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An ice drift model for the Baltic Sea

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ABSTRACT

A dynamical model for simulating and forecasting motion of sea ice fields in the Baltic Sea is presented. The model is based on the equations for conservation of mass and momentum and on an equation describing the distribution of ice mass. In the model, ice drift is caused by wind; currents beneath the boundary layer in the sea are neglected. The material behaviour of sea ice fields is described by a linear viscous law. Mass of sea ice fields is decomposed into concentration, level ice thickness, ridge density and ridge height. The model parameters have been estimated from the data from field experiments in the Baltic Sea. The equations are solved numerically with the finite difference technique; the grid size is 27 km and the time step 6 hours. The model has been in operational use in forecasting ice conditions in the Gulf of Bothnia since winter 1977; 30 h forecasts have been calculated daily and transmitted in chart form to the icebreakers.

1. Introduction

Ice conditions in the Baltic Sea show a very large variation in different winters. In mild winters only the Bothnian Bay and the eastern part of the Gulf of Finland have ice cover, and this area covers about one-seventh of the total area of the Baltic Sea. On the other hand, in very cold winters the whole Baltic is ice covered, as last occurred during the 1940s (Palosuo, 1953). In an average winter, the Gulf of Bothnia, the Gulf of Finland and the Gulf of Riga are ice covered, but in the Baltic proper the ice edge does not extend south of $\sim 59°$ N except for narrow strips in the coastal areas. An example of an average mid-winter situation is shown in Fig. 1. The duration of the ice winter also varies much in the different parts of the Baltic Sea. In the Bothnian Bay it is 5–7 months and in the Baltic proper 0–3 months.

Ice in the coastal area to the depth contour of approximately 10 m is stationary, fast ice. Outside the fast ice zone pack ice fields are found which drift and deform under the influence of winds and currents. Due to the high variability in the driving forces, permanent features in pack ice drift are hard to find. Largest displacements are always connected with high winds which gives, from the point of view of prediction, a large random component to ice drift on a time scale of more than 1 week. One permanent feature, however, the general outflow of ice from the Gulf of Finland to the northern Baltic proper (Jurva, 1942) may be mentioned. This is due mainly to the large difference in the ice production rate between the Baltic proper and the Gulf of Finland. It must be mentioned here that, from the point of view of ice dynamics, the Bothnian Bay and Bothnian Sea are nearly closed basins in contrast to the Gulf of Finland.

Research on the modelling of sea ice drift in the Baltic was begun in Finland in 1973. The aim was to produce real-time ice forecasts for winter navigation. The starting points were the two well-known Arctic sea ice models of Campbell (1965) and Doronin (1970). In the beginning the research was mainly parameter fitting on the basis of numerical simulations. Some new physical ideas emerged from the results of the AIDJEX (*A*rctic *I*ce *D*ynamics *J*oint *Ex*periment) program (Untersteiner, 1980) carried out during the 70s. Information on the ice drift and ice forms in the Baltic was obtained from the experience gained by the Ice Service in the Institute of Marine Research in Helsinki, Finland, and from some earlier studies (e.g. Palosuo, 1953, 1975).

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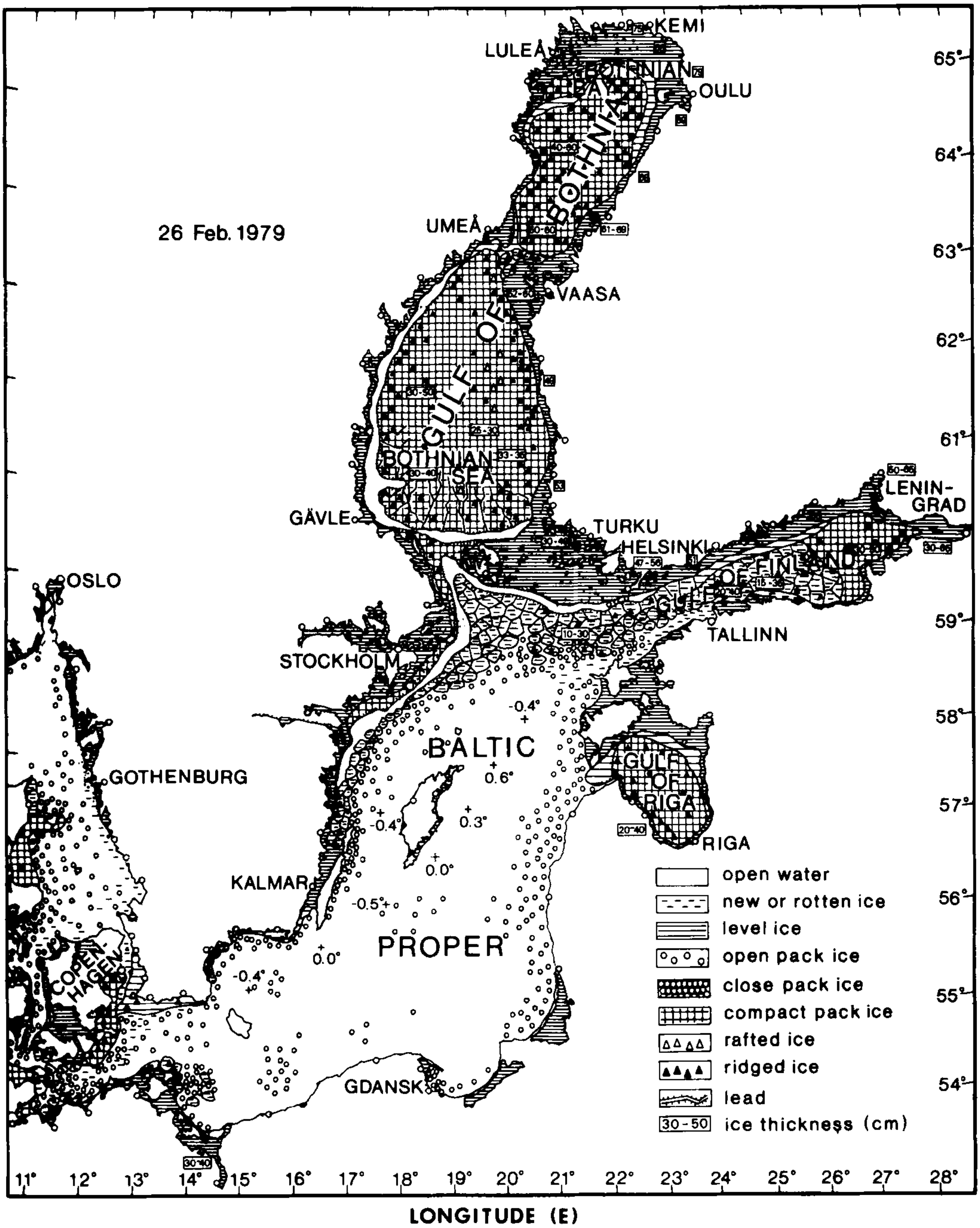


Fig. 1. The ice situation on February 26, 1979.

