MODELLING

OF THE

SEASONAL ICE COVER

OF THE

BALTIC SEA

Jari Haapala

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Modelling of the seasonal ice cover of the Baltic Sea

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Abstract

The sea ice is an important factor in the climate systems of the polar and sub-polar regions. Sea ice may be regarded as a thin and rather rigid film of complex morphology at the ocean-atmosphere interface. The landscape of the ice covered sea is a mixture of open water, level ice and several different types of deformed ice. Ice cover affects both atmospheric and oceanic circulation and largely modifies or even eliminates the ocean-atmosphere heat, radiation and momentum fluxes. This thesis is about the physical description and numerical investigations of the Baltic Sea ice conditions on a seasonal scale.

Time-series analysis of the ice freezing date, thickness, extent and break-up data show that there is a large natural variability in the Baltic. The length of the ice season has shown a decreasing trend and also the probability of annual ice occurrence in Utö is decreasing. Ice thickness and the Baltic Sea annual maximum ice extent do not show clear trends. The minimum season length and ice thickness are 154 day and 48 cm in Kemi, which are very far from those of an ice free season.

A physically based numerical model for the Baltic Sea ice season has been developed. The model takes into account the ice dynamics, thermodynamics and the ice thickness redistribution and it has been coupled to the simple ocean model. Physical ice classes are used to describe the ice thickness distribution. In a model pack ice is decomposed to open water, two different type of undeformed ice, and to rafted, rubble and ridged ice classes. Inherent evolution equations for each ice class are used and mass flux between the ice classes is calculated according to the ice compactness, thickness and velocity divergence dependent functional form. The main advance of the model is a separation of the thermally and mechanically produced ice. Also, the model allows detailed calculation of the exchange of heat and momentum between the atmosphere/ice/ocean interface.

The model has been verified against the observed ice thickness, concentration and velocity data. It has been shown that the main characteristics of the ice season could be produced by the model. Beginning from the same initial conditions and using the observed meteorological forcing, the model simulated realistic annual cycles of sea surface temperature,
ice thickness and coverage and described the inter-annual variability of the ice season well. Also the sub-basin and inter-basin ice characteristics were realized by the model. Deformed ice production was related to storm activity. Most of the deformation was produced at the coastal zone, which is also an important region for thermodynamically produced ice because of the ice growth in leads. Modelled mechanical growth rates of ice were 0.5-3 cm/day on a basin scale which are close to the thermodynamic ice production rates. The deformed ice fraction was 0.2 in mid-winter and increased during the spring to 0.5 - 1.0.

**Key words:** numerical modelling, sea ice, climate, Baltic Sea
1 Introduction

This thesis is about sea ice, the Baltic Sea and the climate. The focus is on the physical description and numerical investigations of Baltic Sea ice conditions on a seasonal scale. Sea ice may be regarded as a thin and rather rigid film of complex morphology at the ocean-atmosphere interface. The landscape of ice covered sea is a mixture of open water, level ice and several different types of deformed ice. The ice thickness variability in a region is a result of the ice motion and varying age of the sea ice, and the life history of the ice pack of the whole season can be seen in the apparent ice conditions (Figure 1). The evolution begins from a cooling of the ocean surface layer. The thicker level ice floes originate from the initial freezing of sea water and the thinner level ice have formed in fractures or leads, formed during a divergent motion sea ice. When the ice field experiences convergent motion it forms different types of deformed ice (Figure 2). The ice floes may override one on top of another and form rafted ice or a broken ice field may form stripe like formations called ridges. In some circumstances pack ice forms rubble fields which are a conglomeration of small ice pieces oriented randomly but comprising of a fairly constant thickness layer of ice. The life history of the ice pack ends in the melting of different ice types according their inherent melting rates.

The ice pack is an important factor in the climate system of the polar and sub-polar regions. The surface properties of the ice, especially its albedo, temperature and roughness, are different than those of open water and sea ice largely modifies or even eliminates the ocean-atmosphere heat, radiation and momentum fluxes. Ice cover directly affects both the atmospheric circulation patterns due the surface albedo and the insulation effect and to the ocean thermohaline circulation via the salt and fresh water fluxes. An ice pack also holds and advects fresh water, heat, atmospheric settling and sediments, and may release them far away from their original source.

Numerical modelling of the Baltic Sea ice conditions began in the early 1970's. Unisitalo (1973) developed a numerical model for the estimation of the heat flux from ocean to ice and calculated the seasonal cycle of the oceanic heat flux. The major activities in the 1970's were the development of sea ice models for the forecasting purposes. In Sweden, Udin and Ullerstig (1973, 1976) developed a dynamic sea ice model. The Udin and Ullerstig (1976) model was a very advanced model at that time. They assumed that an ice field behaves as a viscous fluid when the ice is thin or its compactness is low, in other conditions they assumed that ice behaves as a plastic material. They describe the internal friction of ice with two terms, a viscous part was taken into account in a similar manner as in the Doronin (1970) model. The plastic part was parametrized according the ice thickness, compactness, wind stress and distance from shore dependent form. Also, they paid attention to the deformation processes of ice and to the numerical realization of the model. A comparable model was developed also in Finland by Valli and Leppäranta (1975).
Figure 1: The life cycle of the pack ice. The evolution is characterized by the cooling of the ocean surface, freezing of the sea water, compression and opening of the ice pack, melting of ice and warming of the surface layer. $F_{\text{tot}}$ denotes the total heat flux at the ocean/ice/atmosphere interface and $\text{div}(u)$ the divergence of the ice pack. The SST, $A$, $h$ and $h_{r}$ denote the sea surface temperature, ice compactness, level ice thickness and ridged ice thickness and the arrows indicate whether the variable is increasing or decreasing.

The operational use of that model begin in winter 1976/77 (Leppäranta, 1977) and the final version of the model is presented in Leppäranta (1981a). In that model the physical behavior of ice was assumed to follow the linear-viscous law (Glen, 1970). The main advance of the model was a proper description of the formation of the open water and deformed ice i.e. the redistribution of ice thickness.

