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Spatial and temporal Arctic Ocean depositional regimes: a key to the evolution of ice drift and current patterns

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ABSTRACT

Sediment physical properties measured in cores from all the major ridges and plateaus in the central Arctic Ocean were studied in order to analyze the spatial and temporal consistency of sediment depositional regimes during the Quaternary. In total, six physiographically distinct areas are outlined. In five of these, cores can be correlated over large distances through characteristic patterns in sediment physical properties. These areas are (1) the southern Mendelev Ridge, (2) the northern Mendelev Ridge and Alpha Ridge, (3) the Lomonosov Ridge, (4) the Morris Jesup Rise and (5) the Yermak Plateau. Averaged downhole patterns in magnetic susceptibility, bulk density and lithostratigraphy were compiled to establish a composite stratigraphy for each area. In the sixth physiographic area, the Chukchi Borderland, repeated ice-grounding during recent glacial periods complicates the stratigraphy and prevents the compilation of a composite stratigraphy using the studied material. By utilizing published age models for the studied cores we are able to show that the northern Mendelev Ridge and Alpha Ridge have the lowest average late Quaternary sedimentation rates, while intermediate sedimentation rates prevail on the southern Mendelev Ridge and the Morris Jesup Rise. The second highest sedimentation rate is observed on the Lomonosov Ridge, whereas the average sedimentation rate on the Yermak Plateau is more than twice as high. The close correlation of physical properties within each area suggests uniform variations in sediment transport through time, at least throughout the later part of the Quaternary. The unique stratigraphic characteristics within each area is the product of similar past depositional regimes and are key for furthering our understanding of the evolution of ice drift and current patterns in the central Arctic Ocean.

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1. Introduction

The Arctic Ocean is characterized by comparatively large physiographic variations, in spite of being the smallest of the world oceans. The surrounding wide and shallow continental shelves comprise 53% of the Arctic Ocean's area while the central deep basins are bisected by submarine ridges that account for 16% of the total area compared to only 3% for the other world oceans (Menard and Smith, 1966; Jakobsson et al., 2003b). The rivers on the surrounding continents drain huge geologically distinct areas that deliver varying quantities and compositions of suspended sediment to the shelves and basins (Peterson et al., 2002). These facts argue for large spatial heterogeneity in the composition of sediments within the Arctic Ocean and partially explain why a lithostratigraphic framework for the entire basin has not been possible to establish to-date. For example, Clark et al. (1980)

presented a general lithostratigraphy valid for the Alpha Ridge, in the Amerasian side of the Arctic Ocean, which is not possible to recognize in cores from the Lomonosov Ridge and the Eurasian side (Sellén et al., 2008).

On the other hand, sediment cores from certain regions can be correlated over large distances (e.g. Clark et al., 1980; Jakobsson et al., 2001; Backman et al., 2004; Spielhagen et al., 2004; O'Regan et al., 2008; Adler et al., 2009), sometimes down to sub-decimeter scale changes in the physical properties that reflect subtle variations in sediment composition and grain size (Phillips and Grantz, 2001; O'Regan et al., 2008). Despite the physiographic variability, the modern sea ice cover acts as a homogenizing agent, re-distributing sediments entrained on the margins throughout the basin. This reasoning leads to speculation that sediments deposited under similar surface circulation and subsurface current regimes should display similar compositional and grain size characteristics, which are the main controls on physical property data, namely bulk density, compressional wave velocity and magnetic susceptibility (Hounslow and Maher, 1999; Moran et al., 2007).

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At the cm/ka temporal resolution of most central Arctic Ocean cores collected away from the margins, short-term (decadal through millennial) variations in circulation or sea ice are not preserved in the sediments. However, on longer timescales encompassing glacial/interglacial cycles, changes in sediment entrainment and distribution should be distinguished. In the last 10–15 years there has been a dramatic increase in the number of cores collected from icebreaker expeditions in the Arctic Ocean. Together with advances in rapid non-destructive, high-resolution measurement techniques (principally of physical properties) we are now able to better characterize and compare lithostratigraphic changes in marine sediments and resolve how uniform they are in different regions of the basin.

The ability to recognize and correlate synchronous changes in the physical properties of sediments is especially important in the Arctic due to existing difficulties in dating Arctic Ocean sediments (Backman et al., 2004; Sellén et al., 2009). Age models for Arctic sediments are frequently revised and lack the resolution and certainty attainable in marine sediments from other ocean basins. This poses a particular challenge when two independently dated records are used to interpret spatial and temporal changes in paleoceanographic conditions.

The 2005 Healy-Oden Trans-Arctic Expedition (HOTRAX) recovered piston cores and multi-cores along a pan-Arctic transect extending from the Alaskan to Svalbard margin (Fig. 1) (Darby et al., 2005). Here, we study all piston cores retrieved along the HOTRAX transect together with four cores taken from the Morris Jesup Rise during the Lomonosov Ridge off Greenland (LOMROG) 2007 expedition (Jakobsson et al., 2008b). The objective is to characterize the spatial and temporal variability in depositional patterns using physical property data and through establishing regional stratigraphic frameworks to address the paleoceanographic mechanisms behind different Quaternary sedimentation regimes.

2. Geologic and oceanographic overview

2.1. Physiography

Most coring efforts have focused on the recovery of material from prominent bathymetric ridges that bisect, or protrude into, the deeper basins of the Arctic Ocean; primarily the Alpha and Mendeleev ridges, Chukchi Borderland, Lomonosov Ridge, Morris Jesup Rise and Yermak Plateau (Fig. 1). The length of these cores, generally <10 m, and the overall low sedimentation rates that are

Fig. 1. a) Schematic map of the Arctic Ocean (IBCO 2.0, Jakobsson et al., 2008a) with core sites (yellow dots) from the HOTRAX 2005 expedition and the LOMROG expedition (Darby et al., 2005; Jakobsson et al., 2008b) and cores from earlier expeditions (orange dots) included in this study (see Table 1). Abbreviations used: Ch.P – Chukchi Plateau; NWR – Northwind Ridge; MJR – Morris Jesup Rise; YP – Yermak Plateau. b) The intermediate water circulation (white arrows) and the dominant surface currents (red arrows); the Beaufort Gyre (BG) and the Transpolar Drift (TPD). The dots represent stations where temperature and salinity profiles (see Fig. 2) were recovered in the different basins (Rudels et al., in press).

characteristic for much of the basin (cm/ka) indicate that the recovered sediment only sample the more recent part of the Quaternary. In order to place the sampled stratigraphies in the context of the Arctic Ocean geologic evolution, an overview of the physiography of the submarine ridge systems is required.

The Amerasian Basin of the Arctic Ocean is divided into the Canada Basin and the Makarov Basin by the Alpha and Mendeleev ridges (Fig. 1). This ridge complex stretches more than 1800 km, from the Canadian continental margin off Ellesmere Island to the Eastern Siberian Sea shelf and includes several sea mounts and valleys with depths ranging from 740 to 2000 m water depth (Jakobsson et al., 2003b). Recovered basaltic rocks indicate a Late Cretaceous age for the formation of the Alpha Ridge (Jokat, 2003) and the sediments that drape the ridge crest range in thickness from 500 m to about 1500 m (Jokat, 2003; Lebedeva-Ivanova et al., 2006), although local areas subjected to erosion are characterized by significantly thinner drapes (Jenkyns et al., 2004).

The Chukchi Borderland, which includes the Northwind Ridge and the Chukchi Plateau, is situated within the Canada Basin and extends roughly 600 km out from the Chukchi continental shelf (Fig. 1). These bathymetric highs have steep slopes and relatively flat crests found at generally less than 1000 m water depths (600 m for the Chukchi Plateau) (Hall, 1990; Jakobsson et al., 2003b). Recent geophysical mapping has revealed extensive glacial erosion down to 900 m water depth (Polyak et al., 2001; Darby et al., 2005; Jakobsson et al., 2005; Polyak et al., 2007; Jakobsson et al., 2008c). This information is critical when interpreting the stratigraphy in cores from this area.

