	E_1/E_2	E ($E_2 \times 1.32$) (*cm/yr.*)
North Atlantic	40°N–45°N	20°W–50°W	84.0	88.3	0.95	116.6
North Atlantic	20°N–25°N	25°W–65°W	167.5	94.2	1.78	124.3
North Pacific	40°N–45°N	140°W–160°E	67.7	55.8	1.21	73.7
North Pacific	20°N–25°N	130°W–140°E	148.2	113.2	1.31	149.4

was expected, the values derived by the latter method are lower than those computed by means of the energy equations. The mean annual evaporation for the four areas proves to be 87.9 cm/yr. when computed by equation IV-(5) and 116.8 cm/yr. when computed by means of the energy equations. Most of this difference can no doubt be assigned to the inaccuracies of the humidity measurements at sea for it would be expected that the mean of a large number of such observations would be too high, the tendency of error in the individual observations being in this direction. This would result, of course, in the computed values of evaporation being too low. However, if these differences resulting from the inadequacy of the observational data are consistent, it is possible to prepare a correction factor to be applied to equation IV-(5) such that the two methods can be brought into agreement. The ratios E_1/E_2 given in the sixth column of Table II, indicate that this factor is of the order of 1.32. Thus equation IV-(5) becomes:

$$E = 0.143\ (e_w - e_a)\ W_a \qquad \text{IV-(6)}$$

It now becomes of interest to examine the nature and causes of the remaining differences $(E_1 - E)$.

In the use of the energy equations under section III for computing E_1, it was necessary to disregard the advection term, Q_v, except to the extent that the sea surface temperature data entered into the computations of Q_b and R. For this reason, it would be expected that the evaporation computed by such methods would be too low at high latitudes and too high at low latitudes. An examination of the data in the last column of Table II shows this to be particularly true in the case of the North Atlantic where the dia-latitudinal transport of surface waters is greater than in the North Pacific. Thus in the North Atlantic, on account of the great northerly transport in the west portion, the greater heat transport is from south to north. In the case of the North Pacific, on the other hand, the northerly transport in the west portion is more nearly balanced by the southerly transport in the east portion. If we set

$$Q_v \cong \Delta Q_{1,2} = (E\ L_t + Q_c) - (E_1 L_t + Q_{c1})$$

then for the Atlantic between latitudes 40° N–45° N., Q_v becomes 20533 g. cal. cm²/yr. and for the latitude range, 20° N.–25° N., — 23811 g. cal. cm²/yr. In the Pacific, on the other hand, Q_v is only 444; g. cal. cm²/yr. for the latitude range 40° N.–45° N. and still shows a slight positive net heat transport between latitudes 20° N.–25° N. of 626 g. cal. cm²/yr. The ratios of Q_v for North Atlantic and North Pacific indicate that the heat transport into the latitude range 40° N.–45° N. of the Pacific is only 21% as great as the similar transport in the Atlantic within the same latitude range, while in the latitude range 20° N.–25° N. in the Pacific, the inflow of heat is very nearly balanced by the outflow. These results are in line with present information concerning the transport of surface waters in both oceans. It should perhaps be repeated, that since the areas were so chosen as to exclude the coastal sections, the net heat transport for the entire latiude ranges represented would differ from the figures given above.

V. COMPUTATIONS OF THE SEASONAL EVAPORATION, HEAT EXCHANGE AND TOTAL ENERGY EXCHANGE FOR FIVE-DEGREE SQUARES IN THE NORTH PACIFIC AND NORTH ATLANTIC

Through the use of equation IV-(6), it is now possible to compute the mean seasonal evaporation for five-degree squares in the North Atlantic and North Pacific by means of the previously-mentioned data on sea surface and air temperatures, wet-bulb depressions and wind speeds over the oceans. The values for e_w have been obtained by multiplying the vapor pressures corresponding to the surface temperature t_w, by 0.98 in order to take the effect of salinity into account. The values for e_a have been obtained from the Smithsonian Tables which present vapor pressure as a function of the difference between wet- and dry-bulb temperatures.

Then from the evaporation quantities so determined, and the values of $t_w - t_a$ and $e_w - e_a$, the seasonal and annual values for Q_c and Q_a have been computed for the same areas.

Since the climatic data at hand presented temperatures and wet-bulb depressions in degrees Fahrenheit and the wind speeds in knots, it was convenient to adjust the several equations to the same units such that:

$$E = 0.249\ (e_w - e_a)\ w_a \qquad \text{gms. cm}^2\text{/day} \qquad \text{V-(1)}$$

$$Q_e = 145.4\ (e_w - e_a)\ w_a \qquad \text{gm. cal. cm}^2\text{/day}\ (L = 585) \qquad \text{V-(2)}$$

$$Q_c = 0.01 \left(\frac{t_w - t_a}{e_w - e_a} \right) Q_e \qquad \text{gm. cal. cm}^2\text{/day} \qquad \text{V-(3)}$$

$$Q_a = Q_e + Q_c \qquad \text{gm. cal. cm}^2\text{/day} \qquad \text{V-(4)}$$

in which e is in inches-Hg, t is in degrees Fahrenheit and w_a is in standard knots.

