CHAPTER IV

General Distribution of Temperature, Salinity, and Density

The Heat Budget of the Earth as a Whole

For the earth as a whole, the total amount of heat that is received during one year from the sun at the limit of the atmosphere must exactly balance the total amount that in the same period is lost by reflection and by radiation into space. Otherwise, the temperature of the atmosphere and the oceans would change. The radiation from the hot sun is called short-wave radiation, because the wave lengths which reach the limit of the earth's atmosphere lie roughly between 0.38 μ and 2.5 μ, whereas the dark-heat radiation which is emitted by all objects at ordinary temperatures is called long-wave radiation, being of wave lengths between 5 μ and 20 μ. The part of the short-wave radiation that is *reflected* is of no importance to the heat budget of the earth, and therefore the amount of short-wave radiation that is *absorbed* by the atmosphere, the oceans, and the land must exactly balance the long-wave radiation into space from the entire system. A small part of the heat that the atmosphere receives is transformed into kinetic energy which by friction is transformed back again to heat and ultimately lost into space by radiation. Thus, the transformation of heat to kinetic energy does not lead to any net gain of heat but serves to maintain the circulations of the atmosphere and the oceans.

As is customary procedure, the amounts of heat will be given in gram calories and not in units of work such as ergs or joules. The conversion factors are: 1 gram calorie = 4.183×10^7 ergs = 4.183 joules.

In lower latitudes, heat received by radiation is greater than heat lost by back radiation and reflection, whereas in higher latitudes the gain is less than the loss. Table 24 contains values of heat received and lost by processes of radiation and reflection in different latitudes. The third column, containing the differences between the two quantities, shows that there is an annual net gain of heat in the equatorial regions and a net loss in the polar regions. The mean annual temperatures in different latitudes on the earth remain unchanged from one year to another, showing that within the atmosphere and the oceans there must be a transport of heat from lower to higher latitudes which exactly equals the difference between heat received and heat lost by radiation. Multiplying the average difference between any two parallels of latitude by the area of the earth's surface between these parallels and summing up, starting at the Equator, gives the total amounts of heat that flow from the Equator toward the poles in every latitude. Some of these values are given in the fourth column in the table, from which it is seen that they are of the order of 10^{16} g cal/min. Dividing the numbers by the length of the parallels gives the amounts shown in the fifth column of the table, which represent the average flow of heat across each centimeter of the parallels of latitude. These numbers are of the order of 10^7 g cal/cm/min.

The transport of heat from lower to higher latitudes takes place partly by air currents (winds) and partly by ocean currents. In meteorological literature it is generally assumed that the transport by ocean currents is negligible (Bjerknes *et al*, 1932), although the question has not been thoroughly examined. It can be shown that the assumption is correct when dealing with averages for the whole earth, but in some regions the transport by ocean currents is of considerable importance.

TABLE 24

HEAT BUDGET OF THE EARTH AS A WHOLE AND HEAT TRANSPORT FROM LOWER TO HIGHER LATITUDES

Latitude (°)	Heat received (g cal/cm²/min)	Heat lost (g cal/cm²/min)	Surplus or deficit (g cal/cm²/min)	Heat transport across parallels of latitude (10^{16} g cal/min)	Heat transport across every centimeter of parallels of latitude (10^7 g cal/cm/min)
0	0.339	0.300	0.039	0.00	0.00
10	0.334	0.299	0.035	1.59	0.40
20	0.320	0.294	0.026	2.94	0.78
30	0.297	0.283	0.014	3.58	1.07
40	0.267	0.272	−0.005	3.96	1.30
50	0.232	0.258	−0.026	3.34	1.32
60	0.193	0.245	−0.052	2.40	1.20
70	0.160	0.231	−0.071	1.20	0.88
80	0.144	0.220	−0.076	0.32	0.46
90	0.140	0.220	−0.080	0.00	0.00

The amount of heat transported in a north-south direction by a unit volume of ocean water is equal to $c\rho\vartheta v_N$, where c represents the specific heat, ρ the density, ϑ the temperature of the water, and v_N the north-south component of velocity. The total transport through a certain section of the sea can be found by integration, but for the sake of simplicity we shall assume that this transport can be written $c\bar{\vartheta}\rho T_N$, where $\bar{\vartheta}$ represents the average temperature of the water and ρT_N represents the mass of water passing north through the section in unit time. If the section is taken across an ocean, the mass transport to the north, ρT_N, must equal the mass transport to the south, ρT_S, but the heat transport may differ because the temperature of the water transported in one direction may be higher or lower than that of the water which is transported in the opposite direction. If these temperatures are designated by ϑ_N and ϑ_S, the net transport of heat will be $c(\vartheta_N - \vartheta_S)\rho T$, where ρT now means the transport to the north and the south. As an example, we can apply these considerations to the North Atlantic Ocean along the parallel of 55°N. In the eastern Atlantic about 10 million m^3/sec of warm water flow toward the north, and in the western Atlantic an equal volume of cold water is carried south by the Labrador Current and by the flow of the deep water (p. 684). With $\vartheta_N - \vartheta_S = 5°$, $c = 1$, $\rho = 1$, and $T = 10 \times 10^6$ m^3/sec, we find that the heat transport toward the north through latitude 55° in the Atlantic Ocean is about 0.3×10^{16} g cal/min. The total heat transport across the parallel of 55°N is about 3×10^{16} g cal/min; hence in the North Atlantic the fraction carried by the ocean currents is appreciable.

This example represents an outstanding case of poleward transport of heat by ocean currents. In the Pacific Ocean a transport of comparable magnitude probably takes place in latitudes 30°N to 40°N, but in the southern oceans the north-south circulation and the corresponding temperature contrasts between currents flowing toward or away from the higher latitudes are smaller than those in the northern oceans. A detailed study of the transport of heat by ocean currents has not been made, but it seems certain that by far the major transport of heat is taken care of by the atmosphere.

The Heat Budget of the Oceans

The above consideration applies to the entire system formed by the atmosphere and the oceans, but for the oceans alone we encounter an entirely different picture. On an average, the gain of heat must exactly balance the loss, but the processes involved are not limited to those of radiation, as is evident from the list at the bottom of page 101.

These processes will be discussed in detail, but it can already be stated that of the processes of heating only the first one is important, and the heat budget of the oceans as a whole can therefore be written

$$Q_s - Q_b - Q_h - Q_e = 0,$$

where Q_s is the heat received. Over all ocean surfaces between 70°N and 70°S the average amounts (in g cal/cm²/min) are, according to Mosby (1936),

	$Q_s = 0.221$	$Q_b = 0.090$
		$Q_h = 0.013$
		$Q_e = 0.118$
Total	0.221	0.221

If a specific region is considered, it must be taken into account that heat may be brought into or out of that region by ocean currents or by processes of mixing, and that during short time intervals a certain amount of heat may be used for changing the temperature of the water. The complete equation for the heat balance of any part of the ocean in a given time interval is, therefore,

$$Q_s - Q_b - Q_e - Q_h + Q_v + Q_\vartheta = 0,$$

where Q_v represents the net amount which by currents or processes of mixing is brought into or out of the region, and where Q_ϑ represents the amount of heat used locally for changing the temperature of the sea water.

RADIATION FROM THE SUN AND THE SKY. The short-wave radiation that reaches the sea surface comes partly directly from the sun and partly from the sky as reflected or scattered radiation. The amount of radiation energy which is absorbed per unit volume in the sea depends upon the amount of energy that reaches the sea surface, the reflection from the sea surface, and the absorption coefficients for total energy. The incoming radiation depends mainly upon the altitude of the sun, the absorption in the atmosphere, and the cloudiness. With a clear sky and a high sun, about 85 per cent of the radiation comes directly from the sun and about 15 per cent from the sky, but with a low sun the proportion from the sky is greater, reaching about 40 per cent of the total with the sun 10 degrees above the horizon.

Processes of Heating of the Ocean Water

1. Absorption of radiation from the sun and the sky, Q_s.
2. Convection of heat through the ocean bottom from the interior of the earth.
3. Transformation of kinetic energy to heat.
4. Heating due to chemical processes.
5. Convection of sensible heat from the atmosphere.
6. Condensation of water vapor.

Processes of Cooling of the Ocean Water

1. Back radiation from the sea surface, Q_b.
2. Convection of sensible heat to the atmosphere, Q_h.
3. Evaporation, Q_e.

The incoming energy from the sun is cut down when passing through the atmosphere, partly through absorption by water vapor and carbon dioxide in the air and partly through scattering against the air molecules or very fine dust. The total effect of absorption and scattering in the atmosphere depends upon the thickness of the air mass through which the sun's rays pass, as expressed by the equation

$$I = Se^{-Ta_m m}.$$

Here I represents the energy in g cal/cm²/min reaching a surface which is normal to the sun's rays; m represents the relative thickness of the air mass and is equal to 1 at a pressure of 760 mm when the sun stands in zenith, equal to 2 when the sun is 30° above the horizon ($\sin 30° = \frac{1}{2}$), and so on; S is the solar constant (1.932 g cal/cm²/min); T is the "turbidity factor" of the air; and a_m is $0.128 - 0.054 \log m$.

The sun's radiation on a *horizontal* surface is obtained by multiplication with $\sin h$, where h is the sun's altitude. To this amount must be added the diffuse sky radiation in order to obtain the total radiation on a horizontal surface. Instruments are in use for recording the total radiation and for recording separately the radiation from the sun and from the sky.

When the sun is obscured by clouds, the radiation comes from the sky and the clouds and, on an average, can be represented by the formula $Q = Q_0(1 - 0.071\,C)$, where the cloudiness C is given on the scale 0 to 10, and where Q_0 represents the total incoming radiation with a clear sky. This formula is applicable, however, only to average conditions. If the sun shines through scattered clouds, the radiation may be greater than with a clear sky, owing to the reflection from the clouds, and on a completely overcast dark and rainy day the incoming radiation may be cut down to less than 10 per cent of that on a clear day. Table 25 contains the average monthly amounts of incoming radiation, expressed in g cal/cm²/min, which reach a horizontal surface in the indicated localities (computed from Kimball, 1928). The differences between the parts of the oceans in the same latitudes are mainly due to differences in cloudiness.

Few direct measurements of radiation are available from the oceans, and when dealing with the incoming radiation it is necessary to consider average values which can be computed from empirical formulae. Mosby (1936) has established such a formula by means of which monthly or annual mean values of the incoming radiation on a horizontal surface can be computed if the corresponding average altitude of the sun and the average cloudiness are known:

$$Q = k(1 - 0.071C)\bar{h}(\text{g cal/cm}^2\text{/min}).$$

TABLE 25

AVERAGE AMOUNTS OF RADIATION FROM SUN AND SKY, EXPRESSED IN GRAM CALORIES PER SQUARE CENTIMETER PER MINUTE, WHICH EVERY MONTH REACHES THE SEA SURFACE IN THE STATED LOCALITIES
(After Kimball)

Locality		Month											
Latitude	Longitude	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
60°N	7°E- 56°W	.002	.053	.125	.207	.272	.292	.267	.212	.147	.074	.006	0
60 N	135 -170 W	.005	.078	.155	.208	.269	.260	.242	.185	.127	.077	.015	0
52 N	10 W	.048	.089	.148	.219	.258	.267	.251	.211	.160	.104	.062	.041
52 N	129 W	.053	.091	.135	.185	.246	.250	.230	.214	.158	.097	.058	.039
42 N	66 - 70 W	.094	.138	.212	.272	.306	.329	.302	.267	.230	.174	.115	.086
42 N	124 W	.100	.151	.210	.286	.331	.360	.320	.274	.231	.174	.113	.092
30 N	65 - 77 W	.146	.165	.238	.285	.317	.310	.301	.282	.239	.188	.169	.142
30 N	128 -130 E	.141	.153	.199	.241	.258	.238	.256	.260	.219	.178	.153	.135
10 N	61 - 69 W	.254	.276	.299	.305	.272	.276	.285	.292	.287	.269	.248	.239
10 N	116 E- 80 W	.226	.257	.292	.278	.255	.239	.240	.242	.247	.237	.224	.219
0	7 - 12 E	.239	.248	.244	.230	.210	.196	.188	.194	.220	.240	.239	.235
0	48 W & 170 E	.261	.265	.282	.297	.309	.300	.300	.340	.366	.362	.339	.278
10 S	14 E; 36- 38 W	.329	.328	.301	.254	.219	.206	.232	.278	.312	.324	.317	.320
10 S	72 -171 E	.290	.308	.315	.289	.266	.253	.269	.306	.332	.313	.301	.303
30 S	17 and 116 E	.452	.406	.340	.254	.186	.148	.166	.214	.274	.362	.401	.430
30 S	110 W	.380	.330	.260	.209	.162	.130	.145	.176	.237	.321	.340	.390
42 S	73 W; 147 E	.343	.297	.223	.154	.104	.085	.092	.135	.187	.264	.310	.348
52 S	58 W	.289	.237	.167	.112	.062	.039	.049	.097	.150	.222	.273	.302
60 S	45 W	.213	.171	.105	.056	.011	0	.003	.054	.111	.156	.204	.221

Here $\bar{h}$ is the average altitude of the sun. The factor k depends upon the transparency of the atmosphere and appears to vary somewhat with latitude, being 0.023 at the Equator, 0.024 in lat. 40°, and 0.027 in lat. 70°. Mosby's formula is not valid at $h > 60°$, but gives correct results if, at high altitudes of the sun, the true altitude is replaced by a reduced altitude as follows:

True altitude (°)	60	65	70	75	80	85	90
Reduced altitude (°)	60	62	64	66	68	69	70

The values computed by means of this formula agree within a few per cent with those derived by Kimball in an entirely different manner (table 25).

Part of the incoming radiation is lost by reflection from the sea surface, the loss depending upon the altitude of the sun. When computing the loss, the direct radiation from the sun and the scattered radiation from the sky must be considered separately. With the sun 90°, 60°, 30° and 10° above the horizon, the reflected amounts of the direct solar radiation are, according to Schmidt (1915), 2.0 per cent, 2.1 per cent, 6.0 per cent, and 34.8 per cent respectively. For diffuse radiation from the sky and from clouds Schmidt computes a reflection of 17 per cent. Measurements by Powell and Clarke (1936) gave values on clear days in agreement with the above, but on overcast days when all radiation reaching the sea surface was diffuse, the observed reflection was about 8 per cent. If the fractions of the total radiation from the sun and the sky on a clear day are designated p and q, respectively, and if the corresponding percentages reflected are designated m and n, the percentage of the total incoming radiation that is reflected on a clear day is $r = mp + nq$. Thus, on an overcast day, when all incoming

TABLE 26

PERCENTAGE OF TOTAL INCOMING RADIATION FROM SUN AND SKY WHICH ON A CLEAR DAY IS REFLECTED FROM A HORIZONTAL WATER SURFACE AT DIFFERENT ALTITUDES OF THE SUN

Altitude of the sun (°)	5	10	20	30	40	50	60	70	80	90
Percentage reflected	40	25	12	6	4	3	3	3	3	3

radiation is diffuse, $r = 8$ per cent. Table 26 contains approximate values of r at different altitudes of the sun on a clear day.

The values in the table are applicable only if the sea surface is smooth. In the presence of waves the reflection loss at a low sun is somewhat increased and will be of particular importance in high latitudes. The amount of radiation which under stated conditions penetrates the sea surface is obtained by subtracting the reflection loss from the total incoming radiation.

ABSORPTION OF RADIATION ENERGY IN THE SEA. The radiation that penetrates the surface is absorbed in the sea water. The amounts absorbed within given layers of water can be derived by measuring with a thermopile the energies which reach different depths or by computing these energies by means of known extinction coefficients. Direct measurements of energy have been made in Mediterranean waters only (Vercelli, 1937), but extinction coefficients of radiation of different wave lengths have been determined in many areas (p. 85). For computation of the energy that reaches a given depth, it is necessary to know the intensity of the radiation at different wave lengths; that is, the energy spectrum. The reduction in intensity has to be calculated for each wave length, and the total energy reaching a given depth has to be determined from the energy spectrum by means of integration. The definition of the extinction coefficient for total energy corresponds to the definition of extinction coefficients at given wave lengths (p. 82).

