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Arctic Ocean outflow shelves in the changing Arctic: A review and perspectives

Christine Michel[a,*], Jim Hamilton[b], Edmond Hansen[c], David Barber[d], Marit Reigstad[e], John Iacozza[d], Lena Seuthe[e], Andrea Niemi[a]

[a] Fisheries and Oceans Canada, Freshwater Institute, 501 University Crescent, Winnipeg, Manitoba R3T 2N6, Canada
[b] Fisheries and Oceans Canada, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada
[c] Norwegian Polar Institute, Hjalmar Johansens gt. 14, 9296 Tromsø, Norway
[d] Centre for Earth Observation Science, Clayton H. Riddell Faculty of Environment, Earth, and Resources, University of Manitoba, Winnipeg, Manitoba R3T 2N2, Canada
[e] UiT The Arctic University of Norway, Breivika, N-9037 Tromsø, Norway

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ABSTRACT

Over the past decade or so, international research efforts, many of which were part of the International Polar Year, have accrued our understanding of the Arctic outflow shelves. The Arctic outflow shelves, namely the East Greenland Shelf (EGS) and the Canadian Arctic Archipelago (CAA), serve as conduits through which Arctic sea ice and waters and their properties are exported to the North Atlantic. These shelves play an important role in thermohaline circulation and global circulation patterns, while being influenced by basin-scale and regional changes taking place in the Arctic. Here, we synthesize the current knowledge on key forcings of primary production and ecosystem processes on the outflow shelves, as they influence their structure and functionalities and, consequently their role in Arctic Ocean productivity and global biogeochemical cycles. For the CAA, a fresh outlook on interannual and decadal physical and biological time-series reveals recent changes in productivity patterns, while an extensive analysis of sea ice conditions over the past 33 years (1980–2012) demonstrates significant declines in multi-year ice and a redistribution of ice types. For the EGS, our analysis shows that sea ice export strongly contributes to structuring spatially diverse productivity regimes. Despite the large heterogeneity in physical and biological processes within and between the outflow shelves, a conceptual model of productivity regimes is proposed, helping identify general productivity patterns and key forcings. The different productivity regimes are expected to respond differently to current and future Arctic change, providing a useful basis upon which to develop predictive scenarios of future productivity states. Current primary production estimates for both outflow shelves very likely underestimate their contribution to total Arctic production.

1. Introduction

The extensive Arctic shelves make up approximately half of the total Arctic Ocean area and, together with adjacent sea shelves, represent one fourth of the World Ocean shelves. Among the Arctic shelves, Arctic outflow shelves, as defined according to the typology of Carmack and Wassmann (2006), constitute the transit routes through which Arctic waters and their properties return to the North Atlantic. The two Arctic outflow shelves, i.e. the Canadian Arctic Archipelago (CAA) and the East Greenland Shelf (EGS) extending into Fram Strait, are crucial to thermohaline circulation in the North Atlantic and global circulation patterns. For example, recurring salinity anomalies in the North Atlantic subpolar gyre have been linked to changes in circulation and properties in the Arctic Ocean (Dickson et al., 1988; Curry and Mauritzen, 2005). Changes in the freshwater Arctic outflow to the North Atlantic also have the potential to affect the large-scale meridional overturning circulation (e.g. Aagaard et al., 1985; Häkkinen, 1995, 1999; Holland et al., 2001; Arzel et al., 2008). Over the past three decades (1979–2008), increased sea ice advection into the Atlantic has been linked with a shift in Arctic mean circulation patterns in summer (Kwok, 2009).

As conduits of water properties and ice exported from the Arctic Ocean, the outflow shelves are influenced by the extensive physical and biogeochemical changes taking place in the Arctic. Most notable are the reduced sea ice extent and thickness (Comiso et al., 2008; Kwok et al., 2009; Maslanik et al., 2011;

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* Corresponding author. Tel.: +1 204 984 8726.
E-mail address: christine.michel@dfo-mpo.gc.ca (C. Michel).

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Parkinson and Comiso, 2013). In this respect, recent studies have attributed spatially variable ice loss in the northern hemisphere to dynamic variability in ice export from the western to the eastern Arctic (Kwok, 2008; Ogi et al., 2008). Ice advection from the Pacific to the Atlantic sectors during summer from 2003 to 2007 was found to account for 7–21% of summer retreat in 2003 and 2006, respectively, with enhanced advection in 2007 (Kwok, 2008). Studies of ice age and drift tracks further illustrate the role of changes in sea ice dynamics in the loss of perennial ice, the latter being attributed to increased transport via the Transpolar Drift in response to anomalous atmospheric forcing in 2007 (Nghiem et al., 2007). Recent research (Barber et al., 2015) also show that the Pacific sector of the Arctic responds to summer melt and delayed fall sea ice formation whereas the Atlantic sector responds to Atlantic water heat flux to the sea ice base throughout the winter season.

Additional physical changes taking place in the Arctic and of importance for production processes include increases in light transmission to surface waters (Nicolaus et al., 2012), freshening of the surface waters (Yamamoto-Kawai et al., 2009; Timmermans et al., 2009; Morison et al., 2012), and increased advection of Pacific waters into the Arctic (Shimada et al., 2006; Woodgate et al., 2012). Resulting impacts on the magnitude and type of primary production are evident. These include increases in primary production, implicit to increases in the extent and duration of the open water period (Pabi et al., 2008; Arrigo and van Dijken, 2011; Barber et al., 2015), and a shift towards a dominance of small phytoplankton cells linked to increased surface stratification (Li et al., 2009). In addition, recent evidence of highly productive under ice phytoplankton blooms (Arrigo et al., 2012) and high diatom sinking export in the central Arctic (Boetius et al., 2013) beckons upward revisions of Arctic production.

The Arctic outflow shelves are also experiencing regional changes in oceanographic and sea ice conditions, influencing biogeochemical cycling within the shelf systems and possibly downstream. Changes associated with interior shelves, upstream of the Arctic outflow shelves, will collectively influence outflow shelf processes. For example, in the Beaufort Sea, immediately upstream of the CAA, upwelling and downwelling events impacting production processes have been linked to wind stress and sea ice velocities, with both upwelling and downwelling showing a gradual intensification from 1979 to 2006 (Yang, 2009). Increasing frequency (Sepp and Jaagus, 2011) and strength (Asplin et al., 2012) of Arctic cyclones, together with a longer open water period, would favor upwelling/downwelling and impact local and regional productivity. Recently, a fourfold increase in primary production in the Beaufort Sea was attributed to fall upwelling in this region (Tremblay et al., 2011). Concurrently, increases in fresh water associated with ice melt or riverine input (Peterson et al., 2002), shifts in melt onset (Wang et al., 2013), and in ice age (Barber et al., 2009) and thickness (Laxon et al., 2013) are taking place on Arctic shelves and collectively influence primary production processes and the biogeochemical cycling of materials.

This paper provides a review of the role, structure and function of the outflow shelves in the changing Arctic Ocean. We first summarize hydrocryospheric processes on the outflow shelves, focussing in particular on recent freshwater transport estimates and changing sea ice conditions, as they constitute key forcings on these shelves. We then investigate primary production and ecosystem processes. Based on both commonalities and differences between and within the outflow shelves, we propose a conceptual model of production regimes which can serve for predictive modeling of future productivity states.

2. Materials and methods

2.1. Measurements and analyses on the North East Greenland Shelf and Fram Strait

Many of the results cited in this paper were obtained from an observational program initiated in 1990 at 79°N in Fram Strait (Hansen et al., 2013; Vinje et al., 1998), expanded to its present form in 1997 (Beszczynska-Möller et al., 2012; de Steur et al., 2014) and still on-going. In addition to maintaining an array of moored instruments, a range of variables are measured during the annual ship-based expedition including CTD sections, various tracers (Dodd et al., 2012; Granskog et al., 2012) and biological data (e.g. Svensen et al., 2011). The CTD, chlorophyll *a* (chl *a*) and nutrient observations presented in this paper were obtained during the IPY project IAOOS-Norway onboard the coastguard vessel KV Svalbard, from 12 April to 28 May 2007 and 19 April to 29 May 2008. The observations were carried out using a SBE911 (ship based) and SBE19+ (helicopter based) CTD. Water samples were collected with a SBE 32 Carousel Water Sampler, or Niskin bottles for stations visited by helicopter. All conductivity/salinity measurements with the CTD were calibrated with salinity measurements of water samples (Portasal). Additionally, the SBE19+ was calibrated against the SBE911 at selected depths. Samples for chl *a* analyses were filtered onto GF/F filters, extracted in methanol, and measured on a Turner 10-AU fluorometer (Turner Designs) calibrated with chl *a* standard (Sigma S6144). Samples for nutrient analysis were frozen at −20 °C, and analyzed by standard seawater methods using a Flow Solution IV analyzer from O.I. Analytical, calibrated using reference sea water from Ocean Scientific International Ltd. UK.

2.2. Measurements and analyses in the Canadian Arctic Archipelago

2.2.1. Moored instrumentation

An array of instrumented moorings was maintained across Barrow Strait at 91°W from 1998 to 2011 (Prinsenberg and Hamilton, 2005; Peterson et al., 2012). Different sites across the Strait were instrumented for varying numbers of years. The South site, located along the 150 m contour 8 km from the shores of Somerset Island was instrumented for 10 years, and the North site along the 200 m contour 6 km from the coast of Devon Island, was instrumented for 7 years. The main instruments were upward looking ADCPs for currents and CTDs moored at nominal depths of 40, 80 and 150 m for water temperature and salinity measurements. Integrated pole compasses were used on the ADCPs to provide current direction (Hamilton, 2001) and the bottom tracking feature was implemented to provide ice drift speed. The ADCP backscatter signal was also used to estimate zooplankton abundance. In this paper we report the zooplankton biomass index developed by Hamilton et al. (2013). The authors' derivation and detailed methodology can be found in the aforementioned reference.

In addition to these moorings, a moored Icycler profiler (Fowler et al., 2004) provided profiles of the upper 46 m at the South site during a 2-year period. Icycler consists of a moored winch that allows a sensor float to rise to within a few meters of the ice once a day, taking detailed measurements as it rises. A sonar on the sensor float triggers the termination of the profile and the float is reeled back down to 46 m depth, below potential hazards presented by ice ridges that sweep down through Barrow Strait. The sensor float was equipped with a CTD and fluorometer that was chemically protected from biological growth. Since no *in situ* calibration could be performed, the fluorescence records provide relative estimates of phytoplankton concentration. Small offsets were applied to the fluorescence records in both years to bring the April 1st value to zero, and post deployment checks indicated no significant change in the offset in either year. These moorings provided data for evaluating the magnitude and variability of freshwater, heat and volume fluxes through the Barrow Strait/Lancaster Sound gateway (Peterson et al., 2012), and to explore interannual variability and possible trends in physical and biological properties, as described in this paper.

2.2.2. Sea ice analysis

Weekly regional ice charts for the Canadian Arctic (including the western Arctic, Eastern Arctic, Hudson Bay and the East Coast) were obtained from the Canadian Ice Service (CIS) in digital format (http://www.ec.gc.ca/glaces-ice/). The charts are produced for tactical planning and operational purposes and represent an estimate of the ice conditions at the time of production (Canadian Ice Services, 2005). Production of the ice charts integrate all the information available at the time, including but not limited to weather conditions, visual observations and satellite/aircraft imagery (primarily synthetic aperture radar (SAR) sensors), to characterize the sea ice conditions in the Canadian Arctic. Throughout the study period, the quality and quantity of information used to produce the ice charts has improved, i.e. data sets after 1995 include RadarSat-1 data, which may impact the quality of the sea ice information. Sea ice conditions are charted using the World Meteorological Organization (WMO) egg code, which provides information on sea ice concentration, stage of development or ice type, and floe size or form of development (Canadian Ice Services, 2005). This information is available in vector format, providing the sea ice information in discrete polygons. These data have been used effectively in previous studies examining the spatial and temporal sea ice characteristics in the Arctic (i.e. Stirling et al., 1999; Barber and Iacozza, 2004; Gagnon and Gough, 2005; Howell et al., 2008).

For the current study, ice charts were obtained over a 33-year period, from 1980 to 2012. Ice charts produced prior to 1980 used a different method of coding information, which limits the number of ice characteristics available for analysis. For this analysis, ice charts dated closest to the last week of March (representing maximum ice extent) and the last week of September (representing minimum ice extent) were selected. In September, ice charts were not produced for the east coast of Canada, and therefore are excluded from this analysis. The ice charts were downloaded as vector products and merged in a Geographic Information System (or GIS) to produce a single product for the Canadian Arctic region. Any overlapping areas were removed from the ice charts. The merged charts were then converted to gridded format using a grid spacing of approximately 2 km (true at 67°N, 80°W). Pixels found over land, and those with no data were excluded from further analysis.

For each pixel in the March ice charts, the spatial frequency of multi-year sea ice (defined by the ice charts as ice that has survived a melt season) and first-year sea ice (annual ice defined as a thickness between 30 cm and 120 cm) were estimated. The spatial frequency was defined as the number of times in the 33-year period that a particular pixel consisted of multi-year or first-year sea ice. Therefore a pixel with a frequency of 100% would indicate that the specific ice type was present each year for the entire study period. Decadal trends in ice type concentration in March were computed on a per pixel basis. The ice type concentration was extracted for each pixel and a regression was run over the 33-year study period. A linear line of best fit was computed for the concentration in each pixel and the slope of the line was estimated. Trends (or slopes) in ice concentration were considered significant at the 90% confidence level.

3. Dominant forcings and general hydrography of the outflow shelves

The two Arctic outflow shelves, the EGS and the CAA, are fundamentally different in terms of topography, hydrography and circulation. Yet, there are also similarities between the shelves. The CAA consists of an intricate network of many islands, relatively narrow channels, basins and sills, whereas the linear EGS is nearly 300 km broad at its widest in Fram Strait (Fig. 1). Both shelves extend over a wide latitudinal range which determines the seasonality in irradiance. They are also strongly influenced by ice drift patterns in the Arctic Ocean, i.e. the anticyclonic Beaufort Gyre and the cyclonic Transpolar Drift (e.g. Rigor et al., 2002; Barber et al., 2014). These factors influence productivity patterns and biogeochemical processes, and are discussed further in the next section.

