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Sea Level Variations During Ice-Covered Periods in the Baltic Sea

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Abstract

The investigation analyses sea level data from eight stations along the Swedish coast of the Baltic Sea. The time period considered was from 1975 to 1986, and the data were classified into different sea ice categories. The effects of ice on sea level variations were analysed on the basis of statistical and dynamical methods.

The sea level data illustrated a significant variation during the year with large amplitudes in the autumn and early winter. The amplitudes were reduced during midwinter, spring and summer. When the data were collected into different ice classes, it was observed that the amplitudes were reduced during severe ice conditions. The reason was partly meteorological conditions and partly ice.

The sea levels were also expanded into Empirical Orthogonal Functions (EOF). It was shown that not less than 73.7 % of the total variance was due to a general raising or lowering of the whole surface. The second mode added 20.3 %, and these two modes thus explain 94 % of the total variance. The effect of sea ice on the EOF modes was analysed, but it was difficult to find any general trends.

The effects of sea ice on the sea level variations were most easily detected when analysing the sea surface tilting in the Bothnian Bay. It was shown that the ice cover may reduce the piling-up. The data also indicated that periods with quite high ice strengths (order of 10^5 Nm^{-2}) have occurred in the Bothnian Bay. These periods, not present every year, were associated with severe ice conditions and an almost complete reduction of the piling-up.

1. Introduction

The sea level variations in the Baltic Sea (Figure 1), where tidal variations are negligible, are mainly due to meteorological forcing. The forcing creates in- and outflows to the basin, which causes slow (typically weeks to months) variations of the mean water level. Added to these variations, the wind field over the basin causes more rapid variations, with large amplitudes in the northern and southern part of the basin (wind set-up). Therefore the coupling between the ends of the basin is often well established.

In wintertime when large parts are ice-covered, especially in the Bothnian Bay and Sea, the effect of wind stress is transferred to the sea via the ice drift. The momentum energy input to the sea is then changed. In the case of an open water surface, the air-water drag coefficient is wind speed-dependent (*Garratt*, 1977). In an ice-covered sea, the drag

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coefficient varies depending on ice types and seasonal meteorology (*Overland*, 1985). At high ice concentrations, mechanical energy also dissipates through internal ice friction and ridging processes. In the extreme case, the ice becomes immobile and decouples the atmosphere-ocean momentum exchange completely.

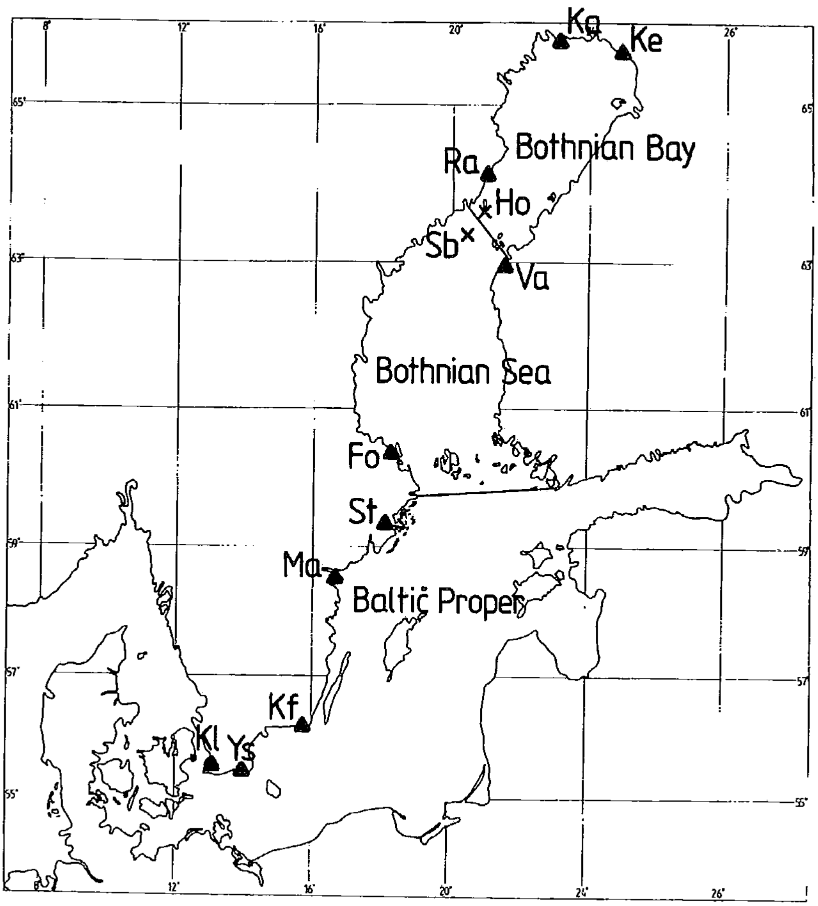

Figure 1.The Baltic Sea with water level and wind stations. The triangles indicate sea level stations, and the crosses indicate meteorological stations. Kl = Klagshamn, Ys = Ystad, Kf = Kungsholmsfort, Ma = Marviken, St = Stockholm, Fo = Forsmark, Ra = Ratan, Ka = Kalix, Ke = Kemi, Va = Vaasa, Sb = Sydostbrotten and Ho = Holmögadd.

The reducing influence on the piling-up of water due to sea ice was studied by *Lisitzin* (1957). Based upon sea level observations from the Bothnian Bay, she showed that the sea level amplitudes were considerably reduced during periods with ice-cover. *Leppäranta* (1981, p. 75) calculated the corresponding reduction in potential energy and argued that this could be explained by the energy loss within the ice cover due to internal friction.

The ice influence on sea level variations when tidal forcing is important was discussed by *Zubov* (1963). In this now classical book several examples are given on how the ice influences the tidal variations. For example, the sea level amplitude was reduced from 3 to 0.68 m during ice-covered conditions in the White Sea in the winter of 1929. Some further studies on tides in ice-covered waters are given by *Sverdrup* (1926), *Hunkins* (1965), *Kowalik* and *Matthews* (1982).

