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Dynamics of the sea ice edge in Davis Strait

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Abstract

Sea ice concentration derived from satellite data were used to quantify sea ice characteristics in the Baffin Bay–Davis Strait–Labrador Sea area. The ice edge in Davis Strait extends from Disko Bay in West Greenland 2500 km south to Newfoundland. The mean intercept at the West Greenland coast between 1979 and 2002 was located at 66.9°N, assuming the ice edge was 85% ice concentration. The shallow banks of West Greenland (>200 m) had, on average, an ice extent covering 30 to 100% of the bank area during March for the 24 year time series. This extent varied in concentration between 39 and 100%. However, intermediate ice concentrations (39–85% ice concentration) covered on average 25% of the banks. The Davis Strait ice edge showed considerable interannual variation correlated with the winter index of the North Atlantic Oscillation and the Arctic Oscillation. No temporal trend in ice extent could be detected over the 24 years. In addition to the ice production on the banks of West Greenland, sea ice produced further north in Baffin Bay was advected to the banks as shown by satellite tracked drifting buoys. Both the local sea ice production and the advected sea ice contributed significantly to sea temperatures and salinities measured during summer on the banks. No correlation between sea ice concentration and plankton abundance could be detected but the recruitment of the offshore cod (*Gadus morhua*) component in South Greenland was negatively correlated to the amount of sea ice in Baffin Bay.

Keywords: Ice edge; Remote sensing; Davis Strait; Sub-Arctic Seas

1. Introduction

Davis Strait located between Greenland and Baffin Island in Canada, displays strong seasonal changes in sea ice extent and concentration, primarily driven by wind and current patterns, and low winter temperatures. Several currents influence Davis Strait. The southward Baffin Island Current along the east coast of Baffin Island contributes cold Arctic water to the western part of Davis Strait. This water mass is transported from the Arctic Ocean through the Canadian high Arctic archipelago and Smith Sound. The eastern part of Davis Strait is influenced by the northward-flowing West Greenland Current, containing mixed cold water from northeast Greenland, and warm Irminger water from the North Atlantic (Reverdin et al., 2003). Together these currents influence sea ice distribution and retreat in the area, reflected by different patterns and extent of sea ice in Davis Strait throughout the sea ice season, where the advection of warm water in the eastern part of Davis Strait results in less pronounced ice coverage along the west coast of Greenland during winter (Parkinson, 1995). This east–west gradient in hydrographic structure creates the ice edge in Davis Strait.

A distinct, mainly north–south oriented sea ice edge beginning at Davis Strait extends from the West Greenland coast southward towards the Labrador Sea.

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Similar to most high latitude ice edges it is a site of high primary and secondary production (Nielsen, 1958). During winter, nutrients are brought up to the shelf areas by upwelling currents and wind forcing on the offshore surface waters (Ribergaard et al., 2004). Annual sea ice cover influences the light conditions, stratification, nutrient availability, and water temperature. The increase in spring daylight induces phytoplankton growth within the nutrient enriched surface layer. This production tends to reach a peak in spring, even though it may continue through summer. Both the strength and longevity of the spring production season are thought to be strongly influenced by the location and extent of the sea ice edge (Overland and Stabeno, 2004). Even though water column mixing before sea ice formation usually provides a sufficient nutrient supply, late break-up of sea ice may still reduce the photosynthetic production (Hansen et al., 2002).

The break-up of the sea ice increases sunlight exposure in the water column which triggers the spring phytoplankton bloom. Thus the timing of the sea ice retreat determines the timing of the primary production bloom (Heide-Jørgensen et al. in press). Early disappearance of sea ice leads to a disadvantageous coupling between phytoplankton and zooplankton, where primary production is not consumed by copepods but rather enters a protozoan-dominated microbial food web (Hansen et al., 2002). It is not known to what extent zooplankton can modify their surface migration and extend their lipid stores to mitigate the effects of variations in the timing of sea ice retreat. Prolonged sea ice coverage increases the albedo and maintains cold sea surface temperatures, which reduce the productivity, as specific temperatures are crucial to the physiology of zooplankton (Huntley and Lopez, 1992).

