The ‘interior’ shelves of the Arctic Ocean: Physical oceanographic setting, climatology and effects of sea-ice retreat on cross-shelf exchange

William J. Williams *, Eddy C. Carmack

Fisheries and Oceans Canada, Institute of Ocean Sciences, 9860 West Saanich Road, Sidney, British Columbia V8V 4L1, Canada

Abstract

The interior shelves of the Arctic Mediterranean are the shelves of the Kara Sea, Laptev Sea, East Siberian Sea and Beaufort Sea. They comprise approximately 40% of the total arctic shelf area (~2.5 × 10^6 km^2) and are distinguished from inflow and outflow shelves by their principal forcing dynamics. Along their southern (continental) boundary the interior shelves are dominated by the major arctic rivers, receiving over 80% of the total freshwater input to the Arctic Ocean. In the mid-shelf region wind and ice motion surface stresses dominate mixing and circulation, resulting in high variability. Along, their northern (seaward) boundary they are forced by upwelling- and downwelling-favourable surface stresses which drive shelf-basin exchanges with Atlantic- and Pacific-origin cyclonic boundary currents over the upper slope. Shelf-basin exchange is further modified by shelf-break morphometry (e.g. canyons, valleys, headlands and bottom slope). Here we review the physical oceanographic settings and forcing of the interior shelves and then focus on shelfbreak exchange and supply of nutrients for new primary production due to upwelling across the shelfbreak. As a proxy for this nutrient supply, we show seasonal and annual time series of along-shelfbreak surface-stress due to wind and ice motion from 1979 to 2011. We apply this analysis to the shallow shelves from the Kara Sea to the Beaufort Sea and comment on recent increases due to atmospheric changes and sea-ice retreat.

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1. Introduction

The supply and disposition of water masses is of particular concern in the Arctic Ocean as it is a mediterranean ocean with generally low tidal mixing coupled with inputs of (1) Atlantic-origin water through Fram Strait and the Barents Sea (~6 Sv), (2) Pacific-origin water through Bering Strait to the Chukchi Sea shelf (~1 Sv), and (3) large freshwater inputs from the surrounding continents of Eurasia and North America (~0.15 Sv) (cf. Carmack, 2000; Serreze et al., 2006). The large and continuous inflows of Atlantic- and Pacific-origin water dominate the Barents Sea and Chukchi Sea shelves respectively and these shelves have been termed ‘inflow shelves’ (Carmack and Wassmann, 2006). Upon entering the Arctic basins, these water masses subduct below the surface mixed layer to below shelf-break depth and move cyclonically around the basin perimeter and ridges as topographically-steered boundary currents (Rudels et al., 1994, 1999; McLaughlin et al., 2002; Aksenov et al., 2011). The inflows of Pacific, Atlantic and riverine water, substantially modified by physical and biogeochemical processes within the Arctic Ocean, must eventually leave the Arctic Ocean, and they do so through Fram Strait and the complex straits and channels of the Canadian Arctic Archipelago (Carmack, 2000; Melling et al., 2008; Holfort et al., 2008; Beszczynska-Möller et al., 2011). These regions have thus been termed ‘outflow shelves’.

Located between the ‘inflow shelves’ and the ‘outflow shelves’ of the Arctic Ocean are the so-called ‘interior shelves’, bounded by the arctic basins to the north and by continental land masses to the south. On the Eurasian side of the Arctic, these shelves are the Kara Sea, Laptev Sea and East Siberian Sea and, on the North American side of the Arctic, the Beaufort Sea which contains the Alaskan Beaufort Shelf and the Canadian Beaufort Shelf (Figs. 1 and 2). The large arctic rivers flow out onto these shelves so they are the source regions for the fresh water that dominates the stratification at the surface of the Arctic Ocean (Aagaard and Carmack, 1989; Proshutinsky et al., 2009).

The inflows of Atlantic and Pacific-origin water bring the vast majority of the dissolved nutrients to the Arctic ocean, as nitrate concentrations are ~12–17 mmol/m³ in the Pacific and Atlantic-origin water (peaking in the Pacific origin water, where it is regionally present; McLaughlin et al., 2011). In comparison, the much lower concentrations in incoming rivers (~3 mmol/m³ in the inflowing rivers during the spring and summer freshet) have a much smaller influence on new primary production (Holmes et al., 2012). But, because the Atlantic- and Pacific-origin waters
are salty relative to surface waters diluted by river water and ice melt, they spill-off their inflow shelves and descend to below the shelfbreak depth in the arctic basins (Timmermans et al., 2014; Aksenov et al., 2011). Thus the primary production of the interior shelves, and also the outflow shelves, depends, to a major degree, on upwelling or mixing of these nutrient-rich waters onto the shelves and up into the freshwater-influenced euphotic zone.

Forcing conditions for the interior shelves are changing rapidly. The well-known decline of Arctic sea ice over the last 30 years (Maslanik et al., 2011) has led to a progressive un-covering of the Arctic Shelves in summer. Multi-year sea ice (ice that survives summertime melt) formerly extended across the Arctic from the region of the oldest ice against the Canadian Arctic Archipelago to over the Siberian Shelves, particularly the Laptev Sea, East Siberian Sea, and the Chukchi Sea (Maslanik et al., 2011). The Arctic sea ice now regularly retreats in summer to the deep Arctic Ocean, beyond the shelfbreak, so that large areas of the Arctic Ocean now only have seasonal ice and so are ice-free for a period in the summer (cf. Bluhm et al. 2015). As an example, Fig. 1 compares the recent record low September sea-ice minima of 2007 and 2012 with the long-term mean minimum for the period 1979–2000. This annual summertime retreat of the ice to beyond the shelfbreak is displayed in Fig. 2 as a time series of ice concentration at the shelfbreak for each of the marginal seas from the Canadian Beaufort Shelf to the Kara Sea and for, 1978–2011. The biggest change has been from the Chukchi Sea to the Laptev Sea where there has been a gradual increase in summer and autumn open water over the shelfbreak with the open water season both beginning earlier in the year and ending later. In particular, the shelfbreak of the East Siberian Sea was ice covered in the, 1980s and now is open up to 3 months in summer. The smallest changes are over the Canadian Beaufort Shelf and the Alaskan Beaufort Shelf with their proximity to the remaining old multi-year ice. We include the Chukchi Sea in this comparison, though it is an inflow shelf, as it forms part of the continuous shelf from the eastern Canadian Beaufort Shelf to the western Laptev Sea.

The oceanography of the interior shelves is thus substantially different than that established by the continuous advection of water across the inflow and outflow shelves, and is also undergoing rapid change. As such, we expect they will respond differently to scenarios of climate change. We here review the physical oceanographic settings of these shelves with a particular focus on shelfbreak exchange and its supply of new nutrients across the shelfbreak for new primary production. We then show seasonal and annual time series of surface-stress at the shelf-break of the shelves from the eastern end of the Canadian Beaufort Shelf all the way to the Kara Sea, and consider how recent changes in the sea-ice and wind could now be providing a significantly different setting. Note the Kara Sea is included, but with complex bathymetry and surrounding islands, the concept of a linear shelf, shelfbreak and slope is less appropriate. A full pan-arctic integration of interior shelves will require more detailed and interdisciplinary

**Fig. 1.** A map of the Arctic Ocean showing the marginal seas including the ‘interior shelves’ of the Kara Sea, Laptev Sea, East Siberian Sea and Beaufort Sea. Superimposed are September minimum sea ice extents using data from the National Snow and Ice Data Center. The two lowest ice extent years 2007 (pink) and 2012 (green) display summertime retreat of ice from the shelves to the deep Arctic basins.
observations of the vast Siberian shelves, particularly the East Siberian Sea.