The purpose of the present paper is to analyse the influence of sea ice on water level variations in the Baltic Sea. For that purpose data from 8 water level stations during 11 years are analysed with statistical and dynamic methods. The material is divided in different groups with respect to the amount of ice, and wind data are also considered in the analysis.

2. *Theoretical considerations*

The wind stress is often the prime driving force of the circulation and causes piling-up when coasts are present. In steady state, only considering variations in the on-shore direction, also neglecting rotation effects as well as bottom friction and ice, one can easily derive an expression for the sea surface slope (*Csanady*, 1984). The slope of the ice-free water surface, S^o, is described according to:

$$S^0 = \frac{\tau_a^0}{g\rho d} \tag{1}$$

where τ_a^0 is the wind stress, g the gravity constant, ρ the water density and d the average water depth.

The wind stress is calculated from a bulk formula:

$$\tau_a^0 = \rho_a C_a^0 V_a^2 \tag{2}$$

where ρ_a is the air density, C_a^0 the air drag coefficient and V_a the wind speed. Following *Garratt* (1977) the air drag coefficient is a function of wind speed (referred to 10 m) according to:

$$C_a^0 = \left(0.75 + 0.067\ V_a\right) \cdot 10^{-3} \tag{3}$$

If the wind set-up, ΔH, is considered over a given distance, the following practical formula can be derived, using equations (1) and (2):

$$\Delta H = \alpha^0 \frac{V_a^2}{d} \tag{4}$$

The factor α^0 is a constant introduced by *Lisitzin* (1957), defined according to:

$$\alpha^0 = \frac{\rho_a C_a^0 \Delta X}{g\rho} \tag{5}$$

where ΔX is the horizontal distance.

The slope of the ice-covered water surface, S^i, can be derived in a similar manner according to:

$$S^i = \frac{\tau_w}{g\rho d} \tag{6}$$

where τ_w is now the stress between ice and water.

If only variations in the on-shore direction are considered (neglecting the acceleration, the Coriolis force and the ocean surface tilt in the ice drift equation), and the ice is assumed to move freely, the ice drift equation (*Omstedt*, 1990) reads:

$$\tau_a^i - \tau_w = 0 \tag{7}$$

where τ_a^i is the wind stress on the ice. The stress reads:

$$\tau_a^i = \rho_a C_a^i V_a^2 \tag{8}$$

where C_a^i is the air drag coefficient above the ice field, which varies typically between $1 \cdot 10^{-3}$ - $4 \cdot 10^{-3}$ according to *Overland* (1985) and *Bruno* and *Madsen* (1989) for smooth to rough ice and for different meteorological conditions. Measurements and ice drift studies in the Baltic Sea in general apply $C_a^i = 1.8 \cdot 10^{-3}$ (*Joffre*, 1984; *Leppäranta* and *Omstedt*, 1990).

The ratio between the slopes then reads:

$$\frac{S^i}{S^0} = \frac{C_a^i}{C_a^0} \tag{9}$$

Different values on the drag coefficients are illustrated in Figure 2. It is interesting to notice that one should expect a larger sea surface slope for ice-covered waters with rough ice at low wind speeds.

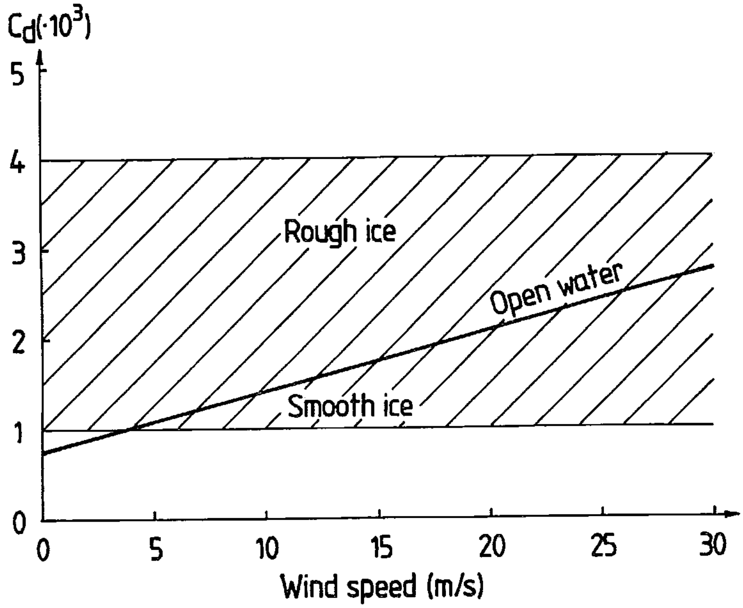

Figure 2. Drag coefficients for an open water surface and for different ice types.

The ice set-up, ΔH^i, over a given distance, ΔX, can be written in a similar manner (using equations (6) - (8)) as:

$$\Delta H^i = \alpha_f^i \frac{V_a^2}{d} \tag{10}$$

where now the factor α_f^i is a constant representing ice conditions with freely drifting ice. The α_f^i factor reads:

$$\alpha_f^i = \frac{\rho_a C_a^i \Delta X}{g\rho} \tag{11}$$

In the case of high ice concentrations one also needs to consider internal ice friction in the ice drift equation. Starting from the water stress formulation:

$$\tau_w = \rho\, C_w \left(U_i - U_w\right)^2 \cong \rho\, C_w\, U_i^2 \qquad (12)$$

where C_w is an ice-water drag coefficient, U_i the ice drift and U_w the current speed which is assumed to be is small compared to the ice drift. Now introducing the linear ice drift model by *Omstedt* (1990), the ice drift can be written according to:

$$U_i = U_i^f - U_i^P \qquad (13)$$

where U_i^f is the free ice drift and U_i^P the plastic ice drift reduction due to ridging. In the ice drift formulation U_i is positive for on-shore drift, and U_i^P is always less or equal to U_i^f. The plastic ice drift reduction reads, within given assumptions:

$$U_i^P = \frac{P\left(X_{dim}\right)}{k\left(X_{dim} - X_f\right)} \qquad (14)$$

where $(X_{dim} - X_f)$ is the ice cover extension, k is a stress coefficient equal to 1, and $P(X_{dim})$ the ice strength at the coast.

