Barents Sea drift ice characteristics*

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A short review of the ice field characteristics in the Barents Sea is given. A digitised ice distribution data set covering the period 1966–88 has been used to calculate the probability distribution of sea ice and temporal trends. The ice-covered area between 68 and 83°N and 10 and 55°E at the end of the melting season is found to have been reduced by 40% during the last 23 years. The maximum extension, however, shows no significant decrease over the same period, and this indicates that there might have been a reduction/increment in the ice thickness/melting effects in the Barents Sea during the last 23 years.

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Introduction

The sea ice conditions in the different marginal seas bordering the Arctic Ocean are determined by dynamic and thermodynamic processes characteristic for each area. In addition, the exchange with the Arctic Ocean may influence, and in some cases dominate, the age-composition of ice in the various seas. The physical boundaries of the different marginal seas are in this connection of major importance. The western and southern parts of the ice fields of the Barents Sea are bounded by the warmer currents spreading north and eastwards from the Norwegian Sea. Towards the east there is a relatively large opening to the Kara Sea between Frans Josef Land and Novaja Zemlja; this facilitates the exchange of ice masses with the neighbouring sea in this area. The boundary towards the north is linked by a number of islands which to some extent govern the exchange of ice with the Arctic Ocean.

The ice drift in the Barents Sea is mainly wind-driven. The exchange of ice with the Kara Sea and the Arctic Ocean may thus vary considerably from year to year in accordance with changes in the general atmospheric circulation. This exchange in turn determines to a large extent the amount of ice that remains in the Barents Sea through the subsequent summer. The composition of the ice fields may therefore at times be very complex, consisting of local and imported ice of various ages.

Glaciers in Frans Josef Land and Svalbard feed icebergs to the area. This occurs at a highly variable rate, mainly determined by the surging activity. The subsequent distribution of icebergs is determined by the prevailing wind field. An extreme event occurred in the spring of 1929 when about 20 icebergs were observed along the coast of Finnmark (Hoel 1962).

Distribution

The denser, warmer modified Atlantic Water fills up the deeper part of the Barents Sea. The southward extension of the sea ice is therefore during the cold season indirectly, topographically controlled in the western part, where the bottom topography shows marked shelf breaks. Similarly well-defined shelf breaks do not exist in the eastern part. Contingent upon the intensity of the governing dynamic and thermodynamic processes, the increasingly cooled water flowing eastward from the Norwegian Sea may here dive under the polar water at quite different latitudes. These features are reflected in the increasing variability of the ice edge position in the eastern part of the Barents Sea (Fig. 1).

The degree of cooling of the warmer water masses entering the southern area from the west as well as the ice formation during the cold season are determined by the atmospheric and oceanic circulation. The position and the intensity of the Barents Sea Low are in this regard of particular importance. Prevailing westerly/easterly winds will for example cause less/more ice formation in

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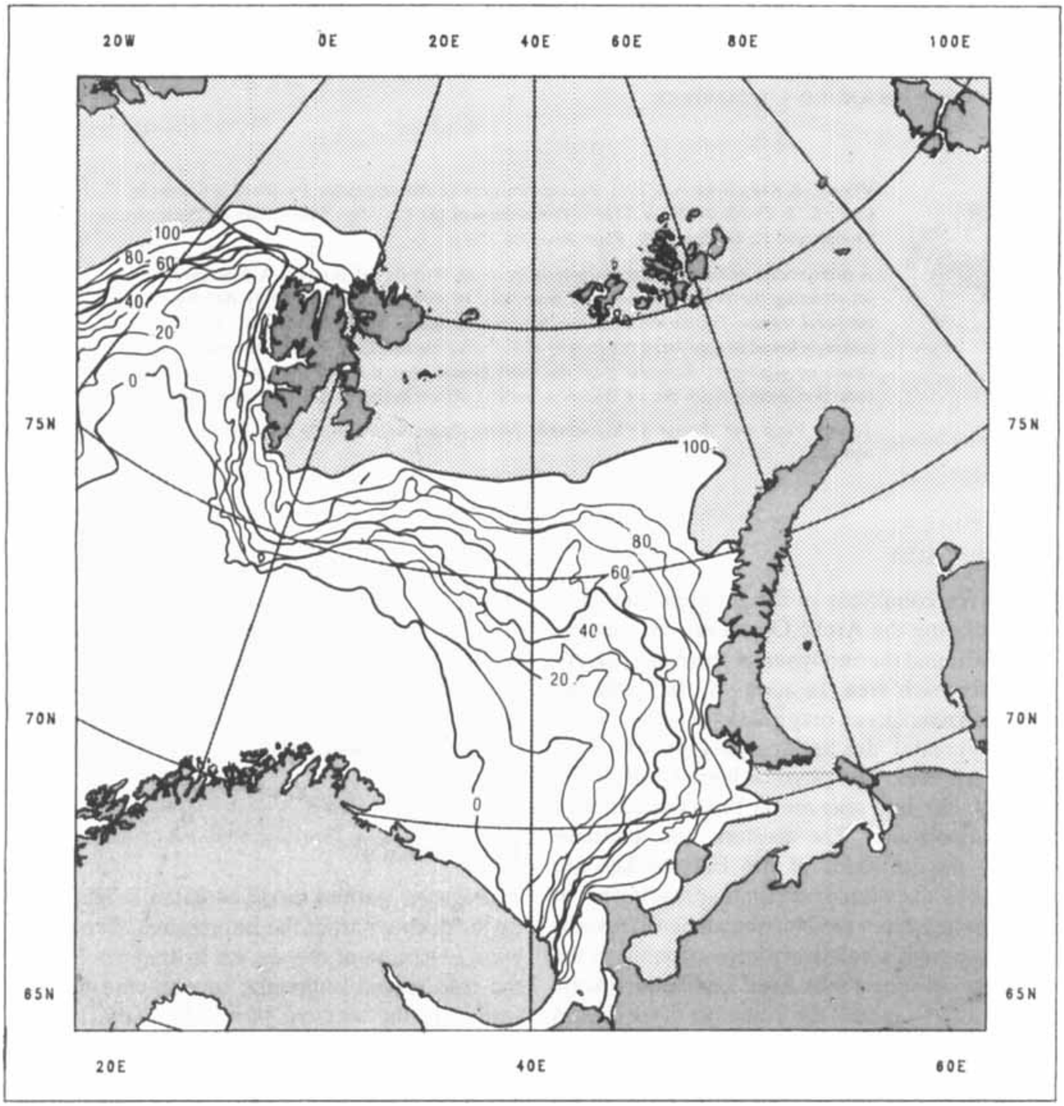