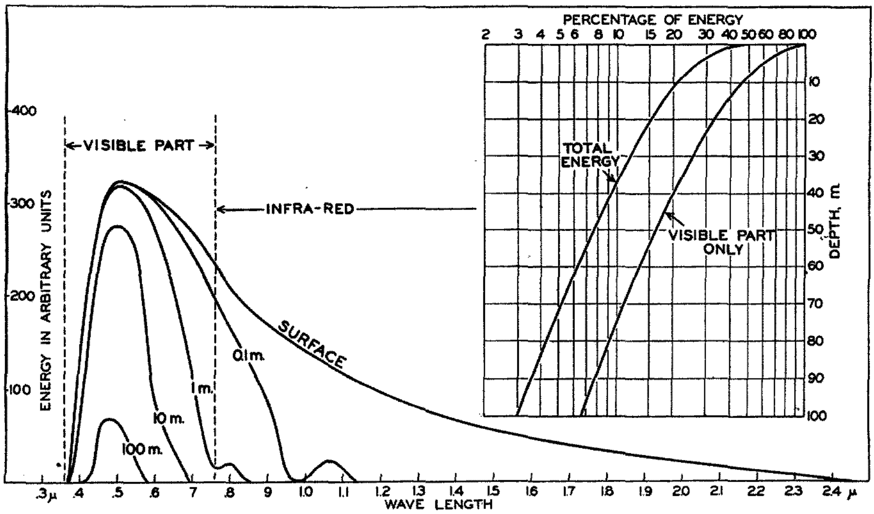


Fig. 21. Schematic representation of the energy spectrum of the radiation from the sun and the sky which penetrates the sea surface, and of the energy spectra in pure water at depths of 0.1, 1, 10, and 100 m. *Inset:* Percentages of total energy and of energy in the visible part of the spectrum reaching different depths.

The spectrum of the energy that penetrates the sea surface is represented approximately by the upper curve in fig. 21, which also shows the energy spectra at different depths in pure water. The total energy at any given depth is proportional to the area enclosed between the base line and the curves showing the energy spectrum. In the inserted diagram the total energy, expressed as percentage of the energy penetrating the surface, as well as the corresponding percentages of the energy in the visible part of the spectrum, is plotted against depth. The figure shows that pure water is transparent for visible radiation only.

For sea water the percentage of the total energy reaching various depths has been computed for the clearest oceanic water, for average oceanic water, for average coastal water, and for turbid coastal water, using the extinction coefficients shown in fig. 20. The results are presented in table 27. In the clearest offshore water, 62.3 per cent of the incoming energy is absorbed in the first meter. The absorption is often increased in the upper one meter owing to the presence of foam and air bubbles. This increased absorption, when dealing with the penetration of light, is referred to as "surface loss." If this process is disregarded, the values clearly demonstrate that the greater amount of energy is absorbed very near the sea surface and that the amount which penetrates to any appreciable depth is considerable only when the water is exceptionally clear. At 10 m, 83.9 per cent has been absorbed in the clearest water and 99.55 per cent in the turbid coastal water.

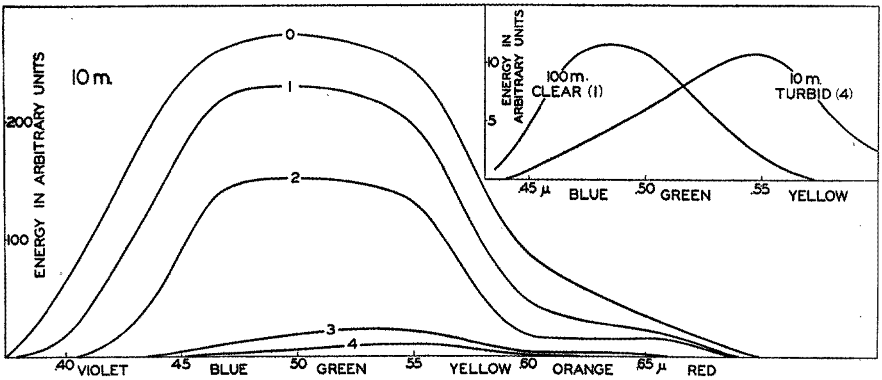


Fig. 22. Energy spectra at a depth of 10 m in different types of water. Curves marked 0, 1, 2, 3, and 4 represent energy spectra in pure water, clear oceanic, average oceanic, average coastal, and turbid coastal sea water, respectively. *Inset:* Energy spectra at a depth of 100 m in clear oceanic water and at 10 m in turbid coastal water.

The absorption of energy is illustrated in fig. 22, which shows the energy spectra in different types of water at a depth of 10 m. At this depth the maximum energy in the clearest water is found in the blue-green portion of the spectrum, whereas in the turbid coastal water the maximum has been displaced toward the greenish-yellow part. This displacement is further illustrated by the inserted curve in the upper right-hand corner of the figure, which shows the energy spectra at 100 m in the clearest water and at 10 m in the most turbid water.

Extinction coefficients of *total* energy have been computed and are entered in table 27. These extinction coefficients are very high in the upper 1 m but decrease rapidly, at greater depth approaching the minimum extinction coefficients characteristic of the types of water dealt with. The smallest values given in the table can be considered valid at greater depths as well.

TABLE 27

PERCENTAGE AMOUNTS OF TOTAL INCIDENT ENERGY AT DIFFERENT LEVELS AND EXTINCTION COEFFICIENTS OF ENERGY IN DIFFERENT INTERVALS OF DEPTH

Depth	Percentage amounts of incident energy					Interval of depth (m)	Extinction coefficients per meter				
	Pure sea water	Oceanic water		Coastal water			Pure sea water	Oceanic water		Coastal water	
		Clearest	Average	Average	Turbid			Clearest	Average	Average	Turbid
0	100	100	100	100	100	0–1	.944	.975	1.080	1.318	1.385
1	38.9	37.7	35.2	26.7	22.8	1–2	.143	.176	.230	.450	.547
2	33.7	31.6	28.0	17.0	13.2	2–5	.062	.095	.159	.351	.452
5	28.0	23.7	17.3	5.95	3.41	5–10	.048	.076	.120	.318	.405
10	22.0	16.1	9.50	1.21	0.449	10–20	.033	.054	.094	.292	.368
20	15.8	9.35	3.72	0.064	0.012	20–50	.024	.042	.083		
50	7.64	2.69	0.311			50–100	.018	.036			
100	3.04	0.452	0.0057								

In fig. 23 the curves marked 0, 1, 2, 3, and 4 represent the percentage amounts of energy which reach different levels between the surface and 10 m, according to the data in table 27. The three curves marked Capri, Trieste, and Venice represent results of measurements in the Mediterranean according to Vercelli (1937), and four other curves represent observed values in lakes according to Birge and Juday (1929). The agreement of the character of the curves indicates that reliable values as to the absorption of energy in the sea can be obtained by means of computations based on observed extinction coefficients.

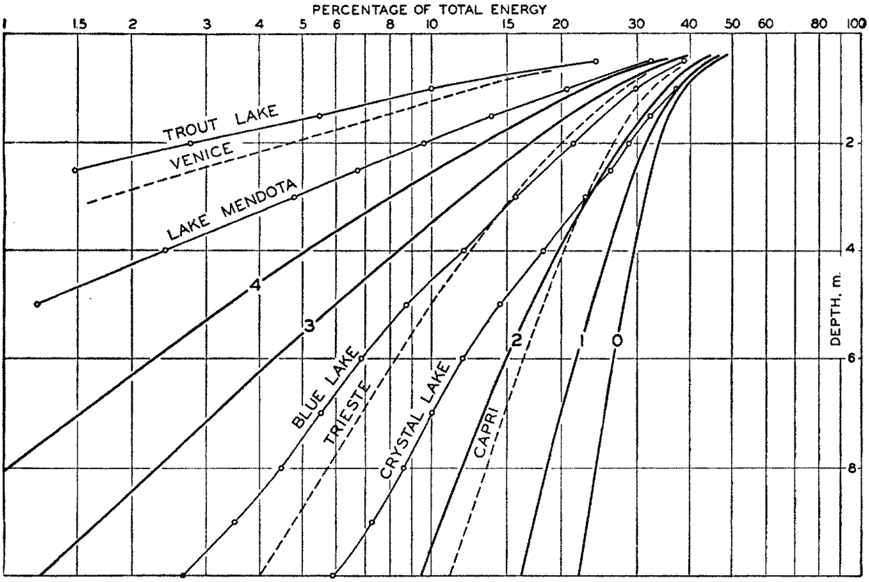


Fig. 23. Percentages of total energy reaching different depths in pure water, clear oceanic, average oceanic, average coastal, and turbid coastal sea water (curves 0, 1, 2, 3, and 4), computed from extinction coefficients, and corresponding directly to observed values in four lakes and at three localities in the Mediterranean.

An idea of the heating due to absorption of radiation can be secured by computing the increase of temperature at different depths which results from a penetration of 1000 g cal/cm^2 through the surface. The results are shown in table 28, which serves to emphasize that the greater part of the energy is absorbed near the surface, particularly in turbid water. If no other processes took place, the temperature between the surface and 1 m would increase in the clearest water by 6.24°, and in the most turbid water by 7.72°. Between 20 and 21 m the corresponding values would be 0.04° and 0.0003°.

The temperature changes recorded in table 28 show no similarity to those actually occurring in the open oceans, where processes of mixing entirely mask the direct effect of absorption, but in some small, land-locked bodies of water the temperature changes at subsurface depths may be governed mainly by absorption of short-wave radiation. Such processes may be observed in the oyster basins on the west coast of Norway, the temperature characteristics of which were described by Helland-Hansen and studied in detail by Gaarder and Spärck (1932). These basins are in communication with the open sea through narrow and shallow openings, but during winter storms a complete exchange of water often takes place between the basins and the outside. In the ensuing spring, after rains, which cause considerable run-off, the surface layer in the basins will be replaced by fresh or brackish water spreading over the sea water in the deeper portion of the basins and forming a cover

TABLE 28

TEMPERATURE INCREASE IN °C AT DIFFERENT DEPTH INTERVALS AND IN DIFFERENT TYPES OF WATER, CORRESPONDING TO AN ABSORPTION OF 1000 G CAL/CM^2

Interval of depth (m)	Oceanic water		Coastal water	
	Clearest	Average	Average	Turbid
0– 1	6.24	6.48	7.32	7.72
1– 2	0.610	0.720	0.970	0.960
5– 6	.236	.282	.164	.120
10– 11	.104	.096	.030	.0140
20– 21	.040	.030	.0016	.0003
50– 51	.0096	.0024	$.0_5 34$	$.0_7 15$
100–101	.0016	$.0_4 11$		

that prevents further exchange between the deeper water and the outside sea. Owing to the difference in salinity the density of the deeper water will be much higher than the density of the surface layer. During summer the incoming radiation will be absorbed both in the fresh water on top and in the underlying sea water, and the temperature will rise within both layers. Within the top layer the ordinary convection currents develop, and the temperature is controlled mainly by the air temperature, but owing to the greater salinity of the lower layer the temperature of the deeper water can rise to high values without leading to unstable stratification, and the effect of absorption becomes apparent, because no other processes are of importance.

Fig. 24 shows the vertical distribution of temperature on June 30 and July 15 in a basin that was examined by Gaarder and Spärck. The days in the period between the stated dates were clear and no rain fell. According to Kimball (1928) the diurnal amount of incoming short-wave radiation was about 740 g cal/cm²/day, or about 11,100 g cal/cm² for the entire period. If 6 per cent is subtracted for reflection, the amount entering the water would be about 10,400 g cal/cm^2. The temperature curves show that of this amount 1630 g cal/cm^2, or 15.5 per cent, was absorbed below a depth of 1 m, and thus 84.5 per cent was absorbed between the surface and 1 m. The latter amount was lost by back radiation, heat conduction to the atmosphere, and evaporation, and at present need not be further considered. It is of interest, however, to point out that the great absorption in the upper meter indicates that the turbidity of the water in the basin was greater than that of ordinary coastal waters (table 27, fig. 23) and approached that of turbid lakes. The heating between 1 and 2 m also indicates water of great turbidity, because, of the total amount of 1630 g cal/cm^2 reaching 1 m, 630 g cal/cm^2 were absorbed in that layer, the corresponding extinction coefficient being 0.488 (cf. table 27). This result agrees roughly with Gaarder and Spärck's statement that a Secchi disc nearly disappeared at a depth of only 2 m. Conditions in this case were unusually clear cut, but even in such basins the effect of absorption is often obscured by processes of heat conduction.

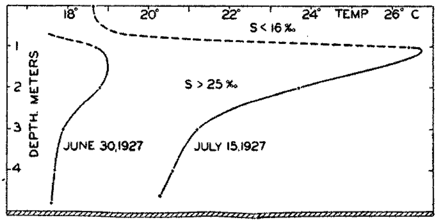


Fig. 24. Vertical distribution of temperature in a Norwegian oyster basin on June 30 and July 15, 1927.

Conduction of Heat Through the Ocean Bottom. It has been estimated that the flow of heat through the bottom of the sea amounts to between 50 and 80 g cal/cm^2/year (Helland-Hansen, 1930). This amount represents less than one ten-thousandth part of the radiation received at the surface and can generally be neglected when dealing with the heat budget of the oceans. In a few basins, where the deep water is nearly stagnant and where no conduction of heat takes place from above or from the sides, the amount of heat conducted through the bottom may conceivably play a part in determining the distribution of temperature, but so far no such case is known with certainty (p. 739).

Transformation of Kinetic Energy into Heat. The kinetic energy transmitted to the sea by the stress of the wind on the surface and by part of the tidal energy is dissipated by friction and transformed into heat. The energy transmitted by the wind can be estimated at about one ten-thousandth part of the radiation received at the surface and can be neglected. In shallow coastal waters with strong tidal currents the dissipation of tidal energy is so great, however, that it may become of some local importance. Thus, in the Irish Channel, according to Taylor (1919), the dissipation amounts to about 0.002 g cal/cm^2/min, or 1050 g cal/cm^2/year. The average depth can be taken as about 50 m, or 5000 cm, and, if the same water remained in the Irish Channel a full year, the increase in temperature would be about 0.2°C, on an average. Such an effect, however, has not been established, and, as it can be expected in shallow coastal waters only, it is of no significance to the general heat budget of the oceans. A possible case of heating due to dissipation of tidal energy or to conduction of heat through the ocean bottom has been discussed by Sverdrup (1929).

Heating due to chemical processes can be completely disregarded.

The convection of sensible heat and the condensation of water vapor will be dealt with in the discussion concerning the exchange of heat and water with the atmosphere.

Effective Back Radiation from the Sea Surface. The sea surface emits long-wave heat radiation, radiating nearly like a black body, the energy of the outgoing radiation being proportional to the

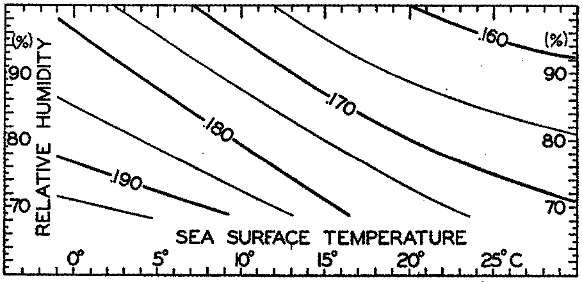


Fig. 25. Effective back radiation in gm cal/cm²/min from the sea surface to a clear sky. Represented as a function of sea-surface temperature and relative humidity of the air at a height of a few meters.

fourth power of the absolute temperature of the surface. At the same time the sea surface receives long-wave radiation from the atmosphere, mainly from the water vapor. A small part of this incoming long-wave radiation is reflected from the sea surface, but the greater portion is absorbed in a small fraction of a centimeter of water, because the absorption coefficients are enormous at long wave lengths. The *effective back radiation* from the sea surface is represented by the difference between the "temperature radiation" of the surface and the long-wave radiation from the atmosphere, and this effective radiation depends mainly upon the temperature of the sea surface and the water-vapor content of the atmosphere. According to Ångström (1920), the latter is proportional to the local vapor pressure, which can be computed from the relative humidity if the air temperature is known. Over the oceans the air temperature deviates so little from the sea-surface temperature that the vapor pressure can be obtained with sufficient accuracy from the sea-surface temperature and the relative humidity of the air at a short distance above the surface.

