GEOFYSISKE PUBLIKASJONER
GEOPHYSICA NORVEGICA
VOL. XXII NO. 7

A STUDY OF EVAPORATION AND HEAT EXCHANGE BETWEEN THE SEA SURFACE AND THE ATMOSPHERE

BY GUNNVALD BØYUM

FREMLAGT I VITENSKAPS-AKADEMIETS MØTE DEN 17DE FEBR. 1961 AV FJELDSTAD

Summary. By means of observations from Ocean Weather Station M (66°N 2°E) collected during a period of ten years, an expression for the heating of the air is obtained, relating the time change of the air temperature to the temperature difference between sea and air.

A similar expression has been found for the water vapour supply to the air.

The equations are based on the assumption that in the lower layer of air moving steadily across equidistant surface isotherms a stationary vertical distribution of temperature and water vapour pressure will finally be established.

The equations are discussed and the time needed to establish the "equilibrium distribution" is estimated.

1. **Introduction.** By means of observations from the Bouvet Island area and from the North Atlantic, MOSBY [4, 6] has shown that the heating of the lower layer of the atmosphere may be expressed as follows:

$$\frac{dt_a}{d\tau} = k(t_s - t_a) - c \tag{1}$$

where k and c are positive constants (see list of symbols). Observations from station M have been used in the following to derive similar equations, as well as the corresponding equations of evaporation. For the derivation of the equations knowledge on the main features of the sea-surface temperature distribution is needed.

Representative results may be expected when examining air-masses which have been moving for a relatively long time over ocean areas, where the surface isotherms are equidistant. In the Norwegian Sea such conditions are more prevalent in the summer than in the other seasons. The surface temperature distribution is also better known for the summer season. The investigation is therefore restricted to the months of June, July, August, and September, and are based upon data from the decennial period 1949—1958.

Most of the observations were made on board the weather-ships "Polarfront I" and "Polarfront II". In July, August, and September 1954, 1956, and 1958 the observations were carried out on board the Dutch weather-ships "Cumulus" and "Cirrus".

About 9.000 sets of observations have been treated altogether, each set including speed and direction of wind, dry and wet bulb air temperature and sea-surface temperature.

Speed and direction of wind were measured by fixed anemometers at an altitude of 10 metres above sea-level, while air temperature was measured by means of an Assman aspiration psychrometer at an altitude of 6—7 metres above sea-level.

For the present study it is assumed that no serious error will be introduced by considering the wind observations being valid at the altitude of the temperature measurements.

The sea-surface temperatures were determined by the bucket method.

2. Sea-surface temperature. Monthly charts of sea-surface temperatures in the Norwegian Sea have been published by KRAUSS [3]. They are based upon most of the observations collected before 1953. The temperature distribution near the Norwegian coast was treated by FROGNER [2], and maps covering the south-western part of the Norwegian Sea were published by EGGVIN and SPINNANGR [1].

Dr. EGGVIN has also kindly put at our dispcsal ten sea-surface temperature maps, based upon data collected on a number of summer expeditions in the period from 1949 to 1958.