In the Baltic Sea the scales of motion of ice are quite different from those in the Arctic Ocean, but there is a high degree of similarity in ice drift and ice forms in these two areas. However, during the modelling work it was recognized that more quantitative data on sea ice dynamics in the Baltic were needed and hence several field experiments were carried out in the late 70s (e.g. Leppäranta, 1980).

Operational use of the developed model started in the winter 1976/77 (Leppäranta, 1977). The results were satisfactory and the use of the model is now part of the routine of the Finnish Ice Service. So far forecasts have been calculated for the Gulf of Bothnia only, but in the near future the operational use will cover the whole Baltic Sea. The modelling work has progressed in close collaboration with the sea ice research group in the Swedish Meteorological and Hydrological Institute. There, a sea ice model somewhat different from ours has been constructed (Udin and Ullerstig, 1976) and put into operational use.

2. Sea ice drift as a geophysical problem

Sea ice fields consist of a large number of individual ice floes of irregular size and shape. Geophysically, sea ice fields are thin layers of discrete material between the atmosphere and the ocean (Fig. 2). However, it is not possible in general to consider each ice floe separately, and hence particles of sea ice fields are defined large enough so that the material behaviour can be described with a suitable continuum model (Rothrock, 1975). In the Bothnian Bay the ice thickness is on the order of 0.5 m with a typical floe diameter of 0.2–2 km (most of the ice mass is in floes with a diameter in this interval). Experience suggests that the continuum length scale is of the order of 10 km. Since the thickest Baltic ice is met in the Bothnian Bay, the ~10 km scale should be realistic for the whole Baltic Sea. In comparison, in the Arctic Ocean, ~100 km is considered suitable (e.g. Rothrock, 1975).

2.1. *Conservation of momentum*

The sources of kinetic energy of sea ice are the surface shear stresses of winds and currents on the upper and lower surfaces of ice (Fig. 2). The former is generally dominant in the Baltic Sea. A third source of minor importance is the gravitational force due to the sea surface tilt. Part of the kinetic energy of drifting ice mass is dissipated in deformation processes in the ice field, while the rest is given to the oceanic boundary layer. Whether the dissipation is important or not largely depends on ice concentration N. Observations of Shirokov (1977) indicate that the transition value, above which dissipation becomes important, is $N \approx 0.8$. The cases where drifting ice gives all its kinetic energy to the sea are called free drift. Observations and numerical solutions of the time-dependent free drift equation show that the inertia of Baltic sea ice is negligible on a time-scale of a few hours or more (Leppäranta, 1980). A rather high correlation exists between the velocities of ice and wind in the Baltic Sea (Fig. 3). However, the ice velocity fluctuates much more than could be expected on the basis of wind data, and ice drift may even be suspended during a moderate wind.

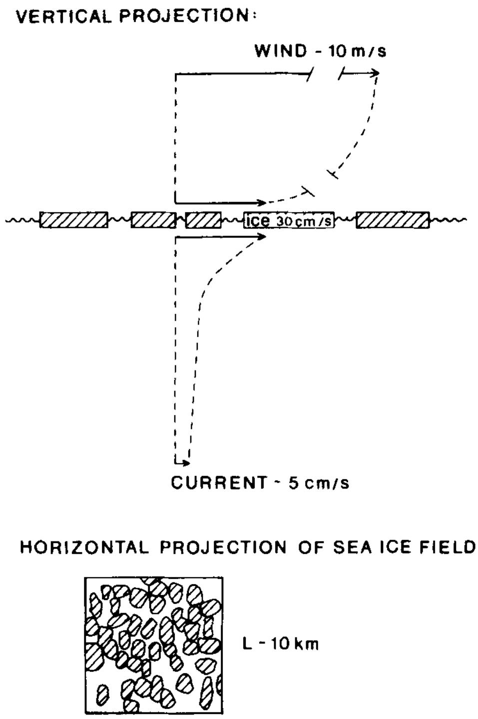


Fig. 2. Atmosphere–sea ice–ocean system.

The equation of motion of sea ice is written (e.g. Campbell, 1965)

$$\rho_i h \frac{d\mathbb{v}}{dt} = \nabla \cdot \boldsymbol{\Sigma} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w + \mathbb{C} + \mathbb{G}, \qquad (1)$$

where ρ_i is ice density, h ice thickness, $\mathbb{v}$ ice velocity, t time, $\boldsymbol{\Sigma}$ the stress within the ice, $\boldsymbol{\tau}_a$ and $\boldsymbol{\tau}_w$ are the tangential stresses of wind and water, respectively, on ice, $\mathbb{C}$ is the Coriolis force and $\mathbb{G}$

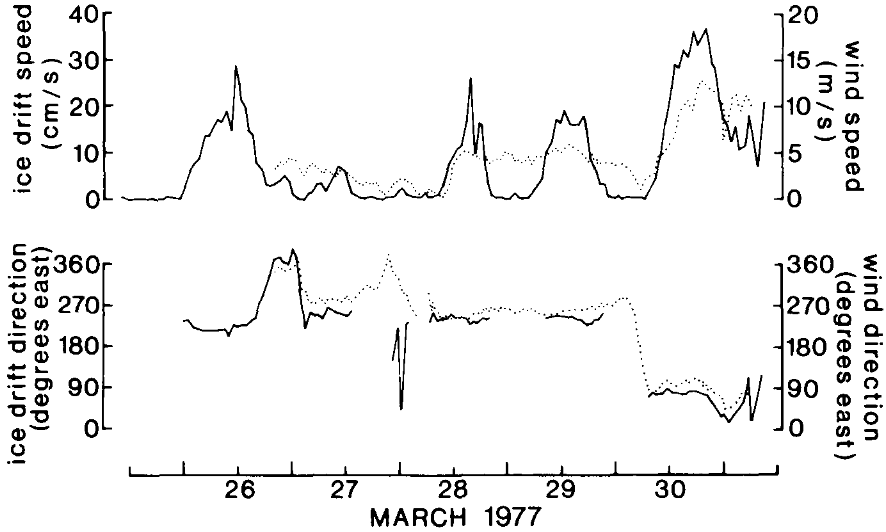


Fig. 3. Observed speed and direction of ice drift (——) and wind (.....) in March 1977, Bothnian Bay.

the gravitational force due to the sea surface tilt. Typical force diagrams are shown in Fig. 4.