Ångström (1920) has published a table that summarizes results of observations of effective radiation against a clear sky from a black body of different temperatures and at different vapor pressures. Fig. 25 has been prepared by means of this table, taking into account the small difference between the radiation of a black body and that of a water surface. The figure shows the effective radiation as a function of sea-surface temperature and of relative humidities between 100 per cent and 70 per cent, but the values that can be read off from the graph may be 10 per cent in error, owing to the scanty information upon which the curves are based. It brings out the interesting fact, however, that, owing to the increased radiation from the atmosphere at higher temperatures (higher vapor pressures), the effective back radiation *decreases* slowly with *increasing* temperature. At a temperature of 0°C and a relative humidity of 80 per cent, the effective back radiation is 0.188 g cal/cm^2/min, and at a temperature of 25° and the same relative humidity it is 0.167 g cal/cm^2/min. At a given temperature the effective radiation decreases with increasing humidity, owing to the increased back radiation from the atmosphere. Thus, at a surface temperature of 15° the effective radiation is about 0.180 g cal/cm^2/min at a relative humidity of 70 per cent, and about 0.163 g cal/cm^2/min at a relative humidity of 100 per cent.

The values of the effective back radiation at higher temperatures, as obtained by extrapolation of Ångström's data (fig. 25) are greater than those computed from Brunt's empirical formula,

$$Q_b = Q'(1 - 0.44 - 0.08\sqrt{e}),$$

where Q' is the radiation of a black body having the temperature of the sea surface and e is the vapor pressure of the air in millibars. However, in this formula the numerical values of the coefficients are uncertain and are applicable only within a range of e between 4 and 18 mb.

The diurnal and annual variations of the sea-surface temperatures and of the relative humidity of the air over the oceans are small, and the effective back radiation at a clear sky is therefore nearly independent of the time of the day and of the season of the year, in contrast to the incoming short-wave radiation from the sun and the sky, which is subjected to very large diurnal and seasonal variations.

In the presence of clouds the effective back radiation is cut down because the radiation from the atmosphere is increased. The empirical relation can be written

$$Q = Q_0(1 - 0.083C),$$

where Q_0 is the back radiation at a clear sky and where C is the cloudiness on the scale 1 to 10. A diurnal or annual variation in the cloudiness will lead to a corresponding variation in the effective back radiation. On an average, the diurnal variation of cloudiness over the oceans is very small and can be neglected, but the annual variation is in some regions considerable. The above equation is applicable to average conditions only, because the reduction of the effective back radiation due to clouds depends upon the altitude and the density of the clouds. Thus, if the sky is completely covered by cirrus, alto-stratus, or strato-cumulus clouds, the effective radiation is about $0.75Q_0$, $0.4Q_0$, and $0.1Q_0$, respectively.

The annual incoming short-wave radiation from the sun and the sky is greater in all latitudes than the outgoing effective back radiation. According to Mosby (1936) the average annual surplus of incoming radiation between latitudes 0 and 10°N is about 0.170 g cal/cm²/min, and between 60° and 70°N about 0.040 g cal/cm²/min. The surplus of radiation must be given off to the atmosphere, and the exchange of heat and water vapor with the atmosphere is therefore equally as important as the processes of radiation in regulating the ocean temperatures and salinities.

The characteristics of the oceans in respect to radiation are very favorable to man. The water surface reflects only a small fraction of the incoming radiation, and the greater part of the radiation energy is absorbed in the water, distributed by processes of mixing over a layer of considerable thickness, and given off to the atmosphere during periods when the air is colder than the sea surface. The oceans therefore exercise a thermostatic control on climate. Conditions are completely changed, however, if the temperature of the sea surface decreases to the freezing point so that further loss of heat from the sea leads to formation of ice, because, when water passes this critical temperature, its thermostatic characteristics are altered in a very unfavorable direction. Sea ice, which soon attains a gray-white appearance owing to enclosed air bubbles, reflects 50 per cent or more of the incoming radiation, and if it is covered by rime or snow the reflection loss increases to 65 per cent, or even to 80 per cent from fresh, dry snow. The snow surface, on the other hand, radiates nearly like a black body, and consequently the heat budget related to processes of radiation, instead of rendering a surplus as it does over the open ocean, shows a deficit until the temperature of the ice surface has been lowered so much that the decreased loss by effective back radiation balances the small fraction of the incoming radiation that is absorbed. The immediate result of freezing is therefore a general lowering of the surface temperature of the ice and a rapid increase of the thickness of the ice. The air that comes in contact with the ice is cooled, and, as this cold air spreads, more ice is formed. Thus, a small lowering of the temperature of the water in high latitudes followed by freezing may lead to a rapid drop of the air temperature and a rapid increase of the ice-covered area. On the other hand, a small increase of the temperature of air flowing in over an ice-covered sea may lead to melting of the ice at the outskirts and, once started, the melting may progress rapidly. In agreement with this reasoning it has been found that the extent of ice-covered areas in the Barents Sea is a sensitive indicator of small changes in the atmospheric circulation and in the amount of warm water carried into the region by currents (p. 662). It has also been computed that, if the average air temperature in middle and higher latitudes were raised a few degrees, the Polar Sea would soon become an ice-free ocean.

Exchange of Heat between the Atmosphere and the Sea. The amount of heat that in unit time is carried away from the sea surface through a unit area is equal to

$$-c_pA\left(\frac{d\vartheta}{dz}+\gamma\right),$$

where c_p is the specific heat of the air, A is the eddy conductivity, $-d\vartheta/dz$ is the temperature gradient of the air (the lapse rate), which is positive when the temperature decreases with height, and γ is the adiabatic lapse rate. Very near the sea surface, γ can be neglected as small compared to $d\vartheta/dz$. The term c_pA enters instead of the coefficient of heat conductivity of the air as determined in the laboratory because the air is nearly always in turbulent motion (p. 92). The state of turbulence varies, however, with the distance from the sea surface, because at the surface itself the eddy motion must be greatly reduced. As a consequence, under steady conditions, when the same amount of heat passes upward through every cross section of a vertical column, the temperature changes rapidly with height near the sea surface and more slowly at greater distance. The product $-c_pAd\vartheta_a/dz$ remains constant, and, as c_pA increases rapidly with height, $-d\vartheta_a/dz$ must decrease.

Detailed and accurate temperature measurements in the lowest meters of air over the ocean have not yet been made, because the hull and masts of a vessel disturb the normal distribution of temperature to such an extent that values observed at different levels on board a vessel are not representative of the undisturbed conditions. The few measurements that have been attempted indicate, however, that the general distribution as outlined above is encountered.

The sea surface must be warmer than the air at a small distance above the surface if heat is to be conducted from the sea to the air. When such conditions prevail, the air is heated from below, the stratification of the air becomes unstable, and the turbulence of the air becomes intense (p. 92). If the sea surface is very much warmer than the air, as may be the case when cold continental air flows out over the sea in winter, the heating from below may be so intense that rapid convection currents develop, leading to such violent atmospheric disturbances as thunderstorms. It is not intended here to enter upon the meteorological aspects of the heat exchange, but the point which is emphasized is that an appreciable conduction of heat from the sea to the atmosphere takes place only when the sea surface is warmer than the air. One might assume that, vice versa, an appreciable amount of heat would be conducted to the sea surface when warmer air flows over a cold sea, but this is not the case, because under such conditions the air is cooled from below, the stratification of the air becomes stable, and the turbulence (and consequently the eddy conductivity of the air) is greatly reduced.

It has been found (p. 128) that on an average the sea surface is slightly warmer than the overlying air and therefore loses heat by conduction. So far, no detailed studies have been made, but Ångström has estimated that only about 10 per cent of the total heat surplus is given off to the atmosphere by conduction and that 90 per cent is used for evaporation. Other estimates indicate that these figures are approximately correct (p. 117). Thus, evaporation is of much greater importance to the heat balance of the oceans than is the transfer of sensible heat. Evaporation will therefore be dealt with in greater detail.

Evaporation from the Sea

The Process of Evaporation. The vapor tension at a flat surface of pure water depends on the temperature of the water. The salinity decreases the tension slightly, the empirical relation between vapor tension and salinity being (p. 66)

$$e_s = e_d(1 - 0.000537\ S),$$

where e_s is the vapor tension over sea water, e_d is the vapor tension over distilled water of the same temperature, and S is the salinity in parts per mille. In the open ocean the relation is approximately $e_s = 0.98e_d$. Table 29 contains the vapor tension in millibars over water of salinity 35.00 $^0/_{00}$ and at the stated temperatures.

Air in which the vapor tension is less than that over water of the same temperature is undersaturated with moisture, and air in which the vapor tension exactly equals that over a water surface of the same temperature is saturated with moisture. In absolutely pure air the vapor pressure can be above the saturation value, but generally the air contains "nuclei" on which the vapor is condensed when the vapor tension reaches the value corresponding to that over water of the same temperature. Under these conditions the vapor tension in the air cannot be further increased, and in meteorology one therefore uses the term "maximum vapor tension" at a given temperature. The maximum vapor tension at which condensation takes place can be reached either by adding water vapor to air of a given temperature or by decreasing the temperature of air of a given moisture content. In the latter case condensation begins at the temperature that is called the "dew point."

TABLE 29

MAXIMUM VAPOR TENSION IN MILLIBARS OVER WATER OF SALINITY 35 ‰

Temperature (°C)	Vapor pressure (mb)	Temperature (°C)	Vapor pressure (mb)
−2	5.19	18	20.26
−1	5.57	19	21.57
0	5.99	20	22.96
1	6.44	21	24.42
2	6.92	22	25.96
3	7.43	23	27.59
4	7.98	24	29.30
5	8.56	25	31.12
6	9.17	26	33.01
7	9.83	27	35.02
8	10.52	28	37.13
9	11.26	29	39.33
10	12.05	30	41.68
11	12.88	31	44.13
12	13.76	32	46.71
13	14.70		
14	15.69		
15	16.74		
16	17.85		
17	19.02		

In discussing the process of evaporation it is more rational to consider not the vapor pressure but the specific humidity, f—that is, the mass of water vapor per unit mass of air. The amount of water vapor, F, which per second is transported upward through a surface of cross section 1 cm² is, then, $-A df/dz$, where A is the eddy conductivity and $-df/dz$ is the vertical gradient of the specific humidity, which is positive when the specific humidity decreases with height. If the vapor pressure, e, is introduced, the equation becomes approximately

$$F = -A \frac{0.621}{p} \frac{de}{dz},$$

where p is the atmospheric pressure. The heat needed for evaporation at the surface is

$$Q_e = -L_\vartheta A \frac{0.621}{p} \frac{de}{dz},$$

where L_ϑ is the heat of vaporization at the temperature of the surface, ϑ (p. 62).

The ratio between the amounts of heat given off to the atmosphere as sensible heat (p. 114) and used for evaporation is

$$R = \frac{Q_h}{Q_e} = \frac{c_p^*}{L_\vartheta} \frac{p}{0.621} \frac{d\vartheta_a/dz}{de/dz} = 0.66 \frac{p}{1000} \frac{d\vartheta_a/dz}{de/dz}.$$

The last expression is obtained by introducing $c_p = 0.240$ and $L = 585$. Thus, the ratio R depends mainly upon the ratio between the temperature and humidity gradients in the air at a short distance from the sea surface. These gradients are difficult to measure but can be replaced approximately by the difference in temperature and vapor pressure at the sea surface and the corresponding values in the air at a height of a few meters:

$$R = 0.66 \frac{p}{1000} \frac{\vartheta_w - \vartheta_a}{e_w - e_a}.$$

This ratio was derived in a different manner by Bowen (1926), and is often referred to as the "Bowen ratio."

Values of the ratio R can be computed from climatological charts of the oceans, but a comprehensive study has not been made. Calculations based on data contained in the *Atlas of Climatic Charts of the Oceans*, published by the U. S. Weather Bureau (1938), show that the ratio varies from one part of the ocean to the other. As a rule, the ratio is small in low latitudes, where it remains nearly constant throughout the year, but is greater in middle latitudes, where it reaches values up to 0.5 in winter and in some areas drops to −0.2 in summer. A negative value indicates that heat is conducted from the atmosphere to the sea. On an average, the value for all oceans appears to lie at about 0.1, meaning that about 10 per cent of the heat surplus that the oceans receive by radiation processes is given off as sensible heat, whereas about 90 per cent is used for evaporation (p. 115).

There are certain points regarding the character of the evaporation which need to be emphasized. If the water is warmer than the air, the vapor pressure at the sea surface remains greater than that in the air, and evaporation can always take place and will be greatly facilitated in these circumstances because the turbulence of the air will be fully developed owing to the unstable stratification of the very lowest layers (p. 92). It must therefore be expected that the greatest evaporation occurs when cold air flows over warm water. If the air is much colder than the water, the air may become saturated with water vapor, and fog or mist may form over the water surfaces. Such fog is common in the fall over ponds and small lakes during calm, clear nights. When a wind blows, the moisture will be carried upward, but streaks and columns of fog are often visible over lakes or rivers and are commonly described as "smoke." The process can occasionally be observed near the coast, but not over the open ocean, because the necessary great temperature differences are rapidly eliminated as the distance from the coast increases.

When the sea surface is colder than the air, evaporation can take place only if the air is not saturated with water vapor. In this case turbulence is reduced and evaporation must stop when the vapor content

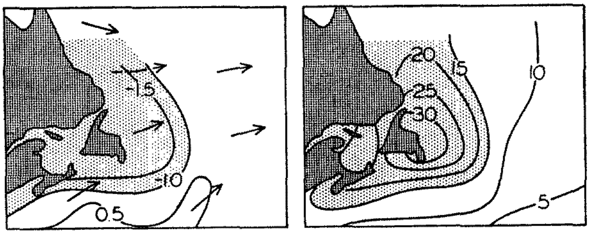


Fig. 26. *Left:* The difference, air minus sea-surface temperature, and the prevailing wind-direction over the Grand Banks of Newfoundland in March, April, and May. *Right:* Percentage frequency of fog in the same months.

of the lowest layer of the atmosphere has reached such a value that the vapor pressure equals that at the sea surface. If warm, moist air passes over a colder sea surface, the direction of transport will be reversed and condensation will take place on the sea surface in such a way that heat is brought to the surface and not carried away from it. Owing to the fact that this process can take place only when the air is *warmer* than the sea and that then turbulence is greatly reduced, one can expect that condensation of water vapor on the sea will not be of great importance, but it should be borne in mind that this process can and does take place when conditions are right. In these circumstances, contact with the sea and conduction lower the air temperature to the dew point for a considerable distance above the sea surface. Condensation takes place in the air and fog is formed, "advection" fog that is commonly encountered over the sea. The relation between the frequency of fog or mist and the difference between sea-surface and air temperatures are well illustrated by charts in the *Atlas of Climatic Charts of the Oceans* (1938). As an example, fig. 26 shows the frequency of fog, the difference between air and sea-surface temperature, and the prevailing wind direction over the Grand Banks of Newfoundland in March, April, and May. It can be concluded that in spring, when the water is colder than the air, no evaporation takes place in this region, but in the fall and winter, when the water is warmer, evaporation must be great.

In middle and higher latitudes the sea surface in winter is mostly warmer than the air, and hence one must expect the evaporation then to be at its maximum. This conclusion appears contrary to common experience that evaporation from heated water is greater than that from cold water, but the contradiction is only apparent, because greatest evaporation always occurs when a water surface is warmer than the air above it, which is exactly what happens in winter.

OBSERVATIONS AND COMPUTATIONS OF EVAPORATION. Present knowledge of the amount of evaporation from the different parts of the oceans is derived partly from observations and partly from computations based on consideration of the heat balance.

Observations have been made by means of pans on board ship, but such observations give values of the evaporation from the sea surface that are too high, partly because the wind velocity is higher at the level of the pan than at the sea surface, and partly because the difference between vapor pressure in the air and that of the evaporating surface is greater at the pan than at the sea surface. Analyzing the decrease of the wind velocity and the increase of the vapor pressure between the average level of pans used on shipboard and a level a few centimeters above the sea surface, Wüst (1936) arrived at the conclusion that the measured values had to be multiplied by 0.53 in order to represent the evaporation from the sea surface.

In computing the evaporation on the basis of the heat balance, one has to begin with the equation (p. 101)

$$Q_s - Q_b - Q_e - Q_h + Q_v + Q_\vartheta = 0.$$

Introducing the ratio $R = Q_h/Q_e$, and converting the evaporation, E, into centimeters by dividing Q_e by the latent heat of vaporization, L, one obtains

$$E = \frac{Q_s - Q_b + Q_v + Q_\vartheta}{L(1 + R)}.$$

In this form the equation representing the heat balance has found wide application for computation of evaporation. The result gives the evaporation in centimeters during the time intervals to which the values Q_s, and so on, apply, provided these are expressed in gram calories.

