Papers

Synoptic variability of characteristics of aerosol formation factors in the Norwegian Sea*

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Atmosphere – sea interaction
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Sea-spray
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Abstract

Synoptic fluctuations of the activity of oceanic regions in the quasi-meridional profile (within *ca* 2°–15°E) in the sector from the Norwegian Sea to about 76°N have been examined. The work aimed at the problem of evaluation of changes in the climatic activity of the ocean on the basis of characteristics determining the oceanic surface whitecap coverage (W). Quantitative estimation of space-time changes of W over a period July–August 1987 and the related fluctuations of characteristics of aerosol water and heat exchange between whitecaps and the atmosphere are presented. The results are discussed and verified on the basis of average climatic characteristics of the region and the contribution of the Arctic front in their formation.

1. Introduction

Among a number of physical characteristics of the atmosphere – sea (A – S) interaction, the wind speed (U), and waving are of fundamental significance from standpoint of aerosol-forming activity of the ocean. Intensity of the activity is evidenced particularly by the degree of oceanic whitecap (W) coverage. On the basis of a few results obtained so far in the Arctic (Lannefors *et al*, 1983; Marks, 1987) and particularly in the Antarctic convergence zone (Garbalewski and Marks, 1985) it is believed that the polar oceanic regions are characterized by considerable variability of wind and aerosol characteristics, unencountered elsewhere on the ocean. These conditions may indicate broad

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possibilities of verification of mathematic models of the effect of interaction A – S on the short-term global fluctuations of climate and the role of polar oceanic regions in climate formation.

The dependence of W on U was investigated by Monahan and O'Muircheartaigh (1980) and Toba and Koga (1986). Earlier Monahan (1971), as well as Toba and Chaen (1973), noticed at the same time the effect of conditions of thermal equilibrium in near water air on the characteristic of W. Simultaneously, the dependence between W and the concentration of marine aerosol in near water air was examined (Monahan *et al*, 1983; Wu, 1986). It was established (Monahan *et al*, 1983) that a positive correlation between the concentration of particles carried out from the sea and the parameter W increases with an increase in droplet radius. When examining the dependence of sea coverage with foam on the wind velocity, Bortkovskiy (1983) observed a significant effect of sea temperature on W. The effect of changes in thermal equilibrium in near water air on the fluctuations of W was also noted (Monahan and O'Muircheartaigh, 1986).

Spillane *et al* (1985) computed monthly isopleths of oceanic whitecap coverage. Garbalewski (1987) calculated monthly isolines of an aerosol emission flux (F_0) from the ocean and made a preliminary comparison of active oceanic fields obtained by Spillane *et al* (1985) from computations of W with oceanic centers of an increased aerosol yield. The comparison aimed at recognition of possibility of utilization of any of the mentioned characteristics (W or F_0) as a criterion estimating the aerosol-forming and climatic activity of the ocean. At the same time Garbalewski and Marks (1987) and Marks (1987), paying attention to a considerable fraction of low-level advection of sea salt particles over the oceanic regions, partly argued usefulness of the characteristic F_0 in estimating climatic activity of specific ocean regions. The quantity F_0 is verified by measurements of aerosol concentration in near water air and, in a majority of cases, it cannot be a characteristics reflecting solely local activity of the ocean. The presence of the advective component also enables occurrence of the effect of distant oceanic regions.

In view of the above statements and determinations it seems appropriate to employ a complementary approach to the problem of synoptic variability of energy-active oceanic regions. Hence, it should be tried in the research undertaken to combine a method of utilization of synoptic aerosol characteristics with a simultaneous determination of activity centers basing on characteristics of variability of whitecap fields in the ocean. It may be anticipated that the parameter W allows to determine unambiguously the energy-active oceanic zones, without fear of disturbing the characteristics obtained in this manner by factors external with respect to the examined regions. The present research constitutes the first attempt at estimating short-term fluctuations of energetic activity of the ocean on the basis of changes in the parameter W on a synoptic scale.

2. Theoretical model of dependence of W on wind

Analysis of the results of studies by Monahan (1969) and Toba and Chaen (1973) allowed Wu (1979) to confirm validity of application of a power-law in a theoretical model of $W(U)$:

$$W = \alpha \cdot U^{\lambda}, \tag{1}$$

where U is the wind speed measured at an altitude of about 10 m above sea level. Earlier Blanchard also formulated the law by approximating an increase of W with the square of wind speed and taking the values of coefficients as: $\alpha = 4.4 \cdot 10^{-4}$, $\lambda = 2$ for the wind speed $U > 5$ m·s^{-1} (when $U < 3$ m·s^{-1}, $W = 0$). Further progress was achieved due to later laboratory experiments (Monahan, 1969, 1971). At the same time it was assumed (Cardone, 1969) that the whitecap coverage of the sea is directly proportional to the amount of energy being transferred from the wind flow to the fully developed spectral components of waving. As a result of such an approach the expression $W(U)$ applicable in the case of fully developed fresh water sea was obtained. However, a disadvantage of laboratory experiments turned out to be the fact that the whitecap lifetime under conditions of fresh water sea is shorter when compared with natural conditions in the ocean. This derives from the effect of salinity and associated with it conditions of water viscosity on W.

Further specification of the model was achieved by Wu (1979), who obtained $\alpha = 2.00 \cdot 10^{-6}$ and $\lambda = 3.75$ by accounting for the dependence of wind velocity on the shear stress and surface drift of water, *ie* on the quantity of work being done by the wind *per* unit surface of the sea. Finally, basing on the data collected by Monahan (1971) and Wu (1979) on the Atlantic and by Toba and Chaen (1973) on the Pacific and using a least squares procedure the dependence was obtained in the following form:

$$W = 2.95 \cdot 10^{-6}\, U^{3.52}. \tag{2}$$

This model of $W(U)$ was found to be the best one by verification based on the data on W (90 observations in total) collected by photographing whitecaps on the ocean. The model permitted to obtain prediction of W with the relatively smallest probable average error ($4.68 \cdot 10^{-5}$), as established in the paper by Monahan and O'Muircheartaigh (1980).