The charts prepared from the mean seasonal and annual values of E and Q_c computed for five-degree squares in the North Pacific and North Atlantic are primarily of interest in the meteorological field which is not only concerned with the amount of heat added to or subtracted from the lower layers of the atmosphere, but also of the amount of latent energy in the form of water vapor which is rendered available for the various meteorological processes. On the other hand, the charts showing the seasonal and regional values of Q_a are of particular interest to the field of oceanography since the loss or gain of energy from the sea surface through either evaporation or conduction is entirely in the form of heat. The charts for E are of interest in this field only to the extent that the evaporation, through its effects upon the surface salinity, serves to alter the distribution of density within the oceans. However, it is not practicable to include the 15 charts necessary to show the seasonal and annual values of E, Q_c and Q_a in the present short paper. It is expected that these will be presented in a later, and more complete, report.

An examination of the charts for E shows that the evaporation in both oceans is nearly everywhere greatest in winter and least in summer; that the regions of greatest evaporation are on the western sides of the oceans and about the southern margins of the North Atlantic and North Pacific semi-permanent high pressure fields. Evaporation on the eastern sides of the oceans is generally smaller. The absolute maximum evaporation for all oceans occurs in the North Atlantic in winter within the Gulf Stream between latitudes 35° N.–40° N. and about 700 miles east of the Virginia Capes, where the average evaporation is about 1.14 gms. cm^2/day. The maximum value in the North Pacific, 0.94 gms., on the other hand, occurs farther south and nearer the coast within the Kuroshio between latitudes 25° N.–30° N. and approximately 300 miles northeast of Formosa. The evaporation in winter along the eastern sides of both oceans is relatively much smaller, for example, in the North Pacific it is only 0.15 gms. cm^2/day off the coast of British Columbia and Central California and 0.14 gms. off the Southern California Coast. Although the values increase slightly southward, the evaporation is still only 0.18 gms. in the Gulf of Panama. In the North Atlantic off the Irish Coast, the winter value is 0.30 gms. and off the northern coast of Africa, 0.36 gms. The average winter evaporation in the Gulf of Mexico and the Caribbean Sea is of the order of 0.40 gms.

In spring the evaporation decreases considerably along the western sides of the oceans, the influence of the Kuroshio upon evaporation in the North Pacific having almost disappeared while the similar effect of the Gulf Stream in the North Atlantic has become considerably less. The evaporation in the south-central portion of the tropical North Pacific, however, increases slightly during this season and reaches its maximum value. The evaporation along the equator in the eastern North Atlantic also increases slightly but at the same time decreases to the minimum seasonal value within the corresponding region in the North Pacific. In the higher latitudes, the central and western portions of the equatorial regions and along the eastern sides of the oceans, the decrease in evaporation is small.

In summer the evaporation reaches its lowest value over most of the North Pacific and North Atlantic. The values, however, increase slightly during this season within the eastern equatorial regions of both oceans and reach their maximum at this time. The decrease in evaporation within the tropical belts of high evaporation is small. The quantities, nevertheless, remain positive for all areas except in the North Pacific north of latitude 55° and west of 160° W., and in the North Atlantic waters immediately surrounding Labrador, where a slight net condensation is computed for this season. The effect of the Kuroshio in the western North Pacific has now completely disappeared and the computed values of evaporation for the eastern and western sides of this ocean are of the same order of magnitude. The effects of the Gulf Stream upon evaporation are, however, still in evidence and the evaporation on the west side of the North Atlantic remains greater than in the eastern portions.

In the autumn the evaporation nearly everywhere increases and is greater than during either spring or summer except in the south-central tropical regions in both oceans where the minimum values for these respective areas occur. Evaporation also decreases from its summer maximum in the eastern equatorial regions of both oceans. The effect of the Kuroshio upon evaporation in the North Pacific again becomes evident and evaporation within the Gulf Stream of the North Atlantic now reaches values as high as 0.88 gms. off the Virginia Capes.

Thus, in general, it is shown that the regions of greatest evaporation are those within which the northerly transport of surface waters is greatest, i. e., along the western sides of the oceans within the Kuroshio and Gulf Stream. Conversely, the regions of least evaporation are those of southerly flow or drift, namely, along the eastern sides of the oceans and the extreme northwestern coastal areas. The only exceptions to this general statement appear within the south-central tropical areas of both oceans where, in the northern part of the trade wind region, a relatively high evaporation is associated with the dry descending air currents accompanying the semi-permanent fields of high pressure, and in the eastern equatorial regions, where relatively low values are associated with the northerly flowing ocean currents which in these regions cross the equator resulting in an influx of cold water from the higher latitudes in the Southern Hemisphere.

In the North Atlantic Ocean, the center of the tropical or trade-wind area of maximum evaporation remains nearly stationary during all seasons except autumn, when it quite largely disappears, with its mean position located between latitudes 10° N.–20° N. and approximately one-third the distance between the coasts of South America and Africa. In the North Pacific Ocean, on the other hand, this tropical area of high evaporation is considerably better developed and its center varies in position with the seasons. In winter the maximum evaporation within this area is 0.70 gms. cm^2/day and the center is located between 5° N.–10° N. and 170° W.–175° W. In spring the maximum increases to 0.72 gms. and the center is displaced northeastward to the area between 10° N.–15° N. and 150° W.–155° W. In summer the maximum evaporation in this region decreases to 0.62 gms. and the center is further displaced northeastward to the area between 15° N.–20° N. and 145° W.–150° W. In the autumn, however, the maximum evaporation within this area falls to its lowest value (0.56 gms.) and the now less well-defined center is displaced southward to the general region on the equator between latitudes 0° N.–5° N. and between 135° W.–150° W. This clockwise migration of the tropical center of high evaporation in the North Pacific appears to correspond to the similar movement of the center of the North Pacific high pressure field.