The Amerasian Basin is separated from the Eurasian Basin by the Lomonosov Ridge, which extends more than 1500 km from the Siberian continental margin to the northern Greenland margin (Fig. 1). Its shallowest part is located north of Canada and Greenland with water depths less than 1000 m. The sediment thickness atop the ridge ranges from a few tens of meters to well over 1000 m (Jackson and Oakey, 1990; Jokat, 2005; Kristoffersen and Mikkelsen, 2006). The Lomonosov Ridge was the target for the Integrated Ocean Drilling Program (IODP) Expedition 302 (the Arctic Coring Expedition; ACEX). Approximately 428 m of sediment draping the ridge, encompassing the past 56 Ma, were cored near 88°N during ACEX (Moran et al., 2006). The recovered Neogene sediment is characterized by oxygenated silty mud with numerous thin sand lenses and dropstones and indicates a generally persistent mode of glaciomarine deposition since about 17.5 Ma (Backman et al., 2008). At the center of the ridge there is a depression, the Intra Basin, with a sill depth of 1870 m towards the Makarov Basin and 2400 m towards the Amundsen Basin (Björk et al., 2007). Results from a detailed ^{14}C study on a core within this basin show a relatively high sedimentation rate of 7–9 cm/ka during the last deglaciation, possibly indicating that the Intra Basin is a local sediment trap (Hanslik et al., *this issue*). Similar to the Chukchi Borderland, the shallow portions of the Lomonosov Ridge (<1000 m) were subjected to ice-grounding, with the oldest dated event being associated with marine isotope stage (MIS) 6 (Jakobsson et al., 2001; Polyak et al., 2001; Jakobsson et al., 2008c).

The Morris Jesup Rise extends northwards from the northern Greenland shelf into the Amundsen Basin and has an area of approximately 5000 km² (Fig. 1). The water depths of the rise range from 500 to 2500 m. Recent sub-bottom profiles and multibeam studies revealed extensive iceberg scouring down to a water depth of about 1050 m on the rise (Spielhagen et al., 2004; Jakobsson et al., *this issue*). On the other side of the Gakkel ridge lies the Yermak Plateau (Fig. 1), a conjugate feature to Morris Jesup Rise, both formed around 34 Ma (Brozena et al., 2003). The Yermak Plateau extends 400 km into the Nansen Basin from the Svalbard Margin with depths from 500 to 1000 m. Whereas the eastern part of the plateau is

blocky and rough, the southwestern and northern parts have a smoother topography (Myhre and Thiede, 1995). Thick sediment drapes the plateau and results from the Ocean Drilling Program (ODP) Leg 151 indicate relatively high sedimentation rates for the Plio-Pleistocene (on average 14 cm/ka) (Knies et al., 2009). Ice has eroded large parts of the plateau at water depths ranging from 450 to 850 m (Dowdeswell et al., *this issue*; Vogt et al., 1994; Kristoffersen et al., 2004).

2.2. Arctic Ocean circulation

The modern Arctic Ocean surface circulation (Fig. 1) is dominated by the Beaufort Gyre in the Amerasian Basin and the Transpolar Drift flowing from the Siberian shelves along the Lomonosov Ridge towards the Fram Strait (Aagaard et al., 1985; Rudels et al., *in press*). These surface currents define the major transport paths for modern sea ice and icebergs, which ultimately distribute sediment from the continental shelves throughout the Arctic Ocean (Pfirman et al., 1997).

Today, there is an exchange of water between the Arctic, Pacific and Atlantic Oceans through the Bering Strait, the Canadian Arctic Archipelago and the Fram Strait/Barents Sea. Low salinity Pacific waters enter the Arctic Ocean through the relatively narrow and shallow Bering Sea (with a sill depth of 50 m) and contribute to the Polar Surface Water (Woodgate and Aagaard, 2005; Rudels et al., *in press*). In addition, the Polar Surface Water receives freshwater from the rivers draining Siberia and North America (Woodgate and Aagaard, 2005). This cold, low salinity layer is separated from warmer water masses by a cold halocline layer (Fig. 2) (Rudels et al., *in press*).

The Arctic Ocean intermediate waters can be divided into the Atlantic layer (250–600 m) and the Upper Polar Deep Water (600–1700 m) (Fig. 2). Relatively warm and saline water from the Atlantic Ocean enters the Arctic across the Barents Sea and through the Fram Strait (Fig. 1). The Upper Polar Deep Water is a mixture of Atlantic water and cold saline shelf water that sinks to this level and cools the surrounding water mass (Rudels, 1995; Rudels et al., *in press*). The intermediate waters circulate anti-clockwise in the Eurasian Basin and flow into the Makarov Basin, between the Lomonosov Ridge and the Siberian shelf (Fig. 1). The flow continues along two paths: one follows the edges of the Canada Basin and the other along the northern slopes of the Alpha Ridge. Portions of this water mass eventually re-circulate to the Eurasian Arctic through an Intra Basin in the Lomonosov Ridge and across the southernmost connection to the northern Greenland shelf (Björk et al., 2007; Rudels et al., *in press*). The Arctic Ocean Deep and Bottom Waters are mainly formed by brine rejection and also flow anti-clockwise through the basins (Rudels et al., *in press*).

The Amerasian Basin water masses join the Eurasian Basin waters through the gap between the Lomonosov Ridge and Greenland before exiting through the narrow Fram Strait (Björk et al., 2007). Due to the depths of the ridges bordering the Arctic Ocean basins, sediment cores collected during the HOTRAX and LOMROG expedition, and additional cores incorporated in this study, sample sediments deposited within the Upper Polar Deep Water or the deep waters (Fig. 2).

2.3. Sediment transportation

Away from the continental shelves, sea ice and icebergs are the dominant mechanism for transporting terrigenous sediments, and are driven by modern surface circulation patterns. As the principal “exit gateway” of ice the Fram Strait receives a mixture from both Amerasian and Eurasian sources. The main area for sea ice production is the Siberian shelves, especially the Laptev Sea,

Fig. 2. All included core sites in relation to the water masses (adapted from Rudels et al., in press), potential temperature and salinity (PSW – Polar Surface Water; UPDW – Upper Polar Deep Water; EBDW – Eurasian Basin Deep Water; CBDW – Canadian Basin Deep Water). The water samples are taken in the Western Svalbard Current (green), the Nansen Basin within the Fram Strait branch (orange), the interior Nansen Basin (red), the Amundsen Basin (light blue), the Makarov Basin (black) and the Canada Basin (blue). The positions of the stations are marked on the map in Fig. 1.

although sea ice is naturally produced along all of the Arctic Ocean continental shelves.

During interglacial periods sea ice is the main carrier of clastic sediment but occasional discharge of icebergs from the surrounding glaciers occurs as well (Bischof, 2000). However, the surface circulation was likely different during glacial times when eustatic levels are estimated to have dropped between 86 and 102 m below present level during MIS 2, 6 and 8 and up to 150 m during MIS 10 and 12 (Rabineau et al., 2007). This sea level change would expose much of the continental shelf areas and close the 40–50 m deep Bering Strait.

Provenance studies of ice rafted debris (IRD) indicate that the Beaufort Gyre was different during late Pleistocene glacial periods. Sediment incorporated into sea ice and icebergs in the south-western Canadian Arctic Archipelago was transported more

directly towards the Northwind Ridge and the Chukchi Plateau and sediment entrained in the Queen Elizabeth Islands towards the central and southeastern Alpha Ridge and Fram Strait (Bischof and Darby, 1997). This enabled icebergs originating from the Laurentide Ice Sheet and the Innuitian Ice Sheet to exit the Arctic Ocean rapidly. Furthermore, the provenance studies of Bischof and Darby (1997) indicate that sea ice during the late Pleistocene continued to drift from the Siberian Shelf region along a similar pattern as the modern Transpolar drift. The inflow of Atlantic water was reduced during glacials, causing the deep-water ventilation that occurs in the North Atlantic today to move further south and the Upper Polar Deep Waters (intermediate water) that today extend to depths of 1700 m, being replaced by waters originating from brine rejection during sea ice formation, mainly in the Kara Sea (Haley et al., 2008).

