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ESTIMATION OF HEAT FLOWS FROM THE OCEAN TO THE ATMOSPHERE IN DROP TRANSPORT

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Estimations of heat flows from the ocean to the atmosphere in drop transport during storms are given. It is shown that: (1) heat transport by sprays generated by strong wind and wave collapse is insignificant, compared with turbulent transport; (2) sprays promote heat transport from the lower to the higher atmosphere.

Heat and mass transfer processes take place continuously between the ocean and the atmosphere. The incoming solar radiation brings the ocean and atmosphere to the thermally nonequilibrium state (the upper 100 m layer of the ocean is warmer than the atmosphere), and irreversible processes of evaporation, IR radiation, and convection originate on the surface of the water-air contact. Evaporation plays a leading part among these processes [1]. Its intensity depends on the temperature of the ocean and, the most essential thing, on the wind speed in the near-water atmospheric layer. As the wind intensifies to the storm level, evaporation grows strongly, and the total "ocean-atmosphere" heat flow increases materially. By the method of direct gradient measurements of temperature at a wind speed greater than 30 m/s flows having a density of about 10 kW m^{-2} were recorded [2].

Meanwhile, there exists a concept that the heat and mass transfer increases because of sprays, whose flow to the atmosphere during the storm becomes considerable. For instance, in [3] it is maintained that already at a wind speed of 20–25 m/s heat and vapor flows from sprays are comparable in magnitude with those occurring on the ocean surface (Fig. 1).

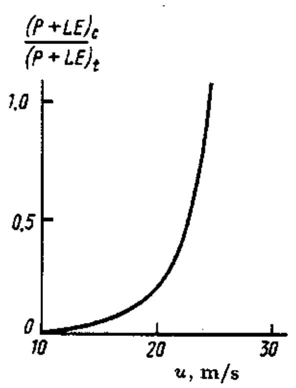

Fig. 1

Relation of vertical heat transport by sprays $(P+LE)_c$ to turbulent flow $(P+LE)_t$ depending on wind speed according to [3].

This point of view gives rise to reasonable doubts, because the heat flow from the ocean surface is ensured by the supply of energy from the active layer of the ocean, whereas the heat flow entrained by drops is provided only by the heat contained in the drops themselves. Therefore, it is of interest to estimate the heat flow provided by the drops purely in terms of energy. The objective pursued by the authors of the present paper is to carry out such estimation.

Let us estimate quantitatively the effectiveness of action of the drop mechanism of heat transport in the ocean-atmosphere system.

During storm air is forced into the water by collapsing wave crests, and a considerable number of air bubbles is formed in the near-surface water layer. Drops are produced as the bubbles burst on the surface of the water: the bubble dome becomes fragmented into a large number of so-called "film" drops (with the diameter $d \sim 10^{-3}$ cm), and then a jet, rising from the center of the air cavity, disintegrates into several "jet" drops, whose size depends on the size of the bubble [4, 5]. Film drops directly carry the water of the surface water layer, the thickness of which is equal to the thickness of the bubble dome film (2-5 μm). The material for the formation of jet drops is first distributed over the inner envelope of the bubble and is connected genetically with the surface layer of sea water having a thickness of the order of the diameter of the sprays [6].

For estimating the amount of heat transported by the drops it is necessary to know the density of the vertical stream of the mass of water, transported by the drops, $\mathcal{E}_w$ (kg m^{-2} s^{-1}). It is obvious that this parameter can be calculated by the formula

$$\mathcal{E}_w = \frac{4}{3}\pi\overline{r^3}\rho_w j_0,$$

where $\overline{r^3}$ is the average-mass radius of the drop; j_0 is the number of drops formed during 1 s on 1 m^2 of the water surface; ρ_w is the density of water.

The data on the average-mass radius of sprays at different wind speeds are reported in [7] (Fig. 2), and the vertical stream of sprays j is estimated in [3, 8, 9].

