On the giant sea-salt particles in the atmosphere

II. *Theory of the Vertical Distribution in the 10-m Layer Over the Ocean*[1]

By YOSHIAKI TOBA,[2] *The University of Chicago and Kyoto University*

(Manuscript received 14th December 1964)

ABSTRACT

The vertical distribution of giant sea-salt particles in the 10-m layer over the sea surface, where humidity, eddy diffusivity, etc., sharply change with height, is analytically studied. A steady state with no horizontal gradient and with a logarithmic distribution of wind speed is assumed, but the vertical distributions of temperature and vapor pressure are taken into account in such a way that each droplet is assumed to instantaneously attain its equilibrium salinity for the environmental air at each level, and to remain as a droplet of supersaturated salt solution when the relative humidity is lower than 75 %.

It is deduced that the vertical distribution of the number concentration of each class interval of mass of salt contained in a droplet should be approximately expressed by a straight line on a logarithmic diagram. Also, the logarithm of the ratio of the concentration of the particles at the sea surface and at the 10-m level is proportional to the two thirds power of the mass of salt, and inversely proportional to the wind speed at the 10-m level. The proportionality factor is approximately a function of the relative humidity at the 10-m level and a friction factor. This proportionality factor is calculated for various values of the variables and the effect of wind speed and relative humidity on the vertical distribution of particles is predicted.

When this result is applied to data of the concentration and the effective production rate of sea-salt particles at the 10-m level compiled in the previous paper, complicated features disappear to give very simple patterns of the production rate or the surface concentration of sea water droplets as a function of wind force and of mass of salt. This represents predicted processes of production of sea-salt particles at the ocean surface, and gives some agreement with the author's 1961 wind flume experiment.

1. Introduction

In the previous paper I[3] we tried to elucidate mechanisms that would rule the distribution of giant sea-salt particles in the atmosphere as a whole by the use of various observational data available from many references. The link, however, between the distribution in the atmosphere as a whole and the process of production of the particles at the very sea surface, is found within the lowest atmospheric layer where the eddy diffusivity, the humidity, etc., sharply change. In other words, the wind speed, the humidity and other factors in this layer seem to have an essential effect in the supply of the sea-salt particles to the overlying atmosphere.

The importance of this layer will be easily recognized if we regard it in the following way. If the humidity of the overlying atmosphere is low, sea water droplets produced at the sea surface quickly become drier and lighter, and more particles can be suspended and supplied to the atmosphere than in conditions when the humidity is higher. Also the distribution of eddy diffusivity near the sea surface is closely related to the wind speed, and the larger the eddy diffusivity near the surface, the more sea-salt particles will be supplied.

In meteorology and oceanography, surface values of wind speed, temperature and other factors are usually referred to the 10-m level, since they attain more regular values at this height. Furthermore, observations at sea are made at or near this height, and 10 m is the WMO international standard for surface wind measurements. In the following sections, the

---

[1] The research in this paper was sponsored by the Atmospheric Sciences Program, National Science Foundation, NSF grant G 22292.

[2] Visiting Research Associate at the University of Chicago from Kyoto University for the period of September 1963 to March 1965.

[3] For convenience the letter "I" is used in referring to the following paper: Y. Toba, On the giant sea-salt particles in the atmosphere. I. General features of the distribution, *Tellus* vol. 17, No. 1 (1965).

atmospheric layer between the 10-m level and the sea surface is considered as the above-mentioned boundary layer, and the effect of wind speed, humidity etc. on the distribution of giant sea-salt particles in this layer will be analytically studied, to obtain a new basis for an estimate of the geographical distribution of the particles over the world ocean.

2. Equation for the vertical distribution of sea-salt particles in the 10-m layer

2.1. STRUCTURE OF THE 10-M LAYER AND THE RELATED PROPERTIES

Since sea-water droplets are produced after the wind speed exceeds some value and the air entrainment into wind waves occurs, we assume that the sea surface is fully rough. Although some deviation exists according to the thermal stratification, we assume, as a first approximation, that the vertical distribution of wind speed, u, is expressed by the logarithmic law for the adiabatic atmosphere:

$$\frac{u}{u_*} = \frac{1}{k} \ln \frac{z+z_0}{z_0}, \tag{1}$$

where k is von Kármán's universal turbulence constant ($k = 0.4$), z the height above the sea surface, z_0 the roughness length, and u_* the friction velocity which is defined by

$$u_* = \sqrt{\frac{\tau_0}{\varrho_a}}, \tag{2}$$

where τ_0 is the shearing stress of the wind over the sea and ϱ_a the density of air. However, we take into account the vertical distribution of temperature itself, together with that of vapor pressure, when we deal with the equilibrium size of droplets. This will be discussed in 2.3 and 2.4.

One of the factors which enter in the following study is the friction factor or the drag coefficient, γ_l^2, which is a nondimensional factor defined by

$$\tau_0 = \gamma_l^2 \varrho_a u_l^2, \tag{3}$$

where the subscript l refers to values at a standard level which the surface wind speed is usually referred to, and which represents in the main part of this paper the top of the 10-m layer ($l = 1000$ cm). Formerly, it had been considered that there was an abrupt change of the value of γ_l^2 as well as other properties at the wind speed of 6–7 m sec^{-1}: the value was about 0.8×10^{-3} for weaker wind speeds and about 2.6×10^{-3} for stronger wind speeds (SVERDRUP, 1946; MUNK, 1947; etc.). BROCKS (1955), however, re-examined data of vertical distribution of wind speed, temperature and water vapor by WÜST (1937), MONTGOMERY (1940) and SVERDRUP (1946), and reported that there was no such abrupt change. DEACON *et al.* (1956), also showed, by observations of wind speed and temperature distributions over the sea, a gradual change in γ_l^2 with wind speed: from $\gamma_l^2 = 1.0 \times 10^{-3}$ for $u_l = 4$–5 m sec^{-1} to $\gamma_l^2 = 2.1 \times 10^{-3}$ for $u_l = 10$–15 m sec^{-1}. BROCKS (1959) showed by observations at sea that $\gamma_l^2 = 1.5 \times 10^{-3}$ and that the variation with wind speed seemed to be insignificant for the range of $u_l = 2$–14 m sec^{-1}. WILSON (1960) synthesized estimates of γ_l^2 by various authors and reported that the average value was $(1.5 \pm 0.8) \times 10^{-3}$ for low wind speeds of 4–5 m sec^{-1} and $(2.4 \pm 0.5) \times 10^{-3}$ for high wind speeds of about 18 m sec^{-1}. KUNISHI (1963) reported as a result of a wind flume experiment that γ_l^2 remained nearly constant with a value of about 1.6×10^{-3} near the wind speed of $u_l = 10$ m sec^{-1}, and it had even a slight trough at 12 m sec^{-1} and again increased. The value of γ_l^2 as a function of wind speed does not yet seem to be established and we will mainly use $\gamma_l = 0.04$ for the present.

Since the value of z_0 is related to γ_l^2 by equations (1), (2), and (3), that is by

$$\frac{1}{\gamma_l} = \frac{1}{k} \ln \frac{z_l + z_0}{z_0}, \tag{4}$$

the value of z_0 corresponding to $\gamma_l = 0.04$ becomes 0.045 cm. If we use $\gamma_l = 0.05$ for higher wind speeds, the corresponding z_0 will be 0.34 cm.

As to the vertical distribution of eddy diffusivity very close to the sea surface, which is aerodynamically rough, there have been different assumptions. SVERDRUP (1951) synthesized five assumptions and compared them with some observational data. It seems that he could not choose one among the five because data were not sufficient, especially as to the value of z_0: he used $z_0 = 0.6$ cm which was estimated by ROSSBY (1936). ERIKSSON (1959) calculated an example of the vertical distribution of relative

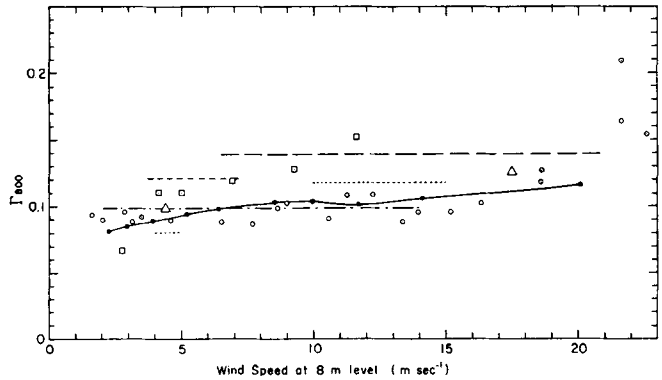

FIG. 1. Comparison of the value of Γ_{800} obtained from various sources (see text). ○, OKUDA & HAYAMI (1959); ●—●, KUNISHI's 1963 γ_l^2 and Sverdrup's assumption No. 4; □, BROCKS (1955); ----, Jacobs' K and Sverdrup's assumption No. 4; △, WILSON's 1960 γ_l^2 and Sverdrup's assumption No. 4; ······, DEACON *et al.* 1956 γ_l^2 and Sverdrup's assumption No. 4; —·—, BROCKS' 1959 γ_l^2 and Sverdrup's assumption No. 4; — —, $z_0 = 0.6$ cm and Sverdrup's assumption No. 4.

humidity by the use of Sverdrup's assumption No. 2: "Next to the surface there is a true diffusion layer of thickness $\delta = 11.5\ \nu/u_{*r}$, where u_{*r} applies to a rough surface and ν is kinematic viscosity of air. From the top of this layer the eddy diffusivity increases at the same rate as the eddy viscosity above a rough surface." According to the present situation, however, it seems that Sverdrup's assumption No. 4 is better than that of No. 2 as inferred below.