Over the past 25 years, Arctic sea ice has shown a negative hemispheric trend (Parkinson et al., 1999; Comiso, 2002). However, the extent of sea ice has increased in Baffin Bay during 1950–2001 (Stern and Heide-Jørgensen, 2003) and locally in areas along the West Greenland coast (Heide-Jørgensen and Laidre, 2004). It is likely that the increase in sea ice, together with lower spring seawater temperatures, has strongly influenced production and biomass of forage fish in West Greenland over the past 50 years (Pedersen and Kanneworf, 1995). The banks and the shelf edge along the West Greenland coast of Davis Strait have sustained and currently sustain a significant commercial fishing industry. The Davis Strait has experienced some of the most dramatic natural changes in the fish fauna observed in any sub-Arctic region as evidenced by the rise and fall of the fishery for Atlantic cod (*Gadus morhua*) during the 20th century (Ottersen et al., 2004). Historically, West Greenland was subject to one of the largest cod fisheries in the Atlantic. Between 1910 and 1995 16 nations participated in the cod fishery in this region and total accumulated landings exceeded 10 million tons. The fishery peaked with about 460,000 tons in 1962 in West Greenland, after which it declined steadily with very low catches after 1990 (Horsted, 2000). These fluctuations were likely due to changes in temperature and current regimes in the Iceland–Greenland system, as cod eggs and larvae are believed to be advected from the Icelandic spawning grounds to West Greenland by the warm and saline Irminger current (Hansen, 1949; Buch et al., 1994). The strength and temperature of the Irminger current is influenced by large scale meteorological events, with the North Atlantic Oscillation playing an important role (Deser et al., 2002; Buch et al., 2004). Currently in Davis Strait, Northern shrimp (*Pandalus borealis*) and Greenland halibut (*Reinhardtius hippoglossoides*) are the most important species, with annual landings of magnitude of 100,000 t and 20,000 t, respectively.

Changes in the physical environment of Davis Strait are transmitted through the food web and eventually have dramatic impacts on the society utilizing marine resources. As sea ice is clearly a critical influence on the hydrographic properties and the sequence of nutrient-phytoplankton–zooplankton transfer in Davis Strait, examination of time series data in the region may have the potential to provide information on the dynamics of the ice edge zone in Davis Strait and possibly explain ecological trends in fisheries over the past 30 years. This study examines the spatial patterns and temporal variability of the Davis Strait ice edge, correlates results to the North Atlantic Oscillation and examines the impact of the sea ice on the temperature regimes and biological production on the banks off West Greenland.

2. Methods

Monthly average sea ice concentration data were obtained from the National Snow and Ice Data Center (NSIDC) in Boulder, Colorado for every March (the most frequent month of maximum concentration) from 1979 through 2002. The sea ice concentration was derived by the Bootstrap algorithm (Comiso, 1995) from the passive microwave data collected by the Scanning Multichannel Microwave Radiometer (SSMR, every second day between 1978–1987) and the Special Sensor Microwave Imager (SSMI, daily on several satellites since 1987). Data for the Northern Hemisphere were mapped to a polar stereographic projection (true at

70°N) at a 25 km pixel size. Sea ice data were also imported to a geographic information system (GIS, ESRI ArcINFO 8.3) as raster grids, where the center of each cell received the estimate of average sea ice concentration in that 25 km pixel.

A time series of the mean sea ice concentration for March was created for 1979–2002 for the area between West Greenland and the Labrador coast to explore the variation in the maximum ice concentration over time.

The latitude at which the sea ice edge intersected the coast of West Greenland was identified based on mean March ice concentration images for each year. Values were obtained in each year for two different sea ice concentrations, <40% and <85%, yielding two latitude values.

The 'intermediate ice zone' was delineated by the 40% and 85% mean ice concentration isopleths. Sea ice concentrations <40% are labeled as 'loose pack ice' and >85% as 'dense pack ice'. The shallow banks along the West Greenland coast (<200 m depth) between 69°N and 60°N were delineated based on bathymetric contours interpolated from depth measurements taken during shrimp trawls (K. Wieland, GINR, unpubl. data) (Fig. 1). The 200 m 'bank region' was used as a spatial mask to quantify the proportion of the bank that fell in each of three categories of sea ice cover during the March composite each year: 0–39% sea ice, 40–85% sea ice, and 86–100% sea ice.

The sea ice edge and its variability were defined following the method of Shapiro et al. (2003). A mean (1979–2002) March ice concentration image was constructed and 16 points along the 30–39% contour between West Greenland and Newfoundland were selected. A cubic spline interpolant was passed through these points and re-sampled at 25 km increments to provide the smoothed mean ice edge (Fig. 2, heavy line). The anomalies from the smooth mean are defined as the perpendicular distance from the mean ice edge to a given March ice edge, at 25 km increments along the mean ice edge (Fig. 3A). Thus the anomaly for each March ice edge consists of 98 values. In case the bisector does not intersect the given ice edge anywhere, a missing value is recorded. Out of 98 × 24 = 2352 potential values, there are 2227 actual anomaly values (94.7%) and 125 missing values (5.3%). The missing values tend to lie at the ends of the ice edge, near land. A retreat of the ice edge relative to the mean is a negative anomaly; an advance is a positive anomaly. The anomalies for a given March constitute a spatial pattern (as in Fig. 3), and the pattern evolves in time. A principal component analysis of the anomalies gives the characteristic anomaly patterns (eigenvectors of the anomaly covariance matrix), the