Provenance studies based on clay and heavy mineral assemblages in sediment cores from the central Lomonosov Ridge and the Morris Jesup Rise point to the Kara, Laptev, and Barents Sea as the major source areas for terrigenous material during the last two glacial cycles (Spielhagen et al., 2004). In sediment cores from the Amerasian Basin, IRD assemblages and individual Fe oxide mineral grains indicate that the main source of IRD is the Canadian Arctic Archipelago and the Banks Island shelf (Darby, 1989, 2003). Lower resolution studies also revealed that terrigenous sediment from both the Siberian and Canadian Arctic shelves coexist within the Neogene–Quaternary section of the ACEX record (Darby, 2008; Krylov et al., 2008; Vogt, 2009).

3. Methods

During the HOTRAX expedition 21 piston cores (Table 1), on average 10 m long and 11.4 cm outer diameter, were retrieved along a transect from the Chukchi Borderland, across the Arctic Ocean, to the Yermak Plateau (Fig. 1) (Darby et al., 2005). In addition, four piston cores that were collected from the Morris Jesup Rise during the LOMROG 2007 expedition are included in this study (Table 1). All sediment cores were digitally photographed and visually described when split.

Prior to splitting, bulk density and magnetic susceptibility were measured shipboard with Stockholm University's Multi-Sensor Core Logger (MSCL, Geotek Ltd., UK). The cores were logged in plastic liners following the procedure described in the Geotek MSCL

manual (Geotek, 2004). The principles of logging cores on the MSCL are outlined in Weber et al. (1997) and Best and Gunn (1999). Magnetic susceptibility was measured using a Bartington loop sensor with a diameter of 12.5 cm for the HOTRAX cores and 10 cm for the LOMROG cores. There was no correction applied for the loop diameter so comparisons between the absolute magnetic susceptibility values are not made in this study. The MSCL estimates bulk density from the attenuation of a gamma-ray beam transmitted by a radioactive source (^{137}Cs). These measurements were made with a 5 mm collimator and using a 15 s count time. Calibration of the estimated densities is performed by measuring the attenuation of gamma rays in standards made of variable thicknesses of aluminum and water (Geotek, 2004). Some of the magnetic susceptibility and bulk density records from different studies display varying magnitudes caused by differences in equipment. In order to stack these records the data was plotted on a standardized scale.

Bulk density is a measurement influenced by grain size, composition and porosity (Weber et al., 1997; Niessen et al., 2007). The reduction in porosity due to downhole compaction, or past stress history, is the over-riding influence on the absolute bulk density measurement. However, relative variations in bulk density that occur on centimeter to decimeter scales are mostly linked to changes in sediment composition and clay fraction Hounslow and Maher, 1999; Moran et al., 2007.

The magnetic susceptibility reflects variation in the concentration and composition of magnetic material in the sample (Verosub

Table 1

All cores included in the study, references used are a) Darby et al., 2005; b) Adler et al., 2009; c) Jakobsson et al., 2008b; d) Stein et al., 2004; e) Spielhagen et al., 2004; f) Jokat, 1999a g) Fütterer, 1992; h) Jakobsson et al., 2001; i) O'Regan et al., 2008; j) Bergmann, 1996; k) Spielhagen et al., 2004; l) Nowaczyk and Frederichs, 2009; m) Aldahan et al., 1997; n) Knies, 1994; o) Jokat, 2000; p) Rack et al., 1996.

Core Name	Position (lat °N; long °E)	Depth (m)	Location	Included measurements
HLY0503-02JPC ^a	74.4925; −159.8898	627	Northwind Ridge	Magnetic susceptibility, bulk density
HLY0503-03JPC ^a	77.2450; −157.0600	594	Northwind Ridge	Magnetic susceptibility, bulk density
HLY0503-04JPC ^a	77.2187; −157.0485	520	Northwind Ridge	Magnetic susceptibility, bulk density
HLY0503-05JPC ^a	78.4406; −162.6624	660	Chukchi Plateau	Magnetic susceptibility, bulk density
HLY0503-06JPC ^a	78.2938; −176.986	802	Mendelev Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-07JPC ^a	78.2987; −176.0877	802	Mendelev Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-08JPC ^{a,b}	79.5928; −172.5016	2792	Mendelev Ridge	Magnetic susceptibility, bulk density, lithology, coarse fraction, foraminifers, inclination, age model
HLY0503-09JPC ^a	79.5934; −172.4611	2783	Mendelev Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-10JPC ^a	81.2261; −177.1935	1865	Mendelev Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-11JPC ^a	83.1436; −174.5372	2644	Mendelev Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-12JPC ^a	83.2911; −171.9577	1585	Mendelev Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-13JPC ^a	84.3056; −160.6792	1400	Alpha Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-14JPC ^a	84.3033; −149.0340	1856	Alpha Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-15JPC ^a	83.9502; −143.1828	2047	Alpha Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-16JPC ^a	84.1690; −150.9582	2506	Alpha Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-17JPC ^a	85.1272; −154.7784	1741	Alpha Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-18JPC ^a	88.4505; 146.5609	2598	Lomonosov Ridge	Magnetic susceptibility, bulk density, lithology
HLY0503-19JPC ^a	88.7151; 169.7857	1023	Lomonosov Ridge	Magnetic susceptibility, bulk density
HLY0503-20JPC ^a	88.8060; 163.5796	2654	Lomonosov Ridge	Magnetic susceptibility, bulk density
HLY0503-21JPC ^a	86.6623; 55.7166	2333	Gakkel Ridge	Magnetic susceptibility, bulk density
HLY0503-22JPC ^a	80.4898; 7.7690	798	Yermak Plateau	Magnetic susceptibility, bulk density
LOMROG07-PC-07 ^c	85.3997; −14.282	1252	Morris Jesup Rise	Magnetic susceptibility, bulk density, lithology
LOMROG07-PC-08 ^c	85.3198; −14.8575	1038	Morris Jesup Rise	Magnetic susceptibility, bulk density, lithology
LOMROG07-GC-09 ^c	85.2945; −14.8922	1016	Morris Jesup Rise	Magnetic susceptibility, bulk density, lithology
LOMROG07-GC-10 ^c	85.2931; −14.8146	1017	Morris Jesup Rise	Magnetic susceptibility, bulk density, lithology
PS51/034-4 ^d	85.3767; −155.4617	2071	Alpha Ridge	Magnetic susceptibility, bulk density
PS51/038-4 ^{e,f}	85.1350; −171.4417	1473	Alpha Ridge	Magnetic susceptibility, bulk density, coarse fraction, age model, foraminifers, inclination
PS2185-6 ^g	87.5367; 144.9267	1051	Lomonosov Ridge	Magnetic susceptibility, bulk density, lithology
96/12-1pc ^h	87.1308; 144.4395	1003	Lomonosov Ridge	Magnetic susceptibility, bulk density, lithology, coarse fraction, foraminifers, inclination, age model
ACEX ⁱ	87.9212; 139.0318	1209	Lomonosov Ridge	Magnetic susceptibility, bulk density, lithology
PS2200-5 ^{j,k,l}	85.3267; 14.0000	1074	Morris Jesup Rise	Magnetic susceptibility, bulk density, lithology, coarse fraction, foraminifers, inclination, age model
PS2213-3/6 ^m	80.4713; 8.2052	897	Yermak Plateau	Magnetic susceptibility, bulk density, lithology, ¹⁰ Be, inclination
PS2123-2 ⁿ	80.1555; 9.8478	573	Yermak Plateau	Magnetic susceptibility, bulk density
PS55/151-2 ^o	80.7245; 8.0220	971	Yermak Plateau	Magnetic susceptibility, bulk density
ODP 151-911 ^p	80.4774; 8.2273	902	Yermak Plateau	Magnetic susceptibility, bulk density, lithology

and Roberts, 1995) and arise from changes in detrital and authigenic magnetic mineral composition as well as the abundance of diamagnetic biogenic material that will dilute the concentration of magnetic minerals (Hounslow and Maher, 1999).