A second method for computing the evaporation from the oceans has been suggested by Sverdrup (1937), who, on the basis of results in fluid mechanics as to the turbulence of the air over a rough surface, established a formula for the evaporation, using in part constants that had been determined by laboratory experiments and in part constants that were obtained from the character of the variation of vapor pressure with increasing height above the sea surface. Similar but more complicated formulae have been derived by Millar (1937) and by Montgomery (1940).

The exact formulae are not well suited for numerical computation, but at wind velocities between 4 and 12 m/sec, the mean annual evaporation in centimeters can be found approximately from the simple relation

$$E = 3.7(\bar{e}_w - \bar{e}_a)\bar{u},$$

where $\bar{e}_w$ represents the average vapor pressure in millibars at the sea surface as derived from the temperature and salinity of the sea, $\bar{e}_a$ represents the average vapor pressure in the air at a height of 6 m above the sea surface, and $\bar{u}$ is the average wind velocity in meters per second at the same height.

Average Annual Evaporation from the Oceans. On the basis of pan measurements conducted in different parts of the ocean, Wüst (1936) found that the average evaporation from all oceans amounts to 93 cm/year, and he considers this value correct within 10 to 15 per cent. W. Schmidt (1915) computed the evaporation by means of the preceding equation for E, in which the terms Q_ϑ and Q_v can be omitted in considering the oceans as a whole. Schmidt introduced high values of R, and on the basis of the available data as to incoming radiation and back radiation he found a total evaporation of 76 cm/year. A revision based on more recent measurements of radiation (Mosby, 1936) and use of $R = 0.1$ resulted in a value of 106 cm/year. The latter value represents an upper limit, and may be 10 to 15 per cent too high, whence it appears that Wüst's result is nearly correct.

It is of interest in this connection to give some figures regarding the relation between evaporation and precipitation over the oceans, the land areas, and the whole earth (according to Wüst, 1936). The total evaporation from the oceans amounts to 334,000 km^3/year, of which 297,000 km^3 returns to the sea in the form of precipitation, and the difference, 37,000 km^3, must be supplied by run-off, since the salinity of the oceans remains unchanged. The total amount of precipitation falling on the land is 99,000 km^3, of which amount a little over one third, 37,000 km^3, is supplied by evaporation from the oceans and 62,000 km^3 is supplied by evaporation from inland water areas or directly from the moist soil. For the sake of comparison it may be mentioned that the capacity of Lake Mead, above Boulder Dam, is 45 km^3.

Evaporation in Different Latitudes and Longitudes. From pan observations at sea, Wüst has derived average values of the evaporation from the different oceans in different latitudes (table 30, p. 123). By means of the energy equation one can compute similar annual values, assuming that the net transport of heat by ocean currents can be neglected. Such a computation has been carried out for the Atlantic

Ocean, making use of Kimball's data (1928) as to the incoming radiation and the observed temperatures and humidities for determining effective back radiation. In fig. 27 are shown Wüst's values of the annual evaporation between latitudes 50°N and 50°S in the Atlantic Ocean and the corresponding values as derived from the energy equation. The agreement is very satisfactory. The low evaporation in the equatorial regions that both curves show can be ascribed to the higher relative humidities and the lower wind velocities of that area, if one considers the processes of evaporation, or it can be ascribed to the effect of the prevailing cloudiness if one considers the energy relations. The great

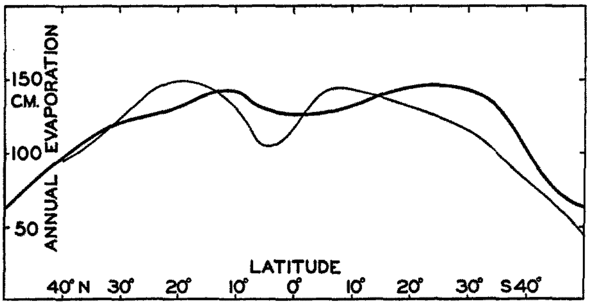


Fig. 27. Annual evaporation from the Atlantic Ocean between latitudes 50°N and 50°S. Thin curve based on observations (Wüst, 1936) and heavy curve on computations, using the energy equation.

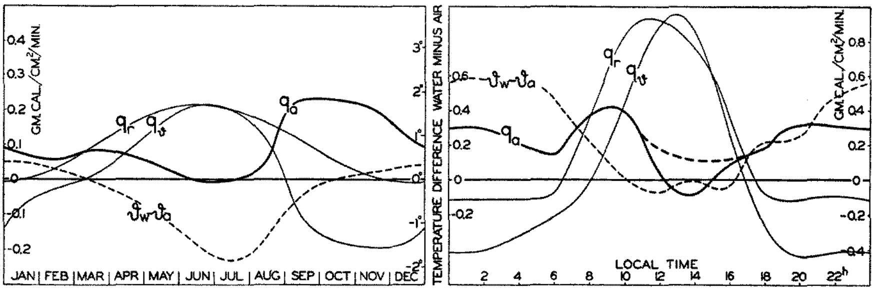
evaporation in the areas of subtropical anticyclones appears clearly, but in the Southern Hemisphere the observations give the highest values of the evaporation nearer to the Equator than do the computations. The discrepancy may be due to the fact that in the course of a year the subtropical anticyclone changes its distance from the Equator and that the observations have not been distributed evenly over the year. The energy equation has also been used by McEwen (1938) for computing values of evaporation over the eastern Pacific Ocean between latitudes 20°N and 50°N. His figures agree with those obtained by Wüst for the same latitudes.

It appears that the average annual values of the evaporation in different latitudes are well established, but the evaporation also varies from the eastern to the western parts of the oceans and with the seasons. These variations are of great importance to the circulation of the atmosphere, because the supply of water vapor that later on condenses and gives off its latent heat represents a large portion of the supply of energy. So far, none of the details are known, but it is possible that approximate values of the evaporation from different parts of the ocean and in different seasons can be found by means of the method proposed by Sverdrup (1937) and used by Jacobs (1942).

ANNUAL VARIATION OF EVAPORATION. The character of the annual variation of evaporation can be examined by means of the energy equation (Sverdrup, 1940):

$$Q_a = Q_e(1 + R) = Q_s - Q_b + Q_\vartheta + Q_v.$$

The quantity Q_ϑ can be computed if the annual variation of temperature due to processes of heating and cooling is known at all depths where such annual variations occur. The annual variation of temperature at the surface has been examined, but only few data are available from subsurface depths, the most reliable being those which have been compiled by Helland-Hansen (1930) from an area in the eastern North Atlantic with its center in 47°N and 12°W (p. 132). The radiation income in that area can be obtained from Kimball's data (1928), the back radiation can be found by means of the diagram in fig. 25, and the transport by cur-



A B

Fig. 28. (A) Annual variation in the total amount of heat, q_a, given off to the atmosphere in an area of the North Atlantic (about 47°N, 12°W). (B) Corresponding diurnal variation near the Equator in the Atlantic Ocean. (For explanation of symbols, see text.)

rents, Q_v, can be neglected. In fig. 28A are represented the annual variation of the net surplus of radiation, Q_r, the annual variation of the amount of heat used for changing the temperature of the water, Q_ϑ, and the difference between these two amounts, Q_a, which represents the total amount of heat given off to the atmosphere. The greater part of the last amount is used for evaporation, and the curve marked Q_a represents, therefore, approximately the annual variation of the evaporation, which shows a maximum in the fall and early winter, a secondary minimum in February, followed by a secondary maximum in March, and a low minimum in summer. In June and July no evaporation takes place. The total evaporation during the year is about 80 cm.

This example illustrates a method of approach that may be applied, but so far the necessary data for a more complete study are lacking. The result that the evaporation is at a minimum in summer and at a

TABLE 30

AVERAGE VALUES OF SALINITY, S, EVAPORATION, E, AND PRECIPITATION, P, AND THE DIFFERENCE, $E - P$, FOR EVERY FIFTH PARALLEL OF LATITUDE BETWEEN 40°N AND 50°S

(After Wüst)

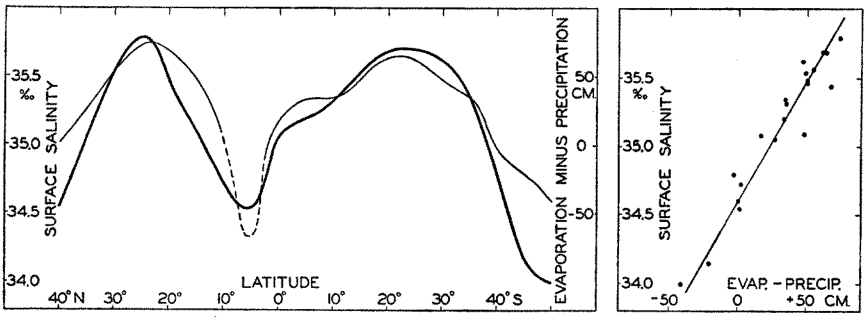
Latitude	Atlantic Ocean				Indian Ocean				Pacific Ocean				All Oceans			
	S (‰)	E (cm/yr)	P (cm/yr)	$E - P$ (cm/yr)	S (‰)	E (cm/yr)	P (cm/yr)	$E - P$ (cm/yr)	S (‰)	E (cm/yr)	P (cm/yr)	$E - P$ (cm/yr)	S (‰)	E (cm/yr)	P (cm/yr)	$E - P$ (cm/yr)
40°N	35.80	94	76	18					33.64	94	93	1	34.54	94	93	1
35	36.46	107	64	43					34.10	106	79	27	35.05	106	79	27
30	36.79	121	54	67					34.77	116	65	51	35.56	120	65	55
25	36.87	140	42	98					35.00	127	55	72	35.79	129	55	74
20	36.47	149	40	110	(35.05)	(125)	(74)	(51)	34.88	130	62	68	35.44	133	65	68
15	35.92	145	62	83	(35.07)	(125)	(73)	(52)	34.67	128	82	46	35.09	130	82	48
10	35.62	132	101	31	(34.92)	(125)	(88)	(37)	34.29	123	127	− 4	34.72	129	127	2
5	34.98	105	144	−39	(34.82)	(125)	(107)	(18)	34.29	102	(177)	(−75)	34.54	110	177	−67
0	35.67	116	96	20	35.14	125	131	− 6	34.85	116	98	18	35.08	119	102	17
5°S	35.77	141	42	99	34.93	121	167	−46	35.11	131	91	40	35.20	124	91	33
10	36.45	143	22	121	34.57	99	156	−57	35.38	131	96	35	35.34	130	96	34
15	36.79	138	19	119	34.75	121	83	38	35.57	125	85	40	35.54	134	85	49
20	36.54	132	30	102	35.15	143	59	84	35.70	121	70	51	35.69	134	70	64
25	36.20	124	40	84	35.45	145	46	99	35.62	116	61	55	35.69	124	62	62
30	35.72	116	45	71	35.89	134	58	76	35.40	110	64	46	35.62	111	64	47
35	35.35	99	55	44	35.60	121	60	61	35.00	97	64	33	35.32	99	64	35
40	34.65	81	72	9	35.10	83	73	10	34.61	81	84	− 3	34.79	81	84	− 3
45	34.19	64	73	− 9	34.25	64	79	−15	34.32	64	85	−21	34.14	64	85	−21
50	33.94	43	72	−29	33.87	43	79	−36	34.16	43	84	−41	33.99	43	84	−41

maximum in fall and early winter is in agreement with the conclusions that were drawn when discussing the process of evaporation in general.

Diurnal Variation of Evaporation. The diurnal variation of evaporation can be examined in a similar manner, but at the present time suitable data are available only at four *Meteor* stations near the Equator in the Atlantic Ocean (Defant, 1932; Kuhlbrodt and Reger, 1938). In fig. 28B the curves marked Q_r and Q_ϑ correspond to the similar curves in fig. 28A, and the difference between these, Q_a, shows the amount of heat lost during twenty-four hours, which is approximately proportional to the evaporation. The diurnal variation of evaporation in the Tropics appears to have considerable similarity to the annual variation in middle latitudes, and is characterized by a double period with maxima in the late forenoon and the early part of the night and minima at sunrise and in the early afternoon hours. It is possible that the afternoon minimum appears exaggerated, owing to uncertainties as to the absolute values of Q_r and Q_ϑ. The total diurnal evaporation was 0.5 cm, but the sky was nearly clear on the four days that were examined and the average diurnal value is therefore smaller. The double diurnal period of evaporation appears to be characteristic of the Tropics, but in middle latitudes a single period with maximum values during the night probably dominates.

Salinity and Temperature of the Surface Layer

The Surface Salinity. In all oceans the surface salinity varies with latitude in a similar manner. It is at a minimum near the Equator, teaches a maximum in about latitudes 20°N and 20°S, and again decreases toward high latitudes.



A B

Fig. 29. (A) Average values for all oceans of surface salinity and the difference, evaporation minus precipitation, plotted against latitude. (B) Corresponding values of surface salinity and the difference, evaporation minus precipitation, plotted against each other (according to Wüst, 1936).

Table 30 contains average values of the surface salinity, the evaporation, the precipitation, and the difference between the last quantities for the three major oceans and for all combined, according to Wüst (1936).

On the basis of these values, Wüst has shown that for each ocean the deviation of the surface salinity from a standard value is directly proportional to the difference between evaporation, E, and precipitation, P. In fig. 29A are plotted the surface salinities for all oceans and the difference, $E - P$, in centimeters per year, as functions of latitude; the corresponding values of salinity and the difference, $E - P$, are plotted against each other. If the values of 5°N are omitted, because they disagree with all others, the values fall nearly on a straight line leading to the empirical relationship

$$S = 34.60 + 0.0175\,(E - P).$$

Wüst points out that such an empirical relationship is found because the surface salinity is mainly determined by three processes: decrease of salinity by precipitation, increase of salinity by evaporation, and change of salinity by processes of mixing. If the surface waters are mixed with water of a constant salinity, and if this constant salinity is represented by S_0, the change of salinity due to mixing must be proportional to $S_0 - S$, where S is the surface salinity. The change of salinity due to processes of evaporation and precipitation must be proportional to the difference $(E - P)$. The local change of the surface salinity must be zero; that is,

$$\partial S/\partial t = a(S_0 - S) + b(E - P) = 0,$$

or

$$S = S_0 + k(E - P).$$

As this simple formula has been established empirically, it must be concluded that the surface water is generally mixed with water of a salinity which, on an average, is 34.6 $^0/_{00}$. This value represents approximately the average value of the salinity at a depth of 400 to 600 m, and it appears therefore that vertical mixing is of great importance to the general distribution of surface salinity. This concept is confirmed by the fact that the standard value of the salinity differs for the different oceans. For the North Atlantic and the North Pacific, Wüst obtains similar relationships, but the constant term, S_0, has the value 35.30 $^0/_{00}$ in the North Atlantic and 33.70 $^0/_{00}$ in the North Pacific Ocean. The corresponding average values of the salinity at a depth of 600 m are 35.5 $^0/_{00}$ and 34.0 $^0/_{00}$, respectively. For the South Atlantic and the South Pacific Oceans, Wüst finds S_0 = 34.50 $^0/_{00}$ and 34.64 $^0/_{00}$, respectively, and the average salinity at 600 m in both oceans is about 34.5 $^0/_{00}$. In these considerations the effect of ocean currents on the distribution of surface salinity has been neglected, and the simple relations obtained indicate that transport by ocean currents is of minor importance as far as average conditions are concerned. The difference between evaporation and precipitation, $E - P$, on the other hand, is of primary importance, and, because this difference is closely related to the circulation of the atmosphere, one is led to the conclusion that the average values of the surface salinity are to a great extent controlled by the character of the atmospheric circulation.

The distribution of surface salinity of the different oceans is shown in chart VI, in which the general features that have been discussed are recognized, but the details are so closely related to the manner in which the water masses are formed and to the types of currents that they cannot be dealt with here.

PERIODIC VARIATIONS OF THE SURFACE SALINITY. Over a large area, variations in surface salinity depend mainly upon variations in the difference between evaporation and precipitation. From Böhnecke's monthly charts (1938) of the surface salinity of the North Atlantic Ocean, mean monthly values have been computed for an area extending between latitudes 18° and 42°N, omitting the coastal areas in order to avoid complications due to shifts of coastal currents. The results of this computation (fig. 30) show the highest average surface salinity, 36.70 ‰, in March, and the lowest, 36.59 ‰, in November. The variations from one month to another are irregular, but on the whole the salinity is somewhat higher in spring than it is in the fall.