The assumption No. 4 was expressed as: "There exists no layer of true diffusion. The eddy diffusivity is at all levels identical with the eddy viscosity,

$$D = ku_*(z + z_0).\text{''} \tag{5}$$

MONTGOMERY (1940) introduced an evaporation coefficient, Γ_l, defined by

$$\Gamma_l = -\frac{1}{\sigma_0 - \sigma_l}\frac{d\sigma}{d\ln z} = -\frac{1}{q_0 - q_l}\frac{dq}{d\ln z}, \tag{6}$$

where σ is the vapor pressure, q the specific humidity and the subscripts o and l signify the values at the water surface and at the l-level stated before, respectively. In this case, the evaporation from the sea surface, E, is expressed by

$$E = \varrho_a k \gamma_l \Gamma_l (q_0 - q_l) u_l, \tag{7}$$

but Γ_l takes various forms according to the assumption as to the vertical distribution of eddy diffusivity, and if we use the assumption No. 4, it must have the form (SVERDRUP, 1946):

$$\Gamma_l = \left(\ln\frac{z_l + z_0}{z_0}\right)^{-1} \approx \left(\ln\frac{z_l}{z_0}\right)^{-1}. \tag{8}$$

In this case, using equation (4), it follows that

$$\Gamma_l = \frac{\gamma_l}{k}. \tag{9}$$

Consequently, we can examine the assumption No. 4 by comparing the value of Γ_l which was obtained from the vertical distribution of vapor pressure by equation (6) and that which was obtained from the vertical distribution of wind speed by use of equation (9). Open circles in Fig. 1 represent values of Γ_{800} obtained from specific humidity distribution by OKUDA & HAYAMI (1959) in a wind flume 21.6 m long. The corresponding values of Γ_{800} from wind speed distribution were calculated from data of γ_l^2 by KUNISHI (1963), which were obtained in the same wind flume, and entered by black circles connected by solid lines. (KUNISHI's data were converted to the condition $l = 800$ cm to meet the data by OKUDA & HAYAMI.) It seems that the assumption No. 4 is correct between $u_l = 4$ m sec^{-1} and 17 m sec^{-1}, although a value of Γ_{800} obtained from $z_0 = 0.6$ cm and equation (8) comes at a different level in the Figure (a broken line). For wind speeds below about 4 m sec^{-1} the water surface was not fully rough (KUNISHI), and for wind speeds above about 17 m sec^{-1}, the evaporation of droplets seemed to have much affected the specific humidity distribution (OKUDA & HAYAMI). It should be noted here that we may expect a possibility that the evaporation of droplets would not greatly affect the specific humidity over the sea in an equilibrium condition of production of sea water droplets and their fallout, because the excess humidity which is given by the evaporation of sea water droplets may again condense on falling particles as they reach the sea surface.

As a reference, Fig. 1 contains Brocks' 1955 estimate of Γ_l entered as small squares. Also, Jacobs' (1942–1951) $K = 0.143$, which appeared in the equation:

$$E = K(\sigma_0 - \sigma_l)u_l \qquad (10)$$

(E in mm per 24 hr, σ in mb and u_l in m sec^{-1}) and was determined by a combination of energy considerations and climatological data, was combined with the assumption No. 4 to give

$$\Gamma_l = \sqrt{\frac{K}{5.37 \times 10^4 k^2 \varrho_a}}$$

and the Γ_l thus obtained is entered as a dashed line. The previously mentioned Wilson's 1960 γ_l^2, Deacon *et al.* 1956 γ_l^2 and Brocks' 1959 γ_l^2 were combined with the assumption No. 4 (equation 9) and they are entered by triangles, dotted lines and a dash-dot line, respectively. All these data are converted to the condition $l = 800$ cm. Brocks' 1955 values seem to coincide with the dashed line from Jacobs' data, but differ somewhat from the wind flume data, and the dash-dot line, as well as the triangles, is much closer to the wind flume data. Although the value of Γ_l itself has yet to be determined, the assumption No. 4 and consequently the equation (5) seem the best to be assumed in this study. The observations that sea water droplets are ejected to some height at the moment they are produced by bursting of air bubbles, plus the fact that, due to their own inertia and the relative motion of the waves, the droplets are caught by the waves, also support the idea of the absence of the layer of true diffusion expressed in assumption No. 4.

2.2. GENERAL PRESENTATION OF THE EQUATION FOR VERTICAL DISTRIBUTION OF SEA-SALT PARTICLES

Since we are dealing with the atmospheric layer of 10-m thickness over the sea surface, we suppose that the mean vertical motion of air may be neglected. We consider a case where a steady state holds with no horizontal gradient, but the fall velocity by gravitational force, w, of sea-salt particles as well as eddy diffusivity, D, change with height, z. Although the fall velocity varies in a complicated way from droplet to droplet, even if we consider droplets of each class interval of mass of salt contained, we assume for simplicity that each droplet instantly attains its equilibrium salinity for the vapor pressure and temperature of the surrounding air, and that the vapor pressure and temperature are given as a function of z. Consequently, we are to suppose that the terminal velocity is a function of z if the mass of salt is given. Since we are concerned with the 10-m layer, we use the concentration (number per unit volume) of saline droplets, θ instead of mixing ratio of droplets with air molecules, and equation of diffusion of droplets is written as follows:

$$\frac{\partial}{\partial z}(w\theta) + \frac{\partial}{\partial z}\left(D\frac{\partial \theta}{\partial z}\right) = \frac{\partial \theta}{\partial t} = u\frac{\partial \theta}{\partial x} = 0,$$

where t represents time, and x the horizontal coordinate in the direction of u. Boundary conditions are that

$$\theta = \theta_0 \quad \text{at} \quad z = 0,$$
$$\theta = 0 \quad \text{at} \quad z = \infty.$$

If w and D are assumed to have finite, non-zero values even at $z = \infty$, the solution becomes

$$\theta = \theta_0 \exp\left(-\int_0^z \frac{w}{D}dz\right) = \frac{F_0}{w_s}\exp\left(-\int_0^z \frac{w}{D}dz\right), \qquad (11)$$

where F_0 is the production rate, per unit area, of sea-salt droplets at the sea surface, and w_s the terminal velocity of sea-water droplets, i.e. droplets having a radius corresponding to a solution of the salinity of sea water.

Introducing equation (5) into (11) and then replacing $z + z_0$ by z, we obtain the following equation:

$$\ln\frac{\theta}{\theta_0} = -\frac{w_s}{ku_*}\int_0^y \frac{w}{w_s}dy, \qquad (12)$$

where

$$y = \ln\frac{z}{z_0}. \qquad (13)$$

TABLE 1. *Some values of vapor-pressure lowering (mm Hg) of sea-salt solutions after* ARONS *&* KIENTZLER *(1954).*

Chlorinity (‰) Salinity (‰)	20 36.1	40 72.2	70 126	100 181	130 235	160 289
Temperature (°C)						
30	0.56	1.25	2.57	4.17	6.09	8.20
20	0.32	0.67	1.36	2.20	3.28	4.49
10	0.16	0.35	0.68	1.10	1.67	2.33
0	0.08	0.17	0.32	0.52	0.79	1.14
−5	—	—	0.22	0.35	0.54	0.79

In this equation, w_s is a function only of the mass of sea salt, m, contained in a droplet, and w/w_s may be expressed as a function of y, or of z, as will be shown in the following sections, if vertical distributions of water vapor pressure and temperature in the 10-m layer are given. The assumption that each droplet has always an equilibrium salinity for the vapor pressure and temperature of the surrounding air, is examined in 2.6.

2.3. HUMIDITY AND EQUILIBRIUM SIZE OF SALINE DROPLETS

In this section, it is shown that the integrand w/w_s in equation (12) may be approximately expressed as a function of relative humidity only, rather than one of both temperature and water vapor pressure.

Table 1 contains some values of vapor-pressure lowering for various temperatures and salinities of sea-salt solutions, measured by ARONS & KIENTZLER (1954). The salinity range covers roughly values from those near the sea water to a saturated sea-salt solution. Table 2 shows the equilibrium relative humidity on sea-salt solutions calculated from Table 1. For temperatures below 0°C, the relative humidity is referred to the saturation vapor pressure over a supercooled water surface. From these tables, we may see that although vapor pressure lowering changes much with temperature as well as with salinity, the equilibrium relative humidity is fairly independent of temperature. Consequently, the equilibrium relative humidity on sea-salt droplets of the giant class is, as a first approximation, independent of temperature, and furthermore is a function of salinity.