The MSCL measured physical property records were used to stratigraphically align cores from individual physiographic areas (see Table 2). This was accomplished using the Correlator software developed by the Lamont-Doherty Earth Observatory for the Ocean Drilling Program (<http://sqlcore.geo.umn.edu/CoreWallDatabase/cwWiki/index.php/Correlator>). The correlations presented are the best estimate of the stratigraphic alignments, however, alternate interpretations may exist. Towards the bottom of the cores, the correlations are less well constrained due to that core disturbances, such as stretching, generally increases downcore. The successfully correlated physical property records were aligned and projected onto a common depth scale, dictated by one of the cores within each physiographic region.

The average depths of the regional tie points (Table 2) were then used to produce an average composite depth scale, referred to as

meter composite depth (mcd) from now on. The composite record for each region was produced by averaging the stratigraphically aligned cores following re-sampling (at 1 cm) and smoothing (5 cm running mean). The resulting stacked record should, to some extent, reduce the effect of small errors in the stratigraphic alignment between cores. The variability of the composite data (expressed as one standard deviation) was also calculated and used to represent the accuracy of the stratigraphic alignment.

4. Results

4.1. Stratigraphic correlation

The results showing stratigraphic alignment of cores using magnetic susceptibility and bulk density are initially presented separately for six different geographical areas (1) the Chukchi Borderland (2) southern Mendeleev Ridge (3) northern Mendeleev Ridge and the Alpha Ridge (4) the Lomonosov Ridge (5) the Morris Jesup Rise and (6) the Yermak Plateau. This was done to first assess

Table 2

Depths (m) of the regional tie points in all cores used in the composite stratigraphies. An asterisk marks the cores for which the measurements were averaged to form the composite physical property records. The scale factor indicates the approximate average number the core lengths need to be multiplied with to equal the length of the composite record for each area.

Southern Mendeleev Ridge										
Core	MR1	MR2	MR3	MR4	MR5	MR6	MR7	MR8	MR9	Scale factor
06JPC*	—	0.88	1.77	3.02	5.07	5.65	6.35	8.58	9.05	1.10
07JPC*	—	2.59	3.93	5.17	6.82	7.23	7.95	10.10	10.60	0.94
08JPC*	0.43	4.25	5.58	6.61	7.84	8.12	8.50	9.97	10.27	0.96
09JPC*	0.40	3.30	4.85	5.95	7.16	7.37	7.75	9.33	9.75	1.02
Av. depth:	0.42	2.76	4.03	5.19	6.72	7.09	7.64	9.50	9.92	
Northern Mendeleev Ridge and Alpha Ridge										
Core	AR1	AR2	AR3	AR4	AR5	AR6				Scale factor
10JPC	0.34	1.36	4.05	5.02	5.66	—				1.20
11JPC	0.38	1.45	5.50	6.98	7.42	—				0.92
12JPC*	0.57	1.39	5.38	7.14	8.40	—				0.81
13JPC*	0.51	1.78	5.72	7.75	8.76	—				0.78
14JPC*	0.32	1.18	4.33	5.30	5.96	9.64				1.00
15JPC*	0.19	0.93	2.94	3.87	4.50	8.17				1.18
16JPC	0.42	1.56	4.72	5.63	6.24	9.38				1.03
17JPC*	0.38	1.41	4.83	6.46	7.53	11.50				0.84
PS51/034-4	0.49	1.85	5.92	—	—	—				0.83
PS51/038-4	0.64	2.67	5.70	—	—	—				0.86
Av. depth	0.42	1.56	4.91	6.02	6.81	9.67				
Lomonosov Ridge										
Core	LR1	LR2	LR3	LR4	LR5	LR6	LR7	LR8		Scale factor
ACEX*	0.40	1.56	4.64	5.58	7.18	8.71	10.65	12.27		0.81
18JPC*	0.75	2.08	4.42	5.22	6.54	7.44	9.34	10.41		0.96
PS2185-6*	0.20	1.25	3.10	3.65	4.90	6.10				1.09
96/12-1pc*	0.35	1.64	2.57	2.84	3.51	4.37	5.76	7.20		1.38
Av. depth	0.43	1.63	3.68	4.32	5.53	6.66	8.58	9.96		
Morris Jesup Rise										
Core	MJR1	MJR2	MJR3	MJR4	MJR5	MJR6	MJR7	MJR8		Scale factor
PC-07*	0.85	1.17	—	—	2.23	2.50	—	—		1.04
PC-08*	0.55	0.83	1.09	1.57	2.20	2.44	2.63	4.07		1.03
GC-09*	0.50	0.68	0.87	—	—	—	—	—		1.13
GC-10*	—	—	0.59	0.95	—	—	—	—		1.54
PS2200-5*	0.60	0.95	1.36	1.85	2.59	2.86	3.13	4.36		0.97
Av. depth	0.62	0.91	0.98	1.46	2.34	2.60	2.88	4.21		
Yermak Plateau										
Core	YP1	YP2	YP3	YP4	YP5	YP6				Scale factor
22JPC*	0.73	1.66	4.27	7	9.46	11.19				1.38
151-911*	1.02	2.45	6.65	12.07	18.48	22.51				0.69
PS2213-3/6*	0.38	1.26	3.98	7.38	10.33	12.68				1.22
PS2123-2*	1.18	2.63	5.65	—	—	—				0.91
PS55/151-2*	0.85	2.28	5.04	—	—	—				1.02
Av. depth	0.83	2.06	5.12	8.82	12.76	15.46

Fig. 3. Magnetic susceptibility (MS – 10^{-5} SI) and bulk density (BD – g/cm^3) for cores 02JPC to 04JPC from the Northwind Ridge and core 05JPC from the Chukchi Plateau.

how well cores can be correlated within different regions and then to construct composite sections, when possible, for comparison between the areas.

Cores 02JPC, 03JPC, and 04JPC were recovered from the Northwind Ridge and 05JPC from the Chukchi Plateau (Fig. 1). The magnetic susceptibility and bulk density records display minor and un-diagnostic downhole variations in cores 04JPC and 05JPC and no correlation could be obtained between them (Fig. 3). 02JPC displays a steep downcore increase in the magnetic susceptibility and bulk density records.

Cores 06JPC and 07JPC were retrieved from a depth of about 800 m on the southern Mendeleev Ridge, while 08JPC and 09JPC were taken about 170 km to the north at a water depth of 2800 m (Fig. 1). In these four cores, nine regional tie points were established (Fig. 4 and Table 2). No lithostratigraphic feature explains the prominent dip in magnetic susceptibility and bulk density between MR7 and MR8 in 06JPC and 07JPC. The large magnetic susceptibility peak in 08JPC with a corresponding dip in bulk density at about 2.3 m matches an interval with pinkish coarse grains. However, neither the peak nor this pinkish layer could be easily recognized in

Fig. 4. Magnetic susceptibility (10^{-5} SI) and bulk density (g/cm^3) for cores 06JPC to 09JPC from southern Mendeleev Ridge. Gray lines and shaded areas indicate correlation points, dotted gray line indicates an alternative correlation.