Fig. 1. The probability of encountering sea ice at the end of April in the Barents Sea. We note the eastward increasing variability in the ice edge position. From satellite-based ice maps during 1966–1988, mainly from the Norwegian Meteorological Institute (DNMI), the NAVY-NOAA Joint Ice Center, U.S.A., and observations from ships and aircraft collected by Norsk Polarinstitutt.

this area and accordingly determine the position of the ice edge (Vinje 1976–1982). The effect of heat transfer from the warmer water masses flowing in from the west is illustrated by a correspondence between temperature anomalies at the Kola meridian oceanographic section and anomalies in the ice extension in the Barents Sea (Sætersdal & Loeng 1984). Loeng et al. (1983) also found that oceanic temperature changes in the eastern part will occur one year later than in the western part. Helland-Hansen & Nansen (1909) found similar indications, mainly that climatic variations in the Barents Sea lagged one year with respect to the Lofoten section. Oceanographic features will accordingly play an important role in long-term ice forecasting.

If the Barents Sea Low center is located in the central part of the Barents Sea, and if the circulation is more intense than normal, we will have an increased surface speed of warmer water as well as relatively high atmospheric temperatures and northward drift in the eastern part of the Barents Sea. This was the case, for example, in March 1985 when there occurred a broad opening, extending up to the northernmost tip of Novaja Zemlja (DNMI Ice Maps). If on the other hand the air pressure is high during the freezing season, there will be a retarding effect on the eastward flowing warmer water, and cold, continental winds may favour the cooling and freezing over of large areas. This happened in 1979 when an extreme western position of the ice edge was observed towards the end of the freezing season (Vinje 1980).

A deviation from the mean ice limit has a traceable persistency in the Barents Sea over a period of 2–3 months (Lemke 1980). This important prognostic feature, here given a statistical number, is so predominant that it has for a long time (on an empirical basis) been used for the planning of expeditions to the area.

The seasonal variation of the ice-covered area of the Barents Sea and the adjacent part of the Arctic Ocean shows great interannual variations (Fig. 2, Area: 68°N–83°N, 10°E–55°E). The seasonal variation of the sea ice distribution shows a maximum/minimum extension in late April/August, respectively.

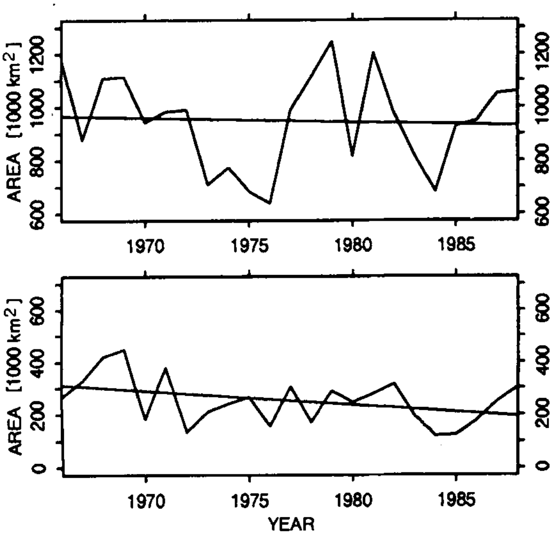


Fig. 2. Annual ice-covered area in the Barents Sea and the adjacent part of the Arctic Ocean (68N–83°N, 10E–55°E), as observed for the period 1966–1988 at the end of April (upper) and at the end of August (lower). The database is the same as for Fig. 1.