The dependence of W on the wind described by equation (2) should be treated as fundamental, which should undergo further specification. Namely, a modifying effect of changes in sea temperature and the related state of low-tropospheric equilibrium in the atmosphere on the value of W should also be taken into account. This is the dependence on $\Delta T = (T_w - T_a)$, *ie* the difference between temperature of surface sea water and temperature of near water air, measured at a height of the ship's deck *ca* 10 m above sea level. The effect was studied by Wu (1979), Monahan and O'Muircheartaigh (1980), and others. According to Blanchard, the effect is associated, among others, with the influence of water temperature on the speed of rising of air bubbles in water column (Monahan and O'Muircheartaigh, 1980). A decrease of kinematic viscosity of water with an increase in temperature (Monahan and O'Muircheartaigh, 1986) and the effect of ΔT on the changes in wind speed are also significant. Hence, these effects should indirectly affect the whitecap lifetime and, thus, their coverage of the sea surface (Monahan and Zietlow, 1969). The results also indicate (Bortkovskiy, 1983) that under conditions of the warm and cold sea just the sea temperature T_w alone is capable of causing distinct differences in the quantity of W. All this indicates the need of development of the model of $W_c(U)$ and its reduction to the expression $W(U, T_w, \Delta T)$. To meet this need Monahan and O'Muircheartaigh (1986) arrived at the present stage, on the basis of many investigations, at the result in the form:

$$W = 1.95 \cdot 10^{-5}\, U^{2.55} \exp(0.086/\Delta T). \tag{3}$$

At the moment, this is one of the best solutions of this problem, although it does not fully describe the complex relation and neglects the effect of T_w.

Returning to the main factor, which is U, it should be noted that already Munk (1947) tried to determine the critical wind speed U_B for the appearance of first whitecaps in the sea. This was to be a speed of 7 $m \cdot s^{-1}$. Subsequently, U_B has been studied by a number of authors (Monahan, 1971; Toba and Chaen, 1973; Monahan *et al*, 1983; Zheng *et al*, 1983) who established that U_B depends on T_w and drops below 4 $m \cdot s^{-1}$ under conditions of the warm sea (17.4 – 30.6°C), increasing to about 7 $m \cdot s^{-1}$ for the cold seas. On the basis of statistical analysis of numerous observations in various oceanic regions made during a number of international experiments, Monahan and O'Muircheartaigh (1986) derived the following relationships for $U_B(T_w)$ and $U_B(\Delta T)$:

$$U_B = 3.36 \times 10^{-0.00309 T_w} [m \cdot s^{-1}], \tag{4}$$

$$U_B = 3.27 \times 10^{-0.0458 \Delta T} [m \cdot s^{-1}], \tag{5}$$

where: $\Delta T_w = T_a$ [°C]. It was also found that under conditions of neutral atmospheric equilibrium in near water air the value $U_B = 3.24$ $m \cdot s^{-1}$ for $\Delta T \approx 0.1$°C (with a deviation up to 2 – 3 $m \cdot s^{-1}$ at unstable and 5 – 6 $m \cdot s^{-1}$ at stable equilibrium).

The above variability of the limiting value of U_B and its dependence on T_w and ΔT were obtained using observation data collected for fixed periods of a day and then averaged for separate, still different physical conditions. Some of the data were obtained for artificial conditions in laboratory experiments. This enabled verification and specification of relationships derived from observations at sea. The relationship proposed should be then completely satisfactory for the periods specified above. It is not known whether and to what degree the principle in the above form is satisfied in the case of space-time changes of W and the conditions determining it, including those described by equations (4) and (5) for fluctuations on a synoptic scale. This problem should be taken into consideration in the studies of synoptic variability of aerosol mass and heat exchange, conducted on the basis of several day averages and monthly observation data. Carrying out profile studies in the polar region, along a sufficiently long sector of a meridian, is very favourable for elucidation of the details, especially those related to temperature changes of surface water of the ocean.

Except of parameters U, ΔT and T_w, variability of wind characteristics in the layer of A–S interactions should also influence considerably the fluctuations of W. In the polar zone winds over the ocean are strong but of short duration. However short, the period of wind intensification followed by the period of its attenuation can be distinguished, the former increasing W. So far, the dependence of the 'patch' growth rate and the whitecap coverage on the duration and frequency of wind intensification periods has not been established. The extent of whitecap patch on the ocean can be assumed to be a function of several factors, and at least:

$$X_w(W, t_u, N, l), \tag{6}$$

where:

t_u, N – average duration of wind intensification and its frequency at time t, respectively,

l – activity extent of intensifying wind.

For polar regions, the occurrence of especially gusty winds and often predomination of long-term squalls *etc* are typical. These winds increase roughness of sea surface and efficiency of aeorosol formation (Garbalewski, Marks, 1985). Wind intensifications are short for gusts, and those accompanying particular gusts affect relatively small, limited parts of sea surface. However, regions of gustiness occurrence of considerable frequency are wide in polar zones. This justifies the necessity to take wind gustiness into consideration in the synoptic studies of the dynamics of aerosol formation characteristics of the ocean, especially characteristics of W.

The factor describing the gustiness of the wind can be calculated (for ΔT = const.) from variability of wind directions and the module of its speed in relation to the speed U_B. Then:

$$k_1 = \frac{n_\varphi}{N_\varphi}, \tag{7}$$

where: n_φ is the number of different wind directions during occurrence of extrems of W, and N_φ is the total number of wind direction notes during the whole period of observation.

The next factor is

$$k_2 = \frac{U}{U_B}, \tag{8}$$

and hence the gustiness factor is presented as

$$g_u = k_1 \cdot k_2. \tag{9}$$

On a synoptic scale, the extremes W are understood as sectors of increase or decrease in whitecap coverage until turning points of W_c values on the cross-section of whitecap coverage patch are reached. The approach proposed has some weak, unascertained aspects (extension of gust notion for the sum of gusts over synoptic periods, estimation of g_U with respect to U_B *etc*). However, it seems that in the scope of studies of synoptic variability, the above assumptions can, at the present state of knowledge, be the clue to further specifications.