Fig. 5. Magnetic susceptibility (10^{-5} SI) and bulk density (g/cm^3) for cores 10JPC to 12JPC from northern Mendeleev Ridge, 13JPC to 17JPC, PS51/034-4 and PS51/038-4 (Jokat, 1999a) from the Alpha Ridge, gray lines and shaded areas indicate correlation points.

the other southern Mendeleev Ridge cores. The correlation from the shallower to the deeper cores is not straightforward as the interval above MR2 is different between the cores. Therefore, an alternative correlation of a prominent peak below MR9 in 06JPC has been marked with a dotted line in Fig. 4. The core description for 07JPC indicated that some of the upper layers are missing and coring disturbance may at least partially explain the lack of correlation towards the top of the cores. Also, an interval between 2.7 m and 3.5 m in 08JPC has been interpreted to represent a slump or gravity-flow deposit (Adler et al., 2009). Given the stronger correlations below MR2 this may also reflect an up-core shift towards a more complex depositional regime.

Cores 10JPC to 12JPC from the northern Mendeleev Ridge compared well to cores 13JPC to 17JPC and the uppermost part of cores PS51/034-4 (Jokat, 1999a) and PS51/038-4 (Jokat, 1999a) from the central Alpha Ridge (Figs. 1 and 9). The cores are spread over a large area, with 550 km separation between 10JPC and 15JPC, and span water depths between 1400 m for 13JPC and 2644 m for 11JPC. There are six regional stratigraphic tie points (AR1 to AR6) established for these cores (Fig. 5), where tie point AR6 is only identified in 14JPC to 17JPC. Especially pronounced in this correlation are the tie points AR3 and AR4 that are easily recognizable in all the HOTRAX cores, but are less recognizable in PS51/034-4 and

PS51/038-4. The general lack of correlation below AR2 in PS51/038-4 is notable and may indicate an overall change in the processes controlling deposition as one moves north towards the Lomonosov Ridge, where a stronger influence of either the Transpolar drift or Atlantic intermediate water re-circulation may exist. There is no indication in the lithostratigraphic description to explain the major peak in magnetic susceptibility with a subsequent drop in bulk density between 8.4 and 8.6 m in core 10JPC.

Cores 96/12-1pc (Jakobsson et al., 2001) and PS2185-6 (Spielhagen et al., 2004) from the crest of the central Lomonosov Ridge were previously correlated to the ACES record by O'Regan et al. (2008). These cores were compared to 18JPC to 20JPC recovered from the Intra Basin (Fig. 1), where eight regional tie points were determined (Fig. 6 and Table 2). However, no correlation was achieved with 19JPC from the inner flank of the Intra Basin, where an eroded core top was noted. In 20JPC, only the tie points LR3 to LR6 were recognized, whereas a good correlation was established between 18JPC and the cores from the ridge crest. The correlations were mainly achieved using the bulk density records and the definition and correlation of LR5 to LR7 are difficult to recognize in the magnetic susceptibility records. Two dips in bulk density, between LR5 and LR6 and below LR6, correspond to two very fine-grained peach colored layers.

Fig. 6. Magnetic susceptibility (10^{-5} SI) and bulk density (g/cm^3) for 96/12-1pc (Jakobsson et al., 2001), PS2185-6 (Spielhagen et al., 2004), ACEX composite core (O'Regan et al., 2008) from the Lomonosov Ridge crest, and HOTRAX cores 18JPC to 20JPC from the Lomonosov Ridge Intra Basin. Gray lines and shaded areas indicate correlation points.

The LOMROG cores from the Morris Jesup Rise were taken at about 1000 m water depth, where GC-09 and GC-10 were situated within an ice scour on the crest of the rise. An additional core, PS2200-5 (Fütterer, 1992; Bergmann, 1996), from 1074 m water depth is also included in the comparison (Fig. 1). A correlation (Fig. 7 and Table 2) between all cores was possible for the upper parts of the records (MJR1 to MJR4). However, the closely spaced MJR1 and MJR2 were not recognized in GC-10 and the erosional horizon in GC-09 and GC-10 prevents further alignment using these

cores. Four additional tie points were found between the longer cores PC-07, PC-08 (MJR5 to MJR8).

Two HOTRAX cores 21JPC from the Gakkel Ridge and 22JPC from the Yermak Plateau were compared to PS2213-3 (Fütterer, 1992), PS2213-6 (Bergmann, 1996), PS2123-2 (Knies, 1994), PS55/151-2 (Jokat, 2000) and the composite stratigraphy from ODP Site 911 (ODP 151-911) (Rack et al., 1996) from the southern Yermak Plateau (Fig. 1). The relatively short core 21JPC does not correlate to any of the cores from the Yermak Plateau or other cores included in this

Fig. 7. Magnetic susceptibility (10^{-5} SI) and bulk density (g/cm^3) for LOMROG cores PC-07, PC-08, GC-09, GC-10 and PS2200-5 (Fütterer, 1992; Bergmann, 1996) from the Morris Jesup Rise. Gray lines and shaded areas indicate correlation tie points.

study. On the Yermak Plateau, a total of six regional tie points (YP1 to YP6) were established between the two longer cores (22JPC and ODP 151-911) and some of the upper tie points were observed in the shorter cores (Fig. 8 and Table 2). The sedimentation rates vary greatly in this area, where tie point YP6 is observed at 11.2 m in 22JPC and at 22.63 m in ODP 151-911.

4.2. Composite stratigraphies

In five of the identified areas, successful correlations were achieved (Figs. 9 and 10). In the sixth area, the Chukchi Borderland, the lack of correlations among the cores recovered from 520 to 660 m, prevented a composite stratigraphy from being developed. However, a study utilizing visual core descriptions and high-resolution bulk density measurements on undisturbed core material (recovered from 690 to 1917 m) from the Northwind Ridge showed a good correlation within the area illustrating a progressive thinning in lithostratigraphic units towards the northern part of the ridge (Phillips and Grantz, 2001). In this section we focus on the composite stratigraphies derived from the five regions within which correlations were established in this study.

The alignment of some of the successfully correlated physical property records is shown in Fig. 10. The averaged depths of lithological layers and key features are included and together with the composite physical property records they form five composite stratigraphies.

4.2.1. Southern Mendeleev Ridge

The four HOTRAX cores from the southern Mendeleev Ridge were used to create the composite stratigraphy for the area (Fig. 10, 11 and Table 2). The uppermost 3.5 mcd is interlaminated brown and yellowish brown sandy mud that transitions into a similar less sandy unit. The bottom unit of these cores is mostly dark brown mud with intervals of lighter colored mud. Dark brown to medium brown cycles occur throughout the cores. Magnetic susceptibility peaks MR2 and MR4 occur within dark brown layers in all of the cores and the dips in magnetic susceptibility at MR5, MR7 and MR8 correspond to medium brown cycles. However, no extensive study of the connection

between the brown cycles and magnetic susceptibility or bulk density patterns was undertaken. A pinkish white layer was consistently found at peak MR1 as well as a sand layer that corresponded to the major dip in magnetic susceptibility between MR3 and MR4. The sedimentation rates for these cores are fairly uniform and the average scale factor by which the core depths need to be multiplied to equal the total length of the composite record are all close to 1 (Table 2).

The uppermost 5 m of the 11.5 m long core 08JPC was studied further by Adler et al. (2009) and coarse fraction, planktic foraminiferal abundance, paleomagnetic inclination and the developed chronology are included alongside the composite record. The age model for this core is based on paleomagnetic inclination, ^{14}C dating of planktonic foraminifera and foraminiferal stratigraphy. The MIS 6/7 boundary (191 ka, Lisiecki and Raymo, 2005) at about 4.0 m is equivalent to 2.6 mcd, just above tie point MR2 in the composite stratigraphy. This yields an average sedimentation rate of approximately 1.4 cm/ka. Just below this depth in 08JPC is a peak in foraminiferal abundance as well as the first downhole reversal in the inclination data, interpreted to represent MIS 7. High coarse fraction abundances are found throughout the 5 m interval studied and published by Adler et al. (2009) although the highest values are found in what is interpreted as MIS 6.