The average annual values of evaporation for the various latitude ranges are given in Figure 18 and are compared with similar values presented by other investigators. The computed annual values appear to be in excellent agreement with those given by Wüst (1936); the ratio $E/E_{\text{Wüst}}$ for all latitudes being 1.00 for the North Atlantic, 1.06 for the North Pacific and 1.03 for both oceans. The agreement with McEwen's (1938) values is also good but the values are lower than those given by Cherubim (1931). The agreement with Mosby's (1936) values is good for the latitude ranges north of 25°, but his values are considerably higher for the lower latitudes. The greatest departures between the present values and those previously determined by other investigators of the problem, appear within the latitude range 45° N.–50° N., where the previous values given have been somewhat higher, and within the latitude range 35° N.–40° N., where the previous values assigned have been lower. In general, these curves show that evaporation in the North Atlantic is at a maximum within the latitude range 35° N.–40° N. with a well defined secondary maximum between latitudes 15° N.–20° N. In the North Pacific, the maximum annual evaporation occurs between latitudes 20° N.–25° N. with evidence of a slight secondary maximum between latitudes 35° N.–40° N. With regard to the curves shown, it is perhaps well to remark that too much dependence should not be placed upon the values north of latitude 55° since the areas involved in the computations for these regions are rather small.

The isolines for Q_c over the oceans show, roughly, the same configuration as shown for E, these values being at their maximum in winter and along the western sides of the oceans, with the major exception that there exist no tropical areas of maximum heat exchange to coincide with the areas of maximum evaporation which appear associated with the North Pacific and Azores high pressure systems.

The absolute maximum value for Q_c (260 g. cal. cm^2/day) occurs in winter in the North Atlantic in about the same location as the center of maximum evaporation. In the North Pacific, the maximum winter value (240 g. cal. cm^2/day) occurs between 35° N.–40° N. and 150° E.–155° E. with a secondary area of maximum heat exchange northeast

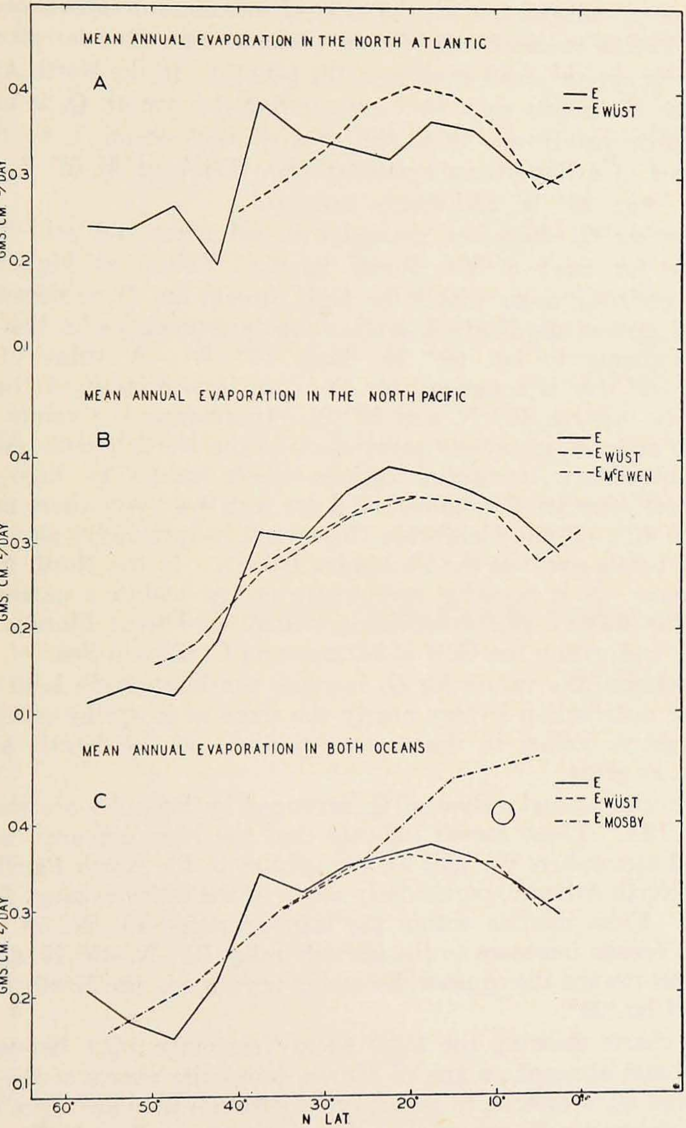

Figure 18. A. The mean annual evaporation in the North Atlantic compared with the values previously determined by Wüst. B. The mean annual evaporation in the North Pacific compared with the values previously determined by Wüst and McEwen. C. The mean annual evaporation in both oceans compared with the values previously determined by Wüst and Mosby for all oceans.

of Formosa coinciding with the area of maximum evaporation. The North Pacific values for Q_c during winter are positive for all regions except for several small areas near the equator. In the North Atlantic, however, the charts show that even during the winter, Q_c is negative for approximately one third the area of that ocean, i. e., roughly southeast of a line drawn northeast from Trinidad, B. W. I., to lat. 40° N., long. 20° W. and thence eastward.

In spring Q_c decreases generally in all areas but still remains positive for most of the North Pacific. Values as high as 116 g. cal. cm^2/day occur within the Gulf Stream but Q_c is negative for all that area of the North Atlantic roughly southeast of a line drawn from Panama to lat. 60° N., long. 10° W. A value of -57 g. cal. cm^2/day is computed for the area immediately off the coast of Africa between 15° N. and 20° N. In summer the values for Q_c become insignificant nearly everywhere in the North Pacific except in the southeastern equatorial regions where values as high as 40 g. cal. cm^2/day are determined, and are negative everywhere north of latitude 40°, off the California Coast and, surprisingly, also off the Asiatic coasts south to the Philippine Islands. In the North Atlantic in summer, Q_c is negative everywhere except within a narrow zone along the western side extending northeastward from Florida to lat. 40° N., and within the Gulf of Mexico and Caribbean Sea.