The EGS receives a steady inflow of sea ice and Arctic water either directly over the shelf north of Fram Strait, or via the East Greenland Current (EGC) flowing southward along the shelf slope. The flow occurs over depths ranging from 2500 m at the EGC ocean front, via 200–250 m at the shelf break and in troughs cutting through the EGS, to 40–50 m on the banks. In terms of annual means, the EGC carries an estimated net volume transport of 3.7–11.1 Sv (de Steur et al., 2009, 2014; Fahrbach et al., 2001; Rudels, 1987). In addition to water masses exported from the upper Arctic Ocean proper in the EGC, the shelf receives warm Atlantic water which may reach all the way to the Greenland fjords. Even where the shelf is at its widest at 79°N, Atlantic water recirculating in Fram Strait cause basal melting on the marine terminating glaciers in the region (Mayer et al., 2000). Long-term moorings (1997 onward) from the shelf west of Svalbard to the EGS slope provide invaluable observations on oceanic and heat fluxes through Fram Strait showing less interannual variability in the ECG than the West Spitsbergen Current (Schauer et al., 2004, 2008).

The general hydrography of the CAA is well described in McLaughlin et al. (2004). Shortly, the CAA is influenced by Pacific waters from the west and Arctic waters flowing through the channels of Queen Elizabeth Islands. Water masses are modified during their transit through the Archipelago. All the Arctic throughflow to the North Atlantic via the CAA passes through four gateways, i.e. Nares Strait with a sill depth of 220 m, Barrow Strait/Lancaster Sound (sill depth of 140 m), Cardigan Strait/Hell Gate (sill depths of 180 m and 125 m, respectively) and to the South, Bellot Strait (<24 m deep). The latter is considered of minor importance with respect to total throughflow (Melling, 2000; Melling et al., 2008). The shallow sill depths in the CAA gateways compared to 2600 m for Fram Strait, allow for exchange of upper ocean water properties only through the CAA. It is in this upper layer that fresher water, comprising variable proportions of sea ice melt water (Yamamoto-Kawai et al., 2010), is concentrated. The complex topography of the CAA strongly influences circulation patterns as well as ice motion (see Section 4). Arctic outflows through the CAA gateways are typically on the right hand side in direction of flow, and although the channels are narrow, they are wider than Rossby radius, allowing for simultaneous counterflow on the opposite side, similar to Fram Strait. However, the sum of the three CAA gateway widths is only 27% that of Fram Strait (Melling, 2000). In addition, waters within the CAA are influenced by strong tidal mixing, which has important consequences for production and biological processes in the ice, open water and in polynyas.

Freshwater and sea ice exports through the EGS and CAA are discussed in more details below. Estimates of freshwater transport are based on a reference salinity of 34.8, which represents the salinity of the Arctic Ocean inflow from the Atlantic (Aagaard and Carmack, 1989). For the CAA, the focus here is on recent volume flux measurements in BarrowStrait/Lancaster Sound. A recent

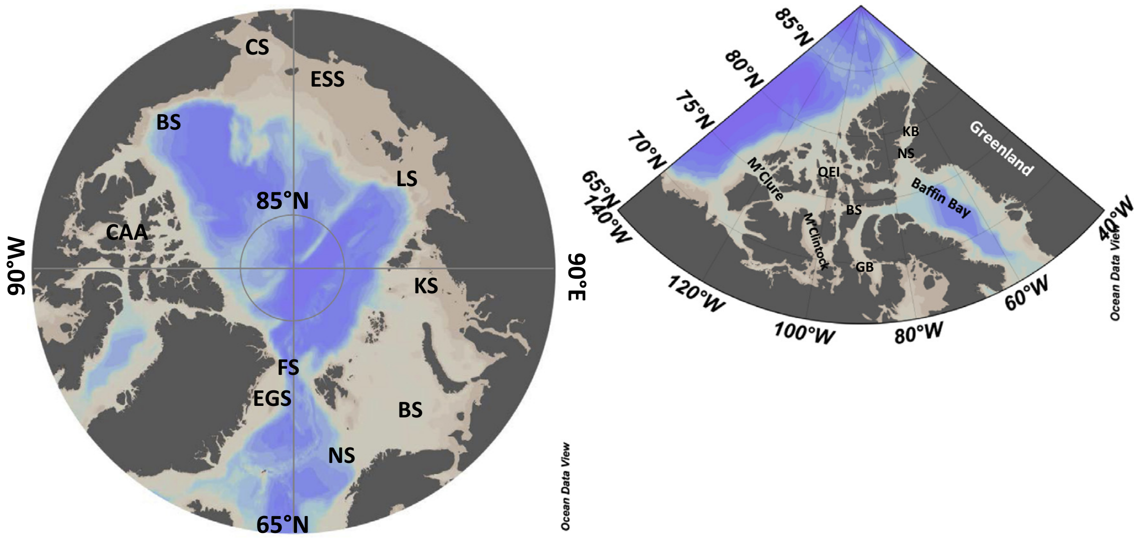

Fig. 1. Map of the Arctic Ocean and the Canadian Arctic Archipelago showing various regions of interest: Beaufort Sea (BS), Chukchi Sea (CS), Eastern Siberian Sea (ESS), Laptev Sea (LS), Kara Sea (KS), Barents Sea (BS), Norwegian Sea (NS), Fram Strait (FS), East Greenland Sea (EGS), Canadian Arctic Archipelago (CAA). Regions of the Canadian Arctic Archipelago are: Queen Elizabeth Islands (QEI), Kane Basin (KB), Nares Strait (NS), Barrow Strait (BS), Gulf of Boothia (GB). This map was created with Ocean Data View; Schlitzer, R., Ocean Data View, http://odv.awi.de, 2014.

synthesis of exchanges in Arctic Ocean gateways (Beszczynska-Möller et al., 2011) provides estimates for the other gateways, summarized as follows. The mean annual volume flux in Nares Strait is estimated between 0.47 and 0.57 Sv for the period 2003–2006, with an additional contribution of 0.25 Sv for the upper water column (0–35 m). Cardigan/Hell Gates account for 0.3 Sv which, together with Barrow Strait accounts for 0.7 Sv, and 2.3–2.6 Sv is exported in Davis Strait (Beszczynska-Möller et al., 2011). Note that slightly different flux estimates are obtained by Peterson et al. (2012) and Curry et al. (2013), with a net volume transport of 0.46 Sv for Barrow Strait (11 years of observations, 1998–2011; Peterson et al., 2012) and 1.6 Sv for Davis Strait (6 years of observations, 2004–2010, Curry et al., 2013).

3.1. The North East Greenland shelf and the East Greenland Current

In the context of coupled physical and ecological characteristics, perhaps the most defining characteristic of the EGS is the advection of sea ice and low salinity water into the region from the Arctic Ocean. The region is at the output end of the Transpolar Drift, each year receiving an area of sea ice corresponding to roughly 10% of the annually-averaged sea ice area of the Arctic Ocean. Of the total Arctic Ocean sea ice and liquid freshwater export, roughly 90% and 40%, respectively, occurs through Fram Strait (Curry et al., 2011; de Steur et al., 2009).

3.1.1. Sea ice and freshwater export

Kwok et al. (2009) estimated the mean annual area outflow of sea ice through Fram Strait at $706 \pm 616 \times 10^3$ km^2 over the 1997–2007 period. There is high interannual and seasonal variability in sea ice export, with the seasonal maximum and minimum occurring in March and August, respectively. No significant trend in the ice area outflow was observed over this 10-year period, a conclusion that remained after revisiting ice flux numbers to include 2009 (Kwok et al., 2013).

The volume export is the product of sea ice concentration (area), drift velocity and ice thickness. Spreen et al. (2009) estimated the ice volume export from satellite data for the period 2003–2008. The 5 year average, estimated at 83 mSv is comparable to earlier estimates of 90 mSv (1990–1996; Vinje et al., 1998), 92 mSv (1950–2000; Vinje, 2001), 70 mSv (1990–1998; Kwok et al., 2004) and to those of Widell et al. (2003) at 76 mSv, and Aagaard and Carmack (1989) at 88 mSv. Similar to previous studies, a strong seasonal cycle shows maximum ice transport in March (251 ± 58 km^3 month^{-1}, 94 ± 22 mSv; Spreen et al., 2009). These recent estimates indicate that any change in ice fluxes since the 1990s remains within the system's variability (Spreen et al., 2009). Overall, the ice volume export estimate through Fram Strait corresponds to ca. 15% of the total annually averaged sea ice volume of the Arctic Ocean.

The export of liquid fresh water through Fram Strait at 79°N was estimated by de Steur et al. (2009) based on a decade of measurements in the EGC. In addition, the authors estimated transport on the shelf based on hydrographic data and model results. Freshwater export occurs above the shelf and in the EGC at the shelf/slope boundary. The 10-year annual mean for the EGC is estimated at 33.6 ± 12.5 mSv, with no significant trend from 1998 to 2008 (de Steur et al., 2009). The shelf component of the total transport is estimated at 25.6 ± 11.3 mSv, for a total transport estimate of 59 mSv. The latter agrees well with estimates based on hydrography and geostrophic transport calculations (Rudels et al., 2008; Marnela et al., 2013). Transports on the shelf and in the EGC appear to be out of phase, resulting in less variability in the total liquid transport than in the individual components. Seasonally, maximum freshwater transport in the EGC and on the shelf occurs during fall, while minimum transport occurs during summer.

Summing the annual mean values for ice (Spreen et al., 2009) and liquid (de Steur et al., 2009) freshwater transports gives an overall average freshwater flux through Fram Strait of 142 mSv, more than half of which in the form of ice.

Dodd et al. (2012) compared 10 sections of concurrent salinity, δ^{18}O, nitrate and phosphate measurements for the period 1997–2011, to assess the contribution of Pacific water, meteoric water, and sea ice melt water in the freshwater outflow through Fram Strait. Meteoric water dominated the outflow, with a typical meteoric to sea ice melt water ratio ranging between −1.5 and −2.8. These authors observed an increased contribution of sea ice melt water in 2009, 2010 and 2011, concurrent with reduced inventories of meteoric water. The 2011 section showed a significant inventory of Pacific fresh water, for the first time since 1998.

Dodd et al. (2012) also found that the time-averaged freshwater composition over the shelf was generally similar to that in the core of the EGC. Maximum Pacific freshwater fractions are typically found in the western part of the section, on the shelf. However, in 2011 the Pacific freshwater maximum was found in the EGC.

3.1.2. Hydrography of the surface waters

Year round hydrographic data from the EGS are sparse. Drifting icebergs and deep pressure ridges represent a hazard to moored instrumentation, generally preventing year round observations of temperature and salinity near the surface. However, instrumentation across the EGC at 79°N in the 45–74 m range (Holfort and Hansen, 2005; de Steur et al., 2009) and on the shelf at the same latitude (Holfort and Hansen, 2005) provide some insight about the hydrography and its seasonal cycle at these sites.

Variation in the EGC front and intrusions of warm recirculating Atlantic water lead to large interannual variability in temperature and salinity in the outer (eastern) fringe of the EGC. Temperatures range from the freezing point to well above 5 °C. Salinity covers a similar range, from values characteristic of Polar Water (as low as 32), to those of Atlantic Water (as high as 35). The seasonal cycle of temperature and salinity on the eastern fringe of the EGC is strongly affected by shifts in the position of the front.

The inner (western) part of the EGC shows little interannual variability in temperature. There is a weak seasonal cycle in temperature which is dominated by the annual cycle of sea ice freezing and melt. The seasonal maximum in temperature occurs in August–September and the minimum (freezing point) extends from late autumn to the winter months. Salinity undergoes a seasonal cycle in the western EGC, with a minimum in October and a maximum in April–May.

In addition to the mooring-based observations referred to above, CTD observations during spring expeditions provide a snapshot of hydrographic conditions across the shelf during the period visited. Fig. 2 shows a section of temperature, salinity and density along 78°50′N during late May 2008. The general physical features discussed above are apparent through the transition from warm, recirculating Atlantic water to cold Polar Water in the surface of the EGC, with cold and fresh water becoming more dominant on the shelf.

3.2. Canadian Arctic Archipelago

3.2.1. Recent trends in transports and freshwater export

The eastward transport of waters though Barrow Strait/Lancaster Sound are confined to the southern side of the Strait. The annual mean easterly flow there is 14 ± 4 cm s^{-1}, and 27 ± 8 cm s^{-1} in late summer, defined as the period between early August and September 21st (Hamilton et al., 2013). Variability in transports through the passages of the CAA has been linked to sea surface slope variability between the Arctic Ocean and northern Baffin Bay (e.g. Prinsenberg and Bennett, 1987; Kleim and Greenberg, 2003). Through modeling, Peterson et al. (2012) showed that 43% of the variability in volume and freshwater transport through Barrow Strait/Lancaster Sound is linked to variability of northeastward winds in the Beaufort Sea, which influence the sea surface slope across the CAA. Wekerle et al. (2013) modeled the variability in monthly transports though Lancaster Sound, with results that are in reasonable agreement with observations ($r = 0.81$) and capture the observed seasonal cycle. These authors identify along-strait sea surface height variability as the main driver of the transports. Furthermore, sea surface height variability upstream of Lancaster Sound is explained by changes in the large scale wind regime in the Arctic Ocean while downstream, in northern Baffin Bay, it is explained by processes influencing sea surface height in the Labrador Sea.