4.2.2. Northern Mendeleev Ridge and Alpha Ridge

The composite stratigraphy for the northern Mendeleev Ridge and the Alpha Ridge was constructed from eight HOTRAX cores (10JPC to 17JPC), PS51/034-4 and PS51/038-4 (Figs. 10 and 12, and Table 2). These cores are dominated by interlaminated brown and yellowish brown mud, becoming sandier towards the top and slightly darker downcore. The alternating brown cycles are found in all of the cores. Two dark brown layers correspond to the magnetic susceptibility highs and bulk density lows AR4 and AR5 and a sand layer is consistently found at the magnetic susceptibility dip and bulk density peak below AR2. Core 13JPC has the highest sedimentation rate of the included cores and needs to be compacted by an average factor of 0.78 to fit the total length of the composite record (Table 2).

Additional published data such as coarse fraction, planktic foraminifera, paleomagnetic inclination and an age model for core

Fig. 8. Magnetic susceptibility (10^{-5} SI) and bulk density (g/cm^3) for cores 21JPC from the Gakkel Ridge and 22JPC, PS2213-3/6 (Fütterer, 1992; Bergmann, 1996), PS2123-2 (Knies, 1994), PS55/151-2 (Jokat, 2000), and ODP 151-911 (Rack et al., 1996) from the Yermak Plateau. Gray lines and shaded areas indicate correlation points.

PS51/038-4 is included in Fig. 12. This core is 7.19 m long, however coarse fraction and foraminiferal data is only available for the top 1.7 m and paleomagnetic inclination data is available down to 2.0 m. The age model is based on paleomagnetic inclination, ^{14}C dating of planktonic foraminifera, biostratigraphy, ^{10}Be and correlations with cores from the Lomonosov Ridge (Spielhagen et al., 2004). The first reversal in paleomagnetic inclination was identified at 1.32 m within MIS 7 and can be inferred to approximately 1.14 mcd. The MIS 6/7 boundary is identified where there is an upward shift to a higher coarse fraction content and the disappearance of planktic foraminifers at about 1.22 m. This boundary

occurs at about 1.00 mcd between tie points AR1 and AR2, which yields an average sedimentation rate of about 0.5 cm/ka for the upper meter of the core. The higher variability in bulk density above AR2 (Fig. 5) corresponds broadly to MIS 1 to 6 according to the age model in PS51/038-4.

4.2.3. Lomonosov Ridge

Core 18JPC from the Intra Basin, 96/12-1pc, PS2185-6 and the ACEX composite section was included in the composite stratigraphy (Figs. 10 and 13 and Table 2). These cores are made up of an interlaminated brown and yellowish brown sandy mud at the top

Additional data is included for the 7.25 m long PS2200-5. The published records of coarse fraction and inclination cover the uppermost 7 m. Planktic foraminiferal abundance data is available down to 4 m. The age model for PS2200-5 relies on paleomagnetic inclination, ^{14}C ages from planktonic foraminifera, biostratigraphy, ^{10}Be and correlations with cores from the Lomonosov Ridge (Spielhagen et al., 2004). The MIS 7.5 (234 ka, Lisiecki and Raymo, 2005) corresponds to a depth in the composite stratigraphy of 3.25 mcd, indicating an average sedimentation rate for the Morris Jesup Rise of 1.4 cm/ka. The interval interpreted as MIS 7 in

Fig. 10. Physical properties for different cores within the same area stretched to the same depth scale as one of the cores. Here the scales refer to the records marked with a black line.

PS2200-5 incorporates a peak in foraminiferal abundance, the first inclination change and an interval of relatively low coarse fraction content.

4.2.5. Yermak Plateau

Two longer cores, 22JPC and the composite section for Site 911 drilled during ODP Leg 151 (Rack et al., 1996), together

with the shorter cores PS2213-6 (Bergmann, 1996; Fütterer, 1992), PS2123-2 (Knies, 1994), and PS55/151-2 (Jokat, 2000) were used for the composite stratigraphy for the Yermak Plateau (Figs. 10 and 15 and Table 2). The cores contain mainly grayish brown to yellowish brown sandy, silty mud, although the available lithological descriptions were limited. The average sedimentation rate for the ODP Site 911 record is

Fig. 11. Composite stratigraphy including general lithostratigraphy (legend for all composites), average magnetic susceptibility and average bulk density for the southern Mendeleev Ridge based on HOTRAX cores 06JPC to 09JPC. The physical property records are plotted with a 5 cm running mean. The variability of the composite data set within one standard deviation of the mean is shown in gray. Also included are coarse fraction, planktic foraminiferal abundance and inclination data for 08JPC as well as MIS from Adler et al. (2009).

significantly higher than the other cores, requiring compression by a factor of 0.69 to equal the total length of the composite record (Table 2).

The ^{10}Be , age and remanent inclination after 40 mT demagnetization for the whole core PS2213-6 (Aldahan et al., 1997;

Schneider, 1996) are included in Fig. 15, where the ^{10}Be record forms the basis for the age model. The MIS 8/9 boundary (300 ka, Lisiecki and Raymo, 2005) can be inferred to 14.55 mcd, which results in an average sedimentation rate of 4.8 cm/ka, which is the highest of all included areas.

Fig. 12. Composite stratigraphy including general lithostratigraphy (for legend see Fig. 11), average magnetic susceptibility and average bulk density for the northern Mendeleev Ridge and Alpha Ridge based on HOTRAX cores 10JPC to 17JPC, PS51/034 and PS51/038 (Jokat, 1999a). The physical property records are plotted with a 5 cm running mean. The variability of the composite data set within one standard deviation of the mean is shown in gray. Also included are coarse fraction, planktic foraminiferal abundance, inclination data and MIS for PS51/038-4 (Spielhagen et al., 2004).

Fig. 13. Composite stratigraphy including general lithostratigraphy (for legend see Fig. 11), average magnetic susceptibility and average bulk density for the Lomonosov Ridge based on HOTRAX core 18JPC, 96/12-pc (Jakobsson et al., 2001), PS2185-6 (Spielhagen et al., 2004) and ACEX drill core (O'Regan et al., 2008). The physical property records are plotted with a 5 cm running mean. The variability of the composite data set within one standard deviation of the mean is shown in gray. Also included are coarse fraction, planktic foraminiferal abundance, MIS and inclination data for 96/12-1pc (Jakobsson et al., 2001).

Fig. 14. Composite stratigraphy including general lithostratigraphy (for legend see Fig. 11), average magnetic susceptibility and average bulk density for the Morris Jesup Rise based on PS2200-5 (Bergmann, 1996; Fütterer, 1992) and LOMROG cores PC-07, PC-08, GC-09 and GC-10. The magnetic susceptibility record is plotted on a standardized scale to account for differences in sensitivity of the measurements and both physical properties are plotted with a 5 cm running mean. The variability of the composite data set within one standard deviation of the mean is shown in gray. Also included are coarse fraction, MIS, planktic foraminiferal abundance and inclination data for PS2200-5 (Spielhagen et al., 2004).

Fig. 15. Composite stratigraphy including general lithostratigraphy (for legend see Fig. 11), average magnetic susceptibility and average bulk density for the Yermak Plateau based on 22JPC, PS2213-3/6 (Bergmann, 1996; Fütterer, 1992), ODP core 151-911 (Rack et al., 1996), PS55/151-2 (Jokat, 2000), and PS2123-2 (Knies, 1994). The bulk density record is plotted on a standardized scale to account for differences in sensitivity of the measurements and both physical properties are plotted with a 5 cm running mean. The variability of the composite data set within one standard deviation of the mean is shown in gray. Also included are ^{10}Be , age and remanent inclination after 40 mT demagnetization for PS2213-6 (Schneider, 1996; Aldahan et al., 1997). Note the different depth scales for the two records.