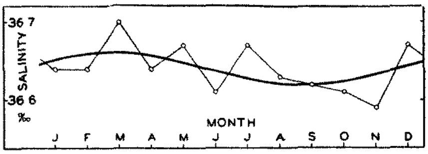


Fig. 30. Annual variation of surface salinity in the North Atlantic Ocean between latitudes 18°N and 42°N.

Harmonic analysis leads to the result

$$S(‰) = 36.641 + 0.021 \cos\left(\frac{2\pi}{T}t - 80°\right),$$

and thus

$$\frac{\partial S}{\partial t} = 0.021\,\frac{2\pi}{T}\cos\left(\frac{2\pi}{T}t - 350°\right).$$

Because $\partial S/\partial t$ is proportional to $E - P$, it follows that the excess of evaporation over precipitation is at a minimum at the end of June and at a maximum at the end of December. This annual variation corresponds closely to the annual variation of evaporation (p. 122), for which reason it appears that in the area under consideration the annual variation of the surface salinity is mainly controlled by the variation in evaporation during the year. For a more exact examination, subsurface data are needed, but nothing is known as to the annual variation of salinity at subsurface depths.

More complicated conditions are encountered in the northwestern part of the Atlantic Ocean, where, according to G. Neumann (1940), between the Azores and Newfoundland the annual variation of salinity has the character of a disturbance that originates to the southwest of Newfoundland and spreads toward the east and east-southeast. Off Newfoundland, the amplitude is 0.37 ‰ and the maximum occurs about March 1 (phase angle equals −60 degrees). Toward the east and east-southeast the amplitude decreases and the maximum is reached later and later, as if the disturbance progressed like a wave that was subject to damping. On the assumption that this damping is caused by horizontal mixing, Neumann finds that the corresponding coefficient of mixing lies between 2×10^8 and 5×10^8 cm²/sec.

From the open ocean no data are available as to the diurnal variation of the salinity of the surface waters, but it may be safely assumed that such a variation is small, because neither the precipitation nor the evaporation can be expected to show any considerable diurnal variation.

TABLE 31

AVERAGE SURFACE TEMPERATURE OF THE OCEANS BETWEEN PARALLELS OF LATITUDE

North latitude	Atlantic Ocean	Indian Ocean	Pacific Ocean	South latitude	Atlantic Ocean	Indian Ocean	Pacific Ocean
70°–60°.....	5.60	.....	.....	70°–60°.....	− 1.30	− 1.50	− 1.30
60 –50......	8.66	.....	5.74	60 –50......	1.76	1.63	5.00
50 –40......	13.16	.....	9.99	50 –40......	8.68	8.67	11.16
40 –30......	20.40	.....	18.62	40 –30......	16.90	17.00	16.98
30 –20......	24.16	26.14	23.38	30 –20......	21.20	22.53	21.53
20 –10......	25.81	27.23	26.42	20 –10......	23.16	25.85	25.11
10 – 0......	26.66	27.88	27.20	10 – 0......	25.18	27.41	26.01

SURFACE TEMPERATURE. The general distribution of surface temperature cannot be treated in a manner similar to that employed by Wüst when dealing with the salinity, because the factors controlling the surface temperature are far more complicated. The discussion must be confined to presentation of empirical data and a few general remarks.

Table 31 contains the average temperatures of the oceans in different latitudes according to Krümmel (1907), except in the case of the Atlantic Ocean, for which new data have been compiled by Böhnecke (1938). In all oceans the highest values of the surface temperature are found somewhat to the north of the Equator, and this feature is probably related to the character of the atmospheric circulation in the two hemispheres. The region of the highest temperature, the thermal Equator, shifts with the season, but in only a few areas is it displaced to the Southern Hemisphere at any season. The larger displacements (Schott, 1935, and Böhnecke, 1938) are all in regions in which the surface currents change during the year because of changes in the prevailing winds, and this feature also is therefore closely associated with the character of the atmospheric circulation. The surface temperatures in the Southern Hemisphere are generally somewhat lower than those in the Northern, and again the difference can be ascribed to difference in the character of the prevailing winds, and perhaps also to a widespread effect of the cold, glacier-covered Antarctic Continent.

The average distribution of the surface temperature of the oceans in February and August is shown in charts II and III. Again the distribution is so closely related to the formation of the different water masses and the character of the currents that a discussion of the details must be postponed.

Difference Between Air and Sea-surface Temperatures. It was pointed out that in all latitudes the ice-free oceans received a surplus of radiation, and that therefore in all latitudes heat is given off from the ocean to the atmosphere in the form of sensible heat or latent heat of water vapor. The sea-surface temperature must therefore, on an average, be higher than the air temperature. Observations at sea have shown that such is the case, and from careful determinations of air temperatures over the oceans it has furthermore been concluded that the difference, air minus sea-surface temperature, is greater than that derived from routine ships' observations. In order to obtain an exact value, it is necessary to measure the air temperature on the windward side of the vessel at a locality where no eddies prevail, but where the air reaches the thermometer without having passed over any part of the vessel. For measurements of the temperature a ventilated thermometer must be used. According to the *Meteor* observations (Kuhlbrodt and Reger, 1938) the air temperature over the South Atlantic Ocean between latitudes 55°S and 20°N is on an average 0.8 degree lower than that of the surface, whereas in the same region the atlas of oceanographic and meteorological observations published by the Netherlands Meteorological Institute gives an average difference of only 0.1 degree. The reason for this discrepancy is that the air temperatures as determined on commercial vessels are on an average about 0.7 degree too high because of the ships' heat. The result as to the average value of the difference, $\vartheta_w - \vartheta_a$, is in good agreement with results obtained on other expeditions when special precautions were taken for obtaining correct air temperatures. Present atlases of air and sea-surface temperatures have been prepared from the directly observed values on board commercial vessels without application of a correction to the air temperatures. This correction is so small that it is of minor importance when the atlases are used for climatological studies, but in any studies that require knowledge of the exact difference between air- and sea-surface temperatures it is necessary to be aware of the systematic error in the air temperature.

The difference of 0.8 degree between air and surface temperatures, as derived from the *Meteor* observations, is based on measurements of air temperature at a height of 8 m above sea level. At the very sea surface the air temperature must coincide with that of the water, and consequently the air temperature decreases within the layers directly above the sea. The most rapid decrease takes place, however, very close to the sea surface, and at distances greater than a few meters the decrease is so slow that it is immaterial whether the temperature is measured at 6, 8, or 10 m above the surface. The height at which the air temperature has been observed on board a ship exercises a minor influence, therefore, upon the accuracy of the result, and discrepancies due to differences in the height of observations are negligible compared to the errors due to inadequate exposure of the thermometer.

The statement that the air temperature is lower than the water temperature is correct only when dealing with average conditions. In any locality the difference, $\vartheta_w - \vartheta_a$, generally varies during the year in such a manner that in winter the air temperature is much lower than the sea-surface temperature, whereas in summer the difference is reduced or the sign is reversed. The difference also varies from one region to another according to the character of the circulation of the atmosphere and of the ocean currents. These variations are of great importance to the local heat budget of the sea because the exchange of heat and vapor between the atmosphere and the ocean depends greatly upon the temperature difference.

It was shown that the amount of heat given off from the ocean to the surface is, in general, great in winter and probably negligible in summer. Owing to this annual variation in the heat exchange, one must expect that in winter the air over the oceans is much warmer than the air over the continents, but in summer the reverse conditions should be expected. That such is true is evident from a computation of the average temperature of the air between latitudes 20°N and 80°N along the meridian of 120°E, which runs entirely over land, and along the meridian of 20°W, which runs entirely over the ocean (von Hann, 1915, p. 146). In January the average temperature along the "land meridian" of 120°E is −15.9 degrees C, but along the "water meridian" of 20°W it is 6.3 degrees. In July the corresponding values are 19.4 and 14.6 degrees, respectively. Thus, in January the air temperature between 20°N and 80°N over the water meridian is 22.2 degrees higher than that over the land meridian, whereas in July it is 4.8 degrees lower. The mean annual temperature is 7.0 degrees higher along the water meridian.

Annual Variation of Surface Temperature. The annual variation of surface temperature in any region depends upon a number of factors, foremost among which are the variation during the year of the radiation income, the character of the ocean currents, and of the prevailing winds. The character of the annual variation of the surface temperature changes from one locality to another, but a few of the general features can be pointed out. The heavy curves in fig. 31 show the average annual range of the surface temperature in different latitudes in the Atlantic, the Indian, and the Pacific Oceans. The range represents the difference between the average temperatures in February and August and is derived for the Atlantic from Böhnecke's tables (1938), and, for the Indian and the Pacific Oceans, from the charts published

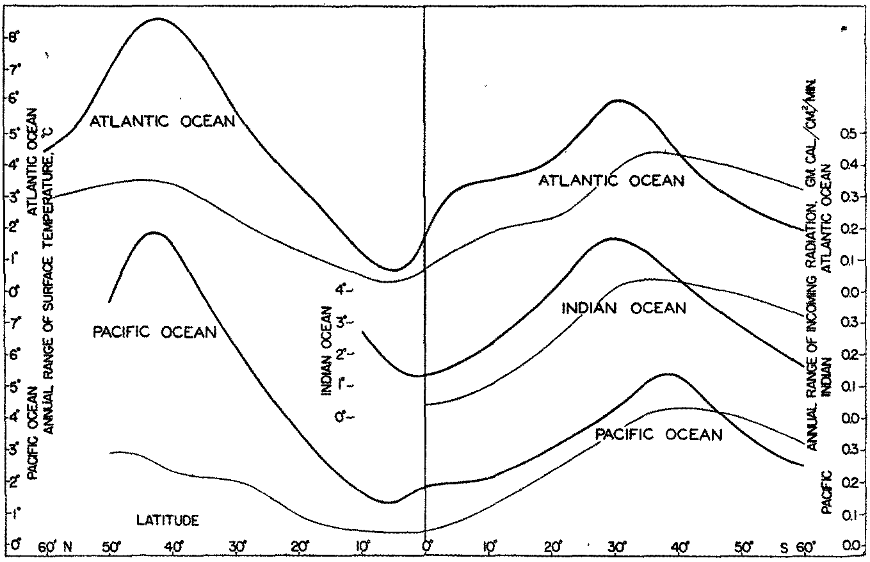


Fig. 31. Average annual ranges of surface temperature in the different oceans plotted against latitude (heavy curves) and corresponding ranges in the radiation income (thin curves).

by Schott (1935). Thin lines in the same figure show the range of the radiation income as derived from Kimball's maps (1928). The curves bring out two characteristic features. In the first place they show that the annual range of the surface temperature is much greater in the North Atlantic and the North Pacific Oceans than in the southern oceans. In the second place they show that in the southern oceans the temperature range is definitely related to the range in radiation income, whereas in the northern oceans no such definite relation appears to exist. The great ranges in the northern oceans are associated with the character of the prevailing winds and, particularly, with the fact that cold winds blow from the continents toward the ocean and greatly reduce the winter temperatures. Near the Equator a semiannual variation is present, corresponding to the semiannual period of radiation income, but in middle and higher latitudes the annual period dominates.

ANNUAL VARIATION OF TEMPERATURE IN THE SURFACE LAYERS. At subsurface depths the variation of temperature depends upon four factors: (1) variation of the amount of heat that is directly absorbed at different depths, (2) the effect of heat conduction, (3) variation in the currents related to lateral displacement of water masses, and (4) the effect of vertical motion. The annual variation of temperature at subsurface depths cannot be dealt with in a general manner, owing to lack of data, but it is again possible to point out some outstanding characteristics, using two examples from the Pacific and one from the Atlantic Ocean. The effects of all four of the important factors are illustrated in fig. 32A, which shows the annual variation of temperature at the surface and at depths of 25, 50, and 100 m at Monterey Bay,

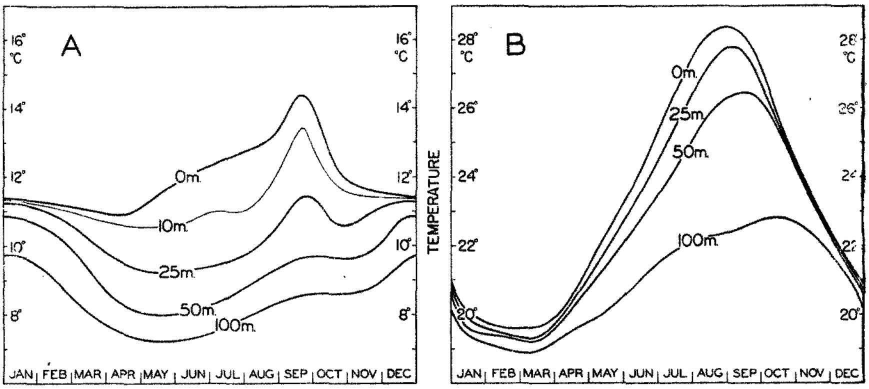


Fig. 32. (A) Annual variation of temperature at different depths in Monterey Bay, California. (B) Annual variation of temperature at different depths in the Kuroshio off the South Coast of Japan.

California (Skogsberg, 1936). Skogsberg divides the year into three periods: the period of the Davidson Current, lasting from the middle of November to the middle of February; the period of upwelling, between the middle of February and the end of July; and the oceanic period, from the end of July to the middle of November. The California Current off Monterey Bay during the greater part of the year is directed to the south, but during winter, from the middle of November to the middle of February, an inshore flow to the north, the Davidson Current, is present (p. 724). The water of this inshore flow is characterized by relatively high and uniform temperature and appears in the annual variation of temperature as warm water at subsurface depths. The upper homogeneous layer is relatively thick, as is evident from the fact that the temperature is nearly the same at 25 m as it is at the surface, and that at 50 m it is only slightly lower. At the end of February the California Current again reaches to the coast and, under the influence of the prevailing northwesterly winds, an overturn of the upper layers takes place that is generally described as upwelling (p. 501). During the period of upwelling, vertical motion near the coast brings water of relatively low temperature toward the surface. Consequently, the temperatures at given depths decrease when the upwelling begins. This decrease is brought out in fig. 32A by the downward trend of the temperature at 25, 50, and 100 m, at which depths the minimum temperature is reached at the end of May. The much higher temperature at the surface as compared to that at 25 m shows that a thin surface layer is subject to heating by radiation, and from the variation of temperature at 10 m, which is shown by a thin line, it is evident that the effect of heating is limited to the upper 10 m. As the upwelling gradually ceases toward the end of August, a sharp rise in temperature takes place both at the surface and at subsurface depths, and the peaks shown by the temperature curves in September are results of heating and conduction and intrustion of offshore water. Thus, the annual march of temperature can be explained from changes in currents, vertical motion associated with upwelling, seasonal heating and cooling, and heat conduction.

The annual variation of temperature in the Kuroshio off the south coast of Japan (Koenuma, 1939), as shown in fig. 32B, gives an entirely different picture. The annual variation has the same character at all depths between the surface and 100 m, with a minimum in late winter and a maximum in late summer or early fall, but the range of the variation decreases with depth, and the time of maximum temperature occurs later with increasing depth. From the course of the curves it may be concluded that the annual variation is due to heating and cooling near the surface and is transmitted to greater depths by processes of conduction (p. 136). This appears to be correct, but the heating and cooling is only partly caused by variations in the net radiation, and it also depends on excessive cooling in winter by cold and dry winds blowing toward the sea (Sverdrup, 1940).

In order to be certain that observed temperature variations are related to processes of heating and cooling only, it is necessary to examine whether the water in a given locality is of the same character throughout the year. For this purpose Helland-Hansen (1930) developed a method that is applicable in areas in which it is possible to establish a definite relation between temperature and salinity (p. 142). He assumed that any temperature value above or below that determined by the temperature-salinity relation can be considered as resulting from heating or cooling of the water, and he used the method within three areas in the eastern North Atlantic. Fig. 33 shows the curves which he determined for an area off the Bay of Biscay with its center in approximately 47°N and 12°W. The character of the curves, the reduction of the range, and the displacement of the times of maxima clearly show that one has to deal with heat conduction. In this case the variation in the heat content corresponds nearly to the variation in net radiation, whereas in the Kuroshio the additional effect of excessive cooling in winter by winds from the continent leads to much greater variations in temperature and heat content.

These examples serve to illustrate different types of annual variation of temperature that may be encountered in different localities and also to stress the fact that conclusions as to the temperature variations associated with processes of local heating are valid only if the data are such that the influences of shifting currents and vertical motion can be eliminated.