The mean freshwater transport through Lancaster Sound estimated based on 13 years of moored measurements is 34 ± 10 mSv (Peterson et al., 2012) and this includes ca. 2 mSv in ice. A conservative estimate of the freshwater transport through

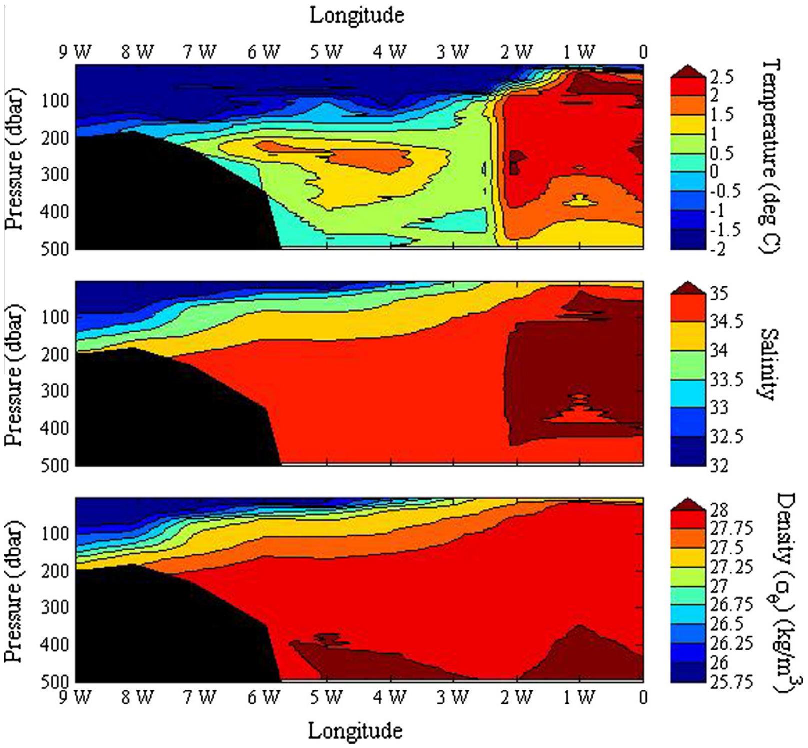

Fig. 2. Temperature (upper panel), salinity (middle panel) and density (lower panel) of a spring (late May 2008) CTD section along 78°50′N, across the East Greenland Current and onto the shelf.

Nares Strait is 28 mSv (Rabe et al., 2010) plus an additional 4 mSv in the form of ice (Melling et al., 2008). Melling et al. (2008) estimate 0.3 Sv for the volume transport through Jones Sound based on 4 years of observations with moored instrumentation. Assuming a freshwater to total water volume ratio that falls between that observed at the other two passages, a crude estimate for the freshwater transport through Jones Sound is ca. 15 mSv. The total freshwater export through the CAA passages is therefore ca. 81 mSv. This is consistent with recent results from six years (2004–2010) of mooring data indicating a net southward freshwater transport through Davis Strait of 93 ± 6 mSv (Curry et al., 2013), with includes export from the CAA and an additional 8 mSv that enters Baffin Bay as glacial melt water. Greenland ice sheet and melt water contribution to the freshwater balance in Baffin Bay is increasing rapidly (Rignot et al., 2011) and there is evidence that the observed acceleration in glacial loss rate is a response to global warming rather than multidecadal natural variability (Hanna et al., 2008).

The aggregate liquid freshwater export through the CAA passages, ca. 75 mSv, is larger than that of Fram Strait but the total (81 mSv, liquid and ice) freshwater transport is ca. 60% of that in Fram Strait. While ice contributes the main proportion of the freshwater export in Fram Strait, the liquid component almost completely dominates freshwater export in the CAA. As proposed by Rudels (2015), this points to the CAA having the largest freshwater export in the context of a diminishing Arctic ice cover and ice export.

The fresh water exported through Fram Strait is carried southward by the EGC and East Greenland Coastal Current entering the western North Atlantic, to be carried northward along the western slope and shelf by the West Greenland Current. Most of this fresh water is lost to the Labrador Sea, so that only about 24 mSv (Curry et al., 2013) passes through Davis Strait into Baffin Bay to circulate cyclonically in the Bay and exit with the Baffin Island Current. Therefore, the export through the CAA dominates the freshwater content of the Baffin Island Current, which flows southward along the western side of Baffin Bay through Davis Strait to merge and impact the properties of the Labrador Current. This points to the importance of freshwater export through the CAA in terms of potential impacts on dense water formation in the Labrador Sea and on the thermohaline circulation.

Freshwater transports estimated from the 13-year (1998–2011) observational program of water properties and transports in Barrow Strait/Lancaster Sound demonstrate high variability both seasonally and interannually. During this period, the overall mean freshwater transport is estimated at 32 mSv ± 6 mSv, with a standard deviation of the monthly mean values of ±24 mSv (Peterson et al., 2012). There are indications of a decreasing trend in freshwater export over the study period (14 mSv per decade; Fig. 3), but the lack of statistical significance due to the high interannual variability precludes definitive conclusions. This emphasizes the challenge of identifying climate-related trends *versus* decadal to multi-decadal variability based on observational series.

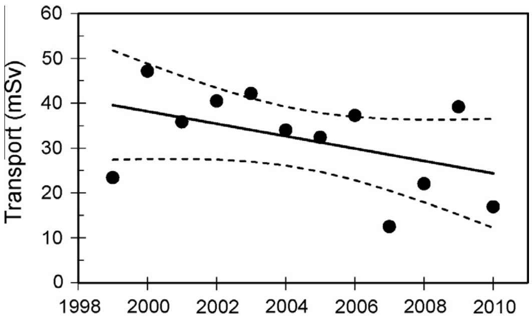

Fig. 3. Annual mean freshwater transports through Barrow Strait from 1999–2011, with a least squares fit to the data showing a decreasing trend in transports of −14 mSv per decade at 87% statistical significance. Dashed lines show 95% confidence interval.

Sea ice is exchanged between the Arctic Ocean and the CAA through its northern and western channels (Melling, 2002; Kwok, 2006; Agnew et al., 2008). Sea ice export from the Arctic Ocean into the Archipelago takes place primarily over the summer months. A recent study quantified areal sea ice exchange between the Arctic Ocean and the CAA via M'Clure Strait and Queen Elizabeth Islands over a 16-year period, from 1997 to 2012, for the months of May to November (Howell et al., 2013). Over this period, the mean flow is estimated at $-1 \pm 21 \times 10^3$ km^2 and $8 \pm 8 \times 10^3$ km^2 at M'Clure Strait and Queen Elizabeth Islands, respectively. Here, note that the negative and positive signs correspond to flow into and from the Arctic Ocean respectively. Of interest is the strong seasonal signal in ice exchange, especially at M'Clure Strait where Arctic Ocean ice inflow occurs between May and September (5×10^3 km^2) and outflow takes place in October–November (7×10^3 km^2). In Queen Elizabeth Islands, most of the flow occurs in August–September, with negligible ice exchange during the other months. Arctic Ocean multi-year ice (MYI) inflow into M'Clure Strait since 2007 is complicated due to the periodic reversal of the Beaufort Sea ice gyre (Lukovich and Barber, 2006). Reversals in the gyre drive MYI ice into McClure Strait only when low ice concentration in the Strait allows for advection. These reversals used to be constrained to August/September but have recently begun to occur through a much longer period of the sea ice annual cycle in the southern Beaufort Sea (Barber et al., 2012). In contrast, increased MYI inflow into Queen Elizabeth Islands is attributed to more open space associated with more open water in the Archipelago, allowing for MYI inflow to take place. Overall, despite reduced Arctic Ocean MYI inflow into M'Clure Strait since 2007, there is continued MYI inflow to the north of the Archipelago at Queen Elizabeth Islands, as well as in Nares Strait (see Section 4). The Queen Elizabeth Islands and Nares Strait MYI is advected southward in the Archipelago channels and in Baffin Bay. The presence and timing of sea ice bridges in Nares Strait control the formation and timing of the North Water (NOW) polynya and also influence the export of MYI freshwater (Barber and Massom, 2007).

3.2.2. Recent trends in oceanographic properties

Analysis of temperature time-series at 40 m, 80 m and near bottom between 2001 and 2011 in southern Barrow Strait where Arctic waters are exported into Baffin Bay, demonstrate high interannual variability in both salinity and temperature over the past decade (Hamilton and Wu, 2013). There are also indications of short-term (decadal) trends. However further analysis is required to establish causal effects and the time-series are short in the context of decadal to multidecadal variability and long-term trends. With respect to temperature, there is a near-bottom warming trend of 0.019 °C y^{-1} ($p < 0.1$), with most of the warming occurring in fall, winter and spring. There is also a suggestion of increased salinity at both 40 m and near-bottom, although in most cases statistical significance is weak. The most robust result is a statistically significant increase in late summer salinity at 40 m (0.05 y^{-1}; $p < 0.05$) over the last decade. There is no statistically significant trend in the timing of break-up or freeze-up on the South side of Barrow Strait, as determined from upward looking moored acoustic Doppler current profilers, over the period of this observational program.

Although the North side of Barrow Strait shares a similar tidal environment with the South side, with M2 and K1 constituents combining to generate typical peak tidal flows of about 50 cm s^{-1}, the mean annual current is westward, and quite weak. Based on

8 years of data, the annual mean current in the upper water column is 2.0 ± 1.3 cm s^{-1} westward (5.8 ± 3.5 cm s^{-1} in late summer). Therefore, rather than representing conditions and properties of waters being exported into the North Atlantic as is the case on the South side, here conditions are representative of the northern side of Lancaster Sound to the east, a highly productive ecosystem. Records indicate that there is also large seasonal variability in current speed and direction, with typically stronger westward currents (2–13 cm s^{-1}) in late summer and fall, and weaker (<5 cm s^{-1}) eastward currents from winter to early summer. Time-series over the past decade or so (1998–2006) show a significant decreasing trend in current speed during late summer and fall (Fig. 4).

Trends in salinity and temperature on the north side of Barrow Strait over the 1999–2006 period are reported by Hamilton and Wu (2013), with lower water column salinity trending higher in both early (0.02 psu y^{-1}) and late (0.04 psu y^{-1}) summer ($p < 0.01$). There is also a warming trend of 0.03 °C y^{-1} ($p < 0.05$) at 80 m through the summer and autumn (Hamilton and Wu, 2013), and Hamilton et al. (2013) report a highly significant warming of the entire water column (average of 40 m, 80 m and 160 m; $p < 0.01$) in early summer. These authors demonstrate a strong connection between early summer water temperature and the timing and productivity of the zooplankton growth season, which will be discussed further in Section 5.3.

4. Sea ice conditions and trends

Ice in the Transpolar Drift may traverse from the Laptev Sea to Fram Strait in less than three years, whereas ice in the Beaufort Gyre may take six or more years to exit the Arctic Ocean through the Archipelago (Rigor et al., 2002). Pfirman et al. (2004) showed a decrease in perennial ice travel times in the late 1990s due to significant loss of thick multi-year ice from the Beaufort Gyre through Fram Strait from 1988 to 1990, also shown by Kwok (2009). The travel time is expected to become shorter as the ice becomes thinner and the drift velocities are increasing (Rampal et al., 2009; Häkkinen et al., 2008). Recent studies of trends in sea ice drift and deformation have attributed accelerated drift both to increased cyclonic activity in the Arctic (Häkkinen et al., 2008), and to changes in internal ice stress associated with a thinner ice cover (Spreen et al., 2011). Rampal et al. (2009) also found acceleration in sea ice motion gradients or deformation. IABP (International arctic Buoy Programme) ice beacon trajectories highlight spatial variability in the Beaufort Gyre and the Transpolar Drift Stream from 1979 to 2002. These dynamical processes associated with the Beaufort Gyre also result in thickening of ice along the Northwestern flank of the CAA where the thickest and oldest ice in the Arctic continues to reside (Barber et al., 2014).

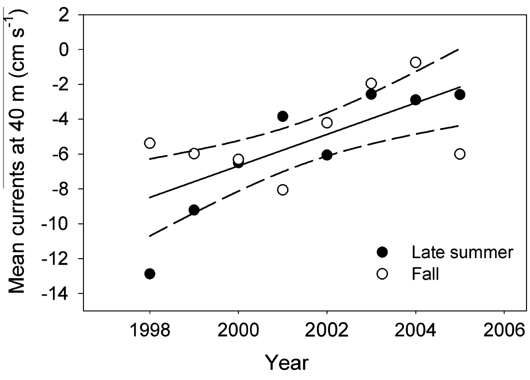

Fig. 4. Late summer and fall mean current speed and direction (positive: eastward, negative: westward) at 40 m on the North side of Barrow Strait over 8 years, from 1998 to 2005. The solid line indicates the slope of the linear regression, with $r^2 = 0.49$, and $p < 0.01$. Dashed lines show 95% confidence interval.

4.1. East Greenland Shelf

Due to the steady advection of sea ice into the region, the seasonality in sea ice extent is less pronounced on the EGS than in other Arctic shelf seas. The location of the summer ice edge is not very different from the winter ice edge (Fig. 5), although ice concentrations are generally much lower during summer than in winter. The marginal ice zone (MIZ) is an important transitional ice edge area on the EGS where intense atmosphere–ice-ocean interactions and biological activity takes place (see Section 5.2). Between 1979 and 2010, during the period of maximum sea ice (i.e. February–April), the Greenland MIZ averaged 98 km in width and narrowed by 43% while moving 158 km west toward the Greenland coast (Strong, 2012). Shifts in the position and width of the MIZ impacts the direction and volume of sea ice drift on the EGS.

The bulk of the sea ice volume exported through Fram Strait originates from the Laptev and East Siberian Seas (Hansen et al., 2013). Export of ice from the central Beaufort Sea occurs in the western part of the Strait, but occurs on a more episodic basis (Pfirman et al., 2004). Hansen et al. (2014) estimated the relative fractions of different ice categories and found that first year ice and ridged ice each constituted, on average, 35–40% of the total amount of ice. Thick ice, surviving at least one melt period, and thin deformed ice constituted the remaining 20–25% of total ice in Fram Strait. Between 2007 and 2011 there was a reduction in the age of pack ice such that the relative amount of thick ridged ice decreased by ca. 25%. The loss of ridged ice was compensated for by an increase in first year ice to 45–50%, and thick ice to 25–30%.

