Late Weichselian to Holocene sedimentation in the inner Kara Sea: Qualification and Quantification of Processes

Klaus Hauke Dittmers


ISSN 1618 - 3193
Late Weichselian to Holocene sedimentation in the inner Kara Sea: Qualification and Quantification of Processes

Sedimentationsprozesse in der inneren Kara See (spät Weichsel bis Holozän)

Klaus Hauke Dittmers
Die vorliegende Arbeit ist die inhaltlich unveränderte Fassung einer kumulativen Dissertation, die im Fachbereich Geowissenschaften der Universität Bremen eingereicht wurde.

Bremerhaven, November 2005

Eine elektronische Version (PDF) dieses Dokumentes kann bezogen werden unter: unter http://hs.pangaea.de/Reports
Content

Summary        IV-VI
Zusammenfassung        VI-IX

Part A

1 Introduction        1
   1.1 Aims of this Thesis        1
      1.1.1 The Project        1
      1.1.2 Overview        1
      1.1.3 Outline and Structure of the Thesis        3

2 Background        5
   2.1 Physiographic Settings of the Study Area        5
   2.2 Hydrography        7
   2.3 Research History        10

Part B

3 Results        15
   3.1 Bathymetry        15
   3.2 Recent sediment distribution patterns and pathways on the inner
      Kara Sea Shelf (Siberia)        21
      1 Introduction        21
      2 Methods        25
      3 Results        26
      4 Discussion        33
      5 Conclusions        37

   3.3 Holocene sediment budget and sedimentary history of the Ob and
   Yenisei estuaries        41
      1 Introduction        41
      2 Materials and Methods        43
      3 Results        45
         3.1 Subbottom profiling and seismic stratigraphy        45
         3.2 Distribution and thickness of Unit I sediments        51
         3.3 Sediment-core lithology        51
3.4 Correlation of acoustic units with sediment cores 54
3.5 Sediment mass 59
4 Discussion 60
  4.1 Late Quaternary sedimentary environments 60
  4.2 Sediment budget 65
5 Conclusions 70

3.4 Acoustic facies on the inner Kara Sea Shelf: implications for late Weichselian to Holocene sediment dynamics 75
  1 Introduction 76
  2 Methods 79
   2.1 Acoustic Data Collection/Acquisition 79
   2.2 Groundtruthing by gravity coring 80
   2.3 Data analysis 82
  3 Results and Interpretation 84
   3.1 Recent sedimentary acoustic facies type 84
   3.2 Lithofacies 89
   3.3 Lithostratigraphy and age control 89
  4 Discussion 92
  5 Conclusion 106

3.5 Late Weichselian Fluvial Evolution on the Southern Kara Sea Shelf, North Siberia submitted to Global Planetary Change 111
  1 Introduction 111
  2 Data Base and Methods 115
   2.1 Geophysical data collection/acquisition 115
   2.2 Morphometric Analysis 117
  3 Results 118
   3.1 Topography 118
   3.2 Channel geometry and morphological parameters 119
   3.3 Spatial distribution of channels on the shelf 122
  4 Discussion 124
   4.1 Isostasy 124
   4.2 Fluvial style of Ob and Yenisei rivers 125
   4.3 Development/evolution of channel geometry 134
   4.4 Paleo-environmental implications 135
  5 Conclusions 141
Part C

Key results
How is the fluvial derived material dispersed on the (recent) shelf 145
The establishment of a Holocene Sediment budget 145
What was the Holocene evolution of the inner Kara Sea Shelf 146
What was the fate of the large west Siberian rivers (Ob and Yenisei) during last glacial maximum sea level lowstand 147

5. Synthesis 148
   Landforms and features diagnostic for an ice-dammed lake 148
   Possible traces of a lake evidence from our data: 149
   Future research perspective 152

Part D

Acknowledgements 157
Literature 159-171
Summary

Recent sediment pathways (Chapter 3.2)

Surface sediments of the Kara Sea display a strong bimodal distribution in the mud fraction. The peaks cluster at $8 \phi$ (4 µm), the dominant mode, and at a secondary peak of $5 \phi$ (25-40 µm). The widespread occurrence of bimodal grain-size distribution mode in the sortable silt fraction leads us to the interpretation that two main processes are responsible for the material transport, strongly suggesting a twofold peak in current intensity. On the basis of our data it is impossible to relate a distinct grain-size fraction with a certain transport mechanism and there is some bias to identify the “exact” processes. We favour a “summer” mode dominated by extensive, suspended sediment-laden freshwater influx, and a “winter” mode characterised by material redistribution and resuspension enhanced by sea-ice and polynia formation and various processes related to them.

Furthermore, Magnetic susceptibility of surface sediments was investigated to reconstruct recent transport pathways of magnetic minerals (terrestrial sediments). The MS signal clearly originates in the Putoran Mountains in the drainage area of the Yenisei branch and values decrease with increasing water depth. We see this as evidence that the channels distribute the MS signal on the shelf. It is evident that sediment pathways follow the morphology of the sea-floor, indicating density-driven bottom layer transport. After the particles have made their way through the water column they settle and accumulate in density-induced flow following the Pleistocene bottom relief. This process is active mainly during summer. During winter resuspension, enhanced particle reorganisation occurs.

Sedimentation in the Ob and Yenisei estuaries: evolution and budget (Chapter 3.3)

Soft sediments in the Ob and Yenisei estuaries can be divided into two major acoustic and lithological Units I and II. Unit II forms the pre-Holocene basement with little acoustic penetration. During the last sea level lowstand, the erosional surface was formed on top of this unit including several incised river channels. The younger Unit I can be divided into three subunits, Ia to Ic. The lowermost subunits Ic and Ib show fluvial to shallow-marine character comparable to typical channel levee complexes representing a fine-grained meandering river depositional environment overlain by a transgressive system tract and finally high-stand system tract after about 5 Calender ka BP.

During the Holocene an estimated total of 50 % and 75 % of the amount of modern river-supplied material extrapolated for the Holocene, accumulated. A progressive
southward migration of the position of the marginal filter in response to the elevating sea level is favoured to account for the “missing” material.

The existence of filled paleo-river channels suggests a pre-Holocene history of the southern Kara Sea Shelf similar to that in the Laptev Sea (Kleiber and Niessen, 1999). These channels must have been eroded during the sea level lowstand and were filled with sediments during the Holocene transgression. River erosion of the exposed shelf of the southern Kara Sea during the Last Glacial Maximum (LGM) implies continuous river transport across the shelf and absence of an ice sheet in the area of channel erosion.

**Acoustic facies on the inner Kara Sea Shelf: implications for late Weichselian to Holocene sediment dynamics (Chapter 3.4)**

Late Pleistocene to Holocene seismic units on the inner Kara Sea Shelf are composed of sedimentary units the deposition of which was strongly controlled by sea level fluctuations. We studied the impact of sea level fluctuations on sediment “dynamics” in the inner Kara Sea by using high-resolution acoustic sub-bottom profiling. The acoustic lines were ground-truthed with dated sediment cores. The first model data on LGM isostatic rebound for this region led us to the first confined interpretation of acoustic units in terms of sequence stratigraphy. We identified and mapped three characteristic sedimentary and depositional surface environments and established sediment masses for each facies type for the Holocene. Furthermore we refined the location of the eastern LGM ice margin, by new sub bottom profiles.

Sea level fluctuations resulted in periods of sub-aerial exposure of huge shallow shelf areas especially in the circum Arctic realm. The Kara Sea Shelf is characterised by a major incised dendritic paleo-river network, that is variably infilled with late Pleistocene to Holocene estuarine sediments topped by marine facies, showing draping geometries.

The Kara Sea can be subdivided into three facies areas A to C: estuaries, the shelf and deeper (> 120 m water depth) lying areas that accumulated a total of 114 x 10^10 t of Holocene sediments.

The shelf channel system was formed during the Weichselian sea level cycle having its lowstand in the LGM. Deposition occurred constantly in Facies C, with higher sedimentation rates during sea level lowstand when all material by-passed Facies A and B. Sediment accumulation progressively shifted towards the modern coastline and the recent estuary of Ob and Yenisei, making up Facies A. There is a significant component of reworked material in Facies B making up at least 50 % of the sediment mass.
Fluvial evolution of the Kara Sea Shelf (Chapter 3.5)

The analysis of numerous channels on the Kara Sea shelf led to the conclusion that they were formed during the last global sea level regression phase, having its peak during the LGM. There is a distinct difference between the Ob and Yenisei branch resulting from the different slope gradients of their paleo-drainage paths and possibly different catchment characteristics. Both systems display filled channels, preferable found in shallow water depths, and open or partly filled channels.

Morphometric analysis of channel dimensions showed that they predominantly belong to the meandering river regime. These findings argue for a time interval of subaerial exposure and channel formation sufficient long enough for rivers to incise into the dry fallen shelf area and to adjust under equilibrium conditions.

The paleo-channels’ dimensions are larger than the modern dimensions of recent gauge stations of Ob and Yenisei rivers on land, probably caused by colder climate conditions during LGM resulting in a distinct peakedness of river run-off.

Most of the channels under investigation plot in the meandering river field, they are quite mature in terms of fluvial development. Although an absolute chronology is missing, the more than 700 km northward extent of fluvial features on the shelf, the good conservation of incised channels and their maturity argue for a fluvial-dominated history of the Kara Sea shelf.

It could be feasible that the larger Yenisei paleo-channels evolved from a rapid melt down of the LGM Putoran ice cap enhancing the fluvial discharge of the paleo-Yenisei effectively for several hundred years. It is remarkable that a sudden decay of a relative thin (300 to 400 m) ice cap, could account for the additional water volume necessary to form the larger dimension of the paleo-Yenisei channels.
Zusammenfassung

Rezenter Sediment Transport (Kapitel 3.2)

Oberflächensedimente der Karasee zeigen eine stark bimodale Verteilung in der Schlickfraktion. Die Maxima finden sich zum einen um 8 $\phi$ (4 µm), der dominanten Fraktion, und des weiteren bei 5 $\phi$ (25-40 µm). Der Großteil aller Proben weist diese bimodale Verteilung in unterschiedlicher Ausprägung auf. Das weit verbreitete Vorkommen der bimodalen Korngrößenverteilung in der Siltfraktion weist auf zwei Hauptprozesse für den Materialtransport hin.


Weiterhin haben wir die Magnetische Suszeptibilität (MS) an Oberflächenproben untersucht. Das MS Signal stammt eindeutig aus dem Putoran Gebirge im Einzugsgebiet des Yeniseis, und die Werte der MS nehmen mit zunehmender Wassertiefe ab. Dies deutet darauf hin, dass unverfüllte Flussrinnen, die den tiefsten Punkt in der Morphologie bilden, das MS-Signal entlang des Schelfs „verteilen“. Die Partikel sinken durch die Wassersäule ab und akkumulieren am Meeresboden. Dichte-induzierte Strömungen folgen dem pleistozänen Relief. Dieser Prozess läuft in erster Linie im Sommer ab. Im Winter treten Resuspension und vermehrte Partikelumverteilung auf.

Sedimentation in den Ästuaren von Ob und Yenisei: Entwicklung und Sedimentbilanzierung (Kapitel 3.3)


Berechnet auf der Basis heutiger Suspensionsdaten, akkumulierten während des
Holozäns zwischen 50 % und 75 % der fluviatilen Suspensionsfracht in den Ob und Yenisei Ästuaren. Eine stetige, südwärts über den heutigen Schelf gerichtete Migration des „Marginal Filters“ (Lisitzin, 1995), dem Übergangsbereich zwischen Fluss und Ozean, als Folge des steigenden Meeresspiegels wird favorisiert, um die „fehlenden“ Anteile zu erklären.


Akustische Fazies des inneren Karasee Schelfes: Implikationen für die Sediment-dynamik während des Spät Weichsel- bis zum Holozän (Kapitel 3.4)


Wir haben drei charakteristische Sedimentablagerungsprovinzen identifiziert und auskartiert, sowie die Sedimentmenge für jeden Facies-Typ ermittelt. Fazies A nimmt den Bereich der heutigen Ästuare ein, Fazies B macht den Schelf bis zu einer Wasser-tiefe von 120 m aus und Fazies C die tieferliegenden Gebiete.

Zusammenfassung


Fluviale Entwicklung des Karasee Schelfes (Kapitel 3.5)


Die Dimensionen der Paläo-Rinnen sind größer als die rezenter Flussläufe an Pegelstationen des Ob und Yenisei im Hinterland der heutigen Mündung, was auf kältere Klimabedingungen während des LGM zurückzuführen ist, die in ausgeprägten Abfluss-Maxima resultierten.

Die Dimensionen der Paläo-Yenisei Rinnen sind größer als die Paläo-Ob Rinnen. Es scheint plausibel, dass die größeren Yenisei-Rinnen sich während eines raschen Abschmelzens der LGM Putoran-Gebirgs-Eiskappe entwickelten, was den Flusseintrag des Paläo-Yenisei während einiger hundert Jahre deutlich verstärkte. Es ist bemerkenswert, dass solch ein rasches Schmelzen der relativ dünnen Eiskappe (300-400 m Mächtigkeit) für das enorme zusätzliche Wasservolumen verantwortlich sein könnte, das zur Erklärung der großen Dimensionen der Paläo-Yenisei Flussläufe nötig ist.
Part A

Introduction
1. Introduction

1.1. Aims of this thesis

1.1.1 The Project

This thesis is part of the joint German-Russian project “Siberian River Run-off” (SIRRO) and includes four manuscripts, generally focusing on the interpretation of acoustic sub bottom profiles, measurements of physical properties of sediment cores and on their sedimentological investigations, as well as investigations of the recent sedimentary processes on the shelf. The project SIRRO was initiated to investigate the nature of continental run-off from Siberian rivers, especially Ob and Yenisei rivers (Stein et al., 2003a and further references therein). The study of its behaviour in the Kara Sea and the adjacent Arctic Basin is the main objective of multidisciplinary research of the SIRRO project. Within SIRRO, six expeditions with RV “Akademik Boris Petrov” were carried out in 1997, 1999, 2000, 2001, 2002 and 2003 (Matthiessen and Stepanets, 1999; Stein and Stepanets, 2000, 2001, 2002; Schoster and Levitan, 2003, 2004). Geological investigations concentrated on characterisation and quantification of terrigenous matter. For multidisciplinary geological studies a large number of surface sediment samples as well as long sediment cores were obtained. The first results of geological, geochemical, biological, hydrological and modelling studies were summarised by Stein et al. (2003a). The present study is closely related to the topics of SIRRO subproject E: “Terrigeneous sediment and particulate organic carbon flux: Sources, pathways, sinks, and variability”, and mainly covers the subjects of sediment characterisation, flux and sediment volume calculations as well as paleo-environmental reconstructions.

1.1.2 Overview

Passive continental margins are an ideal area to study depositional and or erosional processes controlled by eustatic sea level fluctuations. In contrast, active continental margins are strongly subjected to hinterland variations in fluvial sediment delivery (Milliman and Syvitsky, 1992). Sea level change has a profound impact on the behavior of a fluvial system leading to total reorganization of the drainage system (Schumm, 1993). Rivers stretch across the shelf, discharging at the shelf edge, depositing drapes and submarine fans (Piper and Nourmark, 2001).

On the Kara Sea Shelf the following factors acted during the late Quaternary

- Sea level rise (changes)
- Glaciations
- Variable fluvial input
- Variable erosion, subaerial exposure of the shelf
In our working area, the sea level change is further complicated by the existence of ice sheets, which may have physically blocked the free fluvial development. During sea level lowstand rivers emerge over the dry fallen shelf. Dynamic sea level changes, especially during transgression, control the shape of sedimentary units, changing the accommodation space available: sedimentary units aggregate and prograde at a pace dictated by the sea level rise (Posamentier and Vail, 1988), diminishing the amount of material supplied to the lowstand deposits. Phases of deceleration resulted in infilling of coastal plains, promoting formation of deltas (Knox et al., 1993).

The thesis addresses recent and ancient sedimentation processes. The late Weichselian history in the Kara Sea has been reconstructed focussing on shelf shaping processes prior to the Holocene during the sea level lowstand of the last glaciation cycle.

High-resolution acoustic profile surveys with a length of more than 20,000 kilometres in total (obtained by several different sounding systems) offer detailed information about the seafloor topography and the thickness and structure of the upper Quaternary sediment cover of the southern Kara Sea. In combination with dated sediment cores we classified the sedimentary packages in terms of sequence stratigraphy. Furthermore, these data offer a substantial basis to calculate a post-glacial sediment mass balance and provide the basis for other studies, especially geochemical investigations (e.g. Stein et al., 2002a, 2003, 2004; Gebhardt et al., 2004). It also delivers topo-physiographic information for further geochemical (e.g. Kodina et al., 2003), paleontological (e.g. Simstich et al., 2003, 2004), and biological (e.g. Fetzer et al., 2003) studies.

Major objectives addressed in the frame of this thesis are:

• to study the dispersal of the fluvially-derived material on the (recent) shelf: what are the pathways of sediment? (CHAPTER 3.2, partly CHAPTER 3.3)
• to establish a Holocene sediment budget (CHAPTER 3.3 and CHAPTER 3.4)
• to interpretate the Holocene evolution of the inner Kara Sea Shelf and late Weichselian (last glacial) to Holocene change in sedimentary processes, sediment flux (CHAPTER 3.3 and CHAPTER 3.4) in relationship to paleoriver discharge (CHAPTER 3.5) and glaciation in the Kara Sea area (CHAPTER 3.4 and 3.5)
• to reconstruct the pathways of the large west Siberian rivers Ob and Yenisei during last glacial maximum sea level lowstand, and their extended onto the shelf (CHAPTER 3.4 and 3.5)
1.1.3 Outline and Structure of the Thesis

Detailed information on special topics are given in each individual chapter. The background information (Chapter 2) concentrates on the physiography, further detailed in Chapter 3.1 by our own mapping, and the hydrography. Chapter 3.1 and 3.2 can be regarded as an extended introduction, representing the first results of our investigations.

Chapter 3.2: Dittmers, K., Niessen, F. and Stein, R. (submitted to GeoMarine Letters). Recent sediment distribution pattern and pathways on the inner Kara Sea Shelf.

Chapter 3.2 deals with physical processes impacting the modern sedimentation on the Kara Sea Shelf, conducted on surface sediment samples. We studied physical sediment parameters characterised by the silt fraction. Secondly magnetic susceptibility was studied to enlighten sediment distribution and pathways on the shelf as a good provenance indicator. Terrestrially derived sediments can be traced by the magnetic susceptibility, while the study of silt properties yield information about the transport mechanisms.


High-resolution echosounding data and several sediment gravity cores taken in the Ob and Yenisei estuaries allow us to calculate the Holocene sediment budget of both rivers and to reconstruct their sedimentary history. Cores were radiocarbon-dated and linked to acoustic profiles using whole-core physical properties obtained by a multi-sensor-core-logger. The entire sediment fill of the present estuaries reflects a deepening-upward sediment growth structure suggesting that sea level rise exceeded sediment accumulation.


In Chapter 3.4 we utilise the chronostratigraphic framework established for the Yenisei estuary and correlate it with profiles of the whole Kara Sea Shelf. We subdivide the shelf into sedimentary-physiographic domains and differentiate their evolution. The paleo-reconstruction reveals important information about the interaction of the LGM ice sheet and the Ob and Yenisei rivers. An important contribution is made by the first
model results that neglect what field data suggest: no isostatic movements exceeding the order of 15 metres.