2. Shelf morphology and characteristics

2.1. Regional setting

The so-defined Interior Shelves extend un-broken from the eastern boundary of the Barents Sea to the western boundary of the Chukchi Sea (Fig. 3a) and then from the eastern boundary of the Chukchi Sea to the entrance of the Canadian Arctic Archipelago (Fig. 3b), and cover an area of $2.5 \times 10^6$ km$^2$ (Table 1). These shelves receive inflow from eight of the ten largest Arctic rivers or about 88% of the river discharge entering the Arctic Ocean, and four of the twenty largest rivers in the world, with maximum discharge occurring in June (Dai and Trenberth, 2002; Rachold et al., 2004). General features of these seas are described below.

The Kara Sea is bordered from the Barents Sea to the west by Novaya Zemlya and Franz Josef Land, on the south by the northern coast of Russia, and on the east by Severnaya Zemlya where it connects to the Laptev Sea through narrow (50 km wide) and deep (~200 m) Vilkitsky Strait. The deep northwestern boundary is cut by the Santa Anna and Voronin canyons, which drain into the Nansen Basin at approximately 70°E and 85°E, respectively (Aksenov et al., 2011). The average depth of the Kara Sea is ~110 m and its area is $926 \times 10^3$ km$^2$ (Table 1). The water mass characteristics of the Kara Sea derive from the inflow of modified Atlantic Water from the Barents Sea together with major discharges from the Ob and Yenisei rivers, which together supply about 30% of the total discharge into the Arctic Ocean (see Aagaard and Carmack, 1989; Shiklomanov et al., 2000; Carmack, 2000; Rachold et al., 2004; Serreze et al., 2006; Haine et al., 2015, for various budget tabulations). The Kara Sea typically remains colder than 0 °C throughout the year and is ice covered much of the year (Pavlov and Pfirman, 1995; Kulakow et al., 2006). First year ice reaches maximum thickness in May, typically 1.5–2.0 m (Polyakov et al., 2003). Interannual variability of landfast ice extent is significant, and exhibits two dominant modes: one associated with predominantly westward winds (larger fast ice extent) and one associated with eastward winds (smaller extent) (Divine et al., 2003, 2005). Salinity exhibits pronounced temporal and spatial variability related to seasonal fluctuations in river runoff and wind forcing, as well as to ice formation and melting (Hanzlick and Aagaard, 1980a, 1980b; Johnson et al., 1997; Harms et al., 2000; Janout et al., 2015). Circulation features are reviewed by McClimans et al. (2000) and Janout et al. (2015) who note strong topographic steering of flow. Harms et al. (2000) modelled dispersion of river runoff as a potential transport mechanism for dissolved or suspended contaminants and noted that the Kara Sea branch dominated the Eurasian Branch of the Transpolar Drift. Harms and Karcher (1999, 2005) used a regional coupled ice-ocean model to study dispersion of the Ob and Yenesey River plumes over a five year period in relation to sea-level pressure anomalies. They found that a positive anomaly caused a blocking situation of both freshwater export and Atlantic Water inflow from the Barents Sea. Janout et al. (2015) used observations and model results to describe the transport of freshwater through Vilkitsky Strait, noting control of transport by regional summer pressure patterns.

The Laptev Sea is bordered to the west and east by the Severnaya Zemlya and the New Siberian Islands, respectively, yielding a surface area of $498 \times 10^3$ km$^2$. The Laptev Sea is shallow, with depths of less than ~50 m over most of the shelf, and the shelfbreak occurs in water depths ~100 m at distances of 350–400 km from shore. Numerous sea valleys cut the shelf and provide...
cross-shelf transport corridors. The regional hydrography is dominated by the Lena River, and the thickness of the salt-stratified summertime mixed-layer is typically 5–10 m (Pivovarov et al., 2006). The surface circulation pattern is extremely variable and is mainly forced by winds; the average summer surface layer motion is a very slow cyclonic water circulation (Pavlov, 1998). Tides are generally small and assumed not to affect water mass transformation (Padman and Erofeeva, 2004; Pnyushkov and Polyakov, 2012, but see tidal section below). During freeze-up landfast ice develops near the coast, flaw polynyas form offshore at depths ~20 m, and mobile pack ice lies at the offshore limit of the open water area (Zakharov, 1966; Pivovarov et al., 2006) and thus contribute to the suite of flaw polynyas that encircle the Arctic Basin (cf. Timokhov, 1994). Within the Arctic, the Laptev Shelf sustains the greatest ice export, feeding the transpolar drift with sediment-laden first-year ice (Eicken et al., 1997, 2000; Dethleff, 1995).

The East Siberian Sea lies between the New Siberian Islands to the west and Wrangel Island to the east. It is the largest (987 x 10^3 km^2), broadest and shallowest of the Siberian shelves. The bottom topography is intersected by two major sea valleys that extend from the Indigirka and Kolyma rivers. There are two main surface water types in the East Siberian Sea: western surface waters are strongly diluted by direct river input and from the import of relatively freshwater from the Laptev Sea; eastern surface waters reflect the influence of surface waters of the Arctic

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**Table 1**

<table>
<thead>
<tr>
<th>Sea</th>
<th>Large rivers</th>
<th>Annual river discharge, km^3/yr</th>
<th>Shelf area, x 10^4 km^2</th>
<th>Annual discharge/shelf area, m/yr</th>
<th>Approximate Shelf width, km</th>
<th>Shelf break depth, m</th>
<th>Shelf break width, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kara Sea</td>
<td>Ob, Yenisei</td>
<td>1650</td>
<td>77</td>
<td>2</td>
<td>350</td>
<td>~100</td>
<td>5</td>
</tr>
<tr>
<td>Laptev Sea</td>
<td>Lena</td>
<td>822</td>
<td>44</td>
<td>2</td>
<td>~480</td>
<td>100</td>
<td>5</td>
</tr>
<tr>
<td>East Siberian Sea</td>
<td>Kolyma</td>
<td>204</td>
<td>82</td>
<td>0.25</td>
<td>480–850</td>
<td>100</td>
<td>5</td>
</tr>
<tr>
<td>Alaskan Beaufort Shelf</td>
<td>None</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>5</td>
</tr>
<tr>
<td>Canadian Beaufort Shelf</td>
<td>Mackenzie</td>
<td>316</td>
<td>6.2</td>
<td>6</td>
<td>130</td>
<td>80</td>
<td>5</td>
</tr>
</tbody>
</table>
Basin and Pacific inflows (Pivovarov et al., 2006). Near-coastal flows are dominated by the cold, fresh, and highly variable east-flowing Siberian Coastal Current (Weingartner et al., 1999; Munchow et al., 1999). As in the Laptev, recurrent flaw polynyas form at the boundaries between the landfast and pack ice.

The Beaufort Sea shelf is comprised of the Alaskan Beaufort Shelf and Canadian Beaufort Shelf and they are by far the narrowest and geomorphologically linear of the interior shelves (see below). Both these shelves begin and end at large undersea canyons that cross the entire shelf: the Alaskan Beaufort Shelf bounded by Barrow Canyon in the west and Mackenzie Trough in the east and the Canadian Beaufort Shelf bounded by Mackenzie Trough in the west and Amundsen Gulf in the east. River inflow is substantially different between the two shelves. The Canadian Beaufort Shelf is dominated by the inflowing Mackenzie River while the Alaskan Beaufort Shelf does not have a large inflowing river. Instead inflow to the Alaskan Beaufort Shelf is from numerous smaller rivers that drain the north slope of the Brooks Range and also from the brackish Alaska Coastal Current that contains outflow from the Yukon River (Oikkonen et al., 2009; Weingartner et al., 2005).

### 2.2. Geomorphological characteristics of the shelves

Geomorphological characteristics of shelves have direct impact on along-isobath shelf and shelfbreak flow and cross-isobath flow in bottom boundary layer. Important factors are shelf-width, shelfbreak depth, shelfbreak width, bottom slope and along-shelf variation. Each of these factors constrain the regional response to freshwater discharge and wind forcing, and are explored below for each shelf.

