ICE GROWTH ON SEA SURFACE AND DRIFT OF ICE

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Contents

Summary

This paper presents the thermodynamics and dynamics of sea ice, which form the basis of understanding short term and climatic evolution of ice conditions in freezing seas. For its material structure sea ice is a polycrystalline medium in small scale and polygranular medium in mesoscale or large scale. Ice growth and melting is a local vertical process, where first-ice ice may grow up to 2 m and multiyear ice twice that much. It is governed by the radiation balance and coupled with the dynamics of brine pockets in the ice sheet. Thermodynamics models predict the evolution of ice thickness fairly well. Sea ice drift is forced by winds and currents, and the response to forcing is dictated by the internal friction of the ice, which depends on compactness and thickness of ice. In free drifting ice – no internal friction – the velocity is a sum of the ocean current and the wind-driven drift, which is about 2% of wind speed; in the presence of strong internal friction the ice velocity lies between zero and area-averaged theoretical free drift. The complexity of internal friction brings difficulties to theory and modeling

of sea ice dynamics, but basin scale ice circulation can be rather well predicted. The evolution of sea ice conditions is a coupled dynamic-thermodynamic process. Ice growth strengthens the ice while opening of leads accelerates new ice formation due to intense ocean – atmosphere heat losses through open water and thin ice surfaces.

1. Introduction

Sea ice occurs in about 10% of the surface of the global ocean. Perennial sea ice is present mainly above latitude 80°N in the Arctic Ocean, in the East Greenland current, and in the western part of the Weddell Sea in south. In an average sense the seasonal sea ice extends to latitudes 60°, except offshore Scandinavia where to the warm North Atlantic Current protects the sea from freezing, and in lower latitude semi-enclosed shallow basins in the northern hemisphere, even below 40°N in Bohai Sea, China. The renewal time of sea ice is short, as perennial ice is mostly less than ten years old due to melting and advection.

The history of sea ice science initiates from the l800s with polar expeditions collecting sea ice field data (Weeks, 1998b). Weyprecht (1879) reported of the polar ice thickness and inspired Stefan (1891) to develop his elegant analytic ice growth law. It became known that sea ice drifts over long distances, and Nansen (1902) found, in his *Fram* expedition in 1893–96 across the Eurasian side of the Arctic Ocean, that the ice drift speed was 2% of the wind speed and the drift direction deviated 30° to the right from the wind. Several scientific expeditions followed: Makarov's Jermak to the Eurasian Shelf of the Arctic Ocean, Brennecke's Deutschland to the Weddell Sea and Sverdrup's Maud again to the Eurasian Shelf. Soviet Union commenced *North Pole Drifting Stations* programme in 1937, where science camps drifted from the North Pole to the Greenland Sea to be recovered there by icebreakers a year later. Sverdrup stepped into the boundary layer problems above and beneath sea ice, and Finn Malmgren collected the first extensive data set on the salinity of sea ice in the Maud trip. The masterpiece book, *Arctic Ice*, was published by Zubov (1945) presenting the status of sea ice science in Soviet Union.

The 1950s brought much new developments in sea ice science. The fine scale structural models of sea ice were completed in the 1950s (Assur, 1958) and the role of salinity in the parent water body for ice properties became understood. Sea ice thermodynamics research progressed to a better understanding of the annual cycle of ice thickness and temperature, and ice sea ice dynamics problem the closure was completed in the 1950s as rheology and conservation laws were established. Toward the end of 1960s the first numerical models of sea ice thermodynamics and dynamics had been developed. The motivation of the sea ice science was in the cold war after World War II expanding to shipping and oil and gas exploration in later decades.