Chapter 3.5 deals with morphological analysis of widespread channels on the basis of their geometry to deduce the paleo-hydrological conditions. Fluvial dimensions were identified on acoustic profiles. The channels bear valuable information about the riverine extensions onto the Kara Sea Shelf and their interaction with the LGM ice sheet.

The division into different publications focusing on the relevant topic of this cumulative thesis results in some thematic overlaps and repetitions between the chapters of topics, especially in the general introduction/background, which cannot be avoided. Data will be available at the Pangaea database (www.pangaea.de).

In the text numbers of chapters are cross references in this thesis to ease the reader its handling, In the publications they are substituted by the correct citation.

All ages are given in calibrated calender years before present and are abbreviated ka BP (=thousand years before present).

Contributions to other publications


2. Background

2.1 Physiographic Settings of the Study Area

The entire Arctic Ocean (Fig. 2.1) makes up 4.3 % of the world’s ocean area but only 1.4 % of the volume, representing the shallowest (mean depth 1,200 m) of all the major ocean basins of the world (Jakobsson, 2002). Continental shelves occupy more than 52 % of the area of the Arctic Ocean – while average shelves of the worlds oceans only constitute between 9.1 and 17.7 % of the respective areas (Menard and Smith, 1966).

Figure 2.1. Arctic Ocean. Hatched arrows indicate warm Atlantic surface waters, yellow arrows Arctic waters and their circulation pattern (Aagaard and Carmack, 1989). The white line denotes the Last Glacial maximum most proximal ice extent with some uncertainties for the Putoran Mountains (hatched). Major riverine freshwater sources indicated by white arrows. Modified after (Knies and Vogt, 2003) based on IBCAO bathymetry (Jakobsson et al., 2000).
Chapter 2

Figure 2.2 Kara Sea: overview and physiography

In general, the bottom topography of the northern Kara Sea, north of 76° N, is characterised by water depths of more than 50 m (Fig. 2.2) while the South is dominated by shallow waters (with mean depths of 25-30 m) of the “Ob-Yenisei Delta” (Johnson and Milligan, 1968). Two major depressions are found in the northern Kara Sea: the roughly north-south aligned St. Anna Trough (up to 610 m deep) and the Voronin Trough (up to 450 m deep). These troughs with depths greater than 300 m are thought to represent glacial outlet fjords of former glaciations (prior to LGM: Polyak et al., 1997). The Novaya Zemlya Trough is aligned along the east coast of Novaya Zemlya with water depths of up to 350 m and probably represents a sediment starved
The name “Kara” may originate from the Tartian/Kasachian term for “black” possibly due to the Kara Sea’s waters being enriched in humic acids and thus exhibiting a much darker colour than the neighbouring Barents Sea’s clear waters. The Kara Sea (“Karskoye More”) is a semi-enclosed ocean basin bounded by Novaya Zemlya in the North-West, the Siberian Coast in the South and Severnaya Zemlya in the East, while to the North it opens to Central Arctic Ocean waters. In the South-West the Kara Sea is connected to the Barents Sea via the Kara Strait which is 45 km wide, 33 km long and up to 119 m deep. Pavlov and Pfirman (1995) estimate a through-flow of 1,640 km$^3$/y – the same magnitude as the total river run-off into the Kara Sea, while a water volume of 5,000-10,000 km$^3$/y flows from the Barents Sea into the Kara Sea north of Novaya Zemlya (Pavlov and Pfirman, 1995). The Vilkitsky Strait between the Siberian mainland and Severnaya Zemlya connecting Kara Sea and Laptev Sea in the East is characterised by a minimum width of 56 km and depths reaching 210 m. The water flow from Kara to Laptev Sea is estimated to be 4,900 to 11,000 km$^3$/y (Pavlov and Pfirman, 1995).

### 2.2 Hydrography

The whole circumarctic drainage area measures 1.5 times the area occupied by the Arctic Ocean, and is consequently the world’s most land-dominated drainage area (Petersen et al., 2002). The catchment area of the Russian Arctic measuring 12.6 x 10$^8$ km$^2$ makes up 56 % of the area of the former Soviet Union. Run-off totals 2,808 km$^3$/y of which 85% are supplied by Siberian rivers (Romankevich and Artemyev, 1985; Holmes et al., 2002, Romankevich et al., 2003).

River run-off largely contributes to the Arctic Ocean freshwater budget and presently accounts for 50 to 60 % of all freshwater influx into the Arctic Ocean, with saline water import through the Bering Strait between the Asian and American continents, and precipitation in the Arctic Ocean being lower than evaporation accounting for most of the remainder.

The largest river falling into the Arctic based on discharge is the Yenisei river, while
the Ob river is characterised by the largest catchment area. Medianly directed rivers have a characteristic spring-summer flood wave surge moving northwards breaking up ice causing ice dams and high water levels (Aagaard and Carmack, 1989; Carmack, 2000).

Table 2.2. World’s largest rivers’ characteristics (from Degens et al., 1991 and Telang et al., 1991)

<table>
<thead>
<tr>
<th>Rank</th>
<th>Based on drainage area (x1000 km²)</th>
<th>river length (km)</th>
<th>total annual Run-off (km³/y)</th>
<th>Suspended sediment (km³/y)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Amazon</td>
<td>6915</td>
<td>Nile</td>
<td>6670</td>
</tr>
<tr>
<td>2</td>
<td>Congo</td>
<td>3680</td>
<td>Mississippi</td>
<td>6420</td>
</tr>
<tr>
<td>3</td>
<td>Murry</td>
<td>3520</td>
<td>Amazon</td>
<td>6280</td>
</tr>
<tr>
<td>4</td>
<td>La Plata</td>
<td>3100</td>
<td>Yangtze</td>
<td>5520</td>
</tr>
<tr>
<td>5</td>
<td>Ob</td>
<td>2990</td>
<td>Mackenzie</td>
<td>5472</td>
</tr>
<tr>
<td>6</td>
<td>Mississippi</td>
<td>2980</td>
<td>La Plata</td>
<td>4700</td>
</tr>
<tr>
<td>7</td>
<td>Nile</td>
<td>2870</td>
<td>Hiwang Ho</td>
<td>4670</td>
</tr>
<tr>
<td>8</td>
<td>Yenisei</td>
<td>2580</td>
<td>Mekong</td>
<td>4500</td>
</tr>
<tr>
<td>9</td>
<td>Lena</td>
<td>2490</td>
<td>Lena</td>
<td>4400</td>
</tr>
<tr>
<td>10</td>
<td>Niger</td>
<td>2090</td>
<td>Congo</td>
<td>4370</td>
</tr>
<tr>
<td>11</td>
<td>Amur</td>
<td>1885</td>
<td>Niger</td>
<td>4160</td>
</tr>
<tr>
<td>12</td>
<td>Yangtze</td>
<td>1800</td>
<td>Ob</td>
<td>3650</td>
</tr>
<tr>
<td>13</td>
<td>Mackenzie</td>
<td>1790</td>
<td>Yenisei</td>
<td>3490</td>
</tr>
<tr>
<td>14</td>
<td>Ganges</td>
<td>1730</td>
<td>Murray</td>
<td>3490</td>
</tr>
</tbody>
</table>

Table 2.2 gives an overview of the world’s 14 largest rivers, by several criteria. The Ob and Yenisei rivers are characterised by large dimensions and drainage areas, but with moderate discharge and little sediment load. The Ob and Yensiei rivers discharge 404 km³/y and 618 km³/y freshwater on the average, respectively (see Table 2.2 and Figure 2.3 for details). Significant portions of the drainage area of both rivers are dominated by permanent and seasonal permafrost. The recent discharge of total suspended matter by the Ob and Yenisei rivers averages to 15.5 x 10⁶ t/y and 4.7 x 10⁶ t/y, respectively. For the Yenisei, a total suspended matter supply of 14.4 x 10⁶ t/y which is about three times higher than the modern value, has been measured for times preceding the dam constructed near Krasnoyarsk in 1967 (Telang et al., 1991; Holmes et al., 2002, see Table 2.2). In Figure 2.3 the average discharge is compared to data obtained during the SIRRO expedition 2001 (Gebhardt et al., 2004).

The enormous water release into the Kara Sea leads to the establishment of a halocline below a relatively thick low salinity surface layer extending up to approximately 80° N. The total annual run-off would accumulate to a 1.5 m thick layer of freshwater for the Kara Sea area (Gebhardt et al., 2004). Thus, the whole system can be seen as
an estuary. The existence of the giant freshwater layer with its associated halocline has profound impact on the sediment distribution on the shelf, but also enhances sea-ice formation and influences primary production.

Figure 2.3. Water and sediment discharge of Ob and Yenisei rivers after Gebhardt et al. (2004). Long term mean water discharge are given in (a) Yenisei and (b) Ob and sediment discharge for Yenisei in (c) and Ob in (d).
2.3 Research History

Seibold (2001) gives an overview on the exploration and map development along the Arctic coast of Siberia and an example of an early map drawn in 1721 of our working area (Fig. 2.4). The coasts west of the Ob estuary have been fairly well known from about 1600 B.C. on, mainly by merchants, sailors and fishermen and finally by British and Dutch as well as some land-based expeditions. In the beginning of the 16th century first fortified towns along the major rivers evolved (Tobolsk on the Irtish: 1587, Tomsk on the Ob: 1604, Yeniseyesk on the Yenisei: 1619). These towns became

Figure 2.4. Historical map of Seuter from 1721 of the working area taken from Seibold (2001)
exploration centres from where investigations started down river and along the coasts. First scientific data on the Kara Sea were collected by the Swedish expeditions of Nordenskjöld in 1875 (on the Pröven), in 1876 (on the Ymer) and in 1878 (on the Vega). In 1893, the Kara Sea was surveyed by Nansen’s famous Fram, followed by several smaller pushes of several expeditions. Since the early 1920’s, the Russians intensified their investigations strictly keeping other countries from entering the area. Thus the Kara Sea became “Terra Incognita” again for other nations during Soviet times for two reasons: 1. nuclear weapon testing on Novaya Zemlya and dumping of nuclear waste in the Novaya Zemlya trough, 2. the discovery of gigantic gas resources in the South Kara Sea basin situated in the southern Ob estuary, a gas-prone province at least for the Cretaceous and Upper Jurassic section. Russian scientific investigations have been undertaken since the 1950’s but with only little output in western literature (Kulikov and Martynov, 1961). The „Atlas of ground of the Arctic Ocean“ (Treshnikov, 1985) summarises this knowledge, consisting of a large suite of maps including generalised maps and contour plots. However, the data source is uncertain. In 1965, the first „western“ ship entered the Kara Sea again conducting the Northwind expeditions (Johnson and Milligan, 1968). The first multidisiplinary, modern investigation was carried out by RV “Dimitry Mendeleev” in 1993 (Lisitzin et al., 1995).
Part B

3. Results

3.1 Bathymetry
3. Results

3.1 Bathymetry

A detailed high-resolution elevation model was not available for our working area, thus we modelled the Kara Sea on the basis of sea-bottom depth soundings of the ELAC sub bottom profiler, calibrated to mean sea level, conducted during RV “Akademik Boris Petrov” expeditions in 1999, 2000 and 2001 (Stein and Stepanets, 2000, 2001, 2002) (Figs 3.1 and 3.2). Combined with soundings of nautical charts great areas have been completely remapped. The influence of tides is neglectable with a ranging from 0.3 m to 0.5 m (Pavlov and Pfirmann, 1995). In order to combine available data of the Arctic Ocean and our data we merged our data set with the IBCAO data set (Fig. 3.2, Jakobsson et al., 2001). For both datasets, (Latitude, Longitude, and water depth) data points were simply merged and a gridfile created with the Generic Mapping Tool (GMT,

The resulting elevation model (Figs 3.2 and 3.4) was carefully cross-checked with our soundings and a clear improvement in vertical resolution was achieved (Fig. 3.3). In combination with GIS tools, e.g. Global Mapper, was useful for area calculations and estimates of water volume as well as for the construction of morphological sections of the shelf.

Figure 3.2. Isobath map of the new elevation model. Note the hummocky relief at the shelf break and in the NE area of the working area. Equidistance until the 150 m isobath: 10 m. The data presented here are available at www.pangaea.de.

**Data Basis**

A total of 10,000 ship’s soundings and some additional 8,000 soundings obtained from the nautical charts were digitised and used to compute the elevation grid (Figs 3.2 and 3.4). Geo-referenced soundings were gathered generally every 5 minutes.
(ship’s time), with a higher frequency at diversified terrains. The nautical charts were geo-referenced, transformed, rectified, and digitised subsequently by KK+W (Kiel, Germany). Isolines were created using the “tin contour” function of ARC Info. Conflicts arising from the different data sources were addressed as follows: the first problem encountered was the different scale, resolution, and cut-outs from the nautical charts, which is especially problematic in combination with our echo-soundings. Thus, a priority of data sets was established: the primary category used are the echosoundings of our expeditions with RV “Akademik Boris Petrov”. Then, according to their resolution, maps 11133 and 11124, and sheet 10103 at last, displaying the lowest resolution and lowest density of soundings, were used. Additionally, in order to emphasise our soundings, where we have the best confidence in data quality, a buffer of 1.5 km width was established around the ship’s track lines beyond which all other soundings were excluded.

**First Implications from new elevation data**

Figure 3.2 displays the dendritic pattern of the paleo-Yenisei branch: This pattern evolved during times of sea level lowstand when the Yenisei ancestor incised into the seabed of the Kara Sea. This process is described and interpreted in detail in Chapter 3.4 and Chapter 3.5.

The seabed roughness shown in Figure 3.4 gives a first impression of the shape of the
Figure 3.4. Elevation model of our data, visualised with Global Mapper (VERSION 5.10). In this illustration the sea-floor, area that lies underneath 0 m, elevation gradient is shown as shading. Note that large areas of the shelf are rather planar. Formerly glaciated terrain is identifiable by irregular shadings in the NE of the working area north of 76° N and in the west around 70° E.

Kara Sea sea bottom. It is remarkable that the Kara Sea can be subdivided visually into different provinces. In Chapter 3.4 these provinces are related to paleoenvironmental processes that shaped them. Hummocky relief is for example produced by the action of glaciers and ice sheets that constantly advanced and retreated and deformed the surface. This formerly glaciated terrain is in evidence by irregular shadings in the NE of the working area north of 76° N and in the west around 70° E. Whereas the shelf and the modern estuaries are rather planar over large distances.

Data are available at www.Pangaea.de.
3.2 Recent sediment distribution patterns and pathways on the inner Kara Sea Shelf (Siberia)
3.2 Recent sediment distribution patterns and pathways on the inner Kara Sea Shelf (Siberia)

Dittmers, K., Niessen, F. & Stein, R.
Alfred Wegener Institute for Polar and Marine Research, Columbusstrasse 2, 27568 Bremerhaven, Germany

Abstract

We studied the silt fraction on the shallow (20-30 m) continental shelf of the Kara Sea, Siberia, with the Sedigraph 5100 in order to enlighten sedimentary processes. Magnetic susceptibility of surface sediments was investigated to reconstruct recent transport pathways of magnetic minerals indicative of terrestrial sediments. From the distribution of magnetic susceptibility, we reconstructed a gravity driven sediment, guided by incised channels on the shelf, representing topographic lows.

Surface sediment samples exhibit a prominent twofold grain-size distribution. The peaks cluster around $8\phi$ (4 $\mu$m), the dominant mode, and at a secondary peak at $5\phi$ (25-40 $\mu$m). The majority of all samples display both peaks unequally strong pronounced. The twofold grain-size peaks strongly suggest two peaks in current intensity, probably related to at least two different processes.

It is evident that sediment pathways follow the morphology of the sea-floor, indicating density-driven bottom layer transport: magnetic susceptibility values abate away from distributary channels. The distribution pattern of magnetic susceptibility in dependence on water depth can be seen as a dilution indicator for the river-derived material. The susceptibility signal clearly originates in the Putoran Mountains in the drainage area of the Yenisei branch. The susceptibility values decrease with increasing water depth to the North. Basically there are two atmospheric regimes driving the water circulation and thus the sediment distribution in the Kara Sea. this results in enhanced particle input by rivers during the summer and prevailing erosion during winter.

Keywords: surface sediments, silt fraction, terrigeneous fraction, Kara Sea, river run-off

1. Introduction

Continental shelves play an important role for the distribution and “digestion” of fluvially-derived fines, by advection and particle settling (Windom and Gross, 1989). It is estimated that less than 10 % of river-supplied particles are transported across the shelf and reach the deep sea (Drake, 1976; Eisma, 1982; Postma, 1990). Therefore they act as an interface between land, river and ocean (Rudels et al., 2000).

In the Arctic realm terrigeneous sediment supply is controlled by river discharge, sea-ice transport and coastal erosion (e.g. Darby et al., 1989; Stein and Korolev, 1994; Rachold et al., 2003). Particularly the shallow waters of the shelf are subject to reworking by press ice ridges and enhanced water circulation due to ice formation (Reimnitz et al., 1992, 1993, 1994; Dethleff, 1994; Eicken et al., 1997, 2000). On the other hand shallow coastal and estuarine areas are effectively sheltered by fast-ice
from vigour wave-induced currents from autumn to spring (Divine et al., 2003, 2004). Biological productivity in the water column is relatively low compared to other oceans due to the nearly permanent sea-ice cover and low temperatures (Subba Rao and Platt, 1984; Grebmeier, 1995; Stein, 1996). Therefore, the mineralogy parameters of sediments can be used to study the imprint of processes on the sedimentary record without significant distortion (Wahsner et al., 1999; Schoster et al., 2000). So far granulometric investigations of the major components have focused on bulk sand-silt-clay fractions (Steinke, 2002; Stein et al., 2004). In the inner Kara Sea recent fresh water and terrigeneous sediment fluxes are quite well-studied (Telang et al., 1991; Gordeev et al., 1996; Gordeev, 2000; Holmes et al., 2002). The incoming suspended sediment masses are also investigated by Lisitzin et al. (1995) and recently by Beeskow et al. (2003) and Gebhardt et al. (2004). We intend to study and characterise transport ways and mechanisms of fine-grained, riverborne particulate matter. The main focus of this paper is the investigation of suspended sediment dispersal in the inner Kara Sea (up to 78° N).

First of all we studied physical sediment parameters characterised by the silt fraction best suited to study suspended material processes and the physical factors controlling sedimentation. The silt fraction resembles the grain-size fraction that currents are capable to mobilize (McCave et al., 1995). Magnetic susceptibility was studied to highlight sediment distribution and pathways on the shelf, as tracer and important indicator for sediment flux. In this study we expand our existing surface data (Dittmers et al., 2003; Stein et al., 2004) by stations especially north of 75° N. Magnetite, titanomagnetite or maghemite (Thompson and Oldfield, 1986) are the main carriers of magnetic susceptibility. Variations in magnetic susceptibility mainly reflect compositional variations in the terrigeneous source material: the Putoran Mountains-originated fraction of the Yenisei River (Dittmers et al., 2003; Stein et al., 2004) (Fig. 1). The upper Permian to lower Triassic basalts lavas and tuffs of the Putoran mountains are made up of up to 2 km thick basalts covering an area of 1,800 km striking NE to SW and of about 1,000 km width (Sharkov, 1980).