The interior shelves vary enormously in area and width (see Table 1). The narrowest shelf is the Alaskan Beaufort Shelf, at 65 km from the coast to the shelfbreak, and this is half as wide as the as the neighbouring Canadian Beaufort Shelf (130 km wide), but both of these shelves are narrow in comparison to the ~480 km-wide Laptev Sea shelf and the enormous ~700 km-wide East Siberian Sea shelf. The Kara Sea is also huge, approximately 1500 km long and 500 km wide. Deeper water on the western side against Novaya Zemlya (the Novaya Zemlya Trough) suggests that the width of the Kara Sea, from Nova Zemlya to the great Yenisei and Ob Rivers on opposite shore, is best taken as the cross-shelf direction and the length of the Kara Sea as the along-shelf direction, though, as mentioned above, definition of a linear shelf, shelfbreak and slope here is not particularly satisfactory.

The shelfbreak region also varies between these shelves; average shelf-shelfbreak-slope profiles for each of the marginal seas are shown in Fig. 4, and are calculated using the IBCAO dataset (Jakobsson et al., 2012). The shelfbreaks of the Alaskan Beaufort Shelf, Canadian Beaufort Shelf and the Laptev Sea are all very shallow (70, 90 and 100 m respectively) and the shelfbreak region, between the shelf and the slope, is narrow (<5 km wide). The East Siberian Sea is unusual in having a double shelfbreak: the shelf slopes very gently to ~100 m deep, at which point the shelf becomes ~6° steeper up to reaching ~400 m deep about 150 km further offshore. Beyond 400 m deep is the continental slope proper (see Table 1 for a summary). As we are interested in upwelling of nutrient-rich Pacific- or Atlantic-origin water to the continental shelf, and these waters reside beneath the euphotic zone between 100 and 400 m deep, the break in shelf-slope at ~100 m is the most interesting, and thus represents the functional shelfbreak for shallow East Siberian Sea for this study. We expect the dynamics of the East Siberian Sea shelfbreak to be different to the much more pronounced shelfbreaks of the Beaufort and Laptev Seas (see below).

![Fig. 4. Average cross-shelf profiles for the interior shelves of the Arctic Ocean. The shallow and sharp shelfbreaks are the Canadian Beaufort Shelf, Alaskan Beaufort Shelf and Laptev Sea, while the deeper and more rounded shelfbreaks are the East Siberian Sea. Where the shelves are wider than shown, a dotted line is used to indicate continuation to the coast. The profiles have been aligned for easy comparison of the shelfbreaks, so the cross-shelf location of the shelfbreaks are arbitrary. The Kara Sea is also shown, but note this sea is not readily defined interms of a linear shelf, shelfbreak and slope.](image-url)
Laptev Sea the shelf is almost completely blocked by the Taymyr Peninsula, an undersea canyon and an offshore island, so that flow from the Kara Sea tends must pass through the 50 km wide Vilkit-sky Strait (Janout et al., 2015). At the eastern end of the Laptev Sea the New Siberian Islands block about a third of the shelf so the flow between the Laptev Sea and the East Siberian Sea must either be focussed in the straits between the islands (Sannikov Strait), or the islands and Siberia (Dmitry Laptev Strait), or flow along the outer shelf. In the Beaufort Sea, the large canyons break the conti-nuity of the shelf, with Barrow Canyon in the west, Mackenzie Trough in the middle and Amundsen Gulf in the east. These canyons focus and enhance cross-shelf transport and, in addition, cut nearly all the way across the shelf, leading to strong isobath convergence that further enhances cross-shelf transport (see below).

3. Freshwater: Inputs, plumes and coastal flows

3.1. Inputs

The Arctic Ocean is surrounded by Eurasia and North America and large watersheds drain from these continents into the Arctic Ocean providing a huge annual freshwater influx (gauged) of 3740 km$^3$/yr; and additional ~600 km$^3$/yr likely derives from smaller, ungauged inputs (cf. Carmack, 2000). This comprises approximately 10% of the global river runoff (Dai and Trenberth, 2002), and, because the Arctic Ocean is only 4% of the global ocean area, the proportional amount of river runoff into the Arctic Ocean is approximately 2.5× the global average. This leads to the strong stratification by salinity that is so characteristic of the surface waters of the Arctic Ocean (cf. Carmack, 2007).

Most of the freshwater inflow to the Arctic Ocean is via the large rivers that drain into the interior shelves (Carmack, 2000). The largest rivers are the Ob and Yenisei Rivers (~427 and 673 km$^3$/yr respectively) that drain into the Kara Sea, the Lena River (~588 km$^3$/yr) that flows into the Laptev Sea, the Kolyma River (~136 km$^3$/yr) that flows into East Siberian Sea, and the Mackenzie River (~316 km$^3$/yr) that flows onto the Canadian Beaufort Shelf in the Beaufort Sea (Carmack, 2000). The other large Arctic river is the Yukon River (~208 km$^3$/yr) which flows into the Bering Sea south of Bering Strait and contributes to the riverine Alaska Coastal Current that passes through Bering Strait, along with Pacific-origin water, to the Chukchi Sea (Woodgate et al., 2005). These large rivers form about 65% of the Arctic freshwater inflow. There are additional 8 mid-sized rivers in Siberia that constitute another 15% of the inflow and the remaining small rivers the final, 20% (Holmes et al., 2012) Note there has been a 7% increase in discharge from 6 largest Russian Arctic rivers from, 1936 to 1999 (Peterson et al., 2002; Shiklomanov and Shiklomanov, 2003), but no equivalent increase from the Mackenzie River.

The annual freshwater inflow to each of the interior shelves can be divided by the area of each shelf to find how thick the annual freshwater inflow would be if it were spread evenly over the shelf. This is a crude measure of the influence of the river inflow to the shelf. By this measure, the Canadian Beaufort Shelf is the most influenced, receiving ~6 m of river water per year; the Kara and Laptev Seas are intermediate, receiving ~2 m of river water; and the East Siberian Sea is the least, receiving only ~0.25 m of freshwater (see Table 1).

Arctic river flow also has strong seasonality and much of the inflow occurs in late spring and summer. Each river either freezes solid or has low flow in winter (perhaps fed by freshwater lakes), then, in late spring/early summer (end of May/beginning of June), snow melt on the southern watersheds leads to sharp peak in the flow rate by the end of June which then tapers off to the low wintertime flow by November. Examples of the seasonality in flow rate can be found in many papers (e.g. Carmack and Macdonald (2002) for the Mackenzie River).

3.2. Dynamics of River plumes and their cross- and along-shelf transport

The Lena, Kolyma and Mackenzie Rivers all flow into the Arctic Ocean through coastal deltas (Walker, 1998; deltas form due to a combination of high river sediment load and weak tidal mixing and weak wave action). In a delta, river water flows directly out into the coastal ocean rather than mixing in an estuary first leading to a large density difference between more saline water and riverine water in the coastal ocean. A so-called ‘surface trapped’ plume forms in which the majority of the river water tends to ‘pool’ offshore of the delta (Yankovsky and Chapman, 1997; Fong and Geyer, 2002). This pooling of water leads to a 2-layer system with brackish plume water at the surface, marine water underneath and very strong stratification between the two. This plume will tend to stay close to the delta, growing in size as river water flows in, unless it is advected in shelf flows, that are generally along iso-baths, or moved by wind-driven Ekman transport. These plumes are particularly responsive to the wind as the Ekman transport is almost entirely confined to the thin plume layer due to the strong stratification at its base (Fong and Geyer, 2001; Lentz, 2004). If the wind is along-shelf with the coast on the left (upwelling favourable and roughly westward here), the plume is pushed offshore by Ekman transport, often revealing the underlying cold salty water at the coast, and advected along-shelf in the direction of the wind by the wind-driven along-shelf flow (see below). If the wind is along-shelf with the coast on the right (downwelling favourable and roughly eastward here), the plume is pushed onshore by the wind where, if the wind persists, it forms a coastal current that flows rapidly in the same direction as the wind (e.g. Williams and Carmack, 2008; Austin and Barth, 2002; Banas et al., 2009). The along-shelf flows that are generated by along-shelf wind-stress are in the same direction as the stress and have the potential to create strong connections between the interior shelves. Connections are made from east to west during persistent upwelling-favourable surface stress, and rapidly from west to east during downwelling-favourable surface-stress in coastal currents (Carmack et al., 2015).

