Land–ice interaction in the Baltic Sea

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Abstract. An overview of the evolution of landfast ice in the Baltic Sea, especially in the vicinity of Hailuoto Island in the north, is given, and semi-analytical models are presented to examine the vertical and lateral growth and breakage of landfast ice. The outer edge of landfast ice moves further offshore as the ice becomes thicker. Occasionally, landfast ice breaks and moves, forming grounded ridges, scouring the sea bottom, piling up on the shore and riding up onto land up to distances of more than 100 m. According to observations of ice breakage, the yield strength of landfast ice is proportional to the squared ice thickness. In very shallow areas the water may freeze to the bottom, and after sea level rise the ice may drift away and transport bottom sediment. The models provide a first-order approach to understand the evolution of the landfast ice zone from the start to the winter maximum.

Key words: ice, land, Baltic Sea, ice thickness, ice breakage, erosion, sea bottom, shore.

INTRODUCTION

The annual ice season lasts 5–7 months in the Baltic Sea, from November to May. The coastal and archipelago areas are covered by landfast ice, while further out drift ice fields are found. Landfast ice is normally a stable ice sheet, and its outer edge extends further out from the coast in winter, as the ice grows thicker. This edge reaches the isobaths of 5–15 m depending on the severity of the winter (Leppäranta 1981). In extreme forcing conditions, landfast ice may be broken and forced to move horizontally, which drives erosion of the sea bottom and shore. Such ice displacements are of concern to coastal management, first of all for ship traffic, marine and coastal infrastructures, on-ice traffic, and safety issues for work and recreation on ice. In spring, melting starts from the shorelines of mainland and islands, and the last ice floes to disappear are remnants of thick, grounded ridges formed at outer shoals.

In the early history of the Baltic Sea ice science, landfast ice was in the core of the research (Jurva 1937). The extent of landfast ice was also used as an indicator for the ice conditions in the whole basin, since there was no access to examine closer the drift ice. An interesting early work on ice–land interaction was that by Keyserling (1863) from Pärnu Bay, Gulf of Riga. He reported of a case when 70 cm thick ice was pushed to land, riding up to 310 m onshore and in places forming 18 m high ridges. Kraus (1930) reported of a similar event in the same basin. Palosuo (1963) examined the extent, lateral development and thickness of the landfast ice zone. Occasional investigations were later made of grounded ridges and ice ride-up on the shore (Alestalo & Häikiö 1975; Zakrzewska 1981).

The principal motivation of the Baltic Sea ice research was, for long, winter shipping. Ice, in particular drift ice, caused major problems to ships, lighthouses and beacons. With utilization of aerial ice reconnaissance, drift ice research started to grow in the 1940s (Palosuo 1953). Later on, ice ridges were examined by surface-based mapping (Palosuo 1975), and field experiments with drift ice stations were performed on sea ice dynamics (Leppäranta 1981) and remote sensing (e.g., Askne et al. 1992). In recent years, new questions have arisen concerning the environmental and ecological influences of ice formation in the Baltic Sea (Kawamura et al. 2001; Arst et al. 2006). After year 2000, the landfast ice zone has been re-introduced as one focal point in the ice research (e.g., Girjatowicz 2004; Granskog et al. 2004; Goldstein et al. 2009). In particular, a major environmental issue is land–ice interaction, which results in the erosion of shores and sea bottom forced by ice, or briefly, *ice erosion*.

This paper is based on an overview of the physics of landfast ice in the Baltic Sea. The structure and thermodynamics of landfast ice are fairly well known, but for the nearshore zone ice mechanics, only a few dispersed investigations have been published. To improve our understanding of ice–land interaction, first-order mathematical models are presented for the analyses of the evolution of the coastal ice zone. The region between Hailuoto Island and mainland in the Bay of Bothnia is taken as a case study for the literature survey and model applications.

ICE PHASES IN THE COASTAL ZONE

In the Baltic Sea, the landfast ice zone progresses along similar *ice phases* each year, but the timing of the phases varies largely from year to year (Jurva 1937). The idea behind these phases is that the horizontal extent of sea ice progresses in a stepwise manner controlled largely by the heat budget and the topography of the basin. Based on an extensive, long-term database, the characteristic ice situations have been constructed for twenty phases (see Figs 1, 2), and for an actual ice season they act as the guideline of the evolution of the ice conditions (Leppäranta et al. 1988). Eleven phases are progressive and nine are regressive. The morphology of the ice field in the phases is determined by the coastline geometry, islands, bottom topography and ice thickness. The lateral strength of the ice cover increases with thickness, and therefore the thickness of ice is the primary ice property in the evolution of the ice conditions in the phase system.

Figures 1 and 2 show examples of the early and late phases in the northeast part of the Bay of Bothnia, the northernmost basin of the Baltic Sea. The early phases (Nos 1–5) cover on average the period from 15 November to 27 January, altogether 73 days (Fig. 1). Landfast ice grows laterally to around the 10-m isobath encircling the northern archipelago of the Bay of Bothnia. Vertical growth takes place simultaneously, stabilizing the ice sheet. When the coastal ice is still thin, strong wind may break it. This is possible in phases 1–3, but thereafter landfast ice is too thick and heavy ice pressure will result in ridging at the outer boundary of the landfast ice zone. Then fracturing of landfast ice is connected to the distribution of islands and shoals (Goldstein et al. 2004). Figure 3 shows a MERIS image over the Bay of Bothnia from spring (19 April). Snow has mostly melted and therefore the structure of landfast ice is clearly seen, consisting of large ice plates, while drift ice is highly deformed with many ridges and ice floes are small.

The late phases (Nos 18–20) are shown in Fig. 2. On average they cover the period of 5–28 May, altogether 23 days. There are almost no studies regarding the decay of the landfast ice zone. The decay process is fairly short (see Fig. 2), and the ice sheet loses its strength rapidly with melting at the upper and lower surfaces and in the interior. Ice melting starts from shores of islands and mainland, and ice forces decrease due to decreasing fetches and diminishing strength of the ice. Then ice compactness lowers, and the whole ice cover appears more and more as a field of scattered ice floes. Also winds are low in May. Therefore, no strong ice forces arise in spring when landfast ice becomes mobile. In spring with weak winds, landfast ice just rottens and melts where it was formed with no significant drifting, but in a windy spring the ice breaks and may drift away or ride to the shore. Consequently, landfast ice has not caused serious problems in the ice decay period, which explains the low interest in research.