The development of the sea ice forecasting models did not continue until the 1990’s when Leppäranta and Zhang (1992) implemented the viscous-plastic model of Hibler (1979) to the Baltic. Zhang and Leppäranta (1995) coupled the ice model to the storm surge model and they demonstrated clearly how water elevations are reduced due to the internal friction of the ice pack. That model is the present operational sea ice model in Finland (Bai et al., 1995; Cheng et al., 1999). In Sweden, a hybrid model was developed for forecasting purposes (Omstedt et al., 1994, Omstedt and Nyberg, 1995). In that approach the Zhang and Leppäranta (1995) model was coupled to the ice-ocean box model of Omstedt (1990). An other forecasting model was developed by Kleine and Skylar (1995) and Eigenheer and Dalin (1998).

The sea ice forecasting models were developed for short term simulation of the ice conditions. The first climatological ice models were concentrated on the modelling of the thermodynamical growth of ice (Leppäranta, 1983; Stössel, 1985). The pioneering work for climatological modelling in the Baltic Sea was done by Omstedt (1990). He developed a box model for the Baltic Sea. In this model the Baltic was
divided into varios subbasins. The vertical structure of the temperature and salinity was calculated in detail and the horizontal advection of heat and salt was diagnostically calculated. The ocean model was coupled to a one dimensional ice model. The model was further developed by Omstedt and Nyberg (1996). Due the simplicity of the model it allowed the inter-annual simulation of the ice conditions. Omstedt and Nyberg (1996) did decade scale integrations and showed that the ice conditions are largely controlled by the atmospheric forcing and even minor changes in the air temperature can lead to large changes in the ice extent. Haapala and Leppärinta (1996) presented the first seasonal simulations of the Baltic Sea ice conditions where the evolution of sea ice was calculated in two dimensions. Their model was based on the Hibler (1979) viscous-plastic rheology, the Semtner (1976) thermodynamic model and the Leppärinta (1981a) ice thickness redistribution schema. The ice model was coupled to a simple ocean model. The description of the ice thickness distribution and redistribution was further developed by Haapala (2000). Recently, the model was used for estimation of the present and the future ice conditions in the Baltic. Tuomenvirta et al. (2000) did four ten years simulations where the model was forced by the Rossby Center regional atmospheric climate model (Rummukainen et al., 1999). The climatological ice conditions and the interannual variability were realistically produced by the model but the production of the deformed ice was underestimated (Haapala et al., 2000) due the underestimation of the surface winds in the forcing data.

The most advanced Baltic Sea models are the models where the three-dimensional primitive equation ocean model is coupled to the two-dimensional dynamic thermodynamic ice model. The first results were presented by Lehmann and Krauss (1995).
They coupled the high resolution free surface ocean model (Lehmann, 1995) to the viscous-plastic ice model (Hibler, 1979; Stössel and Owens, 1992). Due the large computational requirements of the model the annual simulation has not been available until recently (Lehmann and Hinrichsen, 2000). A comparison of the Lehmann and Hinrichsen (2000) and Haapala (2000) model is presented in Krauss (2000). Both models were forced by the same atmospheric data and the model parameters were similar in both models. Comparison with the observed ice conditions shows that both models reproduce realistic annual evolution of the ice pack. The high resolution model (Lehmann and Hinrichsen, 2000) generates ice tongues and other detailed features on the ice edge, but surprisingly the subbasin features of the ice pack are better represented by the coarser resolution model of Haapala (2000). In addition, the coarser resolution model generated more mobile ice conditions than the high resolution model.

Other high level models are presented by Schrum (1997) and Meier et al. (1999). Schrum (1997) coupled the three-dimensional shelf ocean model (Bachhaus, 1985) to an extended version of Leppä rantan and Zhang’s (1992) model. Meier et al. (1999) approach was to develop a numerical model suitable for parallel computing. They coupled the highly optimized primitive equation ocean model (Webb et al., 1997) to the elastic-viscous-plastic ice model (Hunke and Dukowicz, 1997). Meier (1999) presents the results of the 13-year hindcast simulation and shows that evolution of the ocean temperature and salinity fields and the ice conditions are generally well reproduced by this model.

The summary of my thesis is organized as follows. In this introductory chapter I have described the motivation of the sea ice modelling and have given a review of the Baltic sea ice modelling activities. The background and objectives of this thesis are summarized in the next sub chapter. In chapter 2 I will give an overview of the physical characteristics of the pack ice in the Baltic. Chapter 3 presents the physical principles behind the observed characteristics and discusses how the physical laws are described in the numerical models. Finally, the results of the published articles are summarized and suggestions for further studies are given in chapter 4.

1.1 Background and objectives

This thesis is a product of the two research projects funded by the Academy of Finland (1993-1995) and EC/MAST III programme (1996-1999). When this work begin modelling studies of the Baltic Sea ice conditions were concentrated on the sea ice forecasting (Leppä rantan, 1981a; Stössel, 1985; Leppä rantan and Zhang, 1992; Omstedt et al., 1994) and on climate studies with the one-dimensional box model (Omstedt, 1990) and hence the actual modelling work started from the scratch. The objectives of the project were
• To develop a climatological sea ice model for the Baltic Sea
• To verify the model results
• To examine the variability of the present ice conditions in the Baltic
• To examine future ice conditions in a Baltic

The development and verification work of the model are partly overlapping. The developed model and the verification of the modelled ice thickness and concentration fields are presented in paper II. Verification of the modelled ice velocity fields is presented in paper I. The verified model was used to produce estimates for the future ice conditions based on climate change scenarios. Those results and the analysis of the observed long term variability of the ice conditions are presented in paper III.

The first stage model development work led to an ice climate model comparable to the state of art models of the Arctic i.e. ice dynamics calculated according the Hibler (1979) viscous-plastic rheology, ice thermodynamics based on the Semtner (1976) simplification of the heat conduction law and the ice model was coupled to a simple ocean model. The main advance of the Baltic Sea ice model was that it took account into the redistribution of the ice according the Leppäranta (1981a) scheme. As usual, further developments of the models were found to be necessary. Particularly, a better description of the ice thickness distribution was suggested to accomplish the intended improvements in the model results (Haapala and Leppäranta, 1996, Leppäranta et al., 1998). Modelling of ice thickness distribution is the overall objective of this thesis. It may expressed simply as

• To model the sea ice mass

Although it sounds trivial, the determination of the total sea ice mass in a basin scale has not been solved accurately neither by the observational techniques nor the present ice models because of the uncertainties in determining the ice thickness and the deformed ice portion. Studies of this question have led to a revised Baltic Sea model which is presented in paper IV.