Diurnal Variation of Surface Temperature. The range of the diurnal variation of surface temperature of the sea is not more, on an

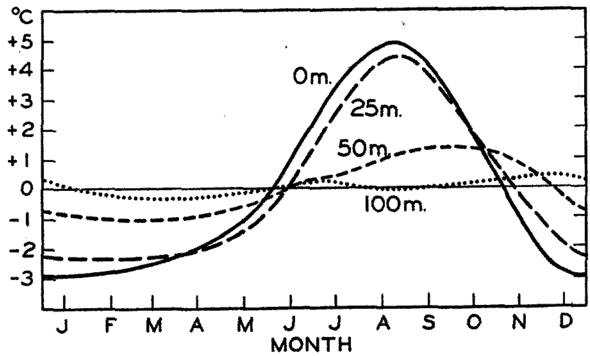


Fig. 33. Annual variation of temperature at different depths off the Bay of Biscay in approximately 47°N and 12°W.

average, than 0.2 to 0.3 degree. Earlier observations gave somewhat higher values, particularly in the Tropics, but new, careful measurements and reexamination of earlier data in which doubtful observations have been eliminated have shown that the range of diurnal variation is quite small. Meinardus (Kuhlbrodt and Reger, 1938, p. 301–302) summarizes his examination of a large number of data by stating, "in general, the diurnal variation of water temperature in lower latitudes can be represented by a sine curve with extreme values between $2{:}30^h$ and 3^h and between $14{:}30^h$ and 15^h, and a range of 0.3° to 0.4°. In higher latitudes the extreme values come later and the range is even smaller." The *Meteor* observations give ranges of only 0.2 to 0.3 degree in the Tropics. The *Meteor* data and the *Challenger* data, which have been discussed by Wegemann (1920), both show that close to the Equator the diurnal variation of surface temperature is somewhat unsymmetrical, the temperature increasing rapidly after sunrise and decreasing slowly after sunset, but at greater distances from the Equator the curve becomes somewhat more symmetrical.

The changes through the year of the range of diurnal variation of surface temperature have been examined in some coastal areas. At forty-four stations around the British Isles, Dickson (see Wegemann, 1920) found that on an average the diurnal range varied between 0.20 degree in December and 0.69 degree in May. At individual stations both the mean annual range and the variation of the range from month to month were dependent upon the exposure of the locality and the depth of the water at which the measurements were made. This annual variation in range is closely related to the annual variation in the diurnal amount of net heat received by processes of radiation.

TABLE 32

RANGE OF DIURNAL VARIATION OF SURFACE TEMPERATURE IN THE TROPICS

Wind and cloudiness	Temperature range, °C		
	Average	Maximum	Minimum
1. Moderate to fresh breeze			
a. Sky overcast	0.39	0.6	0.0
b. Sky clear	0.71	1.1	0.3
2. Calm or very light breeze			
a. Sky overcast	0.93	1.4	0.6
b. Sky clear	1.59	1.9	1.2

The range of the diurnal variation of temperature depends upon the cloudiness and the wind velocity. From observations in the Tropics, Schott (Krümmel, 1907) found the mean and extreme values that are shown in table 32. Similar results but higher numerical values were found by Wegemann from the *Challenger* data. In both cases the numerical values may be somewhat in error, but the character of the influence of cloudiness and wind is quite evident. With a clear sky the range of the diurnal variation is great, but with great cloudiness it is small, at calm or light breeze it is great, and at moderate or high wind it is small. The effect of cloudiness is explained by the decrease of the diurnal amplitude of the incoming radiation with increasing cloudiness. The effect of the wind is somewhat more complicated, but the main feature is that at high wind velocities the wave motion produces a thorough mixing in the surface layers and the heat which is absorbed in the upper meters is distributed over a thick layer, leading to a small range of the temperature, whereas in calm weather a corresponding intensive mixing does not take place, the heat is not distributed over a thick layer, and consequently the range of the temperature near the surface is much greater.

DIURNAL VARIATION OF TEMPERATURE IN THE UPPER LAYERS. Knowledge as to the diurnal variation of temperature at depths below the surface is very scanty. It can be assumed that the depth at which the diurnal variation is perceptible will depend greatly upon the stratification of the water. A sharp increase of density at a short distance below the free-water surface will limit the conduction of heat (p. 477) to such an extent that a diurnal variation of temperature will be present above the boundary surface only.

On the *Meteor* expedition, hourly temperature observations were made at the surface and at a depth of 50 m at a few stations in the Tropics where an upper homogeneous layer was present which had a thickness of 70 m. Defant (1932) has shown that in these cases the diurnal oscillation of temperature at subsurface depths is in agreement with the laws that have been derived on the assumption of a constant heat conductivity (p. 136). At 50 m the amplitude of the diurnal variation was reduced to less than two tenths of the amplitude at the surface, and the maximum occurred about 6.5 hours later.

The diurnal variation of sea temperature in general is so small that it is of little importance to the physical and biological processes in the sea, but knowledge of the small variations is essential to the study of the diurnal exchange of heat between the atmosphere and the sea (p. 124). The data which are available for this purpose, however, are very inadequate at the present time.

Theory of the Periodic Variations of Temperature at Subsurface Depths

Subsurface temperature variations due to processes of heat conduction can be studied by means of the equation (p. 159)

$$\frac{\partial \vartheta}{\partial t} = \frac{\partial}{\partial z}\left(\frac{A}{\rho}\frac{\partial \vartheta}{\partial z}\right), \qquad \text{(IV, 1)}$$

where ρ is the density and A is the eddy conductivity which, in general, varies with depth and time. When one writes $\partial\vartheta/\partial t$, the local change in temperature, in this form it is supposed that heat conduction takes place in a vertical direction only and that advection can be neglected. The terms "local change" and "advection" are explained on p. 157. An integral of equation (1) is easily found if A/ρ is constant, if the average temperature is a linear function of depth, and if the temperature variations at the surface ($z = 0$) can be represented by means of a series of harmonic terms:

$$\vartheta_0 = \bar{\vartheta} + a_1 \cos(\sigma t - \alpha_1) + a_2 \cos(2\sigma t - \alpha_2) + \cdots, \qquad \text{(IV, 2)}$$

where $\sigma = 2\pi/T$ and T is the period length of the first harmonic term. Then

$$\vartheta_z = \bar{\vartheta} + bz + a_1 e^{-r_1 z}\cos(\sigma t - \alpha_1 - r_1 z) + a_2 e^{-r_2 z}\cos(2\sigma t - \alpha_2 - r_2 z) + \cdots, \qquad \text{(IV, 3)}$$

where

$$r_1 = \sqrt{\frac{\sigma\rho}{2A}}, \qquad r_2 = \sqrt{\frac{2\sigma\rho}{2A}} \quad \cdots \tag{IV, 4}$$

Thus, the amplitudes of the harmonic terms decrease exponentially with depth and the phase increases linearly.

Defant (1932) has shown that at the *Meteor* anchor station no. 288 in latitude 12°38′N, longitude 47°36′W, the diurnal variation of temperature in the upper homogeneous layer indicated a constant eddy conductivity. The amplitudes of the diurnal term at the surface and at 50 m were 0.093 and 0.017 degree, respectively, and the phase difference was 6.5 hours. With $\rho = 1.024$, and $T = 24$ hours, he obtained from both the decrease of the amplitude and the difference in phase angle $A = 320$ g/cm/sec.

In cases in which the annual variation of temperature has been examined, the decrease of amplitude and change of phase give different values of A, indicating that A is not independent of depth and time, as assumed when performing the integration. Fjeldstad (1933) has developed a method for computing the eddy conductivity if it changes with depth, provided that the periodic temperature variations are known at a number of depths between the surface and a depth, h, at which they are supposed to vanish. He arrives at the formula

$$\frac{A}{\rho} = \frac{n\sigma}{a_n{}^2 \dfrac{\partial \alpha_n}{\partial z}} \int_z^h a_n{}^3\, dz, \tag{IV, 5}$$

where a_n is the amplitude of the nth harmonic term and α_n is the phase angle.

Fjeldstad has applied the method to the annual temperature variations off the Bay of Biscay, which have been determined by Helland-Hansen (p. 132). He found, with $\rho = 1.025$,

Depth (m)	0	25	50	100
Amplitude, a_1, °C	3.78	3.24	1.24	0.23
Phase angle, α_1	225.1°	235.2°	254.7°	289.3°
Eddy conductivity, g/cm/sec	16.4	3.2	2.1	3.8

Several features show, however, that the observed temperature variations cannot be accounted for by assuming that the eddy conductivity varies with depth only, and variations with seasons also must be considered. Fjeldstad has examined this question and finds that the conductivity reaches a maximum in spring when the stability is at a minimum, but the values remain small throughout the year.

Fjeldstad's method can also be applied to the annual variation of temperature in the Kuroshio, which has been discussed by Koenuma (1939) (fig. 32B, p. 131). It is necessary, however, to make the reservation that in the Kuroshio area the advection term (p. 159) is great (Sverdrup, 1940) and that the use of equation (IV, 5) is therefore correct only if the advection term is independent of time and depth. The harmonic constants and the results, with $\rho = 1.025$, are

Depth (m)	0	25	50	100	200
Amplitude, a_1, °C	4.26	3.97	3.49	2.09	0.71
Amplitude, a_2, °C	0.58	0.49	0.44	0.39	0.14
Phase angle, α_1	250.2°	253.5°	258.7°	271.8°	289.3°
Phase angle, α_2	71.4°	81.0°	100.0°	135.5°	152.6°
Eddy conductivity, A_1, g/cm/sec	78	34	23	22	29
Eddy conductivity, A_2, g/cm/sec	58	43	39	32	26

Both the annual and the semiannual periods have been used for computing the eddy conductivity, and the agreement between the values of A derived from them must be considered satisfactory, in view of the small amplitudes of the semiannual variation. The numerical values decrease with depth, but are much greater than off the Bay of Biscay, as might be expected, because the high velocity of the Kuroshio must lead to intense turbulence. A possible annual variation of the eddy conductivity has not been examined.

In the Kuroshio region, where the velocity of the current is great and the turbulence correspondingly intense, the annual variation of temperature becomes perceptible to a depth of about 300 m, but in the Bay of Biscay it is very small at 100 m. It can therefore safely be concluded that below a depth of 300 m the temperature of the ocean is not subject to any annual variation.

The eddy conductivity off the Bay of Biscay and in the Kuroshio region is much smaller than that in the upper homogeneous layer near the Equator. The difference can be ascribed to the facts that in the former localities the density increases with depth and that the eddy conductivity is greatly reduced where this takes place (p. 477).

Distribution of Density

The distribution of the density of the ocean waters is characterized by two features. In a vertical direction the stratification is generally stable (p. 416), and in a horizontal direction differences in density can exist only in the presence of currents. The general distribution of density is therefore closely related to the character of the currents, but for the present purposes it is sufficient to emphasize the point that in every ocean region water of a certain density which sinks from the sea surface tends to sink to and spread at depths where that density is found.

Since the density of sea water depends on its temperature and salinity, all processes that alter the temperature or the salinity influence the density. At the surface the density is decreased by heating, addition of precipitation, melt-water from ice, or run-off from land, and is increased by cooling, evaporation, or formation of ice. If the density of the surface water is increased beyond that of the underlying strata, vertical convection currents arise that lead to the formation of a layer of homogeneous water. Where intensive cooling, evaporation, or freezing takes place, the vertical convection currents penetrate to greater and greater depths until the density has attained a uniform value from the surface to the bottom. When this state has been established, continued increase of the density of the surface water leads to an accumulation of the densest water near the bottom, and, if the process continues in an area which is in free communication with other areas, this bottom water of great density spreads to other regions. Where deep or bottom water of greater density is already present, the sinking water spreads at an intermediate level.

In the open oceans the temperature of the surface water in lower and middle latitudes is so high that the density of the water remains low even in regions where excess evaporation causes high salinities. In these latitudes convection currents are limited to a relatively thin layer near the surface and do not lead to the formation of deep or bottom water. Such formation takes place mainly in high latitudes where, however, the excess of precipitation in most regions prevents the development of convection currents that reach great depths. This excess of precipitation is so great that deep and bottom water is formed only in two cases: (1) if water of high salinity which has been carried into high latitudes by currents is cooled, and (2) if relatively high-salinity water freezes.

The first conditions are encountered in the North Atlantic Ocean where water of the Gulf Stream system, the salinity of which has been raised in lower latitudes by excessive evaporation, is carried into high latitudes. In the Irminger Sea, between Iceland and Greenland, and in the Labrador Sea this water is partly mixed with cold water of low salinity which flows out from the Polar Sea (p. 682). This mixed water has a relatively high salinity, and, when cooled in winter, convection currents that may reach from the surface to the bottom develop before any formation of ice begins. In this manner deep and bottom water is formed which has a high salinity and a temperature which lies several degrees above the freezing point of the water (table 82, p. 683). A similar process takes place in the Norwegian Sea, but there deep and bottom water is formed at a temperature that deviates only slightly from the freezing temperatures (p. 657).

In the Arctic the second process is of minor importance. There the salinity of the surface layers is very low in the regions where freezing occurs, mainly because of the enormous masses of fresh water that are carried into the sea by the Siberian rivers. Close to the Antarctic

Continent formation of bottom water by freezing is of the greatest importance. At some distance from the Antarctic Continent the great excess of precipitation maintains a low surface salinity, and in these areas winter freezing is not great enough to increase the salinity sufficiently to form bottom water, but on some parts of the continental shelf surrounding Antarctic a rapid freezing in winter leads to the formation of a homogeneous water that attains a higher density than the water off the shelf, and therefore flows down the continental slope to the greatest depths. When sinking, the water is mixed with circumpolar water of somewhat higher temperature and salinity, and hence the resulting bottom water has a temperature slightly above freezing point (p. 611). An active production of bottom water takes place to the south of the Atlantic Ocean, but not within the Antarctic part of the Pacific Ocean.

In some isolated adjacent seas the evaporation may be so intense that a moderate cooling leads to the formation of bottom water. This is the case in the Mediterranean Sea and the Red Sea, and to some extent in the inner part of the Gulf of California, in which the bottom water has a high temperature and salinity and is formed by winter cooling of water whose salinity has been increased greatly by evaporation. Where such seas are in communication with the open oceans, deep water flows out over the sill, mixes with the water masses of the ocean, and spreads out at an intermediate depth corresponding to its density (pp. 670 and 693).

In general, the *water of the greatest density is formed in high latitudes*, and because this water sinks and fills all ocean basins, the deep and bottom water of all oceans is cold. Only in a few isolated basins in middle latitudes is relatively warm deep and bottom water encountered. When spreading out from the regions of formation the bottom water receives small amounts of heat from the interior of the earth, but this heat is carried away by eddy conduction and currents, and its effect on the temperature distribution is imperceptible.

Sinking of surface water is not limited to regions in which water of particularly high density is formed, but occurs also wherever converging currents (convergences) are present, the sinking water spreading at intermediate depths according to its density. In general, the density of the upper layers increases from the Tropics toward the Poles, and water that sinks at a convergence in a high latitude therefore spreads at a greater depth than water that sinks at a convergence in middle latitudes.

The most conspicuous convergence is the Antarctic Convergence, which can be traced all around the Antarctic Continent (fig. 158, p. 606). The water that sinks at this convergence has a low salinity, but it also has a low temperature and consequently a relatively high density. This water, the Antarctic Intermediate Water, spreads directly over the deep water and is present in all southern oceans at depths between 1200 and

800 m. The corresponding Arctic Convergence is poorly developed in the North Atlantic Ocean, where an Atlantic Arctic Intermediate Water is practically lacking, but is found in the North Pacific, where Pacific Arctic Intermediate Water is typically present.

In middle and lower latitudes two more convergences are found, the Subtropical and the Tropical Convergences. These are not so well defined as the Antarctic Convergence, but must be considered more as *regions* in which converging currents are present. The Subtropical Convergence is located in latitudes in which the density of the upper layers increases rapidly toward the Poles. The sinking water therefore has a higher density the farther it is removed from the Equator and will spread out at the greater depths.

In the Tropics the density of the surface water is so low that, regardless of how intense a convergence is, water from the surface cannot sink to any appreciable depth but spreads out at a short distance below the surface. A sharp boundary surface develops between this light top layer and the denser water at some greater depth.