According to Cesium isotopic studies all surface samples are recent (Stepanets et al., 2000 in Stein and Stepanets, 2000) and exhibit no significant older ages.

**Background: Environmental settings**

In the summer sediment transport processes are dominated by a northwest, clockwise directed surface flow consisting of a top flow fed by the strong fluvial freshwater and sediment supply (Fig. 1) and a highly saline, denser and colder “counter-current” intruding into the shallow shelf through depressions formed by channels (Chapter 3.5)
Recent sediment distribution pattern and pathways on the inner Kara Sea Shelf

Figure 1. Working area with position of annual polynia, surface circulation and salinity after Pavlov and Pfirman (1995). Position of Polynia after Polyakov and Johnson (2000) and Divine et al. (2003, 2004) indicated by grey shaded area; arrows indicate prevailing summer surface currents; summer surface salinity isolines indicate relative dilution of fluvial freshwater.

into the estuaries as visible in the bottom salinities (Fig. 2a) (McClimans et al., 2000; Simstich et al., 2003). Winter surface circulation is the opposite with anti-cyclonal flow (McClimans et al., 2000), and reverses the bottom currents to a net outflow (Harms et al., 1999, 2000). The total winter discharge accounts for only 5-10 % of the total river run-off (Antonov, 1970; Holmes et al., 2002). Harms et al. (2000, 2002, 2003) developed an atmospheric driven circulation model for the Kara Sea and also computed bottom current speeds (Fig. 2b). The highest current speed is observed in the shallow waters of Ob and Yenisei, as well as in near-coast regions.

Besides their water loads the Arctic rivers also transport large amounts of dissolved and suspended material towards the ocean (Gordeev et al., 1996; Holmes et al., 2002; Rachold et al., 2003). The modern annual discharge of total suspended matter by the Ob and Yenisei rivers is $15.5 \times 10^6$ t y$^{-1}$ and $4.7 \times 10^6$ t y$^{-1}$, respectively (Holmes et al., 2002; Telang et al., 1991). Based on filtration studies of the 2001 expedition, Gebhardt et al. (2004) estimate that the Yenisei annually delivers $5.0 \times 10^6$ tons of sediment to the Kara Sea, while in contrast the Ob river estuary “absorbs” 75 % of the fluvial
supplied material and releases $3.76 \times 10^6$ tons of sediment into the Kara Sea.

Basically the suspended matter is trapped in the estuaries where a freshwater wedge overlies the sea water. Mixing with salt water promotes rapid precipitation of fine-grained suspension due to coagulation processes, resulting in high sediment accumulation rates (Dittmers et al., 2003). Approximately more than 90% of the suspended matter and about 30% of the dissolved matter accumulate within this “marginal filter” (Lisitzin, 1995). Accumulation rates rapidly decrease north of about 73° 30’ N (Stein et al., 2003b, 2004), as visible on acoustic records (Dittmers et al., 2003). From the Mississippi mouth, for example, it is also well known that concentrations in the river plume are reduced by 90% within 5–10 km of the river mouth (Hill et al., 2000). Recent research showed that estuaries not only act as chemical boundaries, but also influence the settling behaviour of suspended matter due to a rapid change of hydrological/physical settings (Dyer et al., 2004).

Figure 2 a). Bottom salinity gradients of Simstich et al. (2003)

Figure 2 b). Modelled bottom velocity of atmospheric forced shelf model after Harms et al. (2000, 2003)
2. Methods

Surface sediment samples were obtained during three expeditions with RV “Akademik Boris Petrov” carried out in 1999, 2000 and 2001 (Stein and Stepanets 2000, 2001, 2002). Wet surface sediments were obtained with the multicorer, giant box corer and an ocean grab (Stein and Stepanets, 2000, 2001, 2002). Priority of station selection was to sample high Holocene thickness sites.

The bulk organic carbon contents of surface sediments of the Kara Sea Shelf are moderate ranging from 1 % with to values around 2 % in the estuaries, along depression and in the vicinity of the coast (Stein and Fahl 2003, Stein et al., 2004). Towards the North the shelf values lie below 1 % in the working area. This fact eases silt-size sample analysis and preparation (Stein, 1985) and minimizes difficulties due to coagulation effects. Biogeneic opal content in Kara Sea sediment is < 2 % (Stein and Nürnberg, 1995) and can be neglected.

Silt-size analysis
The sand fraction (> 63 µm) was separated by wet sieving, while the remainder fraction (< 63 µm) was collected in 50 ml plastic bottles. From the fine fraction calcium carbonate was removed by 3 % acetic acid and organic content was oxidised by 10 % hydrogen peroxide (standard AWI procedure, for details see Hass, 2002). The carbonate-free samples are regarded as the terrigeneous fraction hereafter. For ideal dispersal after this treatment sediment was put 24 h on a shaker and spiked with sodium polyphosphate. The grain-size distributions was determined using a Micromeritics Sedigraph 5100 over the 1.5 to 63 µm range measuring the grain-size distribution of silt in 1/10 \( \phi \) steps. Analytical details and errors are described by Stein (1985), McCave et al. (1995) and Bianchi et al. (1999).

After grain-size analysis samples were dried and weighed for calculation of weight percentages of the sand, silt and clay fraction. For the silt fraction (McCave et al., 1995) cumulative curves of the sedimentological parameters were plotted and various statistics characterising the grain-size distribution were computed (Pettijohn, 1975). Statistical parameters such as arithmetical mean, modal size (peak occurrence), sorting (= standard deviation) and skewness were determined after the method of moments (Krummbein, 1936).

Magnetic Susceptibility (MS)
Magnetic susceptibility is defined as the dimensionless proportional factor of an applied magnetic field in relation to the magnetisation in the sample of a known volume (expressed in SI units; Thompson and Oldfield, 1986; Pirrung et al., 2002). The volume
magnetic susceptibility of the freeze-dried bulk samples (10 cm³) were measured after the method of Niessen and Weiel (1996) at high (4.6 kHz, hf) and low (0.46 kHz, lf) frequencies in SI units using a susceptibility control unit (MS2B Bartington Ltd, UK), before samples were proceeded for the silt-size analysis.

On the basis of frequency-dependent magnetic susceptibility measurements in situ formation of ultra fine (<0.03 mm) super-paramagnetic and ferromagnetic minerals produced by bacteria and chemical processes (Thompson and Oldfield, 1986; Chang and Kirschvink, 1989; Dearing, 1994) can be neglected. Aeolian input in our working area can be disregarded (Niessen and Weiel, 1996; Shevchenko et al., 2003). In this study we use the porosity-corrected susceptibility, calculated according to Niessen and Jarrard (1998) in order to account for the pure grain signal. MS of the interstitial pore water is negligible (Thompson and Oldfield, 1986, Dearing, 1994).

3. Results

All stations were classified on the basis of high-resolution echo-sounding surveys (Chapter 3.3-3.5) and sub-divided into three sea-floor morphology environments: estuaries, shelf and channels on the shelf (Fig. 3a). This differentiation was made in order to account for different environmental influences on the sedimentary regime. The estuaries are restricted to the Ob and Yenisei river mouths up to 74° N. The shelf occupies the rest, cut occasionally by channels with variable Channel depth (Chapter 3.5).

Grain-size

Particles are fine-grained and silt-size particles are the dominant fraction (Figs 3a and 3b), with contents varying between 38% and 100%. From the total of 74 samples silt is the dominant grain-size.

Figure 3a. Total sand-silt-clay size characterisation (Shepard, 1954) of surface sediments; note difference in symbols of shelf (circles) and estuaries (hatched circles), while channels are represented by triangles; Samples were obtained during year 1999, 2000 and 2001 expeditions (Stein and Stepanets, 2000, 2001 and 2002, respectively) channels on the shelf shown as triangles and samples from the estuaries are hatched, different colors and size of symbols represent grain size classes.
Figure 3b. Overview of stations and total grain size characterisation (Shepard, 1954) of surface sediments; classification after Figure 3a, note that there is no difference in symbols of shelf and estuaries; channels are shown as triangle and estuaries are hatched. Samples were obtained during year 1999, 2000 and 2001 expeditions (Stein and Stepanets, 2000, 2001 and 2002, respectively).
Figure 4. Mass percentage of silt content; Holocene sediment thickness after Dittmers et al. (2003). Position of sediment trap Yen 01 (1999) and Yen 03 (2000) denoted by star, for further informations see Gaye-Haake et al. (2003)
Figure 5. Modal silt-size of surface samples. Stars denote symmetrically distributed samples and thick edges coarse skewed samples. For detailed silt size distribution see Figure 6.
fraction (> 50 %) in 61 samples (Figs 3a, 3b and 4). North of the coastline in the NE working area, a band trending from the SW to NE coast parallel is characterised by relatively low values, while towards the north and west values increase slightly (Fig. 4).

Figure 3a exhibits the grain-size classification and distribution after Shepard (1954). Sand-rich sediments are concentrated in the recent estuaries and on the shelf of Ob and Yenisei rivers displaying the shallowest waters and highest current velocity. The shelf is more mud-dominated. Lowest sand content is found in channels on the shelf. The distribution of the coarse fraction shows the same tendency as the model results of Harms and Karcher (1999) and Harms et al. (2000), where shallow water depths are related to high current speed (Fig. 2b).

**Modal silt-size**

We characterise the silt-size fraction mainly by the modal (dominant) grain-size, i.e. the grain-size making up the main peak. The modal silt-size, as well as the bulk silt content show much variability throughout the working area (Fig. 5). In the estuaries, high values are observed in the southern Ob Estuary south of 70° N and in the Yenisei Estuary up to 73° N. To the North, facing the open shelf, modal grain-size decreases markedly in the estuaries. The area along the southern and north-easterly coast and around the islands is characterised by uniform low modal grain-size > 4 µm. Coarser modal grain sizes are found north of the Ob mouth at 75° N with three samples of 35 µm size, and in the north-western area north of 76° N.

**Statistical parameters**

The mean silt values of all samples range from 14 to 35 µm, which is medium to coarse silt. Again highest values with the highest variation occur in the estuaries, while low values are observed north of the estuaries and in the nearshore NE eastern area. On the shelf medium silt is predominant with values average between 20-30 µm, and shows a very equal distribution. Sorting displays very similar values of 0.2-0.5 (ϕ classes), i.e. very well to well sorted; this is an index of dispersion/scatter of the population. Skewness, an index of symmetry of the grain-size distribution, is symmetrically for the estuaries. North of the Ob, samples are coarse-skewed (negatively), while to the East samples becomes more fine-skewed, i.e. symmetrically to positively skewed.

**Frequency Curve of silt-size sediments**

The frequency curve of silt-size material is shown in Figure 6, ranging from 4 ϕ (finer than sand) to 9 ϕ (clay). In most silt populations there is the tendency to show
a peak in the fines at 8 \( \phi \). Additionally, there is another sub-population at 4-5 \( \phi \) (62-31 \( \mu \)m), variably pronounced.

Figure 6a shows examples of the estuaries displaying a distinct bimodality, and a pronounced maximum on the coarse fraction at 4-5 \( \phi \). In comparison the examples from the shelf are much more fine-tailed with a suppressed secondary mode at 4-5 \( \phi \) (Fig. 6b). The green-coloured curves are examples from the NW corner of the working area, characterised by high water depths. The curves display rather uniform distribution with minor “current shapening”. Examples from channels on the shelf show

![Graphs showing silt-class distributions for shelf, channel, and estuary.](image-url)

Figure 6. Silt-class distributions of all samples. Three areas were differentiated: A) shelf, B) channel, C) estuary. Thick lines show in A) refer to normally distributed samples, situated in the northern shelf area characterised by higher water depths and shown as stars in Figure 5. Coarse skewed examples (negatively) are shown in blue edge thick in Figure 5, in C) hatched lines depict fine skewed examples. The thick black line displays the mean values of all grain-size populations for each differentiated facies provinces.
Figure 7. Distribution pattern of porosity corrected susceptibility
a trend similar to those in Figure 6b for the shelf, with the exception that the scatter between individual samples is much lower and that there is only a faint coarse mode detectable.

**Magnetic susceptibility**

The porosity-corrected magnetic susceptibility of surface sediments shows two major trends (Fig. 7). The first is that values from the Yenisei river are significantly higher than in the Ob area with averages at 3,000 to 4,000 \( \times 10^{-9} \) SI and <1, 000 \( \times 10^{-9} \) SI, respectively. The other important observation is a distinct decrease in values occurring north of 74° N. Despite the distinct decrease off the Yenisei estuary some samples north of 74° N show slightly increased susceptibility compared to other areas of the southern Kara Sea. Values are higher east of 76° E indicating the Yenisei water’s influence. North of 76° N another decrease in the values of magnetic susceptibility is obvious but not as pronounced as for the estuary-ocean transition.

### 4. Discussion

Currents capable of moving/dispersing sediments normally affect the clay and silt fraction in the marine environment. This is clearly underlined by the fact that surface sediments of the Kara Sea are dominated by the fines with generally minor sand components (Figs 3).

In the following section processes effecting the grain-size distribution are reviewed.

**Sedimentation by grain coagulation**

Particles below 10 µm tend to behave cohesively, because clay minerals with their charge imbalances are small enough to be effected by van de Waals forces (McCave, 1985; Weaver, 1989). Therefore, only a measurement of (disaggregated) individual medium to coarse silt particles are a reliable reflection of sediment sorting processes in marine environments (Bianchi et al., 2001). The silt fraction is traditionally used as a paleocurrent indicator (McCave et al., 1995; Bianchi et al., 1999, 2000), concentrating on deep sea environments so far (Hall et al., 1989; Hass, 1996, 2002; McCave and Carter, 1997).

In general the peaks are thought to result from distinct flow speeds responsible for deposition, but nevertheless it is possible that grains aggregate to large flocs resulting in a “virtual” grain-size, with hydraulic behaviour of a larger size. It is known that „marine snow“ is formed by large aggregates of relatively fine, smaller than 10 µm material. The flocculation process is reviewed by McCave (1984) and Eisma (1986). In
estuarine to shelf marine environments at least parts of the suspension is flocculated (Kranck, 1984; Eisma, 1986; Wells and Shanks, 1987). Flocculation should produce an unsorted signal, because particles, regardless of their size coagulate and fall out of the suspension randomly. Fines could be effectively concentrated in “fluid muds”, a process recently observed on river fed shelves (Wright et al., 1990; Kineke et al., 1995). This is a phenomenon of salt induced particle settlement due to coagulation, formerly relegated to estuaries only, as demonstrated in the “marginal filter” (Lisitzin, 1995) of the Ob and Yenisei estuaries.

However, our samples show distinct peaks at defined intervals and contradict a pure flocculation settlement. The abundant fine peak could be the result of preferential agglomeration of the fine fraction.

Environmental factors shaping silt populations

Summer mode: River Run-Off

River-flush events are important to “inject” the suspended material into the system with subsequent particle settlement out of the sediment-plume. Widening of estuaries reduces current speed/flow and results in sedimentation. Reaching the ocean the buoyant river plume is subjected to less friction at its bottom boundary and particle settlement is physically enhanced (Hill et al., 2000; 2001). In addition flocculation accelerates particle settlement. However, although the flocs behave hydrometrically as sand or coarse silt, once deposited they behave like muds, i.e. they show higher resistivity towards eroding currents (Zanke, 1982). River-induced surface currents are too weak for the resuspension of agglomerated bottom sediments. The only possible source of resuspension are the induced bottom counter-currents (Harms et al., 2000). With a maximum speed of 12.5 cm/s they are rather capable of holding the suspension (Harms et al., 2000). However, we think that the discharge peak observed during 3 months, when the Ob and Yenisei rivers release 70-80 % of their annual discharge (Holmes et al., 2002), has profound impact on sediments in the estuaries, because stream power is increased markedly. These events could explain the coarse peak (Fig. 6a) for the estuarine environment. Furthermore, during spring melt ice floes can create ice jams resulting in river damming with subsequent dynamic break-ups as especially known from the Yenisei river (AMAP assessment report 2002). However, these events additionally boost up the stream power at the peak discharge.

During the rest of the year, especially north of the estuaries, suspended particle settling and distribution prevails. The frequency curve of silt-size distribution curves for the shelf and channels (Figs 6b and 6c) show very similar patterns. A marked difference is that the scatter for the shelf is much broader (Fig. 6a) as for examples
from the channels that show very uniform behaviour (Fig. 6b).

**Winter mode: Sea-ice-related Processes**

Shore fast-ice is formed from October to December bound to shallow water depths of 20-30 m (Polyakov and Timokhov, 1994; Divine et al., 2003). The surface water beneath flaw leads is turbulent with convective cells and Langmuir-type circulation (Dethleff, 1994; Dethleff et al., 2000). Current speed may reach 10-15 cm/s strong enough to hold suspension and possibly capable to erode or resuspend deposited material. It is not surprising however, that the highest current speed is observed in late October from current meters attached to sediment traps (Fig. 8). The fine signal at 8 φ (Fig. 6) could either reflect a direct transport signal or an “opportunistic” sedimentary

![Figure 8. Current speed measured at the Yenisei mooring station; the current meter was employed 15 m above the sea-floor; for positions see Figure 4. It must be kept in mind that the current meters were deployed 15 m above the sea-floor and that the vigour of indicated currents is diminishing to the sea-floor, because of boundary frictional of the flow (Zanke, 1982); Note that traps were deployed during slightly different times of the year but that the major trends are similar. Data by the courtesy of B. Gaye and N. Lahajnar, IFBM Hamburg, Germany; Schoster and Levitan, 2004](image-url)
signal from particle settling out of the water column. The estuaries and NE nearshore area are ice covered during most of the year. These conditions result in high mud and bulk silt content in the surface sediments as well as a fine skewed modal size spectrum. These areas are probably effectively sheltered from storm wave action and from large-scale water circulation (Divine et al., 2003, 2004). In contrast north of the Ob River, another facies is established related to more open water conditions during most of the year. Here the modal grain-size values are coarser. This could be related to the fact that the sheltering fast-ice is missing.

Sea-ice can be excluded as transport agent for coarse material (Fig. 6a), because it is freshly formed each year and not of glacial origin with incorporated debris. Evidence from the Laptev Sea (Dethleff, 1994; Dethleff et al., 2000) indicates that preferentially fine material is frozen in during suspension freezing.

Samples north of 76° N are characterised by higher water depths (Fig. 5) and are not as susceptible to surface currents and show a unimodal distribution (green curves in Figure 6). The bimodal grain-size distribution curve strongly suggests twofold peaks in current intensity, probably related to at least two different processes. On the basis of our data it is impossible to relate a distinct grain-size fraction to a single transport process. There is some bias to identify the "exact" processes.

Sediment pathways

The pathways of terrestrial sediment signal and dispersion, mapped by magnetic susceptibility, strikingly resembling the established palaeoriver drainage net indicated by arrows in Figure 7 described in more detail in Chapter 3.5. This is probably related to density-driven sediment flux by a bottom nepheloid layer (McCave, 1972; Nittrouer and Wright, 1994, Pak et al., 1980), predetermined by ancient incised channels. It is proven that grain-size distributions in turbidity currents show the same composition as bottom nepheloid layers on the adjacent shelf (McCave; 1972; Pak et al., 1980). A residual, lag deposits, origin of the relatively high magnetite values along palaeo-river pathways in channels thalwegs, can be excluded, mainly because of the high sediment thickness of the well documented Holocene marine infill as visible in acoustic profiles (Stein et al., 2002, Dittmers et al., 2003, Chapter 3.4 and 3.5).