In addition of papers I - IV some other publications relates to this thesis. Before this project the author was working in the Finnish Institute of Marine Research and analysed varios Baltic Sea temperature and salinity observations. The hydrographic data base (Haapala and Alenius, 1994) has been used for the model initialization and timeseries analysis results (Alenius and Haapala, 1992; Haapala and Alenius, 1993) have been used in the interpretation of the long term variations in the ice conditions (paper III). The database for the forcing and verification data used on
papers II and III was collected in collaboration with other scientist working in the Baltic (Haapala et al., 1996).

1.2 Author’s contribution

The author is fully responsible for paper IV and is the lead author of the papers II and III. In all papers the author has been responsible for the development and implementation of the numerical model. The lead author of paper I was Matti Leppäranta and an analysis of the model simulations was done together with him. Yan Sun provided the ice motion analysis from SAR images and participated in discussions concerning the results. The author’s contribution for the paper I was about 1/3 of the total work. Papers II and III were mostly written by this author. In paper II, Matti Leppäranta participated on the model results analysis work. In paper III Matti Leppäranta did the statistical analysis of the timeseries (Table 1 in paper III) and the analytical modelling work.

2 Baltic Sea ice characteristics

The Baltic is located between the North-Atlantic and Eurasian continental weather systems which is the main reason for the large inter-annual variability of the ice conditions. The maximum annual ice extent of the Baltic (MIB) is rather well correlated to the North Atlantic Oscillation (NAO) winter index (Figure 3). Westerlies bring moist and warm air masses over the Baltic and hence the apparent ice conditions are mild. On the contrary, when the continental weather system prevails the ice season is regarded to be severe. The natural variability of the ice conditions is illustrated in Figure 1 in paper III. The other climatological characteristics of the ice season are not as well related to the NAO-index as the MIB because the NAO and MIB may regarded as the integrals of the winter season, but the ice thickness, freezing and melting dates are more dependent on the local weather conditions (air temperature, precipitation, wind speed and oceanic heat flux). In a climatological sense, the most relevant parameter which describes the ice conditions is the total mass of ice. However, it is most challenging task to to determine the mass of sea ice and an attempt to approach the problem is presented in paper IV.

In the Southern Baltic ice does not occur annually (SMHI and FIMR, 1982). In the Skagerrag and the coastal areas of Germany and Poland the probability of the ice occurrence is 25 - 75 % and the thickness of ice varies from 10 to 50 cm (Girjatowitz,1990; BSH, 1994). Figure 4 shows the ice conditions in the Belt Sea on 10 March 1970. In the western side of the image initially frozen and rafted ice is visible. In the eastern side an ice eddy is apparent. The compactness of the ice field was
Figure 3: Time series of North Atlantic Oscillation index (Rogers, 1984) (line) and the annual maximum ice extent of the Baltic Sea (MIB) (Seinä and Palosuo, 1993) (bars). The MIB is presented as an anomaly of the normalized time series.

Figure 4: A satellite photograph showing the ice conditions in the Southern Baltic Sea on 10 March 1970. The coastline extends from the Fehmahr to Rügen. The photo was taken by U.S. photo-reconnaissance satellite system CORONA

low and hence the internal stress of the ice pack was low. Probably, also the wind stress was low and the main forcing for the ice momentum balance arose from the water stress and because of that the ice field acted as a tracer for an oceanic eddy.

The ice condition are harsher in the Northern Baltic every winter (Jurva, 1937; Palosuo, 1953; Palosuo, 1966). Even in mild winters the ice thickness is 30 - 50 cm in the Bay of Bothnia and the deformed ice production can be extensive. The typical horizontal variation of the ice field is shown in Figure 5. The compact ice field is a mixture of the different types and age of ice. In the middle of the figure is a refrozen lead where part of the ice has been rafted. Ridges and thicker undeformed ice are apparent in the older ice field.
Figure 5: An aerial photograph illustrating the variable surface characteristics of the pack ice. The diameter of the figure is about 1850 m and it was photographed in the Bay of Bothnia on 14 March 1997.

Figure 6: Vertical profile of the ridged ice (Kankaanpää, 1997).

Variations of the deformed ice properties in a vertical space has not studied much in the Baltic. The typical cross-section of the ridged ice field is shown in Figure 6 (Kankaanpää, 1997). The ridges have a clear sail and keel which are rather triangular in shape. The thickness of the ridges in the Baltic is typically 5 - 15 m, and the maximum thickness is about 30 meters (Palosuo, 1975; Leppäranta and Hakala, 1992; Kankaanpää, 1997). The ridge density varies much depending on the location. Close to coastal regions it can be larger than 10 km\(^{-1}\), typically it is 1.4 to 9.5 km\(^{-1}\) (Lewis et al., 1993; Lensu, 1995; Lensu, 1998). The fraction of the deformed ice has been estimated to be about 0.3 in the Baltic (Lewis et al., 1993).

Ice velocities in the Baltic are known to experience free drift speed (Leppäranta and Omstedt, 1990) to highly immobile conditions (Leppäranta, 1981b). Recent
observations of ice velocity in the Bay of Bothnia give a more detailed view of the
ice kinematics in highly compact ice condition (Leppäranta et al., 2000). In the
Baltic, the main factors in the ice momentum balance are the wind stress and the
internal stress of ice pack (Leppäranta, 1981b). Generally, it is known that the
internal stress plays the major role when the compactness of the ice pack is high.
It is negligible when the compactness is $< 0.8$ and then the ice velocity is close
to the free drift velocity. According the plastic law the internal stress of ice is
linearly proportional to the ice strength (which is related on the ice thickness) and
the wind stress is proportional to square of the wind velocity. This implies that in
addition to low compactness situations there is a possibility for the ice velocity to
reach the free drift velocity if the wind stress is considerably larger than the internal
stress. To examine this phenomenon the ice velocity magnitude was plotted against
the observed wind magnitude (Figure 7). The free drift speed was plotted for a
reference. It clearly depicts the ice pack is under immobile conditions when weak
and moderate winds are acting. The ice pack moves when the wind speed is 10 -
15 m/s but at considerable lower speed than would be predicted by the free drift
law due the high internal stress of the ice pack. But, during very strong winds the
observed ice velocities are closer to the free drift speed. The same analysis for the
turning angle (difference between the ice and wind directions) shows no dependence
between the turning angle and wind speed when the wind speed is $< 15$ m/s and
during stronger winds the turning angle seems to reach a constant value such as the
free drift law assumes.
The abovementioned observation of the structure and behavior the Baltic Sea ice conditions have led to several requirements for the numerical sea ice model. The model has to be able

- To simulate the annual cycle of the areal extent and ice thickness in a basin scale.
- To describe the subbasin ice characteristics.
- To simulate the inter-annual variation of the ice conditions.
- To take into account the deformed ice portion.
- To simulate ice motion from free drift to immobile conditions.