In addition to the broad, estuarine structure spanning the width of the shelves, a narrow (~10 km) and shallow (~10 m) coastal-trapped band of low salinity water is persistently observed to flow in a cyclonic (west to east) direction. Carmack et al. (2015) refer to this as the Riverine Coastal Domain (RCD) and note that is a contiguous feature primarily driven by continental runoff and which extends clockwise ~ 10 × 10$^3$ km around northern North America, with a similar feature extending around northern Eurasia. It affects light, nutrient and carbon regimes, and provides a coastal pathway for the dispersal and migration of marine biota.

4. Shelfbreak flows and shelfbreak exchange

The simplest model of upwelling across the shelfbreak of a continental shelf is 2-dimensional, using an ‘average’ cross-shelf profile for bathymetry, which assumes no variation in the along-shelf direction. The ocean is then forced with a uniform surface-stress towards the west caused by either wind or ice motion or a combination of the two (see cartoon in Fig. 5). In this simplified upwelling circulation, the surface-stress spins-up an offshore-directed surface Ekman layer in a few inertial periods, and this offshore flow removes water from the coast creating low
sea level there and a sea surface slope away from the coast. A geostrophically balanced along-shore flow in the same direction of the surface-stress develops in response to the sea-surface slope, and this along-shore flow rubs against the bottom causing the spin-up of upwelling in an onshore-directed bottom boundary layer. The flows all come into balance when the alongshore flow is fast enough to provide a bottom friction equal and opposite to the surface-stress. At this point the onshore transport in the bottom boundary layer equals the offshore transport in the surface Ekman layer so the sea-surface slope stops increasing and the along-shelf flow stops increasing.

Near the top of the slope and over the shelfbreak the flow is modified due to the strong density stratification intersecting the steep slope. The bottom boundary layer over the slope is suppressed by stratification as an arrested bottom-boundary layer (e.g. Garrett et al., 1993) in which tilted-up isopycnals near the bottom reduce the geostrophically balanced along-shelf flow at the bottom and thus reduce the bottom friction that drives the onshore bottom-boundary layer. This arrested bottom-boundary layer has two main consequences: (1) the water that upwells to the shelf across the shelfbreak must be drawn from the interior near the top of the slope instead of the bottom boundary layer, and this water is from sufficient depth to contain high nutrient values; (2) low bottom friction over the slope means that the along-shelf flow there continues to accelerate. Over the upper slope there is less water for the wind to accelerate than in deeper waters farther offshore, and so the acceleration is greater there, leading to faster flow and the shelfbreak jet. This effect is more pronounced for the interior shelves as the shelfbreak is both shallow (there is less water to accelerate over the upper slope) and sharp (the upper slope ends abruptly).

Despite the robust dynamics of this idealised 2-dimensional upwelling it is highly simplified as it requires no along-shelf variation in the shelf topography for approximately 500 km in the Arctic (Pringle, 2002). The interior shelves have large variations in width and so also large along-shelf convergence and divergence of isobaths. The shelves also begin end and have islands, banks and canyons. All these features make surface-driven along-shelf flow, and so upwelling and downwelling, more complex, increasing or decreasing its amplitude and creating local upwelling hotspots (Williams and Carmack, 2008; Williams et al., 2006; Mathis et al., 2012). They are discussed below.

Pringle (2002) considered the amplification and suppression of upwelling circulation due to moderate changes in shelf width. The basic idea is that since along-shelf geostrophically balanced flow is constrained to follow isobaths, wind-driven along-shelf flows must accelerate if the shelf becomes narrower creating greater bottom friction and thus stronger upwelling. Likewise if the shelf becomes wider the along-shelf flow decelerates and there is weaker upwelling. The dynamics are such that the change in upwelling occurs at and to the east of the change in width (for these Arctic shelves). As there are many changes in shelf width and isobath separation in the interior shelves, such as the East Siberian Sea and the New Siberian Islands, or the shelf valleys of the Laptev there is potential for this mechanism to have a significant effect on cross-shelf transport and shelfbreak exchange.

Undersea canyons that cross the shelfbreak are known to enhance shelfbreak upwelling. The most prominent and well known shelfbreak canyons in the interior shelves are Barrow Canyon and Mackenzie Trough in the Beaufort Sea, both of which have enhanced upwelling (Aagaard and Roach, 1990; Carmack and Kulikov, 1998; Williams et al., 2006). Numerous smaller shelfbreak canyons are also present in the IBCAO bathymetry for the Laptev Sea and Alaskan Beaufort Sea, and appear as ‘corrugations’ to the shelfbreak. The effect of these has not been explored.

There are also extreme changes in shelf width on the interior shelves. These are the western end of the Laptev Sea at the Taymyr peninsula, Barrow Canyon, Mackenzie Trough and Cape Bathurst at the eastern end of the Canadian Beaufort Shelf. At these locations isobaths converge by 10 times or more and so are potential locations of extreme acceleration of along-isobath flow and thus enhancement of upwelling. This been documented for Cape Bathurst (Williams et al., 2008) and strong evidence points towards it amplifying upwelling in Mackenzie Trough at Herschel Island (Williams and Carmack, 2012; Mathis et al., 2012). These locations add significantly to the distributed upwelling occurring along the shelfbreak, e.g. it has been estimated localised upwelling at Cape Bathurst roughly doubles the nutrient flux to the Canadian Beaufort Shelf.

We have little direct evidence thus far for upwelling across the shelfbreak in a thin bottom boundary layer over the interior shelves, which we expect is due to the factors mentioned above and difficulty in observing the bottom boundary layer. However, off-shelf Ekman transport during upwelling-favourable
surface-stress is robust, as is the subsequent generation of along-shelf flow. As water that is lost from the shelf in surface Ekman transport must be replaced by on-shelf flow somewhere, we proceed by using along-shelf surface-stress at the shelfbreak as a proxy for the return upwelling flow.

5. Regional shelf, shelfbreak and slope circulation

5.1. Laptev and East Siberian Seas

The wind-driven (surface-stress) responses of surface (plume) waters are evident over the East Siberian Sea and Laptev Sea shelves. Dmitrenko et al. (2005) analysed surface salinity over the shelf and divided the data into years with negative or positive atmospheric vorticity. In years with negative atmospheric vorticity, high pressure centred north of the shelf leads to westward wind over the shelf and offshore and westward transport of the surface plume layer revealing saltier water near the coast and in the east. Plume water crosses the shelfbreak and can be found north of the Laptev Sea (Guay et al., 2001). These are also years in which Pacific-origin waters likely protrude westward from the Chukchi Sea into the East Siberian Sea through Long Strait (Semiletov et al., 2005; Munchow et al., 1999; Weingartner et al., 1999) and freshwater storage on the shelf is low (Dmitrenko et al., 2008). In years when there is positive atmospheric vorticity, the high pressure is shifted east to north of Alaska. In these years summertime winds are weaker and may turn eastward and riverine water tends to be retained on the shelf (Dmitrenko et al., 2008). Eastward wind leads to onshore Ekman transport, accumulation of riverine water against the coast, and eastward alongshelf flow, allowing the Siberian Coastal Current to form (Weingartner et al., 1999).