In autumn, the values for Q_c increase considerably in both oceans and the distribution is very nearly the same as in spring except that the negative values in the eastern portions of the North Atlantic are not as great.

The mean annual values of Q_c arranged by latitudes are shown in Figure 19A. These curves indicate that the heat exchange between sea and atmosphere per unit area is greater in the North Pacific than in the North Atlantic, particularly south of the latitude range 35° N.–40° N. From minima within the latitude range 45° N.–50° N., Q_c in both oceans increases to the latitude range 35° N.–40° N. and then decreases toward the equator, becoming negative in the North Atlantic south of lat. 25°.

The charts showing the total energy exchange (Q_a) between sea surface and atmosphere are, of course, composite charts of the values of Q_e and Q_c, therefore, in general configuration they are quite similar to the charts for E, except that the west-east gradient in the energy exchange is intensified due to the high values for Q_c over the western sides of the oceans and the low values over the eastern sides. Conspicuous areas of maximum energy exchange over the south-central tropical areas of both oceans coincide with the tropical areas of high

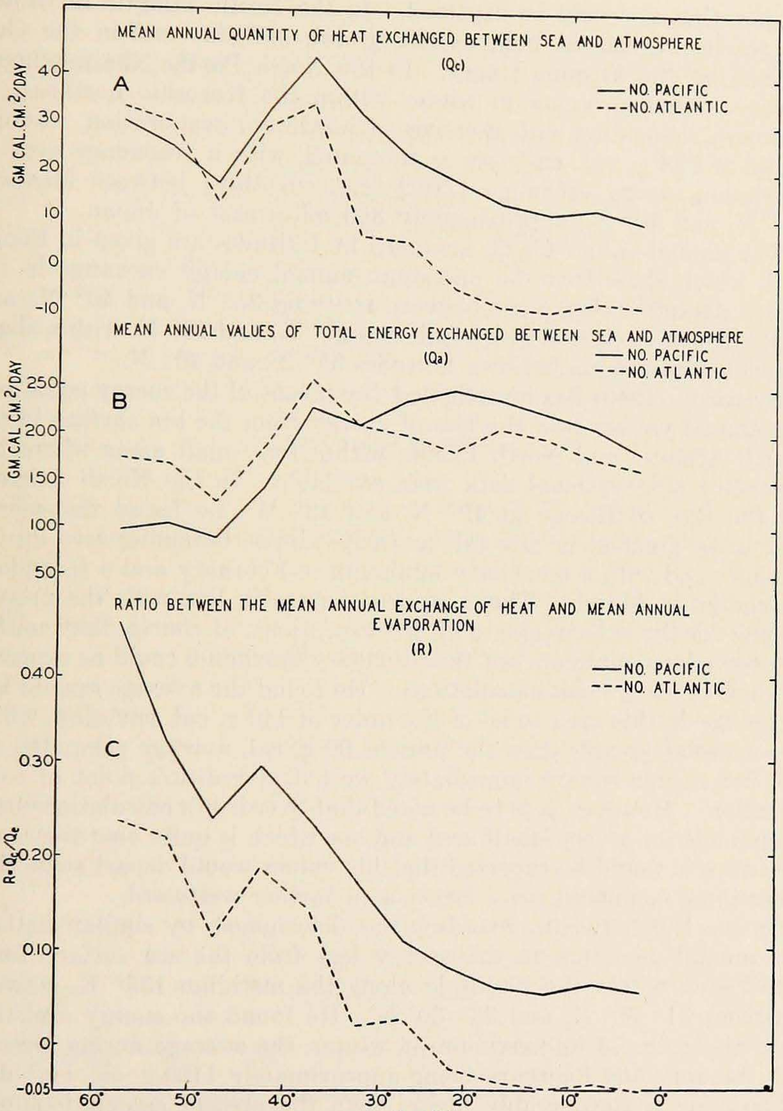

Figure 19. A. Mean annual values for Q_o in the North Atlantic and North Pacific oceans. B. mean annual values for the total energy exchange (Q_a) in the North Atlantic and North Pacific oceans. C. Ratio (R) between the mean annual values for Q_o and Q_e in the North Atlantic and North Pacific oceans.

evaporation as would be expected. In the North Atlantic in winter, Q_a reaches values as high as 937 g. cal. cm^2/day within the Gulf Stream off the Virginia Capes. In the North Pacific, the maximum energy exchange occurs in winter within the Kuroshio northeast of Formosa, coinciding with the area of maximum evaporation, where a value of 728 g. cal. cm^2/day is computed, with a secondary area of maximum energy exchange (718 g. cal. cm^2/day) between latitudes 35° N. and 40° N. approximately 350 miles east of Japan.

The annual values for Q_a arranged by latitudes are given in Figure 19B, which show that the maximum annual energy exchange in the North Atlantic takes place between latitudes 35° N. and 40° N., and in the North Pacific, between latitudes 20° N. and 25° N. with a slight secondary maximum between latitudes 35° N. and 40° N.