3 Modelling

In geophysical modelling, sea ice is assumed to be a continuum of several ice floes. The apparent ice conditions are a result of the thermodynamic and dynamics processes acting in the ice pack (Figure 8). In a closed system the total mass of sea ice is changed only due the freezing and melting of the ice. Dynamics is responsible for the mechanical growth of sea ice but during the deformation processes the total mass of ice remains the same. Indirectly, dynamics have an influence on the ice mass balance due the new ice production in leads. Dynamics also changes the surface characteristics of ice which alter on the heat exchange between the ice/atmosphere and ice/ocean interfaces. The ice thickness distribution has a direct effect both on the thermal and mechanical growth of ice. The thermal growth rate of sea ice is dependent on the floe thickness of ice and the internal stress of ice pack highly depends on the amount of open water, size of the floes and the floe thickness of the ice.

The engine for the sea ice evolution is the heat and momentum exchange between the ocean and atmosphere. Without any external forcing the sea ice system would not be in motion because there are no processes between the ice thermodynamics and dynamics which could maintain any motion like the ocean thermohaline circulation. Thermodynamic growth and the decay of ice is a result of heat loss from the ocean to the atmosphere and absorption of the radiative fluxes inside the ice. In addition, freezing and melting of ice have a direct effect on the ocean thermodynamics via the salt and fresh water fluxes. The ice dynamics is mainly driven by the wind and ocean stresses.
3.1 Ice thickness distribution

The elementary variables of the ice pack are the ice thickness ($h$) and ice compactness ($A$). Ice characteristics in a scale larger than the typical length scale of individual features are described by the ice thickness distribution function ($g(h)$, Thorndike et al., 1975). It is defined as

$$\int_{h_1}^{h_2} g(h) dh = \frac{1}{R} f(h_1, h_2)$$  \hspace{1cm} (1)

where $R$ is the region to be considered and $f(h_1, h_2)$ is the area within the region covered by the ice thickness from $h_1$ to $h_2$. The evolution equation for the $g(h)$ is,

$$\frac{Dg}{Dt} = \Theta + \Psi$$  \hspace{1cm} (2)

where the lhs describes the local change and advection, and $\Theta$ and $\Psi$ are the thermodynamic growth rate and redistribution of ice thickness due the deformations. The ice thickness distribution function can be calculated from several observational data sets (Wadhams, 1998) but only a few numerical model resolve $g(h)$. This is mainly due to difficulties in determining the redistribution function $\Psi$. In principle, $\Psi$ is dependent on $g(h)$ and the strain rate invariants (Thorndike et al., 1975). Numerically eq. 2 is solved by discrete ice thickness on the different thickness categories.
and solving the thermal and mechanical growth rates for each thickness categories separately (Thorndike et al., 1975; Hibler, 1980).

In the classical Hibler (1979) model \( g(h) \) is approximated with two ice thickness categories without any ridging term \( h = h(h_0, H) \), where \( h_0 \) is the thin ice and it is interpreted as open water and \( H \) is the thick ice \( (h \geq 0.5m) \)). The implementation of several ice thickness categories on an entire Arctic model was done by Hibler (1980). Recently, Flato and Hibler (1995) extended the general ice thickness distribution theory to include separate thickness distributions for the ridged and undeformed ice and modelled the entire Arctic ice with 28 thickness categories. In that model the redistribution of ice takes account compressive and shear deformations and the ice strength is calculated according to the energy losses due the deformations. A shortcoming of the Flato and Hibler (1995) implementation is that the model uses equivalent thermodynamics growth rates for the ridged and undeformed ice. Also it assumes that the constant fraction (15 %) of thinnest ice experiences ridging.

Alternatively, \( g(h) \) can be approximated by separating \( h \) and \( A \) in the physical ice classes, in the following manner

\[
h = h(h_a, ..., h_n) \\
A = A(A_a, ..., A_n)
\]

where the subtitles denote the different ice classes (level ice, lead ice, rafted ice, rubble, ridged ice, etc.). The ice class approach have been used by Baltic Sea models (Leppä rantana, 1981a; Zhang and Leppä rantana, 1995; Haapala and Leppä rantana 1996; Schrum 1997) and in the Antarctic model by Harder and Lemke (1994). A model of six thickness categories has been developed by Polyakov et al. (1998). However, in this approach there is not an explicit description of the deformed ice in the model and the redistribution of ice is considered to occur when the model produced ice concentration exceeds unity.

A three-level description (open water, level and ridged ice) was used in papers I - III with the redistribution terms defined by Leppä rantana (1981a). Some shortcomings of the Leppä rantana (1981a) ice redistribution scheme are that the model does not include separate equations for the level ice and deformed ice concentrations and it assumes that ridging occurs only when ice concentration reaches unity during convergence.

In paper IV the description of the deformed ice classes and the deformation processes were further developed. Decomposition of the pack ice is based on the morphological classification of the ice (WMO, 1970). The principal classification is a separation of the pack ice to the undeformed and deformed ice types. Undeformed ice is generally
called level ice and it is furthermore divided according to its phase of development. In Haapala (2000) level ice is used to describe undeformed ice in general. An other undeformed ice class, lead ice, was introduced in order to describe new ice growth. In a model, divergent ice motion decreases ice compactness and new ice growth is calculated if the ice free fraction is frozen. The level ice - lead ice pair is a two-level approximation of the undeformed ice thickness distribution. This approximation enables different thermodynamic growth rates for the thin lead ice and thicker level ice. Numerically the ice growth (both for level and lead ice) is calculated prognostically from the initial thickness of 1 mm.