As would be expected, these maps show numerous differences. Their main features

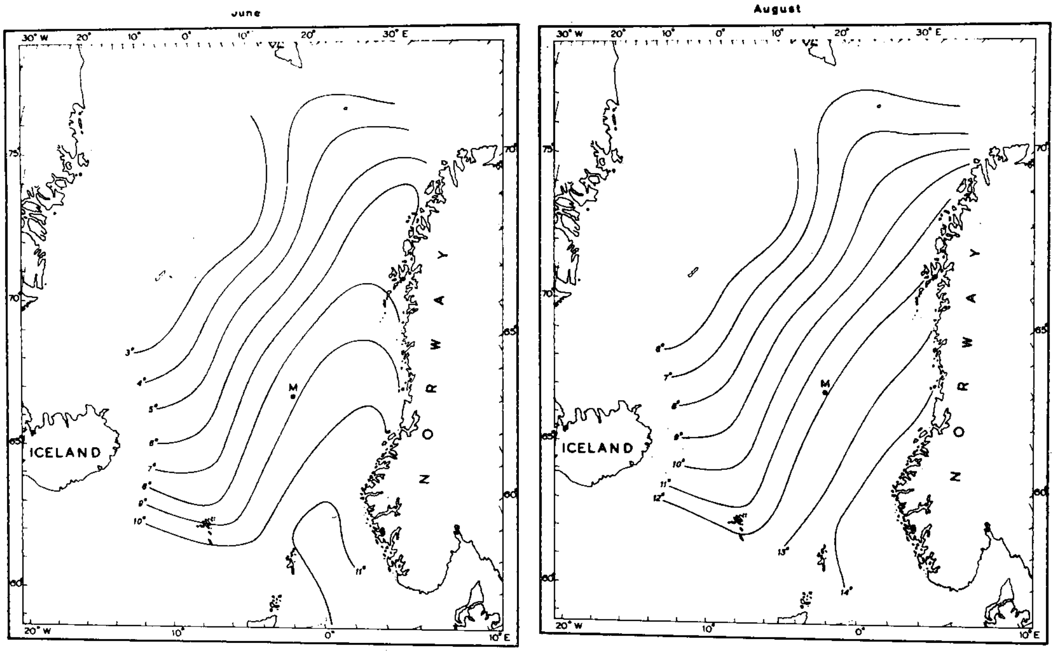

Fig. 1. Approximate sea-surface isotherms for June and August.

however, may be fairly well represented by the smoothed isotherms for June and August drawn in Fig. 1.

Corresponding isotherms for July and September are so similar to those for August that they have not been reproduced.

It is believed that these maps do represent the main features of the surface temperature distribution, and that they may therefore serve as a sufficient basis for the present investigation.

3. Heat transport.

a). *Temperature difference.* In order to show the influence of the wind upon the temperature difference between sea and air, $\Delta t = t_s - t_a$ (see list of symbols), the observations from station M were treated as follows:

For each month the differences Δt were arranged according to wind force in groups 1—5, 5—10, 10—15 .... knots, and within each of these groups they were again arranged according to wind direction in intervals of 20°. Within each interval the mean values of Δt were calculated. These mean values were plotted in a vector diagram, the distance from the centre representing the wind force. The resulting isolines are shown in Fig. 2. Mean direction of sea-surface isotherms is indicated for each month.

The isolines show an irregular form, but certain main features reappear from month to month. When the air moves from colder towards warmer sea, i.e. by northwesterly winds, the air temperature is always lower than the sea-surface temperature. It is seen that $t_s - t_a$ increases with increasing wind force, and the rate of increase depends on the wind direction in relation to the direction of the surface isotherms.

In the diagrams for July and August the direction of the surface isotherms is mainly found to fall between the isolines $\Delta t = 0.5$ and $\Delta t = 1$. In other words, when the wind blows along the isotherms, the temperature difference is positive and approximately independent of the wind force. This conclusion is less obvious from the diagram for June, and still less so from the diagram for September.

If the air moves from warmer towards colder sea, Δt decreases with increasing wind force; but it remains positive until the wind reaches a certain speed that depends on the wind direction.

Reasons for the curved shape of the isolines $\Delta t \geqq 1$ will not be discussed here.

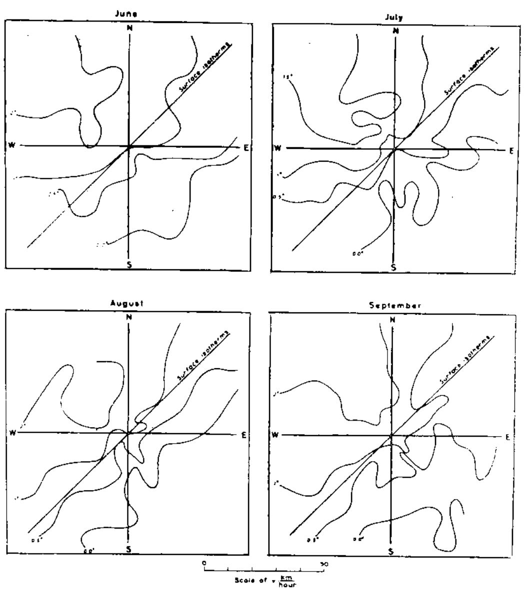

Fig. 2. The temperature difference $t_s - t_a$ in relation to wind force and wind direction.

b). *The irregularities of the isolines.* The air temperature must depend upon the thermal influence to which the air-masses have been exposed during the latest hours or days. The influence again must depend on the motion of the air-masses towards warmer or colder areas of the sea-surface. The rate of movement of the air from one isotherm to another must be expected to play an important part in determining the air temperature.

Air moving over the sea often follows curved trajectories, and the thermal influence from beneath may then be rather complex. This will be reflected in the group mean values of Δt, especially in those based on few observations.

Detailed charts of surface isotherms show that the sea temperature varies from year to year. Air reaching station *M* from a certain direction may thus have been exposed to a different thermal influence each year.

The distribution of single values of Δt appears to be rather asymmetrical. Observations from certain years may dominate one group mean value of Δt, while another group may contain mainly observations from other years within the same decennial period.

Some of the irregularities of the isolines in Fig. 2 may be explained from reasons as mentioned above.

c). *Heating of the air.* In the preceding section it was stated that the difference $t_s - t_a$ depends on the wind force and the wind direction relative to the surface isotherms.

