The role of currents and sea ice in both slowly deposited central Arctic and rapidly deposited Chukchi–Alaskan margin sediments

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A B S T R A C T

A study of three long cores from the outer shelf and continental slope north of Alaska in the Arctic Ocean indicate that localized drift deposits occur here with sedimentation rates of more than 1.5 m/kyr during the Holocene. Currents in this area average about 5–20 cm/s but can reach 100 cm/s and these velocities transport the sediment found in these cores primarily as intermittent suspended load. These high accumulation sediments form levee-like deposits associated with margins of canyons cutting across the shelf and slope. Unlike most textural investigations of Arctic sediment that focus on the coarser ice-rafted detritus (IRD), this paper focuses on the ~95% of the sediment, which is finer than 45 μm. The mean size of this fraction varies between 6 and 15 μm in Holocene sediments from the Chukchi–Alaskan shelf and slope with the higher values closer to shore. Analysis of detailed size distributions of these Holocene deposits are compared to 34 sediment samples collected from sea ice across the Arctic Ocean and to Holocene sediment from central Arctic Ocean cores and indicate that similar textural parameters occur in all of these sediments. Principal components of these size distributions indicate that sea ice is an important link between the shelves and the central Arctic. Factor scores indicate nearly identical components in the clay and fine silt size fractions but very different components in the coarse silt for sea ice sediment and central Arctic ridge sediments compared to shelf and continental slope deposits. Sea ice must contribute to sedimentation in both of these Arctic regions, but bottom currents dominate in the slope region, forming drift deposits.

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1. Introduction

The paleoclimatic history of the western Arctic is still elusive due to the lack of high-resolution sedimentary records. Leg 1 of the Healy–Oden Trans-Arctic Expedition (HOTRAX) of 2005 recovered eight piston cores with accompanying trigger gravity cores and six multi-core stations nearby each piston core from which seven core tubes and piston cores with accompanying trigger gravity cores and six multi.

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early Holocene and even today (McManus et al., 1969; Nelson and Creager, 1977; Naidu and Mowatt, 1983; Keigwin et al., 2006; Ortiz et al., in this issue). Sediment supply to the shelf north of Alaska and Canada is dominated by the Mackenzie River, which provides $1.25 × 10^8$ t a$^{-1}$ of mostly fine-grained sediment and other rivers, which provide another $1.5 × 10^6$ t a$^{-1}$, and the erosion of coastal bluffs supplies $5.62 × 10^5$ t a$^{-1}$ (Hill et al., 1991). The net transport of the Mackenzie plume is generally eastward away from Alaska and then northward into the Canada Basin (Macdonald and Carmack, 1991; Hwang et al., 2008).

The shelf and slope north of Alaska is at the apex of several currents that can transport and deposit sediment under favorable conditions (Weingartner et al., 1998, 2005; Woodgate et al., 2005a). The Alaskan Coastal Current moves north along the western coast of Alaska (Fig. 1). Although some of this current mixes with waters in the central Chukchi Sea, a significant part of the flow turns east near Pt. Barrow with near bottom mean velocities of about 5–15 cm/s and maximum velocities of 50 cm/s, with a tidal constituent of around 10 cm/s near the head of the Barrow Canyon (Woodgate et al., 2005a; Weingartner et al., 2005). Within the canyon, flows are generally stronger with means of 23 cm/s and up to 95 cm/s episodically (Weingartner et al., 2005). Also there can be strong up-canyon flows or upwelling flows (Mountain et al., 1976; Aagaard and Roach, 1990) and there is evidence of the upper layers of the Atlantic Water being advected up Barrow (and other) canyons, to mix on the Chukchi shelf before returning to their density layer in the Arctic Ocean (Woodgate et al., 2005b).

The current structure east of Pt. Barrow is not entirely clear. There is a generally eastward coastal “shelf break” jet (Aagaard, 1984; Pickart et al., 2005), which is variable, but of comparable strength to the Alaskan Coastal Current in the Chukchi Sea, and is fed at least in part by the Alaskan Coastal Current (Fig. 1A). Beyond about the 50 m contour on the shelf and extending to the base of the continental slope the dominant flow appears to be the eastward Beaufort Undercurrent, which is ~10 cm/s becoming stronger apparently with depth (Aagaard, 1984). The relationship between these currents is unclear. The coastal/shelf break jet is variable, strongly influenced by the local wind, and shows significant seasonal variability (Pickart, 2004). The Beaufort Undercurrent is less documented – the upper portions likely are strongly influenced by wind, while the deeper portion may be part of a basin-wide circulation and thus not locally driven (Aagaard, 1984). Both of these currents flow generally in the opposite direction to the surface Beaufort Gyre circulation, which is primarily anticyclonic but variable and strongly wind-influenced (Proshutinsky and Johnson, 1997). West of Barrow, the flow structure is likely similar. The shallower part of the flow is fed by Pacific waters from Herald Canyon, while the deeper part of the flow is likely part of the basin-wide circulation of Atlantic Water. The behavior of these typically contour-following currents in the vicinity of canyons is not well known.

All along the Chukchi slope eddies spin off the eastward flowing coastal current and move offshore into the Canada Basin (Aagaard et al., 1985; D’Asaro, 1988; Plueddemann et al., 1998; Pickart, 2004). These as well as brine enriched density flows from sea ice formation can move sediment offshore (Weingartner et al., 2005) and especially the density flows can transport sediment down the slope (Williams et al., 2008). There is also likely turbulent mixing (probably wind-driven) associated with the upwelling of upper Atlantic water up canyons onto the shelf (Woodgate et al., 2005b). While this can introduce coastal sediment from the east to the Barrow Canyon area, this longshore current does not directly impact on the core sites in this study, which are much farther offshore.

The winter coastal or longshore drift is generally negligible along the northern Alaska coast (Aagaard, 1984; Reimnitz et al., 1988). During summer storms, this longshore drift can increase to 50 cm/s in a westward direction building spits to the west (Short et al., 1974; Hill and Nadeau, 1989).

In addition, melt-out from sea ice contributes an unknown volume of sediment to the nearshore and as far offshore as the summer ice front, several hundred kilometers seaward. Occasional storm waves and associated currents during ice-free intervals can also move sediment offshore. To summarize, typical local current activity is likely in the range of 5–20 cm/s and on rare occasions during storm events or in focused currents can reach values as high as 50–100 cm/s. Flows of this magnitude are capable of transporting sediment in the clay through sand sizes and into the gravel sizes, while most of these currents should be capable of initiating sediment transport in even cohesive silt and sand size sediments typical of the mid to outer shelf areas.