5. Discussion

The standard lithostratigraphic units (A–M) established by Clark et al. (1980) from Amerasian Basin cores were based on lithostratigraphic features such as prominent sand layers and pinkish white carbonate layers. Although these units have been observed in numerous cores from the Alpha and Mendeleev ridges (Clark et al., 1990; Jokat, 1999b; Minicucci and Clark, 1983; Mudie and Blasco, 1985), we are not able to unequivocally identify them in our HOTRAX cores or other records retrieved from the Lomonosov Ridge and the Eurasian Basin (Sellén et al., 2008). There are, however, some lithostratigraphic components from Clark et al. (1980) that are possible to identify in cores from the Alpha and Mendeleev ridges, such as prominent carbonate layers.

The use of physical properties as a tool for stratigraphically aligning cores from different holes and sites is consistently applied within the scientific Ocean Drilling community. This forms the basis for constructing a 'composite' stratigraphic section from overlapping cores recovered in multiple holes. Detailed age models are then constructed through direct or relative dating techniques. Here, we apply the same principles and test to what spatial extent they are applicable within the Arctic Ocean. A lack of correlatable features between two or more records can be due to many depositional and post-depositional processes, but a reliable correlation is a powerful argument for the existence of similar depositional systems when these changes do co-vary through time across large spatial distances (i.e. between cores). The composite stratigraphies, constructed based on similar variations in physical properties, provide a less subjective approach than using lithostratigraphy alone and may contribute to a more robust framework for developing and interpreting chronologies within individual depositional areas and across the Arctic Ocean. As more sediment cores

are retrieved from the Arctic Ocean, and more analyses are made, additional information may be added to revise the composite stratigraphies. This also applies to additional data such as detailed proxies for sediment source and composition, potentially available through analysis of XRF-scanning data.

5.1. Possible circulation and provenance controls on stratigraphy

The close correlation of the physical property data within individual physiographic regions generally illustrates spatially consistent depositional patterns through the later part of the Quaternary. The highest average sedimentation rate is observed in the Yermak Plateau composite record, although there is large variability between the cores in this region. The spatial correlation was limited and cores from the northern part of the plateau and from the flank, such as ODP Site 912, could not be unambiguously linked to the composite stratigraphy. Regional sedimentary patterns may be affected by variable sediment input from the Svalbard continental shelf via sea ice and icebergs, irregular bathymetry and extensive ice erosion in the shallower areas (Vogt et al., 1994; Jakobsson et al., this issue). Additionally, the core sites are also situated within the inflow path of Atlantic waters (Dowdeswell et al., this issue), which must exert a strong influence on sediment deposition.

The intermediate Atlantic waters continue from the Yermak Plateau along the continental slope of the Barents–Kara Sea before turning with one path following the Lomonosov Ridge. Fine-grained sediment from the shelf edge may become suspended and transported to the center of the Lomonosov Ridge (Fahl and Nöthig, 2007), although the magnitude of this transport mechanism is uncertain. Sea ice transported by the Transpolar Drift carries material from the Siberian shelf, mostly from the Laptev Sea, over the Lomonosov Ridge towards the Fram Strait. Although, the central

Lomonosov Ridge receives ice rafted material from both the Eurasian and the Amerasian side of the ridge (Fig. 9) (Bischof and Darby, 1997). According to our results, the Lomonosov Ridge cores have the second highest sedimentation rates in the Arctic Ocean. This may be due to the combined accumulation of sediment transported within both the Beaufort Gyre and the Transpolar Drift. Core 18JPC from the Intra Basin is closely linked to the cores from the ridge crest, indicating that the same sedimentation regime exists, despite being located in different water masses. The poor correlation to the other two cores (19JPC and 20JPC) from the Intra Basin arises from dynamic circulation and reworking of sediment within this basin (Björk et al., 2007).

Some of the Atlantic derived intermediate water spills over the Lomonosov Ridge and flows into the Makarov Basin. Cores located between the Lomonosov Ridge and Alpha Ridge exhibit a slightly higher sedimentation rate than the cores found further into the Canada Basin (Fig. 5). PS51/038-4 is located closest to the Lomonosov Ridge and cannot be correlated to the northern Mendelev Ridge and Alpha Ridge HOTRAX cores below tie point AR2. However, this core has been linked to PS2185-6 by Spielhagen et al. (2004) by coarse fraction content and planktic foraminiferal abundances. This core may therefore be located in a transitional zone between two depositional regimes. The average sedimentation rate for the cores on the Alpha Ridge and northern Mendelev Ridge are the lowest in the Arctic Ocean and may be a result of the location within the Beaufort Gyre. The cores receive material from sea ice and icebergs carried within the sluggish gyre and potentially deliver less sediment due to lower rates of melting.

The southern Mendelev Ridge composite stratigraphy has intermediate sedimentation rates and likely receives more material from the continental shelves and the adjacent Chukchi Plateau via current transport. The melting of sea ice during summers is also more pronounced closer to the shelf edge than in the central part of the ocean basin (Comiso et al., 2008). The sediment sources of the East Siberian Sea and the Laptev Sea influence both the Mendelev Ridge and the Alpha Ridge, but large differences in sedimentation rates make any physical property based correlation difficult. Additionally, icebergs derived from the Laurentide Ice Sheet may not have reached the southern Mendelev Ridge to the same extent as the northern part of the ridge (Fig. 9) (Bischof and Darby, 1997).

The sediment records from the Chukchi Borderland appear heavily affected by glaciogenic processes (Jakobsson et al., 2005; Polyak et al., 2007). The flat downhole density profiles in cores 04JPC and 05JPC are either due to coring disturbance, ice-grounding resulting in homogenization of the sediment from reworking, or represent rapid deposition. The record for 02JPC, which is from a depth and region where mega-scale lineations are present (Jakobsson et al., 2005), has a steep downhole density gradient suggesting erosion or sub-glacial deformation.

The Morris Jesup Rise should record both the Amerasian Basin and Eurasian Basin signal as sediment incorporated in sea ice from the Beaufort Gyre and the Transpolar Drift affects the area. The notable presence of dolomite in core PS2200-5 confirms that sediment from the Canadian Arctic Archipelago or northern Greenland accumulates on the rise (Darby, 1971; Vogt, 1997). Similarly, the waters flowing across the rise are a mixture of Amerasian Basin and Amundsen Basin intermediate waters, possibly bringing suspended sediment from both sides of the Lomonosov Ridge. The Morris Jesup Rise sedimentation regime has similar intermediate sedimentation rates as the southern Mendelev Ridge and is similarly close to a continental shelf source. The lower average sedimentation rates, when compared to the central Lomonosov Ridge, may be attributed to the distance that sea ice has to travel before reaching this location, or a reduction in the amount of melting that occurs in this setting. Ice-grounding on the Morris Jesup Rise, recorded in GC-09 and GC-10 has been precisely

identified in these cores and dated to pre-MIS 5.5 (Jakobsson et al., this issue). This implies that the ice-grounding on the Morris Jesup Rise took place during MIS 6, at the same time as grounding events dated on the central Lomonosov Ridge (Jakobsson et al., 2001; Jakobsson et al., 2008c).

5.2. Links between composite stratigraphies

As we have shown, there are at least five physiographic regions in the central Arctic Ocean, characterized by similar depositional patterns. However, a key question remains as to whether useful stratigraphic links exist between these regions. There are three notable markers that could potentially be used as stratigraphic/chronostratigraphic tie points between the composite sections (Fig. 16).

One of the most conspicuous changes in any of the composite stratigraphies is the clear shift towards a re-current coarse-grained facies that characterize the last two glacial cycles on the Lomonosov Ridge. The longer ACEX record shows that this pattern is not observed previously in the Quaternary sediment record (O'Regan et al., 2009). This change in depositional mode has been linked by Spielhagen et al. (2004) to terrestrial reconstructions of the Barents–Kara ice sheet (Svendsen et al., 2004). The input of large amounts of coarse-grained material is tentatively tied to the marine based part of this ice sheet in the Kara Sea (Jakobsson et al., 2001; Haley et al., 2008). Whether or not this increase in coarse fraction abundance on the Lomonosov Ridge is solely caused by the Barents–Kara ice sheet remains an open question. A reconstruction of the Kara ice sheet indicates that there have been major glacial events during at least two earlier Pleistocene glaciations (Möller et al., 2006), which challenges arguments for a local origin of the grain size shift at MIS 6/7 on the Lomonosov Ridge.