This empirical fact indicates that the heat exchange between the sea and the air is a function of the same quantities. A relation between the heating of the air, the wind force, and the wind direction may be found as follows.

An air particle moving across the sea will have its temperature t_a changed. Generally the temperature of the sea-surface in contact with the air t_s varies along the trajectory of the air particle and the individual change in $\Delta t = t_s - t_a$ is

$$\frac{d(\Delta t)}{d\tau} = \frac{dt_s}{d\tau} - \frac{dt_a}{d\tau} \tag{2}$$

$\frac{dt_s}{d\tau}$ is given by

$$\frac{dt_s}{d\tau} = \frac{\partial t_s}{\partial \tau} + \boldsymbol{V} \cdot \nabla t_s$$

where $\boldsymbol{V}$ denotes the velocity of the air particle which is assumed to be horizontal.

Consider a two-dimensional system in which the velocity is directed along an axis r the direction of which is arbitrary. If we assume $\frac{\partial t_s}{\partial \tau}$ to be small compared to $\boldsymbol{V} \cdot \nabla t_s$ then equation (2) can be written

$$\frac{d(\Delta t)}{d\tau} = v_r \frac{\partial t_s}{\partial r} - \frac{dt_a}{d\tau} \tag{3}$$

In the case of a steady wind blowing across equidistant surface isotherms it is believed that the vertical heat transport will finally be constant and that a stationary vertical temperature profile will be established. When this state of equilibrium is reached the difference $t_s - t_a$ is constant and equation (3) reduces to

$$\frac{dt_a}{d\tau} = v_r \frac{\partial t_s}{\partial r} \tag{4}$$

d). Heat equations. Equation (4) will now be applied to the data from station M and the heating will be related to the temperature difference Δt.

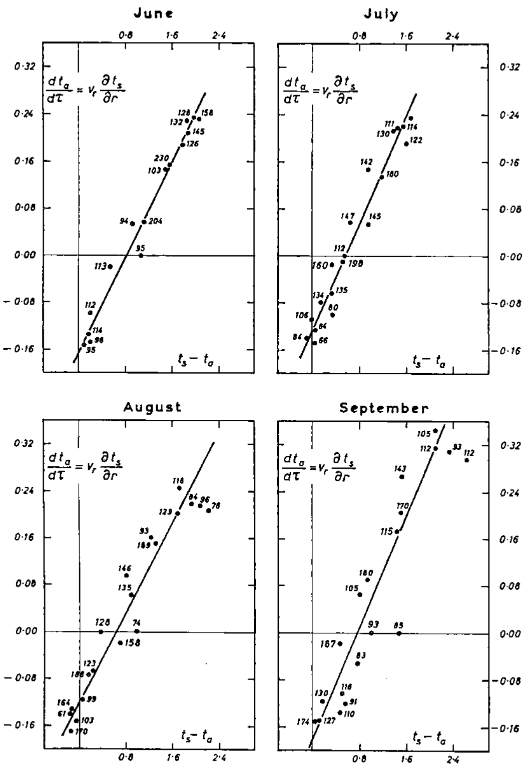

Fig. 3. The relation between the heating $\frac{dt^a}{d\tau} = v_r \frac{\partial t_s}{\partial r}$ and the temperature difference $t^s - t^a$.

The observations are grouped according to the wind direction. Within each group the average wind force v_r in km per hour, the average temperature difference Δt in centigrades, the mean variation of the sea-surface temperature $\frac{\partial t_s}{\partial r}$ in centigrades per km, and the heating $\frac{dt_a}{d\tau}$ are calculated. The resulting values are given in Tables 3, 4, 5, and 6.

The tongue-like distribution of the sea-surface temperature (see Fig. 1) off the Norwegian coast in the month of June implies that air masses moving from easterly directions have first been exposed to a warmer and then to a cooler sea-surface. Because of this complex thermal influence, it must be assumed that in those cases the air will be far from the state of equilibrium when reaching station M. The groups 05—06, 07—08, 09—10 and 11—12 are therefore omitted. The values of $\frac{dt_a}{d\tau}$ are plotted against the corresponding values of the temperature difference in Fig. 3.

As a first approximation a linear relationship is demonstrated in all diagrams, and the values of k and c (see equation 1) are given in Table 1.

From equation (1)

$$\frac{dt_a}{d\tau} = v_r \frac{\partial t_s}{\partial r} = k \, \Delta \, t - c$$

Table 1.