The age of sea ice exported from the EGS has changed over the period where satellites have enabled age classifications to be made (1978–present, Maslanik et al., 2011). In terms of annual averages, the prevailing age of ice exported through Fram Strait during the 1990s was 3–4 y (Hansen et al., 2013). In recent years, the age of exported ice reflects the general loss of perennial ice in the Arctic Ocean (Maslanik et al., 2011). Since 2008, the annually-averaged age of the exported ice has been 1–2 y, with a weak increase in 2011 (Hansen et al., 2013). These results indicate a trend towards a younger age of exported ice in recent years. However, the age of the exported ice is highly variable even on a weekly basis. The ice age data set (Maslanik et al., 2011) comes in relatively high spatial and temporal resolution, with weekly values on a 12.5 grid. Time averaging highlights the long term trend, but masks frequent occurrences of very old ice of 8–9 years of age. Similarly frequent occurrences of much younger ice, i.e. <1 year old, reduces the average age of the ice.

Based on upward looking sonars installed on the seabed, Hansen et al. (2013) described the 1990–2011 thickness evolution of sea ice in Fram Strait. Ice thickness displays large seasonal and interannual variability, with a thinning trend following the recent change in age. During the 1990s the mean ice thickness peaked above 4 m during winter. In 2010 and 2011 the winter maximum was only 2.5 m. In terms of annual averages, the mean ice thickness has decreased from 3.0 m during the 1990s, to 2.0 m in 2010–2011.

The characteristics and fate of ice moving south along the shelf are impacted by several factors that vary seasonally. Southward ice drift is more than twice as fast in winter as it is in summer and ice melt rates range from 0.1 m/month in winter to 0.7 m/month in late summer (Vinje et al., 2002). Observations indicate a decrease in ice thickness moving downstream (Wadhams, 1992; Vinje

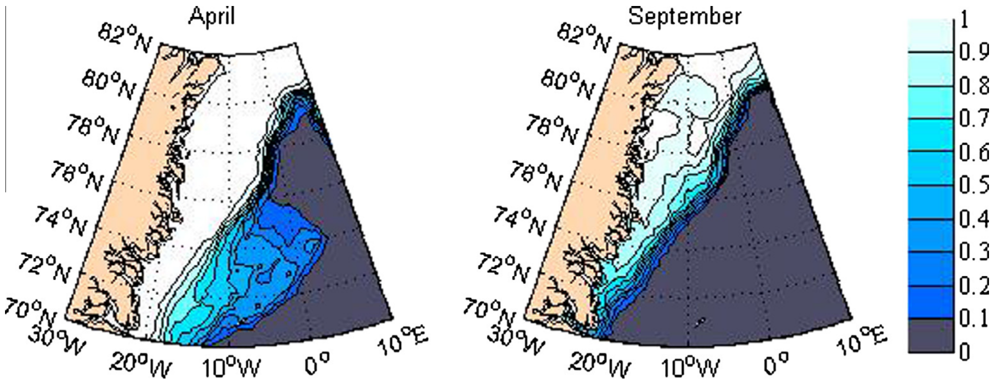

Fig. 5. Probability of encountering ice concentrations greater than 15% on the Northeast Greenland shelf, for April (left) and September (right). Color scale shows probability isolines, from 0.1 (dark blue) to 1 (white). Gray shows the absence of ice. Based on 1978–2006 NSIDC Sea Ice Concentrations from Nimbus-7 SMMR and DSMP SSM/I-SSMIS Passive Microwave Data.

et al., 2002). However, latitudinal differences in ice thickness are blurred during summer due to melting, advection and changing wind patterns (Vinje et al., 2002). As the ice moves southward, a fraction enters the Nordic seas (Dodd et al., 2009) but this quantity is not well known. The ice that remains in the East Greenland Current has largely melted upon reaching the southern tip of Greenland.

4.2. Canadian Arctic Archipelago

The CAA is ice covered during most of the year, with a mixture of perennial MYI and seasonal first-year ice (FYI), the former making up more than half of the total ice-covered area in some years (Canadian Ice Services, 2011).

Fig. 6a shows the MYI ice frequencies, reflecting the dominance of this type of ice in the northern part of the Archipelago (Queen Elizabeth Islands), Western Parry Strait, McClure Strait and M'Clintock Channel. In these regions, ice concentrations often remain high at the end of the melt season (Howell et al., 2013). To the west, the Archipelago is influenced by the presence of MYI in the Beaufort Gyre and in McClure Strait, while localized MYI occurrences are observed in the southern channels due to advection. To the east, MYI exported from the Arctic Ocean can be traced along the western side of Baffin Bay. Fig. 6b, more or less a mirror image of Fig. 6a, shows frequencies of FYI in the CAA. To the west, FYI dominates in the Amundsen Gulf, the Beaufort Sea, Coronation Gulf and Dease Strait. There are significant amounts of FYI in Lancaster Sound and further east in Baffin Bay.

The summer (September) sea ice extent in the CAA has decreased significantly in recent years, likely associated with the shift in ice inflow at M'Clure Strait (Howell et al., 2013). Yet, there is no evidence of statistically significant decreases in MYI due to Arctic Ocean MYI replacing ice lost due to melt (Howell et al., 2009; Tivy et al., 2011; Derksen et al., 2012). Our analysis of sea ice trends over the past three decades (1998–2012) show significant changes in sea ice type extent and distribution in the CAA (Fig. 7). There are significant decreases in MYI extent in the southern and northern (Queen Elizabeth Island) channels of the Canadian Archipelago, especially in M'Clintock Channel and Gulf of Boothia. Following Howell et al. (2013) we surmise that the former is linked to the limited inflow of MYI from the Arctic Ocean into M'Clure Strait, whereas changes in Queen Elizabeth Island are likely related to openings in the channels and the ice becoming more mobile.

There are also smaller decreases in MYI types on the eastern sides of Kane Basin and Nares Strait. These are attributed to changes in advection patterns as long-term ice bridges have not formed in recent years, leading to increased advection of MYI into northern Baffin Bay. As a result, increases in MYI are observed on the western side of Baffin Bay. The latter reflects fundamental changes in the nature of the NOW polynya in recent years. The NOW polynya, historically defined by atmospheric and oceanographic forcings (sensible and latent heat) has transformed, over the past decade, into a region of increased advection of MYI due to the inconsistent formation of the ice bridge in Nares Strait (Barber and Massom, 2007). Such dramatic change is expected to have tremendous impacts on the productive capacity and ecological transfers of this highly productive Arctic region (see Section 7).

Decadal changes in FYI (Fig. 7b) include an increasing trend of FYI types in M'Clintock Channel and Gulf of Boothia, where historically MYI would be confined. There are significant increasing trends in FYI in the circumpolar flaw lead polynya, along the south and west shores of Banks Island, likely due to this polynya expanding in space and time over the past 30 years (Barber et al., 2012). Increasing trends in FYI types to the north, in the Queen Elizabeth Islands, reflect the decrease in MYI in this region. Increasing and decreasing trends in FYI in Baffin Bay and the Labrador Sea are likely due to FYI replacing MYI in the former and increased ablation as sea ice is transported further south with the Labrador Current.

5. Productivity regimes on Arctic outflow shelves

5.1. Controls on primary production

The main controlling factors for primary production in the Arctic Ocean are light and nutrient availability in surface waters, primarily nitrate (Sakshaug, 2004; Codispoti et al., 2013). These key factors are in turn regulated by a complex interplay of processes affecting stratification and mixing, as well as by the presence of sea ice superimposed on seasonal and latitudinal controls on irradiance.

The geographic location of Arctic outflow shelves, which spans >20 latitudinal degrees (60–83°N), determines the seasonality in irradiance as it varies with latitude (Fig. 8). In addition, the temporal and spatial variability in sea ice extent and thickness, the presence of snow, melt pounds, polynyas and leads, and in localized areas riverine input, regulate underwater light conditions.

On Arctic shelves where FYI dominates, the annual cryospheric cycle (i.e. from ice formation to ice melt) imposes a seasonality that is reflected through a seasonal succession in the ecosystem. While the presence of ice imposes limitations on gas and energy transfers at the atmosphere–ocean interface (e.g. Fransson et al., 2009), changes in water column structure associated with ice formation (brine rejection and deep water formation) and melt (surface

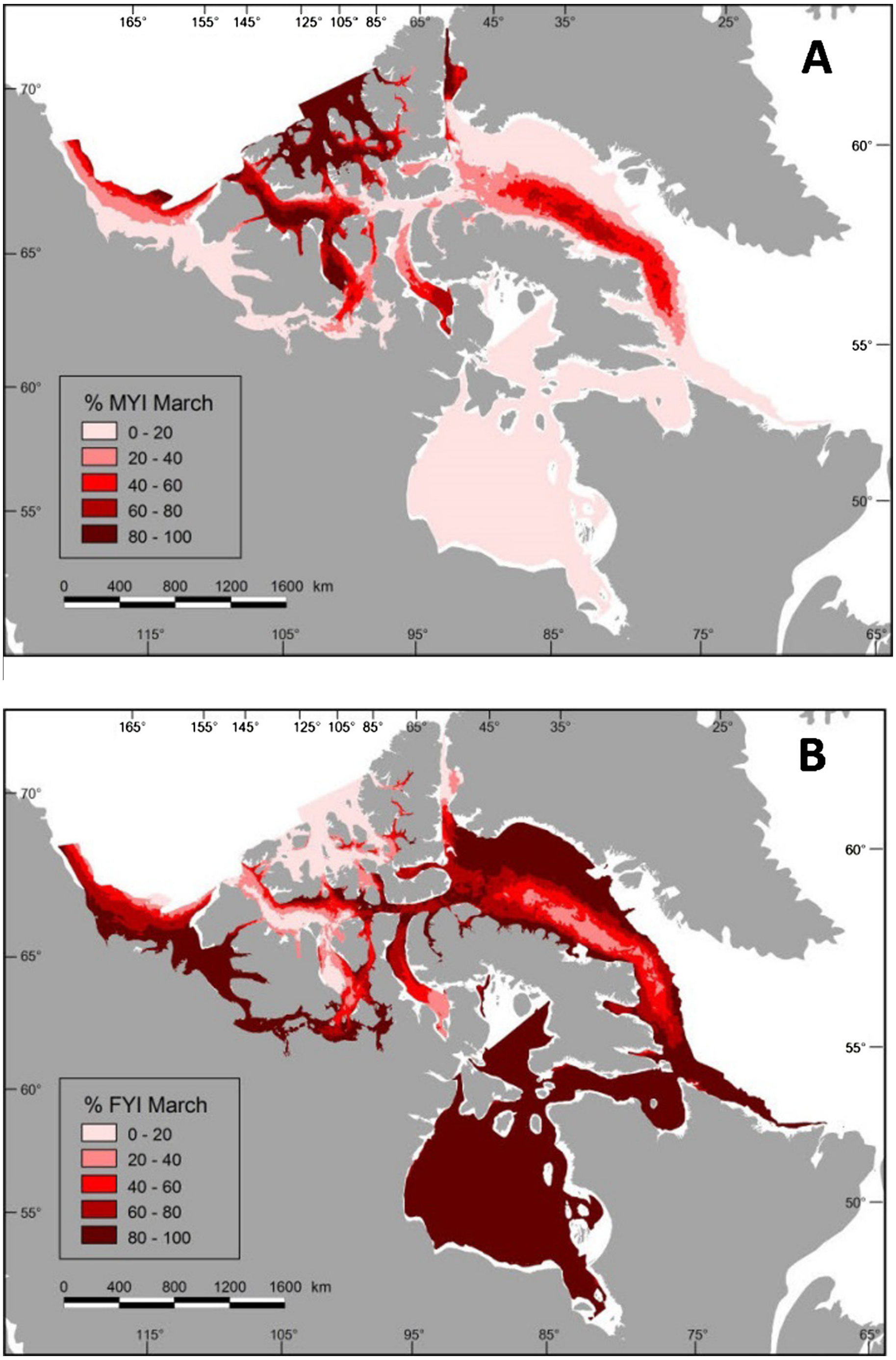

Fig. 6. Frequency of occurrence of multi-year (A), and first-year (B) ice types at the end of March in the Canadian Arctic Archipelago.

stratification, light transmission) provide the backdrop for the growth of primary producers that support pelagic and benthic grazers. In most simple terms, the classic conceptual model of the annual production cycle in a FYI environment goes from the absence of photosynthetically-based production during the dark winter to the initiation of production by ice algae with the spring increase in solar radiation. Ice algae production continues until snow/ice melt prompts their release into the water column under FYI or at marginal ice zones (e.g. Juul-Pedersen et al., 2008; Tamelander et al., 2008). The melt triggers the development of the phytoplankton bloom by: (1) lifting light limitation in the water column caused by the snow/ice matrix and the presence of ice algae, and (2) increasing stratification thereby favoring the maintenance of phytoplankton cells in the surface layer. The maximum production attained during both the ice algal and the phytoplankton blooms has been linked to nutrient inventories, in particular nitrate (Różańska et al., 2009; Tremblay et al., 2008), although silicic acid has also been identified as a limiting nutrient for ice algae production dominated by diatoms that utilize silicic acid for skeletal components (Smith et al., 1988; Lavoie et al., 2005). Nutrient depletion in surface waters is considered a key factor for the termination of the phytoplankton bloom (e.g. Sakshaug, 2004) whereas other secondary factors such as removal by grazing (Olli et al., 2007) or vertical export (Reigstad et al., 2011) can also