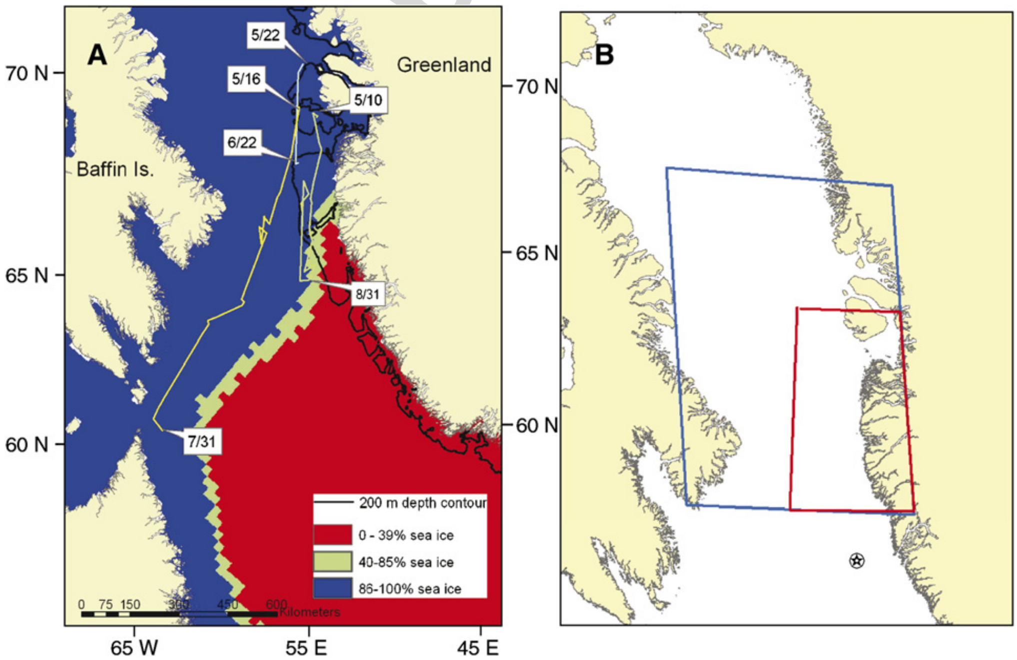

Fig. 1. A. Example of the average ice coverage in three categories in March 2002 in Davis Strait. In addition to the ice edge the dispersal of three drift buoys deployed in May 2002 are shown. The mean speed of the buoys was 0.46 km/h (SD = 0.18) with a mean direction of 186°. B. Delineation of sea ice areas in Baffin Bay and Davis Strait used for quantifying the average sea ice extent in March 1979–2002 by Stern and Heide-Jørgensen (2003). The blue box indicates the large area, red box indicates the medium area and the star shows the position of the hydrographic station.