The overall average values for the estuaries, the channels and the shelf are 1.915 x 10⁻⁹ SI, 1.398 x 10⁻⁹ SI and 1.159 x 10⁻⁹ SI, respectively. We see this as evidence that the channels distribute the MS signal on the shelf. After the particles have made their way through the water column they settle and accumulate in density-induced flow following the bottom Pleistocene relief. This process is active mainly during summer. During winter resuspension, enhanced particle reorganisation occurs. The
resuspended material could settle again and flow downward the bottom topography. Sea-ice formation accelerates circulation, especially at its marginal polynia, while in contrast the coast-near waters appear to be quite well sheltered by fast ice, despite their shallow water depth.

**Channels**

The fact that the highest fine-silt content is found inside the channels means that the flow speed in the channel axis is sufficiently strong to keep the fines in suspension temporarily. It is very likely that these currents decrease very gradually. Each slowing step in the current flow results in deposition of finer material. If the flow turns on again the compaction and bounding forces of the fines are strong enough to counteract/resist the erosional force (Bagnold, 1968; Gibbs et al., 1971; Zanke, 1982). The shelf examples show similar silt distribution, but with a more pronounced coarse mode (Fig. 6b). It is likely that channels are more sheltered from the vigour of currents and additional fines are “collected”.

**5. Conclusion**

In most silt populations of the investigated inner Kara Sea Shelf sediments, there is the tendency to show a peak in the fines at $8\,\phi$ (4 µm). In some samples another sub-population at $5\,\phi$ (25-40 µm) may be variably pronounced. Samples from the estuaries display a distinct bimodality, and a pronounced maximum in the coarse fraction at 30 µm. In comparison, the examples from the shelf are much more fine-tailed with a suppressed secondary mode at 5-6 $\phi$ (30 µm). Examples from channels on the shelf show a trend similar to that for the shelf, but the scatter between individual samples is much lower and there is only a faint coarse mode detectable. The two-fold grain-size peaks strongly suggest two peaks in current intensity, probably related to at least two different processes.

The Ob and Yenisei estuarine grain-size peaks are either transport signals displaying a river run-off shadowing a background sedimentation producing a secondary mode in the fines, or could be related to enhanced flocculation in the mixing zone, producing a peak in the coarse fraction. Samples from the shelf are all dominated by a fine mode at $8\,\phi$ with a minor peak at $5\,\phi$.

The distribution pattern of terrestrial-derived sediment can be demonstrated by the magnetic susceptibility, while the study of silt properties yields information about the transport mechanisms. We strongly promote the existence of a „summer“ and a „winter mode“ of ocean circulation on the shelf, reflecting the bimodal grain-size distribution. In the summer the whole system is controlled/influenced by strong riverine water and sediment influx. The sediment plume is controlled by the prevailing surface currents,
mainly induced by winds, before the fines settle and sediment on the sea-floor. The winter situation is characterised by little direct sediment and freshwater input but due to the high current energy, ice transport and ice erosion the winter rather marks a period of intense sediment redistribution, erosion and resuspension. Coastal, nearshore and estuarine areas are effectively sheltered by the fast-ice from vigour wave-induced currents and finer grain-size populations are dominant.

The denser (re)suspended material follows the deepest points in the bottom topography, especially in the NE. These features are former fluvial channels formed during sea level lowstand, forming a nepheloid layer that is more sediment-saturated than the surrounding water masses. This process may be enhanced during sea-ice freeze-up with the residual ocean water being more dense thus sinking to the sea-floor (“brine formation”). One major task for the future is to examine factors enhancing bottom shear stress and associated sediment resuspension using in-situ measurements covering several seasons and current regimes during the year to characterise the hydrodynamic and sedimentation conditions. A proposal in order to further enlighten sedimentation processes in cooperation with the University of Cambridge (I. McCave) has been submitted (DAAD grant No D/05/10823).

More sophisticated analysis of the polymodal distributions in order to distinguish between different responsible processes will also be useful. The numerical-statistical analysis introduced by Weltje (1997) has proven is an ideal tool for “unmixing” grain-size populations (Frenz, 2003; Holz, 2005). However, this is beyond the main subjects of this thesis, but remains a valuable topic for future research.

**Acknowledgements**

We are grateful to the crew of RV “Akademic Boris Petrov” for their support throughout the cruises. Many thanks to Jens Matthiessen and Christoph Vogt for their helpful comments. Special thanks to Frank Schoster, Matthias Kraus and Christoph Vogt for many fruitful discussions. Ingo Harms is thanked for explaining the physics beyond his model profoundly and patiently. This study has been performed within the German-Russian research project “Siberian River Run-off (Sirro)”, Financial support by the German Ministry of Education, science, Research and Technology (BMBF) and the Russian Foundation of Basic research is gratefully acknowledged. Data are available at www.pangaea.de
3.3 Holocene sediment budget and sedimentary history of the Ob and Yenisei estuaries
3.3 Holocene sediment budget and sedimentary history of the Ob and Yenisei estuaries

K. Dittmers, F. Niessen, R. Stein
Alfred Wegener Institute for Polar and Marine Research, Columbusstrasse 2, 27568 Bremerhaven, Germany

Abstract

High-resolution acoustic data and several sediment gravity cores taken in the Ob and Yenisei estuaries allow us to balance the Holocene sediment budget of both rivers and to reconstruct their sedimentary history. Cores were radiocarbon dated and linked to acoustic profiles using whole-core physical properties.

The Ob and Yenisei estuaries, with their sea water fresh water mixing zone, act as major sediment sinks for fluvially-derived terrigeneous material in Holocene times. Most of the suspended and large amounts of dissolved matter precipitate in this zone termed “marginal filter”. High thickness of Holocene sediments occurs between 72° N and 73°30’N where a distinct decrease in thickness is observed to the north. Two major acoustic units could be differentiated, separated by a prominent reflector interpreted as the base of the Holocene. High-resolution echosound data suggest a fluvial-dominated depositional environment for the early Holocene displaying lateral accretion as point bars and vertical accreted overbank deposits in a fluvial channel-levee-complex. During the early Holocene sea level rise the marginal filter migrated progressively southward (upstream) to its present position forming a typical high-stand system tract in acoustic images. Estuarine sedimentation in a sedimentary environment similar to today started at approximately 5 ka BP.

An estimated total of $14.3 \times 10^{10}$ t and $9.2 \times 10^{10}$ t of fine-grained brackish-marine sediments, in the Ob and Yenisei estuaries, respectively, were accumulated during Holocene times. This is only about 75% and about 50% of Ob and Yenisei estuarine sediment budgets, respectively, estimated by extrapolation of recent river run-off data over the last 7500 years. Filled paleoriver channels indicate active river incision in the southern part of the Kara Sea Shelf prior to the Holocene.

1 Introduction

Sedimentation in the Arctic Ocean is dominated by terrigeneous input which leads to high silicilastic contents in marine sediments (Stein and Korolev, 1994; Stein, 2000). Biological productivity plays a minor role in sediment supply because the availability of nutrients is limited (Grebmeier, 1995). There are two major sources of terrigeneous input: (i) riverine discharge which is prevailing in the western Arctic (MacDonald et al., 1998), and (ii) coastal erosion which is typical for the Laptev Sea (Rachold et al., 2000). At present, total annual run-off sums up to 3330 km$^3$ for the whole Arctic. Approximately 30% of the Arctic Ocean fresh water input is supplied by the rivers Ob ($429$ km$^3$y$^{-1}$) and Yenisei ($620$ km$^3$y$^{-1}$) through their estuaries into the Kara Sea (Fig. 1;
In the Ob and Yenisei estuaries there are generally two factors controlling sediment accumulation and thus the sediment budget. One is sediment input and the other is erosion and/or resuspension.

Sediment input into the system mainly depends on river run-off which varies strongly both seasonally and interannually (Pavlov and Pﬁrman, 1995). The annual maximum discharge rate is observed in June. During this peak, 45 to 65% of the annual freshwater run-off and 80% of the annual sediment masses are released (Gordeev et al., 1996; Shiklomanov and Skakalsky, 1994).

In Arctic coastal waters, resuspension and lateral transport of sediments can be enhanced by several factors and processes. These include brine-induced currents associated with sea-ice formation (Blanchet et al., 1995; Lisitzin, 1995), wave-induced currents during storm events (Løset et al., 1999; Are, 1996; Stein, 2000), sediment erosion by grounding sea-ice (Løset et al., 1999; Barnes and Lien, 1987)
and incorporation of particles into sea-ice during „suspension freezing“ (Reimnitz et al., 1993; Dethleff, 1994). These processes are important on the open shelf with large areas exposed to strong winds and significant sea-ice-formation. They are not expected to play a major role in sheltered estuaries, such as those of Ob and Yenisei.

Estimations of sediment fluxes on arctic continental margins are rare. Stein (2000) quantified the carbon fluxes into the Laptev Sea and Bauch et al. (2000) estimated sediment input in the eastern Laptev Sea shelf. There are some rough flux estimates including the Ob and Yenisei estuaries by Russian authors (e.g. Lisitzin, 1995). Very recently, a synthesis about sediment and organic carbon fluxes in the Arctic Ocean and surrounding marginal seas has been compiled by Stein and Macdonald (2003).

The major aim of this paper is the calculation of a Holocene sediment budget for the estuaries of Ob and Yenisei as a basis for assessment of material fluxes into the Kara Sea and further into the Arctic Ocean. These estuaries may function as major sediment sinks.

In this study, we interpret high-resolution acoustic reflection profiles (2-12 kHz) to map the thickness and lateral extent of fine-grained Holocene sediments between 70° N and 74°30’ N (Fig. 1). High-frequency single channel systems are particularly capable to penetrate these muddy sediments. This, in combination with AMS 14C dating and the physical properties of the retrieved gravity cores, enables us to evaluate the Holocene sediment budget of the area. We further discuss the evolution and depositional history of the Ob and Yenisei estuaries particularly with regard to the post-glacial sea level rise and migration of deposition centers. The data was collected during the expeditions of the RV “Akademik Boris Petrov” in 1999, 2000 and 2001 (Stein and Stepanets, 2000; 2001; 2002; Fig. 1).

In the area of interest extensive surveys have been carried out by Russian authorities and many shallow acoustic lines have been recorded. Unfortunately, this data is not accessible because the region is strategically relevant for military and commercial reasons.

2 Materials and Methods

During three expeditions (1999, 2000 and 2001) more than 5000 kilometers of high-resolution subbottom profiles in the estuaries of Ob and Yenisei were recorded (Fig. 1) using the hull-mounted ELAC system of RV “Akademik Boris Petrov” (ELAC echograph LAZ 72, Honeywell-Nautik, Kiel, Germany) which operates at 12 kHz. In 2001 we amplified and plotted the ELAC signal using a GeoAccoustics receiver and Ultra printer 120, respectively (Stein and Stephanets, 2002). In addition, for higher acoustic resolution and penetration a towed CHIRP system was used (Geochirp,
GeoAcoustics, Great Yarmouth, UK). The CHIRP sweep was set to 2-8 kHz (high-penetration mode). Digitising of acoustic data was achieved with a digital receiver (Octopus 360, Deddington, UK). Analogue data including GPS positions and time were stored on a four-channel DAT recorder. A detailed description of the whole system is given by Stein and Stepanets (2002). Time-to-depth conversion was done using a sound velocity of 1500 m s\(^{-1}\) for water (Hamilton, 1972). Digital processing was carried out using PC–based software REFLEX (HarbourDom GmbH, Köln, Germany).

Sediment cores of up to eight meters in length were retrieved with a gravity corer. Wholecore physical properties were measured in 1 cm depth intervals using a Geotek Multi Sensor Core Logger. The system provides data of density, P-wave velocity and magnetic susceptibility (e.g., Schultheiss et al., 1987; Schultheiss and McPhail, 1989; Weaver and Schultheiss, 1990). Changes in these parameters reflect changes in porosity and/or sediment composition (Thompson and Oldfield, 1986; Weber et al., 1997). The technical setup of the logging system, as used during “Akademik Boris Petrov” expedition in 2001, is described in Stein and Stephanets (2002). After splitting the cores were described (Matthiessen and Stepanets, 1999; Stein and Stepanets, 2000; 2001; 2002) and sampled for determination of sedimentological parameters.

For the determination of magnetic susceptibility of surface samples, freeze-dried samples were densely packed in 12.5 cm\(^3\) plastic vials and measured using a MS2B sensor, Bartington Ltd, UK (for further information see Niessen and Weiel, 1996). Mass-specific susceptibility was calculated by normalising volume-specific susceptibility to 10 cm\(^3\) and dividing by the sample weight (Dearing, 1994).

Linear sedimentation rates were calculated between AMS \(^{14}\)C-dated fix points. All AMS \(^{14}\)C ages presented in this paper are in Calendar kilo years B.P. (ka BP, see Stein et al., 2003b). The acoustic subunits were differentiated after the sequence stratigraphy concept of Vail et al. (1977). Mass accumulation rates (MAR) were calculated according to Stein (1991):

\[
\text{MAR (g cm}^2\text{ ky}^{-1}) = \text{LSR} \times (\text{WBD} - 1.026 \text{GRP}/100)
\]

Sediment porosity was determined by gamma-ray absorption and calculated using the following equation (Weber et al., 1997):

\[
\text{GRP} = \frac{\text{Dgp} - \text{GRD}}{\text{Dgp} - \text{Dw}}
\]

Where: GRP = Gamma-ray Porosity
GRD = Gamma-ray Density
Dgp = Grain Density: 2.65 g cm\(^{-3}\)
Dw = density of (salt) water: 1,024 g cm\(^{-3}\)
LSR = linear sedimentation rate  
WBD = wet bulk density (all values and abbreviations after Weber et al., 1997).

3. Results

3.1 Subbottom profiling and seismic stratigraphy

In general, two major acoustic units (Unit I and Unit II, top to bottom) (Figs 2 to 4) and three subunits (Ia to Ic; Figure 2) were distinguished in the echograph profiles. Major units are separated by a distinct reflector (Figs 2 to 4).

In the CHIRP profile (Fig. 2), located in the central area of the Yenisei estuary (Fig. 1), subparallel reflections of Unit I indicate well-stratified sediments. At its base Unit I drapes concordantly the variable topography of the underlying Unit II (Fig. 2). Subbottom topography is gradually more levelled towards the top of the sediment fill. In both the western and eastern end of the section, two nearly 10 m deep channels characterise the present sediment surface (Fig. 2).

The CHIRP profile (Fig. 2) and two enlarged profile sections (Figs 5 and 6) exhibit important detail information about the seismic stratigraphy of Unit I. This record suggests the definition of three subunits Ia to Ic based on their geometries. In general, all subunits of Unit I are resembling the same back scatter behaviour which is typical of mud (Damuth, 1975; 1978).

Subunit Ia exhibits almost constant total thickness along the whole profile and very little lateral fluctuations of sediment thickness between individual reflectors including the eastern channel. Only in the western channel some thinning of infill strata can be observed. The sediment surface exhibits some incision marks typical of ice gouging, especially in the central section of the profile where water depth is 20 m or less (Fig. 2A).

Subunit Ib has an asymmetrical geometry with higher thickness in the western end of the profile. Between the two channels, the top of this unit forms a nearly horizontal plane at 25 m below present sea level and contains sediment fills of about 5 to 10 m thickness in the eastern channel. In the western area of the CHIRP profile, the asymmetry of Subunit Ib is defined by reflector downlaps onto the surface of Unit II (Figs 2, 5 and 6). Subunit Ib sediments in the eastern channel are characterised by a complex reflector geometry related to channel migration associated with filling (Fig. 6). Subunit Ic shows an asymmetrical, prismatic geometry decreasing in thickness to the west with increasing distance from the eastern channel (Fig. 2C). Subunit Ic sediments are missing in both channel locations and towards the outer edges of the profile so that Subunit Ic is only present in the central area of the cross section (Fig. 2C). Near the eastern channel, Subunit I sediments exhibit inclined reflector downlaps.
Figure 2. High-resolution profiles from the central Yenisei Estuary (see Figure 1 for location). CHIRP (A) and ELAC (B) profiles were recorded simultaneously along the same line. C: acoustic Units and sediment core locations.
onto the upper flank of the channel (Fig. 6). In general, Unit I exhibits the character of a sequence representing a depositional cycle from more asymmetric, channel-levee type of deposits near the base (Ic) to more symmetric type of reflectors (Ia) indicative for sediments draping the cross section at the top. The boundaries between subunits Ia/Ib and Ib/Ic appear to be conformable with the exception of some erosional truncation that is visible near the flank of the eastern channel (Figs 2 and 6). In places, Unit I is affected by diffractions (Fig. 5) typical of „gas blanking“ (Fader, 1997) which mask the
underlying sediments. This may indicate some content of sediment gas possibly related to decomposition of organic matter.

The top of Unit II forms a prominent reflector which is present in the whole working area both in CHIRP and ELAC profiles (Figs 2 to 4). The acoustic impedance contrast between Unit I and Unit II appears to be large because there is almost no penetration
below. The surface of Unit II shows an irregular, rough morphology indicative of an erosional surface. Paleo-channels incised into Unit II including the eastern channel of the CHIRP profile (Figs 2A and 6) and several channels recorded in ELAC profiles (Figs 3C and 4C) which are partly filled with Unit I sediments. In other profiles (Figs 3A and 4A), channels are completely filled and are not, or hardly visible in the present surface morphology. In high exaggerated echosounding profiles, channels appear as typical v-shaped incisions of 5 to 10 m depth. They occur more frequently on the shelf north of 73° N in front of the recent river estuaries (Figs 3A and 4A). Some incisions
appear to be considerably wider than distinct channels and are better described as paleovalleys (Figs 3A, 3B and 4C).

Across the Yenisei estuary (Figs 2A and 2B), the ELAC and CHIRP systems were in operation at the same time which provides a direct comparison between data of both echosounders. The CHIRP system (2-8 kHz) enables higher resolution suitable for subdivision of Unit I although penetration of both systems is about the same (Fig. 2). Both have in common that they can pick up prominent reflectors below the soft sediment such as the top of Unit II (Figs 3 and 4). Therefore, the ELAC system is also capable to identify Unit I including its thickness (Fig. 2).
3.2 Distribution and thickness of Unit I sediments

Acoustic travel times, as recorded in echosounding profiles during “Akademik Boris Petrov” expeditions, are used to determine and map the thickness of Unit I in the area under investigation (Fig. 7). Between profile lines, unit thicknesses were interpolated (Fig. 8). The highest sediment thicknesses occur in the estuaries between 72° and 73°30’ N (Fig. 8). Further north, Unit I is virtually absent on the shallow shelf and deposition is concentrated in morphological depressions and sediment sinks such as channels and paleovalleys (Figs 3A and 4A).

In the narrower Yenisei Estuary, the sediment cover appears more canalised and extends further to the north than in the Ob Estuary (Fig. 8). Maximum sediment thickness of 20 m is found at 73° N in the Yenisei Estuary (Figs 2A and 8). From there northward, an approximately 15 m thick sediment blanket extends to 73°30’ N (Fig. 8). North of this position, sediment thickness decreases distinctly, although in two elongated areas, thicknesses of up to 10 m continue further north up to 74° N (Fig. 8).