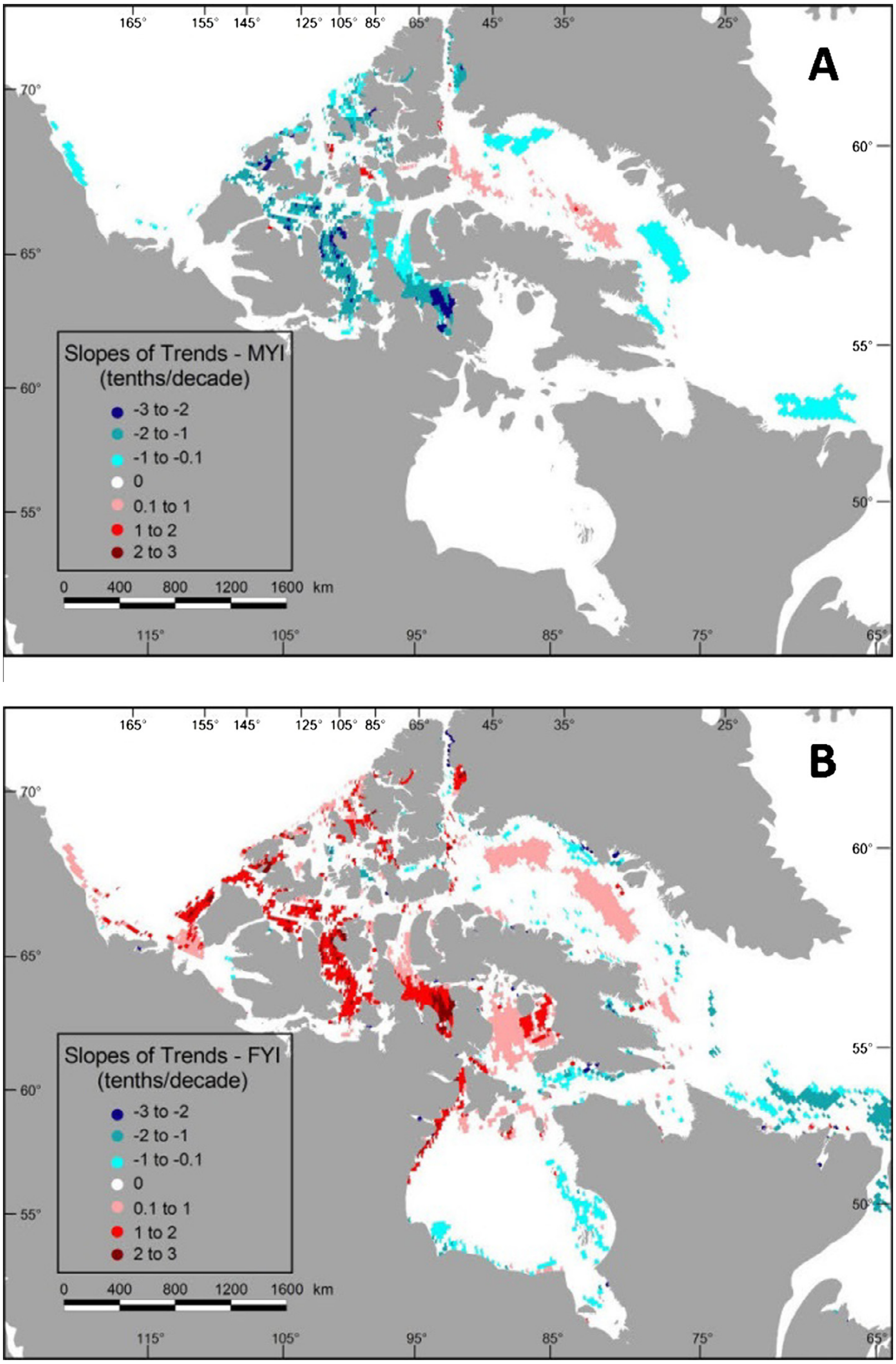

Fig. 7. Trend fit lines ($p < 0.1$) showing the slopes for multi-year (A), and first-year (B) ice types and their spatial trends over the period 1980–2012.

come into play. The duration of the ice algal growth period is, however, largely determined by atmospheric and oceanic heat fluxes acting to deteriorate or flush the ice matrix (Pogson et al., 2011). Tightly linked to this production cycle are herbivorous copepods that depend on ice algal lipids for their reproduction, synchronized for nauplii to benefit from the later phytoplankton bloom (Runge and Ingram, 1991; Søreide et al., 2010).

This seasonality applies, in broad terms, to large parts of the CAA covered by first-year ice but it is less pronounced on the EGS characterized by continuous sea ice export from the Arctic Ocean (see Fig. 5). Both Arctic outflow shelves are also characterized by wide regional variations in terms of types, magnitude and patterns of productivity, supported by their complex hydrocryospheric features, as discussed in Sections 5.2 and 5.3.

For a recent review of primary production in the Arctic Ocean and shelf areas, we refer the reader to three excellent syntheses based on complementary methods, i.e. direct measurements (Matrai et al., 2013), remote sensing estimates (Hill et al., 2013) and nutrient drawdown (Codispoti et al., 2013). Physical–biological coupled models provide additional insights for the EGS (Arctic Ocean Model Intercomparison Project, Popova et al., 2012; SINMOD, Wassmann et al., 2010; Slagstad et al., 2011).

Net community production estimated from nutrient drawdown is ca. 35 ± 15 g C m^{-2} in the CAA and ranges 5–20 g C m^{-2} y^{-1}, with

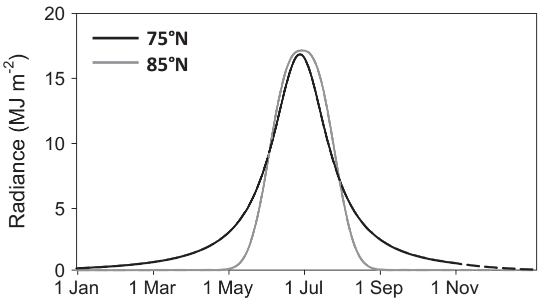

Fig. 8. Regression curves from solar radiation records in Resolute Bay (75°N) and Alert (85°N). Data obtained from Environment Canada record in 2003. Lorentzian regression with 3 parameters was used for Resolute Bay data and modified Gaussian regression with 4 parameters was used for Alert data.

15 g C m^{-2} y^{-1} considered a reasonable estimate for the EGS (Codispoti et al., 2013). Previous estimates include gross primary production at 40–60 g C m^{-2} y^{-1} (Wassmann et al., 2010) and satellite-based primary production estimates <20 g C m^{-2} y^{-1} (Pabi et al., 2008). General circulation-based model intercomparisons for the EGS estimate total primary production between 10 and 30 g C m^{-2} y^{-1}, with one model >50 g C m^{-2} y^{-1}. The wide uncertainty in estimates is largely related to their realism in reproducing observational pre-conditions. Popova et al. (2012) identify high variability in the model performances in reproducing realistic sea ice extent, upper mixed layer depths, and winter nutrient concentrations compared to World Ocean Atlas (WOA) nutrient concentrations (Garcia et al., 2006) or satellite-based estimates of sea ice and upper mixed layer depth. On both shelves, critical parameters governing production are (1) sea ice conditions regulating the incident light, (2) upper mixed layer depth determining the pool of nutrients available for production, and (3) nutrient concentrations, primarily nitrate, available at the onset of, or supplied through, the productive season. In addition, specific to the CAA are riverine input and loading, with local impacts on dissolved and particulate inventories and the spectral transmission of light. Denitrification/nitrification on the shelves and nitrogen fixation, which are unaccounted for in nitrogen budgets will also impact production estimates. On both outflow shelves, nitrification/denitrification occurs in sediments and denitrification in sea ice, the latter corresponding to ca. 27% of denitrification in Arctic sediments (Rysgaard et al., 2004, 2008).

A critical component of the annual primary production estimates is the winter load of nutrients available at the onset of the productive season. The models in Popova et al. (2012) use a winter value of 15 mmol N m^{-3} based on climatology for the EGS (Garcia et al., 2006) but there is no observational basis to support this value (Popova et al., 2012). The SINMOD model uses a value of 12.5 mmol NO_3 m^{-3} as the winter load on the Atlantic Water (AW) boundary (Slagstad et al., 2011). Several studies describe a low initial nitrate concentration of 3–4 μmol L^{-1} in the North East Water (NEW) polynya (Lara et al., 1994; Kattner and Budeus, 1997) and of ca. 4–5 μmol L^{-1} in waters outflowing from the Eurasian Basin in the EGC (Packard and Codispoti, 2007; Codispoti et al., 2013). Transects extending from the EGS outflow to the AW inflow in May 2002 (Codispoti et al., 2013), and across the EGS in April and May 2007 and 2008 (iAOOS Norway) before the onset on the productive season illustrate the strong gradient across the shelf associated with low salinity waters (Fig. 9). The nitrate surface concentrations are >11 μM in the AW inflow (Codispoti et al., 2013), decreasing to <3 μmol L^{-1} in the low salinity waters on the EGS. The very low chl *a* concentrations associated with low nitrate concentrations on the inner EGS support a winter scenario with no nitrate uptake, except at the ice edge off the shelf. Given such a strong gradient in nutrient distribution, the potential primary production in the absence of light limitation could be four times higher on the eastern compared to the western part of the EGS.

The strong east–west gradient in available nitrate at the onset of the bloom on the EGS, illustrates the uncertainty in model estimates of primary production as they are generally based on far too high nutrient pools. On the inner part of the EGS, nutrient limitation associated with the low salinity, low nitrate EGC winter waters (ca. 4 μmol L^{-1}, Codispoti et al., 2013) is also reinforced by the freshening of the surface layer by ice melt (Figs. 5 and 9). On the outer EGS, high salinity, high nitrate AW can support high production unhindered by the presence of ice and its effects on light transmission. Mixing with AW at the outer edge of the EGS will relax stratification and nutrient limitation, generating favorable conditions to support high production in this region.

In the CAA, pre-bloom surface water nitrate concentrations are considered at ca. 10 μmol L^{-1} (Codispoti et al., 2013). Underice surface nutrient concentrations measured in channels of the central/eastern Archipelago (Wellington channel, Barrow Strait, McDougall Sound) in early May can, however be highly variable, with NO_3 concentrations ranging between 2.6 and 11.5 μmol L^{-1} in 2011 and 2012 (results not shown). Multiple factors can explain this variability including biological uptake in/under the ice or in polynyas, local upwelling/mixing, and variable nutrient pools in originating water masses (i.e. Arctic, Pacific). The widespread occurrence of large centric diatoms *Coscinodiscus* spp. (Duerksen et al., 2014) at ice-covered stations in the CAA in early May suggests that pelagic production takes place in areas where favorable light conditions exist, potentially associated with early nutrient drawdown. These large phytoplankton cells may also be advected from polynyas to ice-covered areas in this system where advection dominates. *Coscinodiscus* sp. dominated the phytoplankton assemblage in the NOW early in the season (Lovejoy et al., 2002).

Aside from the large NOW, smaller polynyas are widely distributed in the CAA (Barber and Massom, 2007). Based on results from a tidal mixing model, Hannah et al. (2009) established that the location of polynyas in the CAA corresponds closely to areas of high tidal mixing, lending support to the importance of tidal currents in polynya dynamics in this region. There are no estimates of primary production for these small polynyas and their biological and ecological importance is mainly inferred from their tight association with marine mammal density distribution (discussed in Stirling, 1997). In addition to contributing to polynya formation/-maintenance, tidal mixing is also expected to sustain productivity via nutrient supply to surface waters. The tidally-mixed polynyas of the CAA therefore constitute localized areas of high productivity due to the coincident release of light and nutrient limitation. Primary production in these and other areas of strong mixing and/or episodic upwelling is not accounted for in large-scale primary production estimates for the CAA, nor in the EGS.

Annual estimates of particulate new production in the NOW range 139–152 g C m^{-2}, corresponding to ca. 60–68% of the total annual production (Klein et al., 2002; Tremblay et al., 2006). New production in the NOW is amongst the highest reported in polar marine systems and is much higher than estimates based on nutrient drawdown for the Archipelago (see above). It is also 3–4 times higher than nitrate-based new production in the NEW, estimated at 43 g C m^{-2} y^{-1} (Smith et al., 1997). The very high production in the NOW is explained by the early opening of the polynya, lifting light limitation early in the season, and a prolonged production period sustained by upwelling and nutrient supply (Tremblay et al., 2002).