A linear viscous constitutive law was proposed by Glen (1970):

$$\boldsymbol{\Sigma} = 2\eta(\dot{\boldsymbol{\varepsilon}} - \tfrac{1}{2}\,\mathrm{tr}\,\dot{\boldsymbol{\varepsilon}}\mathbb{I}) + \zeta\,\mathrm{tr}\,\dot{\boldsymbol{\varepsilon}}\mathbb{I}, \tag{2}$$

where η and ζ are the shear and bulk viscosities, respectively, $\dot{\boldsymbol{\varepsilon}} = \frac{1}{2}[\nabla\mathbb{v} + (\nabla\mathbb{v})^T]$ is the strain-rate tensor and $\mathbb{I}$ the unit tensor. Glen stated that ice fields give large resistance to compression but small resistance to tension: $\zeta \sim \eta$ for $\nabla\cdot\mathbb{v} < 0$ and $\zeta \sim 0$ for $\nabla\cdot\mathbb{v} \geq 0$, η being independent of $\nabla\cdot\mathbb{v}$. Numerical simulations with the Baltic ice drift model suggest that ζ for $\nabla\cdot\mathbb{v} < 0$ and η are of the order 10^8 kg s^{-1}, but can vary through about one order of magnitude

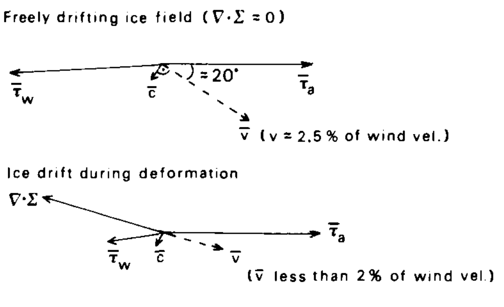


Fig. 4. Typical force diagrams when the wind speed is 5 m s^{-1}.

in the range $N \geq 0.8$. Preliminary results from analysis of observed motion and deformation of sea ice on a 15 km length scale in the Bothnian Bay support the order of 10^8 (Leppäranta and Udin, 1981). In the Arctic ζ and η are 1–3 orders of magnitude larger, depending on the season (Hibler and Tucker, 1977). If the gradients of viscosity are neglected, the divergence of (2) becomes

$$\nabla\cdot\boldsymbol{\Sigma} = \eta\nabla^2\mathbb{v} + \zeta\nabla(\nabla\cdot\mathbb{v}). \tag{3}$$

The AIDJEX modelling group constructed an elastic-plastic constitutive law (Coon et al., 1974) and Hibler (1979) used a non-linear viscous law which approximates the plastic behaviour of ice fields. In the Baltic ice drift model the viscous law (2) has been used, but experience tells us that the plastic nature of sea ice fields with $N \sim 1$ ought to be taken into account through adding some kind of yield criteria.

For the surface stresses $\boldsymbol{\tau}_a$ and $\boldsymbol{\tau}_w$ the quadratic dependencies on velocities are commonly used (e.g. Hibler, 1979; Coon, 1980):

$$\boldsymbol{\tau}_a = \rho_a c_a w[\cos\theta_a \mathbb{w} + \sin\theta_a \mathbb{k}\times\mathbb{w}], \tag{4}$$

$$\boldsymbol{\tau}_w = \rho_w c_w |\mathbb{u}_g - \mathbb{v}|\,[\cos\theta_w(\mathbb{u}_g - \mathbb{v}) + \sin\theta_w \mathbb{k}\times(\mathbb{u}_g - \mathbb{v})]. \tag{5}$$

Here, ρ is density, c drag coefficient and θ the angle of turning and the subscripts a and w refer to air and water, respectively; $\mathbb{w}$ is the wind velocity, $\mathbb{u}_g$

the geostrophic current velocity and $\mathbb{k}$ the unit vector vertically upward. In the present model the surface wind is used (hence $\theta_a = 0$) and the geostrophic current is taken as zero. It is notable that the roughness of the upper surface of ice is about the same as that of the open sea (for the surface wind $c_a \sim 1.5 \times 10^{-3}$, e.g. Joffre, 1978). Thus the input of kinetic energy from the wind is about the same whether the sea is ice covered or not. However, in the Bothnian Bay, where ice concentration typically is above 0.8, water-level variations are considerably damped in winter as compared with similar open season situations (Lisitzin, 1957), which supports the importance of energy dissipation in deformation processes. Observational evidence from the Baltic Sea suggests that $\theta_w \approx 16$ degrees and $c_w \approx 2.5 \times 10^{-3}$ (Leppäranta, 1980). The Coriolis force and the gravitational force are expressed through

$$\mathbb{C} = -\rho_i hf \mathbb{k} \times \mathbb{v}, \tag{6}$$

$$\mathbb{G} = \rho_i hf \ \mathbb{k} \times \mathbb{u}_g, \tag{7}$$

where f is the Coriolis parameter defined as twice the earth's angular velocity times the sine of the latitude. The Coriolis force becomes important when the sea ice Rossby number V/Hf (V and H being the ice velocity and thickness scales) is smaller than $\sim 10^3$. In the Baltic Sea the Coriolis force is generally almost one order of magnitude smaller than the surface stresses (Leppäranta, 1979). Since the model neglects the geostrophic current the force $\mathbb{G}$ becomes zero.

2.2. *Mass of the sea ice field*

Sea ice fields consist of level ice areas, ridges and open areas. It is useful to have all this information separated in the description of ice mass. The total thickness of ice is

$$h = h_l + h_r, \tag{8}$$

where h_l is the level ice thickness and h_r the equivalent thickness of ridged ice. The latter can be written in terms of mean sail height r_s (the height of the ridge above water level) and mean (linear) ridge density μ as (Leppäranta, 1979)

$$h_r = \mu r_s^2. \tag{9}$$

The ridge density describes the number of ridges along linear tracks and the dimensionless quantity (~50) depends on the geometry of ridge profiles and on the spatial distribution of ridges. Then the ice mass per unit area is

$$m = \rho_i hN = \rho_i (h_l + \mu r_s^2)N. \tag{10}$$

This kind of description has some advantages. Firstly, all the mass components have a clear physical meaning. Secondly, information on them is easily found using modern observation techniques. Level ice thickness measurements have been done in the Baltic for a long time on a routine basis. Ice concentration is easily mapped during reconnaissance flights. Use of a laser-profilometer to map the sail height and density of ridges was started in the Gulf of Bothnia in the winter 1978/79 (Leppäranta, 1981); according to the observations r_s is typically 40–70 cm, $\mu \sim 5$–10 km^{-1} and $h_r/h_l \sim 0.2$–0.5 but can occasionally exceed one.