The parameterization of the ice strength follows that by *Hibler* (1979), which reads:

$$P = P_*\, h_i \exp\left(-C\left(1 - A_i\right)\right) \qquad (15)$$

where h_i is the ice thickness, P_* and C ice strength constants, and A_i the ice concentration.

Now introducing equations (12) and (13) into equation (6), the practical formula for the ice set-up reads:

$$\Delta H^i = \alpha_P^i \frac{V_a^2}{d} \qquad (16)$$

where the factor α_P^i is a constant representing ice-covered conditions. The α_P^i factor reads:

$$\alpha_P^i = \alpha_f^i \left(1 - \frac{U_i^P}{U_i^f}\right)^2 \qquad (17)$$

From equation (16) one can notice that for free ice drift, $U_i^P = 0$, the formula is equal to equation (10). In the case of severe ice conditions, the formula approaches zero and the ice set-up is completely reduced.

3. *Materials and methods*

3.1 *Material and ice classes*

The water level data used in the present study are based upon hourly values from eight stations along the Swedish coast (Fig. 1). The time period considered was from August 7, 1975, to August 1, 1986. To reduce the data, three-hour averages were calculated.

Wind information each third hour was extracted from the synoptic weather stations of Sydostbrotten and Holmögadd. The data from Holmögadd were added to the Sydostbrotten data series only when information was missing (a period of about two years). The measuring heights above the sea level at the stations are 36 and 13 meters respectively. To get the standard meteorological level (10 m), the winds from Sydostbrotten were reduced by 10 %.

The studied period was classified with respect to ice on the basis of chart information. The ice charts represent a typical two-day average. The classification divided the time period into four groups and three different sea areas were considered. The ice classification and the number of days for each class is given in Table 1. In Section 4.1, some mean properties of the water level data are discussed.

Table 1. Ice classification for the period 1975 - 1986, where 0 = open water, 1 = open ice (ice concentration 0.1 - 0.6), 2 = close ice (ice concentration 0.7 - 0.8), 3 = consolidated ice (ice concentration 0.9 - 1.0). The different basins are BB = Bay of Bothnia, SB = Sea of Bothnia, BP = Baltic Proper.

Ice class for basin			Number of days
BB	SB	BP	
3	3	1	194
3	3	0	35
3	2	1	113
3	2	0	121
3	1	1	17
3	1	0	427
3	0	0	26
2	2	0	6
2	1	1	8
2	1	0	212
2	0	0	109
1	1	0	67
1	0	0	638
0	1	0	1
0	0	0	2039
Total			4013

3.2 *Statistical methods*

A systematic statistical way to analyse the behaviour of sea level variations is to expand the data into Empirical Orthogonal Functions (EOF). By the requirement of optimized convergence the data are effectively organized into different modes. The mean variance of all data and the contributions to the variance from the different terms in the series are then easy to calculate (*Holmström* and *Stokes*, 1978). The sea level at our stations, $S_i(t)$, i = 1,8, expanded into EOF, reads:

$$S_i(t) = \sum_{n=1}^{8} \beta_n(t)\, h_{ni} \tag{18}$$

where the amplitudes $\beta_n(t)$ and the local response functions h_{ni} were determined to give the series an optimized convergence. It should be noticed that the amplitude functions $\beta_n(t)$ do not depend on i and are therefore common for all stations. Different EOF expansions of the material were performed in the study and are discussed in Section 4.2.

3.3 *Wind and ice set-up*

The influence of sea ice on sea levels was studied by applying the same method as *Lisitzin* (1957), and the theoretical considerations are given in Section 2. The analysis by *Lisitzin* (1957) started from a 15-year period, 1928-1942, where episodes were selected when the average wind speed was at least 14 m/s from a direction between S and SW. With the aid of wind observations, 68 episodes could then be identified during the period 1928-1942. Sea level differences between Kemi and Vaasa were calculated for these episodes, and the ice conditions were classified subjectively. From the water level and wind data during the present studied period 1975-1986, 210 different episodes (99 with northerly and 111 with southerly winds) were selected and classified into different ice classes, see Table 2. In Table 2 we have also classified the data from *Lisitzin* (1957) into the same ice classes, as applied in the present paper. The results are presented in Section 4.3.

Table 2. Number of studied episodes in the study by *Lisitzin* (1957) (a), in the present study with southerly winds (b and c), and with northerly winds (d and e). Mean α-values (s^2) and standard deviations for different ice classes are also given. All α-values are comparable with those given by *Lisitzin* (1957). To get SI units one should divide them with a factor of 100/2.2.

(a) Lisitzin (1957)				
Ice class	0	1	2	3
Number of cases	43	18	1	6
α-mean	2.96	2.16	1.70	0.67
Standard deviation	0.42	0.59	-	0.38
(b) Present study, southerly winds 10-14 m/s				
Ice class	0	1	2	3
Number of cases	31	25	9	9
α-mean	2.16	2.07	1.68	1.15
Standard deviation	0.64	0.55	0.67	0.48
(c) Present study, southerly winds > 14 m/s				
Ice class	0	1	2	3
Number of cases	11	10	8	8
α-mean	2.07	1.54	1.42	1.11
Standard deviation	0.74	0.49	0.27	0.52
(d) Present study, northerly winds 10-14 m/s				
Ice class	0	1	2	3
Number of cases	34	16	5	16
α-mean	1.78	2.04	1.44	1.32
Standard deviation	0.94	0.80	0.55	0.65
(e) Present study, northerly winds > 14 m/s				
Ice class	0	1	2	3
Number of cases	14	5	7	2
α-mean	1.27	1.12	0.87	0.96
Standard deviation	0.40	0.28	0.23	(0.46)

4. *Results and discussion*

4.1 *Mean properties*

The yearly cycle of the water levels at all stations is illustrated in Figure 3. The largest amplitudes are found at the most southerly and northerly stations and during autumn and early winter. The water level amplitudes are reduced during mid-winter, spring and summer, indicating less low pressure activity in the atmosphere during these months.