The modern era in sea ice science commenced in the 1970s based on new technological innovations. Satellite remote sensing revealed finally the character of sea ice fields with their time-space variations, numerical computations made possible to construct realistic sea ice models, and automatic measurement systems became feasible for long-term monitoring of the ice conditions. In the same decade, the *AIDJEX (Arctic Ice Dynamics Joint Experiment)* program was performed with two major findings: plastic rheology

and ice thickness distribution. Recent progress contains granular flow models, ice kinematics mapping with microwave satellite imagery, anisotropic rheologies and scaling laws. Methods for remote sensing of sea ice thickness have been slowly progressing, which is the most critical point for the further development of the theory and models of sea ice dynamics.

Sea ice is a delicate factor in the environmental conditions of ice-covered waters and it plays also an important role in the global climate problem. Sea ice fields are very thin layers on the ocean surface (Figure 1). They grow, drift and melt under the influence of solar, atmospheric, oceanic and tidal forcing. In sizeable basins solid sea ice lids are statically unstable and break into fields of ice floes, undergoing transport as well as opening and ridging which altogether create the exciting sea ice landscape as it appears to a human eye. Thermodynamics and dynamics of sea ice are strongly coupled since ice growth and melting influence the strength of the ice and drift and mechanical deformation modify the sea ice thickness distribution and consequently the thermal insulation capability of the ice.

Sea ice forms of the ocean water and melts into that, and therefore the ice–ocean coupling is strong. The ice and water exchange momentum, heat and salts which in turn influences the ocean surface layer and in places the whole vertical water column. Freezing releases salts to the ocean surface layer and corresponds to evaporation while melting corresponds to precipitation. From the viewpoint of atmosphere–ocean interaction, the presence of ice has qualitative and quantitative consequences. The transfer of solar radiation into the ocean is remarkably lowered, and the forcing of the surface currents by winds and the air–sea heat exchange are strongly modified. The nature of these modifications depends on the structure of the sea ice fields.

This work introduces the science of sea ice geophysics. Chapter 2 presents the material properties of sea ice from small scale to ice floes and fields. The next chapter is about sea ice thermodynamics showing the physics of ice growth and melting and analytical and numerical models. Sea ice dynamics is treated in Chapters 4 and 5. First, the theory and observations are presented and then numerical models with applications for short-term and long-term problems. For references, Doronin and Kheysin (1975) and Wadhams (2000) have published excellent general book on sea ice geophysics, and summer school proceedings are provided by Untersteiner (1986) and Leppäranta (1998). Lake ice modelling has been summarized in Leppäranta (2009). Thermodynamics of sea ice is treated by Maykut and Untersteiner (1971) and Makshtas (1984), while sea ice dynamics is covered by Hibler (1986), Timokhov and Kheysin (1987) and Leppäranta (2010) and sea ice modeling in Hibler (2004).
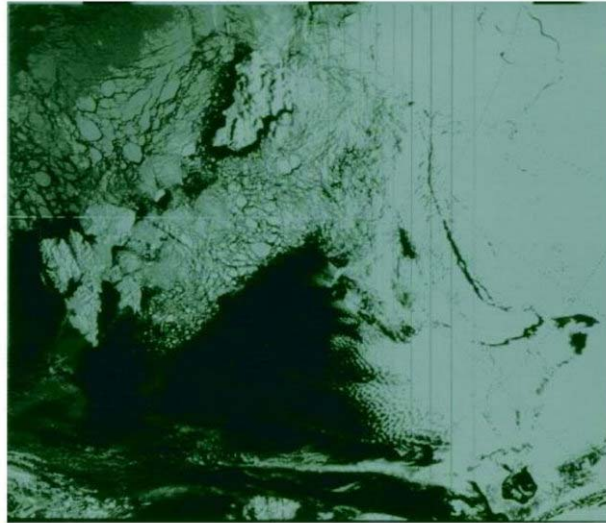
Figure 1. An optical channel NOAA-6 image over the Barents Sea and Arctic Basin north of it, May 6, 1985. Svalbard is shown on the left, Franz Joseph's Land up in the middle and Novaya Zemlya on the right. The east side is cloudy, but elsewhere drift ice floes can be seen. The image was received at the Tromsø receiving station, Norway.