Cross shelf transport across the Laptev Sea shelf has been examined by Bauch et al. (2009) who looked at summertime data in a downwelling-favourable year (1994) and an upwelling-favourable year (1999). They show that, as above, downwelling-favourable wind pushes the plume onshore and generates current to the east, and that the winter remnant of cold salty polynya-formed water underneath the plume also moves this way. Upwelling-favourable wind pushes the plume offshore, the polynya water westward and new salty water floods the bottom of the shelf. Bauch et al. (2011) also used an Ekman model to explain freshwater distributions in 1993, 1995, 2005 and 2006 and demonstrated that they were not correlated with the Arctic Oscillation. Dmitrenko et al. (2010a, 2010b) and Janout et al. (2013) continued these investigations showing that offshore Ekman transport brings halocline waters across the shelfbreak, from near the shelfbreak depth, and onshore across the bottom of the shelf. Slope waters in the Laptev have had similar investigations. Dmitrenko et al. (2006) showed that offshore wind, and offshore surface currents lead to onshore movement of the Atlantic Water current core that forms the source waters for this slope region (Dmitrenko et al., 2011, 2014; Bauch et al., 2014).

5.2. Beaufort Sea

5.2.1. Canadian Beaufort Shelf

Carmack and Macdonald (2002) describe the physical and geochemical seasonality of the Canadian Beaufort Shelf, noting key differences in the joint roles of river discharge, ice disposition and wind-forcing during the winter, spring, summer and fall seasons. Mackenzie River plume in summer is highly variable. During upwelling-favourable wind, the warm Mackenzie Plume is observed to move offshore, revealing cold, salty water near the coast, and along-shelf to the west (Wood et al., 2013). This is wind-driven offshore Ekman transport and westward along-shelf flow. On-shelf transport of water near the seafloor in the shallow Kugmallit Valley during upwelling favourable along-shelf flow has also been documented (Williams et al., 2008; H. Melling, pers. comm.). The along-shelf flow also draws deep nutrient-rich water up to the surface and onto the shelf at Cape Bathurst, where strong convergence of isobaths accelerates along-shelf flow and drives intense localised upwelling (Williams and Carmack, 2008), and at Herschel Island, where Mackenzie Trough and isobath convergence cause upwelling (Carmack and Kulikov, 1998; Williams et al., 2006; Williams and Carmack, 2012; Mathis et al., 2012).

Observations of upwelling across the shelfbreak of the Canadian Beaufort Shelf are sparse. In the absence of wind, there is an eastward current over the top of the slope (Ingram et al., 2008; Melling and Lewis, 1982) which shifts to westward flow during upwelling-favourable surface stress (Melling and Lewis, 1982; H. Melling pers. comm.). In addition, during upwelling favourable surface stress, uplifted isofalines have been observed over the upper slope and are evidence of an arrested bottom boundary layer (H. Melling, pers. comm.) and shelfbreak exchange has been observed (Melling and Moore, 1995; Melling, 2012).

Kulikov et al. (1998) analysed current meter records from the shelfbreak of the Canadian Beaufort Shelf and partitioned the kinetic energy into three main frequency bands: low frequency (seasonal to fortnightly), intermediate or synoptic frequency (fortnightly to 1½ days, and high frequency (<1½ days); within this they found ~65% of the kinetic energy in the low frequency band, 15% in the intermediate frequency band, and ~20% in the high frequency band, with the latter being dominated by tidally related motions. Low frequency variability is highest during the autumn period of intense storms, and lowest in late winter owing to wind shielding and increased frictional damping by the ice cover. Carmack and Kulikov (1998) show that some of this variability is due to an internal Kelvin wave that propagates eastward along the slope and is generated by relaxation of large amplitude upwelling in Mackenzie Trough.

5.2.2. Alaskan Beaufort Shelf

Wind-driven upwelling and along-shelf transport has been well studied on the Alaskan Beaufort Shelf, particularly at 152 W using a high resolution moored array that was deployed from 2002 to 2004 (e.g. Pickart et al., 2009). As with the Canadian Beaufort Shelf, in the absence of wind the flow over the shelf and upper slope is towards the east (e.g. Nikolopoulos et al., 2009) and transports Alaska Coastal Current water and Pacific-origin water. During upwelling favourable surface-stress, the alongshelf flow reverses and becomes strongly westward and there is large amplitude upwelling of Pacific- and Atlantic-origin water across the shelfbreak (Pickart et al., 2009, 2011, 2013b) especially during intense autumn storms associated with the Aleutian Low. On examining an autumn upwelling event in detail, Pickart et al. (2011) then showed that the baroclinic structure of upwelling, and thus, to some extent, the upwelled water, remains after the upwelling-favourable winds have passed and the westward wind-driven barotropic flow has propagated away. Pickart et al. (2013a) examine the long-term trends of upwelling-favourable winds over the Alaskan Beaufort Shelf and identify 9–10 upwelling events per year, but note that the number and strength of events has increased over the last 25 years and that this is due both to an increase Beaufort High and the location of Aleutian Low storm track. Brugler et al. (2014), considering seasonal and interannual variability, report an 80% decrease in the eastward flow of Pacific Origin water at 152 W from 2002 to 2011, and show that it is due to a summertime intensification of the summer Beaufort High and deepening of the summer Aleutian Low (cf. Wood et al., 2013).
5.3. Examples of cross shelf salinity and temperature structure

River inflow is the primary influence on near-surface the stratification in summer for the interior shelves. As examples we show in Fig. 6 two cross-shelf sections of salinity and temperature, one for the Laptev Sea and one for the Canadian Beaufort Shelf (their locations are shown by the red lines in Fig. 3a and b). The sections shown are just one realisation of cross-shelf transects from near the river delta to the shelf-break for these shelves, and there is very large synoptic (wind-driven), seasonal and interannual variability. The Canadian Beaufort Shelf section was taken in early August, 2006, a summer where weak winds meant little mixing or transport and a very fresh plume accumulated offshore of the delta. The Laptev Sea section was taken in September, 2007 and shows a much thicker, saltier and cooler surface layer. It is also necessary, again, to recognise the huge difference in width of the shelves. Though the Canadian Beaufort Shelf section stretches from 30 km offshore of the delta all the way to the shelf-break, the entire section is closer to the Mackenzie Delta than the beginning of the Laptev Sea section is to the Lena Delta. A continuation of the Laptev Sea section towards the delta would reveal stronger influence of the plume. Sections with similar characteristics of river influence and stratification are available for the Kara Sea and East Siberian Sea (M.A. Janout, pers. comm.).

6. Tides and tidally generated flows

Tidal motions are a potential source of vertical mixing which can bring nutrients from depth to the euphotic zone that are then used by primary producers. It is therefore one way in which nutrients brought onto the bottom of the shelf by shelfbreak upwelling can reach the surface. In high latitudes, however, there is weak astronomical forcing of the tides (the main energy bands are the M2, S2, K1 and O1 components) and most of the tidal energy in the Arctic Ocean propagates-in from the Atlantic Ocean (Godin, 1988). Thus tidal energy is generally low (Kowalik et al., 1994; Padman and Erofeeva, 2004) and the interior shelves tend to have weak tidal velocities relative to the continental shelves of the large oceans. Most of the continental shelf from the Kara Sea to the Beaufort Sea has tidal velocities of <3 cm/s (Padman and Erofeeva, 2004; cf. Kagan et al., 2008). Tidal velocities are larger (~10 cm/s) over these shelves in areas of topographic resonance, or a where there is a combination of topographic and inertial resonance near the critical latitude of the M2 (74°28’18’’N) or N2 (70°58’48’’N) tides (Dmitrenko et al., 2012; Kagan et al. 2008; Kowalik et al., 1994; Furevik and Foldvik, 1996). These areas are (from Fig. 3 of Padman and Erofeeva, 2004) the southernmost Kara Sea, the numerous banks of the northern Kara Sea, the southwestern Laptev Sea, around the New Siberian Islands, the outer shelf north of the New Siberian Islands between the Laptev Sea and the East Siberian Sea, the area around Wrangel Island, and the eastern end of the Canadian Beaufort Shelf. In addition, some of the shallow areas with high tidal velocities are predicted to be tidally well-mixed using the Simpson and Hunter (1974) tidal mixing parameter (Chen et al., 2009) and are likely areas where mixing-induced vertical nutrient fluxes feed primary producers. These areas are immediately east of the New Siberian Islands, the southwestern-most Laptev Sea and the southern-most Kara Sea. Note that recent tidal modelling of the Laptev Sea by Janout and Lenn (2014) shows larger tidal velocities than the Arctic-scale model of Padman and Erofeeva (2004) which implies greater vertical mixing and is supported by current meter records from the mid-shelf (Dmitrenko et al., 2012).