	June	July	August	September	Mean
k(hours^{-1})	0.20	0.21	0.19	0.24	0.21
c(°C hours^{-1})	0.16	0.13	0.12	0.18	0.15

it is seen that in calm weather ($v_r = 0$), or when the air moves along the sea-surface isotherms $\left(\frac{\partial t_s}{\partial r} = 0\right)$ the heating $\frac{dt_a}{d\tau}$ vanishes. In this case $\Delta t = \frac{a}{k}$. When substituting from Table 1, $k = 0.21$ and $c = 0.15$ we find that there will be neither heating nor cooling of the lower layer of the air, when the air temperature is about 0.7°C lower than the sea-surface temperature.

e). *Tmperature equilibrium.* When deriving the heat equations it was assumed that, when the air moves across the sea-surface, a stationary vertical temperature profile would finally be established. We will now try to obtain a rough estimate of the time needed to establish this state of equilibrium.

If the relation between $\frac{dt_a}{d\tau}$ and Δt before the state of equilibrium had been known, we might have calculated the time by integrating equation (3)

$$\frac{d(\Delta t)}{d\tau} = v_r \frac{\partial t_s}{\partial r} - \frac{dt_a}{d\tau}$$

assuming $v_r \frac{\partial t_s}{\partial r}$ to be constant. The latter assumption is valid if the air moves with a constant velocity over equidistant isotherms.

On the assumption, however, that $\frac{d(\Delta t)}{d\tau} \ll v_r \frac{\partial t_s}{\partial r}$ and $\frac{d(\Delta t)}{d\tau} \ll \frac{dt^a}{d\tau}$ we may substitute for $\frac{dt_a}{d\tau}$ the equilibrium values given by equation (10).

Equation (3) may then be written

$$\frac{d(\Delta t)}{d\tau} + k\Delta t = v_r \frac{\partial t_s}{\partial r} + c. \tag{5}$$

This relation was obtained by A. Åmot [7].

Integration of equation (5) gives

$$\Delta t = \Delta t_0 e^{-k\tau} + \left(\frac{c}{k} + \frac{1}{k} v_r \frac{\partial t_s}{\partial r}\right)\left(1 - e^{-k\tau}\right) \tag{6}$$

from which is seen that $\Delta t = \Delta t_0$ at time $\tau = 0$. Writing $\Delta t_\infty = \frac{c}{k} + \frac{1}{k} v_r \frac{\partial t_s}{\partial r}$, equation (6) takes the form

$$\Delta t - \Delta t_{\infty} = (\Delta t_0 - \Delta t_{\infty})e^{-k\tau}$$

from which is seen that at time $\tau = \infty$, $\Delta t = \Delta t_{\infty}$. Practically Δt will reach the value Δt_{∞} after a finite time. As an example, the time needed for $(\Delta t - \Delta t_{\infty})$ to be reduced to 10 per cent of $(\Delta t_0 - \Delta t_{\infty})$ is found from

$$\tau = \frac{1}{k} \text{ lognat } 10.$$

Using for k the value 0.21 this gives

$$\tau \approx 11 \text{ hours.}$$

The distance from station M to the Norwegian coast is about 400 km. The average speed of air moving towards station M from easterly or south-easterly directions is seen from the tables to be some 25—30 km per hour. This means that such air masses will reach station M between 13 and 16 hours after passing the coast of Norway. They may, therefore, be expected to be not very far from the state of equilibrium.

Air masses reaching station M from other directions have crossed over much wider ocean areas, and must therefore be expected to be nearer still to the state of equilibrium.

4. Water vapour transport.

a). *Vapour pressure difference.* For a study of the water vapour transport, the sea-surface isotherms may be interpreted as isobars of saturation vapour pressure e_s close to the sea-surface.

The water vapour pressure differences $\Delta e = e_s - e_a$ were treated as explained above for the temperature differences Δt. The relation between Δe and the wind is shown in Fig. 4.

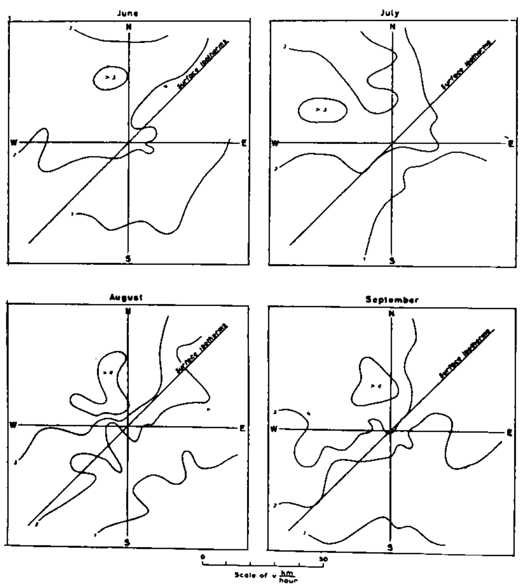

Fig 4. The water vapour pressure difference $e^s - e^a$ in relation to wind force and wind direction.