In order to account for the general features of the density distribution in the sea, emphasis has been placed on descending motion of surface water, but regions evidently must exist in which ascending motion prevails, because the amount of water that rises toward the surface must exactly equal the amount that sinks. Ascending motion occurs in regions of diverging currents (divergences), which may be present anywhere in the sea but which are particularly conspicuous along the western coasts of the continents, where prevailing winds carry the surface waters away from the coasts. There, the *upwelling* of subsurface water takes place, which will be described when dealing with specific areas. The upwelling brings water of greater density and lower temperature toward the surface and exercises therefore a widespread influence upon conditions off coasts where the process takes place, but the water rises from depths of less than a few hundred meters. Ascending motion takes place on a large scale around the Antarctic Continent, particularly to the south of the Atlantic Ocean, where rising *deep water* replaces water that contributes to the formation of the Antarctic bottom water and also replaces water that sinks at the Antarctic Convergence.

It is evident from these considerations that in middle and low latitudes the vertical distribution of density to some extent reflects the horizontal distribution at or near the surface between the Equator and the Poles. It is also evident that, in general, the deeper water in any vertical column is composed of water from different source regions and was once present in the surface somewhere in a higher latitude. Such generalizations are subject, however, to modifications within different ocean areas, owing to the character of the currents, and these modifications will be discussed when dealing with the different oceans.

Subsurface Distribution of Temperature and Salinity

The general distribution of temperature is closely related to that of the density. In high latitudes the temperature is low from the surface to the bottom. The bottom and deep waters that spread out from high latitudes retain their low temperature, but in middle and lower latitudes a warm top layer is present the thickness of which depends partly on the processes of heating and cooling at the surface and partly on the character of the ocean currents. This upper layer of warm water is separated from the deep water by a transition layer within which the temperature decreases rapidly with depth. From analogy with the atmosphere, Defant (1928) has applied the terms *troposphere* and *stratosphere* to two different parts of the ocean. *Troposphere* is applied to the upper layer of relatively high temperature that is found in middle and lower latitudes and within which strong currents are present, and *stratosphere* to the nearly uniform masses of cold deep and bottom water. This distinction is often a useful one, particularly when dealing with conditions in lower latitudes, but it must be borne in mind that the terms are based on an imperfect analogy with atmospheric conditions and that only some of the characteristics of the atmospheric stratosphere find their counterparts in the sea.

So far we have mainly considered an ideal ocean extending to high northerly and high southerly latitudes. Actually, conditions may be complicated by communication with large basins that contribute to the formation of deep water, such as the Mediterranean Sea, but these cases will be dealt with specifically when we consider the different regions. Conditions will be modified in other directions in the Indian and Pacific Oceans, which are in direct communication with only one of the polar regions, and these modifications will also be taken up later. Here it must be emphasized that the general distribution of temperature is closely related to the distribution of density, which again is controlled by external factors influencing the surface density and the type of deep-sea circulation.

The general distribution of salinity is more complicated than that of temperature. Within the oceanic stratosphere the salinity is very uniform, but within the troposphere it varies greatly, being mainly related to the excess of evaporation over precipitation. The distribution of surface salinity, which was discussed on pp. 124 and 125 is, in general, characteristic of the distribution within the troposphere, as is evident from the vertical section in figs. 210 and 212, which will be dealt with in detail later on.

The Water Masses of the Oceans

THE *T-S* DIAGRAM. Water masses can be classified on the basis of their temperature-salinity characteristics, but density cannot be used for classification, because two water masses of different temperatures and salinities may have the same density. For the study of water masses it is convenient to make use of the *temperature-salinity diagram,* which was introduced by Helland-Hansen (1916). Helland-Hansen points out that when in a given area the temperatures and corresponding salinities of the subsurface water are plotted against each other, the points generally fall on a well-defined curve, the *T-S* curve, showing the temperature-salinity relationship of the subsurface water of that region. Surface data have to be omitted, because annual variations and local modifications lead to discrepancies.

The corresponding temperature and salinity values in a water column are found to arrange themselves according to depth. The depths of the observed values can be entered along the *T-S* curve, which then will

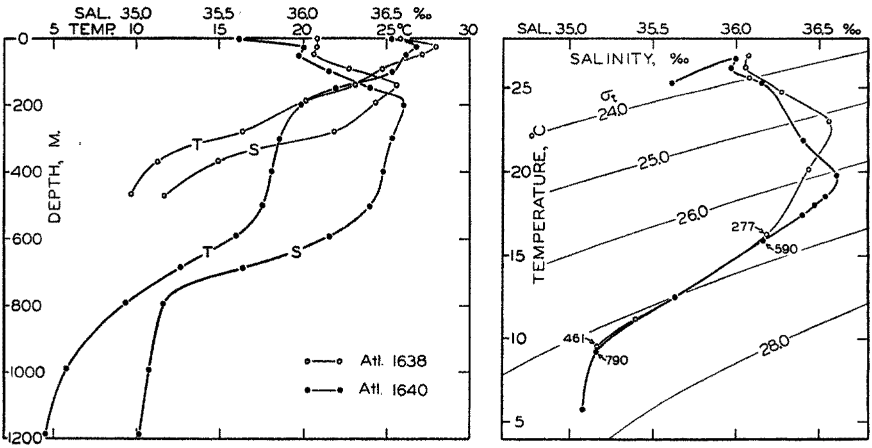


Fig. 34. *Left:* Temperature and salinity at *Atlantis* stations 1638 and 1640 in the Gulf Stream off Onslow Bay plotted against depth. *Right:* The same data plotted in a *T-S* diagram in which σ_t-curves have been entered.

also give information as to variation of temperature and salinity with depth.

Since the density of the water at atmospheric pressure, which is expressed by means of σ_t (p. 56), depends only on temperature and salinity, curves of equal values of σ_t can be plotted in the *T-S* diagram. If a sufficiently large scale is used, the exact σ_t value corresponding to any combination of temperature and salinity can be read off and, if a small scale is used, as is commonly the case, approximate values can be obtained. The slope of the observed *T-S* curve relative to the σ_t curves gives immediately an idea of the stability of the stratification (p. 417).

A *T-S* diagram is shown on the right in fig. 34. On the left in the same figure the observed temperatures and salinities at the *Atlantis* stations 1638 and 1640 in the Gulf Stream off Onslow Bay are plotted against depth, and on the right the same values are entered in a *T-S* diagram. The depths of a few of the observations are indicated. In this case, the temperature-salinity values between 277 and 461 m at station 1638 agree with those between 590 and 790 m at station 1640, indicating that at the two stations water of similar characteristics was present, but at different depths.

The *T-S* diagram has become one of the most valuable tools in physical oceanography. By means of this diagram characteristic features of the temperature-salinity distribution are conveniently represented and anomalies in the distribution are easily recognized. The diagram is also widely used for detecting possible errors in the determination of temperature or salinity (p. 358).

WATER MASSES AND THEIR FORMATION. Following Helland-Hansen's original suggestion, a *water mass* is defined by a *T-S* curve, but in exceptional cases it may be defined by a single point in a *T-S* diagram; that is, by means of a single temperature and a single salinity value. These exceptional cases are encountered in basins where homogeneous water is present over a wide range of depth. A *water type*, on the other hand, is defined by means of single temperature and salinity values, but a given water type is generally present along a surface in the sea and has no thickness. Only in the exceptional cases that were referred to are the terms "water type" and "water mass" interchangeable, but in oceanographic literature the terms have been used loosely and without the distinction that has been introduced here.

In many areas the *T-S* curves are straight lines or can be considered as composed of several pieces of straight lines. Elementary considerations show that a linear relation between temperature and salinity must result if the water types that can be defined by the end points of the straight line mix in different proportions. Similarly, a curved *T-S* relation may result from the mixing of three different types of water. Fig. 35 illustrates in two simple cases how progressive mixing alters the temperature-salinity relation. These considerations are of a formalistic nature, but have in many instances led to the concept that certain water *types* exist and that the *T-S* relations that are observed represent the end results of mixing between the types. This concept presupposes that the water types (often referred to as water masses) are continually renewed, because, if that were not the case, processes of mixing would ultimately lead to the formation of homogeneous water. It is possible, however, to account for the character of the *T-S* curves in the ocean by considering other processes.

In the first place it should be observed that a water mass of uniform temperature and salinity is rarely formed in the open ocean. In high latitudes, where convection currents in winter may reach to the bottom, most of the deep and bottom water will not be uniform, because in some years the density of the surface water will be greater than in other years and the convection current will reach to different depths, depending upon how much the density of the surface water has been increased. As a consequence, even in these areas the density increases toward the bottom. The bottom water is not homogeneous, and shows therefore a definite temperature-salinity relationship. In the second place, sinking at convergences in middle latitudes may lead, as pointed out by Iselin (1939), to the formation of a water mass with a T-S curve that reflects the horizontal distribution of temperature and salinity at the surface. The upper part of fig. 36 illustrates this point. The figure represents a schematic cross section in which are entered isotherms and isohalines

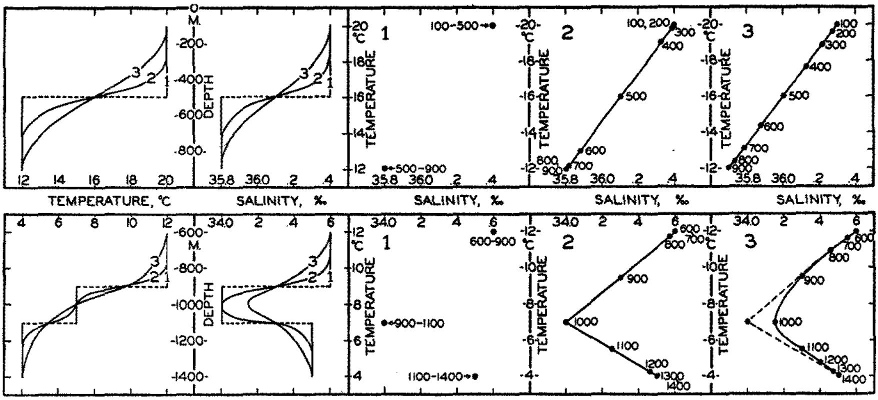


Fig. 35. Diagrammatic representation of results of vertical mixing of water types. To the left the results of mixing are shown by temperatures and salinities as functions of depth, and to the right are shown in three T-S diagrams the initial water types (1) and the T-S relations produced by progressive mixing (2 and 3).

that are all parallel and cut the surface. The σ_t curves have not been plotted, but are parallel to the isolines. The indicated system will remain stationary if sinking of surface water takes place between the lines a and b and if the sinking water remains on the same σ_t surface. It will also remain stationary if mixing takes place along or across σ_t surfaces. These processes will lead to the formation of a water mass that between the curves a and b always shows the same temperature-salinity relation—namely, the relation that is found along the sea surface. Iselin showed that the horizontal T-S curve along the middle part of the North Atlantic Ocean is very similar to the vertical T-S curve that is characteristic between temperatures of 20° and 8° over large areas of the North Atlantic Ocean, and he suggested that processes of sinking and lateral mixing are mainly responsible for the formation of that water. Extensive use of this concept will be made in the chapter dealing with the water masses and currents of the oceans.

However, a similar T-S relationship can be established by different processes, as illustrated in the lower part of fig. 36. It is here assumed that two water types, a and b, are formed at the surface and sink along their characteristic σ_t-surfaces. It is furthermore assumed that at subsurface depths mixing takes place between these two water types, whereas near the surface external processes influence the distribution of temperature and salinity so that there the different curves cross each other. In these circumstances one obtains a T-S relation at subsurface depths that

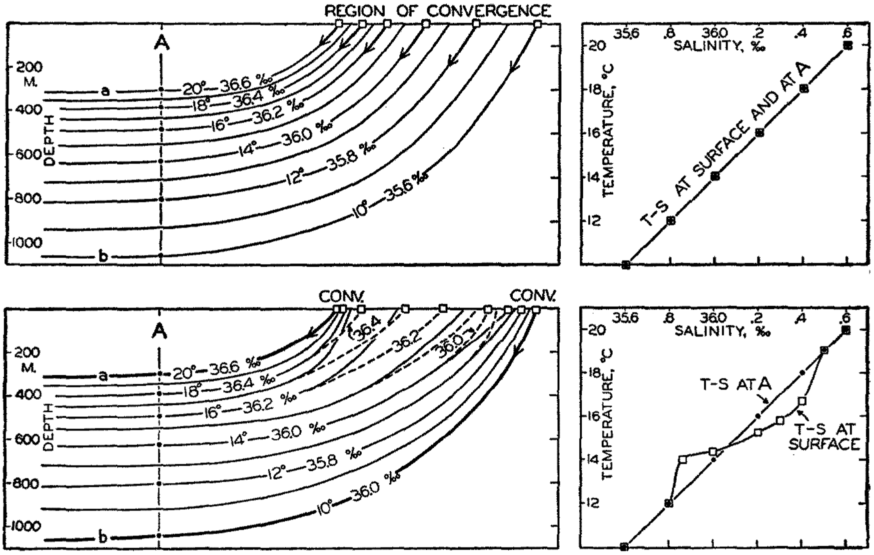


Fig. 36. *Upper:* Schematic representation of the formation of a water mass by sinking along σ_t-surfaces (which coincide with the parallel temperature and salinity surfaces) in a region of convergence. The diagram to the right demonstrates that the vertical T-S relation of the water mass agrees with the horizontal T-S relation at the surface in the region of convergence. *Lower:* Schematic representation of the formation of a water mass by sinking of water types at two convergences and by subsequent mixing. The diagram to the right illustrates that in this case the vertical T-S relation of the water mass need not agree with the horizontal T-S relation at the surface between the convergences.

is similar to the one found in the previous example, but along the sea surface an entirely different T-S relation may exist. The processes of mixing, in this case, must take place across σ_t surfaces in order to establish the T-S relation, but at subsurface depths mixing along σ_t surfaces is not excluded. At present it is impossible to decide what processes are of the greater importance.

The point to bear in mind is that the waters of the oceans all attained their original characteristics when the water was in contact with the atmosphere or subject to heating by absorption of radiation in the surface layers, and that in course of time these characteristics may become greatly changed by mixing. This mixing can either be lateral—that is, take place along σ_t surfaces, or it can be vertical—that is, crossing σ_t surfaces.

An example of lateral mixing between water masses is found off the coast of California (Sverdrup and Fleming, 1941), where the water which flows north close to the coast has a *T-S* relation that differs greatly from that of the water flowing south at some distance from the coast (fig. 199, p. 713). Between these two water masses are found waters of an intermediate character which could not possibly have been formed by vertical mixing and which must have been formed by lateral mixing, probably along σ_t surfaces. An example of modification of a water mass by vertical mixing is found in the South Atlantic, where the Antarctic intermediate water flows north. This water, near its origin, is characterized by a low salinity minimum, but the greater the distance from the Antarctic Convergence the less pronounced is this minimum (fig. 210, p. 748). The change probably cannot be accounted for by lateral mixing, but Defant (1936) has shown that it can be fully explained as a result of vertical mixing.

Wüst (1935) has introduced a different method for the study of the spreading out and mixing of water types, the "Kernschicht-methode," which can be translated "the core method." By the "core" of a layer of water is understood that part of the layer within which temperature or salinity, or both, reach extreme values. Thus, in the Atlantic Ocean, the water that flows out from the Mediterranean has a very high salinity and can be traced over large portions of the Atlantic Ocean by means of a secondary salinity maximum which decreases in intensity with increasing distance from the Strait of Gibraltar. The layer of salinity maximum is considered as the core of the layer in which the Mediterranean water spreads, and the decrease of the salinity within the core is explained as the result of processes of mixing. In this case a certain water type, the Mediterranean water, enters the Atlantic Ocean and loses its characteristic values, owing to the mixing, but can be traced over long distances. The spreading of the water can also be described by means of a *T-S* curve, one end point of which represents the temperature and salinity at the source region and the other end point of which represents the temperature and salinity in the region where the last trace of this particular water disappears. Having defined such a *T-S* curve, one can directly read off from the curve the percentage amount of the original water type that is found in any locality. The core method has proved very successful in the Atlantic Ocean and is particularly applicable in cases in which a well-defined water type spreads out from a source region.