The lateral growth of the landfast ice zone comes to an end in normal winters after Phase 5, having reached the outer archipelago. Then the fetch extends over the whole of the Bay of Bothnia, and very large ice thickness is required to stabilize the ice cover. Based on data from very severe winters, it is known that at least a quasi-stable ice cover is possible when the minimum ice thickness has reached about 50 cm in the entire basin. In such winter, the author stayed in a ‘drifting’ ice station for two weeks in February 1985 without observable displacement (the positioning accuracy was then 40 m) even though the wind speed went up to 17 m s^{-1}. In the same winter it was also observed that icebreaker channels did not deform during a period of about two months in midwinter in the Bay of Bothnia.

In the past, before all-year shipping commenced (1970) in the Gulf of Bothnia, in severe ice seasons ‘ice bridges’ formed across the gulf in the narrow straits between Finland and Sweden, the Northern Quark and the Southern Quark. On-ice travelling was then possible across these straits (Palosuo 1956). There are many documented cases of on-ice crossing of the Northern Quark in the history. In Finland War (1808–1809) the Russian troops crossed the strait from Finland to Umeå, Sweden. In cold winters there was a public ice road across the Northern Quark; the last winter with the ice road was 1966. In the given area, stabilization of the ice cover depends on the wind speed and ice thickness (Palosuo 1963; Leppäranta 1981). An example is given by Fig. 4, where the breakage of ice is shown offshore Oulu in the Bay of Bothnia. Toppila is a small basin close to the shore where much more wind is needed to break the ice than further offshore.

ICE EROSION

Forms of erosion

The growth, decay and movement of landfast ice bring ice to contact with the shore and bottom material. This drives ice erosion, which can be classified into three categories: (i) thermal erosion, (ii) thermo-mechanical erosion and (iii) mechanical erosion.

Thermal erosion is due to thermal expansion of ice. Its influence concerns the nearshore zone where the ice growth has reached the shore or the bottom. Since the linear thermal expansion coefficient is about 5×10^{-5}°C^{-1} (Petrenko & Whitworth 1999), thermal ice ride-up on the shore is limited to ~5 m in ~10-km size basins for ~10 °C temperature changes. It is not very important in the marine environment but can be so in small lakes. However,

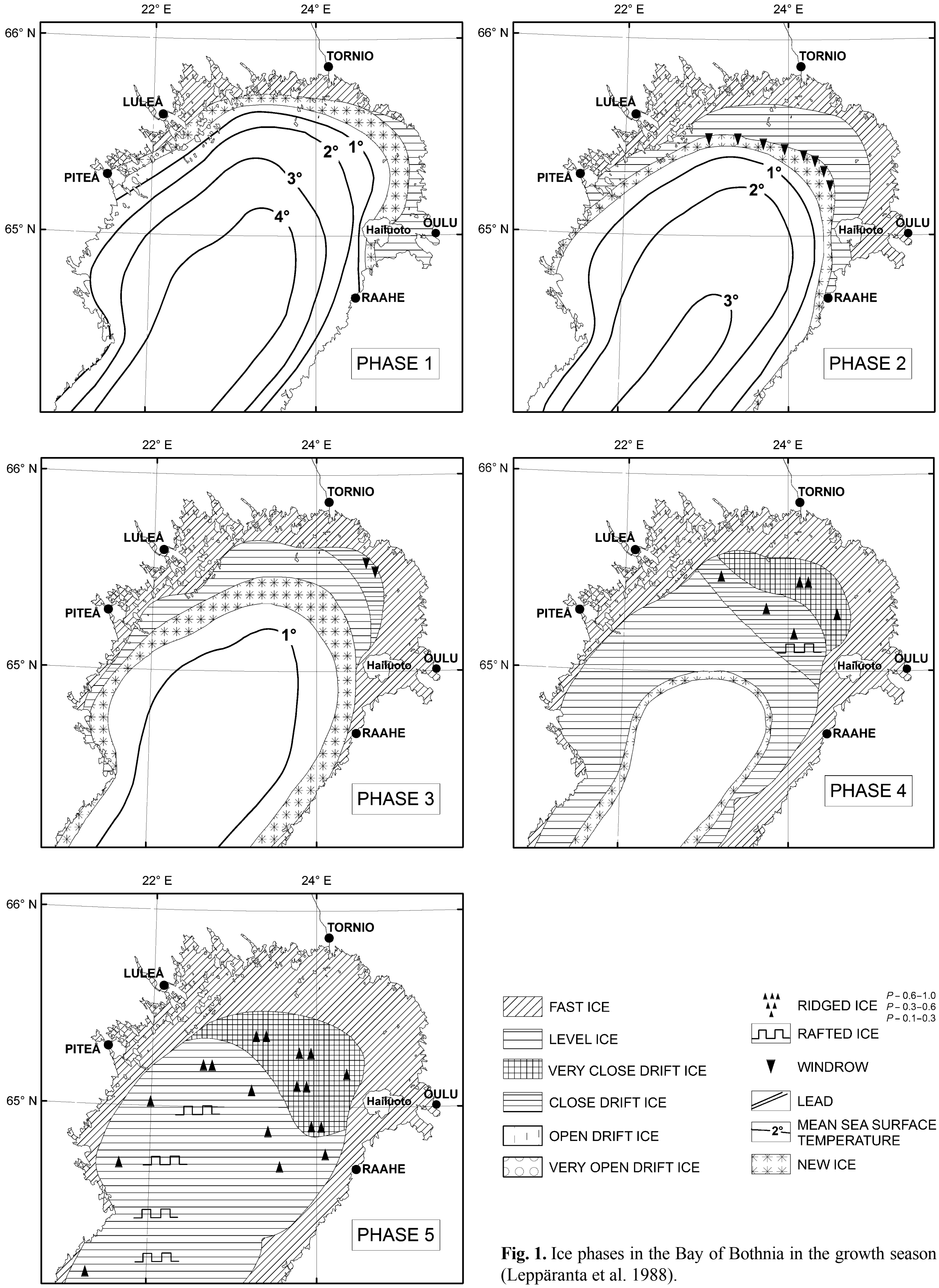

Fig. 1. Ice phases in the Bay of Bothnia in the growth season (Leppäranta et al. 1988).