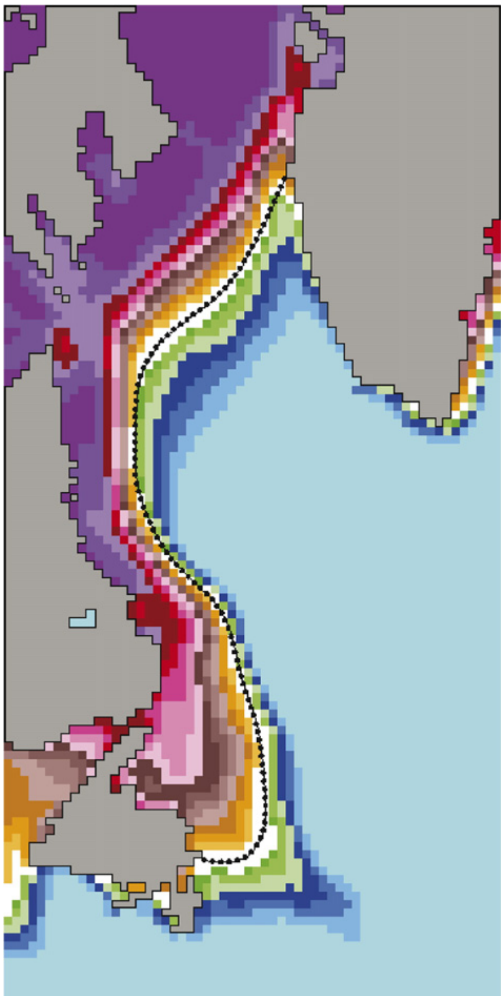

Fig. 2. Mean ice concentration for March in Davis Strait and the Labrador Sea using the average of 24 March (1979–2002) of passive microwave ice concentration data. Greenland is at upper right, Newfoundland at lower left. The ice edge is defined as the 30–39% ice concentration contour. The white pixels have 30–39% ice concentration, and the black dots defining the ice edge are spaced at 25 km. The pixel size in 25 km. The ice concentration is shown in 5% increments from 5 to 100%.

time series associated with each pattern, and the variance (eigenvalue) of each time series, which gives the relative influence of each pattern. There was no clear auto-correlation pattern between years in the ice data, which is probably due to the fact that the sea ice in Davis Strait recedes every summer.

Ice concentration in two large areas in northern Davis Strait and Baffin Bay covering 1953–2001 (Stern and Heide-Jørgensen, 2003) was used for assessing the impact of this pack ice formation on the biophysical parameters on the West Greenland banks (see Fig. 1 for definition of areas and Stern and Heide-Jørgensen (2003) for derivation of data sets for these areas). Sea ice drift in spring was examined by deploying three Argos buoys in water near the pack ice off Disko Island, along the west coast of Greenland, in May 2002. Buoys drifted from the deployment date through August. Mean speed (m/s) and direction (bearing) (weighted by distance) were estimated from all good quality Argos locations (estimated <1 km error).

Hydrographic data from 5 stations across Fyllas Bank (64°N 54°W), Holsteinsborg Deep Channel (66°30N 55°W) and Sukkertoppen (65°30N 54°W) were used to examine the effects of sea ice edge location and residency on sea temperature and salinity. Sea temperatures and salinities in late June and early July during 1979–2002 were averaged over a near-surface depth (0–40 m and 10–50 m) after the recession of the sea ice. Indices of zooplankton abundance including cod larvae from standard stramin net samplings conducted at the hydrographic stations during 1950–1984 were used to examine the effects of sea ice on zooplankton abundance (see Pedersen and Smidt, 2000 for details). A discrete offshore stock of cod in West Greenland, separated from inshore stocks, has shown considerable fluctuations in abundance during the past 100 years (Storr-Paulsen et al., 2004). Recruitment to the Greenland offshore cod stock estimated from Virtual Population Analysis during 1950–1989 (Stein and Borovkov, 2004) was used to identify the relationship between cod recruitment and sea ice conditions. Data on recruitment of Greenland halibut from bottom trawl surveys in Disko Bay and off the West Greenland coast during 1985–2001 (Storr-Paulsen and Jørgensen, 2003) was similarly used to correlate with sea ice conditions. Effects of large-scale atmospheric systems on the ice edge formation were tested by correlations between sea ice extension and winter (December through March) indices of the North Atlantic Oscillation (NAO, Hurrell, 1995) and the Arctic Oscillation (AO, Thompson and Wallace, 1998).

3. Results

The Davis Strait ice edge extended from the West Greenland coast south across the Labrador Sea to Newfoundland. The area that was not covered by the sea ice edge left a coastal field of open water along West Greenland, or the 'land water' (coastal opening bordered by the sea ice to the west), which extended north to Disko Bay (Figs 1 and 2). The >39% ice edge intercepted the West Greenland coast at an average latitude of 65.4°N (range 60.7–66.9°N) between 1979 and 2002. This interception point (at >39% sea ice) demonstrated low variability (SD = 1.45° or 161 km). The >85% ice edge intercepted the West Greenland coast further north at 66.9°N (range, 64.1–68.2 °N, see