Basins

Oceanographically a basin is defined as a depression that is filled with sea water and that is partially separated by land or submarine barriers from the open ocean, with which horizontal communication is restricted to depths less than the greatest depths in the basin. The maximum depth of an entrance from the open ocean to a basin is called the *threshold depth*, or the *sill depth*, of that entrance. By "entrance" is understood a depression in a barrier which limits the basin, and it is immaterial whether or not any part of the barrier rises above sea level. The water in the basin is in more or less restricted horizontal communication with the adjacent sea at all levels above that of the lowest sill depth, but below the sill depth renewal of the water in the basin can take place by vertical motion only. It is therefore characteristic of all basins that below the sill depth the water is nearly uniform and approximately of the same character as the water at the sill depth. Other characteristics of the water below sill depth, which will be called the basin water, depend greatly upon the type of exchange of water with the open ocean.

Basins with Outflow Across the Sill. In nearly closed basins in the semiarid regions of lower latitudes, evaporation greatly exceeds precipitation and run-off, and the salinity of the surface water is increased above that of the adjacent open ocean. The evaporation is at a maximum in winter when the surface temperature is simultaneously lowered under the influence of cold continental winds. In winter the surface density is therefore increased so much that vertical convection currents are developed which, in some years when extreme conditions exist, may reach to the greatest depth and bring about renewal of the bottom water. The basin water which is formed in this manner, owing to its very high salinity, will be of greater density than the water at the same depth outside the sill, and must therefore flow out over the sill, following the bottom slope. At some higher level the oceanic water must flow into the basin, partly to compensate for the outflow and partly to compensate for the excess of evaporation over precipitation and run-off. The Mediterranean Sea, the Red Sea, and the inner part of the Gulf of California represent examples of such basins.

In basins of this character the basin water is always characterized by high salinity and generally by high oxygen content. The amounts of inflow and outflow depend upon the difference between evaporation and precipitation and run-off, and the volumes of in- and outflowing water are many times greater than the excess of evaporation. Under stationary conditions the total amount of water which in a given time, T_i, flows into a region must equal the sum of the outflow, T_u, and the difference, D, between evaporation over precipitation and run-off in the same time: $T_i = T_u + D$. Simultaneously, the amounts of salt carried by the in- and outflowing currents must be equal. In the first approximation (p. 426), $T_i\bar{S}_i = T_u\bar{S}_u$, where $\bar{S}_i$ is the average salinity of the inflowing water and $\bar{S}_u$ is the average salinity of the outflowing water. From these relations one obtains

$$T_i = D \frac{\bar{S}_u}{\bar{S}_u - \bar{S}_i}, \qquad T_u = D \frac{\bar{S}_i}{\bar{S}_u - \bar{S}_i}. \tag{IV, 6}$$

In basins of this character the inflowing water, which comes from the adjacent open sea, has a relatively high salinity, and therefore the difference, $S_u - S_i$, is small. Consequently, the volumes of in- and outflowing water must be great compared to the excess of evaporation over precipitation.

The above considerations are valid only if the entrance is sufficiently wide or deep to permit both inflow and outflow. The Gulf of Kara-Bugaz, on the Caspian Sea, represents an example of a basin which is in such restricted communication with a larger body of water that outflow is

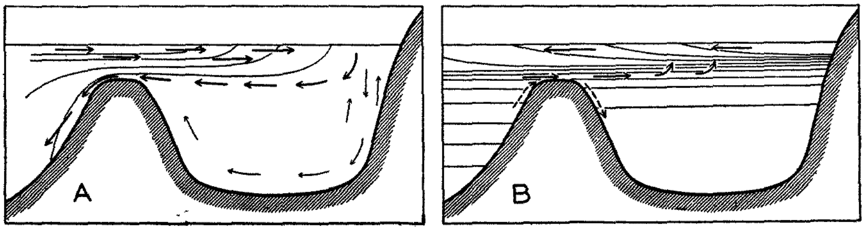


Fig. 37. (A) Basin with local formation of basin water and outflow across the sill. (B) Basin with surface outflow of water of low density and occasional renewal of the basin water by inflow of dense water across the sill.

practically impossible. This gulf is separated from the Caspian Sea by a 60-mile-long bar and has a shallow entrance which is only a few hundred meters wide. The outflow of the saline deep water is so greatly impeded that, owing to excess evaporation, the salinity of that water in 1902 was 164 $^0/_{00}$ as compared with 12.7 $^0/_{00}$ for the Caspian Sea as a whole.

Conditions are encountered which vary from this extreme case to cases in which the excess evaporation for the entire year is zero, but in which seasonal changes may be large enough to permit occasional development of vertical convection currents reaching to the bottom. The essential features to be emphasized is that in basins of this type renewal of the basin water takes place by vertical convection currents which develop in the basin itself and may reach from surface to bottom. Therefore, the water at and below the sill depth has a higher density than the water at sill depth outside the basin, and is not stagnant.

Basins with Inflow Across the Sill. In the nearly closed basins in higher latitudes, precipitation and run-off exceed evaporation. In such basins a surface layer of low salinity and correspondingly low density is developed. Because of the excess of precipitation and run-off there must be a surface outflow of relatively fresh water, and in order to maintain the salt balance there must be an inflow of more saline water.

The exchange of water with the outside sea is small because the difference, $\bar{S}_u - \bar{S}_i$, is great. If the difference is so great that the ratio $\bar{S}_u/\bar{S}_i$ is small compared to unity, the relations that are represented by equation (IV, 6) are reduced to

$$T_i = D\frac{\bar{S}_u}{\bar{S}_i}, \qquad T_u = D\left(1 + \frac{\bar{S}_u}{\bar{S}_i}\right), \tag{IV, 7}$$

where D now means the excess of precipitation and run-off over evaporation. In these circumstances the inflow is only a small fraction of this excess, and the outflow practically equals the excess.

In basins of this type, stagnant water is often found because renewal of the basin water takes place only if the inflowing water is of greater density than the basin water. Outside the sill the density of the water generally increases much more rapidly with depth than it does inside the sill. Renewal of the basin water takes place if the outside water masses are raised so much that the water which flows in across the sill is of such high density that it sinks toward the bottom of the basin. Fig. 37 shows schematically the character of the exchange with the outside and the renewal of the basin water in the two types of basins.

The rapidity of the renewal of the deep water in the basin depends upon the steepness of the vertical density gradient at the sill depth. If this gradient is steep, an occasional large disturbance fills the basin below sill depth with water of great density, and subsequent disturbances must be as great or greater in order to bring about renewal of the basin water. In extreme cases renewal takes place only by major catastrophes. In the intervals between such catastrophes the basin water may become stagnant, because in the upper layers of stable stratification vertical mixing is insignificant. Some mixing takes place, however, which, between major disturbances, reduces the density of the basin water so much that complete renewal can take place when a new catastrophe occurs.

On the other hand, if the density gradient at the sill depth is small, the outside deep water is brought over the sill by any minor disturbance, and stagnation is prevented by intermittent intrusion of outside deep water, and also by vertical mixing, which is more effective owing to the small density gradient.

The water which sinks at sill depth is heated adiabatically, and the basin water is therefore of nearly constant potential temperature. The effective sill depth—that is, the depth at which the potential temperature in the outside water equals that in the basin—is lower on an average than the actual depth of the sill (table 87, p. 738), and the smaller the density gradient in the outside water, the greater is the difference between the effective and the actual sill depth. Great density gradients, if present, are always found near the surface, and a basin with inflow at sill depth is therefore likely to have stagnant water if the sill is shallow. The Black Sea, the Baltic, and numerous Norwegian fjords are examples of basins of this type (Fleming and Revelle, 1939, Ström, 1936).

The sill depth has bearing also on the direction of flow across the sill, and this direction does not therefore depend exclusively on an excess or a deficit of evaporation, which was used to facilitate the discussion. At small sill depths an excess or deficit of evaporation determines the character of the exchange, but at great sill depths inflow across the sill develops in most instances. Oceanic water flows freely in and out of the basin at some distance above the sill depth, but at the sill depth the average flow is directed, as a rule, into the basin, because the density of the basin water remains lower than that of the outside water, owing to more effective vertical mixing in a restricted region. The main in- and outflow takes place, however, at lesser depths, the water often flowing in through one entrance and out through another. The basins of the American Mediterranean Sea serve as examples (p. 639).

In large basins in high latitudes, such as the Norwegian Sea and Baffin Bay, deep water is formed locally, owing to freezing or excessive cooling of high-salinity water, although precipitation exceeds evaporation. In such basins, which might be listed as a third type, stagnating water is not found.

Bibliography

Ångström, Anders. 1920. Applications of heat radiation measurements to the problems of the evaporation from lakes and the heat convection at their surfaces. Geografiska Annaler, H. 3, 16 pp. Stockholm. 1920.

Birge, E. A., and C. Juday. 1929. Transmission of solar radiation by the waters of inland lakes. Wisconsin Acad. Sci., Arts, and Letters, Trans., v. 24, p. 509–80, 1929.

Bjerknes, V., J. Bjerknes, H. Solberg, T. Bergeron. 1932. Physikalische Hydrodynamik. Berlin. Julius Springer, 797 pp., 1932.

Böhnecke, G. 1938. Temperatur, Salzgehalt und Dichte an der Oberfläche des Atlantischen Ozeans. Deutsche Atlantische Exped. *Meteor* 1925–1927, Wiss. Erg., Bd. 5, 2 Lief., 62 pp., 1938.

Bowen, I. S. 1926. The ratio of heat losses by conduction and by evaporation from any water surface. Phys. Rev., v. 27, p. 779–787, 1926.

Defant, Albert. 1928. Die systematische Erforschung des Weltmeeres. Gesellsch. f. Erdkunde, Zeitschrift, Jubiläums Sonderband, p. 459–505, Berlin, 1928.

——— 1932. Die Gezeiten und inneren Gezeitenwellen des Atlantischen Ozeans. Deutsche Atlantische Exped. *Meteor* 1925–1927, Wiss. Erg., Bd. 7, 1 Teil, p. 1–318, 1932.

——— 1936. Ausbreitungs- und Vermischungsvorgänge im Antarktischen Bodenstrom und im Subantarktischen Zwischenwasser. Deutsches Atlantische Exped. *Meteor* 1925–1927, Wiss. Erg., Bd. 6, 2 Teil, Lief. 2, p. 55–96, 1936.

Fjeldstad, J. 1933. Wärmeleitung im Meere. Geofysiske Publikasjoner, v. 10, no. 7, 20 pp., 1933. Oslo.

Fleming, R. H., and Roger Revelle. 1939. Physical processes in the ocean. P. 48–141 in: Recent Marine Sediments, A symposium. Tulsa. Amer. Assn. Petrol. Geol. 1939.

Gaarder, T., and R. Spärck. 1932. Hydrographisch-biochemische Untersuchungen in norwegischen Austern-Pollen. Bergens Museums Aarbok 1932, Naturvidenskapeligrekke, Nr. 1, p. 5–144, 1932.

Helland-Hansen, B. 1916. Nogen hydrografiske metoder. Skand. Naturforsker möte, Kristiania (Oslo). 1916.

——— 1930. Physical oceanography and meteorology. *Michael Sars* North Atlantic Deep-Sea Exped., 1910, Rept. Sci. Results, v. 1, art. 2, 217 pp., 1930.

Iselin, C. O'D. 1939. The influence of vertical and lateral turbulence on the characteristics of the waters at mid-depths. Nat. Research Council, Amer. Geophys. Un., Trans., p. 414–17, 1939.

Jacobs, W. C. 1942. On the energy exchange between sea and atmosphere. Jour. Marine Research, v. 5, 37–66, 1942.

Kimball, Herbert H. 1928. Amount of solar radiation that reaches the surface of the earth on the land and on the sea, and methods by which it is measured. Monthly Wea. Rev., v. 56, p. 393–99, 1928.

Koenuma, K. 1939. On the hydrography of southwestern part of the North Pacific and the Kuroshio. Imper. Marine Observ., Memoirs, v. 7, p. 41–114, 1939. Kobe.

Krümmel, Otto. 1907. Handbuch der Ozeanographie. Bd. 1. Stuttgart. J. Engelhorn, 526 pp., 1907.

Kuhlbrodt, Erich, and Joseph Reger. 1938. Die meteorologischen Beobachtungen. Methoden, Beobachtungsmaterial und Ergebnisse. Deutsche Atlantische Exped. *Meteor* 1925–1927, Wiss. Erg., Bd. 14, p. 215–345, 1938.

McEwen, George F. 1938. Some energy relations between the sea surface and the atmosphere. Jour. Marine Research, v. 1, p. 217–38, 1938.

Millar, F. Graham. 1937. Evaporation from free water surfaces. Canadian Meteorol. Memoirs, v. 1, p. 41–65, 1937. Ottawa.

Montgomery, R. B. 1940. Observations of vertical humidity distribution above the ocean surface and their relation to evaporation. Papers in Physical Oceanogr. and Meteorol., v. 7, no. 4, 30 pp., 1940. Cambridge and Woods Hole.

Mosby, Håkon. 1936. Verdunstung und Strahlung auf dem Meere. Ann. d. Hydrogr. u. Mar. Meteor., Bd. 64, p. 281–86, 1936.

Netherlands Meteorological Institute. Atlas of oceanographic and meteorological observations.

Neumann, Gerhard. 1940. Die ozeanographischen Verhältnisse an der Meeresoberfläche im Golfstromsektor nördlich und nordwestlich der Azoren. Ann. d. Hydrogr. u. Mar. Meteor., Beiheft zum Juniheft, 1. Lief., 87 pp., 1940.

Powell, Wilson M., and G. L. Clarke. 1936. The reflection and absorption of daylight at the surface of the ocean. Optical Soc. Amer., Jour., v. 26, p. 111–120, 1936.

Schmidt, Wilhelm. 1915. Strahlung und Verdunstung an freien Wasserflächen; ein Beitrag zum Wärmehaushalt des Weltmeers und zum Wasserhaushalt der Erde. Ann. d. Hydrogr. u. Mar. Meteor., Bd. 43, p. 111–24, 1915.

Schott, G. 1935. Geographie des Indischen und Stillen Ozeans. Hamburg, C. Boysen, 413 pp., 1935.

Skogsberg, Tage. 1936. Hydrography of Monterey Bay, California. Thermal conditions, 1929–1933. Amer. Phil. Soc., Trans., N.S., v. 29, p. 1–152, 1936.

Ström, K. M. 1936. Land locked waters. Hydrography and bottom deposits in badly ventilated Norwegian fjords with remarks upon sedimentation under anaerobic conditions. Norske Vidensk. Akad. I. Mat. Naturv. Klasse, No. 7, 85 pp., 1936.

Sverdrup, H. U. 1929. The waters on the North-Siberian Shelf. Norwegian North Polar Exped. with the *Maud* 1918–1925, Sci. Results, v. 4, no. 2, 131 + 75 pp., 1929.

———— 1937. On the evaporation from the oceans. Jour. Marine Research, v. 1, p. 3–14, 1937.

———— 1940. On the annual and diurnal variation of the evaporation from the oceans. Jour. Marine Research, v. 3, p. 93–104, 1940.

Sverdrup, H. U., and R. H. Fleming. 1941. The waters off the coast of southern California, March to July, 1937. Scripps Inst. Oceanogr., Univ. Calif., Bull., v. 4, p. 261–378, 1941.

Taylor, G. I. 1919. Tidal friction in the Irish Sea. Roy. Soc., Philos. Trans., Series A, v. 220, p. 1 ff., 1919. London.

U. S. Weather Bureau. 1938. Atlas of climatic charts of the oceans. W. B. no. 1247, 130 pp., 1938. Washington, D. C.

Vercelli, Francesco. 1937. La propagazione delle radiazioni solar nelle acque marine. La Laguna di Venezia, v. 1, parte II, tomo V, 102 pp., 1937. Venice.

von Hann, Julius. 1915. Lehrbuch der Meteorologie. 3rd ed., Leipzig. Tauchnitz. 847 pp., 1915.

Wegemann, G. 1920. Der tägliche Gang der Temperatur der Meere und seine monatliche Veränderlichkeit. Wiss. Meeresuntersuch., N.F., Bd. 19, Abt. Kiel, Abh. 2. Kiel. 1920.

Wüst, Georg. 1935. Die Stratosphäre. Deutsche Atlantische Exped. *Meteor* 1925–1927, Wiss. Erg., Bd. 6, 1 Teil, 2. Lief, 288 pp., 1935.

———— 1936. Oberflächensalzgehalt, Verdunstung und Niederschlag auf dem Weltmeere. Festschrift Norbert Krebs (Länderkundliche Forschung), p. 347–59. Stuttgart. 1936.