In the Ob Estuary, areas with higher sediment thicknesses are wider but do not extend as far north as in the Yenisei Estuary (Fig. 8). Here, the maximum sediment thickness is reached at 72°20’ N. North of 72°50’ N increased sediment thickness occurs along a south to north directing axis following 73° E. There is a second area with lower sediment thickness of up to 10 m extending in a SW-NE direction near 74° E.

3.3 Sediment-core lithology, physical properties and chronology

According to the lithological descriptions of the cores from the area under investigation (Stein and Stepanets, 2000; Stein et al., 2003b), major changes in lithology correspond with both boundaries between acoustic units (Fig. 9) and distinct changes in physical property records (Figs 10 to 12).

Lithological descriptions of sediment units

The predominant lithology of Unit I is a clayey silt to silty clay (Stein et al., 2003b) with quartz and clay minerals as major components (e.g. Stein and Stepanets, 2000). Bivalve shells associated with bioturbation are very common which can grade into horizontal, non-bioturbated bedding with fewer shells in the lower sections of Unit I (e.g. Stein and Stepanets, 2000). After calculating porosity from density data of all cores (Figs 10 to 12), an average porosity of Unit I sediments of 66% was quantified for all stations (Fig. 1). This porosity is used to determine the bulk net sediment mass from the sediment volume of Unit I (Figs 7 and 8; Table 1) determined by sediment
Figure 7. Thickness of Unit I in meters determined by sediment echosounding. Hydrographical conditions after Pavlov and Pfirmann (1995). The trapeze indicates position of Figure 8.

from the sediment volume of Unit I (Figs 7 and 8; Table 1) determined by sediment echosounding (see 3.4). Magnetic Susceptibility:

In general, the upper part of Unit I displays the lowest susceptibilities gradually increasing down-core (Figs 10 to 12). Similar to the density records, the transition from Unit I into Unit II is often characterised by a steep gradient with peak susceptibility at the boundary between Unit I and Unit II (Figs 10 to 11) cores BP99-04, BP99-09, BP00-23 from the Yenisei and BP00-38, BP01-83 from the Ob estuaries (Fig. 12). In some cores higher susceptibility values are associated with increased density as,
for example, in cores BP99-32, BP01-03 (Fig. 11), BP01-83 and BP00-38 (Fig. 12). These values can probably be accounted to decreased sediment porosity and thus more susceptibility bearing material in the core rather than a change in dry sediment composition. Other cores show susceptibility fluctuations independent of the density record as, for example, in cores BP99-39, BP99-04, BP00-23, BP00-14 (Fig. 10) and BP00-22 (Fig. 11) indicative for variations in content and/or composition of magnetic grains.

In the Yenisei Estuary, susceptibility is $100 \times 10^{-5}$ (SI) on average with peak values of $250 \times 10^{-5}$ (SI) and $550 \times 10^{-5}$ (SI) in core BP00-22 and BP00-23, respectively (Figs 10 and 11). Both cores penetrate Unit II and in both cores peak susceptibility values are in Unit II (Figs 9 to 11). In sediment cores from the Ob Estuary, values are significantly lower, averaging about $30 \times 10^{-5}$ (SI). Here, the highest values reach $70 \times 10^{-5}$ (SI)
(Core BP00-38; Figure 12). The mass-specific magnetic susceptibility of recent surface sediments (Fig. 13) shows two major trends. The first is, as observed in sediment cores, that values for the Yenisei river are significantly higher than in the Ob area with averages at 3,000 to 4,000 * 10^{-9} m^3 kg^{-1} and 1,000 to 1,500 * 10^{-9} m^3 kg^{-1}, respectively. The other important observation is a distinct decrease in values north of 74° N which is most pronounced in the Yenisei area. Despite the distinct decrease off the Yenisei Estuary some samples north of 74° N show slightly increased susceptibility compared to other areas of the southern Kara Sea. Core correlation: Core-to-core correlation using physical properties is not possible for the entire area under investigation. Only in case where core locations are in relatively close distances to each other, lateral correlation by density and susceptibility is obvious such as for cores BP99-04 to BP01-25, BP99-39 to BP99-38 (Fig. 10), BP99-32 to BP99-08 to BP01-03 (Fig. 11) and BP00-38 to BP01-83 (Fig. 12).

Chronology

All datings of sediment cores penetrating into Unit I reveal Holocene ages (Fig. 9, Stein et al. 2003). The oldest Unit I age of 9 ka BP is determined at the base of Core BP99-04 located in the norther Yenisei Estuary. The distinct down-core increase of both density and susceptibility at the bottom of core BP99-04 (Fig. 10) suggests that the transition into the underlying Unit II is exposed and dated and the gravity corer got stuck at or shortly above the UnitI/II boundary.

The only dateable core penetrating into Unit II is BP99-05. An age of 15.5 ka BP was determined at a core depth of 3.15 m (Fig. 9). The dated material is a relatively fresh shrub wood suggesting a nearly contemporaneous sedimentation. Though it is driftwood, the good preservation can be used as first evidence for a pre-Holocene age of Unit II.

3.4 Correlation of acoustic units with sediment cores

The coring locations BP00-14, BP99-32, BP99-08 and BP01-03 are on the track of the CHIRP profile (Fig. 2C). BP00-14 cored both Unit I and II. Lithology and physical properties demonstrate that the increase in sand content correlates with higher density. The density gradient defines the major acoustic impedance contrast causing the strong seismic reflector at the boundary between Units I and II. The same is true for a correlation of Unit I/II boundaries cored by BP00-23 and BP99-05 (Fig. 9) with the ELAC profile from the northern Yenisei Estuary (Fig. 4B). This suggests that the transition from Unit I to II observed in cores and acoustic profiles are identical in the entire area.
Figure 9. Stratigraphic correlation of selected cores based upon core description, datings and echo-sounding. Cores are arranged according to their water depth (right hand scale) see Figure 1 for location.
Figure 10. Density and magnetic susceptibility records of Unit I and II sediments from the northern Yenisei Estuary.
Figure 11. Density and magnetic susceptibility records of Unit I and II sediments from the inner Yenisei Estuary.
Figure 12. Density and magnetic susceptibility records of Unit I and II sediments from the Ob Estuary.
The only core, which clearly penetrated the acoustic boundary between subunits Ia and Ib of the CHIRP profile, is BP01-03 whereas BP99-32 and BP99-08 probably only comprise Subunit Ia sediments (Fig. 2C). The boundary between Subunit Ia and Ib is insignificant in physical property records of both density and susceptibility (Figs 10 to 12) and not visible in the lithology (Fig. 9).

Along the CHIRP profile, Subunit Ic was not cored because it is below the maximum gravity-corer sediment-penetration limit of about eight metres (Fig. 2C). Along ELAC profiles, where cores recovered both Unit I and Unit II sediments, Subunit Ic could not be identified due to limited echosounding resolution or may not be present.

### 3.5 Sediment mass

The total sediment volume was calculated from the areal extent of Unit I sediments, compiled by echosounding, multiplied by its thickness (Figs 7 and 8; Table 1). The amount of dry sediment was quantified by subtracting the average volume of pore water (66% porosity) from the total wet sediment volume. The net sediment mass was
then quantified by multiplying the porosity corrected value of dry sediment volume with the average grain density of 2.65 g cm\(^{-3}\) (e.g. Weber et al., 1997). The total estimated sediment yields of the estuaries of Ob and Yenisei differ significantly but are in the same order of magnitude (Table 1). Acoustic Unit I yields 14.3 \(10^{10}\) t and 9.2 \(10^{10}\) t of dry sediment (Table 1) in the Ob and Yenisei estuaries, respectively.

Table 1. Quantified sediment area and volume based on isopach map (Fig.8). The net sediment mass was determined by the sediment volume and the average porosity of sediment cores and the average dry grain density.

<table>
<thead>
<tr>
<th>Sediment Thickness (m)</th>
<th>Area (km(^2))</th>
<th>Volume (km(^3))</th>
<th>Netto Sediment Mass (*10(^{10})t)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Yenisei</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>&gt;15</td>
<td>1815.1</td>
<td>31.8</td>
<td></td>
</tr>
<tr>
<td>&lt;15</td>
<td>2022.1</td>
<td>25.3</td>
<td></td>
</tr>
<tr>
<td>&lt;10</td>
<td>3789.4</td>
<td>28.4</td>
<td></td>
</tr>
<tr>
<td>&lt;5</td>
<td>6512.1</td>
<td>19.5</td>
<td></td>
</tr>
<tr>
<td>total</td>
<td>105.0</td>
<td>9.2</td>
<td></td>
</tr>
<tr>
<td>Ob</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>&gt;15</td>
<td>2452.0</td>
<td>42.9</td>
<td></td>
</tr>
<tr>
<td>&lt;15</td>
<td>3805.4</td>
<td>47.6</td>
<td></td>
</tr>
<tr>
<td>&lt;10</td>
<td>5938.9</td>
<td>44.5</td>
<td></td>
</tr>
<tr>
<td>&lt;5</td>
<td>8709.4</td>
<td>26.1</td>
<td></td>
</tr>
<tr>
<td>total</td>
<td>161.1</td>
<td>14.3</td>
<td></td>
</tr>
</tbody>
</table>

4 Discussion

4.1 Late Quaternary sedimentary environments

Age model

The radiocarbon datings of Unit I and II suggest that the boundary may represent the pre-Holocene transition or unconformity. The susceptibility spike observed near the top of Unit II in cores BP99-05, BP99-09, BP99-38, BP00-23 and BP00-22 (Figs 9 and 10) is similar to a comparatively high spike of susceptibility described in cores from the eastern Kara Sea and Laptev Sea by Kleiber and Niessen (2000).

The susceptibility peak revealed a pre-Holocene age of 10 to 13 ka BP and is interpreted as a result of enhanced input of magnetic minerals during Termination I which have their source in the basaltic Putoran Plateau (Kleiber and Niessen, 2000). The spike is thought to originate from a melt-water run-off peak related to deglaciation of the Putoran Plateau which results in increased delivery of high susceptibility bearing material. If this interpretation is correct, the susceptibility peaks of Unit II observed
in our cores from the Kara Sea are probably related to the same origin (Stein et al., 2003) because the Yenisei is draining part of the Putoran Plateau. The interpretation of magnetic susceptibility peaks near the top of Unit II, together with the radiocarbon age of 15.5 ka BP of Core BP99-05, suggest that Unit II is of late Pleistocene age including Termination I.

Because radiocarbon dating of Unit I confirmed Holocene ages (Fig. 9; Stein et al., 2003), it is likely that the distinct reflector separating Units I and II (Figs 2 to 4) is the base of Holocene deposition. This reflector appears as pre-transgressive, erosional boundary in most locations. Thus, Unit I thicknesses (Figs 7 and 8) are interpreted to represent total Holocene sediment accumulation.

Sedimentary history

Unit II was probably subaerially exposed and partly eroded during sea level low stand. For example, a relatively deep river channel (at least 55 m below present sea level) was incised into the strata of Unit II (Figs 2 and 15: Phase I) during the last sea level lowstand at the location of the present Yenisei Estuary. According to the modern shelf morphology, the paleoshoreline and related estuaries were situated further to the north along the present day outer shelf. Thus, the pre-Holocene paleoenvironment between 72° and 74° N was fluvial or subaerially with erosion as predominant process.

The geometry of Subunit Ic is dominated by typical fluvial architecture elements such as point bars (lateral accretion) and overbank deposits described by e.g. Allen (1965) and Miall (1996 and references therein) in modern fluvial sedimentary environments. In Subunit Ic downlaps onto the upper channel flank are interpreted as accretion of inclined point bars (Fig. 6). This suggests channel migration towards the east (Fig. 6). There are no Subunit Ic sediments preserved inside the channel which indicates high current velocities near the channel axis. Outside the channel, strata grade laterally into sediments with parallel reflectors. This indicates decreased influence of currents resembling typical overbank sediments (Allen, 1965), accumulated in a wedge-shaped structure on a floodplain (Fig. 15: Phase II), typical of a natural levee (Brierly et al., 1997; Bown and Kraus, 1987). According to Galloway and Hobday (1996) horizontal aggradation on a floodplain occurs when sediment-laden water overflows and releases its suspension load due to flow stripping.

In the CHIRP profile, the bottom of Subunit Ic (Fig. 2) is located approximately 40 m below present sea level which is the same depth as the base of core BP99-04 (Figs 2 and 4B) situated some 80 kilometers further downstream of the profile (Fig. 1). AMS 14C dating of the base of the core reveals an age of 9 ka BP (Fig. 9). Sedimentology (Stein et al. 2003) and palynological investigations (Kraus et al. 2003) indicate fluvial influences for the lowermost part of core BP99-04 and changes to a marine environment
Bauch et al. (2001) used age and eustatic sea level data by Fairbanks (1989) for the reconstruction of the Holocene transgression in the Laptev Sea. The sea level curve serves as a rough indication for paleosea levels at a given age, if the area is not affected by isostatic rebound after deglaciation. According to Fairbanks (1989) the eustatic sea level trend was about 30 m below present at 9 ka BP (Fig. 14). Assuming no (or negligible) isostatic rebound of the Yenisei Estuary area during the Holocene, the onset of sedimentation in core BP99-04 took place near the sea level in very shallow water (Fig. 14). If true both locations BP99-04 and the CHIRP profile (Fig. 1) must have been strongly influenced by the paleoriver Yenisei at 9 ka BP. This is consistent with the fluvial character of the lowermost sediments of Core BP99-04 and the interpretation of the seismic pattern above. This suggests a terrestrial, fluvial-dominated depositional environment for Subunit Ic in the CHIRP profile (Fig. 2). The paleoenvironment of Subunit Ic is best explained by a fine-grained meandering river system as proposed e.g. by Miall (1996). Probably, pronounced seasonal flood events similar to those observed in Siberian rivers today during late spring/early summer (Gordeev et al., 1996; Shiklomanov and Skakalsky, 1994) caused the accumulation of overbank deposits on the floodplain.

The age of the boundary between subunits Ib and Ic is difficult to assess because the core, which penetrated this boundary, is not dated yet (Fig. 9). However, the truncations at the top of Subunit Ic in Figure 5 could originate from an erosional transgressional ravinement surface (Thorne and Swift, 1991). This surface possibly formed in response to the landward migration of the shoreline associated with erosion during transgression, for example, as a result of wave action. If the entire Subunit Ic is a relict fluvial floodplain deposit, the marine transgression reached the top of this unit (30 m below present sea level, Figure 2) at about 8.5 to 8 ka BP (Fig. 14) and formed the onset of marine/estuarine Subunit Ib.

Subunit Ib is characterised by sediment accumulation in the entire channel with highest bed thickness at the channel axis that thins towards the flanks resulting in a divergent fill geometry (Fig. 6; Mitchum et al., 1977). This suggests a gradual decrease in current velocity possibly related to increasing water depth during an early Holocene transgression associated with higher settling rates of particles in a shallow and widening estuarine environment.

As a result of rapid sedimentation in the eastern channel, the upper portion of Subunit Ib reached a higher bathymetric level than the channel base at the western end of the profile where no Subunit Ic and Ib sediments accumulated (Fig. 2). Subsequently, near the western deepest point of the profile, a new channel system was established.
where point bar deposits migrated westward forming Subunit Ib downlaps onto Unit II (Fig. 5). Formation of the western channel (Fig. 5) started in the same range of water depth (downlaps on Unit II at approx. 35 to 40 m below sea level; Figure 2) as the termination of current controlled sedimentation in the eastern channel (Fig. 6). The western channel displays sediment layers that are thinner at the channel base than at the adjacent flanks (Fig. 5) which is typical of an aggradating channel-levee-complex (Allen, 1965). The western channel probably represents the recent thalweg because, at present, its channel bottom has the maximum water depth of the entire cross section (Fig. 2C). In addition, the slightly thinner channel-fill suggests that recent sedimentation is affected by currents.

Figure 14. Eustatic sea level trend for the last 15000 years (after Barbados sea level curve by Fairbanks 1989) and reconstructed sediment surface in the Yenisei Estuary after chronological data from cores P99-04 and BP00-14 (Fig.9). Note that channel erosion was before and onset of sedimentation after 10 ka BP. During the Holocene transgression sea level rise exceeded sediment accumulation in the area of the cores.
The entire Subunit Ib is typical for a shallow marine/estuarine channel-levee-complex characterised by high sedimentation rates in both channel and levee areas. The change in the depositional environment compared to Subunit Ic is interpreted as the result of sea level rise, decrease in current velocity and southward migration of marginal filter-type deposition centers. It is interesting to note that the well defined chronology of Core BP99-04, located 80 km to the north of the CHIRP profile also revealed high sedimentation rates for the early Holocene until about 7.5 ka BP which gradually decrease to lower rates until about 5 ka BP (Fig. 9). Consistently high sediment accumulation rates of 650, 250, and 240 (g cm$^{-2}$ ky$^{-1}$) are quantified between 9.5 to 7.5 ka BP in cores BP00-36, BP00-26, and BP00-07 located 390 km, 250 km, and 130 km north of the Yenisei Estuary, respectively (Stein et al., 2003). It is suggested that these early Holocene deposition centers north of the present estuaries are lateral counterparts of the rapidly filling shallow-marine channel-levee-systems of Subunit Ib indicative of a southward migration of the marginal filter during the transgression. Draping subbottom topography and nearly constant thickness of Subunit Ia implies precipitation-type of particle settling out of the water column across the entire estuary, possibly overprinted by more active bottom currents in the western channel. This pattern is typical for the marginal filter setting of today with a relatively stable high sea level where salinity induced processes are important such as particle coagulation followed by rapid settling. Retrieved near the western end of the CHIRP profile from 19.2 m water depth, core BP00-14 reveals a minimum age of 4.8 ka BP for the base of Subunit Ia (Fig. 9). This suggests a major hiatus at the core location with Subunit Ia unconformably overlying Unit II. This is consistent with the seismic stratigraphy (Fig. 2) indicating that subunits Ib and Ic are missing at the coring location. The hiatus was possibly caused by subaerial exposure during the early Holocene until the transgression reached the location.

It seems to be reasonable that the onset of the drape geometry of Subunit Ia is related to the end of the transgression which reached the present sea level by about 5 ka BP. Therefore, we assume that the marginal filter reached its recent position between 7.5 and 5 ka BP.

The distribution pattern of magnetic susceptibility of recent surface sediments (Fig. 13) documents the recent spatial extent of the deposition center in the Yenisei Estuary because magnetic grains derived from the Putoran Plateau trace terrestrial input into the Kara Sea. The susceptibility pattern is consistent with the distribution of Holocene sediment thickness (Fig. 8) demonstrating that a large amount of riverine material is trapped in the estuaries indicative for the marginal filter effect (Lisitzin, 1995). Because there is no similar source of magnetic grains in the Ob catchment, magnetic susceptibility in surface sediments of the Ob Estuary remains low (Fig. 13) despite
increased sediment accumulation (Fig. 8). Therefore susceptibility cannot be used to trace the marginal filter effect in the Ob Estuary.

The proportions of sediment thicknesses of subunits Ia to Ic a were estimated from the transect section in Figure 2 to assess the time-dependent sediment accumulation in the present Yenisei Estuary. Each Subunit represents approximately 1/3 of the Holocene sediments in this particulate section. Subunit Ic covers a time interval ranging from 11.5 to about 8 ka BP. By comparison with the global sea level curve (Fig. 14), it seems likely that significant deposition in Subunit Ic started after 10 ka BP with high sedimentation rates of about 0.33 cm y\(^{-1}\). Subunit Ib comprises some 3000 years from about 8 to 5 kys. BP and Subunit Ia the last 5000 years, resulting in sedimentation rates of 0.2 cm y\(^{-1}\) and 0.14 cm y\(^{-1}\), respectively.