Thorndike et al. (1975) presented ice thickness distribution in terms of relative areal coverage of ice of different thicknesses. In the present formulation the level ice thickness and concentration are areal integrals whereas the ridge characteristics are convenient control parameters which are related to the mass of ridged ice.

2.3. *Conservation of ice mass*

In a drifting ice field the mass changes through advection, deformation and thermodynamic processes. Conservation of mass requires that

$$\frac{\partial m}{\partial t} = -\nabla \cdot (m\mathbb{v}) + \phi \tag{11}$$

(Doronin, 1970), where $-\nabla \cdot (m\mathbb{v})$ describes the effect of dynamics and ϕ that of thermodynamics. In the present work thermodynamics is not considered and hence ϕ is assumed to be zero.

Dividing eq. (11) by ice density ($\approx$ constant) and using eq. (10) the result can be written as

$$\frac{\partial}{\partial t}(hN) = -\mathbb{v} \cdot \nabla(hN) - (hN)\nabla \cdot \mathbb{v}. \tag{12}$$

On the right-hand side the first term describes advection and the second term deformation. Let us write the continuity condition for the four components of ice mass as

$$\frac{\partial}{\partial t}(N, h_l, \mu, r_s) = -\mathbb{v} \cdot \nabla(N, h_l, \mu, r_s) + (\psi_N, \psi_l, \psi_\mu, \psi_s), \tag{13}$$

where ψ_N, ψ_l, ψ_μ and ψ_s are the changes in N, h_l, μ and r_s, respectively, due to deformation. The first term on the right-hand side of (13) describes change due to advection and it has a simple explicit form. The term due to deformation must satisfy the constraint due to conservation of mass. Thus, we have from eqs. (10) and (12)

$$h\psi_N + N\{\psi_l + r_s(2\mu\psi_s + r_s\psi_\mu)\} = -hN\nabla\cdot\mathbf{v}. \quad (14)$$

Three cases of deformation arise: open water areas may form or close. In compact ice cover rafting or ridging occurs during convergence. It is known that thin ice floes override each other without breaking. Parmerter (1975) derived a theoretical value of 17 cm for maximum thickness of rafting in the Arctic. Using the measured values of Määttänen (1973) for the flexural strength (0.59 MN m^{-2}) and Young's modulus (5.7 GN m^{-2}) of sea ice in the Bothnian Bay, Parmerter's equation gives 8 cm for the maximum rafting thickness h_{cr}. The representative value for the whole Baltic should be slightly larger, $h_{cr} \approx 10$ cm.

(i) Opening and closing of leads and polynyas: $N < 1$ or $\nabla\cdot\mathbf{v} \geq 0$. Only concentration changes; formally $\psi_l = \psi_\mu = \psi_s = 0$ and hence $\psi_N = -N\nabla\cdot\mathbf{v}$.

(ii) Rafting: $N = 1$ and $\nabla\cdot\mathbf{v} < 0$ and $h \leq h_{cr}$. Level ice thickness increases and ridges come closer together, which results in an increased μ.

$$\begin{aligned}&\psi_N = 0, \quad \psi_l = -h_l\nabla\cdot\mathbf{v},\\ &\psi_\mu = -\mu\nabla\cdot\mathbf{v}, \quad \psi_s = 0.\end{aligned} \quad (15)$$

(iii) Ridging: $N = 1$ and $\nabla\cdot\mathbf{v} < 0$ and $h > h_{cr}$. Deformation changes only the ridge components; hence $\psi_N = \psi_l = 0$. The continuity condition (14) reduces to

$$2\mu r_s\psi_s + r_s^2\psi_\mu = -h_l\nabla\cdot\mathbf{v} - \mu r_s^2\nabla\cdot\mathbf{v}. \quad (16)$$

The second term on the right-hand side arises as ridges come closer together and thus it increases μ. The first term on the right-hand side is due to build-up of new ridges. The problem arises as to how large ridges will be formed. Data from the Baltic Sea show no correlation between μ and r_s (Leppäranta, 1981) and hence an additional assumption must be made. It is known through experience that the size of ridges largely depends on the thickness of level ice during the time of ridge formation (Palosuo, 1975). We assume a linear form

$$r_s^* = \beta h_l, \quad (17)$$

where r_s^* is the sail height of new ridges and β is just greater than one. Then the change in the mean sail height depends on the relative increase of ridge density, pre-existing mean sail height and the sail height of new ridges. Equations (16) and (17) give

$$\psi_\mu = -\mu\nabla\cdot\mathbf{v} - \frac{1}{\beta^2 h_l}\nabla\cdot\mathbf{v}.$$

$$\psi_s = -\frac{1}{2\beta\mu r_s h_l}\left\{(\beta h_l)^2 - r_s^2\right\}\nabla\cdot\mathbf{v}. \quad (18)$$

3. Ice drift model

3.1. *The system of equations*

The basic assumptions in the model are:

(i) The sea current beneath the boundary layer vanishes, i.e. $\mathbf{u}_g = 0$. The ice drift then becomes purely wind-driven. This assumption may cause large relative errors, when the wind speed is low but for practical applications such cases are not very important.

(ii) The constitutive law for the ice field is the linear viscous law (2).

(iii) Mass changes due to thermodynamics are ignored. This assumption is usually realistic, since the period of forecast is not more than 2 days. However, during periods of very low air temperature in the middle winter the assumption does give rise to errors.

(iv) Steady-state ice drift.