2. Material Properties of Ice on Natural Waters

2.1. Scales and Forms of Ice

Sea ice physics is examined over a wide range of scales. *Microscale* includes individual grains and ice impurities extending from sub-millimeter region to 0.1 m. In the *local scale*, 0.1–10 m, sea ice is a solid sheet, a polycrystalline continuum with sub-structure classified according to the formation mechanism into congelation ice, snow-ice and frazil ice. *Ice floe scale* extends from 10 m to 10 km, including individual floes and ice forms such as rubble, pressure ridges and fast ice. When the scale exceeds the floe size, the sea ice medium is called *drift ice* or *pack ice*, and, as in dynamical oceanography, mesoscale is around 100 km and the scales from 1000 km upward are in the *large scale* regime.

Sea ice growth is a vertical process and a small-scale phenomenon. Sea ice melting is a vertical small-scale process but also a horizontal floe scale process since lateral melting of ice floes is also important. The WMO nomenclature (WMO, 1970) restricts *ice floes* to those ice pieces with size more than 20 m, smaller ones taken as *ice blocks*. This is convenient since then the *aspect ratio h/d* of ice floes is smaller than about 0.1, i.e. they are flat. The floe size ranges from the lower limit to tens of kilometers. In the research of sea ice dynamics we consider the drift of individual floes or the drift of a system of ice floes, a *drift ice field*.

2.2. Small Scale Properties of Ice

Sea ice is a multi-component medium. In sea water ice crystals form from water

molecules and grow into crystal platelets. They are optically uniaxial, with the *optical axis* or the *c*-axis perpendicular to the plate plane. Overlain together they form macrocrystals, which are optically like single crystals but structurally multiple crystals. In the geophysics of sea ice, the term *ice crystal* or *grain* normally refers to these macrocrystals. The size of macrocrystals is $10^{-4} - 10^{-1}$ m. Sea ice crystals are in size and shape as fresh water ice crystals in lakes and rivers, depending on the mode of ice formation. But there are two important differences caused by the dissolved substances in seawater (Weeks, 1998a). In sea ice the crystal boundaries are jagged, and inside macrocrystals between the single crystal platelets there are brine inclusions.

Sea ice forms

Ice formation is based on three main mechanisms resulting in *congelation ice*, *frazil ice* and *snow ice* (Weeks, 1998a; Eicken and Lange, 1990). In Antarctic shelf waters so-called *platelet ice* forms as glacial melt water rises from bottom of ice shelves, becomes supercooled and crystallizes onto sea ice bottom as large platelets. Congelation ice crystals grow down from the ice–water interface, and the crystals are columnar, diameter 0.5–5 cm and height 5–50 cm. The growth is limited by the insulation effect of the ice, and the thicker the ice is the lower the growth rate will be. In the Arctic Ocean congelation ice is the dominant ice type.

Frazil ice forms in open water areas. The crystals are small (1 mm or less), they follow free with the water in turbulent flow and attach later to the bottom of existing ice or join together into a solid sheet at the surface. In shallow and well-mixed waters frazil may also attach into the sea bottom to form *anchor ice*. The growth rate can be fast because of intensive heat losses from open water to cold atmosphere and not strongly limited as long as there is open water present. In the Antarctic seas frazil ice is the dominant ice type. There it is typical that frazil crystals join in the surface to plate aggregates, which collide with each other and become rounded. This is called *pancake ice* due to its visual appearance, and the size of the plates is up to a few meters. Frazil ice crystals are effective in harvesting particles from the water column and they may also accumulate onto the sea bottom and become anchor ice. This anchor ice may rise up due to its buoyancy and bring bottom material to the surface ice.