3. Methodics

A series of methods were applied during the sruise of r/v 'Oceania' to Artic regions (Fig. 1), beginning from routine estimations of power, velocity and direction of wind to visual observations. In the case of wind, particular attention was paid to minimization of measuring errors despite the known difficulties with determination of its accurate characteristics on a vessel. The measurements were carried out on the windward side of the board using a manual cup anemometer. A majority of observations were carried out at measuring stations on the ocean, where the vessel was drifting. Routine observations consisting in determination of the wind force in a Beaufort's scale were carried out during sailing (SHIP observations).

Among a group of thermal characteristics, determinations of sea surface and air temperature were carried out. The $\Delta T = T_w - T_a$ characteristics were accurately determined at measuring stations on the basis of measurements of air temperature carried out on board (at *ca* 6 m asl) and parallel measurements of surface water temperature.

Whitecaps were recorded using a camera equipped with a standard objective ($f = 50$ m) and a UT 21 colour reversible film. Photographs of the foam covered sea surface were taken from the level of the upper deck, *ie ca* 6 m above mean sea level. A single recording involved taking a series of at least 9 pictures in few minute intervals. The largest possible part of sea surface was enclosed with the horizon visible at the edge of a picture. Single pictures were taken rarely, *eg* during storm, when the number of whitecaps was extremely high.

Altogether, data on very diversified anemohydrodynamic conditions characteristic of various stages of wind waving development, and hence also a various degree of sea surface coverage with foam, were acquired. Attention was also paid to typical situations of development of waving (starting from tiny waves to rough surface covered to a large extent with foam) by taking subsequent series of sea surface pictures.

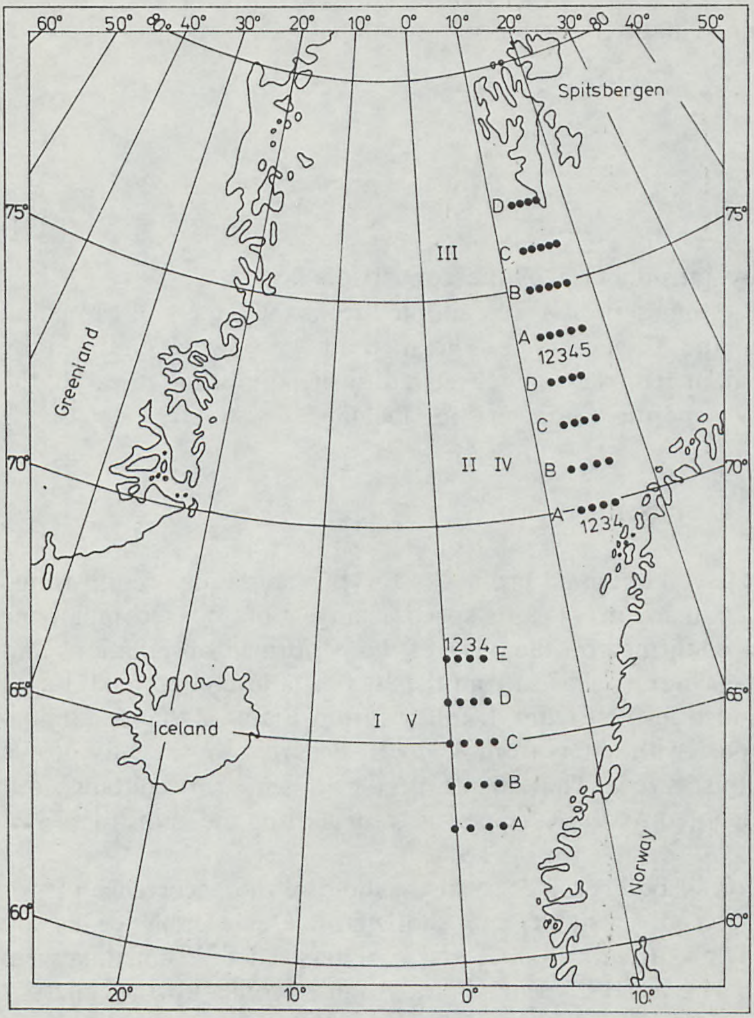

Fig. 1. Map of location of measuring sites in northern regions of the Atlantic included in the polar experiment I, II, III – at the stations designated with dots on the way to Spitsbergen; IV, V – on return trip

Twenty four successful and technically correct recordings were obtained in such a manner. Individual pictures from each series were subsequently subjected to a special optical analysis. Data processing consisted in calculation of visible individual whitecaps or traces in the form of vanishing foam patches. The average number of witecaps $\bar{N}$ was determined on the basis of the results of this analysis for each series of pictures. The value of $[\bar{N}]$, *ie* the average of $\bar{N}$ means for the entire two months' observation period was subsequently calculated and adopted as preliminary July – August characteristics of the examined Arctic regions. The value of $[\bar{N}] = 20.7\ \bar{s}^{-1}$, where $\bar{s}$ is an average sector of sea surface enclosed in a picture, was obtained in such a way. On the basis of the obtained data the value of

$$W_c = \frac{\bar{N}}{[\bar{N}]}, \tag{10}$$

was calculated, and

$$W = k \cdot W_c, \tag{11}$$

was determined by introduction of the conversion factor k.

The method of estimation of W was adopted from the papers of Monahan *et al*, 1986; Toba and Chaen, 1973. The method was partly modified by introduction of some inevitable changes related to the aim of the research, the need for obtaining synoptic characteristics included.