Volume, V, salinity, S, and density, ϱ, of sea-salt droplets are related to those of sea-water droplets, considered as the standard, by

$$\frac{V}{V_s} = \frac{\varrho_s}{\varrho}\frac{S_s}{S}, \tag{14}$$

where the subscript s indicates values for sea-water droplets ($S_s = 35$ ‰). The terminal velocity, w, of droplets of the giant class ($m \leqslant 10^{-8}$ gm) is expressed by Stokes' law:

$$w = \frac{2}{9}\frac{g}{\varrho_a \nu}\varrho r^2, \tag{15}$$

where g is the acceleration of gravity, ϱ_a density of air, ν kinematic viscosity of air, r radius of the droplet. A combination of equations (14) and (15) gives

TABLE 2. *Equilibrium relative humidity on sea-salt solutions* (%) *calculated from Table 1.*

Chlorinity (‰)	20	40	70	100	130	160
Temperature (°C)						
30	98.24	96.07	91.93	86.90	80.86	74.23
20	98.17	96.18	92.24	87.45	81.29	74.38
10	98.26	96.19	92.61	88.05	81.85	74.68
0	98.25	96.28	93.01	88.65	82.76	75.11
−5	—	—	93.05	88.92	82.92	75.02
Range	0.09	0.21	1.12	2.02	2.06	0.88

TABLE 3. *Relations among salinity, specific gravity by* HIGASHI *et al.* *(1931), vapor-pressure lowering by* ARONS *&* KIENTZLER *(1954), equilibrium relative humidity, values of f, and w/w_s for 20°C.*

Chlorinity (‰)	Salinity (‰)	Specific gravity	Vapor-pressure lowering (mm Hg)	Equilibrium RH (%)	f (%)	w/w_s
0	0	0.9982	0	100.00	−1.77	∞
5	9.025	1.0050[a]	0.08	99.54	−1.31	2.45
10	18.05	1.0119	0.15	99.14	−0.91	1.55
19.39	35.00	1.0244[a]		98.23[a]	0.00	1.00
20	36.10	1.0252	0.32	98.17	0.06	0.980
30	54.15	1.0390	0.49	97.21	1.02	0.751
40	72.20	1.0529	0.67	96.18	2.05	0.623
50	90.25	1.0666	0.90	94.86	3.37	0.539
60	108.30	1.0814	1.14	93.50	4.73	0.479
70	126.35	1.0960	1.36	92.24	5.99	0.434
80	144.40	1.1101	1.62	90.76	7.47	0.399
90	162.45	1.1243	1.92	89.05	9.18	0.371
100	180.50	1.1383	2.20	87.45	10.78	0.347
110	198.55	1.1522	2.55	85.45	12.78	0.327
120	216.60	1.1655	2.91	83.40	14.83	0.310
130	234.65	1.1790	3.28	81.29	16.94	0.295
140	252.70	1.1923	3.63	79.38	18.85	0.282
150	270.75	1.2060	4.01	77.13	21.10	0.270

[a] Interpolated values.

$$\frac{w}{w_s} = \left(\frac{\varrho}{\varrho_s}\right)^{\frac{1}{3}} \left(\frac{S_s}{S}\right)^{\frac{2}{3}}. \tag{16}$$

The value w/w_s for 20°C, as the representative temperature, was calculated by equation (16) by the use of values of specific gravity of concentrated sea water by HIGASHI *et al.* (1931), for various salinities. (Values of specific gravity for 20°C were obtained by a smooth interpolation on a graph paper from values for 25°C and 0°C.) A relation between salinity, S, and a relative humidity difference, f, defined by

$$f \equiv \mathrm{RH}(S_s) - \mathrm{RH}(S) = 98.23 - \mathrm{RH}(S), \tag{17}$$

where RH(S) represents the equilibrium relative humidity for salinity S, was calculated from data by ARONS & KIENTZLER (1954), and thus we obtained the relation between f and w/w_s as the following approximate equations, for 20°C:

$$\left.\begin{aligned} &\frac{w}{w_s} = \alpha_1 - \beta_1 \log f + \varepsilon_1 (\log f)^2, \\ &\alpha_1 = 0.760, \quad \beta_1 = 0.480 \quad \text{and} \quad \varepsilon_1 = 0.0820 \end{aligned}\right\} \tag{18}$$

for $f \geqslant 1$, and

$$\left.\begin{aligned} &\frac{w}{w_s} = \alpha_2 - \beta_2 \log (f + 1.77) \\ &\qquad + \varepsilon_2 \{\log (f + 1.77)\}^2, \\ &\alpha_2 = 1.418, \quad \beta_2 = 1.94 \quad \text{and} \quad \varepsilon_2 = 1.02 \end{aligned}\right\} \tag{19}$$

for $1 \geqslant f \geqslant -1$. Some numerical values used in the calculation and relation between f and w/w_s calculated are shown in Table 3. A comparison between calculated values and curves of equations (18) and (19) is seen in Fig. 2.

We will use equation (18) also for relative humidities below about 75% where crystallization occurs in bulk sea water solution, because sea-salt particles very often remain as droplets of supersaturated salt solution until the relative humidity of air goes down to about 30%.

2.4. VERTICAL DISTRIBUTION OF RELATIVE HUMIDITY IN THE 10-M LAYER

In equations (18) and (19), we could express w/w_s in terms of f. In this section, it will be shown that f may be expressed in terms of $y = \ln (z/z_0)$.

Consider as in section 2.2 a condition where a steady state holds, and the wind blows in a

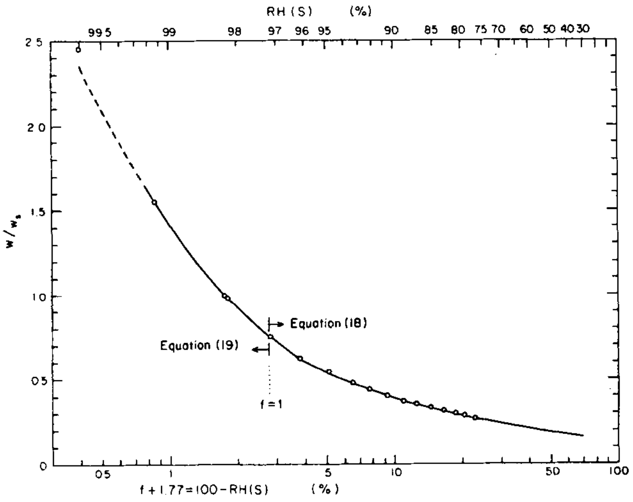

FIG. 2. Relation between equilibrium relative humidity and w/w_s, calculated from data of vapor pressure lowering by ARONS & KIENTZLER (1954) and of specific gravity of sea-salt solutions by HIGASHI *et al.* (1931) (circles) and approximate equations (18) and (19) (a curve).

horizontal direction, where water vapor and heat are diffused only by a vertical eddy diffusion, and thus evaporation from sea-salt droplets does not much affect the distributions of heat and water vapor. In these conditions the vertical distributions of wind speed, heat and water vapor obey a logarithmic law. Under this situation, water vapor transport, or evaporation from the sea surface, E, is expressed by

$$E = -D\frac{d(\varrho_a q)}{dz}. \tag{20}$$

Using equation (5) and replacing $(\varrho_a q)_{z_0}$ by $(\varrho_a q)_0$ and $z + z_0$ by z as in section 2.2, we obtain from equation (20)

$$(\varrho_a q)_0 - (\varrho_a q)_z = \frac{Ey}{ku_*}. \tag{21}$$

Since the water vapor pressure, σ, in c. g. s. units is expressed by

$$\sigma = \frac{pq}{0.622},$$

where p is the atmospheric pressure, and since we may regard ϱ_a as constant because we are concerned with the 10-m layer over the ocean, it follows from equation (21) that

$$\sigma_0 - \sigma = Hy, \quad H = \frac{pE}{0.622\,\varrho_a k u_*}. \tag{22}$$

The H may be regarded as a constant for specific conditions of u_*, E, etc.

A corresponding equation for heat transfer is written as

$$Q = -C_p \varrho_a D\frac{d\theta}{dz}, \tag{23}$$

where Q is the vertical transport of heat by eddy conduction, C_p specific heat of air at a constant pressure, Θ potential temperature, and we have supposed that the eddy diffusivity is the same for water vapor and for heat. Since we are dealing with conditions near the sea surface, we equate Θ with T, absolute temperature, and we obtain

$$T_0 - T = \frac{Qy}{C_p \varrho_a k u_*}. \tag{24}$$

When the Bowen ratio:

$$R_B = \frac{Q}{EL}, \tag{25}$$

where L is latent heat of evaporation, is introduced, equation (24) is reduced to

$$T_0 - T = GHy, \quad G = \frac{R_B L}{C_p}\frac{0.622}{p}. \tag{26}$$

The G may also be regarded as a constant for specific conditions of R_B, etc.

According to KAYE & LABY (1948), the saturation vapor pressure, σ_w, is accurately and conveniently expressed by the Dupré–Rankine formula:

$$\log \sigma_w = A - \frac{B}{T} - C \log T, \tag{27}$$

where A, B and C are constants proper to each substance. The values for water were determined by JULIUSBURGER (1900) as follows: $A = 9.3002742$, $B = 2113.24$, and $C = 0.287710$ for σ_w in mb, up to the critical temperature. When equation (27) is expressed as

$$\sigma_w = 1 \Big/ \left\{ a \exp\left(\frac{b}{T}\right) T^c \right\} \tag{28}$$

and σ_w in dyne cm^{-2}, it follows that

$$a = \exp\left(-\frac{A+3}{\log e}\right) = \exp(-28.3224280),$$

$$b = \frac{B}{\log e} = 4865.91 \text{ (degree)},$$

and $\quad c = C.$

When this expression is used with equations (22) and (26), the difference between the relative humidity at $z = 0$ and that at $z = z$, which is equal to f by the assumption of instantaneous equilibrium, is expressed by

$$\begin{aligned} \frac{f}{100} &= \frac{1}{100}\{\mathrm{RH}(0) - \mathrm{RH}(z)\} = \frac{\sigma_0}{\sigma_{w_0}} - \frac{\sigma}{\sigma_w} \\ &= a\sigma_0 \exp\left(\frac{b}{T_0}\right) T_0^c \\ &\quad - a(\sigma_0 - Hy) \exp\{b/(T_0 - GHy)\}(T_0 - GHy)^c. \end{aligned} \tag{29}$$