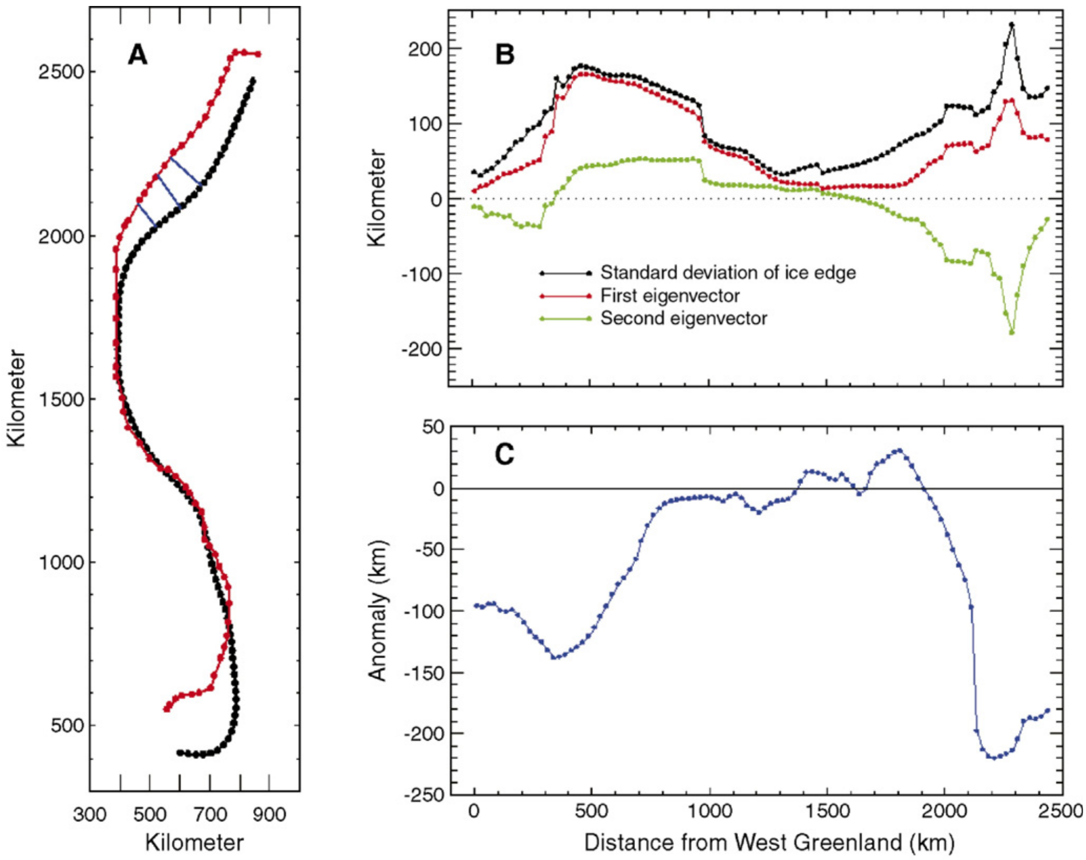

Fig. 3. Procedure for estimating the anomalies from the mean ice edge. A. The black dots are the mean March ice edge, with 25 km spacing. The red dots are the March 1979 ice edge. Three negative anomalies (blue line segments) are shown for illustration. B. Standard deviation (over 24 March) of the anomaly as a function of the distance along the mean ice edge (black curve). The anomaly is measured in kilometers. The figure also shows the first and second eigenvectors. Each eigenvector has an associated time series that gives its evolving contribution to the anomaly pattern. C. The arc length of the mean ice edge is measured from Greenland (0 km) to Newfoundland (~2500 km) and the anomaly of the March 1979 ice edge is shown as a function of the arc length along the mean ice edge.

Fig. 1) and demonstrated even less variability (SD = 0.77° or 84 km). Across the 24-year time series, 1993 was the most extreme ice year as the >85% ice edge first intercepted the coast at 67.0°N (>85% north of 67.0°N). South of that an opening appeared until the >85% ice edge hit the coast again at 63.7°N. Also in 1993 the >39% ice edge reached the southernmost position recorded at 60.7°N. No temporal trend in the position of the ice edge intercept of the coast could be detected.

The March sea ice coverage on the banks (depths <200 m, between 60°N and 68°N and total area 53,750 km^2, Fig. 1) demonstrated large variability in the extent of the intermediate ice zone, ranging from 10 to 58% of the area (average 25%, SD = 12). The dense pack ice coverage (>85%) ranged between 2 to 81% of the area of the banks (average 41%, SD = 16). There was no temporal trend in (>39%) sea ice coverage on the bank region in March, which varied between a minimum of 22,000 km^2 in 1979 and complete coverage (54,000 km^2) in 1983 and 1993. The total area with ice concentration between 39 and 85% of the March ice coverage on the bank region was positively correlated with the winter AO index (ANOVA, $p < 0.05$) yet not significantly correlated with the winter NAO index ($p > 0.05$).

The entire ice edge, or the total number of pixels that are between 40 and 85% ice concentration, in Davis Strait and the northern Labrador Sea (latitude 55°N to 68°N) consisted on average of 17% (range 10–28%, SD = 3.9) of the intermediate pack ice zone category. The total area with intermediate ice conditions in March was strongly correlated with both the AO and the NAO indices (ANOVA, $p < 0.05$). There were no overall temporal trends detected for the 24-year time series.