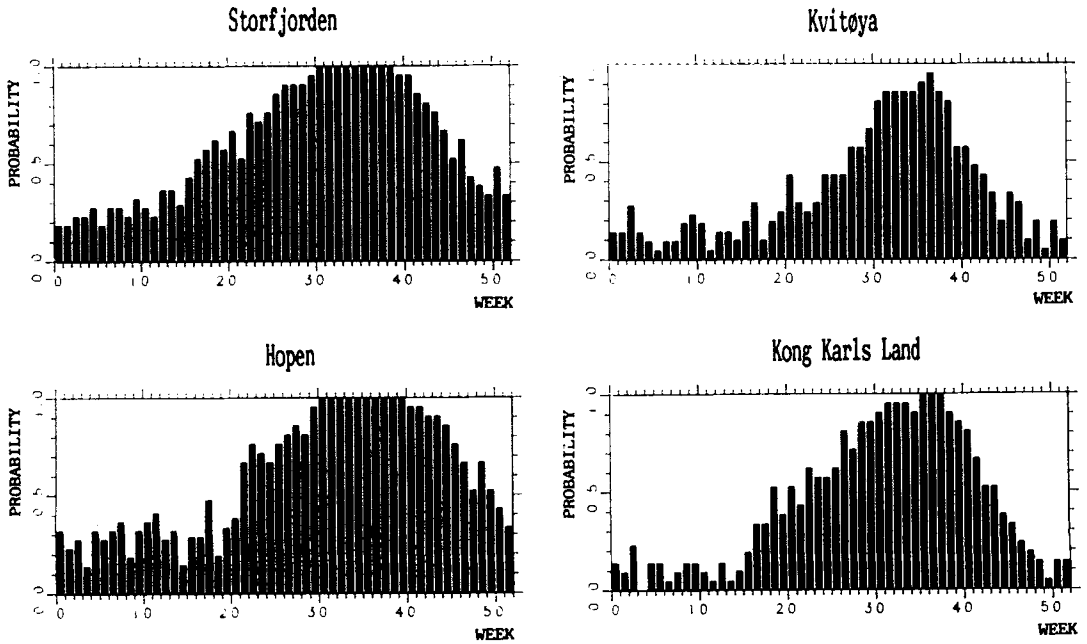
The ice-covered area in late April was during 1973–1976 about 700,000 km^2 and about 1150,000 km^2 in 1969 and 1979, revealing a variation of as much as 400,000–500,000 km^2 in the annual maximum extension over a period of four years. It is noted that the variability in the maximum extension is far greater than the variability in the minimum extension.

The Barents Sea south of 80°N has been completely ice free for one or two months in 1972, 1976, 1984, 1985, and 1986. The long-term variation from the turn of the century indicates that the extension at the end of the melting season has been reduced markedly (Sear 1988). A corresponding yearly reduction of 5.4 ± 2.7 km^2/year is found from correlation studies for the last 23 years (1966–1988) from the area represented by Fig. 2 (late August). Thus, the minimum ice extension in our area has been reduced by as much as 40% during this period. The maximum extension (late April), however, shows no significant change over the same period. The increasing difference between the annual minimum and maximum extension indicates that a reduction/increment of the ice thickness/melting may have taken place in the Barents Sea over the two last decades.

Polynyas

One of the main areas for polynya occurrence in the western Barents Sea is found in the inner part of Storfjorden. When northerly winds prevail, newly formed ice is continuously transported southwards resulting in an opening outside of the rim of land-fast ice. Because of the considerable ice formation that takes place, there occurs a continuous brine production which results in vertical convection and the formation of cold, dense bottom water (Midttun 1985; Quadfasel et al. 1988).

Polynyas are also formed on the lee side of the smaller islands in the archipelago. These are consequently not stable but change position with the wind direction. Hopen, Kong Karls Land, Storøya, and Kvitøya are areas where polynyas may frequently develop. Fig. 3 shows the probability of encountering open water (polynyas) in parts of these areas. Ice-free areas are observed during the winter season 10%–30% of the time,



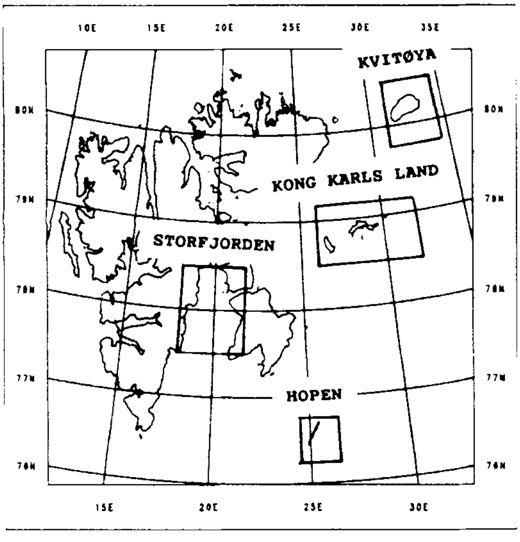


Fig. 3. The probability of encountering polynyas of any size at four various locations in the Barents Sea for the period 1966–1988. The database is the same as for Fig. 1.

and the probability does not change significantly during the winter. (There is, of course, a marked peak during the melting season.)