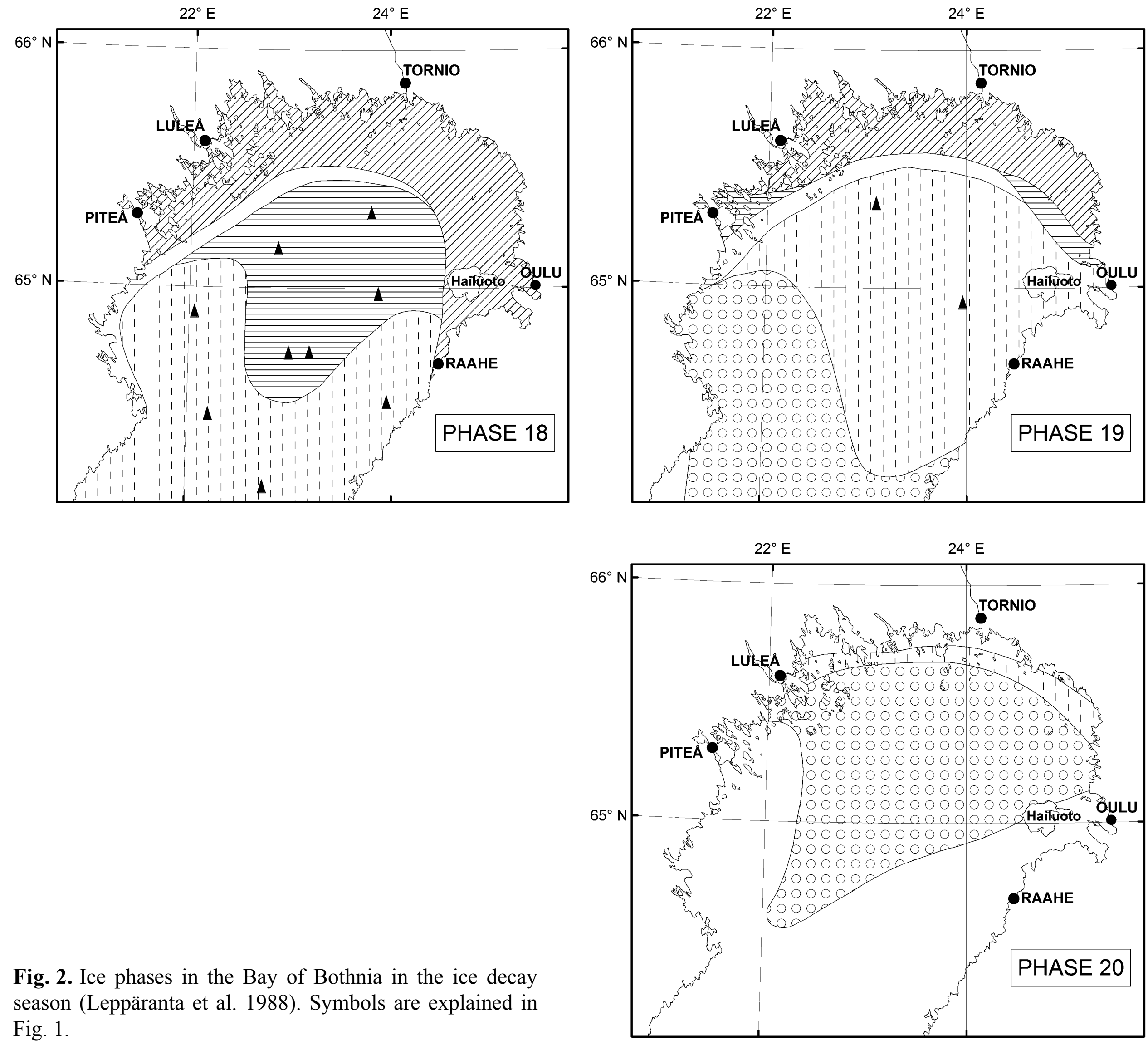

Fig. 2. Ice phases in the Bay of Bothnia in the ice decay season (Leppäranta et al. 1988). Symbols are explained in Fig. 1.

thermal erosion has been suggested to be a significant process at the shoreline of Gdansk Bay (Zakrzewska 1981).

Thermo-mechanical erosion is a sequential process. In very shallow areas the water may freeze to the bottom, and after sea level rise the ice may drift away and transport bottom sediments to another location. Thermal ice growth may reach 120 cm in the northern Baltic Sea (Leppäranta & Myrberg 2009). Whether the ice transports the captured bottom sediments or releases them back at the growth site depends on the spring weather conditions.

Mechanical erosion is the most important ice erosion process in the Baltic Sea. Breakage of landfast ice and the following ice displacements cause interaction between ice and sea bottom and the shore, which has major environmental and societal consequences. These include bottom and shore erosion, and for marine constructions, major forces on structures. The wind may push large ice sheets to land, causing land surface erosion (Kovacs & Sodhi 1980). This pushing results in ice sheet breakage and piling up of ice blocks at the shoreline or riding up of the ice sheet onto land, even to distances of more than 100 m. A few publications have reported the occurrence of these processes in the Baltic Sea (e.g., Keyserling 1863; Kraus 1930; Alestalo & Häikiö 1975; Alestalo et al. 1986).