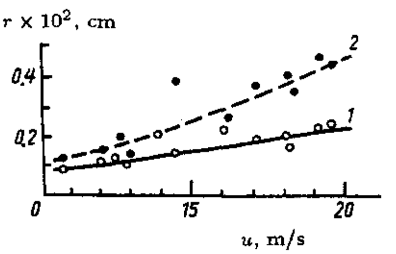

Fig. 2

Average (1) and average-mass (2) radius of sprays at the level $h = 15$ cm [1].

In [8] Toba, when carrying out measurements in an aerohydrodynamic channel for a wind speed from 15 m/s, obtained the following relation for the vertical stream of sprays at 15 cm above the water level:

$$j = j_* \exp\{0.40(u_{10} - 15)\},$$

where j_* is the stream of sprays at the wind speed of 15 m/s, u_{10} is the wind speed.

In [9] the vertical stream of sprays was measured under field conditions at 13 cm above the water level. The setup used by the author of [9] allowed determination of the number of drops having a radius ranging from 4×10^{-3} to 7×10^{-2} cm.

The data on the value of $\mathcal{E}_w$ at the height of 13 cm are reported in [3]. Measurements were also carried out under field conditions and drops having a radius from 0.7×10^{-3} to 0.1 cm were recorded.

Using the data reported by the above-cited authors, we shall present the values of ε_w at 15 cm above the water level at different wind speeds (Table 1).

From Table 1 it is seen that the densities of water streams $\mathcal{E}_w$ at the level of approximately 15 cm above the surface of water, obtained on the basis of the data reported by different authors, are fairly consistent and range from 10^{-6} to 10^{-5} kg m^{-2} s^{-1}. Nevertheless, the vertical stream of sprays at the level of their generation must differ substantially from that measured at the height of 13–15 cm. Therefore, in [7] a model is proposed for estimating the density of the stream of the mass of sprays just on the interface. It is supposed that in the whitecap "falling down" from the crest to the leeward slope the concentration of air bubbles is

Table 1

Wind speed (m/s)	11	16	20	23
$\mathcal{E}_w$, calculated from the data reported by Toba [8] (kg m^{-2} s^{-1})		9.1×10^{-7}		7.8×10^{-5}
$\mathcal{E}_w$, calculated from the data reported by Monahan [9]	9.0×10^{-7}	1.8×10^{-6}		2.8×10^{-5}
$\mathcal{E}_w$, borrowed from Bortkovskii [3]	2×10^{-6}	5.0×10^{-6}	2×10^{-5}	

Table 2

u_{10}, m/s	15	20	30
S_1, %	1.0	2.0	3.0
S_2, %	11.5	22.0	35.0
$\mathcal{E}_w$, kg m^{-2} s^{-1}	7.9×10^{-3}	14.0×10^{-3}	31.5×10^{-3}

maximum, and that in the wake of the whitecap (in the foam) their quantity diminishes as it recedes from the crest. The data on the area occupied by the whitecap and foam, borrowed from [7], are presented in Table 2 (S_1 and S_2 are the relative areas occupied by the whitecaps and foam, respectively).

The mass of sprays generated as the bubbles collapse can be estimated on the basis of the data reported in [7], according to which each bubble generates a drop, whose size is 1/10 of the bubble radius. Consequently, the volume of the drops is equal to 1/1000 of the volume of air in the whitecap and foam. The vertical stream of the mass of sprays at $\rho_w = 1000$ kg/cm^3 will be $\mathcal{E}_w = 3.5 \times 10^{-2} u_{10} S$, where S is the area of the whitecap. Table 2 lists the values of $\mathcal{E}_w$ at the water-air interface at different wind speeds, obtained in [7] with the help of the above-mentioned model.