Particular attention has been paid to areas of tidal amplification in the Laptev Sea and north of the New Siberian Islands. Kowalik et al. (1994) showed that in these areas non-linear interaction of the barotropic tide with the bathymetry, and with the sea ice, create residual tidal currents and ice motion leading to ice production in tidal polynyas on the outer shelf. More recently the role of the baroclinic tide in causing vertical mixing in the Laptev Sea has been examined: Lenn et al. (2011) observed that intense vertical mixing occurred on the outer shelf when the rotating baroclinic tidal shear vector in the pycnocline aligned with the surface stress during a tidal cycle and Dmitrenko et al. (2012) examined the role of the baroclinic tide in causing vertical mixing as the flaw lead polynya opened-up at the edge of the landfast ice. These are both

![Fig. 6. Examples of temperature and salinity sections across river influenced 'interior shelves'. (a) A section across the Laptev Sea shelf in September, 2007 (the section runs from 'near' the Lena River delta to the shelfbreak, and its location is marked in Fig. 3(a), thanks to Igor Polykov for the NABOS data). (b) A section across the Beaufort Sea shelf in early August, 2006 (the section runs from near the Mackenzie River delta to the shelfbreak, and its location is marked in Fig. 3(b)).](image-url)
processes that modulate stratification and vertical nutrient fluxes over the shelf. Janout and Lenn (2014) show that the freshwater distribution from the Lena River, that strongly influences the stratification, also affects the baroclinic tidal structure. They note that a longer open water season will increase mixing and variability of the river plume and so affect the baroclinic tides too. Most recently, Fofonova et al. (2014) has made a high resolution barotropic tidal model of the inner Laptev Sea shelf which paves the way for more baroclinic tidal studies of the Lena-plume influenced region.

Tidal motions across the shelfbreak pulse the stratified water column, resulting in the generation of near-inertial internal waves which subsequently propagate seaward (Baines, 1982). If the generation site is near a critical latitude then the horizontal component of group velocity is near-zero, resulting in wave remaining close to the generation site and subsequent resonant amplification (Kulikov et al., 2004). Shelf areas at these latitudes will exhibit stronger vertical mixing associated with tides (Furevik and Foldvik, 1996).

7. Role of retreating ice cover

The effect of shelfbreak ice cover on wind-driven upwelling/downwelling circulation over a continental shelf was first explored by Carmack and Chapman (2003). They modelled the ice as immobile, so that when ice was present no wind-stress was transmitted to the ocean. The modelling illustrated a basic response: if the shelfbreak and beyond is covered with immobile ice, wind-driven upwelling circulation is restricted to the shelf and the nutrient-rich Pacific or Atlantic-origin water over the upper slope does not upwell to the shelf. If ice retreats beyond a Rossby radius from the shelfbreak upwelling occurs from the upper slope bringing new nutrients to the shelf. (The Rossby radius is defined as \( (gh)^{1/2} f / \), where \( h \) is the water depth, \( f \) is the Coriolis parameter, \( g \) is the reduced gravity = \( (g/\rho) d\rho/dz \) where \( \rho \) is water density and \( z \) is depth. Thus, as the Arctic sea ice has declined and retreated beyond the shelfbreak of the marginal seas from the Beaufort to the Kara for longer periods of time in the summer and autumn, there is the potential for a dramatic wind-driven increase in the nutrient supply to these shelves, if upwelling favourable along-shelf winds are present.

Modelling sea ice as immobile is an over-simplification. Ice moves in response to forcing by the wind and its motion is retarded by internal ice stress. An obvious example of this is from the mobile pack-ice north of the Canadian Beaufort Shelf in winter. In this region, if the wind blows strongly enough from the east and over a large enough area, the ice begins to move towards the west and fractures and opens up as it does so, reducing internal ice-stress and allowing it to move more freely westward. However, if the wind blows from the west, the ice motion is blocked by the Canadian Arctic Archipelago and there is little ice motion due to the wind (Williams et al., 2006; H. Melling, pers. comm.).

The trend towards thinning and reduced concentration of sea ice observed in recent years has resulted in a weaker internal ice stress (Shimada et al., 2006) and, as a result, ice that is free to move under a given wind-stress (Martin et al., 2014). In fact, when ice is not retarded by internal ice stress, and moves freely in response to the wind, it is possible to have greater surface-stress on the ocean than when ice is not present (Williams et al., 2006). This is expected from comparing the drag coefficients in the standard quadratic bulk formulae for wind-water stress and wind-ice stress (Wadhams, 2000; Martin et al., 2014). The ‘drag’ coefficient of air–water stress is \(~1.3 \times 10^{-3}\) whereas the drag coefficient for air-ice stress varies considerably with the surface roughness of the ice, but for ridged seasonal ice is thought to be \(~2–3 \times 10^{-3}\) (Wadhams, 2000). As the drag coefficient is much larger for wind–ice stress than wind–water stress, there will be greater stress over ice than there would be over open water and, if the ice is freely drifting, this can translate to greater surface stress on the ocean. Ice can drift more freely in the marginal ice zone and so this may be a region with elevated surface stress relative to the open water on one side and less mobile pack ice on the other (Wadhams, 2000). Observations over the Alaskan Beaufort Shelf support this when analysing both seasonal variation of upwelling (Schulze and Pickart, 2012) and upwelling during an autumn storm (Pickart et al., 2013b). The ice concentration at which ice begins to free drift is thought to be 80–90% (Martin et al., 2014). Our formulation of surface-stress used in the next section incorporates these effects on surface-stress by using ice velocity when ice is present and wind velocity when wind is present.

Greater surface stress over the interior shelves due to summertime melt and more mobile ice (compared to the presence of rigid multiyear ice) not only drives upwelling and downwelling but also contributes to greater vertical mixing over the shelves. This is important to mixing of nutrients up from depth to the nutrient depleted euphotic zone. Thus, for primary production, upwelling is a two-stage process, where shelf-break upwelling brings nutrient-rich water onto the bottom of the shelf and subsequent vertical mixing brings nutrients to the euphotic zone. Known exceptions to this are Mackenzie Trough and Cape Bathurst on the Canadian Beaufort Shelf where enhancement of upwelling by isobaths convergence brings deep nutrient-rich water directly to the surface.