Snow-ice forms on top of the ice from slush, generated by snow and liquid water available from flooding, liquid precipitation or melt water of snow. The crystals are small as in frazil ice. Growth of snow-ice is limited by the presence of snow and availability of liquid water. The most common formation mechanism is flooding, which becomes possible when the snow weight has forced the ice sheet below the level of the water surface, i.e. when

$$\frac{h_s}{h_i} > \frac{\rho_w - \rho_i}{\rho_s} \qquad (1)$$

where h_s and h_i are snow and ice thicknesses, and ρ_w, ρ_s and ρ_i are water, snow and ice densities. Since $(\rho_w - \rho_i)/\rho_s \approx 1/3$, the thickness of snow needs to be at least

one-third of the thickness of ice for the flooding to occur. The growth limitation is provided by the h_i / h_s ratio; consequently snow ice formation is common in Antarctic seas and in low latitude seas, such as the Baltic Sea and the Sea of Okhotsk, where snow accumulation is large and ice is not too thick. In spring the day-and-night melting-freezing cycle gives additional growth to the snow ice layer.

The impurities of sea ice consist of *brine*, *solid salt crystals*, *gas bubbles* and *sediments*. The first two types result from growing sea ice capturing salts. They constitute the most important impurities. The volume fraction of gas bubbles is $v_a \sim 1\%$, the bubble size is in the range 0.1–10 mm, and they have a significant influence on the properties of the ice. In particular, they have a major influence the transparency of the ice sheet and backscattering of electromagnetic waves. Sediment particles originate from the water body or sea bottom, harvested by frazil ice or anchor ice formation, and atmospheric fallout accumulating on the ice during the ice season. Sediments may influence the properties of ice and they may also have a significant role in transporting matter, especially pollutants.

Sea ice salinity

When saline water freezes, ice crystals form on water molecules, i.e. freezing tends to separate dissolved substances out from the solid phase of water. However, due to constitutional supercooling in the molecular diffusion of salt and heat in the liquid phase, cellular ice-water interface forms and is able to close liquid brine pockets between crystal platelets (Weeks, 1998a). The surface water salinity, where the transition between fresh water ice and sea ice takes place, is within 1–2 ‰ according to observations in the Baltic Sea (Palosuo, 1961; Kawamura et al., 2001). These brine pockets also serve as habitats for ice algae.

The salinity of new ice is a fraction κ of the salinity of the water: $S_i = \kappa S_w$, where $\kappa \approx 0.25 - 0.5$ is the segregation coefficient. The salinity of brine must always correspond to the freezing point of the ambient temperature, $S_b^{-1}(T) = T_f$, and thus the brine salinity and consequently the brine volume change with the temperature. In addition, in low temperature salts start to crystallize from the brine, each in its eutectic temperature point. The most important ones are in temperatures above –25°C sodium chloride (–22.9°C) and magnesium sulfate (–8.2°C). The composition of sea ice at a given temperature can be read from the phase diagram (Figure 2). To estimate the brine volume, the bulk temperature, salinity and density of sea ice are needed. The phase diagram shows the proportions of ice, brine and solid salts, and the phase equations for exact calculations are available in Cox and Weeks (1983) for $T \le -2°C$ and Leppäranta and Manninen (1988) for $-2°C \le T \le 0°C$. In mid-winter the salinity decreases a little due to brine pocket expulsion, but in summer as the ice becomes warm the brine pockets expand into drainage net and much of the brine drains out.

The influence of the salinity on the properties of sea ice is first of all via the brine volume. Solid salt crystals scatter light and therefore influence on the optical properties of sea ice.

Brine pockets serve as biological habitats. They contain nutrients, and with light penetration sufficient conditions for primary production exist. The algae are captured into the ice in the freeze-up process, and their growth in the brine pockets is primarily light-limited. The most active layer is the bottom layer or so-called *skeleton layer*, which may become colored brown-green by the algae.