It is of fundamental importance to point out here that the data presented in Table 2 on the density of the vertical stream of the mass of sprays exceed all the estimates known at present. For determining the magnitude of the heat and moisture streams from the drops in the near-water layer of the atmosphere, it is necessary to take into account the intensity of the generation of drops on the surface of the ocean, their size and speed distribution, the height of ascension. Due to the fact that quantitative estimates of the streams of the mass $\mathcal{E}_w$, reported by different authors [3, 7–9], differ by several orders of magnitude, and there are no direct measurements of $\mathcal{E}_w$ under field conditions, we shall estimate the streams of heat and moisture for all the presented values of $\mathcal{E}_w$.

Heat balance equation for the drop has the form

$$cm\frac{dT}{d\tau} = H + LE + R,$$

where H is the flow of heat in contact heat transfer, LE is the flow of energy for evaporation, and R is the radiation balance of the drop, corresponding to the difference of the radiation absorbed and emitted by the drop.

In this case

$$H = 4\pi r\chi(T_2 - T_1)(1 + 0.23\sqrt{\mathrm{Re}}),$$
$$E = 4\pi r D(q_s(T_2) - q_a(T_1))(1 + 0.23\sqrt{\mathrm{Re}}), \quad (1)$$
$$R = Q - 16\pi r^2 \delta T_1^3(T_2 - T_1),$$

where T_2 is the temperature of the drop, T_1 is the ambient temperature, $q_s(T)$ is the density of water vapor at the temperature T, $q_a(T)$ is the density of water vapor in air, Q is the heat absorbed by the drop from the air, D is the coefficient of diffusion of water vapor in air, χ is the thermal conductivity of water, δ is Stefan-Boltzmann's constant, Re is Reynolds' number.

The order of magnitude of the radiation term R was estimated in [10] for the flux of solar radiation at the upper boundary of the atmosphere and the characteristic drop diameter of 100 mm at $T_2 = 273$ K. In this case the heat delivery by radiation does not exceed 1% of the losses for contact heat transfer, the absorption of solar radiation does not exceed 10^{-6} W/m^2, and the radiation term is negligibly small, compared with H and LE (their value is ~ 10 W/m^2).

Drops are formed from the water of the cold surface film of the ocean, and their temperature T_s is equal to the mean temperature in the drop formation layer. The value T_s can be determined from the data of field observations of the temperature profile in the cold film. Contact heat transfer is constantly taking place between the drop and ambient air. Evaporating in the atmosphere of the near-water layer, not saturated with moisture, the drop cools down, and its ultimate temperature tends to reach the dew point. At the relative humidity of air equal to 90%, most often recorded during storm in the near-water layer of the atmosphere, the dew point is 1–2°C lower than the temperature of the atmosphere. However, as soon as the drop becomes colder than the ambient air, it starts receiving heat from the latter. Combined action of the evaporation and contact heat transfer with the atmosphere leads to that the temperature of the drop becomes equal to the constant value of T_0, which is equivalent in its meaning to the temperature of the wet-bulb thermometer.

Thus, if $LE = H$, where H and E are found from (1), and $q_s(T_0) = q_s(T_a) + \alpha(T_0 - T_a)$, then

$$T_0 = T_a - \frac{DL(q_s(T_a) - q_a)}{\chi + DL\alpha}, \quad (2)$$

where $\alpha = 10^{-3}$ kg m^{-3} K^{-1}, T_a is the temperature of the atmosphere in the near-water layer. We would like to point out that in the determination of T_a it is necessary to take into account the inverse distribution of the temperature in that layer of the air, where the major mass of sprays is concentrated. Evaporation of the drop at T_0 occurs only due to the heat received by the cold drop from the atmosphere.

This gives grounds for concluding that, in the first place, heat delivery from the ocean to the atmosphere in drop transport can be realized only via a change in the internal energy of drops in the atmosphere:

$$\Delta q = cm(T_0 - T_s),$$

and the flow of heat to the atmosphere in this case is

$$\left(\frac{1}{\Delta S}\frac{dQ}{dt}\right)_1 \equiv W_1 = c\mathcal{E}_w(T_0 - T_s), \quad (3)$$

where ΔS is the unit area, W is the density of the heat flow from the ocean to the atmosphere, c is the specific heat capacity of sea water, and T_s is the mean temperature of that layer on the surface of the ocean, from which the drops are formed; in the second place, the drops take part in the altitude redistribution of heat in the atmosphere: cold drops ($T_0 < T_a$) receive atmospheric heat, due to which their evaporation occurs. Since the height of ascension of the drops does not exceed (in the main) 15–20 cm (Fig. 3), they receive heat in the lower atmosphere, whereas condensation takes place at the height over 1 km, i. e., at the level of the formation of the clouds.