8. Along-shelf surface-stress generated by wind and ice motion

Using ice concentration and velocity data from National Snow and Ice Data Center (NSIDC) and wind data from the National Centers for Environmental Prediction (NCEP), daily surface-stress at grid points over the shelfbreak is calculated as ice-water stress when ice is present and wind-water stress when ice-free (for partial ice cover the calculation prorates ice and wind-water stress by the fraction of ice and open water, e.g. Yang, 2009; Martin et al., 2014). We use the standard quadratic drag formulation for stress with a wind-water drag coefficient of 0.0015 and an ice-water drag coefficient of 0.0054 (Wadhams, 2000; Yang, 2009). We then calculate the time series from the beginning of the satellite ice record in 1979–2011. ‘Wind-roses’ for this daily time series of surface stress at the shelfbreak are shown in Fig. 7 for the marginal seas from the Canadian Beaufort Shelf to the Kara Sea, with the Chukchi Sea included, again, for continuity along the shelf. On each of the ‘roses’ the shelfbreak direction of the marginal sea is shown. It is clear that upwelling favourable along-shelf surface-stress predominates from the Canadian Beaufort Shelf to the East Siberian Sea. This is primarily due to relatively persistent Beaufort High that forms north of these shelves in conjunction with the Aleutian Low to the south (Pickart et al., 2013a) though this varies seasonally. (The Beaufort High is strongest in Autumn, Winter and Spring, and tends to weaken in Summer. The Aleutian Low strengthens in Autumn and Winter and weakens in Summer.) The predominance of upwelling favourable along-shelf stress is also enhanced by sea ice moving preferentially towards the west, as westward motion tends to open-up the sea ice in the Beaufort Sea, allowing it to move freely, and eastward motion tends to compress it against the Canadian Arctic Archipelago and its motion is blocked (Williams et al., 2006). West of the East Siberian Sea the wind patterns are very different. Surface-stress in the Kara Sea and Laptev Sea is more evenly distributed in all directions rather than along-shelf, but with offshore winds dominating over onshore winds. For the Laptev Sea, these offshore winds are northerly,
and are due to it being on the boundary between the Beaufort High in the east and the Icelandic Low in the west (Bauch et al., 2009; Dmitrenko et al., 2005). The location and strength of this boundary varies with the Arctic Dipole (Wang et al., 2009; Thibodeau et al., 2014). The wind to the north in the Laptev Sea produces the northward ice motion that both causes the persistent flaw polynya at the edge of the land-fast ice and is the well-known beginning of the wind-driven transpolar ice drift (Krumpen et al., 2013). Because wind directions are evenly spread from upwelling along-shelf to offshore to downwelling along-shelf, there is often focus on northward wind in the Laptev Sea as well as along-shelf wind (e.g. Dmitrenko et al., 2006). We note here that northward wind and ice motion over the Laptev Sea will produce westward Ekman transport which will tend to build sea level against the Taymyr Peninsula and thus provide forcing for northward geostrophic 'cross-shelf' flow. The shape of the Laptev Sea adds complexity to the ocean dynamics and flow.

Upwelling favourable winds can last for a few days (due to low pressure storms) or longer (due to high pressure anticyclones) and so upwelling nutrient fluxes can be persistent or episodic (e.g. Pickart et al., 2009, 2013a; Tremblay et al., 2011; Mathis et al., 2012) The duration and strength of an upwelling-favourable wind event matters as it affects: (1) the depth from which nutrients upwell, (2) the nutrient flux to the shelf, (3) the distance that nutrients can be transported across the bottom of the shelf, (4) the potential for upwelled water masses to mix with shelf water masses and (5) the utilisation of the nutrients in new production. In addition, between upwelling events, intervening periods of downwelling or relaxation from upwelling may also lead to the return of upwelled nutrients back across the shelfbreak to the basin.

This wind-event-based view of upwelling is thus very important for accurate estimation of upwelling fluxes over regional topography and their effects, but is beyond the scope of this integrative study. As offshore Ekman transport generated by along-shelf wind- and ice-motion remains a primary driver of along-shelf flow and upwelling, we show time series of along-shelf surface-stress at the shelfbreak of the interior shelves as a proxy for shelfbreak exchange and upwelling to the shelves. We focus, on seasonal and annual averages of our time series of surface-stress due to wind and ice motion from 1979 to 2011. These average over upwelling and downwelling events and do not include additional upwelling due to canyons and isobath convergence or persistence of nutrients on the shelf due to mixing.

The interannual variation of (a) autumn, (b) winter, (c) spring and (d) summer along-shelf surface-stress for the Canadian Beaufort Shelf to the Kara Sea is shown in Fig. 8. In autumn, from the Canadian Beaufort Shelf to the Chukchi Sea, there is strong upwelling-favourable stress due to cyclones to the south from the Aleutian Low and the Beaufort High to the north. Over the 1979 to 2011 time series, there is an obvious increase in both the total stress and the proportion that occurs over open water in autumn. The increase appears have begun in 1997, which coincides with a shift to a positive phase of the Arctic Ocean Oscillation (see below). Continuing westward to the East Siberian Sea, the diminishing influence of the Beaufort High leads to weaker upwelling-favourable stress and it can be very weakly downwelling in some years. For the Laptev Sea and Kara Sea under the influence of the edge of the Icelandic Low, the stress is similar in size to the East Siberian Sea but downwelling on average. For all these marginal seas the autumn surface stress is a combination of wind over open water and ice-motion, with open water proportion increasing over time.

In winter, near complete ice cover means that the along-shelf stress is almost entirely due to ice motion. Upwelling-favourable stress is again largest from the Canadian Beaufort to the Chukchi and weaker and more variable in the East Siberian Sea. The Laptev Sea and Kara Sea are downwelling favourable. In spring,
Fig. 8. Time series of seasonal averages along-shelf surface stress at the shelfbreak of the Canadian Beaufort Shelf, Alaskan Beaufort Shelf, Chukchi Sea, East Siberian Sea, Laptev Sea and Kara Sea for (a) autumn, (b) winter, (c) spring and (d) summer. Upwelling-favourable surface stress (along-shelf with the coast on the left) is plotted as positive and, for each year, the surface-stress is partitioned into the contribution from the wind blowing over open water (red) and that from ice-motion (blue; note ice is largely set in motion by the wind). The Chukchi is included as part of the progression from the Canadian Beaufort Shelf to the Laptev Sea.
along-shelf surface stress weakens, remaining upwelling favourable for the Canadian Beaufort Shelf to the East Siberian Sea. For the Laptev Sea and Kara Sea the along-shelf surface-stress is very weak.

In summer, the seasonal ice melt means that the majority of the surface-stress is due to wind over open water. Weaker and more variable winds that can be both upwelling-favourable or downwelling-favourable are found on all the shelves due to weakening of the Beaufort High and summertime weakening of the Aleutian and Icelandic Lows. The most striking feature of these time series is the shift towards summertime upwelling favourable wind after 2006 for the Canadian Beaufort to the Chukchi Sea. Winds in July and August are now upwelling favourable. Wood et al. (2013) argue that, since 2006, there summertime climate of the Beaufort Sea has shifted towards persistent upwelling favourable wind due to an intensified Beaufort High that extends over both the Canadian Beaufort Shelf and Alaskan Beaufort Shelf and into the Chukchi Sea. Brugler et al. (2014) also discuss this change over the Alaskan Beaufort Shelf and attribute it, in this location, both to intensification of the summertime Beaufort High and deepening of the summertime Aleutian Low. This shift to greater upwelling favourable wind during the open water season will have: (1) dramatically increased the offshore Ekman transport of surface waters, including the Mackenzie River plume, pushing Mackenzie River water towards the Beaufort Gyre and leaving the Beaufort Shelf saltier at the beginning of freeze-up and thus preconditioning the shelf for dense water production in the flaw lead (H. Melling pers. comm; c.f. Dmitrenko et al. (2010a, 2010b) for the Laptev Shelf); (2) dramatically increased the continuous (rather than episodic) nutrient supply to the Beaufort Shelf, along the shelfbreak and at Cape Bathurst and Mackenzie Trough, in the summertime when primary production that uses upwelled nutrients is possible; and (3) dramatically increased along-shelf flow towards the west over the Beaufort Shelf and upper slope (Brugler et al., 2014).

9. Annual average wind and ice-drift forcing

We summarise the seasonal along-shelf surface-stress time series by combining them in an annual average, shown for all the marginal seas from the Canadian Beaufort Shelf to the Kara Sea in Fig. 9. In this summary, the influence of the Aleutian Low and the Beaufort High is clear from the large-amplitude upwelling-stress from the Canadian Beaufort to the Chukchi Sea. Upwelling-favourable stress continues to the East Siberian Sea but this sea is further away from the centres of the Beaufort High and Aleutian Low and the upwelling stress is much weaker there. In addition, from the Canadian Beaufort Shelf to the East Siberian Sea there is a large increase (up to 100%) in upwelling-favourable stress from, 1997 onwards (green lines in Fig. 9). This is coincident with the change in the Arctic Ocean Oscillation (AOO) from negative to positive (Proshutinsky et al., 1999; McLaughlin et al., 2011) and also includes the effects of the persistence of the Beaufort High in summer since 2006 (Wood et al., 2013).

As described above, the Laptev Sea, Kara Sea and East Siberian Sea behave differently since they are not under the steady influence of the Beaufort High. The Laptev Sea and Kara Sea, under the influence of the eastward extension of the Icelandic Low, have significant wind in the alongshore and offshore directions and along-shore surface-stresses are weakly downwelling-favourable on average.