Deformed ice is separated into rafted ice, rubble ice (or hummocked ice) and ridged ice classes (WMO, 1970). A key question of the ice class approach is a determination of the mass flux between the ice classes. The most common process is that the level or lead ice experiences deformations and produces rafted, rubble and ridged ices. It is also know that rafted ice may be rafted in many times, or rafted ice could form ridged and rubble ices. The unresolved question is in what circumstances rafted, rubble and ridged ices are produced. The deformations begins when the external forces cause ice to fail by flexure, buckling or crushing (Mellor, 1986). Visual observation of the deformed ice field suggest that thin ice predominantly experiences rafting and thicker ice sheets are more often ridged, although also thick ice (h > 1 m) is known to raft (Rothrock, 1979). The Parmeter (1975) model suggest that rafting of an ice sheet is limited by the maximum ice thickness and the cross-over thickness between rafting and ridging relates on the mechanical parameters of the ice. Rubble fields are common close to the coastal region (Kovacs, 1981) but very little data is found about the formation and the detailed morphological characteristics of rubble fields. Parmeter (1975) and Hopkins (1998) suggested that the rubble fields are an extension of ridged ice and rubble are created after ridge formation for as long as convergent motion continues and there is any undeformed ice left.

Recent laboratory and numerical experiments by Tuhkuri and Lensu (1998) and Tuhkuri et al. (1999) have given new insight on the deformations of the sea ice. They simulated deformation processes in an ice tank where an ice sheet was pushed against an other. The major finding of the laboratory experiments was that an ice sheets of uniform thickness did not form ridges but a nonuniform ice sheet generated both rafted and ridged ices. They did not found any cross-over thickness such as suggested by Parmerter (1975) which would determine whether ice sheet is rafting or ridging and actually all ridging events began as rafting events. They distinguish four phases in the development of the ridging events in their conceptual model of ridging (Tuhkuri et al., 1999). Firstly, a force exceeding a threshold value in needed for the initialization of deformation. Then ice sheet begins to raft. Third phase is the keel buildup and the last phase is lateral growth of the ridge. The field observations from Baltic (Tuhkuri et al., 1999) support also that all ridging events begins as a rafting.

Numerical modelling of the deformation processes can be properly modelled only by
the process models (c.f. Hopkins, 1998) which resolve the forces and motion of the ice in a floe scale. Utilization of the process models in a modelling of the seasonal evolution ice pack in the entire Baltic or Arctic region is presently impossible because of the fine time and space discretization of those models. The continuum scale sea ice models resolve an average behavior of the pack ice and the subgrid processes are neglected or taken account in a simplified manner. The following assumptions of the deformation processes in Haapala (2000) model have been done

- Deformed ice is generated only from level and lead ice classes, i.e. rafted ice is not deformed further in the model.

- Cross-over thickness determines whether the undeformed ice is rafted or ridged. This assumption is based on the Parmeter (1975) law and field observations (c.f. Rothrock, 1979)

- Rubble ice is assumed to be formed from lead ice. Via this assumption the model produces rubble fields close to the fast ice regions. These areas are known to be most common regions where the rubble field exist. The horizontal growth of the ridged ice is impossible to take account of in the present model. The model assumes that the rubble fields are formed from an ice sheet which thickness is between the rafting and ridging thicknesses. Via this assumption part of the lead ice mass is converted to the rubble ice and part to the ridged ice during a single deformation event.

- Shape and porosity of the ridges is constant. These assumption are based on the field observations (Timco and Burden, 1997; Kankaanpää, 1997).

- Production of deformed ice and open water due to shear effects are neglected.

The model presented in paper IV describes an average evolution of the pack ice deformation processes. Because the deformation processes are described by the evolution equations of the ice classes and the redistribution functions many deformation processes are possible during a single time step. This mimics the real behavior of the pack ice in a continuum scale. For example, if the compactness of the ice field is 98 % then 37 % of the ice velocity divergence is consumed for the deformations of the ice and rest is used for the compaction of the ice. Also, if lead ice thickness is about 12 cm then 45 % of the deformed fraction of the lead ice is transformed to the ridged ice, 45 % is transformed to the rubble ice and 10 % is transformed to the rafted ice class. On the ultimate ends of the ice thickness space only rafting or ridging could occur.
3.2 Ice thermodynamics

Thermodynamic ice models resolve the vertical structure of the heat and salt within the sea and snow layer. The present models are much based on the Maykut and Untersteiner (1971) model. In the Baltic, extensions of this model has been developed for process studies (Launiainen and Cheng, 1998; Saloranta, 1998; Saloranta, 2000) the primary question is to determine the heat transfer rate inside the ice and the heat exchange between the ice/snow/atmosphere and ice/ocean surfaces. The heat equation inside the ice is,

\[
\frac{\partial}{\partial t}(\rho_i c_i T) = \frac{\partial}{\partial z}(k_i \frac{\partial T}{\partial z}) + q
\]  

(5)

where \( \rho_i \) is the ice density, \( c_i \) is the specific heat of ice, \( T \) is the temperature of the ice, \( k_i \) is the heat conductivity of ice and \( q \) is the internal heat source. The boundary conditions are \( k_i \frac{dT}{dz} = F_{tot} \) at the surface and \( T = T_f \) at the bottom. \( F_{tot} \) and \( T_f \) are the total heat exchange between the ice/snow and atmosphere and freezing temperature of sea water, respectively. Ice growth and melting at the bottom are determined by the balance between the conductive heat flow and the oceanic heat flux

\[
\rho_i L \frac{\partial h}{\partial t} = k_i \frac{\partial T}{\partial z}|_{z=h} - F_{wi}
\]  

(6)

where \( L \) is the latent heat and \( F_{wi} \) is the oceanic heat flux. Equations 5 and 6 can not feasibly resolved in an entire Arctic or Baltic model and some simplification has to done in large scale models. If the internal heat source and diffusion are neglected and heat conduction is assumed to be piecewise linear the formulas are reduced to the Semtner model (Semtner, 1975). A three layer model yields a results close to the Maykut and Untersteiner (1971) model in the Arctic (Semtner, 1984). The equations can be further simplified by assuming that the heat conduction is constant. It reduces to the Semtner 0-layer model which is a valid approximation assuming that the ice is thin (Semtner, 1975). This approach has been used in the all thermodynamic calculations in papers II - IV. The ultimate approximation of the above formulas is to assume that the surface temperature of the ice is equal to the air temperature. This approximation allows an analytic solution called Stefan’s law (Stefan, 1891) and it may used for the description of the overall growth rate of ice (Leppäranta, 1993).

The total heat flux to the surface is,
\[ F_{\text{tot}} = F_{sw} + F_{lw\downarrow} + F_{lw\uparrow} + F_{sh} + F_{ih} \]  

where \( F_{sw} \) is the incoming short wave radiation flux, \( F_{lw\downarrow} \) is the incoming longwave radiation flux, \( F_{lw\uparrow} \) is the outgoing long wave radiation flux, \( F_{sh} \) and \( F_{ih} \) are the turbulent sensible and latent heat fluxes.