It is seen that in air moving from warmer towards colder areas of the sea, the value of Δe decreases with increasing wind force, but it is nearly always positive.

b). *Vapour supply.* Starting from the same simple condition as before, a corresponding expression for $\frac{de_a}{d\tau}$ as for $\frac{dt_a}{d\tau}$ can be derived.

$$\frac{de_a}{d\tau} = v_r \frac{\partial e_s}{\partial r} \tag{7}$$

c). *Evaporation equations.* The values of water vapour pressure have been treated

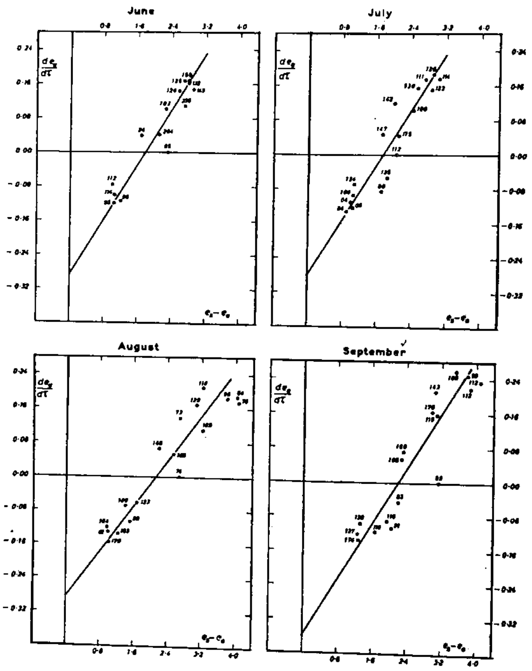

Fig. 5. The relation between the evaporation $\frac{de_a}{d\tau} = v_r \frac{\partial e_s}{\partial r}$ and the difference $e_s - e_a$.

exactly as explained above for the temperature values. The resulting values of Δe, $\frac{\partial e_s}{\partial r}$ and $\frac{de_a}{d\tau}$ for the different directions are given in Tables 7, 8, 9, and 10.

From Fig. 5 it is seen that the relation between $\frac{de_a}{d\tau}$ and $e_s - e_a$ can be represented by the approximate formula

$$\frac{de_a}{d\tau} = v_r \frac{\partial e_s}{\partial r} = b(e_s - e_a) - m \tag{8}$$

where b and m are positive constants. The values of b and m are given in Table 2.

In the case of calm weather ($v_r = 0$), or when the air moves along the sea-surface isotherms $\left(\frac{\partial e_s}{\partial r} = 0\right)$, it is seen from equation (8) that $\frac{\partial e_a}{d\tau} = 0$. We then have $\Delta e = \frac{m}{b}$. Introducing from Table 2, $b = 0.16$ and $m = 0.30$, it is seen that there will be no transport of water vapour to or from the lower layer of the air, when the vapour pressure is about 1.9 mb lower than the vapour pressure close to the surface.

Table 2.

	June	July	August	September	Mean
b(hours^{-1})	0.17	0.16	0.14	0.15	0.16
m(mb.hours^{-1})	0.27	0.28	0.29	0.36	0.30

d). *Vapour pressure equilibrium.* Making the same simplifications as in section 3*c* we find

$$\Delta e = \Delta e_0 e^{-b\tau} + \left(\frac{m}{b} + \frac{1}{b} v_r \frac{\partial e_s}{\partial r}\right)(1 - e^{-k\tau}) \tag{9}$$

where Δe_0 corresponds to $\tau = 0$.

Writing

$$\Delta e_\infty = \frac{m}{b} + \frac{1}{b} v_r \frac{\partial e_s}{\partial r}$$

equation (9) takes the form

$$(\Delta e - \Delta e_{\infty}) = (\Delta e_0 - \Delta e_{\infty}) e^{-k\tau}.$$

The time needed for $(\Delta e - \Delta e_{\infty})$ to be reduced to 10 per cent of $(\Delta e_0 - \Delta e_{\infty})$ is given by

$$\tau = \frac{1}{b} \text{ lognat } 10.$$

Introducing $b = 0.16$, we find

$$\tau \approx 14 \text{ hours}.$$

5. Final remarks. When studying the turbulent processes in the lower layer of atmosphere, several authors have derived formulae for the heat transfer and the evaporation as function of $(t_s - t_a)$ and $(e_s - e_a)$ and wind. Most of these formula e are of the form

(10) $$Q_h = K_1(t_s - t_a)v_a$$

(11) $$Q_e = K_2(e_s - e_a)v_a$$

where K_1 and K_2 are more or less complicated expressions. According to the results obtained in the preceding (section 3*d* and 4*c*) there should be no exchange of heat when $t_s - t_a \approx 0.7°C$ and no exchange of water vapour when $e_s - e_a \approx 1.9$ mb. If this is true, it seems that the introduction of observed values of $t_s - t_a$ and $e_s - e_a$ into such formulae as (10) and (11) must give misleading results. It is, however, difficult to give a satisfactory physical explanation of the above obtained "equilibrium values" of $t_s - t_a$ and $e_s - e_a$. Such an explanation may perhaps throw some light on the complex mechanism of the energy transfer between sea and atmosphere.