Cross-shelfbreak upwelling nutrient fluxes can be very crudely estimated from the annual average surface-stress by calculating the volume of upwelling as the total off-shelf surface Ekman transport and multiplying by an assumed nitrate concentration of the required on-shelf flow. Assuming upwelling from near the top of the slope the nitrate concentration will be from the nitricline that reaches a maximum at either the core of the Pacific Water, for the Alaskan and Canadian Beaufort Shelves, or the core of the Atlantic Water, for the East Siberian Sea and Laptev Sea. For this calculation nitrate flux is simply $N = \rho f$, where $N$ is nitrate concentration, $\rho$ is the along-shelfbreak stress, $\rho = 1000 \text{ kg/m}^2 \text{ s}^{-1}$ is water density, and $f$ is the Coriolis parameter. We use a representative nitrate concentration of 13 mmol/m$^3$. This conversion of the surface-stress to nutrient flux is shown on the right-hand axis of in Fig. 9. We reiterate here that these estimates of upwelling fluxes are quite possibly underestimates, as they assume that upwelled nutrients are returned to depth on downwelling, rather than remaining on the shelf and mixing with shelf water, and do not include localised topographic enhancement of upwelling.

The increase in summer and autumn upwelling-favourable along-shelf surface-stress from the Canadian Beaufort Shelf to the East Siberian Sea, implies:

(1) An increase in new primary production. Though sophisticated estimates of this require estimation of vertical nutrient flux, via wind-driven mixing or localised transport, to the euphotic zone (see Popova et al. (2010, 2012) for numerical modelling of primary production over the shelves), a simple estimate can be made by converting on-shelf nitrate flux to new production via the Redfield ratio. By this method, the increase in nitrate flux to the Canadian Beaufort Shelf after 1997 could represent an increase in new primary production of ~20 gC/m$^2$/yr.

(2) Greater offshore Ekman transport of the riverine surface layer, making the shelf more saline and thus preconditioning the shelf for wintertime dense water production in the flaw lead polynyas.

(3) Increased along-shelf flow from east to west and thus greater connectivity in the direction from the Canadian Beaufort to the East Siberian Sea.

10. A conceptual model of an interior shelf and biological implications

A conceptual model summarising the structure, domains and principle physical forcing mechanisms of interior shelves for the summer and winter seasons, based on the concept of positive (summer) and negative (winter) estuarine structures (Carmack and Wassmann, 2006; Macdonald, 2000), is shown in Fig. 10. In summer, the stratification is dominated by river discharge and seasonal ice melt, which generally weakens seaward from the coast, resulting in a positive estuarine circulation. Riverine waters spread from their sources both as plumes, spreading offshore, or as coastal-trapped gravity currents, spreading alongshore within the so-called Riverine Coastal Domain (Carmack et al., 2015). High levels of turbidity and generally low nutrient loadings, despite moderate subsidies from rivers, act to suppress primary production over the inner shelf domain (Carmack et al., 2004; Tremblay et al., 2014). Turbidity levels in the inner shelf are also increased by recurrent sediment re-suspension during strong wind events. Along the outer shelf, wind-forced upwelling events drive shelf-basin exchange that pushes river plumes offshore (Macdonald et al., 1999) and draws nutrient-rich halocline waters onto the shelf (Carmack and Chapman, 2003). For the Siberian shelves, these halocline waters are Atlantic-origin, whereas, for the Beaufort, Pacific-origin water overlays the Atlantic-origin water. Upwelling may be intensified where canyons intersect the shelfbreak (Carmack and Kulikov, 1998; Williams et al., 2008; Statscewich, 2014). Wind-forcing at the shelf-break also results in the formation of a strong, bottom-intensified, shelf-break jet. Tidal motions are generally weak over the interior shelf domains.
except where amplified by topography or resonance (Kulikov et al., 2004); such sites may be hotspots for marine birds and mammals (Conlan et al., 2013; Walkusz et al., 2012).

In winter (Fig. 10b), river inputs are dramatically decreased, often ceasing completely, except in rivers where water is drawn from large headwater lakes, such as the Mackenzie River. As a result of ice formation and brine release, the estuarine circulation may switch from positive to negative (cf. Melling and Lewis, 1982). The seasonal ice cover (~1.6–2.0 m thick) diminishes the effects of wind forcing. Typically, the ice domain consists of bottom and landfast ice extending seaward to depths of 10–20 m. In the Beaufort the landfast ice is bordered by a heavily ridged zone, known as ‘stamukhi’, which forms at the boundary between the stationary landfast ice and the drifting pack ice due to episodic onshore motion of the pack (Ogorodov et al., 2013; Mahoney et al., 2014). During formation of the stamukhi, ice ridges may be dragged across the bottom, resulting in deep troughs and disruption of benthic communities. The impact of bottom-fast and landfast ice, and especially ridge gouging, often eradicates benthic fauna during winter and spring and new populations must re-establish annually (Conlan and Kvitek, 2005). Incoming river water may be trapped behind the ice ridges of stamukhi, forming a shallow ‘lake’ of low salinity waters (Macdonald and Carmack, 1991). Seaward of the landfast ice, a recurrent flaw lead is often formed by offshore wind events. This open water zone is a site of intense air-sea heat exchange, resulting in rapid ice formation (as frazil ice) and brine-driven convection (Barber and Massom, 2007; Williams et al., 2007). Over the Siberian shelves, offshore wind predominates leading to offshore motion of the pack so that no stamukhi forms and these flaw-lead processes dominate.

For all interior shelves, high freshwater inputs from numerous rivers and streams produce an environment that is decidedly estuarine in character, especially during the late spring and the summer months. Compared to the inflow shelves, new primary production and general biological activity are low (cf. Sakshaug, 2004), and much of the allochthonous matter delivered by rivers is of a refractory nature (Dunton et al., 2006, 2012). Within the inner shelf, high turbidity and nutrient limitation due to stratification are the main causes for the low primary production. Suspended biomass of planktonic organisms is thus low, but that of benthic organisms is relatively high (Conlan et al., 2013; Roy et al., 2014). Some of the food for the benthic organisms is of

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Fig. 9. Time series of the annual average along-shelf surface stress at the shelfbreak of Canadian Beaufort Shelf, Alaskan Beaufort Shelf, Chukchi Sea, East Siberian Sea, Laptev Sea and Kara Sea. Upwelling-favourable surface stress (along-shelf with the coast on the left) is plotted as positive and, for each year, the surface-stress is partitioned into the contribution from the wind blowing over open water (red) and that from ice motion (blue). The alongshelf surface stress can be converted to an estimate of nutrient flux to the shelves; these values are shown by the right-hand axis. The Chukchi is included as part of the progression from the Canadian Beaufort Shelf to the Laptev Sea.
marine origin and derives from the estuarine circulation, some is locally produced and a significant amount derives from littoral and riverine sources (Dunton et al., 2012; Connelly et al., 2014). Natural transport and erosion (permafrost reduction) are significant. The role of terrestrially derived carbon in Arctic estuarine food webs is especially important in view of the current warming trend in the Arctic environment. Coastal erosion and river discharge are largely responsible for introducing high concentrations of suspended sediment from upland regions into the near-shore zone, and sediment is often trapped in the nearshore lagoons that are characteristic of the Beaufort Sea.

The lack of direct grazing pathways among Arctic biota on the Beaufort interior shelves results in the wide range in $^{13}$C values of eastern Beaufort Sea benthic fauna compared to the same species collected in the northeastern Chukchi (Dunton et al., 2012; McTigue, 2013). The wider spread in stable isotope values in the eastern Beaufort Sea also reflects a decoupling between benthic and pelagic components. A relatively low autochthonous production and a rich terrestrial, allochthonous supply means that the latter material is, to a larger extent, incorporated compared to more productive regions (Dunton et al., 2012).

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