As north-northeasterly winds are most frequent over the area during the cold season (Gorshkov & Faleev 1980), a steady divergence occurs in the ice fields with a corresponding formation of leads (Fig. 4). The total open areas in the ice fields of the Barents Sea may therefore be considerably larger than the area of the polynyas. In addition, we also have on a smaller scale the effect of the twice daily tidal divergences in the ice field, in this case particularly over shallow banks or near the islands.

Composition

Thickness

As mentioned above, the ice fields may constitute quite a complex picture, containing ice of various age and origin. As the ice drift in the Barents Sea is mainly wind-driven, the ice exchange with the neighbouring seas is to a large extent determined by the atmospheric stress field. The Barents Sea may on an annual average be an ice source for the Arctic Ocean and an ice sink for the Kara Sea over periods of about a decade (Zacharov 1976). However, on the average, there is a seasonal variation indicating a reversed ice exchange with the neighbouring seas during the warmer season (Vinje 1988).

The dominant ice flow in the Arctic is the Transpolar Ice Drift Stream which pours from 4000 to 5000 km^3 ice per year into the Greenland

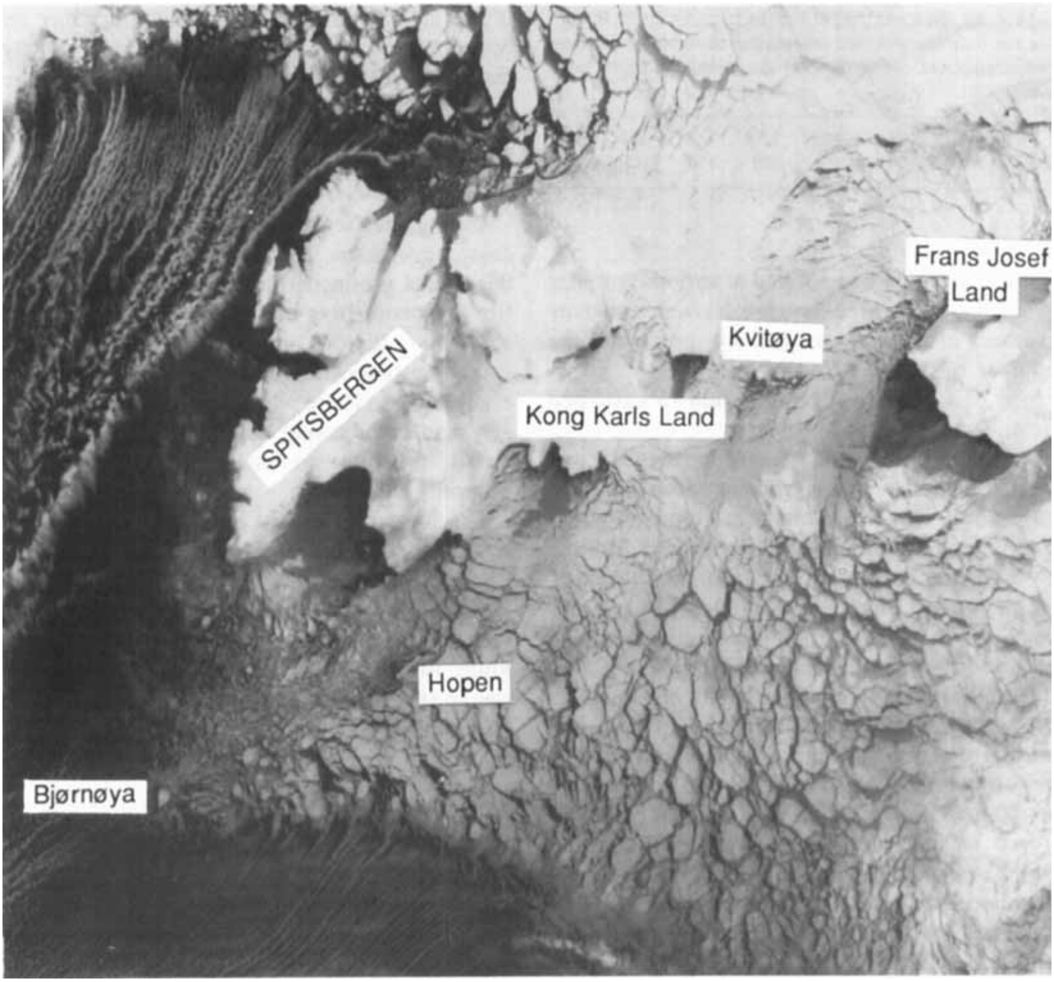


Fig. 4. Infrared NOAA AVHRR image showing lee polynyas and fracture patterns in the Barents Sea during a period with northerly winds, 27 February, 1987.