The calculation of the heat and radiative fluxes are based on empirical functions and several formulas are available (see the reviews by Launiainen and Bin, 1998; Omstedt, 1998; Guest, 1998). Some uncertainties exist in the parameterization of those fluxes and a standard formulas for neither the Arctic nor the Baltic have been recommended. The flux calculations used in the papers I - IV are based on the comparable Arctic models. The calculation of the papers II and III were done with an identical calculation and a revised heat flux calculation methods were used in paper IV. The change of the calculation method was done in order to provide the calculation with the same methods as the other model groups used in the Baltic (Lehmann 1995; Lehmann and Hinrichsen, 2000; Omstedt and Nyberg, 1996). The formulas used in this thesis are summarized in appendix I.

### 3.3 Ice dynamics

The ice dynamics resolves the momentum balance of sea ice. The motion of sea ice is driven by the surface and bottom stresses due the wind and ocean current and sea surface tilt, and it affected by the internal stress of the ice pack and the coriolis force. A review of the climate research oriented numerical modelling of the ice dynamics is given by Hibler and Flato (1992), a detailed mathematical analysis of the ice dynamics is provided by Pritchard (1988) and by Gray and Morland (1994) and an overall review of the problem is given by Leppärinta (1998). The momentum equation in a horizontal plane is

\[ m \left( \frac{D\vec{u}}{Dt} \right) + f \hat{k} \times \vec{u} = A(\vec{\tau}^a + \vec{\tau}^w) + mg\nabla H + \nabla \cdot \sigma \]  

where \( m \) is the total ice and snow mass, \( \vec{u} \) is the horizontal ice velocity vector, \( f \) is the coriolis parameter, \( \hat{k} \) is the upward unit vector, \( \vec{\tau}^a \) is the air stress vector, \( \vec{\tau}^w \) is the water stress vector, \( g \) is the gravitational acceleration, \( H \) is the sea surface tilt and the \( \sigma \) is the internal stress tensor. According to the scaling arguments (Leppärinta, 1998), the nonlinear advective terms and the sea surface tilt can be neglected in the calculation of the momentum balance.

The determination of the internal stress of the pack is the major question of the above
The simplest assumption is the free drift law, i.e. that there does not exist any internal stress. Locally, it may hold (Nansen, 1902; Leppäranta and Omstedt, 1990) but if it is used in a numerical model it leads to a highly overestimation of the ice velocity and dynamic growth of sea ice (Leppäranta et al., 1998). In the first ice models a linear viscous rheology was used (Campell, 1965; Doronin 1970). Later, a plastic rheology for sea ice was suggested and it led to the development of an elastic-plastic model (Coon et al. 1974) and a viscous-plastic model (Hibler, 1979). The viscous-plastic model became the most widely used ice model in climatological sea ice research while the elastic-plastic model has been used mainly for limited purposes. The constitutive law of Hibler (1979) is,

$$\sigma = 2\eta \dot{\varepsilon} + (\xi - \eta) tr \dot{\varepsilon} I + \frac{1}{2} P I$$

(9)

where $\eta$ is the shear viscosity, $\xi$ is the bulk viscosity, $\dot{\varepsilon}$ is the strain rate tensor, $I$ is the unit tensor and the $P$ is the ice strength.

The viscous-plastic ice rheology implies that under very low strain rates the bulk and shear viscosities are constant and the model produces linear viscous behavior, otherwise the viscosities are calculated according to the plastic flow rule (Hibler, 1979). The ice strength parameter links ice dynamics to ice thickness and compactness as follows

$$P = P^* \bar{h} e^{-C(1-A)}$$

(10)

where $P^*$ is the ice strength constant, $\bar{h}$ is the mean ice thickness over the grid cell and $C$ is the compaction constant.

In the present work equation 8 is solved numerically according to the over-relaxation method of Hibler (1979). A recent development of Zhang and Hibler (1997) and Hunke and Dukovich (1997) provides a faster solution of the momentum balance equation.
4 Conclusion

The Baltic Sea ice condition in the past, present and future have been examined by time series analysis and by a numerical model. The time series analysis has been done in order to study the natural variability of the ice conditions. The numerical model was developed for describing the evolution of the ice pack on a seasonal scale. The model was verified against the present ice conditions and estimates of the future ice conditions in the Baltic were made. The results of papers I - IV can be summarized as follows:

- The Baltic Sea ice time series show large natural variability in the ice freezing date, thickness, extent and break-up. The length of the ice season has shown a decreasing trend and also the probability of annual ice occurrence in Utö is decreasing. Ice thickness and the Baltic Sea annual maximum ice extent do not show clear trends. The minimum season length and ice thickness are 154 day and 48 cm in Kemi, which are very far from an ice free season.

- The main characteristics of the ice season could be produced by the moderate resolution ice-ocean model. Beginning from the same initial conditions and using the observed meteorological forcing, the model simulated realistic annual cycles of sea surface temperature, ice thickness and coverage and described the inter-annual variability of the ice season well.

- The analysis of the ERS-1 SAR derived ice velocities showed a considerable stiffening of the ice pack as the minimum ice thickness increased from 10 to 30 cm. This is due to the change in the character of ice deformation under compression from rafting to ridging. Thin ice compressions (rafting) were one order of magnitude larger than thick ice compressions (ridging). The observed ice velocity field could be produced with the viscous-plastic ice model. In all cases the ice field experienced heavy compression. The results supported the assumption of a plastic rheology for thick (more than 30 cm) and compact ice.

- The strength constant $P^*$ is best represented by the value $2.5 \times 10^4 Nm^{-2} \pm 50\%$. The resulting velocity field was sensitive to $P^*$. Reduction to $1 \times 10^4 Nm^{-2}$ lead close to free drift conditions. The motion dropped remarkably due to doubling of $P^*$ and for larger $P^*$ ice motion soon become a slow creep state. Model experiments suggested that the best $P^*$ varied in different ice conditions. A preponderance of thin ice or leads produces almost free drift motion ($P^* = 0$), while for thick compact mid-winter ice the best value for $P^*$ was $5 \times 10^4 Nm^{-2}$.