The sediments of the Chukchi–Alaskan shelf are sandy muds to muddy sands with spotty occurrences of sand and gravel (Barnes and Reimnitz, 1974; Reimnitz et al., 1998). The nearshore (~30 m isobath) is highly variable with mean sizes between 16 and 250 μm that are moderately to well sorted. The central shelf and slope north of Alaska contain fairly well-sorted, fine-grained silt (4–45 μm) but the outer shelf is poorly sorted with patchy gravelly-mud to muddy gravel (Barnes and Reimnitz, 1974). The coarse nature of these patches of sediment here are either palimpsest deposits from a time of lower sea level or due to localized currents that sweep the fines sediments away. The continental slope consists primarily of silty muds except in the channel axes of canyons where coarser sediments often occur.

### 3. Materials and methods

The HOTRAX cores studied in this paper are designated HLY0501-JPC5, -JPC8 (Darby et al., 2005), and HLY0203-JPC16 (Keigwin et al., 2006) herein referred to as JPC5, 8, and 16, respectively (Fig. 1). All of these cores are missing sediment in the core tops due to bypassing near the surface during core barrel entry. Trigger cores for JPC5 and 8 (TC5 & 8) or a multicore (MC14) for JPC16 are used to obtain the uppermost sediment sections. These surface sediments are correlated to the piston cores using radiocarbon dates, core logs (such as density and reflectance logs), and mineral abundance peaks. The estimated offset is 51 cm in JPC8 and TC8, 75 cm in JPC5 and TC5, and 25 cm in JPC16 and MC14 (Table 1). The Holocene sediment of these cores is compared to the Holocene in several box-cores obtained during the Arctic Ocean Section expedition (AOS94) (Darby et al., 1997; Poore et al., 1999). Two cores are from the Mendeleev Ridge (94BC12 and 94BC19, henceforth BC12 and BC19, Fig. 1). One is from the Podvodniki Basin (BC21; formerly Wrangel Abyssal Plain) between the two major ridge systems in the Amerasian half of the Arctic; and two cores (BC25 and BC28) are from the Lomonosov Ridge. An added tube from the same box core as BC28 (BC28-B) is included for replication. In order to assess the role of sea ice, 34 sediment samples collected from modern sea ice sampled at 24 locations across the Arctic were analyzed (Fig. 1). In 2005, twenty-nine samples were collected during the HOTRAX expedition, including 15 from north of Alaska (Fig. 1a). In 2007, 16 samples were collected in the central Arctic mostly near the North Pole, 12 samples at six sites between Fram Strait and the North Pole during the Lomonosov Ridge off Greenland expedition (LOMROG), and four samples at two sites from the mouth of M’Clure Strait in 2007 (Fig. 1).

Radiocarbon age dating by Accelerator Mass Spectrometry (AMS) was performed on shelf material, mostly mollusks and benthic foraminifers from cores JPC 5, 8, and 16 (Table 1). The ages were corrected for marine reservoir effects and stable carbon isotope fluctuations with CALIB5.0.2 (Stuiver et al., 2005) using ΔR values of 0 and 460 for comparison because the magnitude of the Arctic Ocean reservoir effect is poorly constrained (Dyke et al., 1996; Bauck et al., 2008).
2001a,b). The higher $\Delta R$ is favored by studies of mollusks from the near shore or coastal environment near Barrow, Alaska (McNeely et al., 2006) but there is no consensus on samples from farther off shore in the Amerasian half of the Arctic Ocean (Barletta et al., 2008). Several dates in the lower several meters of JPC16 were close to 40 ka and possibly transported offshore from older deposits (Table 1). No lithologic change is observed above this anomalously old interval and the texture also remains unchanged, leading us to suspect the old ages here as transported probably by ice-rafting during the waning stages of the last deglaciation. Only the uppermost unit composed of soft, bioturbated, olive-gray mud is investigated here. This unit is the Holocene based on radiocarbon dating discussed later. Most box cores from the central Arctic Ocean were sampled so as to include the fine-grained, uppermost interval (typically 0–10 cm), which has abundant foraminiferal and was identified as Holocene (Darby et al., 1997). These cores were selected to represent a geographic distribution across the central Arctic (Fig. 1). Paleomagnetic intensity changes were used to supplement radiocarbon dates throughout JPC16 (Table 1).

Samples, more or less evenly spaced down each core, were wet-sieved after thorough dispersal by sonication. Sieving was at 250, 63, and 45 $\mu$m and detailed size analyses using a Malvern Mastersizer 2000 laser particle size analyzer (Syvitsky, 1991; O’Neal et al., 1998) were performed on the $<45 \mu$m size fractions in order to remove the highly variable coarse ice-rafted debris (IRD). While most sedimentologic studies in the Arctic Ocean focus on the coarse IRD fractions, the finer fractions can provide valuable insight into the deposition of the most abundant fraction—the clay to silt-sized fraction. We set the Mastersizer to measure irregularly shaped particles of opaque quartz rather than idealized spheres. Due to the fact that some of the grains sieved through a 45 $\mu$m sieve were probably elongated, the laser analysis measures a small percentage of grains larger than this size. The size distributions were analyzed by principal component analyses (PCA).

4. Results

4.1. Sedimentation rates

Cores from areas of fine sediment deposition on the shelf and slope north of Barrow, Alaska, have very high sedimentation rates for the Holocene that exceed 1.5 m/kyr (Table 1). Core JPC8 is from a depositional lens at 90 m water depth on the shelf along the western