Positive anomalies from the mean ice edge in Davis Strait were correlated with positive anomalies in the Labrador Sea ($p = 0.01$). Davis Strait was unique as it

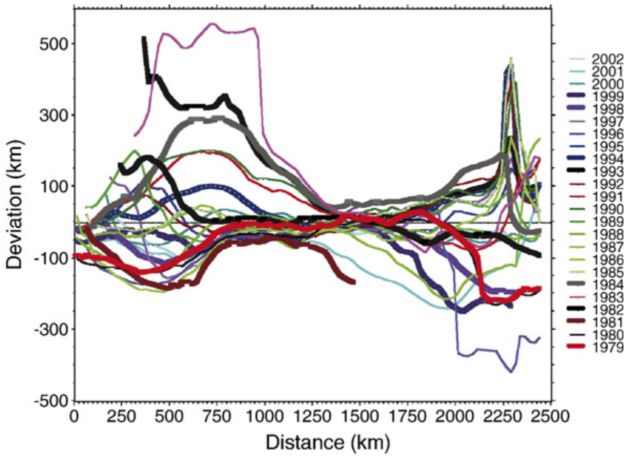

Fig. 4. Anomalies of all 24 March ice edges from 1979 through 2002 shown as a function of the arc length along the mean ice edge. The large positive anomalies at a distance of 500–1000 km correspond to the years when the ice advanced farther than average down the coast of West Greenland. The large positive spike at 2300 km in some years is a lobe that extends east–southeast from Newfoundland. In other years the ice doesn't even reach Newfoundland. Notice that in the middle range of distance (1200–1800 km) there is little variability. This is the north–south portion of the ice edge off the coast of Labrador.

showed twice the interannual variation from the mean anomaly (SD=5155 vs. 2896) detected in the Labrador Sea (Fig. 4). The sum of the 98 ice edge anomalies showed a strong positive correlation with the 24 annual NAO indices ($p<0.01$) and AO indices ($p<0.05$) (Fig. 5). No overall trend for the 24 years of sums of ice anomalies was evident.

The ice edge anomalies are functions of the distance along the mean ice edge, and time (year). A principal component analysis of the anomalies gives a leading eigenvector (spatial pattern) with a large anomaly between 300 and 900 km along the mean ice edge, and another peak around 2300 km, similar to the patterns for the heavy ice years seen in Fig. 4. This leading pattern accounts for 60% of the variance in the ice edge anomalies. The associated time series is significantly correlated (0.57, Students *t*-test) with the winter (Dec.–Mar.) NAO index, pointing to the atmosphere as the main driver of ice edge variability. This indicated that a large part of the March ice edge anomaly was explained by the current winter's atmospheric conditions. There was a less pronounced correlation (0.43) of the previous winter's NAO index with the first principal component time series.

Drifting buoys deployed in Baffin Bay (<71°N) in spring all moved southwest with the surface current (Fig. 1). Regressing summer temperatures (June–July) of the near-surface (10–50 m) and winter ice extent in the March composite for 1979–2002 displayed strong negative correlations ($p<0.05$) along West Greenland (see Fig. 1B for definition of medium area) (Fig. 6). The temperature regimes on Fyllas Bank and at Sukkertoppen were also negatively correlated with the anomalies of the ice edge ($p<0.01$), but the correlation was not significant for Holsteinsborg ($p=0.26$). Similarly the salinity in late June on Fyllas Bank was negatively correlated with the sea ice extent on the bank region in March ($p<0.05$).

The time series of zooplankton abundance indices was discontinued in 1984 and examination of the relation with sea ice conditions was restricted to the pre-satellite data from 1950 to 1978 or SSMR satellite data for 1979–1984 (see Stern and Heide-Jørgensen, 2003). No clear relation between sea ice extension, zooplankton displacement volume, copepod or cod larvae abundance indices could be detected. The recruitment

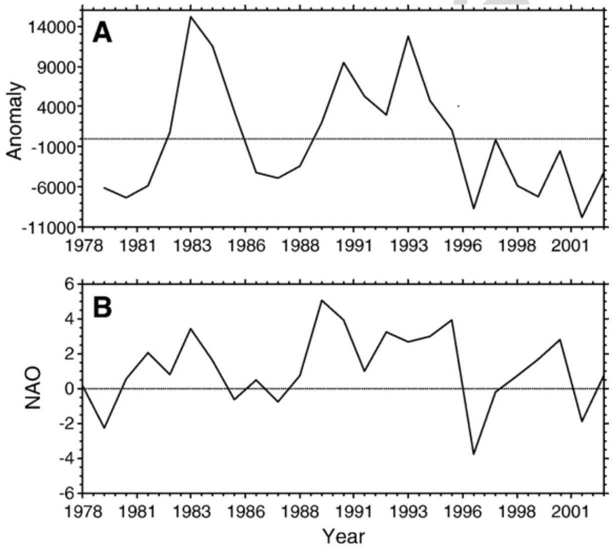

Fig. 5. Time series of (A) the ice edge anomalies between 1978 and 2002, and (B) the North Atlantic Oscillation index (December through March).