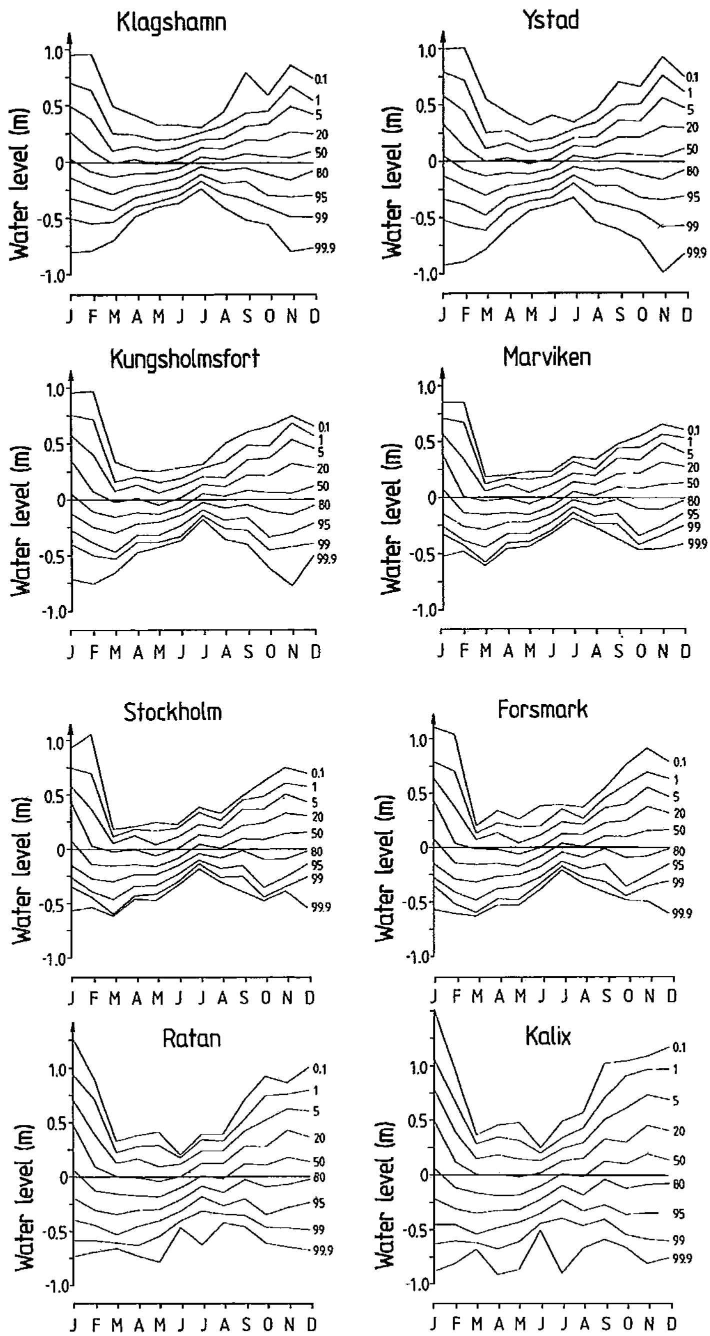

Figure 3.The yearly cycle of sea levels at studied stations. The different curves are for the percentiles indicated, i.e. the value indicated by the curve is exceeded by so many percent of the observations.

The effects due to sea ice is analysed in Figure 4. Only water level data from the Bothnian Bay are considered, and the data are collected with respect to different ice classes. It should be noticed that the number of observations are different for different classes, which may explain some apparent irregularities in the extreme percentiles. From the figure one can notice that ice-free periods have higher sea levels compared with the case when ice is present in the Bothnian Bay. The reason is different meteorological conditions during the year, as illustrated in Figure 3. In the case of severe ice conditions the amplitudes are significantly reduced, partly due to the meteorological conditions and partly to ice.

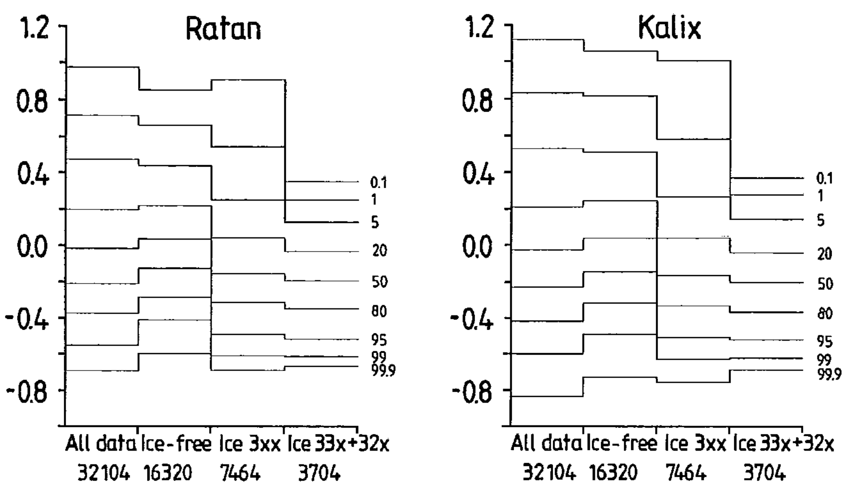

Figure 4.The sea level amplitudes at Ratan and Kalix for different ice classes. The ice classes are: ice-free = no ice, ice 3xx = consolidated ice in the Bothnian Bay and all other ice types in the Bothnian Sea and Baltic Proper. Ice 33x + 32x = consolidated ice in the Bay of Bothnia together with close or consolidated ice in the Sea of Bothnia. The numbers below each class indicate numbers of observations.