Due to the salinity the properties of sea ice are different from those of fresh water ice. The difference is largest in the electromagnetic properties and strength. The electromagnetic properties are normally not so relevant as such but they are used indirectly to observe other ice properties. *E.g.,* microwave backscatter depends on the dielectric constant, and the salinity of ice consequently shows up in radar mapping of sea ice. Due to the presence of brine pockets sea ice is porous, and therefore the strength of sea ice is less than that of fresh water ice. The flexural strength is (Figure 3) is a good reference since there is a lot of *in situ* data.
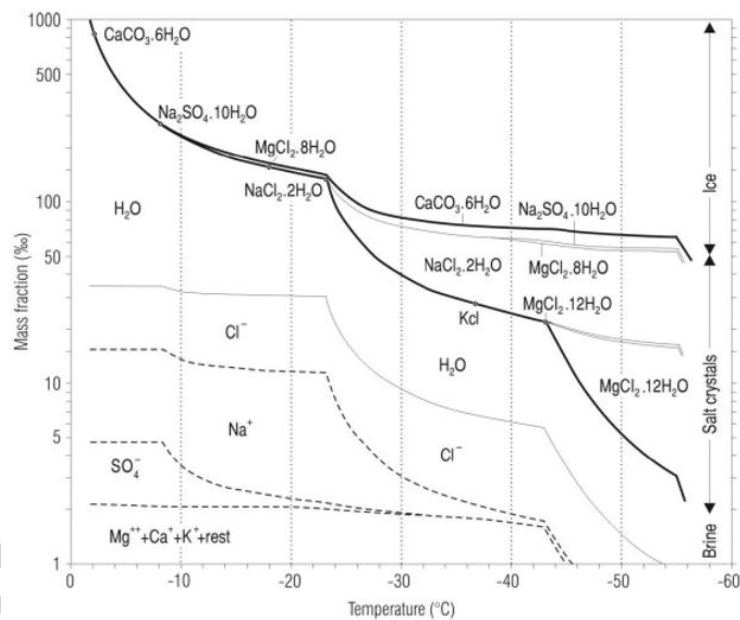

Figure 2. Sea ice phase diagram with the proportions of ice, brine and solid salts as a function of temperature. The chemical symbols show the eutectic points of the main salts. Redrawn from Assur, A. (1958) Composition of sea ice and its tensile strength. In: Thurston, W. (ed.), *Arctic Sea Ice*, pp. 106–138. U.S. National Academy of Science/National Research Council: Publ. No 598
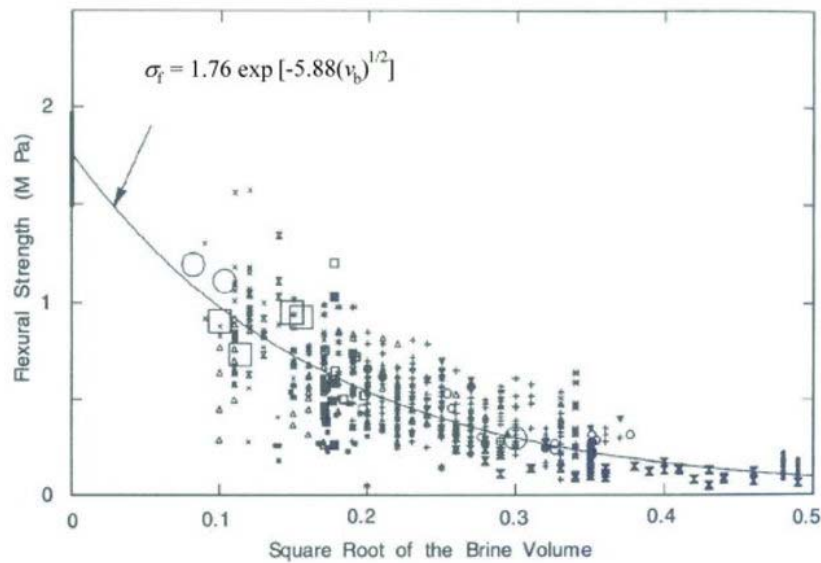
Figure 3. Flexural strength of sea ice as a function of brine content. Redrawn from Weeks, W.F. (1998a) Growth conditions and structure and properties of sea ice. In: Leppäranta, M. (ed.), *Physics of Ice-Covered Seas,* Vol. 1, pp. 25–104. Helsinki University Press, Helsinki.