4. Results and Discussion

The obtained quasi-meridional profiles of the W_c magnitude are illustrated in Figure 2. A comparison of time-space changes of W_c accompanying synoptical changes confirmed on the basis of the performed observations and analysis of facsimile synoptic charts (from the 1000 hPa level) obtained during this period, is included in the Figure. It follows from Figure 2 that 5 dominating W_c peaks denoted with letters from A_1 to E_1 occurred on the July profile (on the way to Spitsbergen). The August profile (return trip) contains four similar W_c peaks, denoted A_2 to E_2, respectively (neglecting the undistinguished D_2).

Closer inspection of both profiles seems to indicate the occurrence of two types of W_c peaks, *ie* quasi-stable and short-term. Quasi-stable peaks are indicated in the July profile with $B_1 - C_1$ and E_1 letters. These W_c maxima seem to correspond to $B_2 - C_2$ and E_2 peaks in the August profile. Further analysis proves identity of these W_c peaks. On the other hand, the increase of W_c denoted as D_1 turned out to be distinctly ephemeric. It is hardly visible in the August profile.

Further examination (Fig. 2) revealed that E_1 and E_2 peaks were related to atmospheric front. Everything indicates the influence of the Arctic Front in the polar latitudes on the W_c increase. A southward shift of the Arctic Front during the observation period is visible in the examined profile (Fig. 2) and its particular sections (Fig. 3). The shift proceeded with the average rate C_F (Table 1).

The Table 1 lists front shift rates estimated in meridional direction, hence it is on the average a normal to the front surface in the case of the Arctic Front.

Southern peripherals of the Arctic Front extend in summer (July) to 70 – 75°N latitude and in winter (January) move southward reaching 60 – 65°N (Dzerdzeyevskiy, 1945; Khromov, 1948). It is a quasi-stationary front and its meridional shift is relatively slow. The shift rate is related to two components,

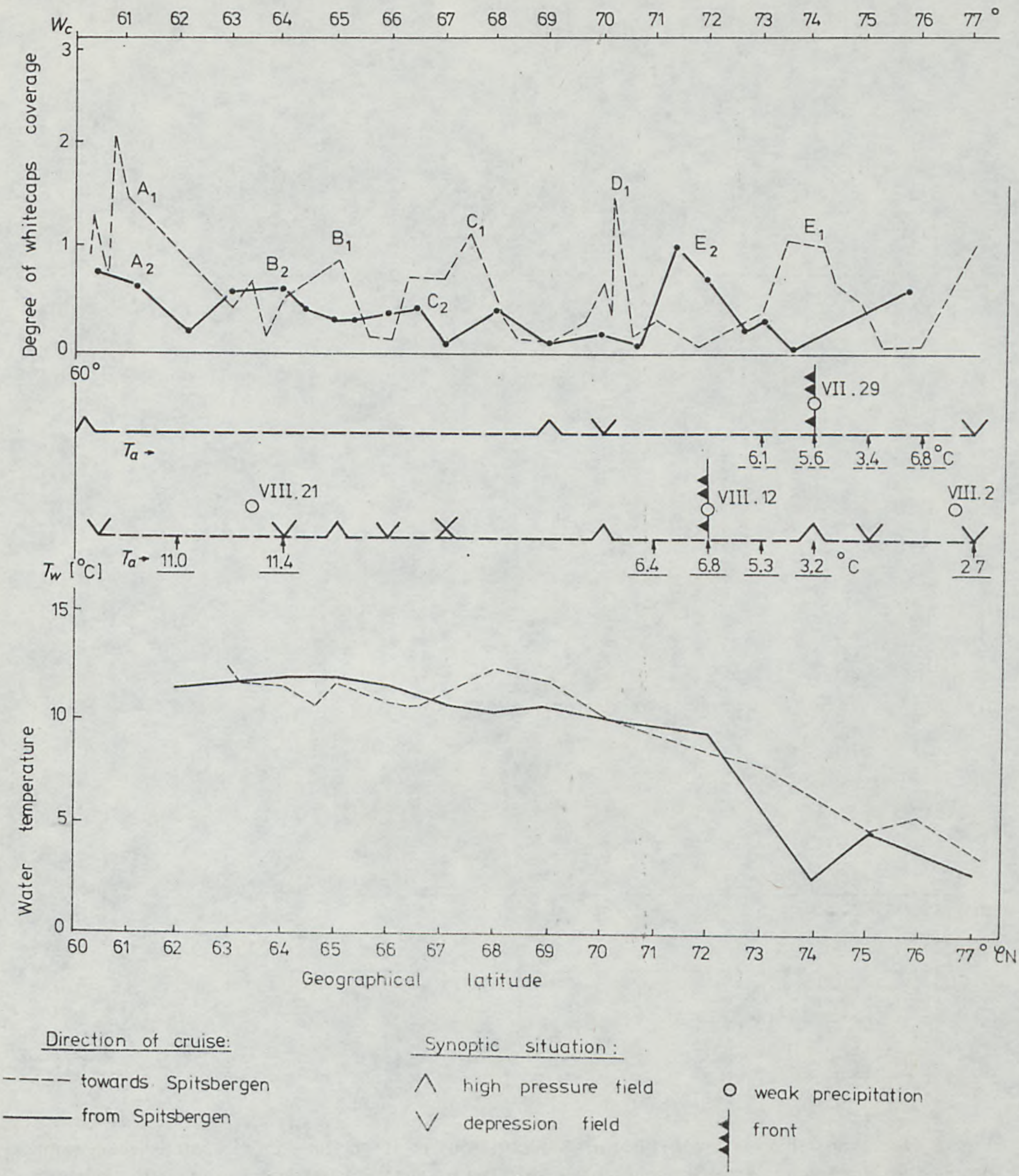

Fig. 2. Quasi-meridional profiles in the belt of latitudes from the Norwegian Sea to Spitsbergen, of normalized whitecap coverage ($W_c = \bar{N}/[\bar{N}]$), and sea surface temperatures T_w obtained in July and August 1987

viz the seasonal front shift and wavy motion of the frontal surface. Cold front velocity related to wavy motion usually varies within the range from 70 to 90% of geostrophic wind velocity, hence it is equal to *ca* 0.5° h^{-1} on the average, *ie* *ca* 50 $km \cdot h^{-1}$.