The system of equations of the model is written

$$\nabla\cdot\boldsymbol{\Sigma} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w + \mathbb{C} = 0, \quad (19)$$

$$\boldsymbol{\Sigma} = 2\eta(\dot{\boldsymbol{\varepsilon}} - \tfrac{1}{2}\,\mathrm{tr}\,\dot{\boldsymbol{\varepsilon}}\mathbb{I}) + \zeta\,\mathrm{tr}\,\dot{\boldsymbol{\varepsilon}}\mathbb{I}, \quad (20)$$

$$m = \rho_i(h_l + \mu r_s^2)N, \quad (21)$$

$$\frac{\partial}{\partial t}(N, h_l, \mu, r_s) = -\mathbf{v}\cdot\nabla(N, h_l, \mu, r_s) + \boldsymbol{\psi}, \quad (22)$$

Equation (19) is the momentum equation, eq. (1), where the inertia and gravitational force have been neglected; eq. (20) the constitutive law of sea ice fields, eq. (2); eq. (21) the mass distribution function, eq. (10); eq. (22) the mass continuity equation, eq. (13), where $\boldsymbol{\psi} = (\psi_N, \psi_s, \psi_\mu, \psi_s)$. The external forces in the momentum equation are expressed by

$$\boldsymbol{\tau}_a = \rho_a c_a w_{10} \mathbf{w}_{10}, \tag{23}$$

$$\boldsymbol{\tau}_w = -\rho_w c_w v(\cos\theta_w \mathbf{v} + \sin\theta_w \mathbf{k}\times\mathbf{v}), \tag{24}$$

$$\mathbb{C} = -\rho_i hf \mathbf{k}\times\mathbf{v}. \tag{25}$$

$\boldsymbol{\tau}_a$ is the wind stress, eq. (4), with $\mathbf{w} = \mathbf{w}_{10}$ the surface wind velocity and $\theta_a = 0$; $\boldsymbol{\tau}_w$ the water stress, eq. (5), with $\mathbf{u}_g = 0$; $\mathbb{C}$ the Coriolis force, eq. (6). The viscosities η and ζ depend on the large-scale concentration of the ice field, though the exact functional shape is not known. Numerical experiments suggest that η decreases about one order of magnitude from the middle winter to spring. The values for the parameters used in the model are shown in Table 1.

Examples of other models are found in Campbell (1965) and Doronin (1970). Doronin's model is the first complete dynamic-thermodynamic sea ice model. These, like most models, have been used for the Arctic regions. Sea ice models for the Antarctic have been presented by Parkinson and Washington (1979) and Ling et al. (1980). The most advanced models at present are Hibler's model (Hibler, 1979) and the AIDJEX model (Coon, 1980), which are both used for the Arctic. A Swedish Baltic sea ice model has been developed by Udin and Ullerstig (1976).

3.2. *Numerical solution*

The system of equations (19–22) is solved in a Cartesian grid using the finite difference technique. The grid size is 15′ in latitude and 30′ in longitude (approximately 27 km) and the time step is 6 h. The advection terms are treated through taking directed derivatives in the direction where the ice is coming from, as shown in Doronin (1970). E.g. let q be any mass component. Then, using the approximation $\Delta x \approx \Delta y$,

$$\begin{aligned}\mathbf{v}\cdot\nabla q = \frac{1}{\Delta x}\{&u(i,j)[q(i-\operatorname{sign}(u(i,j)),j)\\ &-q(i,j)] + v(i,j)[q(i,j-\operatorname{sign}(v(i,j)))\\ &-q(i,j)]\},\end{aligned} \tag{26}$$

where $(u, v) = \mathbf{v}$ and sign $(x) = 1, 0$ or -1 for $x > 0$, $x = 0$ or $x < 0$, respectively. Attention must be paid to the discontinuities of ice mass components. Problems arise when, e.g. ice is drifting towards open water. Then the advection of ice thickness and ridges is treated as simple transport, i.e. they are taken from the grid point(s) where the ice is coming from, and the amount of transported ice is reflected

Table 1. *The parameters in the ice drift model*

Parameter	Notation	Value	Reference
Shear viscosity of ice	η	from $5\cdot10^8$ kg s^{-1} (middle winter) to $5\cdot10^7$ kg s^{-1} (late spring)	numerical experiments with the model; Leppäranta and Udin (1981)
Bulk viscosity of ice	ζ	as η for $\nabla\cdot\mathbf{v} < 0$; 0 for $\nabla\cdot\mathbf{v} \geq 0$	
Density of ice	ρ_i	910 kg m^{-3}	well-known
Ridge mass coefficient	α	50	Leppäranta (1979)
Ridge height coefficient	β	1·2	Leppäranta (1981) (a rough estimate)
Maximum thickness for rafting	h_{cr}	10 cm	Parmerter (1975) (adapted)
Density of air	ρ_a	1·3 kg m^{-3}	well-known
Drag-coefficient for the surface wind	c_a	$1\cdot5\times10^{-3}$	Joffre (1978)
Density of water	ρ_w	10^3 kg m^{-3}	well-known
Drag-coefficient for the geostrophic current	c_w	$2\cdot5\times10^{-3}$	Leppäranta (1980)
Boundary layer angle in the water	θ_w	16 degrees	Leppäranta (1980)
Coriolis parameter	f	$1\cdot3\times10^{-4}$ s^{-1} (60° N)	well known

in concentration on the basis of conservation of mass. Formally this means the use of eq. (12) with h = constant. The numerical technique gives satisfactory solutions in a time-scale of a few days. In a longer time-scale, numerical diffusion, especially at discontinuities, causes problems. This could be solved by including in the discrete model the first order moments of the spatial distribution of ice mass in each cell. Such a technique, however, demands larger memory space from the computer than is at present available.

At the boundary between pack ice and open water, it is assumed that ice velocity is constant, which implies that ice stress Σ is zero. This assumption is often used but in general it would be more realistic to set only the normal stress equal to zero. At the boundary of fast ice and pack ice, ice velocity is put equal to zero or constant, if the pack ice is drifting towards or away from the boundary, respectively. The model needs as input data the initial situation for the wind field. Then the time-evaluation of ice velocity field and mass components can be calculated.

In practice the initial ice data are obtained from the routine observation program of the Ice Service. For the wind field, the estimates of regional surface winds given by the Finnish Meteorological Institute are used. These do not cover the southern Baltic Sea and hence during severe winters the surface winds must be calculated from the surface atmospheric pressure fields. An example of initial ice data, wind field and calculated ice velocity field is shown in Figs. 1 and 5. In Fig. 5a the regions are shown for which the surface wind estimates are given. The calculated ice velocity field shows strong eastward–northward motion which agreed well with observation.

3.3. *Objective verifications of the model*

Usually verifications have been made on the basis of observed changes in ice concentration. The changes in the other mass components are generally not known accurately enough to be used in verification. However, it is known through experience that the model gives the amount of ridged ice qualitatively correctly. Checking of the ice concentration shows that model predicts well the overall displacements in the ice field. Hence forecasts for opening and closing of leads can be considered reliable.