In thermodynamics, sea ice differs qualitatively from fresh water ice. Along with temperature evolution, the brine volume changes with associated phase changes at brine pocket boundaries. This brings an interesting viewpoint: sea ice has no definite melting point but always there is melting involved when sea ice warms in order to dilute the brine. (Sea water has a well-defined freezing point, which depends on the salinity; it is –1.8°C for the salinity of 35 ‰.) Thus brine pocket dynamics has strong influence on the apparent heat capacity of sea ice, especially when the temperature is higher than –5°C. In addition, the thermal conductivity of brine is a little less than that of ice, and thus thermal conductivity of sea ice is slightly smaller than that of fresh water ice.

The optical properties of sea ice are influenced by the gas bubbles and the salt content (Perovich, 1998). Gas bubbles are similar to those in fresh water ice and they are good scatterers. The brine pockets absorb light and add little scattering, and salt crystals scatter light but their number becomes significant only at very low temperatures. Especially in summer, chlorophyll in the brine pockets absorbs strongly in the bands 430–440 nm and 660–690 nm. The penetration of sunlight can be described using the Beer's law as in the case of liquid waters:

$$Q(z) = (1-\alpha)\gamma Q(0)\exp(-\lambda z) \qquad (2)$$

where Q is solar irradiance, α is albedo, $\gamma \approx 0.45 - 0.50$ is the optical band fraction in the solar radiation, and λ is the attenuation coefficient; $\alpha \sim 0.75$ and $\lambda \sim 0.2 \text{ cm}^{-1}$ for snow and $\alpha \sim 0.5$ and 0.02 cm^{-1} for ice. The typical attenuation distance can be taken as $3\lambda^{-1}$ or 15 cm in snow and 150 cm in ice (Perovich, 1998). The euphotic depth is usually taken as the depth where the irradiance

level has gone down to 1% of the surface level. Consequently, bare first-year sea ice allows enough light to penetrate through for primary production but snow cover of more than about 20 cm shuts the light out.

2.3. Ice Floes and Ice Fields

Drift ice is a peculiar geophysical medium (Figure 4). It is granular (ice floes are the basic elements, grains) and the motion takes place on the sea surface plane as a two-dimensional system. The compactness of floe fields may easily change, i.e. the medium is compressible, the rheology shows highly non-linear properties, and by freezing and melting an ice source/sink term exists

A sea ice landscape consists of ice floes with ridges and other morphological features, and leads and polynyas. It can be divided into zones of different dynamic character (Weeks, 1980). *(Land)fast ice* is the immobile coastal sea ice zone extending from the shore to about 10–20 m depths (in Antarctic grounded icebergs may act as tie points and extend the fast ice zone to deeper waters). Next to fast ice is the *shear zone* (width 10–200 km), where the mobility of the ice is restricted by the geometry of the boundary and strong deformation takes place. Thereafter comes the *central pack*, free from immediate influence from the boundaries; its length scale is the size of the basin. *Marginal ice zone (MIZ)* lies along the boundary to open sea. It is loosely characterized as the zone, which "feels the presence of the open ocean" and extends to a distance of 100 km from the ice edge. Well-developed MIZs are found along the oceanic ice edge of the polar oceans. They influence the mesoscale ocean dynamics resulting in ice edge eddies, jets and upwelling/downwelling. MIZ and shear zone can be fairly wide, and therefore a well-defined central pack exists only in large basins.
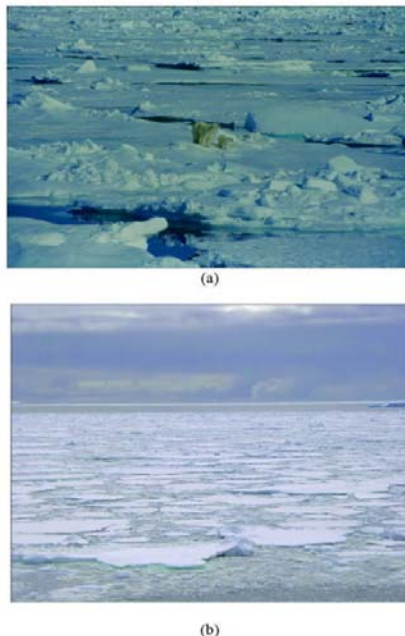