The above data indicate that the determined rate of movement of the

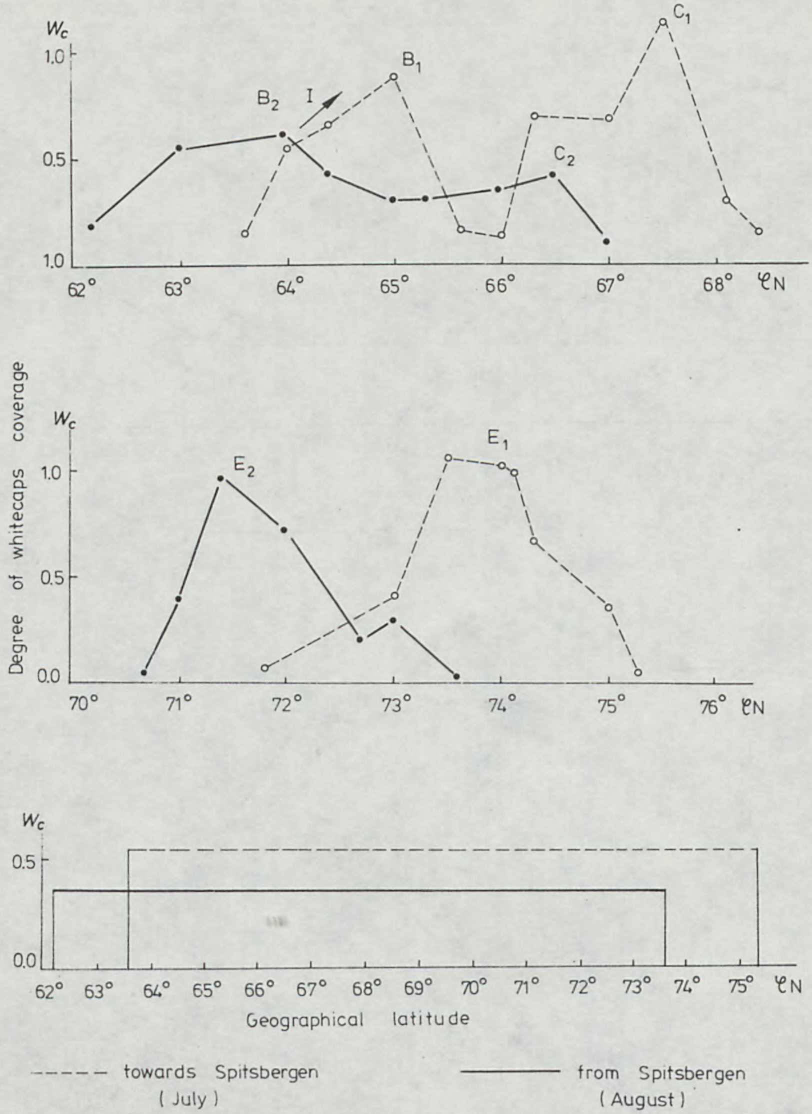

Fig. 3. Distinguishable cycles of space-time fluctuations of W_c in the sectors of quasi-meridional profile the Norwegian Sea – Spitsbergen as well as (in schematic representation at the bottom) general average monthly changes (over a period July – August 1987) of average W_c changes in the whole profile

Table 1. Shift rates of Arctic Front estimated from the transposition of W_c peaks

Section	°φ}	$C_F\,[km \cdot h^{-1}]$
$E_1 - E_2$	74 – 72	0.45
$B_1C_1 - B_2C_2$	66 – 65	0.24
$B_1E_1 - B_2E_2$	(64 – 75) – (62 – 74)	0.29

Atlantic Arctic Front during the July–August period is related to mean climatic southward shift occurring in the latitude belt from *ca* 60 to *ca* 75°N. A typical rate of this movement is equal to *ca* 0.15 $km \cdot h^{-1}$ (Khromov, 1948).

Taking into account deviations from the average climatic value in particular periods, the determined C_F values ranging from 0.24 to 0.45 $km \cdot h^{-1}$ can be treated as generally corresponding to a characteristic rate of seasonal movement of the Arctic Front. However, the fraction of wavy component of the movement rate should also be taken into consideration. It seems that the contribution of waving of the frontal surface in the overall motion can be evidenced by W_c maxima occurring at the forefield of the Arctic Front, *viz* B_1 and C_1 in July and B_2 and C_2 in August (Fig. 3). They allow to put forward a hypothesis on the identity of July and August peaks and on the occurrence in the polar Atlantic region of a single quasi-stationary 'patch' of a distinctly increased whitecap coverage of ocean, caused by the Arctic Front. It is evidently a patch related to the main front, distinguishing itself in Figure 2 by a perceptible decrease in temperature of near water air and by appearance of precipitation, as well as in expected occurrence of secondary fronts being traces of front wavings withdrawing northward. The observed cases of precipitation at 63–64°N latitudes, *ie* at a distance of *ca* 10°N from the main front (Fig. 2) can be explained by the latter effect, the more that this precipitation is not related to such a distinct air temperature decrease in the meridional profile as in the case of the Arctic Front latitudes–*ca* 74°N in July and *ca* 72°N in August. Therefore, it is also possible that an increase of C_F during the observation period could also have been related to wavy motion of a frontal surface in the initial or final stage of its southward movement.