Similar changes in the abundance of coarse fraction material can be observed in core PS2200-5 from the Morris Jesup Rise since MIS 6 (Spielhagen et al., 2004), however, there is no general shift to coarser grain sizes at MIS 6/7 boundary, which may be due to its relative proximity to the Greenland continental shelf. In the Amerasian basin cores, a coarse-grained interval dated to MIS 6 is consistently observed in published grain size records. In 08JPC from the southern Mendelev Ridge, this does not seem to represent a comparable shift in the long-term abundance of coarse-grained material since MIS 6. PS51/038-4 from the Alpha Ridge shows an apparently dramatic increase in the coarse fraction abundance at the MIS 6/7 boundary, but long grain size records are largely lacking. Therefore, while MIS 6 sediments often appear to be enriched in coarse material, it is only on the central Lomonosov Ridge that this occurs uniquely in MIS 6 or younger sediments. However, a potential problem when interpreting grain size records from low sedimentation rate regimes in the Amerasian Basin is the influence of authigenic precipitates of Fe–Mn nodules. Clark et al. (1990) show that 20–30% of the coarse fraction content in sediment from the northern Alpha Ridge is composed of these authigenic minerals. Accounting for the authigenic component of the coarse fraction material would facilitate comparing the downhole changes in grain size between these regions.

A more obvious stratigraphic marker may be the first prominent paleomagnetic inclination change. This change is observed on the Mendelev Ridge, Alpha Ridge, Lomonosov Ridge and Morris Jesup Rise (Jakobsson et al., 2000; Spielhagen et al., 2004; O'Regan et al., 2008; Polyak et al., 2009) (Fig. 16), and is consistently used to-date the MIS 6/7 transition. In many of the analyzed cores where this inclination shift is recognized, there are independent chronostratigraphic data that confirm an age at least older than MIS 5 and most likely older than MIS 6. This evidence includes amino acid racemization (08JPC and 96/12-1pc, Kaufman et al., 2008); OSL

Fig. 16. Comparison of composite magnetic susceptibility (10^{-5} SI) and bulk density (g/cm^3) records and potential correlation points based on a shift to coarser grains at MIS 6/7 (blue), foraminiferal peak at MIS 7.1 (green) and first paleomagnetic inclination change (red). The position of the gray layer has been marked for the Lomonosov Ridge and the Morris Jesup Rise composite stratigraphies.

dating (96/12-1pc, Jakobsson et al., 2003a); diagnostic calcareous nannofossil assemblage and abundance data (see the summary in Backman et al., 2009), and the basin-wide restricted occurrence of the benthic foraminifera *Bulimina aculeata* in MIS 5.1 (96/12-1pc, Jakobsson et al., 2001; GreenIce 10, Nørgaard-Pedersen et al., 2007; NP-26 cores, Polyak et al., 2004; NWR 5, Poore et al., 1994) and *Pullenia* spp in MIS 7 (96/12-1pc and NWR 5, Backman et al., 2004).

A recent study of the magnetic mineralogy of HOTRAX core 06JPC from the southern Mendelev Ridge offers insights into the cause of the recognized anomalous paleomagnetic inclination data recorded in many cores from the Arctic Ocean. Channell and Xuan (2009) show that the paleomagnetic stratigraphy of this core is likely affected by oxidation processes that result in the development of so-called self-reversals. They propose that titanomagnetites are transformed to titanomaghematites during seafloor oxidation events in areas with low sedimentation rates, which acquire a chemical remnant magnetization that is partially self-reversed and not resolvable using alternating field demagnetization techniques. Channell and Xuan (2009) suggest that changes in the amount and type of organic matter delivered to the seafloor may be one explanation for the apparent basin-wide correlation of these long-duration negative excursions (Spielhagen et al., 2004; O'Regan et al., 2008). Another possibility is that variations in the inflow of Atlantic, Pacific and freshwater affects the ventilation of the basin on glacial-interglacial timescales. Whether the inclination changes record the time-dependant behavior of the geomagnetic field, or the ventilation conditions within the Arctic Ocean, both possibilities provide useful stratigraphic and/or paleoceanographic information. The absence of the anomalous excursion behavior in sediment younger than MIS 6/7 suggests a rather fundamental and basin-wide change in paleoceanographic conditions, organic carbon delivery and/or overall sedimentation rates.

The third boundary that may be used for correlation is the last prominent peak in planktic calcareous foraminiferal abundance (Fig. 16). However, recent results show that even within the central Lomonosov Ridge the preservation of calcareous foraminifera may vary from core to core. In the ACEX record, planktic foraminifera from MIS 9 were observed (Cronin et al., 2008), whereas in 18JPC from the Intra Basin only calcareous foraminifera from MIS 3 are

preserved (Hanslik et al., this issue). In order to use planktic foraminiferal peaks for correlation, the age of the lowermost peak must be firmly established. Where the resolution does not allow distinguishing between MIS 7 interstadials, we assume that the lowermost foraminiferal peak corresponds to MIS 7.1, and this is consistently observed between the coarse fraction shift and the first inclination change (Fig. 16).

A potential fourth stratigraphic marker may be the prominent gray unit identified on both the central Lomonosov Ridge (Jakobsson et al., 2001) and the Morris Jesup Rise, but assumes that this unit originates from a particular event in time and requires additional work on the mechanism driving the diagnostic color change in the sediments. A correlation between the Morris Jesup Rise and the Lomonosov Ridge based on this layer indicates higher sedimentation rates on the Lomonosov Ridge, which is consistent with previous stratigraphic links (Fig. 16).

Perhaps one of the most interesting aspects is that the first three outlined stratigraphic markers fall close to one another in the different physiographic regions. This observation suggests that identifying all three markers may be the best way to correlate between the composite stratigraphies. Also, there may be a common and underlying cause for (a) the increased variability and amount of coarse fraction material delivered to these regions, (b) the basin-wide change in magnetic mineralogy within the sediments and (c) the preservation/deposition of calcareous foraminifera throughout the Arctic Ocean. The link between these observations and paleoceanographic changes and/or sea ice conditions remains unknown. However, while these markers seem to be correlatable across the central Arctic Ocean, none appear to be found in sediments deposited on the Southern Yermak Plateau (Knies et al., 2009). This may be explained by the stronger influence of inflowing Atlantic water, which maintains less intensive sea ice cover and more ventilated conditions.

6. Conclusions

In summary, we can conclude that non-destructive physical property records, such as magnetic susceptibility and bulk density are useful for correlating widely spread core sites within similar

depositional environments. Based on correlations of cores from all of the major ridges and plateaus, six areas were demarcated and represent unique sedimentation regimes. The sediment transport to each area must have varied uniformly through time, at least for the latter part of the Quaternary. The proposed five composite stratigraphies may be used to further develop age models and constrain interpretations of past ice transport and current patterns. By applying available age models to the composite stratigraphies, we can deduce that the highest general sedimentation rates are found on the Yermak Plateau and the lowest on the northern Mendelev and Alpha ridges, especially towards the Canada Basin. Although no single stratigraphic feature can be unambiguously used to link cores from different physiographic regions, an up-core shift towards larger and more frequent variations in the coarse fraction content, the appearance of calcareous foraminifera and a change in the magnetic mineralogy (inferred from paleomagnetic inclination records) all point to a significant shift in paleoceanographic conditions occurring around the MIS 6/7 boundary.

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Appendix. Supplementary information

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