Experiments on shore pile-up and ride-up were made by Sodhi et al. (1983). The goal was to be able to predict which one takes place in a given case. The number of experiments was limited, as expressed by the authors, but there was an indication that for a fixed flexural strength of ice, ride-up results when thickness is large enough or the coefficient of ice–land friction

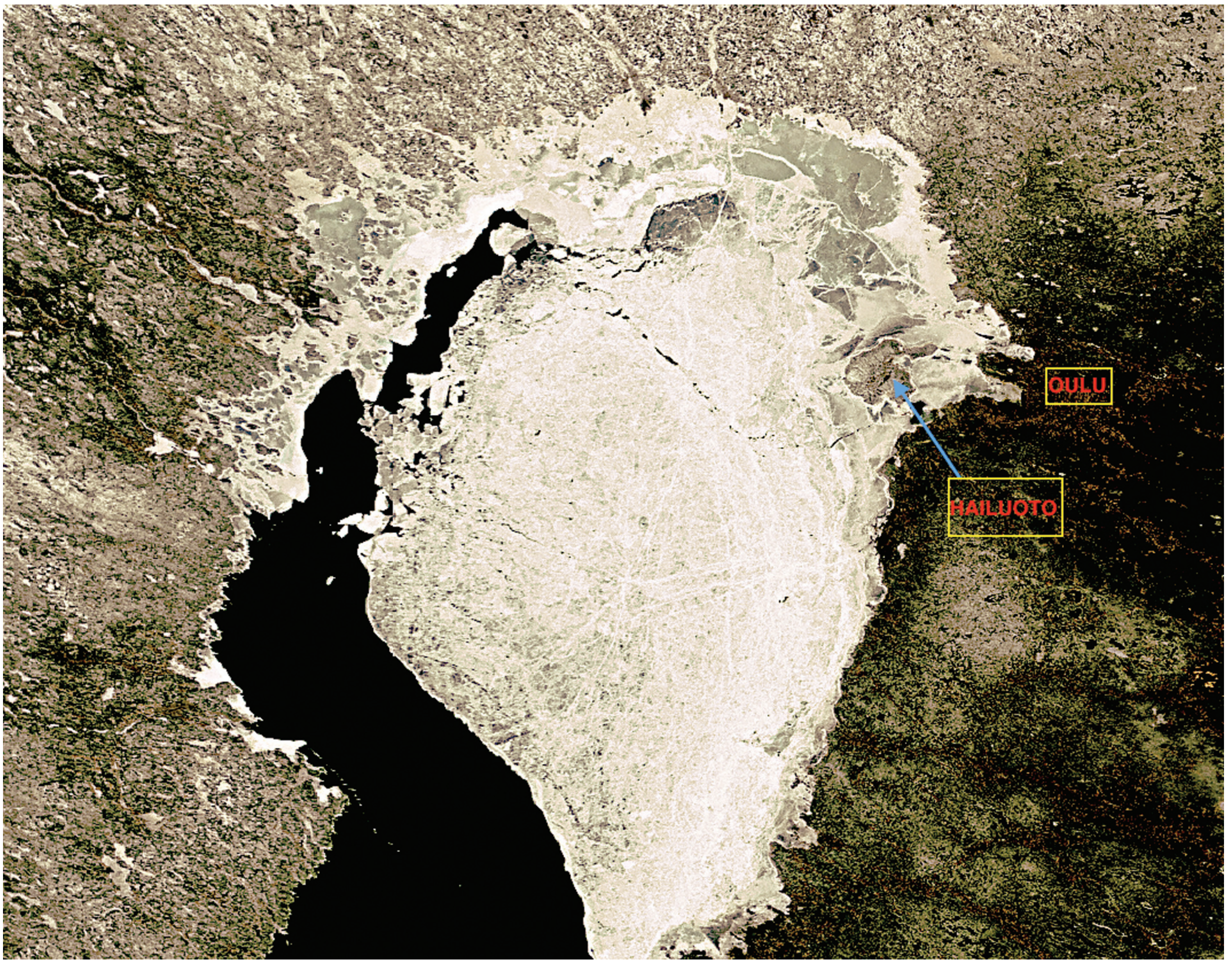

Fig. 3. Envisat MERIS image over the Bay of Bothnia, 19 April 2011. © ESA

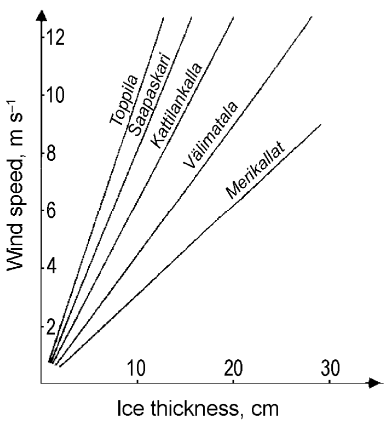

Fig. 4. A nomogram of the stability of landfast ice offshore Oulu (Palosuo 1963). The figure shows the dependence of the breakage of ice on ice thickness (horizontal axis) and wind speed (vertical axis); the lines *Toppila* to *Merikallat* mark sub-basins from the shore line to the outer edge of the archipelago. © Finnish Environment Institute, Helsinki.

is small. The ice force required to push an ice sheet must overcome the gravitational and frictional forces (Kovacs & Sodhi 1980). This force per unit width is proportional to $hh_{\mathrm{f}}(1+\mu\cot\alpha)$, where h is the ice thickness, h_{f} is the freeboard, α is the slope angle and μ is the coefficient of friction. The force per unit width required to fail an ice sheet in bending is proportional to $\sigma_{\mathrm{f}}h^2/L$ (Croasdale et al. 1978), where σ_{f} is the flexural strength and L the characteristic length of the ice sheet. The force per unit width required to fail an ice sheet in buckling is proportional to L^2 (Sodhi 1979).

Bottom scouring by ridge keels has not been investigated in the Baltic Sea. Ice ridges have typically 5–15 m deep keels in the Baltic Sea; the maximum observed keel depth has been 28 m in the Bay of Bothnia (Palosuo 1975). The breakage of the ice depends on the thickness of ice, fetch and wind speed (Palosuo 1963). With increasing thickness, longer and longer fetches are needed to break the ice, which results in the shifting of the landfast ice boundary further offshore during winter (see Fig. 1). Which zones are the most likely to experience bottom scouring is not clear.