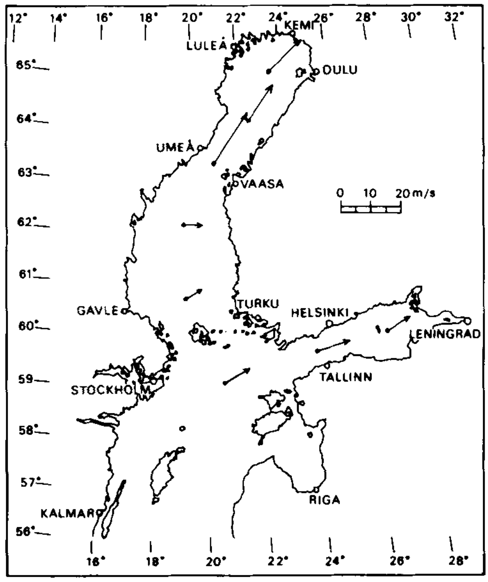


Fig. 5a. Estimated regional surface winds on February 27, 1979 06 GMT (The Finnish Meteorological Institute).

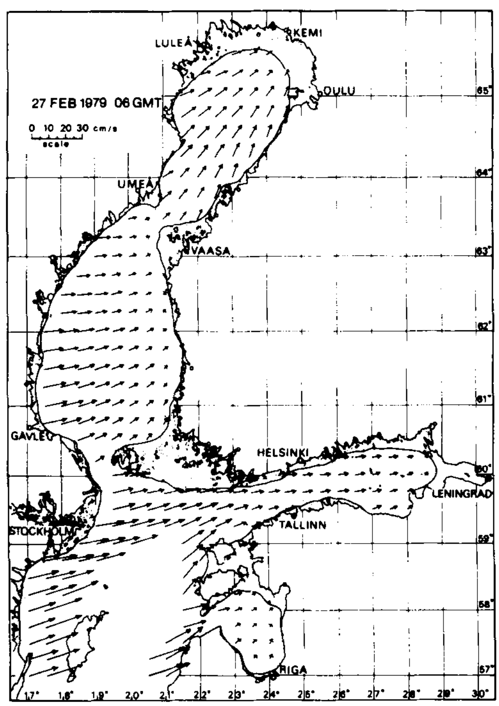
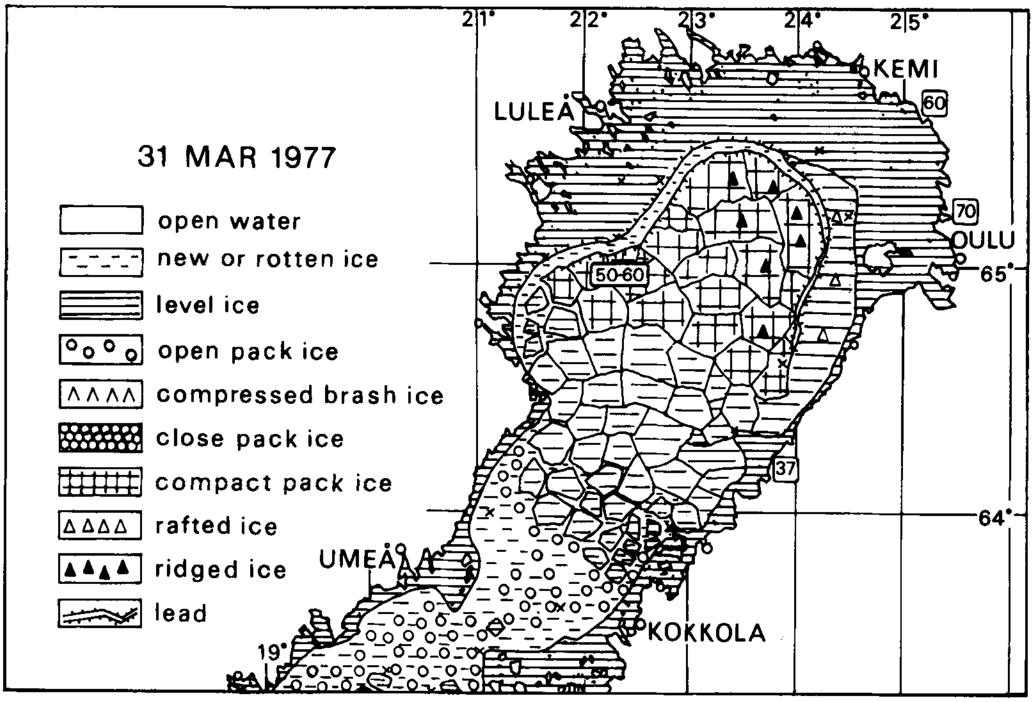


Fig. 5b. Ice drift on February 27, 1979 06 GMT calculated with the model.



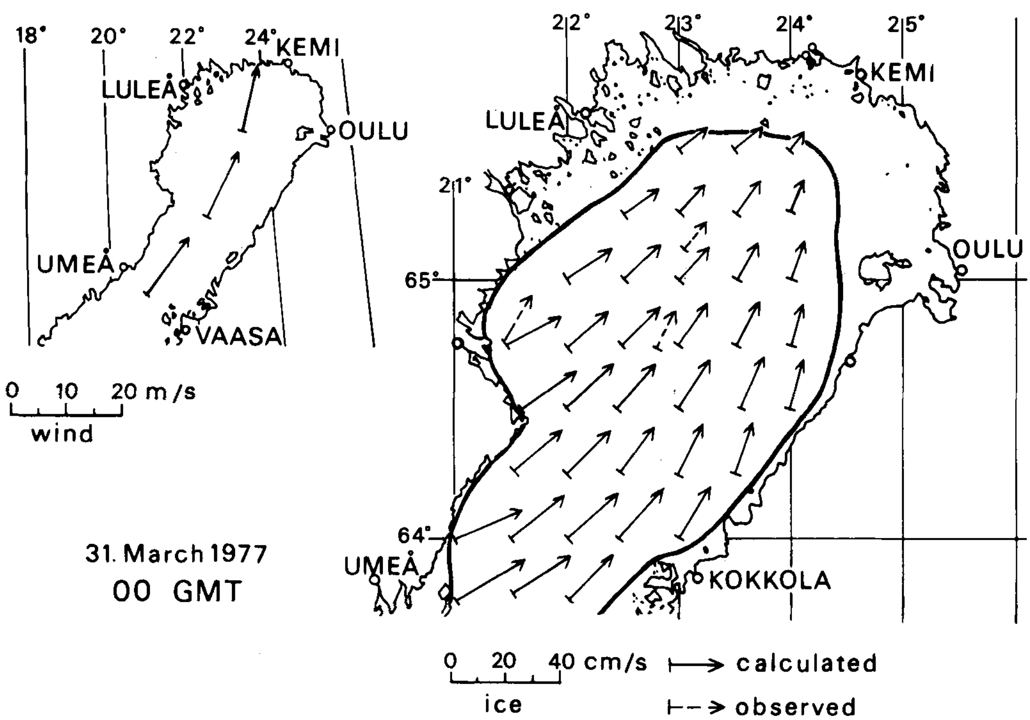


Fig. 6. Verification of calculated ice velocities.