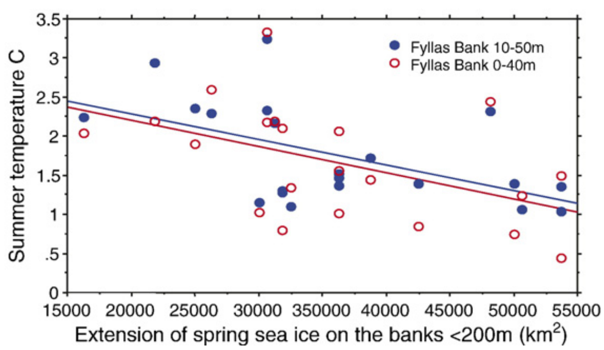

Fig. 6. Near-surface temperature measured in late June at Fyllas Bank (63.967N 52.733W) in relation to the mean March sea ice coverage on the banks for 1979–2002 (temperature data available from GINR www.natur.gl).

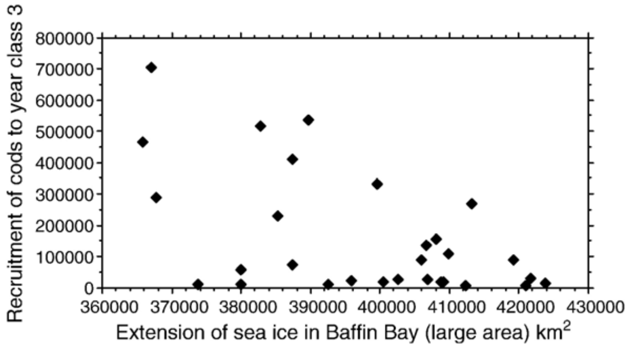

Fig. 7. The relation between the recruitment to the Greenland–Iceland cod stock during 1953–1989 and the ice coverage in Baffin Bay (large area in Fig. 1B, $r^2=0{,}28$). Cod data from Stein and Borovkov (2004).

to the cod stock lasted until 1989 where after the offshore component of cod in Greenland virtually disappeared, and only local inshore stocks remained. The recruitment to this offshore cod component showed a strong negative correlation with the sea ice extension in West Greenland ($p<0.05$, medium area) and in Baffin Bay ($p<0.01$, $r^2=0.28$, large area, Fig. 7) for the period 1953 to 1989. However, the very low recruitment after 1965 makes it obvious that there are some other factors involved in explaining the variation in recruitment. No similar relation between sea ice and Greenland halibut recruitment (age 1, age 2 and age 3+) could be detected.

4. Discussion

The Davis Strait–Labrador Sea ice edge was clearly responsive to changes in the AO and NAO indices, suggesting that these high latitude pressure indicators have a direct effect on the ice edge in the Davis Strait. A strong NAO brings lower temperatures and stronger northerly winds to Baffin Bay–Davis Strait–Labrador Sea, establishing the ice edge farther south. Stern and Heide-Jørgensen (2003) found a strong correlation between the March ice concentration and the one-year lagged winter NAO index in Baffin Bay. The sub-Arctic Davis Strait area is probably more directly affected by the NAO than the high Arctic area of Baffin Bay, however Partington et al. (2003) found that "the winter sub-polar seas respond clearly to the NAO after a lag of 1 year". The heat loss from the ocean to the atmosphere during autumn reduces the ocean surface temperature to the freezing point. The heat loss in the preceding autumn could be caused by a NAO driven transfer of cold low saline upper water masses (Deser et al., 2002). Evidently the spatial evolution of sea ice and sea surface temperature anomalies is to a large extent accounted for by heat fluxes governed by atmospheric forcing. However, frontal zones like the West Greenland part of Davis Strait will often show local and seasonal variability that directly affects the primary production.