Sverdrup (1940) has investigated (by means of the energy equation) the annual variation of the loss of energy from the sea surface in the North Atlantic and North Pacific within two small areas where the necessary observational data were available. In the North Atlantic off the Bay of Biscay at 47° N. and 12° W., he found the energy loss to be greatest in late fall or early winter, becoming zero during summer and with a secondary minimum in February and a secondary maximum in March. These conclusions are in line with the present results for the same region with the exceptions, of course, that neither the secondary minimum nor the secondary maximum could be expected to show in the present calculations. He found the average annual loss of energy in this area to be of the order of 140 g. cal. cm^2/day, which is somewhat greater than the present 90 g. cal. average computed for the five degree square immediately west of Sverdrup's point of computation. However, it is to be noted that Sverdrup's calculations have been made for a very small area and one which is quite near the coast, therefore it would be expected that his values would depart somewhat from those computed for a larger area farther westward.

In the North Pacific, Sverdrup has determined, by similar method, the annual variation in the energy loss from the sea surface for a small area within the Kuroshio along the meridian 135° E. between latitudes 31° 50′ N. and 33° 30′ N. He found the energy exchange here also to be at its maximum in winter, the average during December, January and February being approximately 1100 g. cal. cm^2/day. This value is considerably higher than the present calculations give for the same general area, i. e., about 600 g. cal. However, again it must be considered that Sverdrup's computations have been made only for a very small area within which the temperature gradients were found to be very steep and should, therefore, give greater values than those computed for the five degree square.

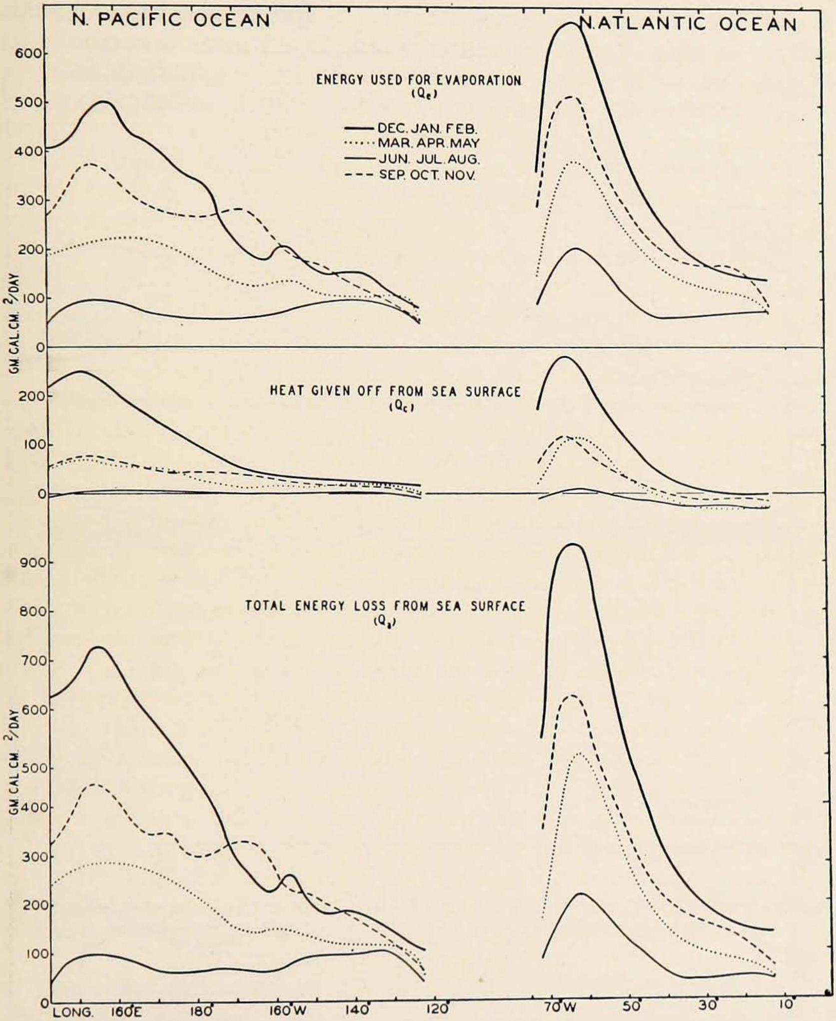

Figure 20. Seasonal variations in evaporation, heat exchange and the total energy exchange within the latitude range 35° N.–40° N. in the North Atlantic and North Pacific oceans.