- The aspect ratio of the yield ellipse ($e$) was found to be probably within the range 1.5 - 4. The compaction constant ($C$) could not be studied in detail because of the lack of accurate ice compactness data, but the model results
proved that the sensitivity of the strength to compactness is high, which means that $C \gg 1$ and $C = 20$ is reasonable.

- According to the SILMU central air temperature scenario for the Baltic Sea (temperature rise 0.6 °C/decade for winter, 0.4 °C/decade for spring and autumn and 0.2 °C/decade for summer, Carter et al., 1995) the expected mean ice conditions would change considerable by 2050. In an average winter the ice would be formed only in the Bay of Bothnia, the Archipelago Sea, the end of Gulf of Finland and on the Estonian west coast. The freezing date would be shifted to about 20 days later and the break-up date 10 earlier. The ice thickness would become about 20 cm thinner. The model produced a 30-year range for the annual maximum ice thickness of 30 - 65 cm in the Bay of Bothnia in 2050. The length of the ice season varies from 3.5 - 6 months. The latest freezing occurs in January and the ice break-up happens at earliest in April.\(^1\)

- An ice thickness redistribution model has been formulated where the pack ice is decomposed to open water, two different type of undeformed ice, and to rafted, rubble and ridged ice classes. The benefits of the physically based ice thickness distribution model are that ice classes are prognostic variables and specific thermodynamic growth and decay are calculated for each ice class. The main advance of the model is a separation of the thermally and mechanically produced ices.

- Most of the deformation were produced at the coastal zone, which is also an important region of thermodynamically produced ice because of the ice growth in leads. The modelled mechanical growth rates of ice was 0.5-3 cm/day in a basin scale which was close to the thermo-dynamical ice production rates. Deformed ice fraction increased during the season. In the early and mid-winter is was about 0.2 and during the spring in increases to 0.5 - 1.0 due the new deformations and the melting of the undeformed ice.

- The ice class approach gives more information of the surface properties of the ice pack than the widely used 2-level model of Hibler (1979) and its recent developments (Harder and Lemke, 1994). The ice concentration, surface temperature, albedo and surface roughness are the primary factors governing the atmospheric and oceanic boundary layers and since these parameters are explicitly resolved for each ice class it allows detailed calculation of the exchange of heat and momentum between the atmosphere/ice/ocean interface.

\(^1\)NOTE: These estimates were based on SILMU air temperature scenarios which are not valid anymore. According what is now known the increase of air temperature will be about 2.6 °C in the year 2100 from the present level (Rummukainen et al., 1999). The revised estimated of the future ice conditions have been done by Tuomenvirta et al. (2000) and some results can be found Appendix II. The estimates according to the SILMU scenarios for the ice conditions on 2050 can now understood to represent the ice conditions around 2100.
Suggestions for further studies

- Determination of the sea ice mass balance on a seasonal scale is the key question for the sea ice research. As a simple question such as the calculation of the total amount of sea ice in Arctic or Baltic can not solved accurately neither by observational techniques nor the present ice models because of the uncertainties in the determining of the deformed ice portion. Direct and indirect measurements of pack ice thickness are suitable for characterizing the ice extend or the local ice thickness characteristics but can not determine accurately the ice mass of a large region. In principle, numerical models resolve the undeformed and deformed ice mass but many uncertainties exist in a parameterization of the sea ice deformation processes in continuum models. More emphasis has to put on the monitoring of the ice thickness and drift on basin size and seasonal scales. Extensive observational data can be used for model verification and further developments of the sea ice models.

- Many of the sub-grid processes of the model are based on empirical formulas derived from limited observational data or even on ad hoc estimates. More physically based description of the ice internal stress and the redistribution processes may provide more accurate numerical results. For example, the present model utilizes the Parmeter law (1975) for the determination of the cross-over thickness between rafting and ridging. Recent laboratory experiments by Tuhkuri and Lensu (1998) and model studies by Hopkins et al. (1999) suggest that there is not such a cross-over thickness and the good consistency with the field observations is due the non-homogeneity of the ice pack in the continuum scale. Ice tank and field experiments (Tuhkuri et al., 1999) may provide a physically based equations for the calculation of the ice thickness redistribution instead of the presently used redistribution functions.

- The author has tested the calculation of the internal stress according to the suggestion that the weakest ice class determines the ice strength (Leppäraanta et al., 1998). The numerical model generated sharp velocity gradients as recognized from the satellite images but long term simulations were unstable because of the numerical instabilities in the calculation of the ice thickness continuity equations. This is a common problem in finite difference discretization and presently there is no clear solution to overcome these numerical problems.

- Although the air temperature has a direct effect on the ice conditions it is not obvious that warmer climate would led to the milder ice conditions in respect of the ice thickness distribution (Wadhams, 1998). Even if the wind conditions would remain same the production of deformed ice and hence the total mass of sea ice can be larger in mild winters than for average conditions due the strong dependence of the ice thickness and internal stress of the ice pack. Numerical experiments to examine the variability of production of the deformed ice mass should be made to test the above hypothesis.
• In the Baltic Sea several ice models have been used (Omstedt, 1990; Kleine and Skylar, 1995; Haapala and Leppäraanta, 1996; Lehmann and Hinrichsen, 2000; Launiainen and Cheng, 1998, Schrum, 1998, Eigenheer and Dahlin, 1998, Saloranta, 1999, Meier et al., 1999). Model results are not easily comparable because the calculation methods of the surface fluxes have varied from model to model. Also, the parameterization of the albedo and oceanic heat flux, and the calculation of the snow evolution is not similar in the models. To establish the effect of parameterization schemes and the different forcing data on the model results model inter-comparison and sensitivity studies should be made.