This assumption clearly demonstrates that those units comprising acoustic evidence for enhanced current activities (active channel-levee-complexes as subunits Ia and Ib), and thus higher rates of sediment by-pass, are also characterised by higher sedimentation rates than the marginal filter of the younger Holocene (Subunit Ia). We suggest that the location of channel-levee-complexes are strongly controlled by the location of the paleoshoreline and sea level, therefore of small spatial extent and steadily migrating southward during the early Holocene transgression. In contrast, the marginal filter setting after 5 ka BP remains in relatively stable steady state. Sediment accumulation took place in deeper water and thus over a relatively large area (Figs 2 and 8) resulting in overall lower sedimentation rates.

### 4.2 Sediment budget

The Holocene sediment budget from seismic data and sediment cores as presented in this paper is difficult to test because a number of uncertainties remain which affect the sediment accumulation. This includes additional input from coastal erosion long-term variations in the riverine discharge, and transfer rates from the estuaries to the shelf. The effect of these processes are hard to quantify with an appropriate spatial and temporal resolution.

For example, Rachold et al. (2003) conclude for the Laptev Sea that the amount of coastal erosion (44 \(*\ 10^6\ \text{t}\ y^{-1}) is almost twice as high as the riverine input (25 \(*\ 10^6\ \text{t}\ y^{-1}). However, this estimate was made for the open shelf. In the sheltered estuaries of Ob and Yenisei, we expect lower rates of coastal erosion because erosional energy by waves is minor and thus the coastline is more stable. Furthermore, there is a different type of permafrost along the Kara Sea coast (International Permafrost Association, Data and Information Working Group, comp. 1998) because no ice complexes are found as described from the Laptev Sea (Schirrmeister et al., 2002). Such ice
complexes commonly contain up to 80% of water which melts during the summer thereby forming unstable sediment cliffs susceptible to coastal erosion. Missing ice complexes in association with a low relief (Gordeev et al., 1996; Milliman and Syvitzky, 1992) suggest that significant coastal erosion is unlikely in the sheltered Ob and Yenisei estuaries.

Figure 15. Late Quaternary evolutionary model for the high-resolution CHIRP profile (Fig. 2).
If coastal erosion is not important, one simple way of testing our budget is to compare the total net sediment masses estimated from this study with the modern average discharge of suspended and dissolved matter (Gordeev et al., 1996; Telang et al., 1991) extrapolated back over the last 7500 years (Table 2), the time period during which we believe the establishment of the modern marginal filter system has occurred.

In order to be closer to Holocene conditions not affected by anthropogenic change, we include data from before A.D. 1950 (Table 2; Telang et al., 1991; Holmes et al., in press) for the Yenisei River. This takes into account dam building (e.g., the Krasnoyarsk Dam) during the last 50 years in the catchment area of the Yenisei Estuary.

Table 2. Recent sediment discharge for the Ob and Yenisei rivers. II: Discharge amount of suspended and dissolved matter during the last 7500 years quantified by linear extrapolation of I. III: Total amount of material expected as accumulated sediments present in the marginal filter according to Lisitzin (1995);

<table>
<thead>
<tr>
<th></th>
<th>I. suspended matter (10^6 t a^-1)</th>
<th>dissolved matter (10^6 t a^-1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ob</td>
<td>16.5(2)</td>
<td>34(1)</td>
</tr>
<tr>
<td></td>
<td>13.4(1)</td>
<td>15.5</td>
</tr>
<tr>
<td></td>
<td>5.9(2)</td>
<td>43.2(1)</td>
</tr>
<tr>
<td>Yenisei</td>
<td>14.4(1)</td>
<td>14.4</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>II. total discharge in 7500 years (10^10 t 7500 a^-1)</th>
<th>total discharge in 8000 years (10^10 t 7500 a^-1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ob</td>
<td>11.6</td>
<td>25.5</td>
</tr>
<tr>
<td>Yenisei</td>
<td>10.8</td>
<td>32.4</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>III. 90 % of II (10^10 t 7500 a^-1)</th>
<th>30 % of II (10^10 t 7500 a^-1)</th>
<th>Total (10^10 t 7500 a^-1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ob</td>
<td>11.0</td>
<td>7.7</td>
<td>18.7</td>
</tr>
<tr>
<td>Yenisei</td>
<td>10.3</td>
<td>9.7</td>
<td>20.0</td>
</tr>
</tbody>
</table>
The extrapolation suggests that the Ob River released $11.6 \times 10^{10}$ t of suspended matter and $25.5 \times 10^{10}$ t of dissolved matter during the last 7500 years (Table 2). The Yenisei River discharged $10.8 \times 10^{10}$ t of suspended matter and $32.4 \times 10^{10}$ t of dissolved matter (Table 2). According to Lisitzin (1995), 90% of the total suspended matter and 30% of the dissolved matter precipitate and accumulate in the marginal filter at the river-sea interface. If true, this would result in a total sediment accumulation of $18.7 \times 10^{10}$ t for the Ob River and $20 \times 10^{10}$ t for the Yenisei River (Table 2). These numbers are significantly higher than the Holocene sediment masses estimated from mapping of Unit I sediment thickness in the Ob and Yenisei estuaries which results in $14.3 \times 10^{10}$ t (about 75% of the modern-based calculated value), and $9.2 \times 10^{10}$ t (about 50% of the modern-based calculated value), respectively (Table 1). It may suggest that the study of Lisitzin (1995) describing the marginal filter as efficient sediment trap for fluvially-derived material underestimates sediment by-pass and/or export into the Kara Sea. The discrepancy between our and Lisitzin's approximation is highly relevant for both assessment of material exported from the Ob and Yenisei catchments into the Kara Sea and the sediment budget of the entire Arctic Ocean. Thus, the possible reasons, which can account for the discrepancy, need to be discussed in further detail.

One reason may be that the ELAC system used for our quantification does not penetrate Unit I down to the top of Unit II in all locations thereby underestimating Unit I thickness and mass. This is not very likely because the strong reflector marking the top of Unit II is visible in most profiles (Figs 4 and 5). Moreover, comparison along lines, where both CHIRP and ELAC data were recorded during the expedition BP01, demonstrate that both systems revealed the same thickness for Unit I (Stein and Stepanets, 2001; Figure 2).

Another reason may be that the extrapolation of recent run-off data as summarised in Table 2 may not be correct and/or not valid for the last 7500 years because climate change could have caused variation in run-off during the Holocene (Stein et al., 2003b). Here it is interesting to note that our assessment of sediment accumulation for Subunit Ia for the last 5000 years matches almost exactly recent sedimentation rates in the marginal filter using Cs 137 records (Stepanets et al., 1999). Combining the average porosity (66%) with the mean Subunit Ia sedimentation rate (0.2 cm y $^{-1}$) and the average sediment surface porosity (74%) with near-surface sedimentation rates of 0.3 cm y $^{-1}$ (Stepanets et al., 1999), respectively, results very similar in mass accumulation rates (MAR) of 0.21 g cm$^{-2}$ ka$^{-1}$ in surface sediments and 0.2 g cm$^{-2}$ ka$^{-1}$ in the entire Subunit Ia.

Thus, the most reasonable explanation for this discrepancy is that our study does not quantify early Holocene accumulation correctly until ca. 5 to 6 ka BP, the time
when sea level reached the present level. The southward migrating position of the marginal filter during the transgression is difficult to compile from the data available. Possibly, there are additional depositional centers further north, which are outside the area of data acquisition and thus not yet mapped (Stein et al., 2003b). The character of a deepening upward sedimentary sequence in the estuaries in association with acoustic pattern indicative for currents in the early Holocene (Figs 2, 5 and 6) suggest that current-induced by-passing of sediments onto the open Kara Shelf has decreased during the Holocene. In turn, this implies higher sediment export out of the area mapped in this study before 5 ka BP. This export must have been significant because Subunit Ic sediments, interpreted as fluvial deposits possibly older than 8 ka BP, are included in the budget based on mapping. Therefore, our approach using the total thickness of Unit I sediments should overestimate the actual budget for the last 7500 years thereby increasing the discrepancy between the different ways of quantification.

Although early Holocene by-pass probably accounts for most of the budget discrepancy discussed above, sediment export out of the estuaries may also play some role today and during the last 5000 years. Is there evidence for sediment erosion and lateral transport?

Ice gouges visible in the acoustic images are evidence for grounded ice floes (Fig. 2) which can disperse sediment at the bottom (Barnes and Lien, 1987; Reimnitz et al., 1972; 1978) when drifting sea-ice ridges plough through the sediment. Ice thickness can reach up to 20 m in pressure ridges (Løset et al., 1999; Barnes and Lien, 1987). However, at least for the area of Ob and Yenisei estuaries sediments appear mostly undisturbed in acoustic images and ice gouges only occur locally (Fig. 2). Therefore, ice gouging is expected to play only a minor role in the sediment budget of the Ob and Yenisei marginal filter systems.

Sediment by-pass may be enhanced by bottom currents which can be provoked by tidal activity, changes in the river discharge (especially at the spring discharge peak), and brine formation by freezing of seawater (Reimnitz et al., 1993; Dethleff, 1994; Løset et al., 1999; Barnes and Lien, 1987; Lisitzin, 1995; Pfirman et al., 1995; Are, 1996). Evidence from this and other studies suggest that, at least for the Yenisei Estuary, considerable by-pass has to be considered for the present situation of the marginal filter. Material export along certain pathways out of the Yenisei Estuary is documented by relatively high values of magnetic susceptibility in surface sediments on the open Kara shelf, mainly concentrated in morphological depressions and/or channels (Fig. 13; Stein et al., 2003b). This is consistent with the fact that in cores from the area north of the Yenisei marginal filter (BP0036, BP00-26, and BP00-07) there is still accumulation today, but with smaller rates of 7, 17 and 40 (g cm$^2$ ka$^{-1}$),
respectively, compared to the early Holocene. Moreover, according to our estimates, there is a smaller Holocene sediment mass in the Yenisei Estuary (Table 1) than in the Ob Estuary although they receive nearly the same influx (Table 2). Thus, accumulation rates, magnetic susceptibility in surface sediments and the sediment budget indicates some unquantified sediment export from the Yenisei Estuary to the shelf probably active during the entire last 5000 years.

Higher export rates of particulate matter out of the Yenisei Estuary could be related to different estuarine morphologies and hydrological regimes. It is interesting to note that inflow of sea water into the Yenisei Estuary is enhanced by its deeper bathymetry compared to that of the Ob. Harms et al. (2003) describe a strong sea water underflow in the Yenisei as a result of channelled river-water inflow into the Kara Sea which is of the same direction as the main summer winds thereby enforcing Ekman transport. Thus, through the Yenisei Estuary surface waters are transported relatively far to the north. In contrast, the mouth of the Ob Estuary is characterised by a tide-wave interference zone with relatively high kinetic energies which may influence sediment transport, mixing, and deposition in relatively unstratified waters (Harms et al., 2003). These different characteristics may result in a larger sediment export to the open shelf out of the Yenisei Estuary as compared to the Ob.

5 Conclusions

The sediments in the Ob and Yenisei estuaries can be divided into two major acoustic and lithological Units (I and II). Unit II forms the pre-Holocene basement with little acoustic penetration. During the last sea level lowstand, the erosional surface was formed on top of this unit including several incised river channels. The younger Unit I can be divided into three subunits, Ia to Ic. Their formation is related to post-glacial sea level rise after about 12 ka BP and the southward migration of the river-sea marginal filters to their recent position in the estuaries between 72° to 73° N. The lowermost subunits Ic and Ib show fluvial to shallow-marine seismic features comparable to typical channel-levee-complexes representing a fine-grained meandering river depositional environment overlain by a transgressive system tract and finally high stand system tract after about 5 ka BP.

The entire sediment fill of the present estuaries reflects a deepening-upward sediment growth structure suggesting that sea level rise exceeded sediment accumulation although migration of the marginal filter during the Holocene shifted depositional centers south.

The present estuaries of the Ob and Yenisei rivers are the major Holocene sediment sinks of terrigeneous material supplied by rivers. An estimated total of $14.3 \times 10^{10}$ t and
9.2 \times 10^{10} \text{ t of dry sediment have accumulated during the Holocene, which is about 75\% and 50\% of the material expected from extrapolation of modern run-off data, respectively.}

Sediment deposition in early Holocene times outside the recent position of the marginal filter is proposed to be the reason for the differences in the sediment budgets. A progressive southward migration of the position of the marginal filter in response to the elevating sea level is favoured to account for the “missing” material. This would mean that a significant amount of material was deposited on the present shelf and not in the position of the recent marginal filter. Extensive sediment reworking is not observed but significant sediment by-pass through the marginal filter system cannot be excluded. Further studies are necessary to improve the temporal and spatial resolution of Holocene sediments for a better assessment of the early Holocene budget particularly in the central and northern areas of the Kara Sea.

The existence of filled paleoriver channels suggests a pre-Holocene history of the southern Kara Sea Shelf similar to that in the Laptev Sea (Kleiber and Niessen 1999). These channels must have been eroded during the sea level lowstand and were filled with sediments during the Holocene transgression. River erosion of the exposed shelf of the southern Kara Sea during implies continuous river transport across the shelf and absence of an ice sheet in the area of channel erosion.

**Acknowledgements**

Many thanks to J. Matthiessen and V. Rachold for their helpful comments. A special thanks to Frank Schoster and Mathias Kraus for many fruitful discussions. This study has been performed within the German-Russian research project “Siberian River Run-Off (SIRRO)”, Financial support by the German Ministry of Education, science, Research and Technology (BMBF) (grant no. 03G0547A) and the Russian Foundation of Basic research is gratefully acknowledged. We are also grateful to the crew of RV “Akademik Boris Petrov” for their support. Review comments of L. Polyak and R. Endler were highly appreciated and improved an earlier version of the paper considerably.

Data are available at www.pangaea.de.
3.4 Acoustic facies on the inner Kara Sea Shelf: implications for late Weichselian to Holocene sediment dynamics
3.4 Acoustic facies on the inner Kara Sea Shelf: implications for late Weichselian to Holocene sediment dynamics

Dittmers, K.*, Lambeck, K.+, Niessen, F. *, Stein, R.*

*Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany
+ The Australian National University, Canberra, Australia

Corresponding Author kdittmers@awi-bremerhaven.de

Abstract

We studied the impact of sea level fluctuations on sediment dynamics in the inner Kara Sea by using high resolution acoustic sub-bottom profiling, combined with the physical properties of AMS $^{14}$C-dated sediment cores.

The first model data on LGM isostatic rebound for this region led us to confined interpretation of acoustic units in terms of sequence stratigraphy. The acoustic lines were ground-truthed with dated sediment cores. We identified and mapped three characteristic sedimentary/depositional environments and established sediment masses for each facies type. Furthermore we refined the location of the eastern LGM ice margin.

Sea level fluctuations resulted in periods of sub-aerial exposure of huge shallow shelf areas. The Kara Sea Shelf is characterised by a major incised dendritic paleo-river network that is variably infilled with late Pleistocene to Holocene estuarine facies topped by marine sediments, having draping geometries.

During sea level lowstand sediments by-passed the shelf, accumulating in water depths exceeding 120 m, while Holocene sedimentation concentrated on the recent estuaries. The shelf channel system was formed during phases/episodes of sea level lowstand during the Last Glacial Maximum and subsequent marine post-glacial sea level rise across the Kara Sea Shelf. During transgression the shallow shelf was subjected to intensive reworking and erosion.

The Kara sea can be subdivided into three Facies areas: estuaries, the shelf and deeper (> 120 m water depth) lying areas that accumulated a total of $114 \times 10^6$ t of Holocene sediments.

Continuous (sub)marine channel-levee complexes located at the LGM ice margin and are most probable of late glacial to early Holocene age.

Keywords: high resolution seismic, sequence stratigraphy, sea level, shelf deposits, isostasy, Kara Sea
Chapter 3.4

1. Introduction

Glaciations played a major role in the sedimentary history of northern Siberia and in particular the Kara Sea (e.g. Mangerud et al., 2003; Svendsen et al., 2004) (Fig. 1). Ice sheets covered Scandinavia and parts of northern Eurasia and its adjacent shelf areas. In particular, the Barents and Kara Seas were covered by thick grounded ice sheets during the Early and Middle Weichselian but this cover was less extensive during the Last Glacial Maximum (LGM), being largely restricted to the Barents Sea (Lambeck et al., 1995; Svendsen et al., 2004). The Barents-Kara Sea ice sheet was grounded, and may therefore have been susceptible to rapid changes due to a sea level rise (Lindstrom and MacAyeal, 1993), possibly triggered by deglaciation in the southern or the western hemisphere. After the LGM there were well-dated brief intervals of accelerated sea level rise (originating in rapid deglaciation phases) at 14.2-13.7 ka BP and 12.6-11.4 ka BP, named melt water peak 1A (MWP1A) and 1B (MWP1B), respectively (Fairbanks, 1989).

After 14 ka BP there is evidence for increased Arctic melting documented by ice-

![Figure 1](https://example.com/figure1.png)

**Figure 1.** Working area, general overview and glaciation limits

- ice sheet limits (according to Svendsen et al. 2004)
- ice margin ca. 90-80 krys BP
- ice margin ca. 60-50 krys BP
- LGM ice margin ca. 20 krys BP
rafted debris (IRD) events (Clark et al., 1996; Norgaard-Pedersen et al., 1998, 2003). The sea level curve for Weichselian times is characterised by an asymmetric shape with relatively slow regression preferentially preserving lowstand features in the sedimentary record and higher rates of transgression (Posamentier et al., 1992; Lambeck and Chappell, 2001; Coe et al., 2003). As examples of other well studied shelf areas show, sea/base level fall results in fluvial incision phase(s) during lowstand often followed by amalgamated channel infill during subsequent transgressions (Suter and Berryhill, 1985; Miall, 1993; Shanley and McCabe, 1993; Ashley and Sheridan, 1994).

In the Siberian realm of the Arctic investigations of sea level change during the Late glacial and Holocene period have been sparse, except for regional sea level reconstructions for the Laptev Sea by Bauch et al. (2001) and Stein and Fahl (2004) who describe the sedimentary evolution of the shelf with a diachronous reduction in accumulation rates from north to south related to the Holocene transgression at the Eurasian continental margin. Kleiber and Niessen (1999) identified major riverine extensions on the shelf, while Kleiber et al. (2000) reconstructed the LGM glaciation history. On the Kara Sea Shelf, Stein et al. (2003b, 2004) reconstructed sedimentary records and interpreted their formation as a consequence of the time transgressive southward shift of the depocentre during the post-LGM sea level rise. The studies were based on dated sediment cores and no acoustic data were included.