Since the value $GHy = T_0 - T$ is generally less than 1 or 2% of T_0, we may use the approximation:

$$\begin{aligned} \exp\{b/(T_0 - GHy)\} &\approx \exp\left(\frac{b}{T_0} + \frac{bG}{T_0^2} Hy\right) \\ &\approx \exp\left(\frac{b}{T_0}\right)\left(1 + \frac{bG}{T_0^2} Hy + \frac{b^2G^2}{2T_0^4} H^2y^2\right) \end{aligned}$$

and $\quad (T_0 - GHy)^c \approx T_0^c\left(1 - \frac{cG}{T_0} Hy\right)$

and equation (29) is reduced to

$$\begin{aligned} \frac{f}{100} = a \exp\left(\frac{b}{T_0}\right) T_0^c Hy \Bigg[1 &- \left(\frac{b}{T_0} - c\right)\frac{G}{T_0}\sigma_0 \\ &+ \left\{ \left(\frac{b}{T_0} - c\right) - \left(\frac{b}{2T_0} - c\right)\frac{b}{T_0}\frac{G}{T_0}\sigma_0 \right\} \frac{G}{T_0} Hy \\ &+ \left\{ \left(\frac{b}{2T_0} - c\right)\frac{b}{T_0} + \frac{b^2c}{2T_0^2}\frac{G}{T_0}\sigma_0 \right\} \frac{G^2}{T_0^2} H^2y^2 \\ &- \left\{ \frac{b^2c}{2T_0^2} \right\} \frac{G^3}{T_0^3} H^3y^3 \Bigg]. \end{aligned} \tag{30}$$

When we consider magnitudes of the factors

$$\left(\frac{b}{T_0} - c\right) \approx \frac{b}{T_0} = 16.5, \quad \frac{G}{T_0}\sigma_0 \leqslant 3 \times 10^{-2}$$

(cf. Table 4)

and $\quad \dfrac{b^2c}{2T_0^2} \approx 42,$

and omit small terms in equation (30), we obtain finally that

$$\begin{aligned} f = 100a \exp\left(\frac{b}{T_0}\right) T_0^c Hy \Bigg[&\left\{ 1 - \frac{bG}{T_0^2}\sigma_0 \right\} \\ &+ \left(1 - \frac{1}{2}\frac{bG}{T_0^2}\sigma_0\right)\frac{bG}{T_0^2} Hy + \frac{b^2G^2}{2T_0^4} H^2y^2 \Bigg] \end{aligned} \tag{31}$$

and we may see that the vertical distribution of relative humidity over the sea surface is expressed by a cubic equation of $y \equiv \ln(z/z_0)$. It can be expressed by a linear equation of y, or a logarithmic law holds only when the terms containing y and y^2 in the brackets may be neglected.

Values of G and $(bG/T_0^2)\sigma_0$ were calculated from mean sea surface temperatures for each 10° zone for both hemispheres, and by the use of values of Bowen ratio that were estimated by JACOBS (1942) and corrected by SVERDRUP (1951). The values are listed in Table 4. According to the table, the value of $(bG/T_0^2)\sigma_0$ is between 0.24 and 0.29 (mean value $= 0.27$) for a large range of latitude between 40° S and 60° N, and it is much smaller south of 40° S. The value of $Hy = \sigma_0 - \sigma$ is, as may be seen from observational data of vertical vapor pressure distribution by MONTGOMERY (1940), SVERDRUP (1946), and FLEAGLE *et al.* (1958),

TABLE 4. *Mean sea surface water temperature, T_0 (after* DEFANT, *1961), Bowen ratio, R_B, estimated by* JACOBS *(1942) and corrected by* SVERDRUP *(1951) and values of G and $(bG/T_0^2)\,\sigma_0$ calculated from these values.*

Latitude	T_0(°K)	R_B	G (10^{-4} cm^2 dyne^{-1} degree)	$\frac{bG}{T_0^2}\sigma_0$
(°N)				
60–70	276	0.45	6.9	0.34
50–60	279	0.31	4.7	0.28
40–50	284	0.21	3.2	0.25
30–40	291	0.15	2.3	0.28
20–30	297	0.11	1.7	0.27
10–20	300	0.10	1.5	0.28
0–10	300	0.10	1.5	0.29
(°S)				
0–10	299	0.10	1.5	0.28
10–20	298	0.10	1.5	0.26
20–30	295	0.11	1.7	0.24
30–40	290	0.14	2.1	0.24
40–50	283	0.17	2.3	0.17
50–60	276	0.20	3.1	0.15
60–70	272	0.23	3.5	0.13

smaller than, and generally much smaller than one third of σ_0 in some mean values, although individual observed values on rare occasions reach a half of σ_0. Consequently, the middle term in the brackets of equation (31) is usually smaller than 0.085, and the last term in the brackets is smaller than 0.01. So the approximation that the vertical distribution of relative humidity in the 10-m layer over the ocean obeys the logarithmic law will not lead us to a great error.

2.5. MODIFICATION OF THE EQUATION FOR VERTICAL DISTRIBUTION OF SEA-SALT PARTICLES

The vertical distribution of wind speed for neutral conditions is expressed by equation (1):

$$\frac{u}{u_*}=\frac{y}{k}. \tag{32}$$

Since we saw, in the previous section, that the vertical distribution of f approximately obeys a logarithmic law, we now express it, as in equation (32), by

$$\frac{f}{f_*}=\frac{y}{k} \quad (\text{for } f_l \neq 0), \tag{33}$$

where f_* is given by

$$f_*=\frac{u_* f_l}{u_l}=\gamma_l f_l. \tag{34}$$

From equations (33) and (34) it follows that

$$f=\eta y, \tag{35}$$

where

$$\eta=\frac{\gamma_l f_l}{k}. \tag{36}$$

Putting, further,

$$\zeta=\log|\eta| \quad (\text{for } f_l \neq 0), \tag{37}$$

we obtain, for $f \geq 1$,

$$\log f=\zeta+\log y \tag{38}$$

and for $1 \geq f > 0$ and $0 > f \geq -1$,

$$\left.\begin{aligned} \log(f+1.77)&=\zeta+\log|y_2|,\\ y_2&=y+\frac{1.77}{\eta}. \end{aligned}\right\} \tag{39}$$

We replace $\log f$ and $\log(f+1.77)$ in equations (18) and (19) by equations (38) and (39), respectively, to obtain for $f \geq 1$,

$$\frac{w}{w_s} = \varphi_1 + \psi_1 \log y + \varepsilon_1 (\log y)^2 \tag{40}$$

and for $1 \geq f > 0$, and $0 > f \geq -1$,

$$\frac{w}{w_s} = \varphi_2 + \psi_2 \log |y_2| + \varepsilon_2 (\log |y_2|)^2, \tag{41}$$

where

$$\left.\begin{aligned} \varphi_i &= \alpha_i - \beta_i \zeta + \varepsilon_i \zeta^2 \\ \psi_i &= 2\varepsilon_i \zeta - \beta_i \end{aligned}\right\} i = 1 \text{ and } 2. \tag{42}$$

Since from equation (35), for $f = 1$,

$$y = \frac{1}{\eta} \quad \text{and} \quad y_2 = \frac{2.77}{\eta}$$

and since $dy_2 = dy$, the integration in equation (12) may be calculated, according to ranges of f_l, as follows:

for $f_l \geq 1$, $\ln \dfrac{\theta}{\theta_0}$

$$= -\frac{w_s}{ku_*}\left[\int_{1.77/\eta}^{2.77/\eta} \{\varphi_2 + \psi_2 \log y_2 + \varepsilon_2 (\log y_2)^2\}\, dy_2 + \int_{1/\eta}^{y} \{\varphi_1 + \psi_1 \log y + \varepsilon_1 (\log y)^2\}\, dy\right], \tag{43}$$

for $1 \geq f_l > 0$, $\ln \dfrac{\theta}{\theta_0}$

$$= -\frac{w_s}{ku_*}\int_{1.77/\eta}^{y+1.77/\eta} \{\varphi_2 + \psi_2 \log y_2 + \varepsilon_2 (\log y_2)^2\}\, dy_2, \tag{44}$$

for $0 > f_l \geq -1$,

$$\ln\frac{\theta}{\theta_0} = -\frac{w_s}{ku_*}\int_{1.77/\eta}^{y+1.77/\eta} \{\varphi_2 + \psi_2 \log |y_2| + \varepsilon_2 (\log |y_2|)^2\}\, dy_2$$
$$= -\frac{w_s}{ku_*}\int_{|y+1.77/\eta|}^{|1.77/\eta|} \{\varphi_2 + \psi_2 \log |y_2| + \varepsilon_2 (\log |y_2|)^2\}\, d|y_2|. \tag{45}$$

For $f_l = 0$, w is always equal to w_s, and the equation (12) is reduced to

$$\ln\frac{\theta}{\theta_0} = -\frac{w_s}{ku_*}\int_0^y dy = -\frac{w_s}{ku_*} y. \tag{46}$$

After performing the integrations, we reach an expression for θ/θ_0 such that

$$\log\frac{\theta}{\theta_0} = -m^{\frac{2}{3}} u_l^{-1} \chi(f_l, \gamma_l, z), \tag{47}$$

where logarithm of the ratio θ to θ_0 is expressed as being proportional to the two thirds power of mass of salt contained in a droplet, $m^{2/3}$, and inversely proportional to the wind speed at the 10-m level, u_l, with a factor χ which is a function of the difference of relative humidity between the sea surface and the 10-m level, f_l, friction factor, γ_l, and height, z. The χ has the following forms:

for $f_l \geq 1$,

$$\chi(f_l, \gamma_l, z) = \frac{\Lambda M}{k\gamma_l}\left[\frac{K_1}{\eta} + U_1 - U_2\zeta + U_3\zeta^2\right],$$

$$K_1 = [(\alpha_2 + M\beta_2 + 2M^2\varepsilon_2) - (\alpha_1 + M\beta_1 + 2M^2\varepsilon_1) - (\beta_2 + 2M\varepsilon_2)(2.77 \log 2.77 - 1.77 \log 1.77) + \varepsilon_2\{2.77 (\log 2.77)^2 - 1.77 (\log 1.77)^2\}],$$

$$U_1 = (\alpha_1 + M\beta_1 + 2M^2\varepsilon_1)\, y - (\beta_1 + 2M\varepsilon_1 - \varepsilon_1 \log y)\, y \log y,$$

$$U_2 = (\beta_1 + 2M\varepsilon_1 - 2\varepsilon_1 \log y)\, y,$$

$$U_3 = \varepsilon_1 y, \tag{48}$$

where K_1 is a constant and U_1, U_2 and U_3 are functions of y, or of z.