Verification of the ice velocity field is a more sensitive test, but the lack of quantitative data imposes restrictions. There is one period of 24 h, at the end of March 1977, when three ships were moored to ice floes in the Bothnian Bay and their drift was measured with the ships' Decca-apparatus (Leppäranta, 1980). The model calculations agree reasonably well with the observations in the Central Bothnian Bay, but there is a deviation of 35 degrees in drift direction near the west coast (Fig. 6).

Since thermodynamics is not included in the model, the time scale for which the model can be used is limited to a few days. In mid-winter and in late spring, when the thermal changes of ice mass are rapid, the model can fail in 1–2 days in extreme cases.

4. Operative use of the model

4.1. *Routine in daily forecasting*

The links in daily routine are the icebreakers in the Baltic Sea, the Ice Service and the Finnish Meteorological Institute (Fig. 7). The icebreakers send information on ice conditions to the Ice Service. Combining that with other information the Ice Service produces daily charts such as shown in Fig. 1. The correct ice situation is digitized every morning to serve as the initial data for the model. The Meteorological Institute gives a 30 h prediction for the regional surface winds with the temporal resolution of 6 h (Lange, 1973). The regions are the same as shown in Fig. 5a. Then a 30 h ice forecast is calculated. As forecast output,

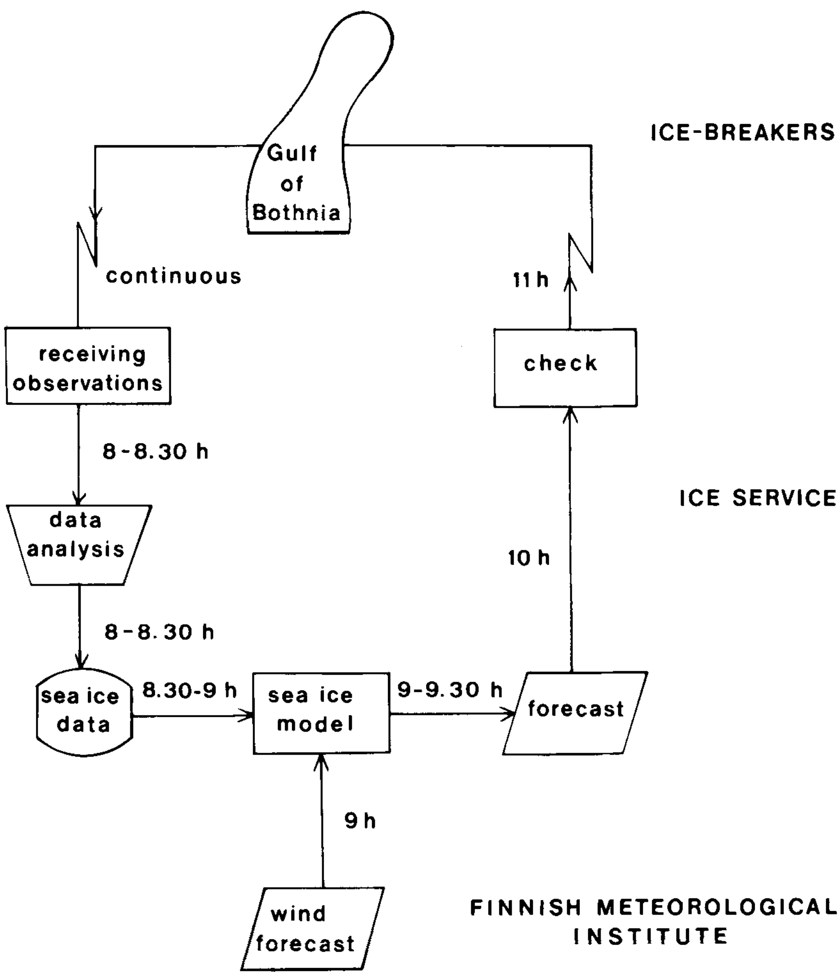


Fig. 7. Operative scheme for daily ice forecast. Time is given as local time.