Table 1

<table>
<thead>
<tr>
<th>Core type</th>
<th>Depth (cm)</th>
<th>Adj Depth</th>
<th>Material</th>
<th>Wt. (mg)</th>
<th>14C age (BP)</th>
<th>CALIB5.0.2 del R = 400 (Kyr BP)</th>
<th>Age Interval for Sed. Rates</th>
<th>Sed. Rates (cm/Kyr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HLY0501-JPC8 &amp; TCR</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>JPC</td>
<td>51</td>
<td>102</td>
<td>Bivalve Macoma</td>
<td>72</td>
<td>3216±37</td>
<td>2.486</td>
<td>0.2–5.0</td>
<td>40.4</td>
</tr>
<tr>
<td>JPC</td>
<td>133</td>
<td>181</td>
<td>Bivalve Asteria</td>
<td>185</td>
<td>4590±30</td>
<td>4.187</td>
<td>2.5–4.2</td>
<td>46.4</td>
</tr>
<tr>
<td>trigger core</td>
<td>185</td>
<td></td>
<td>Bivalve Asteria</td>
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<td>4591±42</td>
<td>4.192</td>
<td>2.5–4.2</td>
<td>49.2</td>
</tr>
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<td>JPC</td>
<td>327</td>
<td>378</td>
<td>Bivalve Nuculana (fragments)</td>
<td>34</td>
<td>5210±30</td>
<td>4.973</td>
<td>4.2–5.0</td>
<td>245.8</td>
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<tr>
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<td>510</td>
<td>561</td>
<td>Bivalve Macoma (fragment)</td>
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<td>789</td>
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<td>5.1–5.9</td>
<td>354.1</td>
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<tr>
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<td>851</td>
<td>902</td>
<td>Undist. bivalve fragments</td>
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<td>6.065</td>
<td>5.9–6.1</td>
<td>446.0</td>
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<tr>
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<td>1115</td>
<td>1166</td>
<td>Undist. bivalve fragments</td>
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<td>7265±35</td>
<td>7.334</td>
<td>6.1–7.3</td>
<td>208.0</td>
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<tr>
<td>JPC</td>
<td>1115</td>
<td>1166</td>
<td>Undist. bivalve fragments</td>
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<td>7285±35</td>
<td>7.348</td>
<td>6.1–7.3</td>
<td>205.8</td>
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<tr>
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<td>1150</td>
<td>1201</td>
<td>Gastropoda*</td>
<td>210</td>
<td>7760±51</td>
<td>7.765</td>
<td>7.3–7.8</td>
<td>83.9</td>
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<tr>
<td>JPC</td>
<td>1153</td>
<td>1204</td>
<td>Undist. bivalve fragments</td>
<td>460</td>
<td>7415±35</td>
<td>7.467</td>
<td>7.3–7.8</td>
<td>319.3</td>
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<td>HLY0501-JPC5</td>
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</tr>
<tr>
<td>JPC</td>
<td>37</td>
<td>112</td>
<td>Bivalve- Thyasira</td>
<td>1930±45</td>
<td>1.011</td>
<td>0.1–1.0</td>
<td>104.2</td>
<td></td>
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<td>JPC</td>
<td>484</td>
<td>559</td>
<td>Bivalve- Yoldia</td>
<td>4465±40</td>
<td>4.006</td>
<td>1.3–3.9</td>
<td>161.3</td>
<td></td>
</tr>
<tr>
<td>JPC</td>
<td>560.5</td>
<td>644.5</td>
<td>Bivalve- Thyasira</td>
<td>4820±70</td>
<td>4.510</td>
<td>3.9–4.4</td>
<td>161.3</td>
<td></td>
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<td>688.5</td>
<td>764.5</td>
<td>Bivalve- Yoldia</td>
<td>5220±40</td>
<td>5.007</td>
<td>4.4–5.1</td>
<td>161.3</td>
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</tr>
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<td>800</td>
<td>875</td>
<td>Bivalve- Portlandia</td>
<td>5885±40</td>
<td>5.804</td>
<td>5.1–5.8</td>
<td>161.3</td>
<td></td>
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<td>JPC</td>
<td>880.5</td>
<td>955.5</td>
<td>Bivalve- Portlandia + Thyasira</td>
<td>6395±45</td>
<td>6.343</td>
<td>5.8–6.3</td>
<td>161.3</td>
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<td>HLY0203-JPC16 &amp; MCI14D</td>
<td>0.5</td>
<td>6.5</td>
<td>Thyasira fragment*</td>
<td>10</td>
<td>527±33</td>
<td>0.05</td>
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<td>Multi-core</td>
<td>52</td>
<td>52</td>
<td>Thyasira fragment</td>
<td>9</td>
<td>1497±34</td>
<td>0.583</td>
<td>0.05–0.543</td>
<td>40.30</td>
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<td>240</td>
<td>265</td>
<td>Bivalve</td>
<td>2110±45</td>
<td>2.125</td>
<td>0.6–1.2</td>
<td>374.6</td>
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<td>1008</td>
<td>1033</td>
<td>Thyasira fragment</td>
<td>4.5</td>
<td>4529±41</td>
<td>4.106</td>
<td>1.2–4.1</td>
<td>264.7</td>
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<td>1055</td>
<td>1081</td>
<td>Undist. bivalve fragment*</td>
<td>5.2</td>
<td>2314±34</td>
<td>5.090</td>
<td>4.1–6.0</td>
<td>162.8</td>
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<td>1331</td>
<td>1356</td>
<td>N. juradonica</td>
<td>6</td>
<td>750±40</td>
<td>6.090</td>
<td>4.1–6.0</td>
<td>162.8</td>
</tr>
<tr>
<td>JPC</td>
<td>1331</td>
<td>1356</td>
<td>N. pachyca left</td>
<td>−4.4</td>
<td>7000±40</td>
<td>5.990</td>
<td>4.1–6.0</td>
<td>171.4</td>
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<tr>
<td>JPC</td>
<td>1652</td>
<td>1677</td>
<td>Thyasira fragment*</td>
<td>2.8</td>
<td>39,800</td>
<td>4412±4110</td>
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<tr>
<td>JPC</td>
<td>1964.5</td>
<td>1989.5</td>
<td>Benthic forams*</td>
<td>2.8</td>
<td>39,800</td>
<td>4412±4110</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Not used in aged model.

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edge of the Barrow Canyon. It has exceptionally high sedimentation rates of 202–1109 cm/kyr from 8 to 4 ka, especially around 5 ka (average of 348 cm/kyr). Then the rate decreases dramatically to 37 cm/kyr at 4 ka, although still high by marine standards. This is the most variable rate of the cores studied. JPC5 from the upper slope (415 m) has sedimentation rates that are much more consistent throughout the Holocene with rates of 159 cm/kyr or 161 cm/kyr depending on the age model used (Fig. 2).