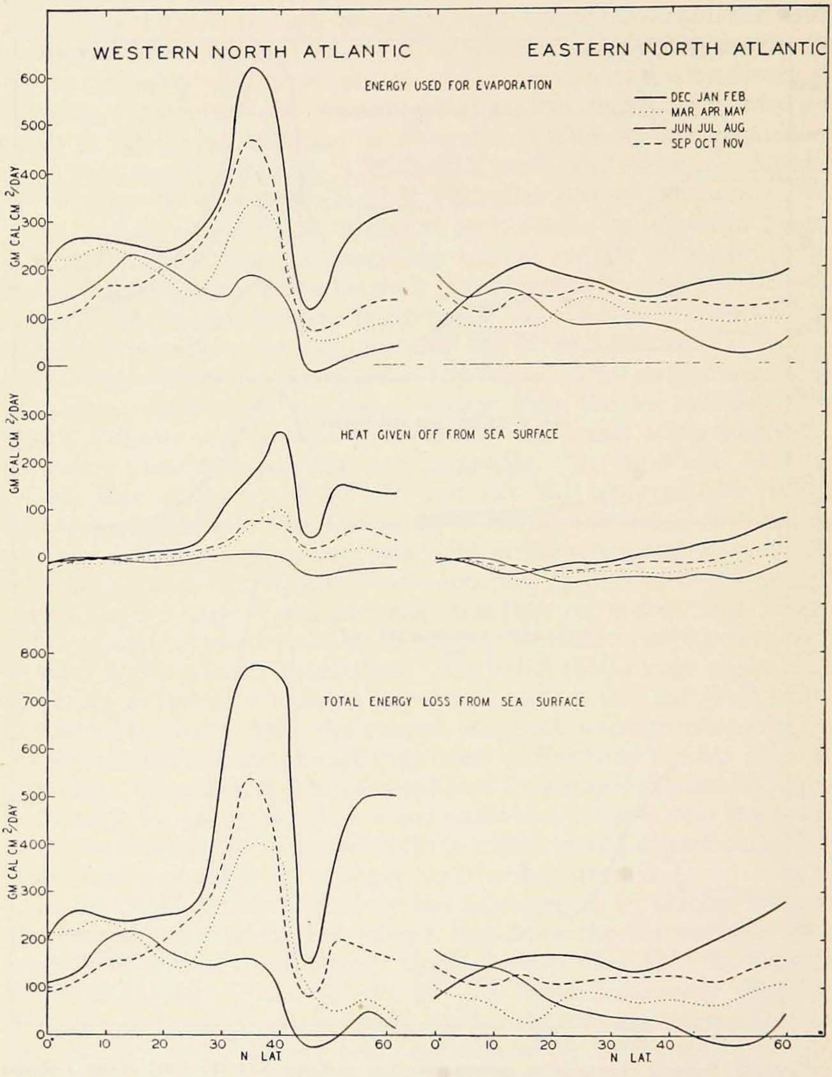

Figure 21. Seasonal variations in evaporation, heat exchange and the total energy exchange according to latitude along the western and eastern sides of the North Atlantic ocean.

The seasonal variations in the quantities Q_e, Q_c and Q_a across the North Pacific and North Atlantic within the latitude range 35° N.–40° N. are shown in Figure 20. These curves serve to illustrate the large seasonal variations in these quantities over the western portions of both oceans, i. e., within the Kuroshio and Gulf Stream, and the relatively much smaller seasonal variations over the eastern portions.

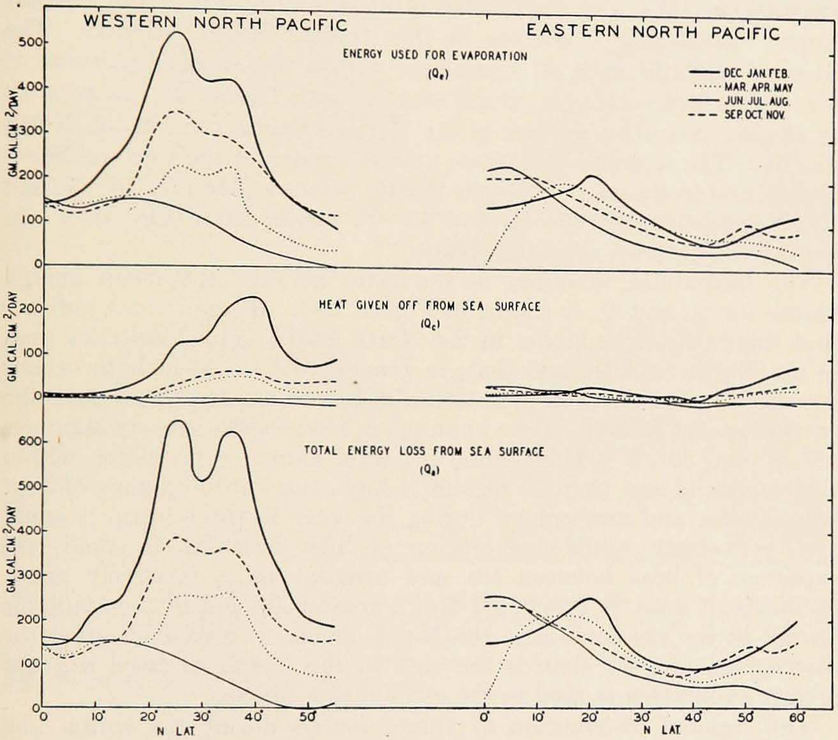

Figure 22. Seasonal variations in evaporation, heat exchange and the total energy exchange according to latitude along the western and eastern sides of the North Pacific ocean.

The seasonal variations in evaporation, heat exchange and the total energy exchange according to latitude along the western and eastern sides of both oceans are illustrated in Figures 21 and 22.[1] These curves show that, in general, the greatest seasonal variations of all

[1] In view of the facts that these north-south cross sections tend to parallel the isolines for Q_e, Q_c and Q_a, and that the selections of the lines of cross section were arbitrary ones, the curves shown in Figures 21 and 22 have been smoothed somewhat in order to eliminate the minor, non-significant variations.

three quantities occur in mid-latitudes along the western sides of the oceans; that the seasonal variations in the equatorial regions in both oceans are small with the exception that in the eastern North Pacific a very decided minimum in spring is shown. It is also indicated that in the western portions of both oceans, the areas of maximum heat exchange (Q_c) occur farther north than do the areas of maximum evaporation (Q_e), the displacement being 15 degrees in the western North Pacific and 5 degrees in the western North Atlantic. The seasonal variations in all quantities within the eastern portions of the oceans are relatively much smaller and, except at the equator as already described, greater in the North Atlantic than in the North Pacific. The seasonal variations at the equator in the western North Pacific and in the eastern North Pacific between latitudes 10° N. and 15° N. and between 25° N. and 45° N. are so small that they can scarcely be considered as significant.