(a)

(b)

Figure 4. Sea ice landscapes. (a) The heavy pack ice in the Arctic Ocean, north of Svalbard, and (b) First-year ice floe field in the Weddell Sea, Antarctica. Photographs have been taken by the author.

The horizontal structure of a sea ice cover is well revealed by optical satellite images (Figure 1). Ice floes are described by their thickness h and diameter d, and we may examine the drift of an individual floe or a field of floes. For continuum models to be valid for a floe field the size of continuum material particles D must be satisfy

$$d << D << \Lambda \tag{3}$$

where Λ is the gradient length scale. The right side inequality is to apply the linear deformation theory. The ranges are in nature $d \sim 10^1 - 10^4\,\mathrm{m}, D \sim 10^3 - 10^5\,\mathrm{m}$ and $\Lambda \sim 10^4 - 10^6\,\mathrm{m}$. As $D \to \Lambda$ discontinuities build up, and as $D \to d$ a single floe or a few floes system appears. The vertical dimension is described by the thickness of ice.

In the continuum approach in sea ice dynamics, an *ice state J* is defined for the material description, $\dim(J)$ being the number of levels. The state variables contain the drift ice properties, which are needed in the drift ice rheology and consequently in modeling the internal friction of drift ice fields. The first attempt was $J = \{A, H\}$, where A is ice compactness and H is mean thickness (Doronin, 1970). Three-level ice states $J = \{A, H_{\mathrm{u}}, H_{\mathrm{d}}\}$ decomposing the ice into undeformed ice thickness H_{u} and deformed ice thickness H_{d} have been used (Leppäranta, 1981). The fine resolution approach is to take the thickness distribution for the ice state (Thorndike et al., 1975). This distribution consists of fixed thickness classes h_{k}, $k = 0, 1, \ldots$, arbitrarily spaced, and it represents the ice state:

$$J = \{\pi_0, \pi_1, \pi_2, ...\} \tag{4}$$

where π_k's are the spatial densities of the thickness classes. Usually $h_0 = 0$ and thus $A = 1 - \pi_0$.

In granular flow models the horizontal floe characteristics in addition to their thickness is needed. The first models, based on stresses transfer due to ice floe collisions, assumed circular ice floes (Shen et al., 1986). Later, discrete particle models have taken both circular and polygonal floes (Løset, 1993; Hopkins, 1994).

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Biographical Sketch

Matti Leppäranta was born 1950 in Helsinki, Finland. He received MSc degree in mathematics in 1976 and PhD degree in geophysics (sea ice) in 1981 in the University of Helsinki. He has worked as a research physicist in the Finnish Institute of Marine Research in 1975-1991 and as a professor in geophysics in the University of Helsinki since 1992. His science career has been around sea ice geophysics from its beginning, with sea ice dynamics as the leading theme. The research has covered short-term sea ice forecasting for winter shipping in the Baltic Sea, marginal ice done dynamics, climatic sea ice modeling, scaling questions, sea ice structure and thermodynamics, and drift of oils spills in ice conditions. He has given sea ice dynamics lectures in several advanced study institutes and published the books *The drift of sea ice* (2005) and *Physical oceanography of the Baltic Sea* (2009) with Kai Myrberg. Matti Leppäranta has acted as a post doc physicist in Cold Regions Research and Engineering Laboratory, Hanover, NH, visiting teacher in Universitetstudiene i Svalbard (UNIS), and visiting scientist in Hokkaido University, Japan, and Dalian University of Technology, China. He is an adjunct professor in University of Sherbrooke, Québec, Canada and Polar Research Institute of China, Shanghai.