For $1 \geq f_l > 0$,

$$\chi(f_l, \gamma_l, z) = \frac{\Lambda M}{k\gamma_l}[\{(\alpha_2 + M\beta_2 + 2M^2\varepsilon_2) - (\beta_2 + 2\varepsilon_2 M)\zeta + \varepsilon_2\zeta^2\}\, y - (\beta_2 + 2\varepsilon_2 M - 2\varepsilon_2\zeta) \times (\xi_2 \log \xi_2 - \xi_1 \log \xi_1) + \varepsilon_2\{\xi_2 (\log \xi_2)^2 - \xi_1 (\log \xi_1)^2\}], \quad \xi_1 = \frac{1.77}{\eta}, \quad \xi_2 = \xi_1 + y, \tag{49}$$

for $f_l = 0$,

$$\chi(\gamma_l, z) = \frac{\Lambda M}{k\gamma_l} y, \tag{50}$$

and for $0 > f_l \geq -1$,

$$\chi(f_l, \gamma_l, z) = \frac{\Lambda M}{k\gamma_l}[\{(\alpha_2 + M\beta_2 + 2M^2\varepsilon_2) - (\beta_2 + 2\varepsilon_2 M)\zeta + \varepsilon_2\zeta^2\}\, y - (\beta_2 + 2\varepsilon_2 M - 2\varepsilon_2\zeta) \times (|\xi_1| \log |\xi_1| - |\xi_2| \log |\xi_2|) + \varepsilon_2\{|\xi_1| (\log |\xi_1|)^2 - |\xi_2| (\log |\xi_2|)^2\}]. \tag{51}$$

TABLE 5. *Values of Δt (sec) corresponding to a distance of fall of droplets in a unit time.* $z = 1$ *m*, $\gamma_l = 0.04$.

		m			
f_l(%)	RH$_l$(%)	10^{-11} (gm)	10^{-10}	10^{-9}	10^{-8}
−1	99.23	1.5	6.4	33	150
5	93.23	.074	.36	1.6	9.4
15	83.23	.017	.088	.41	1.8
30	68.23	.0081	.034	.16	.71
50	48.23	.0034	.017	.074	.31

In these expressions,

$$M \equiv \log e \approx 0.4343$$

and

$$\Lambda \equiv \frac{w_s}{m^{\frac{2}{3}}}.$$

Since Stokes' law holds for giant sea-salt particles ($m \leqslant 10^{-8}$ gm), w_s is expressed by

$$w_s = \frac{2}{9}\frac{g}{\varrho_a \nu}\varrho_s r_s^2 = \frac{2}{9}\frac{g}{\varrho_a \nu}\left(\frac{3000}{4\pi S_s}\right)^{\frac{2}{3}} \varrho_s^{\frac{1}{3}} m^{\frac{2}{3}} = \Lambda m^{\frac{2}{3}}, \quad (52)$$

hence $\Lambda = 4.3 \times 10^6 (\text{cm sec}^{-1} \text{gm}^{-\frac{2}{3}}). \quad (53)$

2.6. Examination of the assumption of instantaneous equilibrium of droplets to the environmental air

In this section the basic assumption of this article that each droplet instantaneously attains its equilibrium salinity for the humidity of the surrounding air is examined. For this purpose, it is useful to determine the time required for a droplet to adjust its size to the change of environmental humidity, which results from its falling at the terminal velocity.

The time rate of change of radius of a droplet is approximately expressed by a well-known equation:

$$\varrho_w r \frac{dr}{dt} = -D' \varrho_a \Delta q, \quad (54)$$

where D' is the diffusivity of water vapor in air, Δq is the difference of specific humidity between the surface of the droplet and the environmental air, and ϱ_w is the density of pure water. Consider a case where a droplet, which has been in an equilibrium with air of a certain level, falls a distance Δz, with its terminal velocity, in a *unit time*. A value of Δq appears corresponding to Δz, and a value of Δr^2 corresponding to Δq, which may be considered as a change in r^2 which should occur for the droplet to reach equilibrium with the air at the new level. Then the value Δt which is derived from equation (54),

$$\Delta t = \frac{-\Delta r^2 \varrho_w}{2 D' \varrho_a \Delta q} \quad (55)$$

is considered as the time required to reach a new equilibrium by the initial evaporation or condensation speed. Whether it is larger or smaller than 1 gives us the criterion.

The values of Δt for $z = 1$ m and $\gamma_l = 0.04$ for various mass of salt contained in a droplet and for various humidities at the 10-m level have been calculated by the use of equations (13), (15), (18), (19), (33), (34), (52), (53), and Table 3, and are shown in Table 5. From Table 5 we see that the speed of adjustment to a new environment is larger for small masses of contained salt and for small relative humidities of the air, and that the assumption of instantaneous equilibrium is valid in most parts of the table. Since, in turbulent diffusion, those droplets that come from above are usually more saline, those from below less saline than the equilibrium droplets at the level considered, and since we are concerned with an average condition, the range of validity may be extended a little more. For $f_l = 0$ or RH$_l = 98.23$, the assumption is exactly valid, because all droplets have the salinity of sea water, although the value of Δt cannot be determined by equation (55), and the assumption will be nearly valid for the f_l close to 0, although the value of Δt

is quite large in Table 6. Consequently, it is concluded that the basic assumption is in general substantiated. However, it is considered that the assumption may not be so valid very close to the sea surface, although the estimate of Δt becomes difficult there. But in this region droplets are being ejected by bubbles, being caught by waves, and a more exact treatment will be extremely difficult.

3. Results of the calculation and discussion

3.1. CALCULATION OF χ AND FEATURES OF THE VERTICAL DISTRIBUTION OF SEA-SALT PARTICLES IN THE 10-M LAYER

Values of χ in equations (48) through (51) were calculated for $\gamma_l = 0.04$ for various relative humidities as a function of z. The results are illustrated in Fig. 3. Values of χ for $z = 10$ m as a function of $f_l + 1.77$ or of RH_l for both $\gamma_l = 0.04$ and $\gamma_l = 0.05$ are illustrated in Fig. 4.

Values of θ/θ_0 were calculated from Fig. 3 and equation (47) for three cases of the value of $m^{\frac{2}{3}}\ u_l^{-1}$. These are shown in Fig. 5 through Fig. 7, and some possible combinations of m and u_l for the three cases are shown in Table 6. It contains some extremely large values of u_l where our calculation for $\gamma_l = 0.04$ may no longer be valid, and very small values of u_l where the basic assumption of the steady state may become unreasonable; these values are entered in parentheses. But these should be interpreted as showing that the concentrations of particles of $m = 10^{-11}$ gm have very little vertical gradient in the 10-m layer for a wind speed larger than a few m sec^{-1}, and that the concentrations of particles of $m = 10^{-8}$ gm have a very large vertical gradient for wind speeds less than those shown in the table.

TABLE 6. *Some combinations of m and u_l for the three cases of the value of $m^{\frac{2}{3}} u_l^{-1}$ illustrated in Figs. 5, 6, and 7.*

	$m^{\frac{2}{3}} u_l^{-1}$ (c. g. s.)		
m (gm)	1.7×10^{-9} (Fig. 5)	1.0×10^{-9} (Fig. 6)	0.5×10^{-9} (Fig. 7)
10^{-11}	$u_l = (0.27)$ (m sec^{-1})	(0.46)	(0.97)
10^{-10}	1.27	2.15	4.31
10^{-9}	5.88	10.0	20.0
10^{-8}	(27.3)	(46.4)	(92.8)

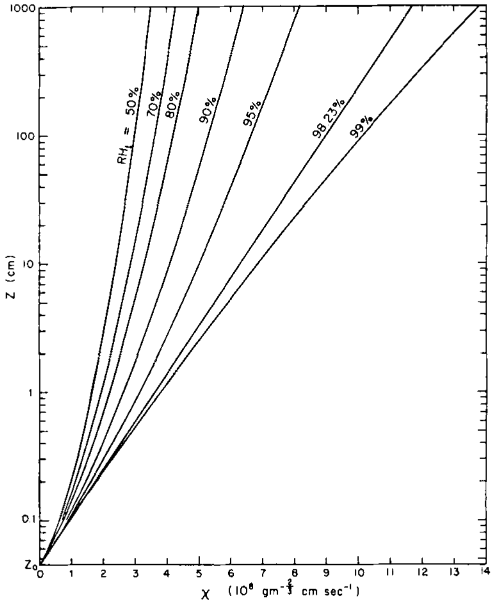

FIG. 3. Values of χ for $\gamma_l = 0.04$.