The latitudinal variation in the ratio between the mean annual values for Q_c and Q_e is shown in Figure 19C. These curves indicate that this ratio, *R*, is higher in the North Pacific at all latitudes than in the North Atlantic and that, in general, the ratios in both oceans decrease from maxima at high latitudes to minima at or very near the equator except for well-defined minima in both oceans between latitudes 45° N. and 50° N. Thus, while the total annual evaporation within the equatorial and tropical regions is high, the total exchange of heat between sea and atmosphere during the year in these areas is small and, conversely, while evaporation at high latitudes is small, the exchange of heat between sea and atmosphere is relatively great. Similarly, it must be concluded that a greater amount of the available energy at the sea surface in the North Pacific is used in heating the atmosphere directly than is the case in the North Atlantic where a greater proportion is used in the evaporation process.

The regional distribution of values for Q_a during the winter and summer seasons is shown in Figures 23A and 23B. The charts illustrated are quite similar in configuration to those for *E* during the same seasons and also, in a qualitative manner, quite similar to those for Q_c except that in the latter case, as mentioned previously, there exist no maximum areas of heat exchange within the tropical trade-wind region to correspond to those areas of maximum *E* and Q_a within this belt.

The locations of the principal frontal zones as given by Petterssen (1940) are also shown in Figure 23B and it is interesting to note that these zones correspond quite closely with the zones of maximum energy exchange. However, the lines or zones of frontogenesis (cyclogenesis) themselves, will coincide more nearly with the zones of the maximum temperature gradient thus, in the western portions of the oceans, the principal frontogenetic areas will be located along the eastern coasts of the two continents northwestward from the mean position of the polar fronts themselves. This, according to Petterssen (loc. cit.), is due to the fact that the zones of frontogenesis will tend to move with the air currents toward the axis of outflow and will assume a mean

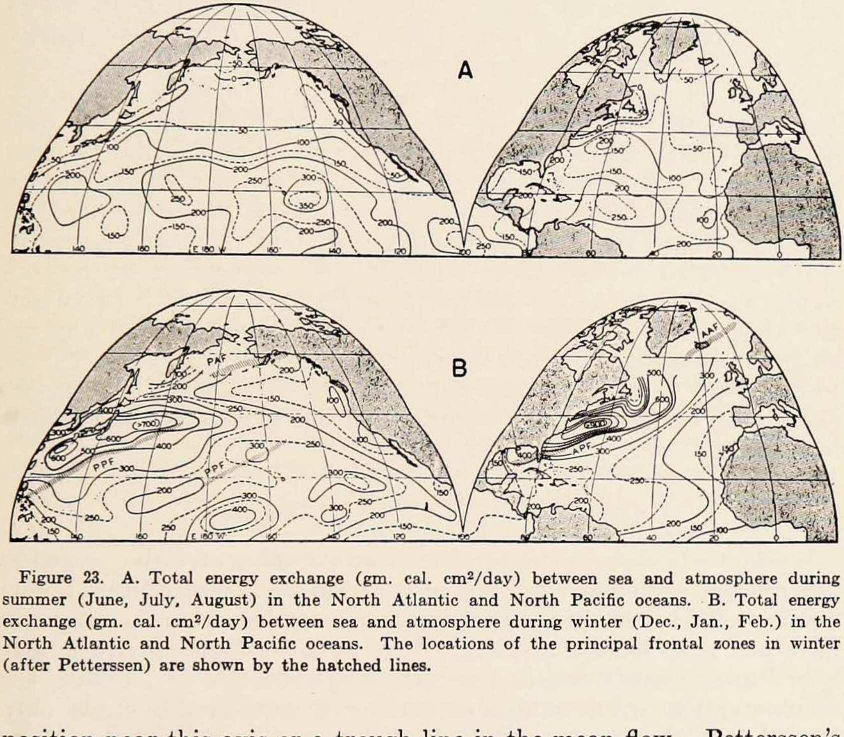

Figure 23. A. Total energy exchange (gm. cal. cm²/day) between sea and atmosphere during summer (June, July, August) in the North Atlantic and North Pacific oceans. B. Total energy exchange (gm. cal. cm²/day) between sea and atmosphere during winter (Dec., Jan., Feb.) in the North Atlantic and North Pacific oceans. The locations of the principal frontal zones in winter (after Petterssen) are shown by the hatched lines.

position near this axis or a trough line in the mean flow. Petterssen's position for the principal frontogenetic zone in the western North Pacific corresponds very closely with the position of the zone of maximum cyclone frequency as determined by Richardson (1936). It is found then, that in both oceans the zones of frontogenesis (cyclogenesis) will in every case lie to the northwestward of a zone of maximum energy exchange and it will be noted (see Figure 23B), in addition, that the major axes of the zones of maximum energy exchange will parallel the principal axes of the frontal zones.

Before concluding the discussion of results, it should be emphasized that the mean seasonal data presented in this report can by no means be taken as representative of instantaneous conditions over the oceans. It is expected that over given intervals of time of a year or less, the actual values for E, Q_c and Q_a will depart widely from the values given herein, particularly over the western portions of the oceans, and that the positions of the areas of maximum or minimum energy exchange will also vary. Thus over the Kuroshio and Gulf Stream, at the times of outbreak of cold continental air in winter, these values may occasionally be extremely high; conversely, during periods in winter when the polar fronts are located near the coast and the air masses over the neighboring oceans are of southerly origin, and quite warm and moist, the energy exchange in these areas must reach low values. Also, along the eastern shores, the energy exchange in winter may reach high values at the times of outbreaks of dry continental air from the interior. So, just as there are important non-periodic variations in the positions of the principal centers of action in the atmosphere, the polar fronts, etc., with the resulting time or regional changes in humidity, wind, temperature, cloudiness, etc., so must there be corresponding changes in the amount of energy exchanged between sea and atmosphere. In light of such reasoning, it appears that any attempts toward a solution of the problems of long-period weather forecasting must be concerned not only with the dynamics of the atmosphere (and sea) but of the thermodynamical aspects of the problem as well.