Sea (Vinje & Finnekåsa 1986). (This corresponds to a fresh water supply to the Greenland Sea of the same order as the Amazon discharge.) In comparison, the average annual transport from the Kara Sea to the Barents Sea is on the order of 500 km^3, and from the Barents Sea to the Arctic Ocean only 30–40 km^3 per year (Vinje 1988). Although the latter two figures are small, their variations play an important role for the stability of the water column as well as for the local heat budget, particularly because of the shallowness of this sea.

The thickest ice encountered in the Barents Sea is generally imported from the Arctic Ocean. This influx of perennial or multi-year ice may be extra large during conditions with persistent northerly winds in the region. This will affect the heat budget of the area to a great extent, resulting in a larger extent of sea ice the following summer season. Such a succession of events took place, for instance, in 1988–89 when the ice fields at the end of July 1989 showed a 60-years' extreme southern extension of multi-year ice (Volkov & Vinje 1990).

The thickness of the land-fast ice at Hopen reached a maximum of 173 cm at the end of April during the extreme cold season of 1968–69 (Vinje 1985). Based on the relationship between the ice thickness and the accumulated degree-days at this site, we have calculated the long term average ice

Table 1. Ice thickness [cm] at various latitudes in the Barents Sea based on the observed relationship between ice thickness and accumulated degree-days for the period 1967–1980 (after Vinje 1988).

Latitude	75°N	76°N	77°N	78°N	79°N	80°N
Thickness (cm)	95	105	120	130	145	170

thickness that may be formed at various latitudes (Table 1). The degree-days are derived from daily averages of temperature and are the sum of the number of degrees Celsius below zero experienced each day during the period of sustained frost.

Table 1 illustrates the effect of the considerable meridional temperature gradient in the area. We have noted that the extreme at Hopen, 76.5°N, was 173 cm. Comparing this figure with Table 1 gives us an indication of the maximum deviation we may expect from the mean (60 cm). The multi-year ice drifting in from the Arctic Ocean is generally 2–4 m thick. In addition there is a considerable formation of ice in the lee polynyas to the south of the islands during conditions with northerly winds. Typically, 20 cm of new ice may be formed per day during cold spells. This may result in an accumulation of about 6 m of newly-frozen ice per month in the steadily unveiled water in a polynya. Martin & Cavalieri (1989) calculate a new ice formation of 10–17 m in the polynya near Frans Josef Land for the whole freezing season. If northerly winds prevail for some time, the result may be long lee-belts of thinner ice, bordered by thicker winter ice or multi-year ice from the Arctic Ocean (Fig. 4). Thus the polynya effect may influence the ice thickness distribution in the Barents Sea to a considerable extent.

An additional ice producing agent is the tidal effect which twice daily opens and closes the space between the ice floes. This may cause an accumulation of frazil ice along the rim of the ice floes, illustrated by the often observed raised ice floe edges.

Floe size and fast ice distribution

Wave effects produce small-sized ice floes of 10–50 m in the outer 10–50 km of the ice margin. This active disintegration zone is most narrow in areas with great water depths, while it becomes particularly broad in areas where the waves are increased by a rising shelf break, for instance on Spitsbergenbanken. The wave energy is rapidly dissipated and further inward from the active wave zone only the long waves with periods of 15–25 s penetrate; accordingly, the floe size also increases from several hundred metres to several kilometres.

During the freezing season off-ice winds will cause divergence and refreezing between the floes, thus producing larger floes of ice breccia (ice floes consisting of smaller floes of different age). In addition, a formation of ice bands extending perpendicular to the main edge generally occurs. There are a number of hypotheses and models indicating how these bands are produced, but so far no single hypothesis has been generally accepted.

The formation of fast ice in the Barents Sea occurs in sheltered areas. The most important places are the inner part of the Storfjorden, the triangle between Nordaustlandet, Kong Karls Land and Barentsøya, and the archipelago of Frans Josef Land. A smaller, frequently recurring, fast ice area is found over the shallow shelf just south of Kongsøya.

Ridging

Ridging events may take place with onshore winds. Rubble fields, possibly due to this effect, are often observed in the NW part of the Barents Sea, and ice piles 10–15 m high have been observed on the shore of the islands. The height of the ridging is also illustrated by the fact that fairly large accumulations of ice blocks may be observed on top of 7–15 m high grounded icebergs (Løset et al. 1988; Vinje et al. 1989). Ice ridging also occurs in convergences caused by wind, currents, and land constraints.