From these figures, it is seen that the vertical distribution of sea-salt particles in the 10-m layer has a curve very close to a straight line in a logarithmic diagram, and it becomes a straight line when there is no vertical gradient in relative humidity ($\mathrm{RH}_l = 98.23\%$). Consequently if we draw the distribution on a z, $\log\theta$ diagram, it should have a concave curve. It is also seen that in the range of giant particles,

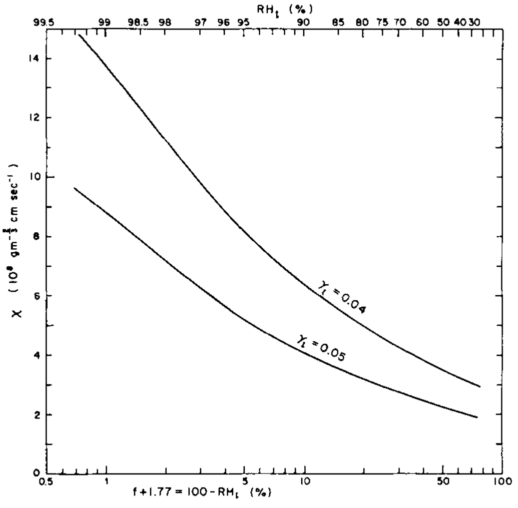

FIG. 4. Values of χ for $z = 10$ m.

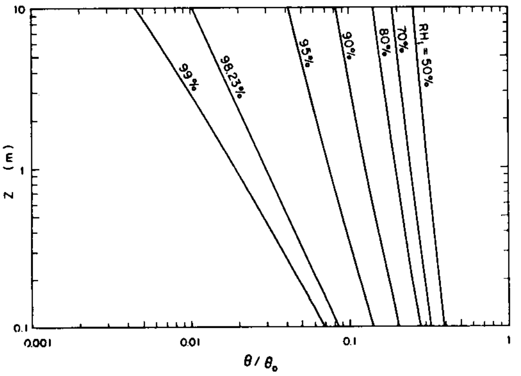

FIG. 5. Values of $\theta/\theta_0\,(\gamma_l = 0.04)$. Case I: $m^{\frac{2}{3}} u_l^{-1} = 1.7 \times 10^{-9}$ (c.g.s.).

f_l as well as u_l effectively affects the concentration of particles at the 10-m level relative to that at the sea surface, θ_l/θ_0. Values of θ_l/θ_0 as a function of m for various wind speeds, u_l for $\mathrm{RH}_l = 80\,\%$ as an example ($\gamma_l = 0.04$) are shown in Fig. 8.

3.2. APPLICATION TO THE DATA OF θ_l AND DISCUSSION

In the previous paper I, the mechanism of distribution of giant sea-salt particles in the atmosphere as a whole was discussed and their concentration, θ_l, and effective production rate, F_l, at the 10-m level were estimated by use of the data of WOODCOCK (1953) obtained at cloud base levels. Since we have predicted features of their distribution in the 10-m layer in this study, now we may subtract the effect of the boundary layer of 10 m thickness from the θ_l and F_l, and estimate the θ_0 and F_0, which may represent the very sea-surface processes. Fig. 9 shows the θ_0 which were calculated from Fig. 10 of I and Fig. 8 of the present paper. (Woodcock's 1953 measurements were made over the sea area where the average RH_l was 80%.) According to Fig. 9, there is a strong tendency that the weaker the wind force, the more extreme high value appears for larger m. However, we cannot accept this tendency as real. Comparing the figure with Fig. 10 of I, we recognize that the excessive values of θ_0 exactly correspond to the values of θ_l smaller than 5×10^{-3} cm^{-3}, and the smaller the θ_l, the more extreme the value of θ_0. This fact leads to the thought that the small number of large particles that came out

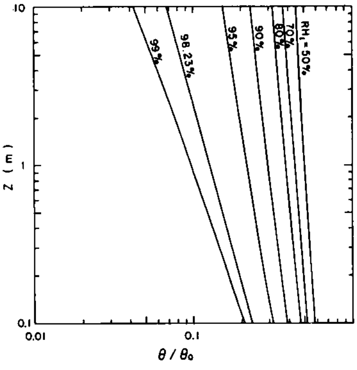

FIG. 6. Values of $\theta/\theta_0\,(\gamma_l = 0.04)$. Case II: $m^{\frac{2}{3}} u_l^{-1} = 10^{-9}$ (c.g.s.).

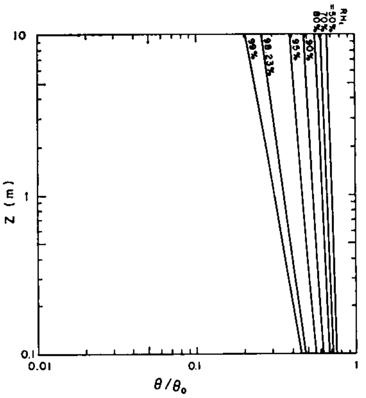

FIG. 7. Values of $\theta/\theta_0\,(\gamma_l = 0.04)$. Case III: $m^{\frac{2}{3}} u_l^{-1} = 5 \times 10^{-10}$ (c.g.s.).

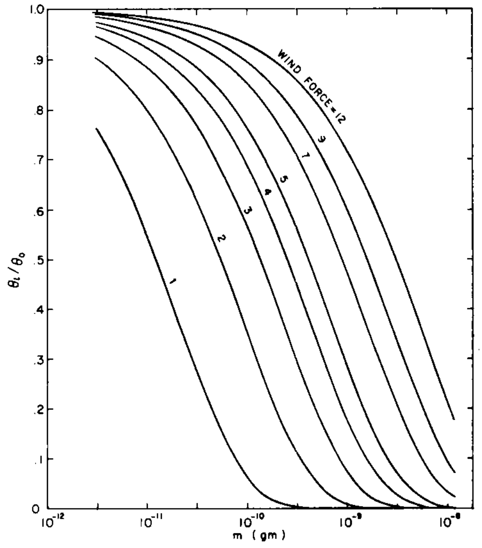

FIG. 8. Values of θ_l/θ_0 as a function of u_l and m for $\mathrm{RH}_l = 80\,\%$ ($\gamma_l = 0.04$).

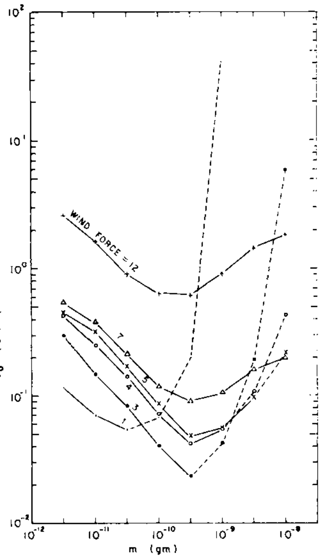

FIG. 9. Estimated concentration of sea water droplets at the sea surface θ_0, calculated from Fig. 10 of I, and from Fig. 8. Values are entered for the range of $\log m = 0.5$.

by coalescence of particles in clouds at the top of upward current that was discussed in I, greatly affected the result for θ_0 when the absolute concentration of the particles or θ_l was low. Points which correspond to $\theta_l < 5 \times 10^{-3}$ cm^{-3} are shown by linking with dashed lines in Fig. 9. If we exclude the dashed lines, we may see that curves for various wind forces have now essentially a similar trend, and that a trough of curves at $m \approx 10^{-9.5}$ gm, which might divide jet drops and film drops[1] as was pointed out in I, is now clear.

Figure 10 shows the estimated production rate at the sea surface, F_0, calculated from Fig. 9 by $F_0 = w_s \theta_0$. We may see the same features as in Fig. 9. Besides, the F_0 increases with log m exponentially in the range of 10^{-8} gm $> m > 10^{-9.5}$ gm for the jet drops. Since 10^{-8} gm corresponds to a sea water droplet of 81 μ diameter, which may be ejected from a bubble of 0.75 mm diameter (BLANCHARD, 1963), it seems that the production rate of bubbles increases with the bubble diameter at least up to 0.75 mm. As was discussed in 2.6, the assumption used in the calculation should have led to some error for larger particles such as $m = 10^{-8}$ gm very near the sea surface. But this error presumably has a tendency to lead to smaller values of θ_0 or F_0, because, very near the sea surface, more droplets must be produced and more droplets must fall back to the sea because of their larger sizes due to the lag from the instantaneous response to the environment; consequently, this error does not influence our conclusion. There might be some possibility, however, that the effect of coalescence of particles has also entered to some extent for larger wind speeds, but as long as we use the resulting values of θ_0 or F_0 for an estimate of the geographical distribution of giant sea-salt particles, it is better to accept these values as they are.

Figure 11 shows the wind dependence of F_0 for various sizes of the particles. In this figure, points corresponding to $\theta_l < 5 \times 10^{-3}$ cm^{-3} are excluded. It is very interesting to see that a curious behavior of the curves of wind depend-

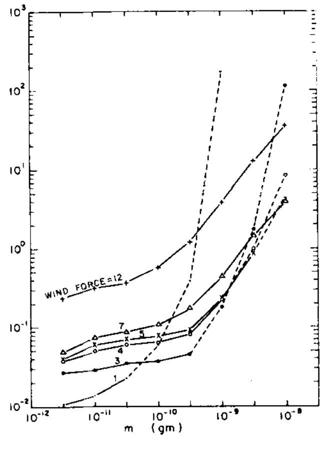

FIG. 10. Estimated production rate of sea water droplets at the sea surface F_0, calculated from Fig. 9. Values are entered for the range of $\log m = 0.5$.