JPC16 is farther down the slope in 1300 m water depth and is located on the east side of Barrow Canyon, nearly on the divide between this and a smaller canyon immediately east. The dated part of the core, ca. 0–8 ka, has a higher average sedimentation rate of 234 cm/kyr than core JPC5. There are two 14C calendar ages in JPC16 at 32 and 1081 cm that are much older than the paleomagnetic intensity ages for these depths (Table 1). In addition the two 14C ages below 6 ka (1356 cm) are both older than 39 ka despite the lack of any lithologic change or apparent hiatus. The benthic shells used for these older dates are suspected of being transported and thus they and the dates at 32 and 1081 are not used in the age model (Fig. 2). The paleointensity data for this core shows several events that can be correlated to other independently dated paleointensity records from high latitudes (Fig. 3; Lund and Schwartz, 1999; Lund, 2007; Lise-Pronovost et al., in this issue). The dates derived from the paleomagnetic events produce a near linear fit that merges with the radiocarbon age/depth curve below 2–4 ka where a $\Delta R$ of 460 is applied but indicates that the radiocarbon dates are even older than a correction of this magnitude at 4 and 6 ka (Fig. 2). The average sedimentation rate for this core using both calendar 14C Ages based on a $\Delta R$ = 460 and the paleomagnetic dated intensity peaks is 234 cm/kyr (Fig. 2).

In summary, the high sedimentation rates north of Alaska vary from about 40–1200 cm/kyr with average rates between 150 and 313 cm/kyr. There is no indication of rapid sedimentation events such as turbidites, hyperpycnites (sediment-laden fluvial discharges), mass flows, or storm deposits in this core. Correlations to other HOTRAX cores from the upper slope farther to the west indicate similar to lower deposition rates in the Holocene suggesting patchiness to the areas of highest sedimentation (Darby et al., 2005). USGS core P1-92AR-P1...
sediment in BC25 is unimodal at 7.5 μm (henceforth P1, Fig. 1) located on the shelf edge near the Northwind Ridge in 205 m water depth shows somewhat lower sedimentation rates of 30–44 cm/kyr except for an interval of low sedimentation (5 cm/kyr) between 2 and 6 ka (Darby et al., 2001; Darby and Bischof, 2004). Between the HOTRAX core sites and the Mackenzie River, core P1–89AR–P45 (henceforth P45, Fig. 1) from the upper slope (405 m) is missing the uppermost 7 ka sediment interval, possibly due to recent mass movement like slumping, but has early Holocene sedimentation rates of about 110 cm/kyr (Andrews and Dunhill, 2004).

In contrast to the high sedimentation rates found in many sites along the Chukchi–Alaskan margin, sedimentation rates in the central Arctic ridges and basins where turbidites are infrequent are usually less than 2 cm/kyr (Darby et al., 1997; Bachman et al., 2004). Sedimentation rates increase slightly from the western Arctic or Amerasian half to the eastern, Eurasian half from about 1 cm/kyr on the Mendeleev–Alpha Ridge to nearly 2 cm/kyr on the Lomonosov Ridge (Jakobsson et al., 2001).

4.2. Sediment texture

The average Holocene size frequency plots for each of the three cores from the Chukchi–Alaskan Margin (JPC5, 8, & 16) are unimodal, poorly sorted, and nearly symmetrically-skewed, with mean sizes between 5 and 8 μm (7–8 φ; very fine silt; Folk, 1974; Fig. 4). JPC8, on the continental shelf, is slightly coarser with just slightly more fineskewness than the other two cores. The mean sizes and the size frequency distributions of these cores are similar to the average of 34-highly variable sea ice sediment samples collected from the Beaufort Sea off North America and across the central Arctic (Figs. 1 and 5). The finest sediment is found in Holocene deposits in the box cores from the central Arctic. Except for less material between about 5–7 φ (8–31 μm) in these central Arctic cores, their average size distribution is similar to that of sea ice sediment and JPC5 and 16. The sea ice sediment is more variable in mean size than any of the cores (Fig. 4).

In general, there is better sorting as the mean grain size decreases (Fig. 5) for all samples from the cores and sea ice. The slightly coarser nature of JPC16 compared to JPC5 is apparent as the samples from both cores plot parallel each other. JPC8 is coarser than either of these cores and more poorly sorted. Holocenic sediment from the central Arctic box cores (except for BC25) is finer than any of the core sediments from the Chukchi–Alaskan margin. Core BC25 from the Lomonosov Ridge is coarser and more poorly sorted than the other central Arctic cores. This is the only central Arctic core with a bimodal size distribution in the upper 4 cm (Fig. 6). One mode is at 8.3 μm (3 μm) and the coarser at 5.3 μm (25 μm). The preceding Holocene sediment in BC25 is unimodal at 7.5 μm (5.5 μm). Of significance is the distribution of sea ice samples compared to all the core sediments in the mean size versus sorting plot (Fig. 5). These sea ice samples overlap all of the core samples, particularly the central Arctic core samples.

The three Chukchi–Alaskan margin cores show fairly uniform mean grain-size down-core in the Holocene unit (Fig. 7). While there is slightly more variability in the mean size in the lowermost 2 m of JPC16 and 1 m of JPC5, there is no correlation between the mean size of these <45 μm size fractions and the percentage of coarser than 63 μm IRD in JPC16 (Fig. 7). In fact, the mean size of the <45 μm size distribution is finer in several samples within the lowermost 2 m of JPC16 where several samples contain more than 4–16% >63 μm. Above these intervals the mean size remains mostly unchanged with a slight trend to increase in size near the top of JPC16 and decrease in JPC5. This independence of the >63 μm and the <45 μm size fractions in JPC16 suggests that the coarser material is added as a separate process to that which deposits the bulk of the sediment finer than 45 μm.

In order to determine the various processes that may be involved in depositing these Arctic sediments we employed varimax-rotated

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Principal Component Analysis (PCA) using the correlation matrix of grain size spectra for grain sizes ranging from 0.3 to 60 μm as the input matrix. This method allows us to separate out orthogonal modes (independent grain-size spectral components), which we relate to potential input functions. As an initial step we performed separate PCA analysis on the sea ice, central Arctic, and Chukchi–Alaskan margin data sets. This allowed us to test the null hypothesis that different processes control the sedimentation in these three data sets. This analysis indicated that while the sea ice and central Arctic sediment had the same data structure as seen by nearly identical factor loading models, the Chukchi–Alaskan margin sediment was somewhat different in that it contained a coarse silt factor component absent in the central Arctic Holocene sediment and sea ice. We thus combined the sea ice and central Arctic data sets, and analyzed these separately from the Chukchi–Alaskan margin data set for the analysis we present.

The combined sea ice and central Arctic sediment PCA analysis is based on a total of 91 samples. We refer to this as the central Arctic PCA. (Fig. 8). Only the rank of factors changes when the PCA is run separately on the sea ice and central Arctic Holocene size distributions, probably due to the smaller sample size for these separate runs. The three leading modes in the central Arctic PCA explain 98.7% of the variance in the data set. The model communality as a function of grain size was equally high with values that averaged 0.99±0.011, and a minimum communality of 0.95. The first mode accounts for 35.0% of the variance; the second for 34.8% and the third for 28.9% of the variance. The implication is that the processes generating these modes have a similar impact on the overall variance in the data set.