The most frequent ridge height observed from the meteorological station on Hopen is 2 m. The most frequent ridge density in the ice fields of the Barents Sea as observed from a long term project using aircraft is about 10 km^{-1}. The frequency distribution is fairly broad, with a maximum of 26 km^{-1} (Vinje 1985).

Ice bottom topography

The ice bottom topography has been studied extensively using a scanning sonar (Johnsen 1989). The measurements indicate that the multi-year ice has a relatively smoother bottom top-

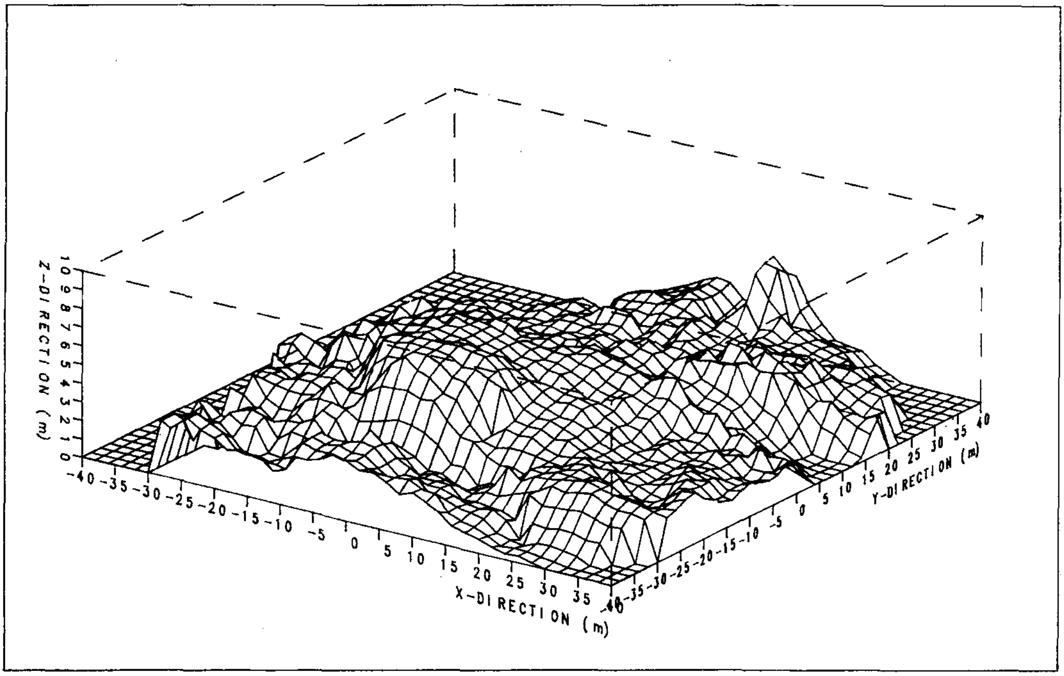


Fig. 5. Typical bottom topography features of multi-year ice as mapped with a scanning sonar at 78.8°N and 29.7°E September 1989. The dimensions of the box are 40 m × 40 m × 10 m. The flat areas in the corners are areas with no data due to ship or ridge shadows. The grid spacing is 2 m.

ography than the first-year ice in the marginal ice zone. This may reflect the effect of thermodynamic processes having acted over a longer time period for the former ice type. A typical sample of a multi-year ice bottom topography is given in Fig. 5.

An ice field is considered to have a very high oil pooling capacity due to capture in the leads, under and upon the ice floes. Johnsen & Vinje (1987) estimated that the domes under the ice floes 10–90 km from the outer ice edge could trap as much as 0.08 m^3/m^2 on an average. Martin et al. (1976) found in a laboratory test that 0.004 m^3/m^2 oil could be pumped upon pancake ice. Pancake ice is a very common ice type in the outer 10 km of the ice margin during the freezing season.

Terrigenous and biogenous material in sea ice

When the freezing starts in September/October, the coastal water masses contain a maximum of suspended material delivered by the run-off from land and glaciers. The material in the upper water layers will be embedded in the new frozen ice to a variable extent, dependent upon the concentration of suspended material and the intensity of the freezing. The occurrence of sediment-laden ice is quite common in the Arctic Ocean (Nansen 1904, 1906; Thiede 1988) as well as in the Barents Sea. Elverhøi & Solheim (1983) report that the material in the latter area is both of local and external origin. This is in accordance with the influx of ice to this region from other shelf areas in the Arctic. As mentioned above, the greater part of the potentially sediment-laden ice import generally comes from the Kara Sea. However, temporary deviations from the normal flux pattern indicate that considerable amounts may at times also come from other parts of the Siberian shelf. This was the case in 1989 (Volkov & Vinje 1990).

Another phenomenon which was observed on Spitsbergenbanken in March 1987 is the accumulation of material in ice ridges which may act as traps for material brought up under the ice by turbulent currents over shallow banks. When these newly formed, nonconsolidated ridges are broken up, mud-accumulation between the ice blocks and a number of dark brown ice surfaces along the track of the ship may be observed. The observation of scattered muddy ice floes in the Barents Sea ice fields is not a rare event.