Upwelling at ice edges (Mundy et al., 2009) or at the shelf-break (Williams and Carmack, 2008; Tremblay et al., 2011) can significantly increase primary production regionally, with 2–4-fold

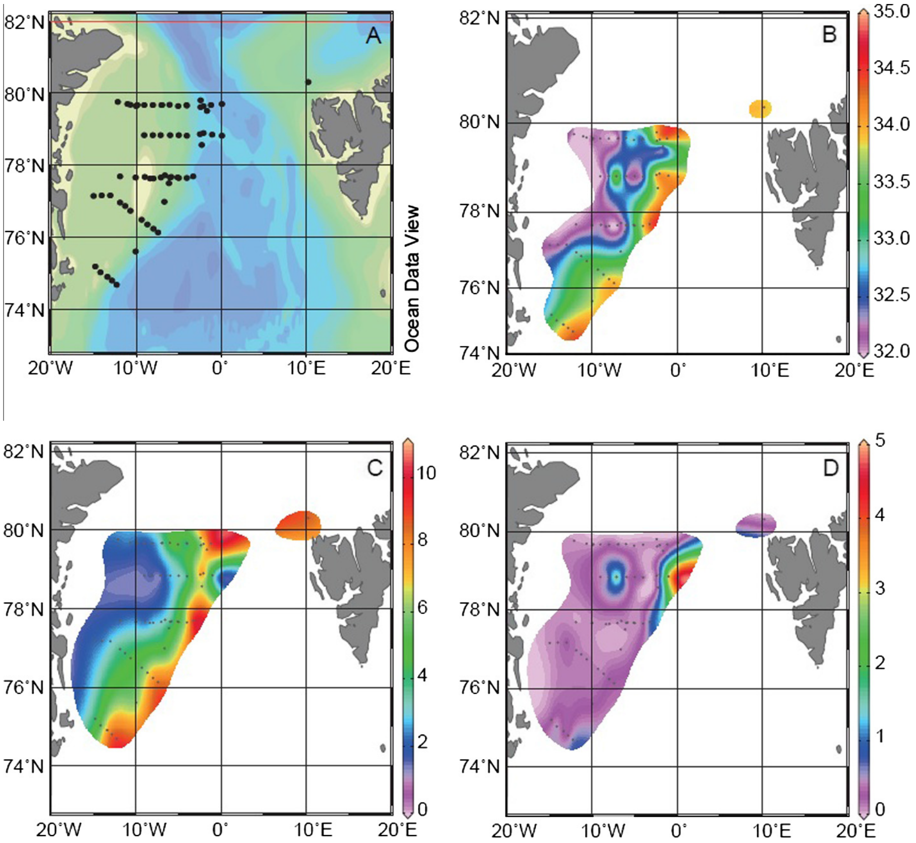

Fig. 9. Location of sampling stations (A), contours of surface salinity (B), $NO_3 + NO_2$ concentrations (μmol L^{-1}) (C), and chl *a* (mg m^{-3}) (D), on the Eastern Greenland Shelf in April/May 2007 and 2008. Figures were created with Ocean Data View; Schlitzer, R., Ocean Data View, http://odv.awi.de, 2014.

increases documented in the Canadian Beaufort Sea. Other processes also contribute to the nutrient supply into productive surface waters. Vertical turbulent diffusion was estimated to contribute up to 60% of the nitrate supply under the ice in the Amundsen Gulf, although high interannual or regional variability has been identified (Bourgault et al., 2011).

The contribution of ice algae to total annual production in coastal Arctic waters and on outflow shelves is still poorly constrained. This often leads to the assumption that ice-associated production is negligible, contributing only a minor fraction of primary production (e.g. Martin et al., 2013). Considering the wide range of temporal and spatial variations in ice algal production on Arctic shelves (Arrigo et al., 2010), and as suggested by Matrai and Apollonio (2013), it is very likely that ice-associated production is underestimated on Arctic shelves. Annual estimates of ice algal production ranging 10–20 g C m^{-2} are common in the CAA (13 g C m^{-2}, Michel et al., 1996; 15 g C m^{-2}, Smith and Herman, 1991; 23 g C m^{-2}, Smith et al., 1988), and combined ice and sub-ice production estimates can reach 29–30 g C m^{-2}, representing up to 50% of total primary production (Matrai and Apollonio, 2013). Highest ice and sub-ice algal production and biomass accumulation is observed in regions of high mixing or during periods of upwelling (e.g., Cota et al., 1987; Smith et al., 1988).

On the outflow shelves, initial nutrient inventories available for primary production depend on the annual maximum depth of the upper mixed layer and water mass distribution and variability. While regenerated production can take place in the upper water column without an additional nutrient source (nutrient recycling), new and export production (Eppley and Peterson, 1979) is limited by the annual nitrate supply. The maximum depth of the upper mixed layer on the EGS is considered 30–50 m by the WOA (Garcia et al., 2006). New production estimates for the EGS and CAA are based on variable integration depths ranging 0–50 m on the shelf (Codispoti et al., 2013) to 0–70 m in the NEW (Wallace et al., 1995), reflecting the depth range where nutrient depletion was observed or assumed. In late summer sea ice melt, combined with more local riverine input in the CAA, create a strong stratification regulating the nutrient supply to the surface waters. However, earlier in the season the distribution of the Pacific and Atlantic water is predominant. Initial nutrient inventories are expected to be strongly linked to Pacific Water inflow on the CAA outflow shelf (Michel et al., 2006; Apollonio and Matrai, 2011). On the EGS, there is large interannual variability in the contribution of nutrient-rich Pacific Water. There is, however, evidence of a recent increase in Pacific Water based on the 2011 inventory which was comparable to that observed in 1998 (Dodd et al., 2012). The Bering Strait throughflow of Pacific Water has increased by ca. 50% between 2001 and 2011 (Woodgate et al., 2012). Changes in the Pacific Water throughflow influence heat and freshwater volumes in the Arctic Ocean. Maximum oceanic heat fluxes, in 2007, likely influenced the early ice retreat in the western Arctic (Shimada et al., 2006; Woodgate et al., 2010). We expect changes in Pacific Water throughflow to impact directly and indirectly nutrient dynamics on outflow shelves. However, these impacts should be different on the two outflow shelves.

Although water masses are significantly modified during their transit, changes in Pacific Water inflow are expected to modify nutrient inventories and ratios in the CAA (McLaughlin et al., 2004), thereby impacting primary production, and as suggested from the observed large interannual variability (Michel et al., 2006; Apollonio and Matrai, 2011). In contrast, on the EGS, changes in Pacific Water inventories would have limited impact on the nutrient concentrations due to the long residence time and biological consumption upstream. In this region, varying Pacific Water inventories would have a larger impact on nutrient ratios with possible effects on the type rather than the magnitude of primary production. Finally, indirect effects of changes in Pacific Water inflow on sea ice conditions, freshwater budgets and stratification are expected on both shelves.

5.2. Food web structure in different biogeographic regions of the East Greenland Current

The hydrocryospheric and geographical features associated with the EGC create three biogeographic climate zones with (1) a region of low production under the year-round pack ice leaving the Arctic Ocean, (2) a relatively high productive belt along the MIZ to the east of the EGC, and (3) more productive spots along the Northeast Greenlandic coast where the sea ice opens up, and especially in the NEW. These three regions are by no means static as they change with wind, ice extent, water column stability and eddy formation. Nevertheless, overall, the three biogeographic regions differ profoundly in the onset and magnitude of their biological production and food web structure (e.g. Spies, 1987; Barthel, 1990; Hirche et al., 1991).

Primary production increases first along the MIZ, where the retreating ice edge leads to increased water column stability due to fresh water release by melting sea ice. Within the upper layer of this strongly stratified water column phytoplankton biomass accumulates, with peak bloom chl *a* concentrations of 1–2 mg m^{-3} according to SeaWiFS observations (Perrette et al., 2011). The strong stability of the water column in the MIZ impedes replenishment of inorganic nutrients from below the pycnocline (Smith et al., 1987). Subsequently the surface waters become depleted in nitrate (Kattner and Becker, 1991; Smith et al., 1985), leading to a deepening of the chl *a* maximum down to the nutricline at 30–50 m (Kattner and Becker, 1991; Gradinger and Baumann, 1991) by June–July. At the same time of the year, nutrient concentrations in the surface waters in the pack ice region of the EGC are still fairly high (ca. 4–6 μM NO_3, Kattner and Becker, 1991; Hirche et al., 1991) and chl *a* levels low (Gradinger and Baumann, 1991), most likely due to light limitations of phytoplankton under the dense pack ice (Smith et al., 1987), keeping the planktonic assemblage in a stage of a prolonged "winter–spring transition" even during summer (Hirche et al., 1991).

Taxonomic studies nicely demonstrate the late development of the plankton community in ice-covered waters of the EGC compared to the MIZ during summer. By June/July, the phytoplankton assemblage in the MIZ is dominated by centric diatoms, such as *Thalassiosira* spp. and *Chaetoceros* spp., dinoflagellates and *Phaeocystis pouchetii*, suggesting the succession to a summer plankton community (Spies, 1987; Smith et al., 1987; Gradinger and Baumann, 1991). In contrast, in June/July the phytoplankton assemblage under the pack ice still consists of extremely low standing stocks of nanoflagellates, small dinoflagellates and the ciliate *Mesodinium rubrum* (Gradinger and Baumann, 1991), and hence appears largely unchanged from the plankton community described from the same region in April/May (Seuthe et al., 2011).

The few measurements of primary production from the water column underneath the pack ice zone of the EGC demonstrate the low productivity of that region. In May, oxygen production measurements estimated gross primary production to range from 4.5 to 18 mmol O_2 m^{-3} d^{-1} (Seuthe et al., 2011; Vaquer-Sunyer et al., 2013), while ^{14}C-uptake primary production estimates were <50 mg C m^{-2} d^{-1} for two ice-covered stations within the EGC in June/July (Hirche et al., 1991). In contrast, integrated primary production in the MIZ averaged 426 mg C m^{-2} d^{-1} during summer after nutrients had become depleted in the surface layer (Smith et al., 1987).

The difference in primary productivity between the MIZ and the ice-covered EGC is also reflected in the secondary production of the two regions. Bolms (1986) found large numbers of copepod eggs and calanoid nauplii stage I and II in the MIZ but not under the pack ice of the EGC. Correspondingly, copepod egg production was found to be elevated in the MIZ compared to the ice-covered EGC (Hirche and Bohrer, 1987). In the EGC in June/July, *Calanus glacialis* did not spawn *in situ*, but spawning could be induced experimentally by offering the copepods improved feeding conditions (Hirche and Bohrer, 1987), indicating that *in situ* feeding conditions are inadequate for the copepods to reproduce under the ice of the EGC. This conclusion is supported by different studies, demonstrating that herbivorous copepods in the EGC have still not matured by July (Diel, 1991), have very low ingestion rates (Barthel, 1986, 1988, 1990), as well as low digestive enzyme activity (Hirche, 1989). Smith (1988) found copepods under the pack ice of the EGC to metabolize protein and not store lipids, agreeing with observation that EGC overwintering stages of *Calanus* had lower protein content (Diel, 1991) and body weights (Barthel, 1990) than overwintering stages of *Calanus* from the more productive Atlantic-influenced region of Fram Strait. Based on these findings, Barthel (1986, 1990) concluded that at least the larger herbivorous copepods cannot sustain themselves in the pack ice region of the EGC at food concentrations of 50 μg C l^{-1}. But why then do we find large herbivorous copepods in the EGC? The answer is most likely multifaceted, with advection of organisms being an important part of it.

The major source for zooplankton within the EGC is the upstream Arctic Ocean. Yet, boreal fauna is also transported onto the EGS along with Atlantic Intermediate Water entering onto the shelf through troughs and canyons (Bourke et al., 1987; Hirche et al., 1991; Hirche and Kwasniewski, 1997). This on-shelf transport of organisms explains the presence of the Atlantic *Calanus finmarchicus* on the shelf (Ashjian et al., 1995), but also in part the populations of *Calanus hyperboreus* and *Metridia longa* (Hirche et al., 1991). The contribution of Atlantic copepod species diminishes to the west (Diel, 1991) and north (Ashjian et al., 1995), as does the overall zooplankton biomass. In fact, total zooplankton biomass on the EGS appears to be only about 20% of that found off the shelf break to the east (Hirche et al., 1991). This indicates that advection processes are not strong enough to maintain large zooplankton populations on the EGS (Ashjian et al., 1997), as well as that autochthonous secondary production within the EGC is low, most likely due to food limitation.

How strongly primary and secondary production in the ice-covered waters of the EGC are light and food limited, respectively, is best explained by looking at the planktonic community developing in areas of the EGC largely free from sea ice, such as the NEW. The NEW is highly heterogeneous and hence biological development is heavily steered by local topography and currents (e.g., Budeus et al., 1997; Schneider and Budeus, 1997). For a detailed description of the polynya see Berreville et al. (2008). In general, the NEW is an area of elevated primary production within the EGC (Hirche et al., 1991; Smith, 1995; Wallace et al., 1995). During the 3–4 months that the NEW is open, the onset and development of phytoplankton may differ slightly between various regions of the polynya, but phytoplankton communities dominated by larger cells (Barthel, 1990), such as diatoms, appear to prevail throughout the productive season (Pesant et al., 1996; Lara et al., 1994; Gradinger, 1986; Gradinger and Baumann, 1991; Booth and Smith, 1997; von Quillfeldt, 1997). This creates a good feeding ground for calanoid copepods, which show highly increased digestive enzyme activities in the polynya compared to the surrounding ice-covered areas (Hirche, 1989; Diel, 1991). The favorable feeding conditions in the NEW are also reflected in high copepod egg production rates (Hirche and Bohrer, 1987; Hirche et al., 1994; Hirche and Kwasniewski, 1997; Ashjian et al., 1995), and high abundance of copepod nauplii in the NEW (Bolms, 1986). Diel (1991) observed a significant shift from dominance of copepodite stage CI and CII towards older developmental stages in *C. glacialis* and *C. hyperboreus* in the polynya from mid-July to early August, indicating that the NEW supports the recruitment of a new copepod overwintering population.

Despite the moderate-to-high ingestion rates of herbivorous copepods (Barthel, 1988) and protozooplankton (Pesant et al., 1998), phytoplankton biomass accumulation in the NEW does not appear to be controlled by grazing, owing to the low abundance of grazers (Daly, 1997; Hirche et al., 1991, 1994; Hirche and Kwasniewski, 1997; Ashjian et al., 1995). This results in many occurrences (in space and time) of chl *a* concentrations >1 mg m^{-3} within the polynya, with sometimes a relatively homogenous vertical distribution of chl *a* from the surface to the seafloor (Gradinger and Baumann, 1991). Consequently, sedimentation rates can be moderately high during summer (116–327 mg C m^{-2} d^{-1}; Legendre et al., 1994). It appears that 20–60% of the pelagic primary production is either vertically exported to the seafloor or laterally advected out of the polynya (Pesant et al., 1996; Pesant et al., 2002).

The large fraction of ungrazed phytoplankton biomass reaching the seafloor fuels the benthic community (Hobson et al., 1995; Rowe et al., 1997), such that pelagic and benthic pigment concentrations are the most important predictors for benthic density in the NEW (Ambrose and Renaud, 1995; Piepenburg et al., 1997; Brandt, 1995, 1997). Benthic density and composition varies between regions in the polynya (Piepenburg, 1988; Piepenburg et al., 1997), but are generally moderately high. On the shallow banks, epibenthic organisms such as brittle stars can be highly abundant (340 ind. m^{-2}, 1.8–10.5 g AFDW m^{-2}; Piepenburg and Schmid, 1996), while suspension feeders predominate in the troughs.