In the central Arctic PCA, the first loading score has a positive broad peak at 2 μm and an anti-correlated negative peak at 10 μm, the second has a positive peak at 5 μm and strongly anti-correlated loadings at about 45 μm. The third Factor contains a peak at ~0.5 μm and moderately strong negative loadings above 8 μm.

Analysis of the 171 Chukchi–Alaskan margin samples by PCA followed the same procedure as used for the central Arctic samples (Fig. 9). This model also extracts three PCA modes, which account for 98.1% of the variance in the data set. The model communality as a function of grain size for the Chukchi–Alaskan margin PCA was also very high with values that averaged 0.98±0.026, and a minimum communality of 0.86. The first mode accounts for 45.3% of the variance, the second for 38.8% of the variance and the third for 14.1% of the variance in the data set. The leading two modes in the Chukchi–Alaskan margin PCA are very similar in rank and shape to the leading two modes of the central Arctic PCA. The third mode in each of the two PCA models however is distinct (Figs. 8 and 9).

The first two Factors in the two PCA models are similar enough that they very likely arise from the same processes in each region. The leading Factor in the Chukchi–Alaskan margin PCA exhibits stronger factor loadings below 2 μm than the leading factor in the central Arctic PCA. This could arise from a regional difference or may be a function of the small sample size for the Central Arctic data. The coarser, anti-correlated negative peak is anchored near the same grain size (8 vs. 10 μm, respectively). The second Factor in the Chukchi–Alaskan margin PCA has a positive grain size peak similar to that of the second Factor from the central Arctic (4.2 vs. 4.8 μm). The strong anti-correlated negative loadings in Chukchi–Alaskan margin PCA are better resolved as a trough centered on 20 μm, while in the central Arctic PCA there is no indication of decreasing negative loadings with increasing grain size. As with the first Factor, this difference could arise from a regional difference indicating that the sea ice samples are finer than the Chukchi–Alaskan margin samples with respect to this Factor or the difference may represent a bias resulting from the small sample size of the central Arctic data set.

The third mode in the two PCA models is unique to each region. In the central Arctic, this mode represents extremely fine particles, which peak at 0.3 μm. On the Chukchi–Alaskan margin, the third mode represents rare coarse particles, which peak at 53 μm. The only similarity between these two modes is that their inflection points both occur around 6 μm. Alternative PCA models that extracted up to six modes from each data set failed to find the coarse mode in the central Arctic data set, or the fine-mode in the Chukchi–Alaskan margin data set. This indicates a fundamental difference in the grain size spectra for the two sample sets. When all three Factors are considered, the PCA model for the Chukchi–Alaskan margin exhibits a greater variation in grain size than the central Arctic PCA. This is consistent with the observation that the margin sediment is coarser than the central Arctic sediment.

5. Discussion

5.1. Sedimentation rates

The extremely high sedimentation rates in the sediment cores from the shelf and continental slope north of Alaska suggest some form of current deposition. This was previously inferred based on the spatial distribution of sediments—both surface sediment composition...
Fig. 7. Down-core change in mean size (<45 μm) for the Chukchi–Alaskan margin cores. The mean size of the bulk sample (all size fractions) is very similar to that of the <45 μm fraction in JPC5. The same is true for the other cores but is not shown. The coarse IRD (>63 μm sand fraction) in JPC16 is generally less than 2–3% except for a few samples, especially in the lowermost 1.5 m of the core. Note that many of the samples with the largest percentage of sand in JPC16 are the finest mean size.
velocity. The upwelling currents noted earlier might mix with the down-canyon flows to create eddies or decrease net currents thus promoting deposition. Obviously fine-grained sediment is being deposited and not sand so the current regime must be weak enough at the deposition sites to allow clay-size sediment to settle out (Fig. 4). There is no indication that the canyon axis is shifting position due to inflowing from the west. In contrast, the slope is aggrading, forming a 6 km wide, lens-shaped deposit (Fig. 10). This lens is thickest closer to the canyon where the bottom begins to slope into the canyon and is akin to a levee deposit.

JPC5 and JPC16 are also located on the sides of canyons but on the upper and mid-slope areas, respectively (Fig. 1a). These cores were also raised from lens-shaped deposits that are possible levee deposits, although the quality of the CHIRP data does not allow for clear determination as to whether these are accreting or aggrading. Thus the sub-bottom data does not indicate whether deposition is due to a separation of flow at the edge of these canyons due to an eastward geostrophic contour current or a down-canyon density flow that is depositing in back-peedies at suitable locations along the sides of the canyon, or a combination of both.

The shelf area near JPC8 is in the physical regime of a shallow shelf sea and is strongly affected by surface forcing (such as wind) and by the coastal currents that veer offshore and down canyons (Alaskan Coastal Current) or dense brine currents during sea ice formation. In contrast, sites JPC5 and 16 lie deeper, and in the physical regime of mostly along- topography, slope currents (the generally eastward flowing Pacific Waters from Herald and Barrow Canyon, and the Beaufort Undercurrent), although these sites are also influenced by upwelling/downwelling (often wind-related) and dense outflows. Direct wind influence is weaker here, and eddies are more important. It is possible that the deeper Beaufort Undercurrent might be a more constant factor during the Holocene and result in more constant sedimentation rates in the slope regions. The location of JPC5 on the west side of a canyon supports this because the Beaufort Undercurrent flows eastward and would produce a separation of flow as it passes over the canyon margin allowing for deposition. JPC16 is on the east side of Barrow Canyon but still can receive sediment deposited by the Beaufort Undercurrent if most of the sediment is transported in suspension. Separation of flow in a bottom current would seem to be less important on the east side of a canyon under an eastward current. Besides suspension deposition, eddies from down-canyon flows might be important here. Due to Coriolis effects, the dense outflows moving down the canyons along the Chukchi–Alaskan margin favor flow down the east side of the canyon, but the flows are so strong that

**Fig. 8.** a: Factor scores from the combined PCA analysis of the size distributions (<45 μm) for the central Arctic cores and sea ice samples. Factor 1 has a broad peak around 2 μm and negative loadings near 10 μm. Factor 2 peaks at 5 μm and Factor 3 peaks in the finest sizes (<0.5 μm). Communality is high throughout b: Factor scores for the sea ice samples and central Arctic box cores for separate PCA analysis. The order of the factors is different but the three factors from each sample set (box cores and sea ice) are nearly identical.