The influence of ice erosion on the ecosystem and coastal management

Ice erosion is an important phenomenon in the Baltic Sea environment. There are large, shallow archipelago areas as well as relatively flat shore areas where ice has major impact on the shoreline and sea bottom. In the northern part of the Baltic Sea, the geology of the basin is also highly dynamic due to land uplift up to 1 m in a century. In the Finnish western lowlands, the shoreline has progressed by 1 km in a century since the Weichselian glaciation.

The ecosystem in the Baltic Sea coastal zone has adapted to regular or rare occurrence of ice erosion events. Even more, parts of it have become dependent on ice erosion. Shoreline areas can be kept clean of bushes by ice ride-up on the shore. Spots where ride-up is more common are exposed to long fetches for dominant directions of strong winds. In the Gulf of Bothnia strong winds are primarily from the sector between south and west, and thus the eastern coast of the basin can experience heavy ice forces. The terrain is quite flat on the eastern side, and therefore shore ice ride-up can reach long distances. Also small, low islets can be kept clean by the ice movement.

Vascular plants, which can survive only due to ice erosion keeping the shores clean, grow in the Gulf of Bothnia. Otherwise the shore area would be covered by bushes, which would destroy the ecological niche for these plants. A few vascular plant species have been classified as endangered. Migrating birds use open shore areas for their stopovers, which are conserved in such state due to ice erosion. For some bird species ice erosion is highly important. Nesting of a few bird species is conditioned by the presence of bare rock islets, without any bushes or trees, existing due to recurring ice erosion.

Landfast ice influences the spreading of river waters by weakening the turbulence. Baltic Sea coastal waters are saline. In the absence of mixing, fresh river waters can stay in a thin layer just beneath the ice. Thus the transport and dispersion of pollutants in the coastal zone differ in quality between ice and open water seasons. Also the formation of the thin upper layer supports the development of winter blooms, since phytoplankton can stay in the layer where sunlight is available (see Myrberg et al. 2006). Unless the snow cover is very thick, there is light enough for primary production below a half-metre ice cover in the Baltic Sea (Arst et al. 2006).

For coastal management issues, the main problems are the ice forces, bearing capacity of ice and ice displacements. The bearing capacity is quite well understood. It is proportional to ice thickness squared, with the scaling coefficient as about 5 kg cm^{-2} serving as a good approximation until spring when ice becomes porous and loses its strength. The potential mobility of landfast ice introduces a more unpredictable risk, which is of major concern for constructions and traffic in the coastal zone. The moving ice causes forcing on offshore and onshore structures and ships and is a danger to people on ice. The breakage of landfast ice takes place under certain extreme conditions and can be avoided for traffic, however. Due to potential bottom scouring, cables and pipes need to be buried deep, and wrecked ships are harmed. The size and strength of ice ridge keels become a critical factor.

In spring, when the solar radiation becomes stronger, the ice sheet is deteriorated in its interior. The strength of the ice decreases and the contact between ice and land becomes loose due to melting starting from the shoreline. Ice is more easily breakable but, on the other hand, in springtime strong winds are rare and the melting process is fast. Therefore serious ice displacement cases are not likely to occur (no documented case is known to the author).

Hailuoto case

Landfast ice and its mechanics are of major concern in the northern Bay of Bothnia. The greatest ice forces in the Baltic Sea are recorded in this region, since here the thickness of ice is at largest and strong and stormy winds, which blow dominantly from the southwest sector, have the longest fetches. Hailuoto is the largest island in the Bay of Bothnia, located in the northeastern part of the bay offshore Oulu, with the shortest distance of 8 km from the mainland. The island and its surroundings are close to the location of the land uplift maximum. The altitude is very low and, consequently, land uplift results in rapid geological evolution of the region.

The traffic connection between Hailuoto and the mainland has been kept by a ferry and in winter also by an ice road, which goes not far from the ferry line. In the distant past, the island was blocked out from the mainland by weak ice in autumn and spring. The winter ferry traffic was opened in 1968, but until 1989, apart from mild winters, it was closed for 2–3 months each winter. Full, all-year operational traffic started in 1989, when a larger and more powerful ferry was taken into use.

The evolution of the ice conditions in the vicinity of Hailuoto is illustrated by the ice phases in Figs 1, 2 (see also Seinä & Peltola 1991). The area between Hailuoto and the mainland is very shallow, less than 10 m, and therefore it becomes rapidly ice-covered at the beginning of the ice season. Landfast ice extends up to 30–40 km out from the shoreline encircling Hailuoto (Fig. 1). When ice thickness has reached 30 cm, the ice is no longer mobile (Fig. 5).

During ice displacements, bottom scouring may be significant further offshore where ridges are formed. However, there is not much information of scouring in this region. It is known that transport of bottom sediments frozen into the ice sheet in shallow regions is a significant process and affects the spreading of sea grass.

There are several documented cases of the breakage of landfast ice between the mainland and Hailuoto, when the thickness has been less than 30 cm. The typical progress of an event is as follows:

(i) *Preconditioning*. The wind piles up water, rising the sea level enough to release the ice from contact with the shoreline and sea bottom. This means that the wind must be from southerly directions, which may pile up the sea level to more than 1.5 m at Oulu.

(ii) *Ice breakage*. With strong wind and long fetch, the ice cover breaks and is set in motion if its thickness is less than about 30 cm.

(iii) *Ice erosion*. Ice displacement has three possibilities:
- piling up at the shoreline: thin ice (thickness less than about 15 cm) breaks into smaller pieces and may form shoreline ridges;
- ride-up on the shore: thick ice (thickness more than about 15 cm) may ride up longer on the shore if the ice is a solid sheet (the necessary force is 10–50 N m^{-1}, the wind speed must be more than 20 m s^{-1} to achieve that);
- ice ridging: ice breaks within the ice area and forms a ridge, which further may scour the sea bottom.

(iv) *Slowdown*. The process ends due to increasing frictional resistance to erosion and due to decreasing fetch.