The flow of heat W_2 (W/cm^2) from the lower to the higher atmosphere is equal to the contact flow of heat from the air to the cold drops:

$$\left(\frac{1}{\Delta S}\frac{dQ}{dt}\right)_2 \equiv W_2 = 4\delta r\chi(T_a - T_0)n; \quad n = \frac{P(v)h}{(4/3)\pi\rho_w r^3}, \quad (4)$$

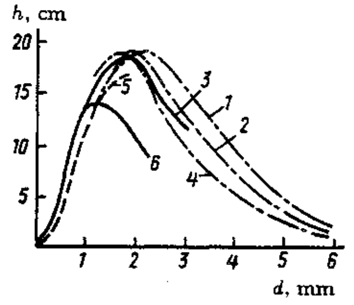

Fig. 3

Dependence of the height of ascension of the drop on the diameter of the bubble upon its collapse at $T_w = 4°$ (1), 16° (2), 22–26° (3), 30° (4), 4° (5), and 4° (for fresh water) (6) [11].

where r is the mean radius of the drops, $P(v)$ (g/cm^3) is the water saturation of the near-water layer of the atmosphere, $h \approx 20$ cm is the thickness of the near-water layer of the atmosphere, in which the major mass of the drops is contained, n is the average number of drops above 1 m^2 of the ocean surface, ρ_w is the density of water.

In accordance with the foregoing, we shall estimate the magnitude of the heat flows, transported by drops in the ocean-atmosphere system under stormy conditions. The characteristic vertical temperature profile near the ocean-atmosphere interface is presented in Fig. 4. The values of the density of the heat flow from the ocean to the atmosphere W_1, caused by the carrying away of the sprays, are listed in Table 3.

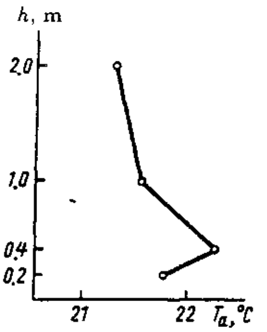

Fig. 4

Characteristic temperature profile in the near-water two-meter atmospheric layer [12].

It is necessary to point out here that the values W_1, calculated by us from the data on the rate of generation of the drops, borrowed from [7], differ essentially from the value $W_1 = 500$–1000 W/m^2, obtained in [7]. It is natural that so strongly overestimated values obtained in [7–9] suggest that the main role in the heat transfer between the ocean and the atmosphere during storm is played just by the drop transport. We think that several factors are responsible for the serious error in the calculations. First, the authors left out of account the existence of a cold film on the surface of the ocean, they did not take into consideration changes in the temperature and humidity of air over the height of the near-water layer. Second, it is erroneous to consider the air humidity during storm to be 80% when we deal with the layers of air directly adjacent to the surface of water. These inaccuracies lead to an almost ten-fold overestimation of the resultant value of the flow W_1. Furthermore, in [7] it is assumed that the coming of heat delivery from water to air, caused by other mechanisms (besides the drop mechanism) during storm increases only a little, whereas direct gradient measurements of the temperature in the cold surface cold film demonstrated that at wind speeds $u \sim 30$ m/s it increases to 10 000 W/m^2 and over [2] and the contribution made by the drops to the transport of energy from the sea to the atmosphere remains insignificant as before.