**Fig. 9.** PCA analysis combined factor scores for the size distributions (<45 μm) of the three Alaskan Margin cores. Note that Factors 1 (anchor ice) and 2 (nepheloid transport) are the same as for the combined sea ice and central Arctic sediment PCA (Fig. 8a). Factor 3 (suspension freezing) is completely opposite of Factor 3 in the central Arctic containing a peak in the coarsest sediment and interpreted as the intermittent suspended load.
ageostrophic dynamics are also important and allow eddies to spin off (D’Asaro, 1988, Pickart et al., 2005), which may deposit sediment along the west side of the canyons or on the divides between canyons such as the location of JPC16. Perhaps both of these factors work to produce the very high sedimentation seen in these cores, but the more consistent sedimentation rates on the slope suggest a greater influence by the Beaufort Undercurrent, which only influences the slope core sites.

5.2. Sedimentation processes

Several processes operate in the Arctic to transport sediment. The most important of these include: suspended and bed load transport by currents, and suspension freezing and anchor ice formation and transport by sea ice (Reimnitz et al., 1987, 1992; Nürnberg et al., 1994). Transport by turbidity currents is important on the abyssal plains (Campbell and Clark, 1977), but our cores were sited to minimize the influence of turbidity currents. The sedimentary structures and textures in our cores also rule out turbidity currents as a significant source of sediment transport at these locations. Likewise, we assume that Aeolian and volcanic input provides minimal influence on the observed grain size spectra at these locations because evidence for either is lacking. The biogenic component is also low overall, except for the enrichment of the sand-size fraction with foraminiferal tests in Holocene sediments in the central Arctic Ocean with very low background lithogenic sedimentation rates.

To interpret these factors, we make use of the fact that many of our samples were collected in the field from sea ice. We surmise that the sediment incorporated into this ice must have been included through either suspension freezing or anchor ice production and we hypothesize that these two processes should have distinct spectral grain size patterns (Reimnitz et al., 1998). Sea ice sediment entrained by suspension freezing should be skewed toward fine-grain sizes and deficient in coarser silt and sand while anchor ice sediment should be more variable in size and sorting reflecting the shelf sediment from which it is obtained. While one process or the other will dominate some of the sea ice samples we collected, others should represent sediment generated by a combination of these processes.

Extreme samples were identified from cross plots of the factor scores and their size distributions, and then transformed to z-scores for comparison with the factor score patterns for each factor. A z-score transformation is appropriate because the factors are based on the correlation structure of the data, which can be determined from the cross product of the z-scores of the data (Kachigan, 1986).

Analysis of sea ice samples with extreme factor scores for the combined central Arctic factor model indicates that the first factor in both the central Arctic and Chukchi-Alaskan margin PCAs has the most positive factor scores for a coarse-grained sample with a fine-grained tail and minimum factor scores for a sample with a unimodal, fine, silt-sized peak (Fig. 11a). This indicates that the first factor is associated with poorly-sorted samples, similar to what might be expected for anchor ice.

The second factor in both the Central Arctic and Chukchi-Alaskan margin PCAs has its most positive factor score for a well-sorted, unimodal sample with a peak in the silt-sized range (Fig. 11b). The most negative factor scores for the second factor in the Central Arctic occur for an extremely coarse sample, which is depleted in fine, silt-size grains. This indicates that the second factor is associated with sediment in the sortable-silt size range.

The third factor in the Central Arctic PCA is inversely correlated with the first factor in the Central Arctic PCA. The same set of sea ice samples that serve as the extreme samples for Factor 1, do so for Factor 3, but with the minimum and maximum samples reversed (Fig. 11a, b). The most positive factor scores for Factor 3 in the Central Arctic are associated with the fine-grained tail of the coarse sample, while the most negative factor scores for Factor 3 in the Central Arctic PCA are associated with the sample that exhibits the unimodal, fine, silt-sized peak. This indicates that the variance associated with the third factor arises from the finest particles in the sea ice and central Arctic core samples and that they are uncorrelated with sediment coarser than silt-sized.

We infer that the first and third Factors in the Central Arctic PCA represent two different sea ice related entrainment processes due to

![Fig. 10. CHIRP (~12 kHz) seismic profile line for the JPC8 site in 85 m water depth. The N-S profile line is shown on Fig. 1a. Note the lens-shaped nature of the uppermost unit (Holocene), which is about 15 m thick at JPC8.](image-url)
to entrain due to their greater mass (McCave, 1984). The fact that this Factor is associated with sea ice samples is probably because it comprises a significant portion of the shelf sediment entrained by both suspension freezing and anchor ice.

Only a few bottom current measurements have been made on the ridges and basins in the western Arctic and these indicate weak currents of 4–6 cm/s on the ridges and only about 1 cm/s or less in the Canada Basin (Hunkins et al., 1969). Based on 3 years of sediment trap data, subsurface nepheloid clouds of fine sediment are transported from the shelf near the Lena River at least as far as the Lomonosov Ridge at 81° N latitude (Fahl and Nöthig, 2007). Similar fine-grain transport probably occurs in the Canada Basin (Hwang et al., 2008). It is uncertain as to whether these transported fine sediments contribute to the 5 μm Factor or the finer, <0.5 μm Factor or both. Our size analyses are of the dispersed sediment. The sediment collected by these sediment traps on the Lomonosov Ridge or in the Canada Basin are aggregated sediment (Fahl and Nöthig, 2007; Hwang et al., 2008). However, these data have not been analyzed using the methods we present in this paper, and thus we cannot be certain how they relate to our central Arctic Factors.

The influence of these three end-member processes can be observed in a plot of Factors 2 and 3 (Fig. 11b). The high degree of variability in size and sorting among the sea ice samples (Fig. 5) is probably due to the fact that anchor ice is not selective and entrains whatever is on the bottom. Yet the locations and water depths in which this process occurs might restrict the size characteristics somewhat because anchor ice occurs in depths of less than ~30–50 m (Reimnitz et al., 1987). Ice attached to the bottom in depths of ~<10 m (nearshore environment) would fall into the category of fast ice and would not usually become part of the drifting pack ice until late in the melt season so that it tends to melt before it can drift far.

The third Factor on the Chukchi–Alaskan margin probably represent a component that is regionally limited to processes on the shelf and slope not involving sea ice since this “coarse” Factor is not seen in the sea ice samples or in the Central Arctic PCA Factors. The lack of associated sedimentary structures lets us rule out mass wasting and turbidity currents as potential mechanisms for transport of this sediment. This third Factor from the PCA of the Chukchi–Alaskan margin sediment is strongly correlated to the coarsest sediment (~53 μm) in the size fraction analyzed and thus likely represents down slope or along slope sediment transport in response to local bottom currents.