The accumulation of bottom material trapped in newly formed ridges is a phenomenon that surely occurs during freezing in all shallow areas. When these ridges are consolidated during the following summer, storage of sediments as well as of nutrients, bacteria, or other biological material such as eggs and larvae will occur. Being transported over long distances, this storage may form a basis for new biological growth in new areas during the melting season, then mainly in the marginal ice zone. The potential of an ice field to carry trapped terrigenous or biogenous material is indicated by the above-mentioned great number of ridges observed per km. Garrison et al. (1989) observed that a high concentration of organisms in sea ice may be the result of accumulation of frazil ice. The highest concentration of organisms should accordingly be observed along the downwind border of a polynya or along the edge of ice floes where we will have an accumulation of frazil ice.

Another structure-forming agent is the melting and freezing that takes place in perennial ice. Freezing takes place at the bottom due to the conducted radiative heat loss during winter. Melting mainly takes place on the surface due to the positive radiation balance during the summer. This repeated melting and freezing over a period of some years results in an upward transport of sediments, nutrients, and biological material. This may explain the frequent observations of very muddy surfaces or marked layering of dark material in the column of perennial ice.

Ice drift

Tidal or inertial effects play an important role for the daily movement of the ice, particularly over the shallow shelves, in passages, or along the coast. The movement of the ice fields due to tidal or inertial effects is clearly illustrated by the trails formed by grounded icebergs (Fig. 6). The diameters of the loops vary between 5 km and 15 km, depending on the corresponding water depth and

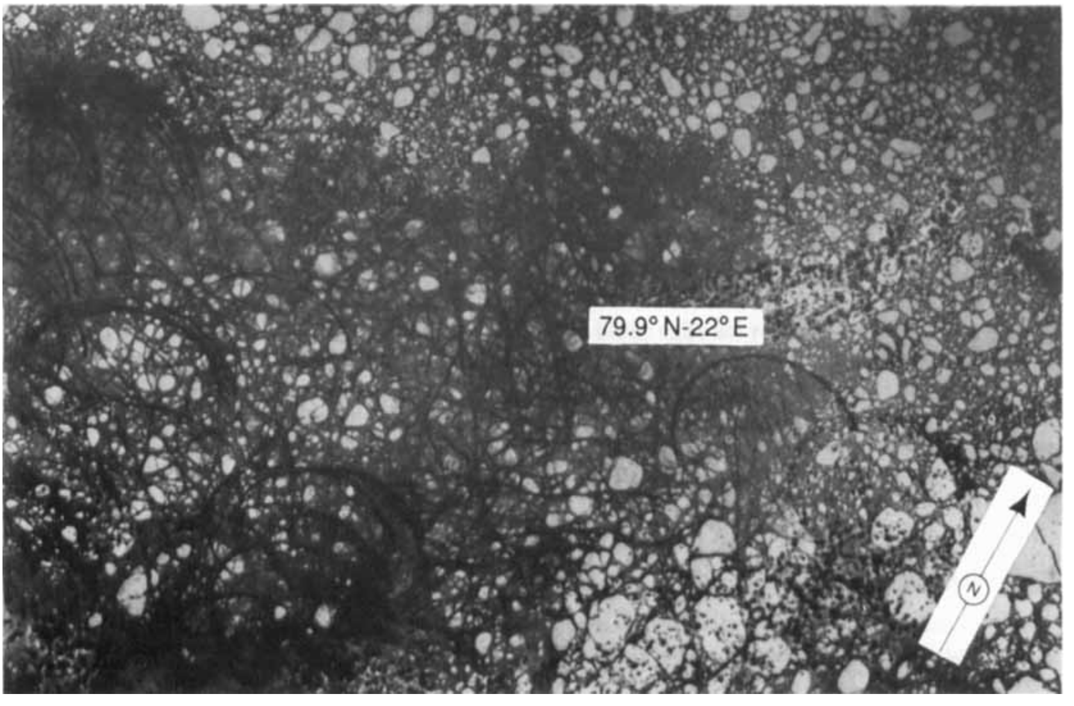


Fig. 6. Trails after grounded icebergs on Spitsbergenbanken caused by tidal or inertial effects in the moving ice field. Landsat TM image 1 June, 1988. The loop diameter is about 15 km.

wind stress. If the number of grounded icebergs is large, the grinding of ice floes may have an effect on the floe size distribution in the area.

The twice-daily tidal wave causes a periodic divergence and convergence over shallow banks, near the coast or in narrow passages. This is a well-known and marked phenomenon which has been taken advantage of for a long time by ice navigators. The tidal wave may also at times produce double, mushroom-shaped eddies. This has repeatedly been observed near the southern tip of Hopen (Vinje 1985).