Table 3

u, m/s	11	16	23	$\mathcal{E}_w$ is borrowed from Toba [8]
W_1, W/m^2	3.1×10^{-4}	3.8×10^{-3}	0.33	
u, m/s	11	16	23	borrowed from Monahan [9]
W_1, W/m^2	3.1×10^{-3}	8.1×10^{-3}	0.12	
u, m/s	11	16	19	borrowed from Bortkovskii [7]
W_1, W/m^2	8×10^{-2}	0.21	0.84	(experimental data)
u, m/s	15	20	30	borrowed from Bortkovskii [7]
W_1, W/m^2	33	58	132	(theoretical data)

The flow of heat from the lower to the upper atmosphere, calculated by formula (4) with recourse to the data of field measurements of water saturation (Fig. 5) and to the size distribution of the drops (see Fig. 2) is 15 W/m^2 at $u = 5$ m/s and 130 W/m^2 at $u = 15$ m/s.

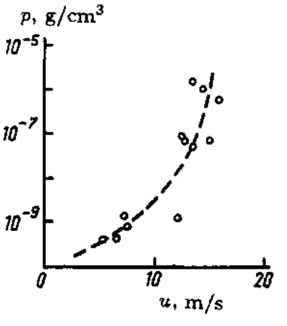

Fig. 5

Dependence of the mass of drops (water saturation) on wind speed [9].

It is also of interest to estimate the influence produced by foam and whitecaps, which cover during storm an appreciable part of the ocean surface (see Table 2), on the processes of heat transfer in the ocean-atmosphere system. We think that foam, whose heat conductivity is low, thermally insulates portions of the ocean, thus obstructing the heat transfer between the ocean and the atmosphere.

The estimate of the heat flows from the ocean to the atmosphere through the foam can be derived in the following manner:

$$W_p = \lambda_p \frac{T_1 - T_2}{h_p},$$

where λ_p is the heat conductivity of the foam, h_p is the thickness of the foam layer, T_1 and T_2 are the temperatures of the upper and lower surfaces of the foam, respectively. We consider that $T_1 = T_0$ and T_2 is the temperature of the ocean.

The equation for determining the thermal conductivity coefficient of the foam [13] is

$$\lambda_p = \frac{2}{3\beta}\lambda_w + \lambda_a,$$

where $\beta = (V_a + V_w)/V_w$ is the foam factor (V_w and V_a being the volume of water and air in the foam).

The data on the disperse structure of sea foam can be found in [14], where the estimate $\beta \sim 10$ is given.

The relation for h_p at $\delta = 35\%$, obtained in laboratory measurements, is presented in [15]:

$$h_p = 10.5 + 0.21\,T_2,$$

where T_2 and h_p are measured in degrees Centigrade and in millimeters, respectively.

With these data taken into account, the value W_p is approximately 30 W/m^2, whereas without the foam the same portion of the ocean surface would transfer about 10 kW/m^2 to the atmosphere. The tabulated data on the area occupied by the foam show what part of the ocean surface is "idle" in the heat transfer with the atmosphere.

We shall now formulate the main conclusions made in this study.

1. The heat transport by sprays generated in the case of strong wind and the collapse of waves is insignificant.

2. Since the drop is detached from the main source of heat (from the ocean), the transport of heat by the drops from the ocean to the atmosphere under storm conditions, compared with the heat flow directly from the ocean surface, does not exceed 2%.

3. At a wind speed of up to 15 m/s the drops make a contribution to the heat transport from the lower to the upper atmosphere not exceeding 130 W/m. The drops, ejected from the ocean during storm, evaporate and during the evaporation they receive heat from the lower atmosphere, because the amount of heat accumulated in the drop itself is not sufficient for its evaporation. This heat is released in the course of vapor condensation at the height of cloud formation.

4. The foam which is formed as wind-driven waves collapse on the surface of the ocean and which is by itself a source of sprays, obstructs heat transport from the ocean to the atmosphere. Thus, at the wind speed of 30 m/s about 30% of the heat flow from the ocean to the atmosphere is cancelled.

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11 July 1991

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