Acknowledgement. The writer owes his interest in the problems presented i this paper to Professor MOSBY, whose advise and encouragement have been indispensable.

Table 3. *June.*

Wind direction	Number of obs.	v_s $\frac{\text{km}}{\text{hour}}$	$(t_s - t_a)$ °C	$\frac{\partial t_s}{\partial r}$ °C/km	$\frac{dt_a}{d\tau}$ °C/hour
Calm	95	0.0	1.08	—	0.000
01—02	230	28.4	1.55	0.0055	0.156
03—04	204	29.5	1.10	0.002	0.059
05—06	142	24,8	0.92	—	—
07—08	96	21.2	1.00	—	—
09—10	67	19.1	0.67	—	—
11—12	71	22.7	0.00	—	—
13—14	95	23.4	0.11	—0.0065	—0.152
15—16	98	24.1	0.20	—0.0061	—0.147
17—18	114	24.1	0.16	—0.0055	—0.133
19—20	112	24.8	0.19	—0.0039	—0.097
21—22	113	22.3	0.53	—0.0009	—0.020
23—24	94	21.6	0.84	0.0025	0.054
25—26	103	25.6	1.48	0.0057	0.146
27—28	126	24.8	1.78	0.0076	0.188
29—30	158	24.8	2.05	0.0094	0.233
31—32	128	25.2	1.98	0.0094	0.237
33—34	132	27.0	1.83	0.0085	0.230
35—36	145	27.0	1.87	0.0077	0.208

Table 4. *July.*

Wind direction	Number of obs.	v_r $\frac{\text{km}}{\text{hour}}$	$t_s - t_a$ °C	$\frac{\partial t_s}{\partial r}$ °C/km	$\frac{dt_a}{d\tau}$ °C/hour
Calm	112	0.0	0.55	—	0.000
01—02	180	23.9	1.18	0.0055	0.134
03—04	175	24.6	0.92	0.0022	0.054
05—06	198	23.5	0.52	—0.0004	—0.009
07—08	135	23.5	0.32	—0.0027	—0.063
09—10	80	24.7	0.35	—0.0040	—0.099
11—12	66	24.0	—0.09	—0.0057	—0.140
13—14	84	23.5	0.01	—0.0063	—0.148
15—16	84	22.0	0.02	—0.0057	—0.125
17—18	106	23.5	—0.01	—0.0046	—0.108
19—20	134	26.0	0.14	—0.0031	—0.081
21—22	160	25.8	0.30	—0.0006	—0.015
23—24	147	24.1	0.64	0.0024	0.058
25—26	142	26.8	0.95	0.0056	0.146
27—28	130	26.1	1.39	0.0082	0.214
29—30	120	24.1	1.67	0.0097	0.234
31—32	111	22.5	1.44	0.0097	0.219
33—34	114	24.6	1.53	0.0090	0.221
35—36	122	23.4	1.60	0.0075	0.102

Table 5. *August*.

Wind direction	Number of obs.	v_r $\frac{\text{km}}{\text{hour}}$	$(t_s - t_a)$ °C	$\frac{\partial t_s}{\partial r}$ °C/km	$\frac{dt_a}{d\tau}$ °C/hour
Calm	74	0.0	1.02	—	0.000
01—02	189	27.6	1.35	0.0055	0.152
03—04	185	27.9	0.89	0.0022	0.061
05—06	158	27.4	0.70	−0.0007	−0.019
07—08	123	24.6	0.23	−0.0027	−0.066
09—10	99	23.5	0.00	−0.0050	−0.118
11—12	61	22.3	−0.17	−0.0063	−0.140
13—14	103	23.7	−0.08	−0.0067	−0.145
15—16	170	27.0	−0.15	−0.0063	−0.170
17—18	164	28.0	−0.13	−0.0046	−0.128
19—20	188	27.1	0.15	−0.0028	−0.076
21—22	128	29.1	0.35	0.0000	0.000
23—24	146	29.0	0.80	0.0027	0.078
25—26	73	26.8	1.20	0.0060	0.161
27—28	118	26.8	1.76	0.0085	0.243
29—30	78	20.5	2.22	0.0100	0.205
31—32	96	21.6	2.08	0.0100	0.216
33—34	84	24.3	1.94	0.0090	0.218
35—36	129	24.9	1.69	0.0080	0.200

Table 6. *September*.