Using sequence stratigraphic methods (Mitchum et al., 1977; Vail et al., 1977) Dittmers et al. (2003) distinguished three Subunits in the Holocene succession of the Yenisei area. At its base Unit I fills the underlying topography and shows signs of a fluvial environment grading into a deeper water draping facies at the recent sea-floor. They form a classical deepening upward sequence, reflecting a change in depositional environment in response to post-glacial sea level rise. The chronostratigraphic frame has been previously established for the restricted area of the Ob and Yenisei estuaries south of 74° N (Dittmers et al., 2003) and in the present paper we extend this onto the shelf, using the original results of Stein et al. (2002) and Dittmers et al. (2003) and incorporating more than 8,000 km of echosounding surveys of the 2001 (Stein and Stepanets, 2002) and 2003’s PARASOUND survey (Dittmers and Schoster, 2004) as well as the results for the physical properties of 14 sediment cores from the shelf, 12 of which are AMS 14C-dated.

This present investigation is mainly based upon the interpretation of widespread acoustic profiles with dense spatial coverage of the Kara Sea Shelf. In particular the focus of the seismic analysis is on:

1. mapping the spatial distribution of acoustic surface facies provinces,
2. identification and characterisation of key surfaces responsible for reflector ter-
Regional setting

The Kara Sea is a typical marginal Arctic shelf sea which is ice-covered during nearly nine months of the year (Blanchet et al., 1995; Pfirman et al., 1995). It is characterised by high rates of fluvial run-off, i.e. 404 and 620 km$^3$/y from the Ob and Yenisei rivers, and suspended matter discharge of $15.5 \times 10^6$ t/y and $4.8 \times 10^6$ t/y, respectively (measured after dam construction in the upper reaches Telang et al., 1991; Holmes et al., 2002; Rachold et al., 2003). The recent relief of the sea-floor is shaped by fluvial activity intermingling with the complex Weichselian glaciation history (e.g. Svendsen et al., 2004). A more detailed description of the physical environment is given in Dittmers et al. (2003, Chapter 3.2) and Stein et al. (2003b, 2004).

Glaciation history

The recently published outcomes of the QUEEN project (Quaternary environments of the Eurasian North (e.g. Mangerud et al., 2004; Svendsen et al., 2004) exclude a massive glaciation of the Kara Sea Shelf as proposed by Grosswald and Hughes (1999).
and limit ice thickness to 300 m for the working area.

The eastern boundary of the LGM Barents-Kara Sea ice sheet extent is now well studied with the exception of some small uncertainties in the North-East of the Kara Sea Shelf (Fig. 1). However, these areas are of major importance, because the seafloor of the western side of the LGM ice sheet limit is characterised by an uneven glacially imprinted morphology (Dittmers et al., subm., Chapter 3.5; Stein et al., 2002), whereas the inner Kara Sea to the east and south of the mapped ice extent is characterised by relatively flat relief and shallow water depth, associated with several fluvial truncations (Dittmers et al., subm., Chapter 3.5).

The interaction of different forcing factors on the Kara Sea Shelf resulted in a fairly complicated sedimentation history. These factors are a) variable sediment supply due to natural variations in both fluvial discharge and coastal erosion, b) the heterogeneous relief of the Kara Sea, c) a non-uniform sea level rise and d) the direct neighbourhood of the LGM ice sheet.

Sea level history

For the inner Kara Sea there is no field evidence of a late or post-glacial coast-line emergence (Mangerud et al., 2004). Studies of raised beaches concentrating on Novaya Zemlya (Forman et al., 1995, 2004; Zeeberg et al., 2001) reported only moderate uplift of maximally 20 m during the Holocene. This implies that isostatic rebound of the Kara Sea coast is less than eustatic rise. Thus, the maximum thickness of any ice on the shelf and adjacent coastal plain is unlikely to have exceeded about 400 m (Lambeck et al., 1996; Svendsen et al., 2004). We included AMS $^{14}$C-dated cores BP99-04 (73°24.9 N; 79°40.4 E) and BP00-38 (73°11.8 N; 73°14.3 E) and the exact depth of the terrestrial-marine transition based on pollen studies of Kraus et al. (2003) into elastic earth rebound models, defined in Lambeck et al. (2002) (Fig. 2). Sea level predictions for the two Kara Sea sites indicate a maximum rebound of 20 m before the Holocene and allow us to neglect major isostatic effects.

2. Methods

2.1 Acoustic Data Collection/Acquisition

This multidisciplinary study was carried out in the southern and inner Kara Sea (Fig. 3a) within the frame of the German-Russian research project SIRRO (“Siberian River Run-off”, Stein et al., 2003a and references therein). Data were collected during the expeditions of the RV “Akademik Boris Petrov” in 1999, 2000, 2001 and 2003 (Fig. 3a) (Stein and Stepanets, 2000, 2001, 2002; Dittmers and Schoster, 2004).
ferent echosounding systems were used. During the 2000 and 2001 echosounding surveys we employed the hull-mounted ELAC echograph of the ship (ELAC echograph LAZ 72, Honeywell-Nautik, Kiel, Germany) operating at a frequency of 12 kHz. Since 2003 RV “Akademik Boris Petrov” is equipped with a parametric ATLAS PARASOUND system (Dittmers and Schoster, 2004) and profiles presented in this study are examples from the Parasound survey of 2003 (Schoster and Levitan, 2004). This narrow-beam system utilizes the parametric effect for signal generation which is a non-linear interaction of sinusoidal waves or wavelets (Westervelt, 1963). Grant and Schreiber (1990), Rostek et al. (1991) and Spieß (1993) describe the predecessor models and the functioning in more detail. For the data presented here, a 3.5 kHz frequency was chosen as source signal (Dittmers and Schoster, 2004).

Surveys were archived as paper scrolls printed online at the time of acquisition and in digital format on CD-ROM (Niessen and Dittmers, 2002; Dittmers and Schoster, 2004). PARASOUND profiles were reproduced using SENT Version 1.0. Details are given by Dittmers and Schoster (2004). Time-to-depth conversion was done using a standard sound velocity of 1500 m/s for water (Hamilton, 1972).

2.2 Groundtruthing by gravity coring

The Gravity Corer (core diameter: 12 cm) used for our studies has a penetration weight of 1.5 t and is used with up to 8 m long core barrels. Physical properties of sediment cores were logged using a Geotek Multi Sensor Core Logger (MSCL 25). P-wave travel time, attenuated gamma counts, magnetic susceptibility and core temperature were measured. Using Geotek software the raw data were processed for determination of wet bulk density, porosity and p-wave velocity. For technical details of the Multi Sensor Core Logger set-up see Dittmers and Niessen (2002). Magnetic susceptibility is defined as the dimensionless proportionality factor of an applied magnetic field in relation to the magnetisation in the sample and has been measured with a BARTINGTON MS2C coil sensor. Calibration was carried out according to the method of Weber et al. (1997) and Best and Gunn (1999). After splitting the cores into work and archive halves, X-ray radiographs of the intermediate 5-10 mm slabs were conducted to reveal sedimentary structures (Grobe, 1987). The chronostratigraphic framework of sediment cores is based on several AMS $^{14}$C radiocarbon dates (Stein et al., 2001, 2002, 2003b; Table 1). The base of Unit I is determined on the basis of linear sedimentation rates of dated cores (Fig. 3b). The sedimentary record and physical property data were used to correlate sediment cores as well as for sequence stratigraphic interpretation. The magnetic susceptibility data reflecting the changing source signal proved to be especially helpful in this interpretation (Dittmers et al., 2003; Stein et al., 2004).
Figure 3a. Overview of cruises of RV “Akademik Boris Petrov” during several years, and position of sediment cores under investigation.
2.3 Data analysis

Sedimentary sequences have been analysed with respect to the geometry of their internal reflection patterns, their external morphology and their physiographic positions in terms of system tracts according to Posamentier and Vail (1988) and Vail et al. (1991). GPS positioning provided accurate location and orientation of the profiles and sediment stations/cores. The GIS programme Global Mapper (version 5.10) was used for visualisation of our elevation model based on our soundings in combination with nautical charts (Dittmers, 2001; www.pangaea.de) and for determination of the spatial extent of the different identified acoustic facies.

Sediment mass calculations were also based on those area measurements in combination with GRAPE (Gamma Ray Attenuated Porosity Evaluation) data of Dittmers et al. (2003) obtained by whole-core logging during the BP 2000 and 2001 expeditions (Dittmers and Niessen, 2002; Dittmers et al., 2003). Thickness of the hanging wall sediments of a prominent pre-Holocene unconformity (sub sea-floor thickness) was determined on sub-bottom profiles (see below). This average thickness value in con-

<table>
<thead>
<tr>
<th>Station</th>
<th>Lat (° N)</th>
<th>Long (° E)</th>
<th>Water depth (m)</th>
<th>Recovery (m)</th>
<th>Unit I (m) base</th>
<th>HST (m) base</th>
<th>TST (m) base</th>
<th>date of base of Unit I</th>
<th>AMS date source</th>
</tr>
</thead>
<tbody>
<tr>
<td>BP99-04/7</td>
<td>73° 24.9</td>
<td>79° 40.4</td>
<td>32.0</td>
<td>7.95</td>
<td>5.40</td>
<td></td>
<td></td>
<td>9,500</td>
<td>Stein et al. 2002</td>
</tr>
<tr>
<td>BP99-05/1</td>
<td>73° 30.1</td>
<td>80° 00.7</td>
<td>38.0</td>
<td>4.1</td>
<td>1.10</td>
<td>0.65</td>
<td>1.20</td>
<td>av. 10,000</td>
<td>Stein et al. 2002</td>
</tr>
<tr>
<td>BP00-07/7</td>
<td>74°39.5</td>
<td>81°08.5</td>
<td>38.0</td>
<td>7.23</td>
<td>5.60</td>
<td></td>
<td></td>
<td>Stein et al. 2002</td>
<td></td>
</tr>
<tr>
<td>BP00-14/3</td>
<td>72°55.8</td>
<td>79°47.4</td>
<td>19.0</td>
<td>2.43</td>
<td>1.80</td>
<td>1.60</td>
<td>1.95</td>
<td>&gt;6,000</td>
<td>Stein et al. 2002</td>
</tr>
<tr>
<td>BP00-22/4</td>
<td>72°33.9</td>
<td>79°54.9</td>
<td>11.0</td>
<td>4.36</td>
<td>2.40</td>
<td>1.90</td>
<td>2.50</td>
<td>no data</td>
<td></td>
</tr>
<tr>
<td>BP00-23/6</td>
<td>73°28.5</td>
<td>79°51.3</td>
<td>33.0</td>
<td>4.17</td>
<td>2.75</td>
<td>2.40</td>
<td>3.30</td>
<td>no data</td>
<td></td>
</tr>
<tr>
<td>BP00-26/4</td>
<td>75°42.5</td>
<td>77°57.6</td>
<td>68.0</td>
<td>3.96</td>
<td>2.10</td>
<td>1.80</td>
<td>3.30</td>
<td>8,545</td>
<td>Stein et al. 2002</td>
</tr>
<tr>
<td>BP00-29/4</td>
<td>76°56.2</td>
<td>85°45.8</td>
<td>68.0</td>
<td>3.29</td>
<td>2.50</td>
<td>2.30</td>
<td>2.80</td>
<td>11,400</td>
<td>Stein et al. 2002</td>
</tr>
<tr>
<td>BP00-36/4</td>
<td>76°57.7</td>
<td>81°57.8</td>
<td>66.0</td>
<td>5.63</td>
<td>2.50</td>
<td></td>
<td></td>
<td>Stein et al. 2002</td>
<td></td>
</tr>
<tr>
<td>BP00-38/2</td>
<td>73°11.8</td>
<td>73°14.3</td>
<td>20.0</td>
<td>6.52</td>
<td>6.00</td>
<td></td>
<td></td>
<td>10,000</td>
<td>Stein et al. 2002</td>
</tr>
<tr>
<td>BP01-39/2</td>
<td>77°06.7</td>
<td>86°44.9</td>
<td>106.0</td>
<td>5.38</td>
<td>4.30</td>
<td></td>
<td></td>
<td>Stein et al. 2002</td>
<td></td>
</tr>
<tr>
<td>BP01-42/2</td>
<td>75°26.2</td>
<td>86°03.7</td>
<td>49.0</td>
<td>5.72</td>
<td>4.40</td>
<td></td>
<td></td>
<td>Stein et al. 2003</td>
<td></td>
</tr>
<tr>
<td>BP01-61/7</td>
<td>76°12.9</td>
<td>75°53.2</td>
<td>111.0</td>
<td>5.66</td>
<td>3.20</td>
<td></td>
<td></td>
<td>Stein et al. 2003</td>
<td></td>
</tr>
<tr>
<td>PS2719/1</td>
<td>77°36.0</td>
<td>97°32.0</td>
<td>135.0</td>
<td>5.18</td>
<td>3.42</td>
<td>2.21</td>
<td>4.30</td>
<td>9,640</td>
<td>Stein et al. 2001</td>
</tr>
</tbody>
</table>

1) inferred from linear sedimentation rates; Fig. 3c
2) inferred by Pollan sudies of Kraus et al., 2003)
junction with the areal extent/coverage of each facies type, provides an estimate of the sediment volume of each differentiated facies area. In order to calculate the sediment mass (grain density) from the estimated volumes we use the average porosity of 66%, determined by Dittmers et al. (2003), based on GRAPE estimates from a total of 95 sediment cores (Dittmers and Niessen, 2002; Dittmers et al., 2003).
3. Results and Interpretation

3.1 Recent sedimentary acoustic facies types (surface) and their spatial distribution

Based on the interpretation of acoustic profiles we divided the working area into three distinct (surface) provinces (Facies A to C) characterised by: (1) sediment penetration above the pre-Holocene unconformity, (2) sea-bottom roughness and (3) geometry of seismostratigraphic units. These are:

Facies province A: a high mud accumulation, fluvial to marine succession with laterally consistent reflectors. It is restricted to the modern estuaries up to 74° N.

Facies province B: covers the shallow and flat central shelf to water depths up to 120 m. Sediment thickness is variable and highest in depressions. Outside the depressions sediment penetration is minor. It is restricted to the shelf.

Facies province C: This shows the highest relief energy associated with little sediment accumulation and high sediment thickness with draping reflectors in channel-levees complexes. It occurs north and westward of Facies B and generally below the 120 m isobath.

Facies province A

The recent estuaries of the Ob and Yenisei rivers (Fig. 4a) are characterised by shallow water depths between 20 m and 30 m. Further to the North the facies rapidly grade into Facies province B with a pronounced decrease in sediment thickness above the basal (unconformity) reflector in the central estuary. Sediment penetration above the unconformity is high, up to 20 m in places, indicating relative rapid sediment accumulation. Sea-floor and depositional packages have smooth surfaces and a prolonged continuously stratified, layered internal assembly with draping character in the upper part of the succession. Internally, there are several discontinuities accomplished as downlapping surfaces. Acoustic profiles strikingly display the threefold division of Holocene sediments above the unconformity (Dittmers et al., 2003, Chapter 3.3). Their vertical succession represents a deepening upward sequence with fluvial character at its base (Fig. 4b).

Facies province B

Facies B covers the inner Kara Sea Shelf areas extending to the 120 m depth isobath. It is characterised by impermanent Holocene sediment penetration above the regional unconformity, due to variable accumulation of sediment. Incised fluvial channels, with different sedimentary infill-stages (Fig. 4c) are characteristic features (Dittmers et al., in press, Chapter 3.5). Underneath the prominent basal reflector, probably
due to energy attenuation by reflection, there is little penetration and only some weak local stratification, interpreted as bedrock or basal tills (see below), can be seen. The relief of the Facies B surface is smooth and almost plain, characterised by a low slope gradient of 1:6,000 on average. Several channels have been incised into the shelf and were subsequently filled such that sediment thickness increases drastically measuring up to 20 m in the channel axis (Fig. 4c). In places, well-stratified older strata show deformation and intense folding (Figs 4d, 5). The dislocations could be related to glaciotechnic ground moraine deformation during mid-Weichselian ice sheet advance (Svendsen et al., 2003) when an ice sheet advanced from the Kara Sea onto the mainland, forming a prominent chain of end-mo-
Figure 5. Spatial distribution of Facies types differentiated in Figure 4; and division into major Facies areas. Area measurements were conducted with the GIS tool Global Mapper: a-g refer to sections in Figure 8.
raines known as the “Markhida line” (Mangerud et al., 1999).

**Facies province C**

Facies C covers the area deeper than the 120 m depth contour and is characterised by an irregular morphology (Fig. 4e). There is a distinct difference in the facies between the western region with high sediment thickness in the levee/overbank area, and the north-eastern region (Fig. 5) which is acoustically well-stratified along profiles and which is also of greater lateral extent (Niessen and Dittmers, 2002; Stein et al., 2002). The overall slope gradient is steeper than 1:1,000. Associated with channels there is a well-stratified, layered levee overbank facies along the channel margins. These levees are characterised by their typical external wedge shape and drape-like character, with internal sub-parallel reflectors, thinning away from the channel axis (Fig. 4f). They are often fringed by moraines. Laterally the channels grade into overbank facies deposits, which in turn are often bordered by morainic ridges accompanied by a rapid decrease in thickness of the sediment cover (Fig. 4e). They appear in water depths between 79 and 120 m (Dittmers et al., in press, Chapter 3.5). Besides glaciated, hummocky (ground) morainic terrain, channel, levee complexes are the most prominent sedimentary features in Facies C.

Channel-levee complexes often overlie formally glaciated terrain, of probably mid Weichselian age, but their dating proves difficult. Their genesis is discussed in Dittmers et al. (subm., Chapter 3.5) in more detail.

**Features commonly observed in all facies provinces**

A few acoustic features are observed in all facies described above. These include acoustic voids, ice gouges and fluvial channels. A widespread phenomenon is the occurrence of an acoustic transparent zone in the lower part of acoustic images. Apparently, the acoustic voids, are post-sedimentary in origin cutting across layering and bedding or other structures, with no penetration below (Fig. 4f). The strong backscatter at this front implies a major difference in acoustic impedance and suggests a transition to material with major difference in density. We interpret this feature as gas-saturated sediment. In the sediment cores these acoustic voids coincide with an occurrence of meta-stable mineral Ikaite (CaCO$_3$·H$_2$O) which is formed by meteoric methane (Kodina et al., 2003). This effect is encountered in all facies types at different grades (Fig. 5).

Approaching the shelf/slope break and greater water depths, intense but locally restricted ice-gouging is visible possibly originating from drifting icebergs and/or press ice ridges that plough through the sediment (Fig. 4g) (Reimnitz et al., 1972, 1978; Barnes and Lien, 1987; Løset et al., 1999).
Unit Ia bioturbated
Unit Ia partly laminated/bioturbated
Unit Ib non-bioturbated lamination
Unit II cross lamiation including erosive surfaces

BP99-05  185-210 cm
BP99-04  670-695 cm
BP99-04  245-280 cm
BP99-04  470-495 cm

Figure 6. Radiographs of main sedimentary units; each slab measures 25 cm in length; Sediment cores of the Yensie estuary; for location see Figure 3.
Outstanding features are the channels in all facies regimes; Dittmers et al. (subm., Chapter 3.5) examine these features and their evolution/development in more detail. Within Facies A, the channels are buried at the base of the sequence (Figs 4f and 4b), or on top of the succession, indicative of recent current activity (Dittmers et al., 2003). In Facies B fluvial channels are abundant and show different levels of infill. Facies C is characterised by channel-levee complexes, indicative for underwater formation, in contrast to the subaerial formation of the channels in Facies A and B.