SUMMARY AND CONCLUSIONS

While it has long been known that evaporation over the oceans is governed by humidity, sea surface temperature and wind speed, empirical methods used in the past to relate evaporation to these factors have not been successful, largely because the equations either contained unevaluated functions, or were constructed to fit a special set of observations. The multiplicity of such equations, each of only limited use, indicates the difficulty of the evaporation problem. The present investigation, however, offers a method for the proper evaluation of such an equation through a comparison with evaporation computed by means of the energy equations when using long-period climatic data over the oceans. The practicability of the technique has been demonstrated and the following evaporation equation presented:

$$E = 0.143\ (e_w - e_a)\ W_a \quad \text{mm/24 hours.}$$

where the vapor pressure is in millibars and the wind speed in meters per second at the height $a = 6$ meters above the sea surface.

It is further shown that the quantity of heat exchanged between sea and atmosphere can be computed by means of the following equation:

$$Q_c = 0.65 \left(\frac{t_w - t_a}{e_w - e_a} \right) Q_e \quad \text{where } p \text{ is const.} = 1000 \text{ mb.}$$

By means of these equations and using available seasonal climatic data over the North Pacific and North Atlantic, computations have been made of the amount of evaporation and of the quantity of heat and total energy exchanged between sea and atmosphere within each five-degree square during the several seasons. Some of the more important conclusions drawn from these computations are that:

1. Evaporation over the North Atlantic and North Pacific is nearly everywhere greatest in winter, least in summer and somewhat greater in autumn than during spring. The only major exceptions are on the equator in the eastern portions of both oceans, where the maximum occurs in summer, and in the south-central tropical North Pacific, where there is a slight maximum in spring.

2. The regions of greatest evaporation are those of northerly flowing ocean currents, i. e., along the western sides of the oceans within the Kuroshio and Gulf Stream. The only exceptions in this case occur within the tropical trade wind belts of both oceans where a high evaporation is associated with the dry descending air currents in these regions, and in the eastern equatorial belts of both oceans where a low evaporation is associated with the northerly flowing cold currents from the Southern Hemisphere which cross the equator in these regions. In winter, evaporation reaches values as high as 1.14 gms. cm^2/day in the Gulf Stream and 0.94 gms. cm^2/day within the Kuroshio. In mid-latitudes along the eastern sides of the North Pacific, the winter evaporation is of the order of 0.15 gms. cm^2/day, and in the eastern North Atlantic, 0.25 gms. cm^2/day.

3. The zones of maximum annual evaporation in the North Atlantic occur farther north than in the North Pacific, viz., within the latitude range 35° N.–40° N. in the North Atlantic and within the latitude range 20° N.–25° N. in the North Pacific.

4. The seasonal variations of evaporation are greatest in mid-latitudes and along the western sides of the oceans and somewhat greater in the North Atlantic than in the North Pacific.

5. Evaporation is a positive quantity for all oceans except during summer for several small areas at high latitudes in the northwest portions.

6. The distribution of values representing the amount of sensible heat exchanged between sea and atmosphere is similar to the dis-

tribution of values for evaporation with the major exception that there exist no tropical areas of maximum heat exchange to coincide with the tropical areas of high evaporation. Also, except during summer, the center of the area of maximum heat exchange in each ocean is situated farther north than the centers of maximum evaporation, the displacement being 15° in the North Pacific and 5° in the North Atlantic.

7. The values for Q_c are generally positive in the North Pacific except for large areas in summer; in the North Atlantic, these quantities are negative for large areas during all seasons, even in winter.

8. The ratio between the amount of heat exchanged between sea and atmosphere and evaporation is greatest at high latitudes and decreases toward the equator; is greater in the North Pacific at all latitudes than in the North Atlantic.

9. The total evaporation and energy exchange is greater in the North Atlantic than in the North Pacific at the higher latitudes (above 30° N.–35° N.) but the reverse is true at lower latitudes.

10. And, as suggested by 7 and 8, the total amount of sensible heat exchanged between sea and atmosphere is greater in the North Pacific than in the North Atlantic at all latitudes below 50° N.

11. It is shown that the locations of the principal frontal zones over the oceans correspond closely with the areas of maximum energy exchange. It is also shown that the principal frontogenetic regions in every case lie to the northwestward of a zone of maximum energy exchange and that their axes are parallel to the major axes of the zones of maximum energy exchange.

12. It is suggested that for every major meteorological change over the oceans there must be a corresponding change in the energy relationships between sea and atmosphere. Thus, it would seem that any attempts toward solution of long-period weather problems must be concerned with thermodynamical as well as dynamical factors.

ACKNOWLEDGMENTS

The writer wishes to gratefully acknowledge the assistance and advice given by Dr. H. U. Sverdrup of the Scripps Institution of Oceanography at all stages of the investigation. The writer is also indebted to the Washington, D. C., and the New Orleans offices of the U. S. Weather Bureau for the loan of manuscript charts presenting recently compiled climatic data over the oceans.

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