Wind direction	Number of obs.	v_r $\frac{\text{km}}{\text{hour}}$	$(t_s - t_a)$ °C	$\frac{\partial t_s}{\partial r}$ °C/km	$\frac{dt_a}{d\tau}$ °C/hour
Calm	85	0.0	1.50	—	0.000
01—02	115	29.9	1.45	0.0058	0.173
03—04	105	30.2	0.81	0.0022	0.063
05—06	93	29.9	0.98	0.0000	0.000
07—08	83	29.2	0.78	−0.0018	−0.053
09—10	116	29.5	0.51	−0.0035	−0.103
11—12	91	29.5	0.58	−0.0041	−0.121
13—14	110	27.7	0.46	−0.0048	−0.133
15—16	127	28.8	0.09	−0.0052	−0.150
17—18	174	29.9	0.04	−0.0050	−0.150
19—20	130	31.7	0.17	−0.0037	−0.117
21—22	187	33.5	0.42	−0.0006	−0.020
23—24	180	32.8	0.94	0.0027	0.089
25—26	170	33.5	1.52	0.0061	0.204
27—28	143	32.4	1.56	0.0082	0.266
29—30	105	34.6	2.11	0.0100	0.346
31—32	93	31.0	2.33	0.0100	0.310
33—34	112	29.2	2.62	0.0100	0.292
35—36	112	31.7	3.11	0.0086	0.273

Table 7. *June.*

Wind direction	$e_s - e_a$ mb	$\frac{\partial e_s}{\partial r}\frac{mb}{km}$	$\frac{de_a}{d\tau}\frac{mb}{hour}$
Calm	2.82	—	0.000
01—02	2.71	0.0039	0.111
03—04	2.10	0.0014	0.041
05—06	1.89	—	—
07—08	1.92	—	—
09—10	1.64	—	—
11—12	1.11	—	—
13—14	1.06	—0.0051	—0.120
15—16	1.20	—0.0048	—0.117
17—18	1.06	—0.0042	—0.101
19—20	1.01	—0.0031	—0.077
21—22	1.28	—0.0007	—0.016
23—24	1.66	0.0018	0.039
25—26	2.28	0.0040	0.102
27—28	2.58	0.0058	0.144
29—30	2.74	0.0067	0.166
31—32	2.70	0.0067	0.169
33—34	2.73	0.0061	0.165
35—36	2.89	0.0055	0.149

Table 8. *July.*

Wind direction	$e_s - e_a$ mb	$\frac{\partial e_s}{\partial r}\frac{mb}{km}$	$\frac{de_a}{d\tau}\frac{mb}{hour}$
Calm	2.15	—	0.000
01—02	2.44	0.0043	0.103
03—04	2.10	0.0018	0.040
05—06	1.90	—0.0003	—0.007
07—08	1.86	—0.0023	—0.054
09—10	1.74	—0.0035	—0.086
11—12	1.05	—0.0050	—0.123
13—14	0.93	—0.0055	—0.129
15—16	1.01	—0.0050	—0.110
17—18	1.08	—0.0040	—0.094
19—20	1.10	—0.0027	—0.070
21—22	1.36	—0.0004	—0.010
23—24	1.73	0.0020	0.048
25—26	2.00	0.0046	0.119
27—28	2.55	0.0065	0.156
29—30	2.89	0.0078	0.188
31—32	2.73	0.0078	0.176
33—34	3.02	0.0072	0.177
35—36	2.86	0.0060	0.152

Table 9. *August.*

Wind direction	$e_s - e_a$ mb	$\frac{\partial e_s}{\partial r}\ \frac{mb}{km}$	$\frac{de_a}{d\tau}\ \frac{mb}{hour}$
Calm	2.65	—	0.000
01—02	3.32	0.0047	0.130
03—04	2.54	0.0019	0.053
05—06	2.18	—0.0004	—0.011
07—08	1.71	—0.0025	—0.062
09—10	1.52	—0.0046	—0.108
11—12	1.00	—0.0058	—0.129
13—14	1.22	—0.0062	—0.135
15—16	1.02	—0.0058	—0.156
17—18	0.97	—0.0043	—0.120
19—20	1.42	—0.0026	—0.070
21—22	1.65	0.0000	0.000
23—24	2.19	0.0023	0.067
25—26	2.68	0.0051	0.137
27—28	3.24	0.0073	0.209
29—30	4.08	0.0085	0.174
31—32	3.81	0.0085	0.184
33—34	4.02	0.0077	0.187
35—36	3.68	0.0068	0.169