The mode of transport can be determined using the Rouse number, \( R_o \), the ratio of the settling velocity to the shear velocity (\( R_o = W_s / \kappa U^* \)) where \( W_s \) is the settling velocity of the grains, \( \kappa \) is the von Karman constant (0.4), and \( U^* \) is the boundary shear velocity. The \( R_o \) has a critical value of >2.5 for bedload transport, 2.5–12 for 50% suspension transport, 1.2–0.8 for 100% intermittent suspension transport, and 0.8 for fully suspended wash load. A calculation of the \( R_o \) for 6.4 μm (mean size in Chukchi–Alaskan margin core sediments) and 53 μm (Factor 3 size, Fig. 9) particles based on the extremes of observed regional current strength (\( U = 5–100 \) cm/s at 8 m off the bottom in sea water of 0 °C and 35 psu), yields values ranging from 0.3 to 0.02 for 6.4 μm particles and 2.7 to 0.1 for 53 μm particles, respectively for 5 and 100 cm/s flows (U), using settling velocities of 0.03 cm/s for 6.4 μm particles and 0.25 cm/s for 53 μm particles. These low Rouse numbers suggests that all of the 6.4 μm particles would be transported by fully developed suspension and 50–100% of 53 μm particles would only be transported by intermittent suspension with these currents (5–100 cm/s). In fact, only sand (5 μm) would be transported as bedload (\( R_o > 2.5 \)) and then only by the lower current velocity (5 cm/s). According to Shields Diagram, a critical \( U^* \) for 6.4 and 53 μm particles is 0.02 and 0.4 cm/s, respectively and this corresponds to 9–10 cm/s currents that would be required to initiate transport assuming no cohesion. This is comparable to the velocity of the Beaufort Undercurrent. Shear velocities (\( U^* \)) for 5 and 100 cm/s currents are 0.3 cm/s (\( U = 5 \) cm/s at 8 m above the bed) and 6 cm/s (\( U = 100 \) cm/s at 8 m above the bed).

![Fig. 11. a: Plot of factor scores 1 and 3 for all the central Arctic core samples and sea ice sediment. Here the coarsest and finest sea ice samples appear to plot together but are not in the same plane as shown by their separation along Factor 2 in Fig. 11b. The sea ice samples extend from the suspension freezing pole to the anchor ice pole, somewhat biased toward the latter, showing perhaps the relative importance of these two processes in sea ice entrainment. b: Plot of Factor 2 and 3 scores for the sea ice and central Arctic core data sets (see Fig. 8). These are the two fine-grained factors (2 μm) and (<0.5 μm). While not plotted, Factor 1 (anchor ice) is shown as a minimum for these other two factors and the three sea ice samples with the coarsest mean sizes and the poorest sorting taken from the end-members in Fig. 5 plot closest to this end-member Factor 1. The three sea ice samples with the finest mean size and best sorting plot closest to the nepheloid or sortable silt Factor maximum. The sea ice samples with >1% sand are shown but do not correspond to any one factor.](image-url)
above bed) assuming a smooth bed (roughness < 1 cm). Coarse silt (45–63 μm) in the Chukchi–Alaskan sediment suggests Kc values between 0.8 and 1.2, the intermittent suspended load at typical velocities for the bottom currents on the shelf and slope here (~10 cm/s). Since this velocity is also the critical entrainment velocity, small changes in these currents control both erosion and deposition of the coarse silt, representing Factor 3. We cannot rule out that this sediment is not transported by other means such as sea ice, but such an interpretation would imply two populations of anchor ice, which seems implausible, as we find no evidence for this in the central Arctic sediment samples or the sea ice sediment samples.

Provenance of the sand fraction in these sediments aids in our interpretation of depositional processes. Based on 22,700 microprobe analyses of Fe oxide mineral grains in the 45–250 μm fraction from the entire length of JPC16, only 10% of these grains are traced to the Chukchi Sea (mostly near Bering Strait) and a mere 1% to the northern Alaskan coast or the Chukchi Shelf (Darby and Bischof, 2004; Keigwin et al., 2006). Previous provenance investigations (see Darby, 2003 for details on the Fe grain provenance technique). Thus, the Coastal Jet and the Alaskan Coastal Current that might transport such grains from the northern Alaskan coast or the Chukchi Sea, respectively (Fig. 1a), only account for 11% of the sand fraction at this core site. Even this sand may get offshore to JPC16 site via sea ice. The remainder of Fe grains in this core are traced to more distant sources (Naidu and Mowatt, 1983). More recent work suggests that the sediment in a core very near JPC5 changes around 6 ka from a more illite and smectite rich clay mix to a more chlorite rich clay mix suggesting a greater input of clay from the Pacific through Bering Strait between ca. 4–6 ka (Ortiz et al., in this issue).

The precise sources of the sand fraction in these sediments aids in our interpretation of depositional processes. Based on 22,700 microprobe analyses of Fe oxide mineral grains in the 45–250 μm fraction from the entire length of JPC16, only 10% of these grains are traced to the Chukchi Sea (mostly near Bering Strait) and a mere 1% to the northern coastal shelf (inside the 50 m isobath) east of Barrow Canyon (Table 2; see Darby, 2003 for details on the Fe grain provenance technique). Thus, the Coastal Jet and the Alaskan Coastal Current that might transport such grains from the northern Alaskan coast or the Chukchi Sea, respectively (Fig. 1a), only account for 11% of the sand fraction at this core site. Even this sand may get offshore to JPC16 site via sea ice. The remainder of Fe grains in this core are traced to more distant sources including northern Canada and the Russian shelves indicating that sea ice is the primary transport agent for the small amount of sand in this core.

The clay mineral content of the <2 μm size fraction on the Chukchi–Alaskan margin has been related to the Yukon River clays as well as more local sources (Naidu and Mowatt, 1983). More recent work suggests that the sediment in a core very near JPC5 changes around 6 ka from a more illite and smectite rich clay mix to a more chlorite rich clay mix suggesting a greater input of clay from the Pacific through Bering Strait between ca. 4–6 ka (Ortiz et al., in this issue). The precise sources of the fine fraction in this shelf area are difficult to determine but available evidence including both the clay mineralogy and Fe grain provenance suggests multiple sources.