4.2 Empirical Orthogonal Functions (EOF)

The local response functions h_{ni} for all data are plotted in Figure 5. It should be noticed that the first mode covers not less than 73.7 % of the total variance. The signature of this response pattern indicates a general increase or decrease of the sea levels in the whole Baltic Sea. The second response function adds 20.3 % to the total variance, and thus these two modes together cover 94 % of the total variance. The signature of the second response mode illustrates a close correlation between an increase or decrease in the northern part with a decrease or increase in the southern part of the Baltic Sea. A covariance as large as this indicates a close correlation to the meteorological forcing (air pressure and winds). In the higher response functions more complex response patterns are collected. Together they add only 6 % to the total variance and are not further discussed.

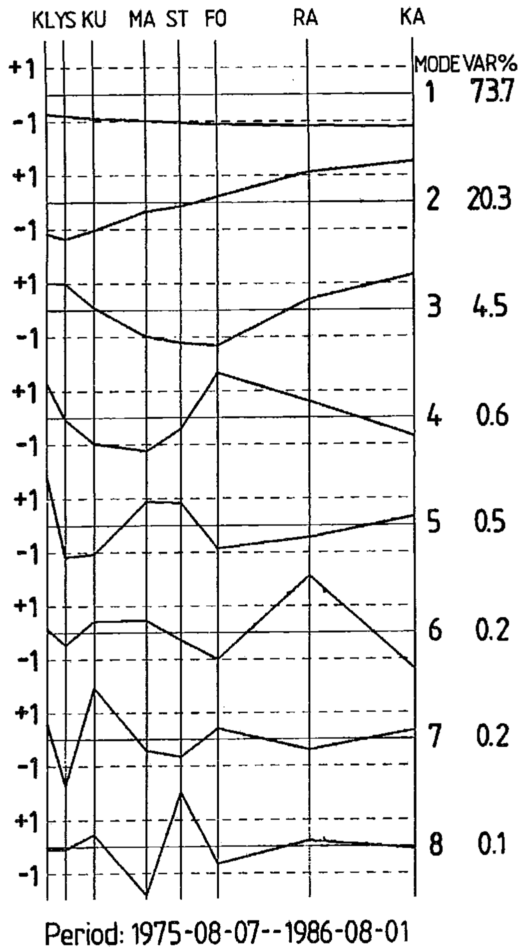

Figure 5. Normalized h_{ni}-functions in the case of all data. For the abbreviations, see Figure 1.

For the study of the effects of ice on sea levels several different EOF expansions were made. These were based upon different time periods and different ice classes. What we expected was that the influence of ice on sea levels should be detected mainly in the second mode. In general it was, however, difficult to find any general trends, and different response patterns were given for different winters. Two examples will be given and discussed below.

The first case is from March 1980 (Fig. 6). During that period consolidated ice was present in the Bothnian Bay and Sea, and open ice in the Baltic Proper. The water levels show a high correlation without any pronounced tilting of the sea surface. The short period oscillations at Klagshamn indicate tidal motions. The local response modes for the same period are given in Figure 7. During this period no less than 91 % of the total variance are due to a general lowering or raising of the whole surface. The winds were very weak during this period.

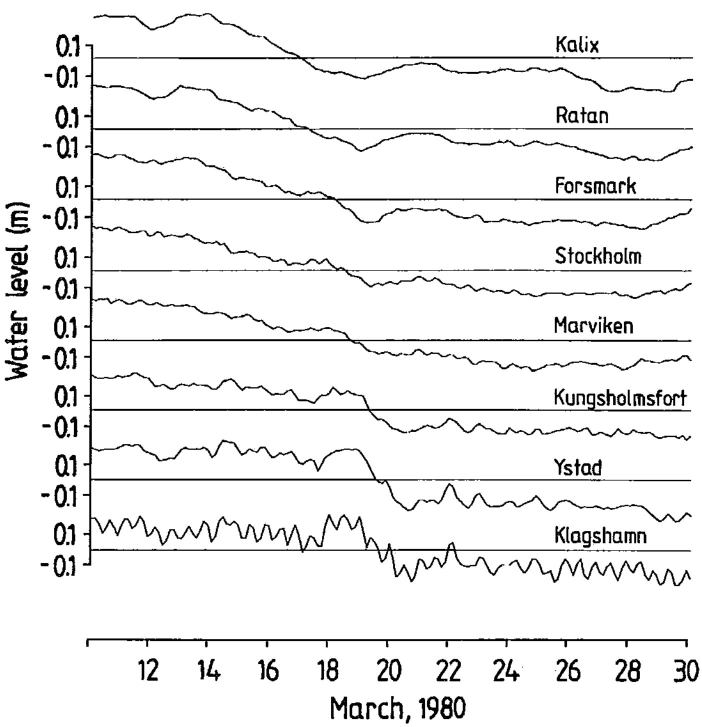

Figure 6. Sea level data from the period 800310 - 800329.

The second case is from February 1985 (Fig. 8). The ice classes are almost the same as in the first case, but now the winds are stronger. The water levels in the Bothnian Bay and Sea are closely correlated. In the Baltic Proper the water levels are more variable. The local response modes for the corresponding period are given in Figure 9. The signatures of the response modes are now quite different, with a tilting structure as the first mode (tilting is mainly in the Baltic Proper). The first mode now explains only 65.8 % of the total variance. The stations in the Bothnian Bay and Sea contribute very little to this mode.