There is no discernible temporal trend in the extent of sea ice on the banks, in the intermediate ice zone, in the summed anomalies, or in the coastal intercept of the ice edge. Stern and Heide-Jørgensen (2003) and Heide-Jørgensen and Laidre (2004) found that significant increases in sea ice conditions could be detected in several local areas in Baffin Bay. Parkinson (1995) and Parkinson et al. (1999) found increasing trends in sea ice concentration for the region including Baffin Bay, Davis Strait and the Labrador Sea and similarly, Deser et al. (2000) found increasing ice coverage west of Greenland. Most likely the difference between the ice conditions in Baffin Bay and northern Davis Strait is due to a decrease of the surface air temperature in this region compared to an increase in air temperature south of 69°N (Rigor et al., 2000; Hanna and Capellen, 2003).

The retreat of the sea ice determines the timing of the primary production and the ambient temperature during the zooplankton blooms. The drifting buoys indicate transport by the surface current of winter ice over at least the northern part of the shallow banks (<200 m) in West Greenland. The Davis Strait sea ice contributes through melting to the cooling and freshening of the near-surface waters on the banks in West Greenland in spring. This is evident from the correlation between sea ice coverage in eastern Davis Strait and summer temperatures on the banks. The hydrographic measurements were conducted at variable dates between 1 June and 31 July, and break-up of sea ice varied similarly between March and May. Most of the variation around the regressions was probably due to different periods between the disappearance of the sea ice and the temperature or salinity measurements.

The cold and low salinity water that the sea ice leaves on the near-surface waters on the banks increases the spring stratification of the water masses. A critical component of the link between the extension and residency of sea ice and the production of forage fish is the success of the herbivore plankton (Hansen et al., 2003), especially during the spring period where the sea ice presumably has a cooling effect on the sea surface temperature. Models of the pelagic food chain show that the timing of the break up of sea ice is critically important for the success of the consumption of the spring bloom by upper trophic levels (Hansen et al., 2003). At the same time, both physiological (e.g., McLaren, 1963) and empirical studies (e.g., Huntley and Lopez, 1992) show that growth and reproduction of copepods is sensitive to temperature. Even direct observation of spring temperatures show that copepod growth is more limited by temperature regime than prey availability (Smith and Vidal, 1986; Coyle and Pinchuk, 2002) and a number of studies show that zooplankton production is augmented in warm years (Walsh and McRoy, 1986; Loeng, 1989; Hunt et al., 2002). The spring coupling between primary production and zooplankton seem to be the major transfer of energy to other levels of the pelagic food chain including the commercially important fish larvae (Hansen et al., 2002).

There are not many time series that can be used to assess the biological effects of sea ice coverage in the West Greenland–Davis Strait area and the relevant ones have probably been exhausted in this study. Perhaps the main biological effect of annual sea ice cover is its influence on stratification manifested through changes in energy flux and productivity during the spring phytoplankton bloom. However, there does not appear to be a straight forward relation between sea ice and zooplankton indices in West Greenland, but several factors impact the compatibility of the zooplankton surveys including differences in sampling design, insufficient number of sampling stations, plankton drift and predatory processes. The annual success of the zooplankton is likely better integrated by the recruitment to the cod stock since cod and their larvae are more efficient samplers. A confounding problem with the cod data is that recruitment is determined for the entire east and West Greenland stock, however most of the larval drift goes to the banks in West Greenland where also most catches are taken (Horsted, 2000; Wieland and Hovgård, 2002). The cod recruitment has generally been very low the past 35 years which renders any relationship to physical variables difficult to assess. The recruitment to the offshore component of cod in Greenland is, despite the good correlation with sea ice, primarily dependent on advection of cod larvae from southeast Greenland and Iceland entrained in the Irminger current, influence of warmer currents on the West Greenland banks, fishing morality, and the predation level (Buch et al., 1994). Sea ice on the banks will primarily affect the cod recruitment through cooling and decreased salinities in early summer and perhaps by reduced zooplankton abundance and uncoupling between phyto- and zooplankton. However, on a larger temporal scale several of these effects are clearly also correlated to the sea surface temperature and the time series of atmospheric conditions, as revealed by the NAO.

This study did not address major questions about the spatial evolution of sea ice and the timing of the retreat of the ice. Even though some correlation can be expected between the maxima of ice extent in March and the timing of the build up and retreat of the sea ice, it may still be worthwhile examining the variations in sea ice phenology by integrating the temporal and spatial development of the ice production. Furthermore, the data that are available for elucidating the relationship between sea ice coverage and near-surface temperatures in early summer are less than optimal. Here we used data from standard hydrographic surveys conducted in summer dispersed over two months with different time span from the ice retreat as proxies for measuring the effects of ice on sea temperature. More focused hydrographic sampling are needed to examine to what extent the sea ice coverage determines the near-surface temperature in the spring period that is so critical for the pelagic food web in sub-Arctic seas.

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