• In the near future it will be possible to utilize remote sensing and model data for sea ice research to a considerably greater extent since remote sensing ice classification algorithms also resolve ice classes. Hence, the remote sensing data can be used for the verification of the redistribution functions and even for the assimilation of the ice model.
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References


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Notation

$A$  ice compactness
$\alpha$  surface albedo
$c$  specific heat
$C$  compaction constant
$C_H$  transfer coefficients of sensible heat
$C_E$  transfer coefficients of latent heat
$e$  aspect ratio of yield ellipse and water vapor pressure in heat flux equations
$\eta$  shear viscosity of ice
$\varepsilon$  water emissivity
$\dot{\varepsilon}$  strain rate tensor of ice
$f$  coriolis parameter
$F_{tot}$  total heat exchange between ice and atmosphere
$F_{sw}$  incoming short wave radiation
$F_{lw\downarrow}$  incoming long wave radiation
$F_{lw\uparrow}$  outgoing long wave radiation
$F_{sh}$  sensible heat flux
$F_{th}$  latent heat flux
$F_{wi}$  ice-ocean heat flux
$g$  gravitational acceleration
$g(h)$  ice thickness distribution function
$h$  ice thickness
$H$  sea surface tilt
$I$  unit tensor
$k$  heat conductivity
\( \hat{k} \)  upward unit vector  
\( L \)  latent heat  
\( m \)  mass  
\( N \)  total cloudiness  
\( P \)  ice strength  
\( P^* \)  ice strength constant  
\( \Psi \)  redistribution of ice  
\( \rho \)  density  
\( \sigma \)  internal stress tensor of ice  
and Stefan Boltzman constant in the heat flux equations  
\( U \)  wind velocity magnitude  
\( q \)  specific humidity of air  
\( R \)  enthalpy of vaporization  
\( t \)  time  
\( T \)  atmospheric turbidity  
\( T_a \)  air temperature  
\( T_0 \)  surface temperature of ice or water  
\( T_f \)  freezing point temperature of ice  
\( T_w \)  water temperature  
\( \Theta \)  thermodynamic growth rate of ice  
\( \theta \)  solar altitude  
\( \vec{u} \)  velocity vector  
\( \vec{\tau}^a \)  air-ice stress vector  
\( \vec{\tau}^w \)  water-ice stress vector  
\( \xi \)  the bulk viscosity of ice
Appendix I Calculation of heat fluxes

Incoming clear sky shortwave radiation \( (F_{sw0}) \) is calculated according the astronomical formulae which give the solar altitude in a function of time and location (Iqbal, 1983). The effect of the atmospheric turbidity and clouds are taken account via the following parametrization (Reed, 1977)

\[
F_{sw} = (1 - \alpha) \cdot F_{sw0} \cdot T \cdot (1 - 0.62N + 0.0019\theta)
\]  
(11)

where \( F_{sw} \) is the solar radiation gained by the surface, \( \alpha \) is the surface albedo (used values, see paper II), \( T \) is the atmospheric turbidity, \( \theta \) is the solar altitude and \( N \) is the total cloudiness.

The longwave radiation have been parametrized according to the Brunt-type formulas (Berland and Berland, 1952; Omstedt, 1990)

\[
F_{lw\uparrow} - F_{lw\downarrow} = \begin{cases} 
\varepsilon\sigma T_0^4 - \varepsilon\sigma T_a^4(0.39 + 0.058 \cdot \sqrt{e}) \cdot (1 - N^2) \\
\varepsilon\sigma T_0^4 - \varepsilon\sigma T_a^4((0.68 + 0.0036 \cdot \sqrt{e}) \cdot (1 + 0.18 \cdot N^2))
\end{cases}
\]  
(12)

where \( \varepsilon (= 0.97) \) is the water emissivity, \( \sigma \) is Stefan Boltzman’s constant, \( T_0 \) is the surface temperature, \( T_a \) is the air temperature, \( e \) is the water vapor pressure, and \( N \) is the cloud coverage. The first formula was used in papers II and III and the second formulae in paper IV.

Turbulent sensible and latent heat fluxes are calculated by the bulk methods:

\[
F_{sh} = \rho_a c_p C_H (T_0 - T_a) U
\]  
(13)

\[
F_{lh} = \rho_a R C_E (q_0 - q_a) U
\]  
(14)

where \( C_H \) and \( C_E \) are the transfer coefficients, \( U \) is the wind velocity magnitude, \( R \) is the enthalpy of vaporization and \( q \) is the specific humidity. A constant transfer coefficient was used in papers II and III and a functional form (Isemer and Hassé, 1987) was used in paper IV.

\[
C_H, C_E = \begin{cases} 
1.32 \cdot 10^{-3} & : \text{constant} \\
C_H (U, T_0 - T_a), C_E = C_H & : \text{63 values, Isemer and Hassé 1987}
\end{cases}
\]  
(15)
The oceanic heat flux is calculated prognostically by the bulk method since the ice model is coupled to the ocean model.

\[ F_{wi} = (\rho c)_w C_{wi}(T_w - T_{fp}) \]  \hspace{1cm} (16)

where \( C_{wi} \) is the bulk heat exchange coefficient \( 2.8 \cdot 10^{-5} \) and \( T_w \) is the water temperature (Omstedt and Wettlaufer, 1992).
Appendix II Revised estimates of the future ice condition

The model results of the future ice conditions in the Baltic presented in paper III were based on SILMU air temperature scenarios (Carter et al. 1995). Revised and more accurate climate change scenarios for the Baltic Sea region have been produced in the framework of the Swedish Regional Climate Modelling Programme (Rummukainen et al., 1999). They used regional atmospheric model forced by the global climate model to obtain downscaling simulation of the climate change in the northern Europe. Tuomenvirta et al. (2000) used the output of the regional atmospheric model (RCA1) in the estimation of the effect of the climate change to the Baltic Sea ice conditions. They used the ice model presented by Haapala (2000) and did two ten year integrations for the control and scenario simulations. All the model parameters and calculation of the surface fluxes were exactly the same as in Haapala (2000). The control simulation reflects the ice conditions during the pre-industrial time and the scenario simulation represents the ice conditions around year 2100. Figure 9 depicts the mean level ice thickness of the simulated ice winters during the mid-March. In the control simulation ice covers all the northern subbasins and the Danish straits. Ice thickness varies from 70 cm in the Bay of Bothnia to 10 cm in the southern regions. In the scenario simulation the extent of the ice covered area and the thickness of ice is largely reduced. Only the Bay of Bothnia is fully ice covered. Ice is also formed in the coastal areas of Bothnian Sea, in the eastern parts of the Gulf of Finland and the Gulf of Riga. Ice thickness is about 20 cm less than in the control simulation. The model produced a range for the maximum annual ice thickness in Kemi from 20 to 58 cm. The previous results presented by Haapala and Leppäranta (1997) are in consistency with the present result. The estimates according the SILMU scenarios for the ice conditions in 2050 can now understood to represent the ice conditions around and past the year 2100.
Figure 9: Modelled mean level ice thickness on 1-10 March. The left panel presents the ice thickness during the pre-industrial climate and the right panel is an estimate of the ice thickness around the year 2100 (Haapala et al., 2000)