---

[1] According to B. J. MASON (*Nature*, 174, 470, 1954), there are three groups of droplets which are produced by bursting bubbles: (1) jet droplets, which have been studied by rather many investigators, and have a diameter 10–15 % of those of the original bubbles; (2) droplets which have a mass of salt of 5×10^{-10}–2×10^{-12} gm (30 μ–5 μ in diameter) and are ejected at a low angle to the horizon; (3) droplets of 2×10^{-14}–10^{-15} gm of salt. The second group of droplets was also reported by D. M. NEWITT *et al.* (*Trans. Instn. Chem. Engrs.* 32, 244, 1954) for fresh water bubbles, and according to TOBA (1961) the number of these droplets was nearly 50 per cm^2 of the area of the original bubble film cap. K. ISONO's experiment (*Japanese J. Geoph.*, 2, 1, 1959) also showed this group of droplets. Although this second group of droplets seems not to have received much attention, it should be pointed out here that the upper limit of their range (mass of salt of 5×10^{-10} gm) coincides with the position of the trough of curves in Fig. 9.

ence of F_l or θ_l for different values of m in I is now eliminated in Fig. 11, and here also the curves essentially have a single shape.

Since curves in each of these figures have similar shapes, it would be possible to express these values approximately by a more simple way, in other words, by three curves which are representatives of the three figures. As to the concentration, θ_0, and the production rate, F_0, we adopted, for $m \leqslant 10^{-9.5}$ gm, the curves for wind force 4, at which Woodcock's observation is considered to have been performed under a steady state, because it is nearly the mean wind speed; and for $m \geqslant 10^{-9.5}$ gm, the slopes of curves for wind force 7, at which the concentration of larger particles such as $m > 10^{-9.5}$gm must have been subjected to meteorological control at the time as discussed in 2.4 of I. As to wind dependence, we used, as a best representative, the gradient for $m = 10^{-9}$ gm for wind forces greater than 5 and that for $m = 10^{-9.5}$ gm for wind forces smaller than 5. These representative curves are shown in Fig. 12.

Since we have here predicted the sea surface processes for F_0 and θ_0, the results may be compared with an experimental study. Toba (1961) carried out an experiment by use of a wind flume 21.6 m long, and obtained figures of F_0 and θ_0 from an observation of distribution of bubbles in wind waves, droplets in the airflow at a height range of 8 cm to 25 cm, etc. Comparing these results, we may find some common and some differing points between them. The common points are: (1) that there is a trough in the curves which would divide jet drops and film drops; (2) that the shape of the curves essentially remains unaltered when wind speed varies; (3) that the production rate of bubbles increases with bubble diameter up to 0.75 mm; (4) roughly the same orders of magnitude. The discrepancies are: (1) the shape of curves; (2) the position of the trough; (3) the gradient of the curves of the wind dependence. It seems that these discrepancies arose from the facts that Toba's 1961 study was a wind-flume experiment and that tap water was used instead of sea water. The common points, however, seem to be real. When we estimate the distribution of sea-salt particles in the atmosphere, it seems more realistic to use the estimated curves of the present study.

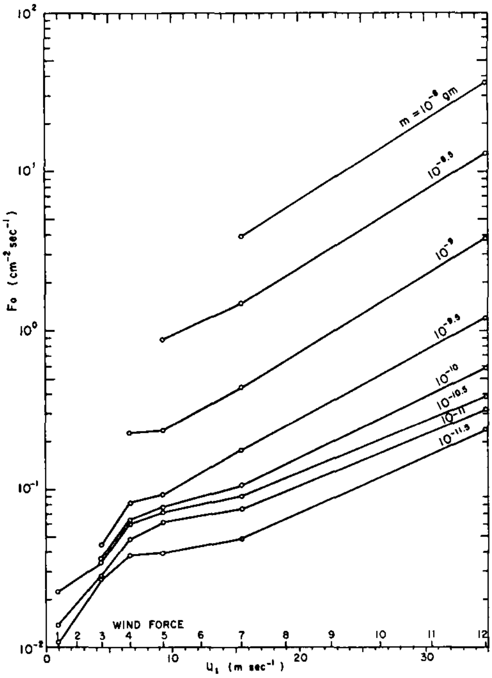

Fig. 11. Wind dependence of the production rate, F_0. Values are entered for the range of $\log m = 0.5$.

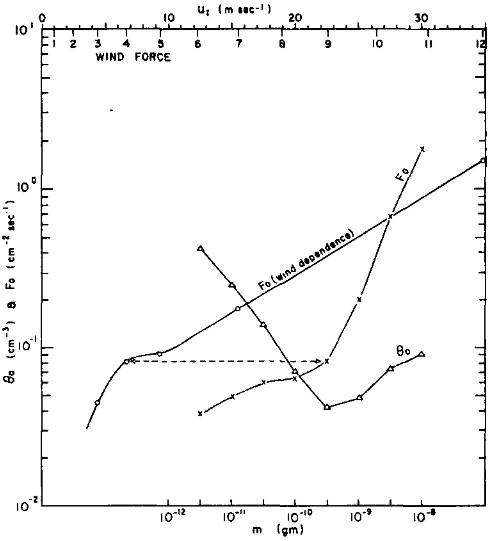

Fig. 12. Representative curves for the production rate, F_0, the concentration, θ_0, and wind dependence of F_0. Values of F_0 and θ_0 are entered for wind force 4, and the values of wind dependence of F_0 are entered for $m = 10^{-9.5}$ gm. Values are also entered for the range of $\log m = 0.5$.

Something should be mentioned here about the shape of the curve of the wind dependence of the production rate. We may see from the figures that there is a sharp rise in F_0 at wind speed below 6 or 7 m sec^{-1} where white caps begin to appear; then the curve becomes flat until the wind speed reaches 10 m sec^{-1} or so, and then it goes up exponentially with wind speed. According to a measurement of the friction factor, γ_l^2 in the wind flume experiment by KUNISHI (1963), the γ_l^2 has a similar trend although the range of the change is very small: it increases with wind speed but has a flat part or rather a small trough at about 12 m sec^{-1} in wind speeds of the 10-m level. Using the same wind flume, TOBA (1961) found that air bubbles became entrained into wind waves by instability of small wavelets of a few to several cm in wavelength, which were superimposed upon the gravity waves, and in which the surface tension exerted a considerable effect, instead of the breaking of gravity waves themselves, when the wind speed at the 10-m level reached 12 or 13 m sec^{-1}. Both investigators thought of this wind speed as one of the critical wind speeds. We could not now conclude, however, that the flatness at about 10 m sec^{-1} in Fig. 12 really corresponds to the above phenomena because there still remains room for suspicion that the critical wind speed found in the wind-flume experiments might correspond to several meters per second on the actual sea surface at which white caps began to appear, in other words, that such white caps that appear at sea at the wind speed of several meters per second are caused by the instability of the small wavelets and not of the gravity waves.

TOBA & TANAKA (1963) made an observation of dry fallout of giant sea-salt particles at several stations along the direction of the winter monsoon flow in Japan, and explained the observed distribution by use of the figure of F_0 of TOBA (1961) and a model distribution. However, it seems that two defects of the too-large gradient in the curve of wind dependence of F_0, and of the model of non-ground-sink of the particle distribution over land, incidentally canceled each other, and consequently that the TOBA & TANAKA estimate of D became smaller than that in I by about two orders of magnitude.

4. Summary

Features of the vertical distribution of giant sea-salt particles in the 10-m layer over the ocean have been analytically studied. The assumptions used are as follows:

(1) A steady state holds and there is no horizontal gradient in wind speed, particle concentration, etc.

(2) The vertical distributions of wind speed, water vapor pressure and temperature obey a logarithmic law.

(3) Transport of sea-salt particles is carried out by vertical eddy diffusion and fall by terminal velocity.

(4) Eddy diffusivity, D, in the atmosphere in the 10-m layer is expressed by equation (5):

$$D = ku_*(z + z_0),$$

consequently, there is no layer of true diffusion next to the sea surface. (This assumption was discussed in 2.1.)

(5) Each droplet instantaneously attains its equilibrium salinity, or equilibrium size, for the vapor pressure and temperature of the surrounding air. (This assumption was examined and in general substantiated in 2.6.)

(6) Sea-salt particles remain as supersaturated saline droplets when the relative humidity of the air is lower than about 75%.

The following preliminary results have been shown:

(1) The effect of vapor pressure and temperature on the equilibrium size of saline droplets may be approximately represented by one factor of relative humidity (2.3).

(2) The relation between the ratio of the terminal velocity of saline droplets to that of sea water droplets of the same mass of salt, w/w_s, and the difference between relative humidity at a level and at the sea surface, f, may be expressed by equations (18) and (19).

(3) Vertical distributions of relative humidity approximately obey a logarithmic law when vapor pressure and temperature obey a logarithmic law (2.4).

From a combination of all these, we have attained equation (47) from which we may see that $\log(\theta/\theta_0)$ is proportional to $m^{2/3}$ and u_l^{-1} with a factor χ, which is a function of f_l, γ_l and z. The value of χ has been calculated for various values of the variables and it has been shown that f_l as well as u_l affects greatly the vertical distribution of sea-salt particles in the 10-m layer, and also that, when the vertical distribution of the number concentration of sea-salt particles of each class interval of mass of salt is plotted on a logarithmic diagram, the result is an almost straight line, as is shown in Figs. 5, 6 and 7. When the results have been applied to the data of the concentration, θ_l and the effective production rate, F_l, of sea-salt particles at the 10-m level obtained in the previous paper I, the complicated features in θ_l or F_l have disappeared to give a very simple pattern of the production rate, F_0, or the surface concentration of sea water droplets, θ_0, as a function of wind force or of mass of salt, as a predicted sea-surface process (Fig. 12). This result gives some agreement with the author's 1961 wind flume experiment.