The tight pelagic–benthic coupling and resulting moderate benthic biomass appear to sustain small populations of benthic feeders, such as eiders and walruses (Weslawski et al., 1997). However, a larger survey on seabird communities utilizing the NEW found that surface-feeding birds, such as Fulmars and gulls, dominate the polynya (Falk et al., 1997). These birds depend on small fish and zooplankton rather than on benthic production. The presence of breeding seabirds in the NEW illustrates that the polynya supports bird recruitment (Falk and Moller, 1997) in an otherwise deserted high Arctic region. However, the small population size of seabirds (<14000 birds) is evidence of the relatively low carrying capacity of the NEW (Falk et al., 1997).

Using seabirds and marine mammals as indicators of marine production, their summer distribution provides further evidence for the different productivity of the three biogeographic climate zones within the EGC. Harp and hooded seals, as well as different seabird species, accumulate at the eastern MIZ, where their food intake is substantially higher than in the low productivity pack-ice zone of the EGC (Joiris, 1992; Joiris and Falck, 2011). The pack ice, however, is used by marine mammals for rearing (e.g. harp seals, Haug et al., 2006; hooded seals, Salberg et al., 2008), and hence plays a crucial role in the life history of many animals, with potentially devastating consequences if the pack ice were to disappear (Kovacs and Michel, 2011).

The emerging picture of the marine ecosystem within the EGC is that of a highly heterogeneous system, where different biogeographic climate zones exist at the same latitude. These climate zones primarily result from the year-round advection of sea ice from the Arctic Ocean, which substantially alters the underwater light climate, and hence production regime, between the different regions within the EGC. By June/July, a pronounced east-west gradient in plankton development stage exists across the EGC, with (1) a microbial summer plankton community along the MIZ and eastern EGC front, with production mainly from regenerated nitrogen sources (*f*-ratio averaged 0.31, Keene et al., 1991); (2) a prebloom plankton community under the heavy pack ice of the central EGC; and (3) a vernal bloom plankton community in the NEW, with production largely based on nitrate (*f*-ratio averaged 0.65, Smith et al., 1997).

5.3. Regional productivity regimes and responses to change in the Canadian Arctic Archipelago

Recent multidisciplinary research efforts in the CAA (e.g. ArcticNet, C3O, CFL) combined with modeling approaches, have provided new insights into primary production patterns, their drivers and responses to change. Here we provide an overview of some key findings and discuss, in more detail, the decadal timeseries in Barrow Strait.

Ship-transects in the CAA during three consecutive years (2005, 2006, 2007) define a shelf dominated by high spatial variability, with two main productivity regimes characterized by distinct phytoplankton community structure, biomass and production (Ardyna et al., 2011). These authors identify an oligotrophic flagellate-based system extending mainly over the Beaufort Sea and the central region of the Archipelago and a eutrophic diatom-based system extending over Baffin Bay, Lancaster Sound and the central Gulf of Amundsen. The high diatom abundances and biogenic silica inventories observed in Lancaster Sound in summer 2007 and 2008 (Wyatt et al., 2013) provide additional evidence of the high productivity of this region. The main factors explaining the different productivity regimes are stratification and nutrient concentrations at the base of the euphotic zone and, to a lesser extent, incident irradiance. Of importance for predictions of future productivity in the changing Arctic, the two productivity regimes were found to respond differently to the decrease in sea ice cover (Ardyna et al., 2011). A large increase in primary production was observed in eutrophic regions whereas little change was found in oligotrophic regions, supporting previous modeling evidence that extending the open water period would not increase primary production in oligotrophic regions where nutrient limitation prevail (Lavoie et al., 2010).

Recent evidence of the widespread occurrence of subsurface chl *a* maxima in the Canadian Archipelago and other Arctic regions (Martin et al., 2010; Hill et al., 2013) also provides new insights into the seasonal dynamics of primary production in this region. Following from the general seasonal cycle described earlier, the emergent view is that phytoplankton initially develops in the surface layer, followed by a subsurface chl *a* maxima developing later on, associated with the nutricline. As pointed out earlier, there can be large regional variations in the mixed layer nutrient inventory at the onset of the phytoplankton growth season. There are also large regional variations in mixed layer depth and the strength of the pycnocline. To exemplify extremes in conditions, on the one hand, melt and freshwater input in the Beaufort Sea can create a strongly stratified surface layer isolated from nutrient-rich deeper waters by a robust pycnocline (e.g. Simpson et al., 2008). During the overwintering CASES study in 2003–2004, Tremblay et al. (2008) documented very low NO_3 concentrations (<1 μmol L^{-1}) in the surface layer at the time of ice breakup, followed by the development of a subsurface chl *a* maximum deepening with the nutricline until August. On the other hand, high spring nutrient inventories (up to 10–12 μmol L^{-1} NO_3, Matrai and Apollonio, 2013) and strong mixing are documented for the eastern Archipelago. The annual phytoplankton cycle in these regions is described in more details below.

The annual time-series of fluorescence and zooplankton biomass derived from backscatter for the south side of Barrow Strait clearly show a tight association between the development of the phytoplankton bloom (maximum fluorescence signal, Fig. 10a) and the timing of ice break-up. There is also clear evidence that the development of the spring/summer phytoplankton bloom is linked to the shallowing of the surface mixed layer (Fig. 11), as in other seasonally ice-covered Arctic regions including the NOW (Sakshaug, 2004; Tremblay et al., 2006) and the MIZ of the EGC (see Section 5.2). These results support the classic paradigm of the summer phytoplankton bloom developing in association with the ice break-up, as observed in earlier decadal time-series (1983–1993) in Barrow Strait (Michel et al., 2006). However, in contrast to the earlier time-series that reliably showed a unimodal phytoplankton bloom reaching maximum biomasses at the end of July–early August (Michel et al., 2006), the annual cycles for 2004 and 2008 reveal a 7-week difference in the timing of ice break-up between mooring years, and a similar difference in the ensuing spring–summer fluorescence increase (Fig. 10). These recent annual time-series show a strong subsurface fluorescence maximum developing as early as the beginning of June (2004) and in mid-July (2008) during years of contrasting ice cover break-up (Fig. 11). High interannual variability in the timing of ice melt and the associated release of ice algae, with a range up to ca. 4 weeks over a decade of measurements, has been documented under landfast ice in this region (Fortier et al., 2002; Michel et al., 2006). However, the extreme variability in the timing of the phytoplankton bloom reported here is not, to our knowledge, documented in historical records. This may indicate that earlier phytoplankton growth in response to changes in sea ice conditions in the Canadian Archipelago has occurred over the past decade. Yet, this may also reflect shortcomings in the observational database as there are limited systematic annual data series needed to resolve temporal variability at interannual and decadal scales.

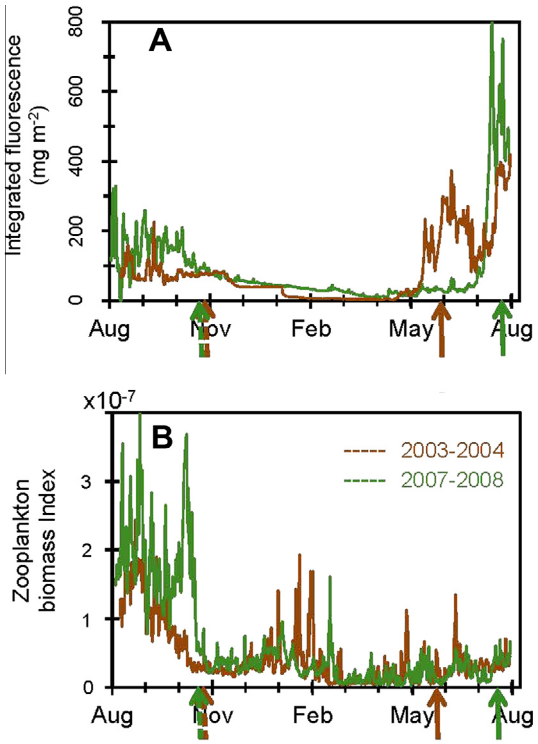

Fig. 10. Yearlong records of depth-integrated relative fluorescence from daily Icycler profiles over a depth interval of 3–46 m (A), and zooplankton biomass index computed from ADCP backscatter over its profiling range of 10–70 m (B), on the South side of Barrow Strait in 2003–2004 and 2007–2008. The two records are indicators of the relative abundance of phytoplankton and zooplankton in the upper water column. The timing of freeze-up (dashed arrows) and break-up (solid arrows) is indicated for both years.

Observations from ship-based expeditions on outflow shelves are biased towards the end of summer-fall due to constraints imposed by the presence of ice, in particular multi-year ice. Consequently the early spring–summer bloom (including ice-associated and under ice) is not often captured, being rather inferred from derived variables (e.g. Tremblay et al., 2008; Matrai and Apollonio, 2013). Ardyna et al. (2011) observed widespread fall phytoplankton blooms in the CAA and, more recently, Ardyna et al. (2014) report an increase in occurrence of fall phytoplankton blooms overall the Arctic. Fall blooms are documented in polynyas, linked to a prolonged open water period and nutrient input to surface waters during fall mixing events (Caron et al., 2004; Forest et al., 2008). For the central Archipelago, the fall bloom can be related to long open water periods dictated by late freeze-up rather than early break-up (Ardyna et al., 2011). Informative results from the annual time-series help elucidate aspects of the dynamics of the so-called fall phytoplankton bloom (see Fig. 10). Firstly, both time-series during contrasting years show the development of a bloom towards the end of summer, in August. However, this bloom persists into the fall only during the year of late ice break-up (2008, Fig. 10). During that year, the summer-fall bloom is the unique annual phytoplankton bloom event, whereas during the year of early ice break-up (2004), the August bloom is the second seasonal bloom event. It follows that the development of the “fall” bloom does not relate to the timing of either ice formation or break-up. Rather, the occurrence of the “fall” bloom as a second seasonal bloom is contingent upon a sufficiently long open water period, in this case determined by early ice break-up. Widespread observations of a fall bloom may therefore reflect a unique or a second seasonal bloom, depending on regional, or internannual, differences in the length of the open water period. Interestingly, our results also show much stronger fluorescence signals for the unique fall bloom (2008, Figs. 10 and 11), possibly reflecting the phytoplankton inability to use nutrients earlier in the season during years of heavy and/or enduring ice coverage, or interannual variability in nutrient inventories. Unfortunately, these hypotheses cannot be resolved with the current observational database. Our annual time series substantiate nutrient replenishment in surface waters fueling a second bloom late in the season when ice conditions allow (August 2003–2004, Fig. 10). A corollary is that primary production based on nutrient drawdown is underestimated in the CAA due to the key role of tidal mixing and upwelling in this region, as previously suggested (Codispoti et al., 2013).

During both observation years the fluorescence signal started to increase 1–2 weeks prior to the ice break-up (Fig. 10a). This may reflect underice phytoplankton growth or the release of ice algae during melt, both of which have been observed in the CAA in June/July (Michel et al., 1996; Fortier et al., 2002). Underice phytoplankton blooms are deemed to occur under favorable light conditions linked to increased light transmission through the ice matrix after snow melt or in the presence of melt pounds (Frey et al., 2011; Perovich and Polashenski, 2012). These blooms can account for a large proportion of Arctic primary production unaccounted for by remote sensing estimates (Arrigo et al., 2012). In the CAA, the development of underice phytoplankton blooms has been linked to atmospheric forcings (rain, warm temperatures) causing early melt (Fortier et al., 2002). The signal of ice algae released into the pelagic environment during melt can be notoriously difficult to distinguish from that of underice phytoplankton in the Canadian Arctic. Stable isotope signatures can help discriminate ice-associated and pelagic producers (Pineault et al., 2013) but species overlap between the ice and underice habitats and the fact that ice

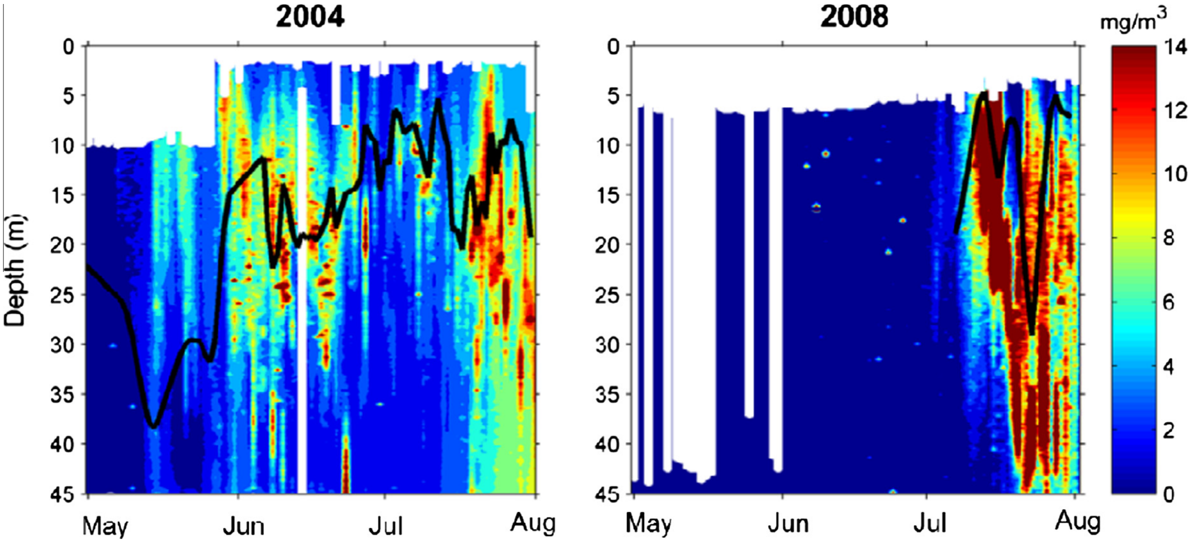

Fig. 11. Daily fluorescence profiles collected with the moored profiler, Icycler, from May to August 2004 and 2008. The black line in each panel represents the depth of the maximum density gradient over the 45 m interval, smoothed with a 3-day running mean. Interpolated value were used when the maximum gradient for an individual profile was <0.1 kg m^{-4}.

algae can be photosynthetically competent after their release from the ice in spring (Michel et al., 1993; Yamamoto et al., 2014) complicates matters further. While underice phytoplankton blooms are expected to be more widespread than previous thought (Mundy et al., 2009; Arrigo et al., 2012), favorable light conditions, associated with advanced melt and/or ice ablation causing the disappearance of the optically dense ice algal and snow layers, are prerequisite for their development in landfast ice covered areas.