An analysis of time-space W_c variations (Table 2) at active sectors of the

Table 2. Analysis of space-time fluctuations of W_c with time Δt in active sectors of quasi-meridional profile in July–August 1987

Symbols of sectors	Increase in W_c				Decrease in W_c			
	ΔU $[m \cdot s^{-1}]$	ΔT_a [°C]	Δt [h]	ΔW_c	ΔU $[m \cdot s^{-1}]$	ΔT_a [°C]	Δt [h]	ΔW_c
B_1	6.0	0.9	54	0.73	−6.3	−0.8	27	−0.75
C_1	8.3	0.9	51	1.02	−8.5	1.0	28	−1.03
B_2	2.6	1.4	27	0.32	−3.8	−1.1	43	−0.44
C_2	3.9	0.1	6	0.34	−1.3	1.4	38	−0.13
E_1	8.5	0.3	42	0.09	−9.0	−2.4	60	−1.1
E_2	9.0	1.6	48	0.96	−8.5	0.6	24	−0.93

profile revealed a distinct dependence of this parameter on ΔU. An increase of W_c was observed solely in cases of positive values of ΔU corresponding to time intervals of the ship's route, and conversely–a decrease of U (negative ΔU) was usually accompanied by a decrease of W_c. This could already be predicted from theoretical assumptions, since the dependence of the degree of coverage of a sea with whitecaps on U is known and was found to have the following form in our studies (Fig. 4):

$$W_c = 24.0506 \cdot 10^{-3}\, U^{1.60932} \tag{12}$$

with residual analysis error equal to 4.58717.

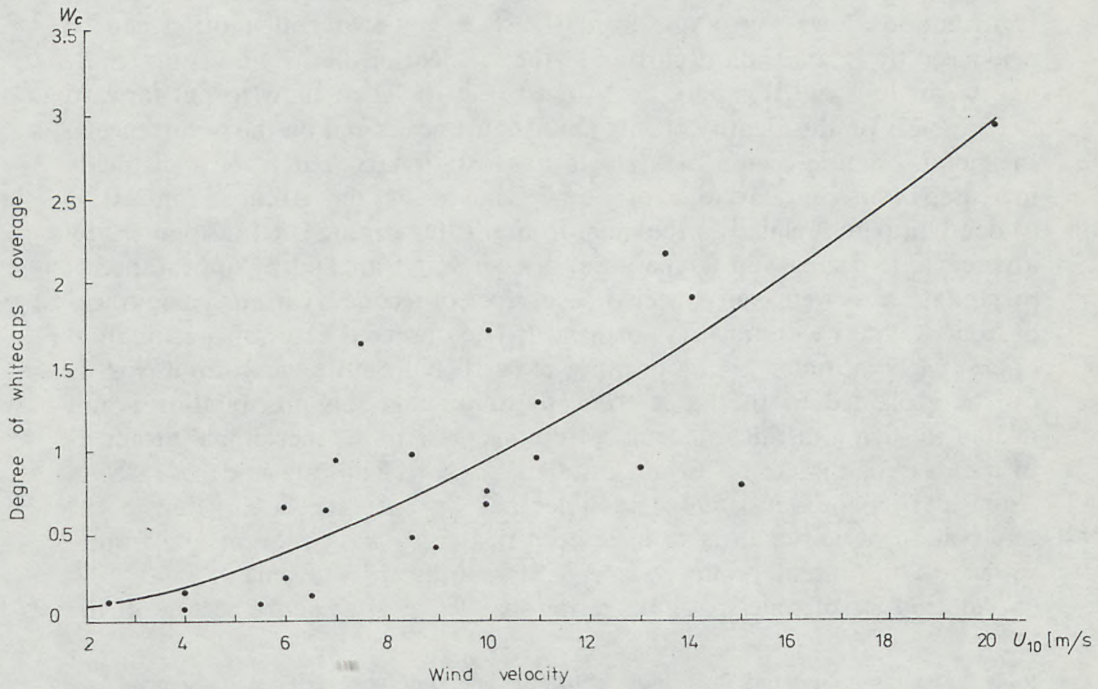

Fig. 4. Exponential relationship $W_c(U_{10})$ derived from measurements at I–V measuring sites

The above results were supplemented with investigations on the influence of conditions of wind gustiness at the segments of occurrence of W_c extrema on the ocean. Table 3 presents the values of the g_U coefficient for the regions of increase and decrease of W_c in a profile crossing polar latitudes, estimated according to equation (9). It follows from these data that W_c maxima at the regions of wind velocity increase should be influenced also by the increased wind gustiness.

On the other hand, no distinct effect of water temperature on W_c as depending on whitecap frequency was observed. This effect is considerable when relative to W measured in a small scale, and particularly laboratory conditions (Bortkovskiy, 1983). The statement that no effect of T_w on W_c occurs in synoptic scale seems to be true, since no tendency of W_c to decrease in the

Table 3. Average values characterizing conditions of wind gustiness in the sectors of occurrence of W_c extremes in the Norwegian Sea – Spitsbergen profile

Extremes of W_c	U [$m \cdot s^{-1}$]	U_B [$m \cdot s^{-1}$]	k_1	k_2	g_U
Decrease in W_c to minimum	3.65	2.85	0.88	1.28	1.13
Increase in W_c to maximum	10.85	3.04	0.97	3.57	3.46

60 – 75°N profile (moving north) can be observed. This fact is confirmed by comparison of $\Delta W_c/\Delta t$ changes accompanying $\Delta T_w/\Delta t$ variations and variations of wind acceleration $\Delta U/\Delta t$ (Fig. 5A, B) performed in the paper. The following relationship was obtained for the distinct effect of $\Delta U/\Delta t$:

$$\frac{\Delta W_c}{\Delta t} = 13.556 \cdot 10^{-2} \frac{\Delta U}{\Delta t}. \tag{13}$$

Donelan and Pierson's (1985) conclusion on the effect of a decrease of W (accompanying a decrease of T_w), due to an increase of viscous energy dissipation in the sea at the expense of a decrease of energy dissipation through breaking of waves, should be no concern of whitecap frequency, and in synoptic scale rather valid at larger T_w differences, hence for latitude profiles longer than those studied in this paper.