In conclusion it should be said that the most plausible rate, in the present situation, of production of giant sea-salt particles on the sea surface, which may be expressed as a function of wind speed at the 10-m level and the mass of salt in a particle, has been obtained. Also, the effect of the wind speed, humidity, etc., on the equilibrium distribution of the particles in the layer just above the sea surface, has been predicted, to give a new basis for an estimate of the geographical distribution of the particles over the world ocean.

Acknowledgements

The present study has been performed at the Cloud Physics Laboratory of the University of Chicago under the supervision of Professor Horace R. Byers. The author wishes to express his thanks to Professor Byers for arranging his visit to the United States, his kind advice, suggestions, and discussion throughout this study. Special thanks are due to Professor Shōitirō Hayami of the Geophysical Institute, Kyoto University, Japan, who permitted the author to visit the United States on leave of absence, and who has always given encouragement.

List of symbols

A, B, and C: constants in Dupré-Rankine formula (B: degree^{-1})

a, b, and c: constants in modified Dupré-Rankine formula (b: degree^{-1})

C_p: specific heat of water at constant pressure (cal gm^{-1} degree^{-1})

D: eddy diffusivity in the atmosphere (cm^2 sec^{-1})

D': diffusivity of water vapor in air (cm^2 sec^{-1})

E: evaporation from the sea surface (gm cm^{-2} sec^{-1} unless specified)

F_0: production rate of sea-salt particles (cm^{-2} sec^{-1})

$f = \mathrm{RH}(S_s) - \mathrm{RH} = 98.23 - \mathrm{RH}$ (%)

$G = \dfrac{R_B L}{C_p} \cdot \dfrac{0.622}{P}$ (cm^2 dyne^{-1} degree)

g: acceleration of gravity (cm sec^{-2})

$H = \dfrac{pE}{0.622\, \varrho_a k u_*}$ (dyne cm^{-2})

K: Jacobs' evaporation factor defined by equation (10)

k: von Kármán's universal turbulence constant ($= 0.4$)

L: latent heat of evaporation of water (cal gm^{-1})

$M = \log e \approx 0.4343$

m: mass of salt contained in sea-salt particles or droplets (gm)

p: atmospheric pressure (dyne cm^{-2})

Q: vertical transport of heat (cal cm^{-2} sec^{-1})

q: specific humidity

$R_B = Q/EL$: Bowen ratio

r: radius of saline droplets (cm)

RH: relative humidity (%)

S: salinity of saline droplets (‰)

T: absolute temperature of air (°K)

t: time (sec)

u: wind speed (cm sec^{-1} unless specified)

$u_* = \sqrt{\dfrac{\tau_0}{\varrho_a}}$: friction velocity (cm sec^{-1})

V: volume of a droplet (cm^3)

w: terminal velocity of fall of sea-salt particles (cm sec^{-1})

$y = \ln(z/z_0)$

z: height above the sea surface (cm)

z_0: roughness length of the sea surface (cm)

α_1, β_1, ε_1, α_2, β_2, and ε_2: constants in equations (18) and (19)

Γ_l: evaporation coefficient defined by equation (6)

γ_l^2: friction factor defined by equation (2)

$\zeta = \log \eta$

$\eta = \dfrac{\gamma_l f_l}{k}$

Θ: potential temperature of air (°K)

θ: concentration of sea-salt particles (cm^{-3})

$\Lambda = w_s \cdot m^{-2/3} = 4.3 \times 10^6$ (gm$^{-2/3}$ cm sec^{-1})

ν: kinematic viscosity of air (cm^2 sec^{-1})

$\xi_1 = 1.77/\eta$

$\xi_2 = \xi_1 + y$

ϱ: density of sea-salt solution (gm cm^{-3})

ϱ_w: density of pure water (gm cm^{-3})

ϱ_a: density of air (gm cm^{-3})

σ: water vapor-pressure (dyne cm^{-2} unless specified)

σ_w: saturation water vapor-pressure (dyne cm^{-2} or mb: specified in the text)

τ_0: sheering stress by wind over sea water (dyne cm^{-2})

$\varphi_1 = \alpha_1 - \beta_1 \zeta + \varepsilon_1 \zeta^2$

$\varphi_2 = \alpha_2 - \beta_2 \zeta + \varepsilon_2 \zeta^2$

χ: a factor appearing in equation (47)

$\psi_1 = 2\varepsilon_1 \zeta - \beta_1$

$\psi_2 = 2\varepsilon_2 \zeta - \beta_2$

subscript 0: values at the sea surface

subscript l: values at the 10-m level

subscript s: values for sea water droplets of salinity 35 ‰

REFERENCES

Arons, A. B., and Kientzler, C. F., 1954, Vapor pressure of sea-salt solutions. *Trans. Amer. Geophys. Union*, **35**, pp. 722–728.

Blanchard, D. C., 1963, The electrification of the atmosphere by particles from bubbles in the sea. *Progress in Oceanography, vol.* **1**, New York, The Macmillan Company, pp. 71–202.

Brocks, K., 1955, Wasserdampfschichtung über dem Meer und "Rauhigkeit" der Meeresoberfläche. *Arch. Meteor. Geophys. Biokl. Ser. A. Meteor. Geophys.* 8, pp. 354–383.

Brocks, K., 1959, Measurements of wind profiles over the sea and the drag at the sea surface. *International Oceanogr. Congress Preprints*, Wash. D.C., Amer. Assoc. Adv. Sci., pp. 742–744.

Deacon, E. L., Sheppard, P. A., and Webb, E. K., 1956, Wind profiles over the sea and the drag at the sea surface. *Austral. J. Phys.*, **9**, pp. 511–541.

Defant, A., 1961, *Physical Oceanography*, vol. **1**, Oxford, New York, Pergamon Press, p. 141.

Eriksson, E., 1959, The yearly circulation of chloride and sulfur in nature; meteorological, geochemical and pedological implications, Part I. *Tellus* **11**, pp. 375–403.

Fleagle, R. G., Deardorff, J. W., and Badgley, F. I., 1958, Vertical distribution of wind speed, temperature, and humidity above a water surface. *J. Marine Res.* **17**, pp. 141–157.

Higashi, K., Nakamura, I., and Hara, R., 1931, The specific gravities and the vapour pressures of the concentrated sea water at 0–175°C. *J. Soc. Chem. Ind. Japan*, **34**, pp. 166–172.

Jacobs, W. C., 1942, On the energy exchange between sea and atmosphere. *J. Marine Res.*, **5**, pp. 37–66.

Jacobs, W. C., 1943, Sources of atmospheric heat and moisture over the North Pacific and North Atlantic Oceans. *Ann. N.Y. Acad. Sci.*, **44**, pp. 19–40.

Jacobs, W. C., 1951, The energy exchange between sea and atmosphere and some of its consequences. *Bull. Scripps Instn. Oceanogr. Univ. Calif. LaJolla, Calif.*, **6**, pp. 27–122.

Juliusburger, P., 1900, Über das Dupré-Rankine'sche Dampfspannungsgesetz. *Ann. d. Physik.* vierte Folge, bd. **3**, pp. 618–659.

Kaye, G. W. C., and Laby, T. H., 1948, *Tables of Physical and Chemical Constants*, London, New York, Toronto, Longmans Green, p. 54.

Kunishi, H., 1963, An experimental study on the generation and growth of wind waves. *Disaster Prevention Res. Inst. Bull.* **61**, Kyoto, Kyoto Univ. pp. 1–41.

Montgomery, R. B., 1940, Observations of vertical humidity distribution above the ocean surface and their relation to evaporation. *Pap. Phys. Ocean. Meteor. Mass. Inst. Tech. Woods Hole Ocean. Inst.* **7**, (4), pp. 1–30.

Munk, W. H., 1947, A critical wind speed for air-sea boundary processes. *J. Marine Res.* **6**, pp. 203–218.

Okuda, S., and Hayami, S., 1959, Experiments on evaporation from wavy water surface. *Rec. Oceanogr. Works Japan*, **5**, pp. 6–13.

Rossby, C.-G., 1936, On the momentum transfer at the sea surface. I. *Pap. Phys. Ocean. Meteor. Mass. Inst. Tech. Woods Hole Ocean. Inst.* **4** (3), pp. 1–20.

Sverdrup, H. U., 1946, The humidity gradient over the sea surface. *J. Met.* **3**, pp. 1–8.

Sverdrup, H. U., 1951, Evaporation from the oceans. *Compendium of Met.*, Boston, Amer. Met. Soc., pp. 1071–1081.

Toba, Y., 1961, Drop production by bursting of air bubbles on the sea surface (III). Study by use of a wind flume. *Memoirs Coll. Sci., Univ. Kyoto, Ser. A* **29**, pp. 313–344.

Toba, Y., 1965, On the giant sea-salt particles in the atmosphere. I. General features of the distribution. *Tellus*, **17**. pp. 131–145.

Toba, Y., and Tanaka, M., 1963, Study on dry fallout and its distribution of giant sea-salt nuclei in Japan. *J. Met. Soc. Japan, Ser. II* **41**, pp. 135–144.

Wilson, B. W., 1960, Note on surface wind stress over water at low and high wind speeds. *J. Geophys. Res.*, **65**, pp. 3377–3382.

Woodcock, A. H., 1953, Salt nuclei in marine air as a function of altitude and wind force. *J. Met.* **10**, pp. 362–371.

Wüst, G., 1937, Temperatur- und Dampfdruckgefälle in den untersten Metern über der Meeresoberfläche. *Meteor. Z.*, **54**, pp. 4–9.