Areas of multi-year ice and the Sverdrup Basin in general, are poorly studied and may, to some extent, follow a similar seasonal pattern as described above. Our recent results show that ice algal biomass is 2–3 orders of magnitude lower in multi-year ice at >80°N than in bottom first year ice at 75°N in the CAA. A similar pattern is observed in first year ice cores, with 2 orders of magnitude differences in spring ice algal biomass between latitudes >80°N and 75°N. This indicates that the seasonality in light (see Fig. 8) may regulate primary production at higher latitudes on the outflow shelves. Therefore, the shift from multi-year ice to first year ice in the northern channels of the CAA (see Fig. 7) may not lead to increases in ice-associated production commensurate to production measurements in the more southern channels. Additionally, the very short phytoplankton growth season, defined as the overlapping time window between sufficient seasonal PAR availability and partial (MYI) or complete (FYI) melt, likely limits potential phytoplankton production in the northern CAA.

The annual time-series reveal that there is no tight coupling between the development of the zooplankton community and the timing of the phytoplankton bloom (Fig. 10). This may be due to the integration of different zooplankton species with different life cycles, in the present estimate. Yet, our results agree with the overall seasonal development of populations of calanoid copepods, expected to dominate the biomass signal in this region (Conover and Siferd, 1993; Fortier et al., 2002). The four dominant large calanoid species in this region are *Pseudocalanus acuspes*, *C. glacialis*, *M. longa* and *C. hyperboreus*. Except for the latter species which starts to reproduce during winter, other species start to reproduce in spring. A mixture of developmental stages is found later in the season with Calanus stages III and IV overwintering to reach maturity the next year. The winter biomass signal probably reflects the presence of these overwintering copepod populations. The lack of coupling between zooplankton biomass indicators and phytoplankton abundance is not surprising in the context of the key importance of ice algae in the copepods life cycle in the Arctic. Spring reproduction has often been linked to the accessibility of ice algae as a primary food source, with abundant evidence of grazing on ice algae (Michel et al., 1996, 2002; Søreide et al., 2006; Tamelander et al., 2008). Recent studies showing that ice algae provide essential fatty acids for zooplankton consumers (Søreide et al., 2006, 2008) add to the collective evidence that ice algae play key ecological role in Arctic marine food webs (Leu et al., 2011; Kovacs and Michel, 2011).

The northern side of Barrow Strait where mean currents are very weak and westward, experienced a positive trend in summer water temperature, increasing lower water column salinity, and a trend towards earlier freeze-up (see Section 3.2.2). Unfortunately, there are no fluorescence records for this region but the development of the resident zooplankton population was shown to be tightly linked to early summer water temperatures (Hamilton et al., 2013). Significant changes in the phenology of initial population growth have been observed over the last decade, advancing from mid-July in the late 1990s to mid-May in 2006 and estimated at ca. 8 days per decade (Fig. 12). Comparable advances in the phenology of marine species are reported throughout the World Ocean (Poloczanska et al., 2013). These changes indicate that phenotypic plasticity would allow large changes in phenology in response to

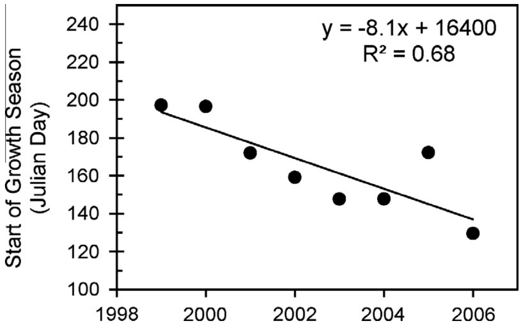

Fig. 12. Trends in the timing of the seasonal increase in zooplankton biomass, defined as a five fold increase in backscatter relative to winter values, between 1999 and 2006 on the north side of Barrow Strait.

environmental variability, making it possible for zooplankton species to respond and adapt to on-going and, possibly future, change.

6. Conceptual model and future perspectives

Notwithstanding the distinct hydrocryospheric and geographical features of the two Arctic outflow shelves, commonalities can be drawn, helping identify general productivity patterns and key forcings. We define four simple conceptual models of productivity regimes representing conditions observed on both outflow shelves (Fig. 13). For simplicity, the main focus here is on the seasonal dynamics of the open water phytoplankton bloom.

The first model applies to regions of high initial nutrient inventories, where the development of highly productive spring/summer phytoplankton bloom is associated with the seasonal shallowing of the pycnocline. This type of regime, defined as high nutrient–high stratification (HN–HS), was observed in Barrow Strait of the CAA during this study (2003–2004 time series, Fig. 10) and also applies to the MIZ of the EGS. In this regime, production would be strongly linked to initial nutrient inventories due to the absence of nutrient resupply and/or slow remineralization during the growth season.

The second type of regime differs from the previous one in that the initial nutrient inventories are lower, thereby limiting maximum primary production attained during the growth period. This regime, defined as low nutrient–high stratification (LN–HS) is typical of the EGS and the Beaufort Sea in the absence of upwelling.

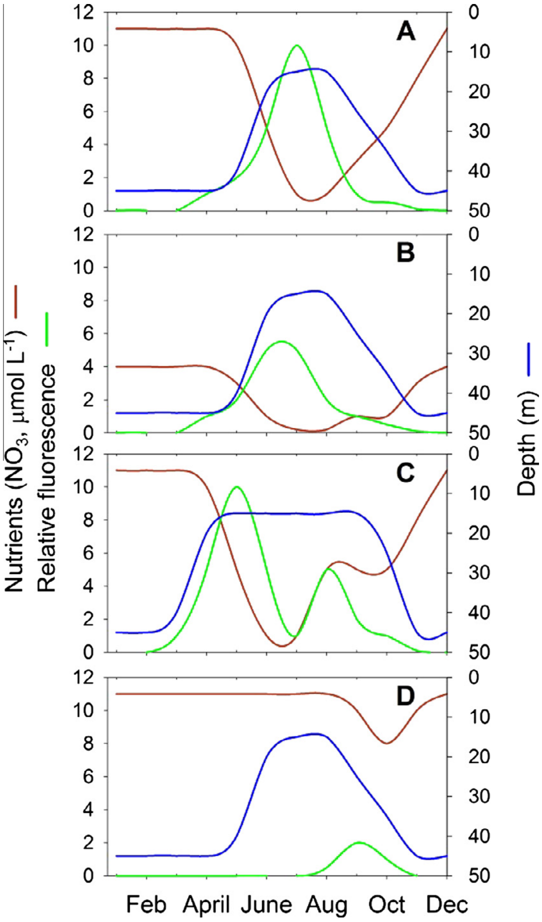

Fig. 13. Conceptual model of four productivity regimes observed on Arctic outflow shelves. High nutrients and high stratification (A), low nutrients and high stratification (B), high nutrients and high mixing (C), and variable nutrients and low light (D).

The third regime characterizes regions of highest production, with high initial nutrient inventories combined with upwelling and mixing events sustaining production after the initial drawdown and as long as light conditions allow. The sketch presented here reflects the development of a second phytoplankton bloom, as observed in the 2007–2008 Barrow Strait time series and in the NOW (see previous section). However, short episodic nutrient supply sustaining primary production over longer periods, rather than two separate blooms, is another manifestation of this high production regime defined by high nutrients and high mixing (HN–HM). Ice edge and wind-driven upwelling sustaining high production as observed in the Beaufort Sea (Mundy et al., 2009; Tremblay et al., 2011) would fall under this regime.

The fourth production regime applies to regions of extensive ice cover and/or extreme high latitudes, where light limitation precludes high annual production regardless of nutrient inventories. In our example, we present a high initial nutrient inventory to emphasize the importance of the short growth period in this regime, as it constrains the seasonal depletion of nutrients in the upper mixed layer. This regime is defined as variable nutrient–low light (VN–LL).

The development of phytoplankton blooms under the ice is not represented schematically in the conceptual models presented here as these blooms are expected to fall within the regimes proposed, with the key difference of advancing the growth season and therefore shifting seasonality. Underice phytoplankton blooms would not substantially change overall production under the HN–HS and LS–HS regimes unless increased nutrient recycling during a protracted growth season would increase later regenerated production. There is evidence of a seasonal shift from new to regenerated N-based production under current HN–HS regimes in the Beaufort Sea (Tremblay et al., 2008). Here, the protracted growth season could magnify regenerated production and modify export ratios. Under the HN–HM regime, underice phytoplankton blooms would contribute to increase total primary production. In these regions, the declining sea ice cover is expected to cause significant increases in production due to the absence of nutrient limitation, as suggested by Ardyna et al. (2011). Potentially large increases in primary production are anticipated in areas where a protracted open water period combined with increasing storm frequency and intensity can favor more frequent episodic wind-driven upwelling and mixing, as observed in the Canadian Beaufort Sea during the 2007 minimum ice year (Tremblay et al., 2011). In this context, regions currently characterized as HN–HS and LN–HS could shift to HN–HM. In contrast, under the VN–LL regime, potential increases in primary production would be associated with a shift from light to nutrient limitation.

On both outflow shelves, light is expected to remain an important factor regulating primary production in areas that are heavily ice covered during most of the year, especially in the northern regions where light availability is highly constrained seasonally. Yet, the current trends in decreasing ice thickness, smaller ice floes, and increased presence of melt ponds allowing more light to reach the primary producers suggest that light limitation may be reduced in the foreseeable future. Under such scenario, concomitant potential increases in primary production and an earlier and more northward consumption of nutrients along the advective flow down the EGS and in northern channels of the Archipelago are plausible.

In this context, the changing ice cover is likely to modify productivity regimes on outflow shelves, through aggregated effects on light transmission and euphotic zone depth, as well as on surface stratification and mixing processes, both intimately linked to nutrient dynamics in the productive surface layer. Regions presently or formerly considered oligotrophic could become eutrophic due to increased mixing and nutrient input into surface waters.

Conversely, eutrophic regions would become even more productive with longer productive seasons, given than nutrients are not limited.

7. Conclusions

This study highlights the heterogeneous nature of the Arctic outflow shelves, both in terms of physical forcing and biological processes. The two Arctic outflow shelves experience a broad radiative context, diverse ice conditions ranging from 0% to 100% ice cover and from FYI to MYI, wide gradients in mixed layer depth, advection patterns, and nutrient inventories. Our results show that this large heterogeneity operates at various spatial scales, from local scales (less than a few kilometers, CAA) to more regional scales (10s of kilometers, EGS). Latitudinally-influenced seasonality affects irradiance and ice conditions in particular, but it also impacts biogeochemical processes and signatures of shelf waters as biologically-spent Arctic waters are advected southward on the EGS and the CAA.

Linked to their function in the Arctic Ocean system, conditions on outflow shelves are influenced by basin-wide changes in atmospheric, oceanographic and sea ice patterns which determine upstream conditions. These provide initial conditions to the outflow shelves, but cascading impacts and additional changes taking place within the shelves ultimately determine future trends in productivity patterns and ecosystem transfers. In this respect, the two outflow shelves are expected to respond differently to on-going and future Arctic changes. For example, sea ice loss in the Arctic basin may trigger significant increases in production in the CAA (see Section 5.3) but may not elicit readily observable change in productivity in the EGS, due to continuous ice export. In the latter, changes in ice characteristics may have a stronger influence on productivity patterns than changes in ice extent and duration. This example underscores fundamental differences between the two ouflow shelves, with advection being a dominant forcing on the EGS whereas variable nutrient inventories and mixing processes are especially important in the CAA.

For the CAA, we propose that the collective evidence of: (1) phytoplankton blooms occurring earlier than in the past, (2) favorable conditions for upwelling/mixing under the ice and in open waters becoming more frequent, and (3) earlier melt and overall thinning of the ice, dictates an upward re-evaluation of primary production estimates, positioning the CAA as one of the highly productive Arctic shelves. For the EGS, nutrient limitation may become more prominent as incident light increases with thinning ice. In addition, the ecosystem response to variable input and transports of Pacific Waters is unknown as is the future location and magnitude of the productive MIZ.

On the outflow shelves, the structure of the pelagic food web is not only linked to primary production but also to the time available for building up a trophic succession and for species to reproduce. Advection is therefore determinant. In addition, phenological adaptations of zooplankton species are evidenced, implying phenotypic plasticity and the capacity to adapt to environmental changes.

One of the current challenges faced by the Arctic science community is to better resolve the heterogeneity within outflow shelf systems, for consideration in estimates of Arctic primary production. The conceptual models attempt to capture some of this heterogeneity and variability, providing a useful basis upon which to develop predictive scenarios of future productivity states. Key questions arise though, which are outside the scope of this paper and the simple models presented here. For example, the respective contributions of ice algae and phytoplankton under the ice and at depth, still poorly constrained, are likely to change in the future. Since these different contributors to total production are distinct in space and/or time, their pelagic–benthic couplings within food webs are likely to differ, suggesting different pathways of transfer to higher trophic levels and potentially, harvestable resources. The latter being commensurate to new production, it is also important to elucidate the role of new *versus* regenerated production in the changing Arctic.

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