Table 10. *September.*

Wind direction	$e_s - e_a$ mb	$\frac{\partial e_s}{\partial r}\ \frac{mb}{km}$	$\frac{de_a}{d\tau}\ \frac{mb}{hour}$
Calm	3.19	—	0.000
01—02	3.09	0.0046	0.138
03—04	2.31	0.0018	0.054
05—06	2.73	0.0000	0.000
07—08	2.27	—0.0016	—0.047
09—10	1.97	—0.0031	—0.092
11—12	2.05	—0.0036	—0.106
13—14	1.68	—0.0042	—0.116
15—16	1.28	—0.0046	—0.121
17—18	1.32	—0.0045	—0.108
19—20	1.34	—0.0031	—0.069
21—22	1.54	—0.0005	—0.017
23—24	2.33	0.0022	0.072
25—26	2.98	0.0049	0.164
27—28	3.06	0.0066	0.210
29—30	3.55	0.0080	0.259
31—32	3.80	0.0080	0.248
33—34	4.08	0.0080	0.234
35—36	3.84	0.0069	0.219

APPENDIX

After having finished this work, the author became aware of a paper which may support the results arrived at in this investigation.[1]

Simultaneous measurements of the radiation temperature and the conventional thermometric temperature demonstrate the existence of a thin layer on the ocean's surface, averaging about 0.6°C colder than the average "surface temperature" found by conventional methods. The two scientists disclose that the layer is less than 0.1 millimetre thick.

The measurements were made with a special instrument, called an infrared radiometer. The radiometer had a special sensitivity in the band from 6 to 20μ, a region in which the absorption in water is so high that 98 per cent of the radiant flux originates in the first 0.1 millimetre. The radiation from the thin layer was used to calculate its temperature.

The authors of the paper conclude that the cold layer may be expected from evaporation effects and also from long-wave radiation from the surface.

When determining the "surface temperature" by the bucket method, one gets the mean temperature of a relatively thick water layer instead of the temperature of the thin surface film from which the heat transfer and evaporation take place. In fact, the "sea-surface temperature" arrived at will be too high.

Consequently, when the air temperature is measured to be about 0.6°C lower than the "surface temperature", there will be no vertical temperature gradients and therefore no vertical heat transport in the lower layer of the air.

By the methods used in this investigation we found that no vertical heat transport occurred when the air temperature was about 0.7°C lower than the "surface temperature". In view of the approximate method applied in our investigation the difference of 0.1°C is probably not significant.

Similar reasonings may be applicable with respect to humidity. The saturation vapour pressure e_s close to the sea-surface will be too high, when calculating it from the "surface temperature" determined by the bucket method.

[1] Ewing and MacAlister: "On the Thermal Boundary Layer of the Ocean". Science May 1960.

REFERENCES

1. Eggvin, J., and F. Spinnangr: Fog forecasting and sea-surface temperature. *Reports on Norwegian fishery and marine investigations*, **VII**, No. 9, 1944.
2. Frogner, E.: Means and extremes of sea temperature by the Norwegian coast. *Geof. Publ.*, **XV**, No. 3, Oslo 1948.
3. Krauss, W.: Temperatur, Salzgehalt und Dichte an der Oberfläche des Atlantischen Ozean. *Wiss.schaft. Ergebn. d. Forsch. u. Vermess. Schiff «Meteor» 1925—1927,* **V**, 3. Lief., Berlin 1958.
4. Mosby, H.: Temperaturvekslinger i Nordatlanteren. *Naturen*, Nr. 4—1942, Bergen.
5. Mosby, H.: The evaporation from the oceans. *Comptes rendus* **XI**, *Assemblé général de l'U.G.G.I.*, Toronto 1957.
6. — The sea-surface and the air. *Scientific results of the Norwegian antarctic expeditions 1927—1928.* No. 10, 1933, Oslo.
7. Åmot, A.: On the temperature difference between the air and sea-surface and its applicability in the practical weather analysis. *Met. Ann.* **1**, Nr. 19, 1944.

LIST OF SYMBOLS

a = altitude of observations above sea-surface.
t_s = sea-surface temperature.
t_a = air temperature at altitude a.
Δt = $t_s - t_a$.
e_s = saturation vapour pressure at the temperature t_a.
e_a = vapour pressure at altitude a.
Δe = $e_s - e_a$.
v_r = wind velocity along the axis r.
v_a = wind velocity at altitude a.
τ = time.
Q_e = energy used for evaporation.
Q_h = sensible heat exchanged between sea and atmosphere through convection.