Southerly-westerly directions are dominant for strong winds in the Baltic Sea, and thus the water level is often high in the northern coasts of the Bay of Bothnia. Similar preconditioning was also reported by Keyserling (1863) and Kraus (1930) for the Gulf of Riga. If a strong wind persists, the ice sheet may break and move; however, then, after preconditioning, the wind may blow from any direction, breaking the ice if the fetch is sufficient. What happens next depends on the ice cover structure and ice thickness. Ice ride-up on the shore and islets is the most critical for nature conservation, and it necessitates the ice thickness to be at least about 15 cm. Thus, for ride-up we must have: sea level rise, wind speed more than 20 m s^{-1}, ice thickness 15–30 cm and the ice field large enough to give the needed fetch. The period for the required ice thickness range is normally 2–4 weeks, and it is clear that the necessary wind conditions do not occur every year.

In the last 50 years major ice breakage events have happened about once in a decade. Apart from the cases of Keyserling (1863) and Kraus (1930), the highest grounded ridge measured in the Baltic Sea area was formed on 23 March 1986. The wind was from the south with a speed of 27 m s^{-1}, and 50-cm thick ice floes accumulated to a 15-m high, grounded ridge close to

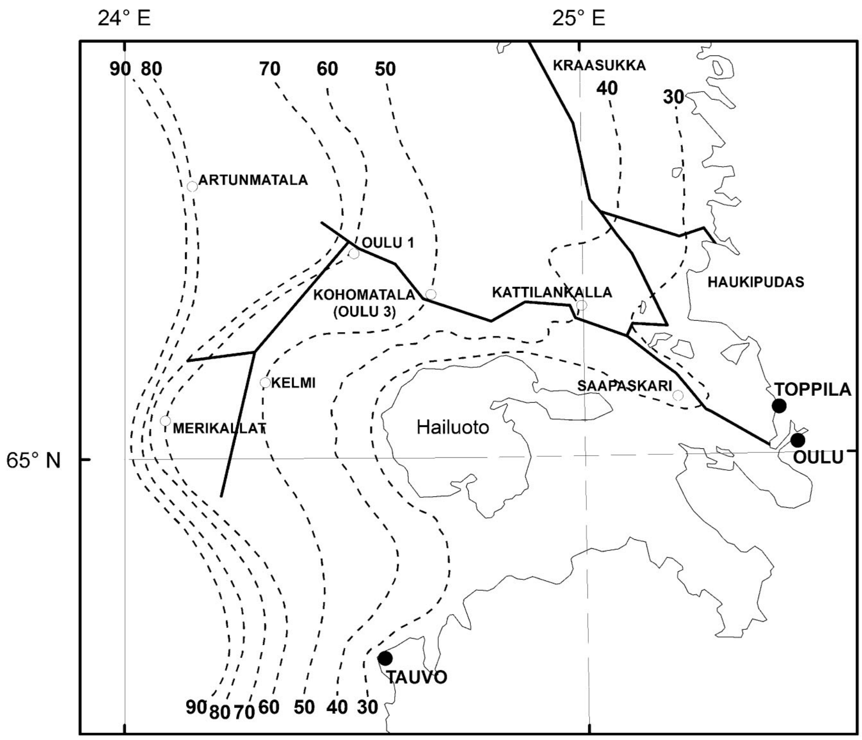

Fig. 5. Mobility of ice in the vicinity of Hailuoto. The maximum thickness of mobile ice is shown (Palosuo et al. 1982).

Marjaniemi lighthouse (Alestalo et al. 1986). Shore ride-up cases have occurred at Hailuoto in the autumn. When the coastal ice was still thin, strong wind could break it and push it onshore to distances of more than 100 m from the shoreline in the mainland next to Hailuoto (Alestalo & Häikiö 1975). Ice thickness was then 18–25 cm, and the wind was from the southwest with a speed of 15–21 m s^{-1}. The last authors also reported that trees were less than 17 years old in the nearshore zone of 100 m width, which refers to frequency of major ride-up events in the past.

The most destructive shore ride-up case known so far took place on 19 December 1992 (Markku Tönkyrä, pers. comm.) (Fig. 6). The speed of the southern wind was 28 m s^{-1}, and the ice thickness was 40 cm between

A

B

Fig. 6. Ice ride-up on the shore in Hailuoto on 19 December 1992: **A**, ice sheet breaks onshore; **B**, ice pushes a house close to the shoreline. © Courtesy of Mr Markku Tönkyrä, Hailuoto.

the mainland and Hailuoto. First the sea level rose to 150 cm above mean, and then the ice sheet was riding onshore over long distances on the eastern side of the island, Santonen Peninsula, destroying several buildings (Fig. 6B). It is the only time this thick ice (40 cm) is known to have ridden up, but on the other hand, the wind speed was almost at the all-time record level. Six days later the ice that was left in the basin was pushed to the mainland shore by the westerly wind with a speed of 25 m s^{-1}.

After year 2000, two major events have occurred in Santonen (Markku Tönkyrä, pers. comm.). On 23 November 2004 there was a strong southwesterly storm. The sea level rose to 130 cm above mean, and 20 cm thick ice rode onshore and accumulated into high ridges. Sidemarks of the main road in Santonen suffered from ice forces. A similar case developed two years later, on 25 November 2006 (Fig. 7).

The diary of the Hailuoto ferry of 1999–2009 tells about four disturbances caused by ice ride-up and movement. These are all from the early ice season, November–January. Ice rode up onshore close to the ferry terminal on 25 January 2006, and the piles of ice blocks needed to be removed by a tractor. On 25 November 2006 (the same storm as in Santonen above) the ice was moving and the ferry traffic was partly inactive during a 3-h period. On 31 January 2007 one trip was cancelled due to the ice, and on 5 January 2009 the ice caused serious delays for the whole day.

According to the local population in Hailuoto, the opening of the winter ferry traffic in 1968 and the consequent persistent channel between the island and the mainland changed the mobility of the ice. Ice breakage and displacements are said to have been absent before that for the ice thickness greater than 20 cm. Also, according to the local population, another increase in the ice mobility took place in winter 1989, when a larger ferry started up and the ferry traffic became denser and the ferry channel widened. This statement is somewhat difficult to judge since the winters 1989–1993 were very mild and the average annual ice thickness has decreased since 1989.