On the other hand, a slight influence of the difference between water and air temperature ($\Delta T = T_w - T_a$) on W_c was observed. This influence is related to changes in conditions of local thermal equilibrium in lower layers of atmosphere accompanying changes of ΔT. Such an effect was already observed (Monahan, O'Muircheartaigh, 1986). The analysis of this influence was carried out according to a principle applied in a paper by Spillane *et al* (1986). The following classification was carried out: conditions when $T_w < 12.5$°C – cold sea; $12.5 < T_w < 14.0$°C – temperate sea; $T_w > 14.0$°C – warm sea; and respectively, when $\Delta T < -0.4$°C – stable equilibrium; $-0.4 < \Delta T < 0.6$°C – neutral equilibrium; $\Delta T > 0.6$°C – unstable equilibrium. The results of such an analysis prove, however, that the occurrence of ΔT differences on long profile sectors would be necessary for the effect of ΔT to be more noticeable on a synoptic scale. Profile characteristics obtained by us correspond rather to neutral equilibrium conditions, typical for remote ocean regions (Monahan, O'Muircheartaigh, 1986). Hence, it is difficult to find larger ΔT differences in the investigated profile. The above statement seems to be true, particularly in view of the dominating role of wind velocity in W_c forming on a synoptic scale.

Synoptic W_c variations found in this study constituted a basis of sensible heat calculations, of mean approximately monthly (comprising not full monthly observation periods, but 70 – 80% of them) characteristics of aerosol heat and moisture exchange through oceanic whitecaps (Table 4). The value of

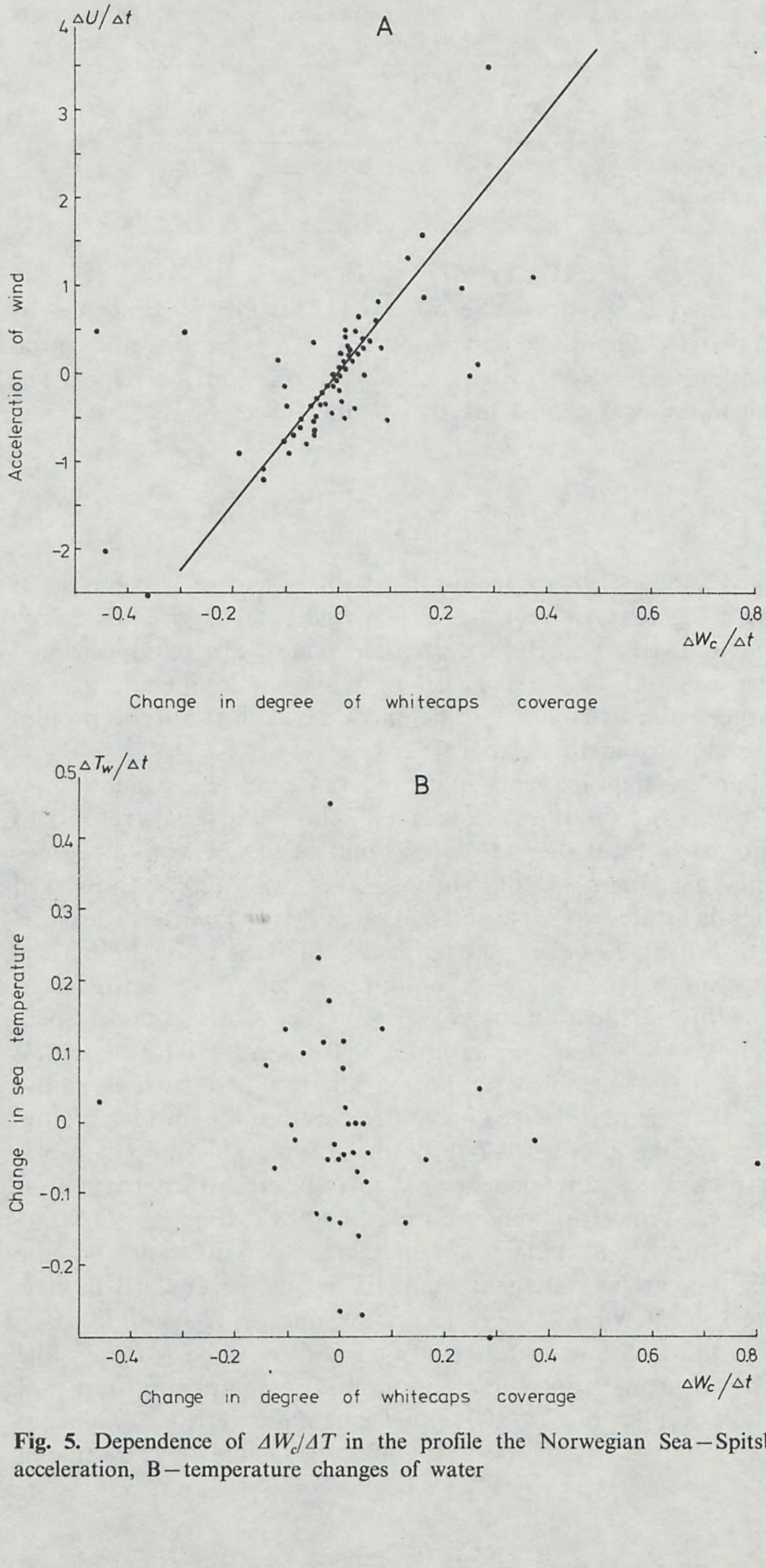

Fig. 5. Dependence of $\Delta W_c/\Delta T$ in the profile the Norwegian Sea – Spitsbergen on A – wind acceleration, B – temperature changes of water

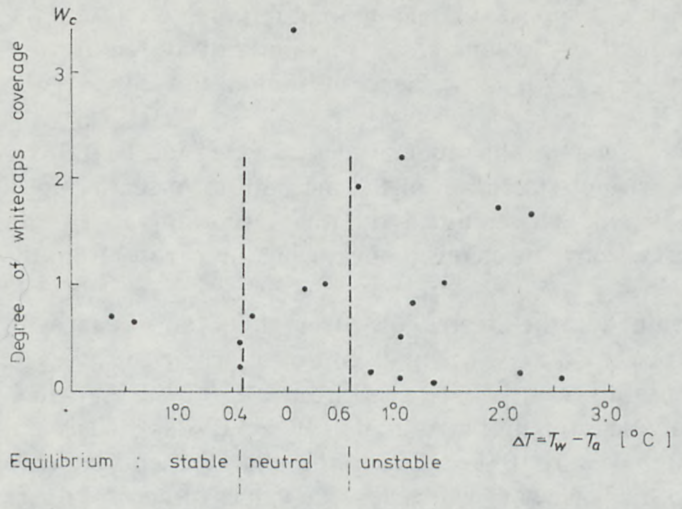

Fig. 6. Dependence of W_c on the conditions of local atmospheric equilibrium estimated on the basis of $\Delta T = T_w - T_a$