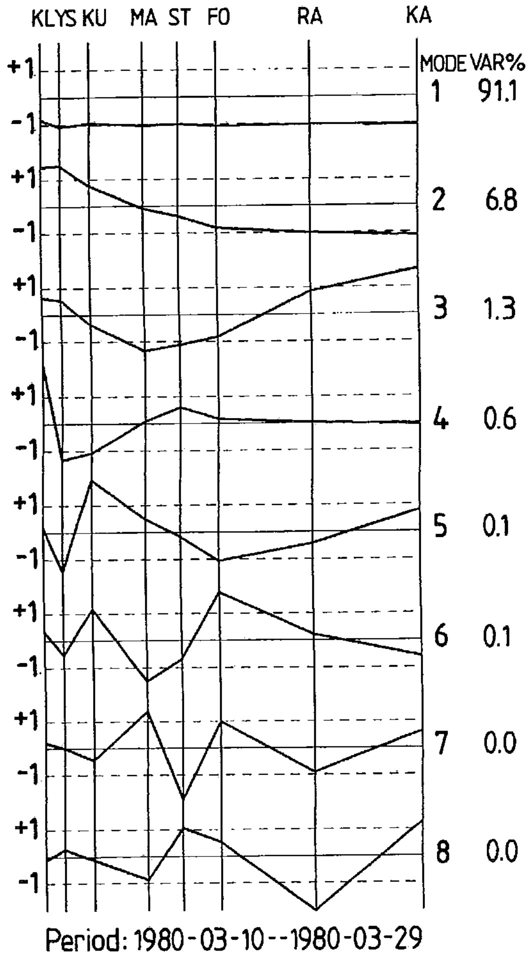

Figure 7. Normalized h_{ni}-functions for the period 800310 - 800329. For the abbreviations, see Figure 1.

The two first modes together, however, explain 92.3 % of the total variance. This case illustrates how severe sea ice damps the piling-up and decreases the correlation between water levels in the northern and southern part of the Baltic Sea. During February 1985 ice drift measurements were made in the Bothnian Bay (*Leppäranta*, 1987). The ice pack was consolidated and about 0.60 m thick. The maximum wind velocity during the period was 17 m/s, and the ice showed no motion at all. The stagnation of the ice indicates a considerable ice strength during this period. The average ice strength was estimated to more than $2.2 \cdot 10^5$ Nm^{-2} (*Leppäranta*, 1987).

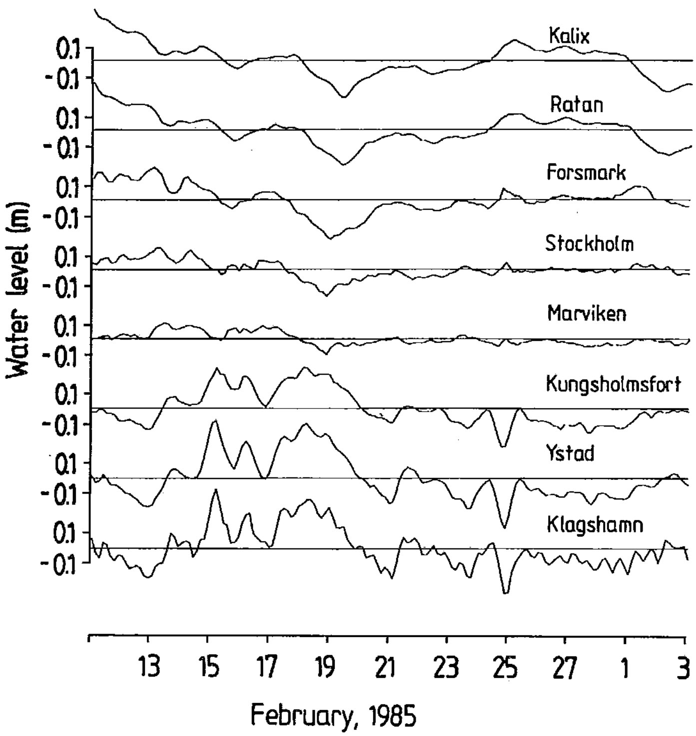

Figure 8. Sea level data from the period 850211 - 850302.

4.3 *Wind and ice set-up*

The piling-up between Kalix and Ratan is illustrated in Figure 10. The calculations were based upon the difference in sea level between the two stations, referred to local means. From the figure one can notice how the amplitudes are reduced when ice is present in the Bay. To interpret the reduction in piling-up due to ice we recall the theoretical considerations in Section 2. On the basis of the equations derived in Section 2 we could estimate the α-values for different idealized conditions. In Figure 11 the calculated variations of α^0 and α_f^i due to winds and ice types are illustrated. The air drag coefficient over open water is calculated from equation (3), while the air drag coefficient over pack ice is varied between 1.10^{-3} and $4 \cdot 10^{-3}$.

Due to variable wind speed or ice type, the α-values vary from 1 to 5, with increasing values for increasing wind speed or increasing roughness. As we only have considered cases with wind speed higher than 10 m/s, the variations should probably be somewhat smaller, but α-values less than 1 are not predicted in Figure 11.