FIRST-ORDER MATHEMATICAL MODEL SYSTEM

Analytic tools provide a first-order approach to the evolution of the landfast ice zone. Ice formation and growth models provide the lateral and vertical growth of the landfast ice zone, determined by the depth of the sea and air temperature. The breakage of ice then depends on the wind speed, fetch and ice thickness. For the deterioration of landfast ice in spring neither good analytical nor good numerical models are available.

Ice formation and growth

Ice formation begins in the Baltic Sea in the autumn at the shoreline. After the air temperature has crossed below the freezing point temperature, freezing of the surface water follows with a time lag depending on the depth of the sea. For the brackish water of the Baltic Sea, the freezing point is 0.1–0.3 °C below 0 °C, but for simplicity, it is assumed here that the freezing point temperature equals 0 °C. With time, the lateral boundary of the ice cover expands further out and the ice grows thicker. Cooling and vertical ice growth can be approximated in

Fig. 7. Ice ride-up on the shore in Hailuoto on 25 November 2006. © Courtesy of Mr Markku Tönkyrä, Hailuoto.

the coastal zone by simple models where the forcing is provided by the air temperature T_a. The basic equations are for the cooling (T – water temperature) and vertical ice growth (h – ice thickness), respectively:

$$\frac{dT}{dt} = \lambda(T_a - T), \quad \lambda = \frac{K_a}{\rho_w c_w H}, \tag{1a}$$

$$\frac{dh}{dt} = \frac{a}{2}\frac{T_f - T_a}{h + d}, \quad a = \frac{2k_i}{\rho_i L_f}, \tag{1b}$$

where t is time, λ is the relaxation constant of the mixed layer (λ^{-1} is equal to the memory length of the system), K_a is the air–sea heat transfer coefficient, ρ_w is water density, c_w is specific heat of water, H is the depth of the mixed layer, a is the freezing-degree-day coefficient, T_f is the freezing point temperature, d is a parameter describing the insulating effect of the atmospheric surface layer (in terms of equivalent ice thickness), k_i is thermal conductivity of ice, ρ_i is ice density and L_f is latent heat of freezing. Equation (1a) describes how the lateral ice edge expands out from the shoreline, and Eq. (1b) shows the vertical ice growth after freezing. In the landfast ice zone the depth of the mixed layer can be taken equal to the sea depth. Based on the work by Palosuo (1963) with Hailuoto data, the estimates for the parameters in Eq. (1) are $\lambda \sim 1 \times H^{-1}$ m d^{-1}, $a \sim 5$–10 cm^2 d^{-1} °C^{-1} and $b \sim 10$ cm.

The solutions of Eqs (1) are written as (Zubov 1945; Rodhe 1952)

$$T(t) = T(0)\exp(-\lambda t) + \int_0^t \exp[-\lambda(t' - t)]T_a(t')dt', \tag{2a}$$

$$h(t) = \sqrt{a^2 S + d^2} - d, \; S = \int_{t_0}^{t} \max[T_f - T_a(t'), 0\,°\mathrm{C}]\,dt', \tag{2b}$$

where S is the sum of freezing-degree-days and t_0 is the time of freezing. The time zero is arbitrary. Combining Eqs (2), the full lateral–vertical ice growth is obtained as a function of depth.

Examples of the solution of Eqs (2) are given in Fig. 8 in the case of linear atmospheric cooling by 6 °C per month, which corresponds to the conditions in Hailuoto. Freezing takes place at the time $t_0 = \lambda^{-1}$ after 0 °C down-crossing of air temperature, i.e., the delay is proportional to depth, so that, roughly speaking, each additional metre gives one day more of delay (Leppäranta & Myrberg 2009). The period of ice thickness upcrossing 15 cm is much shorter than the period of freeze-up in the zone from the shore to 30 m depth. Thus the period when the

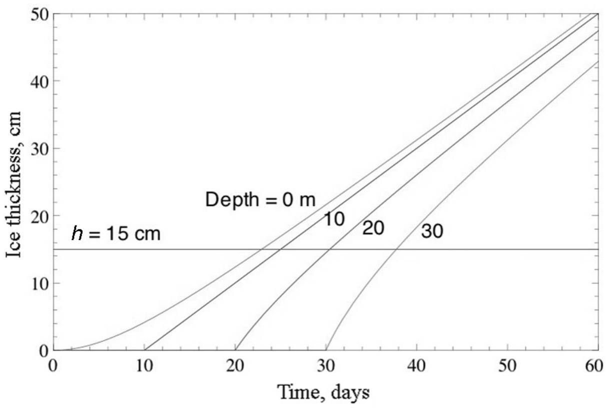

Fig. 8. Ice thickness evolution along the coastal zone for the depth zones of 0, 10, 20 and 30 m. Air temperature decreases linearly by 6 °C per month.

landfast ice zone is strong and mobile becomes fairly short. The profile of ice thickness from the ice edge to the shore would become convex upwards, with a maximum at the shore and zero at the edge.

The coastal zone can be described as a mosaic of basins, with depth and size increasing with distance from the coastline. In places, such as the Latvian–Lithuanian coast in the Gotland Sea, the coastal zone is just a sloping bottom without islands, while wide archipelago areas exist in most of Finland. If air temperature is taken locally as a constant, the freezing front progresses perpendicular to the depth contours capturing deeper basins. New ice is thin and breaks easily, but once the critical thickness has been achieved, landfast ice forms. This basin-by-basin advance of the landfast ice zone was recognized by Jurva (1937) as a series of phases, which follow the same pattern each year but the timing of the phases differs from year to year.

Breakage of landfast ice

Landfast ice forms a stable ice cover until spring but occasionally, under extreme forcing conditions, it can break and experience displacements of tens of metres to kilometres. The necessary forcing is provided by stormy winds, preceded by build-up of high sea level releasing the ice cover from contact with boundaries. Because of the specific chain of events, displacements of landfast ice are very rare.