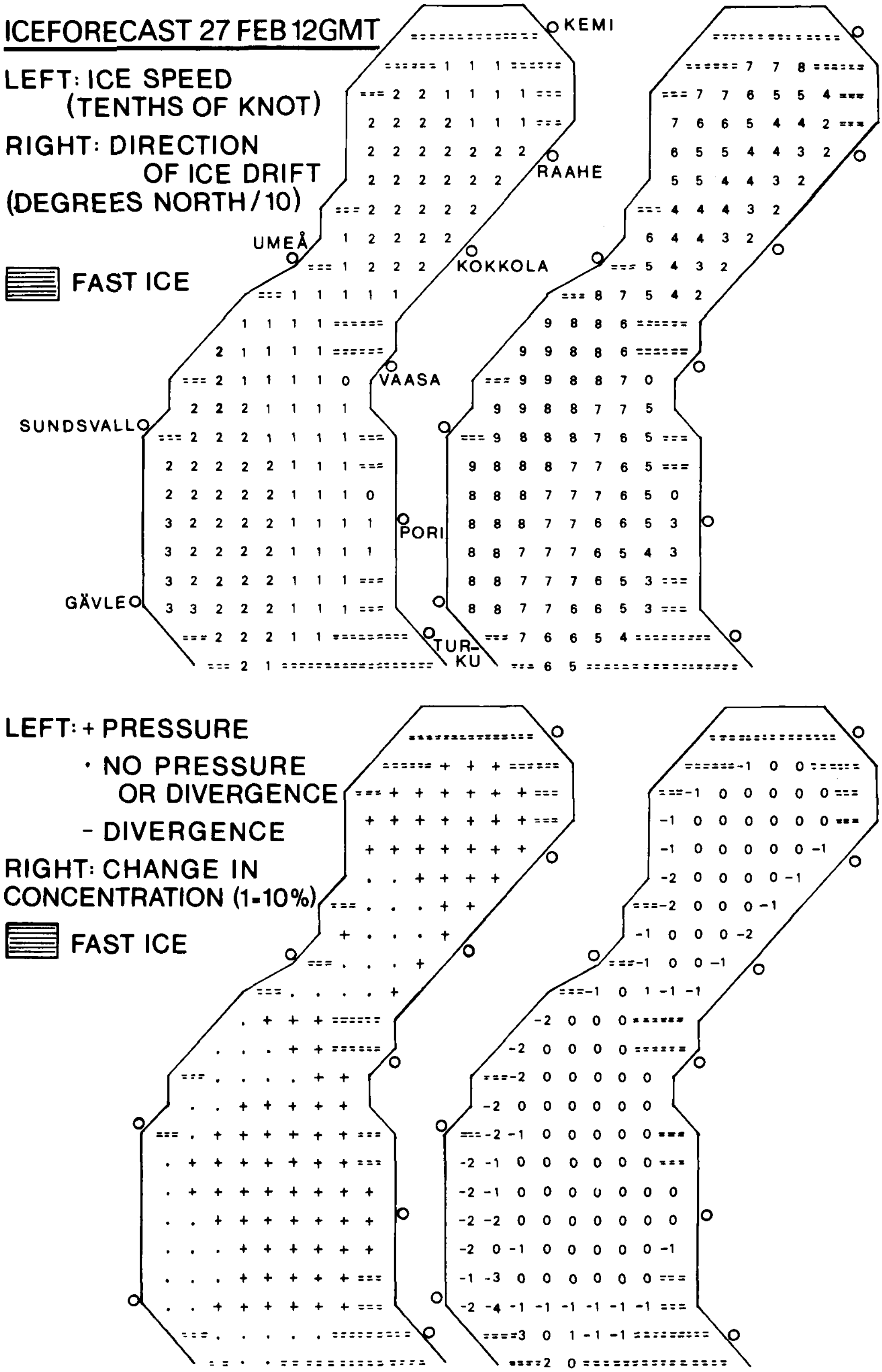


Fig. 8. Ice forecast chart.

ice velocity field, pressure and change in ice concentration are listed (Fig. 8) for two points of time. The forecast is manually checked in the Ice Service and then sent by facsimile to the ice-breakers. In average winters, the daily routine is performed throughout January–May.

The forecast period is restricted to 30 h due to the reliability of the meteorological forecast. Simulations with real wind data show that the ice drift model, in March–April situations when thermal changes of ice mass are slow, could give reasonable predictions for 1 week, if such a wind forecast were available.

4.2. *Subjective verification of the forecasts*

After the first winter, when the model was in operational use, a written inquiry was sent to the icebreaker officers to seek their opinions on the necessity and accuracy of the forecasts (Table 2). The score scale was as follows. Necessity: 1—unnecessary, 2—a little necessary, 3—necessary, 4—very necessary. Accuracy: 1—completely wrong, 2—bad, 3—satisfactory, 4—good. From the total of 60 inquiries 20 were returned; 15 and six gave their opinion on the necessity and accuracy, respectively. It is notable that ice drift direction was considered very necessary, whereas the change in ice concentration was considered the least necessary. This is probably due to the fact that on a time-scale of 30 h, the icebreaker officers can themselves estimate rather well the change in ice concentration.

In the winter 1978/79 checks were made daily in the Ice Service (Table 3). The score scale was: 1—completely wrong, 2—no more useful, 3—useful and 4—essentially correct. Verification was done by two officials (I) and (II) of whom (I) had several years experience whereas (II) was on duty for his second year only; the more experienced gave lower scores. The number of cases was 19 and 22 for (I) and (II), respetively. To make comparisons, a general view of the fit of persistence was asked from the officials after the winter, and again the more experienced one gave lower scores.

Table 2. *Results from the written inquiry sent to the ice-breaker officers in 1977. Mean values for the necessity and fit of ice forecasts*

	Drift speed	Drift direction	Pressure	Change in concentration
Necessity	3·1	3·9	3·1	2·6
Fit	2·7	2·8	3·0	2·2

Table 3. *Verification of ice forecasts made by the Ice Service in the winter 1978/79. Mean values for the fit of ice forecasts and persistence for officials I and II*

	Drift speed	Drift direction	Pressure	Change in concentration
I				
Forecast	2·8	2·9	2·5	2·7
Persistence	2·2	2·3	2·3	2·5
II				
Forecast	3·3	3·6	3·3	3·2
Persistence	3·3	3·1	2·5	2·6

5. Concluding remarks

The dynamical model for ice drift in the Baltic Sea presented here is suitable for short-term simulation and prediction except when thermal changes of ice mass are very rapid. The physical principles on which the model is based are the conservation of mass and momentum, linear-viscous constitutive law for sea ice fields on the geophysical scale and decomposition of ice mass into concentration, thickness, ridge density and ridge size.

The important external factors affecting the momentum of sea ice fields are the surface stresses of wind and water on ice and the Coriolis force. Currents beneath the boundary layer in the sea are neglected; this can cause severe relative errors during periods of low winds. Although the constitutive law is rather simple, lead opening and amount of ridging can be predicted quite well. These are, however, quantities of a time integral character and a more sensitive test involving the fields of deformation rates could reveal deficiencies in the model. Coupling the ice drift model with a thermodynamic model and possibly a sea current model is needed to extend the ability of simulation to more than a few days.

The model has been used for operational 30 h prediction in the Gulf of Bothnia. The results have been satisfactory. The daily routine has worked well thanks to good communication between the Ice Service, which produces daily ice charts, the weather forecast centre and the ice forecast centre.

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МОДЕЛЬ ДВИЖЕНИЯ ЛЬДА В БАЛТИЙСКОМ МОРЕ

Представлена динамическая модель для моделирования и прогнозирования движения ледяных полей в Балтийском море. Основными уравнениями модели являются уравнения законов сохранения массы и количества движения, а также описывающие распределение ледяной

массы. Движение льда рассматривается как вызванное ветром; морские течения ниже слоя трения не учтены. Материальная сторона поведения льдов описана линейным уровнением вязкой структуры. Масса ледяных полей характеризуется концентрацией, толщиной ледяного покрова, а также, количеством и высотой торосов. Физические парпметры модели оценены на основе результатов полевых исследований Валтийского моря. Система уровнений решается в цифровом виде методом конечных разностей при чем шаг сетки составляет 27 километров и шаг времени 6 часов. Модель прменяется с 1977 года оперативно при составлении прогнозов движения льдов на Ботническом заливе. Ежедевно составляется прогноз на 30 дней, который в виде карты направляется ледоколам.