5.3. The role of sea ice

If sea ice-rafting played a major role in sedimentation at the Chukchi–Alaskan margin core sites, then we should expect to see the same size PCA Factors for the sea ice and these margin cores. Factors 1 and 2 are nearly identical but because the third Factors are so different between the central and margin areas, this suggests that different processes or sources of sediment are involved. Sea ice entrains sediment from the inner shelf (Reimnitz et al., 1998). The three Chukchi–Alaskan margin cores probably receive sediment eroded from these general morphologic areas. However, sea ice can entrain sediment from many distant shelves including those in Siberia (Nürnberg et al., 1994; Darby, 2003) and the sea ice-rafted component in JPC16 bears this out (Table 2). We assume that the clays transported by sea ice are derived from the same source as the sand component. The Chukchi–Alaskan margin cores most likely contain sediment reworked from nearby on this shelf or the Chukchi Shelf (Darby and Bischof, 2004; Keigwin et al., 2006). Previous provenance investigations of sea ice indicates that most Arctic sea ice sediment is derived from the Siberian shelves such as the Laptev Sea and from northern Canadian shelves and only minor amounts originate from the Chukchi–Alaskan shelf (Pfirman et al., 1997; Eicken et al., 1997, 2000; Darby, 2003; Eicken et al., 2005). Thus similar sources for the central Arctic sea ice and the Chukchi–Alaskan margin cores are difficult to establish.

The similarity of the PCA factors thought to be due to sea ice in the central Arctic and margin core areas must indicate that at least some portion of the shelf sediment is originally deposited from melting sea ice.
5.4. Holocene changes in grain size factors

The sediment texture in the central Arctic appears to have changed very little with respect to the anchor ice Factor, with only a slight increase towards the recent (Fig. 12a). This might be related to the cooling during the last 6 kyr in the western Arctic where the Holocene thermal maximum terminated earlier (at 8–10 ka) than in the Greenland and Canadian Archipelago region (Moore et al., 2001; Kaufman et al., 2004). We assume that colder temperatures would favor anchor ice formation, but little is known about the factors critical for this process on shelves in water depths of 10–30 m. Like the anchor ice Factor, the suspension freezing Factor (Factor 3, associated with abundant fine sediments, <0.5 μm; Fig. 12b) is unchanged during the Holocene with a possible slight decline toward the recent. The Holocene temperature and more marine permafrost that would promote sea ice entrainment analyzed further implies that the processes of sea ice entrainment produce similar sediment textures, regardless of where they occur (Reimnitz et al., 1998; Stierle and Eicken, 2002). This is especially true for suspension freezing.

6. Conclusions

The areas of high sedimentation on the outer shelf and slope north of Alaska appear to be localized and associated with canyons. Many of these cores are located on the Lomonosov Ridge in 2125 m water depth in a small intra-ridge basin, the walls of which are swept by currents (Björk et al., 2007). This might explain the unique texture of BC25 (igs. 5, 6, 11a and 12). The Factor related to the winnowed silt fraction, which we liken to nepheloid transport by weak currents, is mostly negative in BC25 suggesting that sortable silt is removed (Fig. 12b).

The Chukchi–Alaskan margin cores show a similar flat trend after 7.5 ka in the anchor ice Factor as the central Arctic cores, except that JPC16 and 8 have much lower Factor 1 scores than JPC5, which is more similar in this Factor loading to the central cores (Fig. 13a). There is a marked increase in the anchor ice Factor scores prior to 7.5 ka in JPC16 (ka) and JPC8. This higher component of anchor ice in the early Holocene in JPC8 & 16 and possibly in JPC5 might reflect colder bottom temperatures and more marine permafrost that would promote anchor ice formation prior to 7.5 ka. The nepheloid or sortable silt Factor (Factor 2) in these cores also shows JPC8 has much lower loadings for this Factor after 7.5 ka (Fig. 13b). Here near the head of the Barrow Canyon where currents average about 14 cm/s and can reach nearly 100 cm/s, current transport and deposition dominates that of sea ice as shown by the somewhat higher loadings for the third and coarsest Factor (Fig. 13c). The fact that all of the margin cores show this Factor 3 to be unchanged and nearly equal in all three cores throughout the Holocene suggests that currents have played an important role in all of these core sites throughout the Holocene.

Only the third Factor in the margin cores, coarse silt, is not present in the central Arctic or sea ice sediment. This coarse silt fraction around 53 μm must be enhanced in the margin areas by winnowing because this third Factor is not present in sea ice or the central Arctic cores. The presence of the first two Factors in all sea ice sediment analyzed further implies that the processes of sea ice entrainment produce similar sediment textures, regardless of where they occur (Reimnitz et al., 1998; Stierle and Eicken, 2002). This is especially true for suspension freezing.

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The areas of high sedimentation on the outer shelf and slope north of Alaska appear to be localized and associated with canyons. Many of these cores are located on the edge of a canyon (possibly levee deposits) and thus the rapid sedimentation might be associated with eddies developed from down-canyon flows or to eastward flows along the shelf or slope such as the Beaufort Undercurrent (Aagaard, 1984). Regardless of which current system is involved, the sediment size distributions in these areas throughout the Holocene indicate a current is involved. Nearly all of the sediment is deposited from intermittent suspension or fully developed suspension and not bed load according to the Rouse numbers for the mean size and coarsest particle sizes associated with PCA factors from the Chukchi–Alaskan margin sediments, Global Planet. Change (2009), doi:10.1016/j.gloplacha.2009.02.007
margin area sediments (6.4 and 53 μm, respectively) and the range of currents likely encountered. The sand fraction in these margin cores is nearly all deposited from sea ice melt-out even though shelf currents are strong enough to transport this size fraction because only about 11% of the sand is derived from potential sources of entrainment by these currents. Despite the dominance by current deposition for the clay and silt fractions, sea ice contributes enough sediment to impart a unique textural component as indicated by the similar PCA factors with sea ice sediment collected from Arctic Ocean pack ice.

The central Arctic Holocene sediment texture appears to have a strong association with sea ice sediment. This is not surprising in light of the high transport capability of sea ice (Eicken et al., 1997). The fact that the anchor ice factor accounts for the largest amount of variance suggests that it is more important than previously thought. The most surprising result of this study is the fact that Holocene sediment in both the margin and central Arctic areas share two of the PCA texture factors and that these factors are found in sea ice sediment. Because sea ice might entrain this sediment from many different shelves, the similarity is probably not source-related but process-related. Thus sea ice entrains similar size sediment regardless of the source and deposits it throughout the Arctic, including other shelves far removed from the source.

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