The breakage of landfast ice was examined by Palosuo (1963) based on empirical data. He showed that in a given basin, the stability or breakage of the ice sheet depends on ice thickness and wind speed. He did not consider the precondition of high sea level. However, this precondition has been noted in all well-documented cases of major displacements of landfast ice. The results of Palosuo (1963) can be formulated as

$$h_c = f(L)U_a, \tag{3}$$

where h_c is the critical thickness between stable state and breakage, U_a is wind speed and f is a function of the size of the basin L. It is clear that f is a monotonously increasing function of L. On the other hand, plastic sea ice mechanics theory tells that the ice will move when (e.g., Leppäranta 2011)

$$\tau_a L = P, \tag{4}$$

where τ_a is wind stress and P is the two-dimensional strength of the ice sheet, which depends on the thickness and length scale. Wind stress is proportional to the square of the wind speed (e.g., Andreas 1998): $\tau_a = \rho_a C_a U_a^2$, where ρ_a is air density and C_a is air drag coefficient. Then

$$\rho_a C_a U_a^2 L = P(h; L). \tag{5}$$

By Eq. (3), ice thickness at breakage is proportional to wind speed. Therefore ice strength must be proportional to the thickness squared (f is independent of wind speed) and can be written as

$$P = P_2^* \psi(L) h^2, \tag{6}$$

where P_2^* is a strength constant (equal to the strength of ice of unit thickness) and the function $\psi = \psi(L)$ provides the dependency of the strength on the length scale. The results of Palosuo (1963) show that the strength decreases with the length scale in a sub-linear manner. Sanderson (1988) prepared a diagram of the ice stress dependency on the length scales from 10^0 m to 10^5 m based on results from small-scale experiments and mesoscale and large-scale modelling. His results show that the strength decreases roughly in proportion to the square root of the length scale. The power of $-\frac{1}{2}$ can be taken as a hypothesis of the strength–length scaling law. Ice breakage cases in Hailuoto suggest that $P_2^* \sim 3\times10^5$ N m^{-3} at $L \sim 10$ km. Then we have

$$h_c = \kappa^* U_a \sqrt{L}, \ \kappa^* = \sqrt{\frac{\rho_a C_a}{P_2}}, \tag{7}$$

where the parameter κ^* can be taken as a constant, $\kappa^* \sim 10^4$ m$^{-1/2}$ s. Thus, when the thickness of ice increases, the maximum distance between support points for stabilization increases quadratically. For example, for $h_c \sim 20$ cm we have $L \sim 10$ km; and for $h_c \sim 60$ cm we have $L \sim 100$ km.

From Eq. (7) it can be understood that landfast ice extends easily up to the outer islands. As the ice thickness increases, the length between stabilizing support points may increase and the landfast ice zone becomes wider. But the nonlinearity of Eq. (6) tends to keep the boundary of landfast ice at the outer islands. To span a stable ice cover across the Baltic Sea, much thicker ice is needed. In very cold winters, the Bay of Bothnia ice cover can reach a stationary state. With $U_a = U_{a,max}$, the maximum wind speed in a given period of time, the minimum thickness of ice in the landfast state is obtained.

CONCLUSIONS

An overview of the landfast ice zone in the Baltic Sea was given, and analytic models were presented to examine the vertical and lateral growth and breakage of landfast ice. The area between Hailuoto Island and the Finnish mainland in the Bay of Bothnia was taken as the case study. The ice conditions undergo evolution along similar phases each year (Leppäranta et al. 1988). In extreme conditions, the stability of landfast ice is broken and the ice sheet moves driving ice erosion, i.e. onshore pile-up and ride-up of ice, ridging and scouring, and transport of bottom sediments.

Vertical and lateral growth of landfast ice can be described by analytical freezing and ice growth models. For a given weather history, the freezing date depends on the depth of the sea providing the lateral growth of the ice edge, and the vertical growth can be estimated by Zubov's (1945) analytical model. Landfast ice can be considered as a two-dimensional plastic medium, which breaks when the wind forcing reaches the yield limit. The parameters of these simple models were estimated from Hailuoto data. The ice growth law is straightforward since Zubov's (1945) law has only two parameters: the freezing-degree-day coefficient (5–10 cm^2 °C^{-1} d^{-1}) and the ice–air boundary layer insulating buffer effect (10 cm ice equivalent). A quadratic law $P \propto h^2$ was obtained for the lateral strength of a two-dimensional ice sheet, with inverse sub-linear dependency on the length scale. The estimated strength coefficient of 3×10^5 N m^{-3} at the length scale of 10 km was obtained. The power of the sub-linear length scale law was of the order of $-\frac{1}{2}$ as in the scaling law of Sanderson (1988); this power provides a good hypothesis but the question needs further investigations. This result should be applicable to sea ice models for coastal zone applications (e.g., Leppäranta & Wang 2002) and to improve their ice stress fields in general. The obtained strength law is valid for early or mid-winter situations; another question is the breakage of landfast ice in spring during ice deterioration.

The role of the ice in the coastal zone has not been much examined in the Baltic Sea, but it is becoming more important for management and environmental protection of coastal areas. Infrastructure constructions in coastal areas may change the ice conditions and state of the ice cover significantly. Due to, e.g., landfills and wind farms, the geometry of the coastal ice zone may change, modifying potential ice displacements and wind fetch and consequently ice forces would be modified. Ice erosion is an active agent in the ecology of coastal regions and dependent on the weather conditions and morphology of the coastal zone.

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Maa-jää vastasmõju Läänemere rannikul

Matti Leppäranta

On antud ülevaade kinnisjää moodustumisest Läänemerel Hailuoto saare põhjaranniku näitel. On rakendatud pool-numbrilisi mudeleid kinnisjää kujunemise ja lagunemise kohta. Varasematele töödele vastavalt on kinnisjää voolavuspiir proportsionaalne jää paksuse ruuduga, st kinnisjää piir nihkub jää moodustumisel oluliselt avamere poole. Samas kinnisjää tuule toimel aeg-ajalt murdub, moodustades kohati merepõhja uuristavaid valle, mis kuhjuvad rannikule ja võivad liikuda kuni 100 meetrini merepiirist kaugemale. Madalas vees võib vesi külmuda põhjani, mere-taseme tõusmisel hakkab jää liikuma, võttes endaga jällegi põhjasetteid kaasa. Olemasolevad mudelid annavadki esimest järku lähenduse kinnisjää tekkimisest ja arenemisest kuni talveharjani.