F_c and H_c obtained for the observation period, *ie* water and heat fluxes carried by marine aerosol in the investigated profile, were compared with data on mean climatic F_w- and H_w-aerosol fluxes of water and heat, respectively, obtained for the world ocean in the 60–70°N belt by Spillane *et al* (1986). It should be noticed that Spillane *et al* (1986) calculated isolines of W on the basis

Table 4. Characteristics of aerosol water [$\mu g \cdot m^{-2} \cdot s^{-1}$] and heat [$mcal \cdot m^{-2} \cdot s^{-1}$] exchange between whitecaps and the atmosphere in the Norwegian Sea–Spitsbergen profile

Period of 1987	°φN	T_w [°C]	W [%]	Profile averages		Climatic averages*	
				F_c	H_c	F_w	H_w
July	60°15′–77°00′	9.2	1.05	11.8	0.108	8.9	0.082
August	77°00′–60°30′	9.8	0.64	7.3	0.071	8.8	0.086

*Fluxes F_w and H_w were calculated on the basis of climatological estimations of the ocean in the cross-section 70°N–70°S (Spillane *et al*, 1983) with regard to whitecap coverage and lifting spray from the ocean

of an atlas of mean wind characteristics, while the results in this study were obtained on the basis of direct observations at the ocean. It follows from Table 4 that the calculated climatic averages are generally similar to experimental averages at the profile. Although the latter are fragmentary compared to climatic averages, since they concern July and August of 1987, while isolines of

W calculated by Spillane *et al* (1986) were based on long-term material from the last century (designation used by Spillane *et al*), yet similarity of the average values of sea-spray fluxes (F_c, F_w) is close, hence it should also concern the related heat fluxes (H_c, H_w).

Undoubtedly of great interest are the magnitudes of water and heat fluxes carried out of the ocean with co-operation of the mechanism of scattering of droplets by whitecaps. It should be pointed out that according to current assumptions it is the most effective mechanism of carrying out aerosol from the sea.

The characteristic value of the degree of coverage of an ocean with whitecaps is estimated to be equal to *ca* 1% of its entire surface on the average. In our case the mean monthly value of W varies from 0.64, to 1.05% which corresponds to mean climatic estimations. Assuming $W = 1.0\%$ as a basis and calculating the fluxes for $W = 100.0\%$, *ie* assuming that the entire ocean surface emits droplets with efficiency of whitecaps, the values of aerosol fluxes of water in July and August 1987 equal to 1.12 and 1.14 $\mathrm{mg \cdot m^{-2} \cdot s^{-1}}$, respectively are obtained. It is worthwhile to compare such values with average monthly values of sea-spray fluxes carried out of the real ocean surface. In the regions of the Atlantic Ocean included in the investigated profile, the mean July values of F_0 calculated from long-term climatic data on wind field vary between 0.01 and 0.03 $\mathrm{mg \cdot m^{-2} \cdot s^{-1}}$ (Garbalewski, 1987). Hence, they are fifty times smaller. This comparison, however, seems to confirm the validity of the obtained values, since the ratio of maximum assumed efficiency of the ocean (at $W = 100\%$) to the observed efficiency (at $W = 1\%$) proves that the total efficiency of aerosol production is approximately only two times higher than in the case of the above assumption.

5. Conclusions

Observations on the Norwegian Sea – Spitsbergen profile point out the effect of the Arctic Front on aerosol-forming activity of the ocean. The results of investigations on this effect allow to put forward a hypothesis on the occurrence of a single quasi-stationary 'patch' of distinctly increased degree of coverage of ocean with whitecaps in the Atlantic – polar region of the Arctic Front. It has been established on the basis of investigations on time-space changes of W_c that during the period July – August 1987 the Arctic Front moved from north to south with an average rate equal to *ca* 0.29 $\mathrm{km \cdot h^{-1}}$. This rate is comparable with the average climatic rate of meridional movement of the Arctic Front (*ca* 0.15 $\mathrm{km \cdot h^{-1}}$). A deviation from long-term mean equal to *ca* 0.14 $\mathrm{km \cdot h^{-1}}$ is not strange taking into account that our investigations concerned a particular, randomly chosen period. It is also possible that an increase of the effective velocity C_F could be additionally due to a component related to wavy motion of the frontal surface.

Among the factors normally responsible for variations of W, *viz* $U, \Delta T$, and T_w the decisive significance for the processes taking place on a synoptic scale in the investigated polar region should be attributed to wind acceleration $\Delta U/dt$, including that related to gusts. In contrast to $W(U)$, a linear increase of $\Delta W_c/\Delta t$ is observed with wind acceleration changing from negative to increasing positive values of $\Delta U/\Delta t$. Due to the predominating effect of this factor no significant influence of changes in sea surface temperature reaching even 10°C was observed on the scale of the investigated profile. Larger time-space T_w differences would be necessary to expand this influence on a synoptic scale.

The average monthly values of W obtained for July and August 1987 (1.05 and 0.64%, respectively) can be assumed to correspond to the average climatic estimations (Spillane *et al*, 1986: 0.85 and 0.84% respectively) taking into account that they are only random values. The same concerns comparison of aerosol water and heat fluxes calculated on the basis of our observations at the ocean. These comparisons, being the only ones currently available, confirm validity of the obtained values and their mean monthly changes in the investigated profile.

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