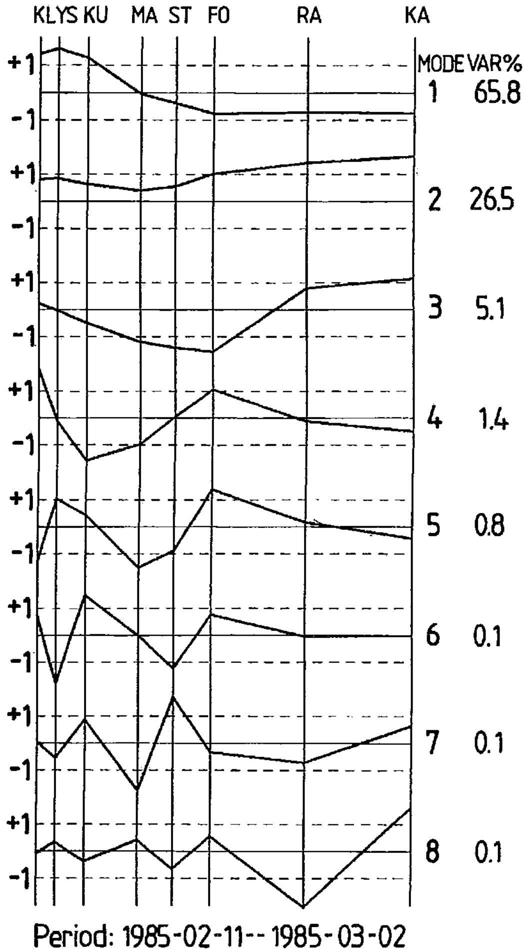

Figure 9. Normalized h_{ni}-functions for the period 850211 - 850302. For the abbreviations, see Figure 1.

The values of $\alpha_P{}^i$ for different ice strengths are given in Figure 12. In the calculations, the ice thickness is put equal to 0.5 m and the ice concentration to 0.99. Also in the figure an α_f^i-value is given corresponding to a drag coefficient typical for the Baltic Sea ($C_a^{\ i}$ = $1.8 \cdot 10^{-3}$). From the figure one can notice that with increasing ice strength the $\alpha_P{}^i$-values decrease and become zero for low winds. At higher winds the $\alpha_P{}^i$-values show wind dependency for constant ice strength. At lower ice concentrations, the reduction of piling-up becomes less and is insignificant for ice concentrations less than 0.8, even if the thickness is put equal to 1 meter. We may thus expect that reduction of the piling-up due to ice ridging will only occur during consolidated ice conditions, defined as when the ice concentration is between 0.9 and 1.0.

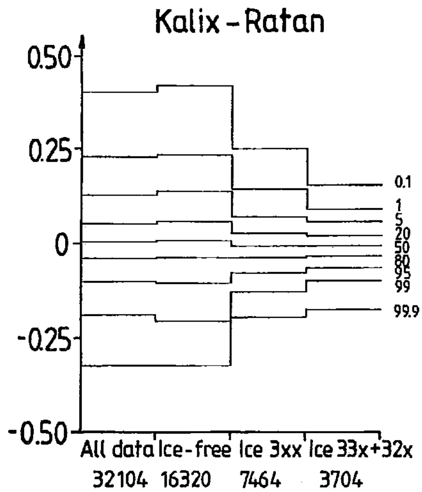

Figure 10. The sea level differences between Kalix and Ratan for different ice classes. The ice classes are explained in Table 1 or Figure 4. The numbers below each class indicate numbers of observations.

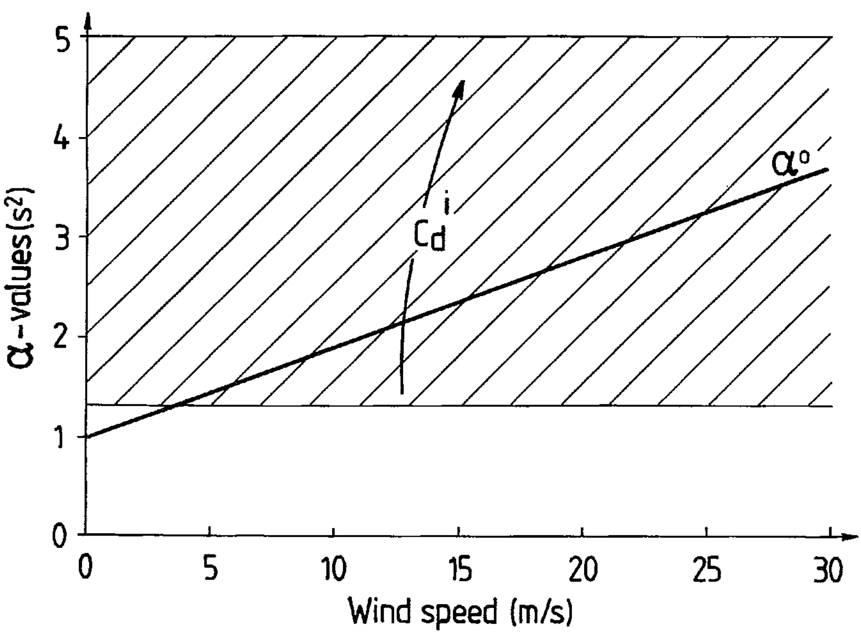

Figure 11. α^0- and α_f^i-values calculated from Equations (5) and (11), putting $\rho_a = 1.3$ kg m^{-3}, $\Delta X = 2.2 \cdot 10^5$ m, $g = 9.81$ m s^{-2} and $\rho = 10^3$ kg m^{-3}. All α-values are comparable with those given by *Lisitzin* (1957). To get SI-units one should divide the α-values with a factor of 100/2.2.