Information on the ice drift in the Barents Sea has been collected mainly from ice drift buoys since 1975. To illustrate the variance caused by the passage of lows, the drift of three buoys deployed at the same time and longitude but at different latitudes can be considered (Fig. 7). The

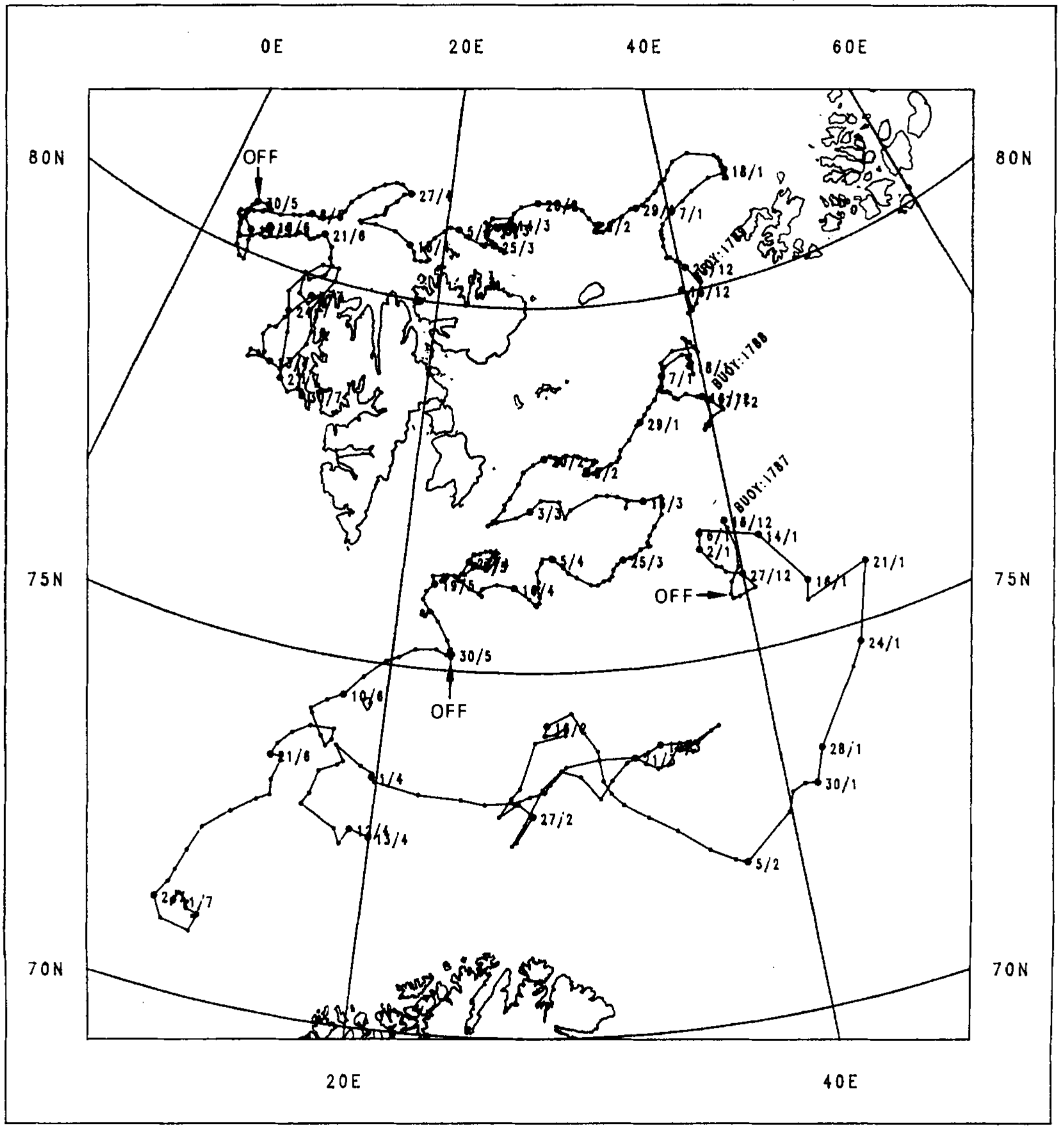


Fig. 7. Drift of ICEXAIR-buoys in 1986/87. The buoys were parachuted on ice floes 16 December 1986 at 76.8°N, 78.5°N, and 80.0°N along 40°E. OFF indicates when the buoy became afloat. Note the great variety in the drift directions. The dates in the figure are day/month. (After Johnsen & Vinje 1987).

average simultaneous ice velocity in the outher margin is observed to be about 0.2 m s^{-1} with a mean net displacement of 0.06 m s^{-1} towards southwest (Vinje 1988). According to buoy drift observations, the ice velocity amounts to 1–2% of the geostrophic wind speed (Johnsen & Vinje 1987). This is in agreement with wind-induced ice edge displacements such as those observed on satellite images (Vinje 1977). The wind effect on the ice increases towards the margin, probably because of the increased form drag due to wave effects (Stokes drift) as well as to the scattering of ice floes.

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