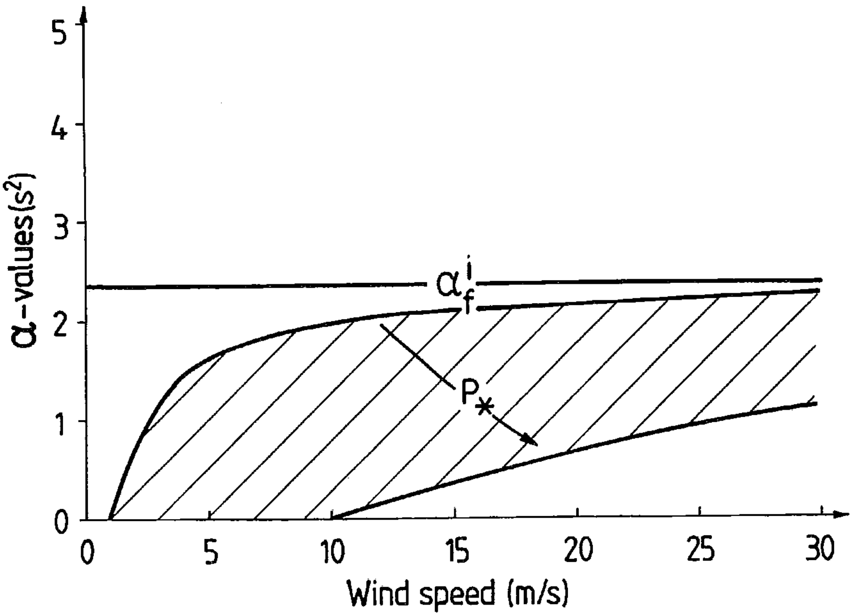

Figure 12. $\alpha_P{}^i$-values calculated from Equation (17), with P_*-values from $1 \cdot 10^4$ to $1 \cdot 10^5$ Nm^{-2}. The $\alpha_f{}^i$-value corresponds to a drag coefficient $C_a{}^i = 1.8 \cdot 10^{-3}$. All α-values are comparable with those given by *Lisitzin* (1957). To get SI-units one should divide the α-values with a factor of 100/2.2.

The results up to now only consider the theoretical values predicted by the equations introduced in Section 2. We now turn our interest towards the α-values calculated from observed sea level data. From the water level difference between Kalix and Ratan, analysed on the basis of different ice classes and two classes of strong winds, the corresponding α-values were calculated using equation (4). The results are presented in Figure 13 and in Table 2. The data show quite a large scatter; however, in the case of southerly winds there is a clear reduction in the α-values. The reduction is mainly in the ice class 3, corresponding to severe ice conditions with ice concentration between 0.9 and 1.0. This is in accordance with the theoretical consideration and illustrates the importance of considering ice dynamics and ridging processes in the calculation of the piling-up. The reduction of α-values below 1 in ice class 3 indicates considerably high values of the ice strength, of order 10^5 Nm^{-2}. It should be noticed that ice strength values of this magnitude are ten times larger than those generally used in ice models. The results indicate, however, that periods with such high ice strengths may exist in the Bothnian Bay, which has also been reported by *Leppäranta* (1987).

For northerly winds, the overall values are lower, probably due to the absence of a well defined lee coast. The α-values for northerly winds also show less reduction at high ice concentrations; this indicates that less energy is put into ice ridging when the wind is blowing from the north, compared to south.

Two different wind speed classes have been used in the analysis; surprisingly, the low winds seem to give slightly larger α-values than the high winds. A possible explanation for this might be that some wind episodes were of too short a duration, so that the sea surface slope was not fully developed and therefore the quadratic dependence on wind speed in equation (4) not quite attained.

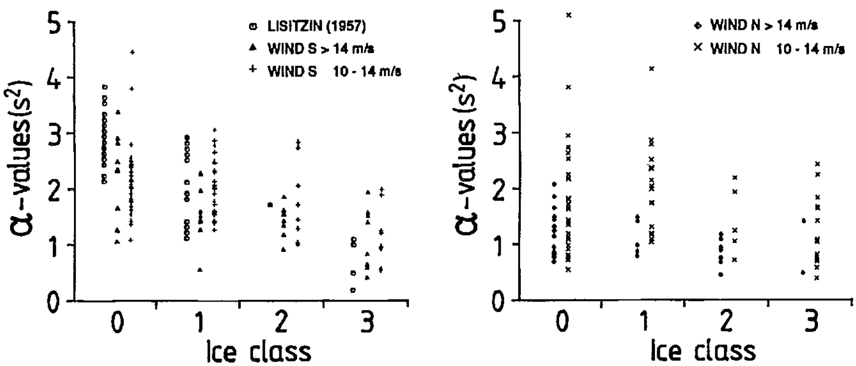

Figure 13. α-values for different ice classes in the Bay of Bothnia, calculated on the basis of water level differences between Kalix and Ratan, and in the case of *Lisitzin* (1957) between Kemi and Vaasa. All α-values are comparable with those given by *Lisitzin* (1957). To get SI-units one should divide the α-values with a factor of 100/2.2.

5. *Summary and conclusions*

The investigation analyses sea level data from eight stations along the Swedish coast of the Baltic Sea. The time period considered was from 1975 to 1986, and the data were classified into different sea ice categories. Based upon statistical and dynamic methods, the effects of ice on sea level variations were analysed.

The sea level data illustrated a significant variation during the year with largest amplitudes in the autumn and early winter. The amplitudes were reduced during midwinter, spring and summer. When the data were collected into different ice classes, it was observed that the amplitudes were reduced during severe ice conditions. The reason was partly meteorological conditions and partly ice.

The sea levels were also expanded into Empirical Orthogonal Functions (EOF). It was shown that not less than 73.7 % of the total variance was due to a general rising or lowering of the whole surface. The first tilting mode added 20.3 %, and these two modes thus explain 94 % of the total variance. The effect of sea ice on the EOF modes was analysed, but it was difficult to find any general trends. However, during severe ice conditions it was observed that the variations were reduced at the northern stations, and the correlation between the northern and southern parts of the Baltic Sea was decreased. The effects of sea ice on the water level variations were most easily detected when analysing the sea surface tilting in the Bothnian Bay. Following the method by *Lisitzin* (1957) it was shown that the ice cover may reduce the piling-up. The present analysed period thus confirms the results by *Lisitzin* (1957). From theoretical considerations it was shown that factors as wind speed, ice types and ice strength may influence the reduction. The data also indicated that periods with quite high ice strengths (order of 10^5 Nm^{-2}) have occurred in the Bothnian Bay. These periods, not present every year, were associated with severe ice conditions and an almost complete reduction of the piling-up.

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