3.2 Lithofacies

Two major siliciclastic lithofacies, Units I and II have been previously identified in the cores, expressed by a distinct change in sediment composition (Stein et al., 2002, 2003b; Dittmers et al., 2003) as dividing the sediments that accumulated before and after the Holocene transgression. This boundary between Unit I and II correlates with the strong reflector seen in the acoustic profiles.

In Unit I, the amounts of clay, silt and sand are in the range of 50-80%, 25-50%, and < 10%, respectively (Steinke, 2002; Stein et al. 2004). In some cores at the base of Unit I a subunit is distinguished that is visually characterised by a distinct lamination that has not been disrupted by bioturbation (Fig. 6). In Unit II, the sand fraction values may reach up to 75% of the grain size spectrum. Unit I is mostly unfossiliferous and bioturbation is very rare (Stein and Stepanets, 2000). It has only been cored at stations BP99-05, BP00-29, BP00-26, BP00-23, BP00-22 and in PS-2719 (Fig. 3) (Kleiber and Niessen, 2000; Stein et al., 2001). In this Unit cross-stratification is visible and individual beds are erosively capped by overlying laminated strata (Fig. 6).

3.3 Lithostratigraphy and age control

Cores penetrating Unit I are of Holocene age (Fig. 7a), except for cores BP01-39 and BP00-26. In these latter cores there is no visible distinct facies shift during sea level rise, probably related to the relatively great water depths of 109 and 66 m, respectively. However, the susceptibility signal (Fig. 7b) shows higher values at the base of both cores, indicating a transition to Unit II. The transition from Unit I and Unit II is marked by a distinct change in magnetic susceptibility with markedly higher values in Unit II (Fig. 7b). The boundary has been AMS $^{14}$C -dated to approximately 10 ka BP (Dittmers et al., 2003; Stein et al., 2002). Unit II could be dated in cores PS-2719 (11.2 ka BP, Kleiber and Niessen, 2000, Stein et al., 2001), BP00-29 (12.2 ka BP, Stein et al, 2004) and BP00-26 (10.0 ka BP, Stein et al., 2004) with confidence and in core BP99-05 (15.5 ka BP) was determined from a piece of well-preserved driftwood (Dittmers et al., 2003; Stein et al., 2003b).
Figure 7a. Sediment key cores aligned according to water depth and distance from the recent coastline, lithological logs indicated. AMS 14C dates indicated by arrows; major facies divisions according to Stein et al. (2004), division into acoustic units according to Dittmers et al. (2003)
Acoustic Facies on the inner Kara Sea Shelf: implications...

Figure 7a. Sediment key cores aligned according to water depth and distance from the recent coastline, lithological logs indicated. AMS 14C dates indicated by arrows; major facies divisions according to Stein et al. (2004), division into acoustic units according to Dittmers et al. (2003)

Figure 7b. The sequence of cores shown in Figure 5a); magnetic susceptibility curves shown in 10^-5 SI sensor units, on the basis of which subdivision into sequence stratigraphic units has been conducted. Depositional units shown in dependence of water depth and age.

Inferred LST/FSST TST HST
4. Discussion

Seismic Sequence Stratigraphy

Sea level changes can be used for regional correlation of stratigraphic units (Vail et al., 1977; Posamentier and Vail, 1988; Posamentier et al., 1988) and act as a good link between geomorphology, sedimentology and stratigraphy (Schumm, 1993). A prominent feature visible in profiles across the region is a strong irregular uniform reflector, separating two major acoustic units (Unit I and Unit II) (Figs 4) (Stein et al., 2002, Dittmers et al., 2003). Dittmers et al. (2003) interpreted this feature as an erosional surface representing an unconformity similar to the “upper regional unconformity” (URU) described by Solheim and Kristoffersen (1984) for the Barents Sea. A similar feature was reported by Polyak et al. (2000) from the Pechora and western Kara Sea and by Kleiber et al. (2001) for the Laptev Sea.

After Mitchum and Van Wagoner (1991) seismic sequences are defined as genetically related successions of strata without internal unconformities, arranged in system tracts and bounded by unconformities and thus representing depositional sequences/environments. Ideally a sequence consists of four system tracts: low stand system tract (LST); transgressive system tract (TST); highstand system tract (HST), succeeding a regressive system tract (RST) or forced RST (FRST), if it is related to a global sea level fall (Coe et al., 2003). The major difference between the RST and LST units is that the LST sediments belong to the new sequence and lie above the sequence boundary. These latter comprise subaerially exposed areas during the onset of sea level rise whereas the RST is associated with falling sea level which, in our study area, is the time interval before the LGM.

The FRST is represented by the regional unconformity on the shelf and its deposits can only be found in deeper areas that have not been subaerially exposed. Polyak et al. (2002) report a high sediment accumulation beyond the 120 m depth contour for the Kara Sea that could represent a lowstand wedge in the sense of Wetzel and Allia (2000).

Figures 8 a-c give examples of the transition of the estuarine Facies A towards Facies B. The TST is characterised by moderate sea level rise manifested in a phase of non deposition/sediment by-pass (Vail 1987; Van Wagoner et al., 1987). The subsequent HST is equally well developed, with constant sediment thickness that indicates a relatively deeper water environment at time of deposition.

Facies B (Fig. 8d-f) is best preserved in depressions, created by channels cut into the Kara Sea Shelf during sea level lowstand, in the older sediments (Chapter 3.5). Infilling started earliest with the LST fluvial and possibly estuarine facies as in the lower
reaches of Facies A. In Figure 8d an example of a filled channel is shown with parallel reflectors and traces for lateral channel migration in the RST (dark grey). In Figure 8e the same evolution is visible for the upper part of the succession, but in the lower part, making up the LST strata, beds are much more inclined as compared to Figure 8d. This can be explained by relatively strong currents. We interprete those formations as depositional current controlled cross bedding. Figure 8e shows a filled channel with evidence for weak current activity during LST sedimentation since there has been no levelling of the old relief. The TST and HST grade into draping deposits almost leaving no relief differences to the surrounding sea-floor indicative of deeper water. Figure 8f shows an example of an unfilled channel with lateral accretion, indicative of paleo-fluvial activity that formed a meander bend with the undercut slope side to the east and typically inclined point bar sedimentation to the west. Only a thin cover of HST draping

![Interpreted sections of differentiated key Facies A and their sequences stratigraphically interpretation showing depositional units; system tracts have been developed in the response to changing sea level](image-url)

Figure 8. Interpreted sections of differentiated key Facies A and their sequences stratigraphically interpretation showing depositional units; system tracts have been developed in the response to changing sea level
sediments overlies the channel, indicating relative sediment starvation. All successes display deepening upward conditions.

A channel-levee complex typical for Facies C is shown in Figure 8g. Boundaries can only be inferred in Unit I and between Unit I and Unit II, because as discussed above, it is possible that they contain FRST sediments and there is no unconformity visible. Furthermore the exact dating remains unclear, because they overlie formerly glaciated terrain of probably early to mid Weichselian age and on the other hand they are often bordered by LGM moraines, which could imply co-existence of the channel-levees and the ice sheet. However, Dittmers et al. (subm.) in Chapter 3.5 deal with the fluvial evolution of the Kara Sea and their underwater prolongations (channel-levee complexes) in more detail.

**Ground truthing of acoustic units by sediment core data**

In the inner Kara Sea it is difficult to directly link recovered gravity cores to seismo-stratigraphic key horizons because the majority of the cores consists of homogeneous marine to estuarine muds with only little variation in the grain size composition and uncharacteristic changes in sediment colour. Furthermore the limited penetration depth of the gravity corer does not always cover the complete sedimentary interval of interest, with the additional problem that areas of non-soft sediment deposition cannot be cored. AMS $^{14}$C datable material, especially in formerly glaciated terrain is very rare.

With respect to this sedimentary homogeneity the investigation of physical properties, in particular the analysis of magnetic susceptibility, sheds new light on the sedimentary record, including important information about the sediment source. As outlined by several workers, magnetic susceptibility is a useful tool for the determination of terrestrial influx component, especially from the Yenisei branch (Dittmers et al., 2003; Stein et al., 2002, 2003b). High concentrations of ferromagnetic materials, especially magnetite, lead to high values of low-field magnetic susceptibility (Robinson, 1993). Magnetic susceptibility is commonly used as an indicator for the magnetic mineralogical composition of sediments and proved to be valuable for lateral core correlation (e.g. Kleiber and Niessen, 2000). High values of magnetic susceptibility characterise the Kara Sea, because magnetite minerals are supplied via the Yenisei River from the basaltic Putoran Mountains (Stein et al., 2002, 2003b, 2004; Dittmers et al., 2003).

The main stages of the flooding and the sequence stratigraphic division of sedimentary units are characterised as follows:

Unit II is characterised by the highest values of magnetic susceptibility, often coinciding with an increased sand content. These sediments belong to the FRST and LST. They are characterised by high-energy transport mechanisms and relative prox-
Figure 8. Interpreted sections of differentiated key Facies B and C and their sequences stratigraphically interpretation. showing depositional units; system tracts have been developed in the response to changing sea level
imity to the source inferred from the high magnetic susceptibility values. To the top the transition to Unit I is characterised by higher values of magnetic susceptibility. Frequently a spike occurs at the base of Unit I, as visible in cores PS-2719, BP99-05, BP00-22 and BP00-23 (Fig. 7b), and described by Kleiber and Niessen (2000) for the Laptev Sea. This spike is generally dated between 10-13 ka BP and reflects a depositional peak. The peak is explained by the enhanced decay of a continental (Putoran) ice sheets and increased fluvial discharge rates due to climatic warming (Kleiber and Niessen, 2000; Dittmers et al., 2003), coinciding with accelerated sea level rise (Fairbanks, 1989). During the deposition of Unit Ia water depth is greatest in relation to the preceding units. The absolute distance to the high susceptibility bearing material is greater. Transport energies are generally lower resulting in finer particles. The upper part of Unit I displays the lowest susceptibility values and is related to the HST. The maturity of sediments progressively indicating, reflected in finer particles manifested in an elevated clay content.

Sequences stratigraphic evolution and sea level cycle

The interaction of the global transgression and the smooth slope gradient of the Kara Sea Shelf resulted in very high lateral transgression rates, exposing huge areas to intense reworking and subaerial erosion, particularly during times of large meltwater release such as MWP1 and MWP2 (Fairbanks, 1989), averaging to 100 m/y for the Holocene. This physical process resulted in a superimposition of system tracts rather than in a migrating depositional equilibrium surface as suggested by Swift and Thorne (1991) leading to a facies succession in the sense of Walther`s Law (Middleton, 1973). It is very likely, for example, that during the LGM sea level lowstand an estuarine environment similar to that of the Facies A in the present Ob and Yenisei estuaries prevailed at the present 120 m isobath in the Facies C sediments.

An interpretation of the entire seismic unit configuration in dependence of the relative/eustatic sea level is given in Figure 8 and Table 2. The whole flooding history of the shelf is reconstructed in Figure 9. All terms used in the following discussion are used in the sense of Vail et al. (1987) and Posamentier and Vail (1988).

Regressive and lowstand system tracts (RST) 15 ka BP (Fig. 9 a)

We interpret the top of Unit II as a sequence bounding unconformity eroded during the regression, representing a Type 1 sequence boundary in the sense of Van Wagoner et al. (1988). The top of this Unit is irregular (Fig. 8). During the sea level lowstand the present shelf areas were transformed into alluvial plains acting mostly as non-deposition or sediment by-passing zones with sediment accumulation beyond the
a) Late Weichselian Regressional/Lowstand system tracts (RST and LST) before 15 ka BP

b) Transgressive system tract (TST) 15 ka BP- 8 ka BP

c) Transgressive/Highstand system tract (TST/HST) 8 ka BP-5 ka BP

d) Highstand system tract (HST) since 5 ka BP

Figure 9. Major steps of sea level states and distribution of land and sea during flooding/transgression history of the Kara Sea. Major steps in conjunction with stratigraphic units. The portion of the global eustatic sea level display in each illustrator is marked by the red bold line
a) depicts period when shoreline moved seaward during last regression
b) LGM shelf subaerially exposed, major phase of channel incision
c) onset of the transgression shoreline progressively moves southward towards the Siberian mainland; channels and incisions start to get filled
d) modern situation
shelf edge. Channels were formed on the shelf erosively in Facies A and B while channel-levee complexes developed in Facies C’s deeper water environment.

In sediment cores this is manifested by high transport energies indicated by high sand content (Fig. 7a), cross lamination (Fig. 6), high susceptibility values (Fig. 7b), and erosional surfaces.

The sequences boundary is a discontinuity on the shelf and grades conformably into a non-distinguishable correlative conformity in Facies C (Coe et al., 2003).

Transgressive system tract (TST) 15 ka BP- 5 ka BP (Figs 9 b and c)

During late lowstand relative sea level began to rise, forming first onlapping strata onto Unit II in a transgressive surface as well as the base of Unit Ic, the transgressive system tract (Fig. 8).

Sediments on the shelf (< 120 m water depth) comprising this time interval are of fluvial origin. Shortly after the LGM, shoreline transgressed over the shelf area involving a sea level rise of more than 80 m. Submarine accommodation space was rapidly increased so that sedimentation could keep pace with sea level rise, resulting in a deepening upward sequence. In facies A-C this is manifested by high accumulation rates

Table 2. Summary of seismic stratigraphy/depositional window of differentiated units, the age estimates refer to the area of the Ob and Yenissei estuaries (Dittmers et al., 2003) and represent a rough framework; note that for deeper water depths major surfaces (RS & MFS) yield older ages.

<table>
<thead>
<tr>
<th>characteristic features</th>
<th>Sequences stratigraphic interpretation</th>
<th>rel. sea level/vertical location</th>
<th>depositional time window</th>
</tr>
</thead>
<tbody>
<tr>
<td>- whole unit has draping character</td>
<td>HST Unit Ia</td>
<td>maximum flooding surface (MFS)</td>
<td>5 ka BP</td>
</tr>
<tr>
<td>- well stratified, downlaps at the base</td>
<td>TST Unit Ib</td>
<td>transgressive surface (TS)/ ravinemet surface (RS)</td>
<td>8-8.5 ka BP</td>
</tr>
<tr>
<td>- acoustic masking due to gas in the sediment</td>
<td>LST Unit Ic</td>
<td>sequence boundary</td>
<td>15 ka BP</td>
</tr>
<tr>
<td>=&gt; deposition in estuary setting with low current influence</td>
<td>RST Unit II</td>
<td></td>
<td></td>
</tr>
<tr>
<td>- well stratified downlapping reflectors at the base</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- on the shelf often assymetric channel infills</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- sediment thickness increases away from channel axis</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>=&gt; deposition in alternating marine-fluvial environment with moderate current influence</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- draping irregular top of Unit II</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- assymetric depositional bodies; indicative for current influences on sedimentation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- occurrence at the topographic lowest part of incised channels</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- top characterised erosional surface</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- irregular highly reflective toplayer</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- seldom internal stratification</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>=&gt; older sediments incised during sea level lowstand</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- whole unit has draping character</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- well stratified, downlaps at the base</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>- acoustic masking due to gas in the sediment</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

=> deposition in estuary setting with low current influence

- well stratified downlapping reflectors at the base
- on the shelf often assymetric channel infills
- sediment thickness increases away from channel axis

=> deposition in alternating marine-fluvial environment with moderate current influence

- draping irregular top of Unit II
- assymetric depositional bodies; indicative for current influences on sedimentation
- occurrence at the topographic lowest part of incised channels
- top characterised erosional surface
- irregular highly reflective toplayer
- seldom internal stratification

=> older sediments incised during sea level lowstand
expressed by high relative thickness of this Unit (Fig. 8). The transgression formed the transgressive surface due to the action of waves, a shore face ravinement in the sense of Swift (1968), eroding underlying strata. This surface defines the boundary between subunit Ic to subunit Ib, dated to 8.5 ka BP in the Yenisei area (Table 2). Sediments exhibit a distinct susceptibility spike at the base of the TST grading into elevated values (Fig. 7b), and are commonly finely laminated without bioturbation (Figs 6 and 7a).

To the top the TST is bound by the maximum flooding surface at ca. 6 ka BP marking the highest rate of available accommodation space between sea and sediment surface and relatively deepest water conditions during sea level rise, manifested in high accumulation rates.

**Highstand system tract (HST) since 5 ka BP (Fig. 9 d)**

Since approximately 6 ka BP sea level approached a relatively stable position. The deposition of draping HST (Fig. 8) started at least 5 ka BP. The topmost sediments in all facies areas have draping character, implying weak currents in relatively deep water. During this depositional episode sediments prograde-agrade above the maximum flooding surface and marine shelf sediments associated with low relative sea level form the “condensed section” as a result of slow rate of sediment accumulation of (hemi-) pelagic sediments as introduced by Mitchum (1977). Sediment cores show low magnetic susceptibility values (Fig. 7b), parallel bedding often obscured by bioturbation and are often rich in bivalves (Figs 6 and 7a).

**Evolutionary scenario and paleo-environmental reconstruction**

Especially for Facies C, as noted earlier, there is no existing confined age model. Channel-levee complexes often overly formally glaciated terrain, of probably mid Weichselian age, but their dating proves difficult. A key to answer this question is the interrelation of channel-levee complexes with the bordering LGM morphology (Fig. 10). The channel levee complexes overly glaciated ground (Fig. 10) and therefore postdate the LGM. A striking feature is the fact that the levees often terminate abruptly towards the contact to LGM morainic terrain with the typical hummocky relief of LGM origin (Polyak et al., 2002; Stein et al., 2002). In these key sections sedimentation on the overbank facies takes place on higher ground than the ice sheet occupied area.

On the otherhand it is possible that the terminal moraines acted as barriers. However, it is not exactly clear when the marine ice sheets decayed, but they could have existed until 15 ka and even have survived a major part of the marine transgression, which would give an early Holocene age.

The widespread association of channel levee complexes with the LGM margin (Fig.
11) displays a complex interrelationship. The general geometry of the channels and their morphology hold clear evidence for the sedimentation under current influenced conditions (e.g. Flood and Damuth, 1987). This evidence neglects a genesis in an ice dammed lake setting. This in turn, if our interpretation of the relative chronological succession is correct, implies that there was a steady (suspension) flow through the channels axis during levee/overbank formation. Our interpretation of the channels in Figure 11 seems to give evidence of a submarine flow direction somewhere to the north. This implies that the ice sheet could not have been closed the passage to the north, also

![Figure 10. Acoustic section of channel-levees bordering glacially overprinted terrain. Note the abrupt change in sediment thickness in the transition from channel-levee facies to LGM ice overlain area. Note that channel deposition (shaded area) did not occur in the left side of the section although it lies topographically lower.](image)

![Figure 11. Enlargement of Last Glacial Maximum ice margin. We refined our former results and conclude a much more lobate form of the ice border.](image)
indicating a post LGM date for the channel levee complexes. Moreover the fact that some channels are situated westward of the maximal ice extent (Fig. 11) implies a formation after the LGM.

However, their genesis is discussed in Chapter 3.5 in more detail. It is not possible to establish absolute dates but interpretation of acoustic profiles reveals important information about the relative chronology.

Figure 12 summarises the evolution of the Kara Sea Shelf schematically in a Wheeler-type diagram (Wheeler, 1958). The widespread occurrence of fluvial channels on the whole shelf, up to 700 km north of the recent coastline (Chapter 3.5), speaks in favour of a long lasting subaerial erosional period. These channels are partly unfilled and well preserved. They are quite mature in terms of auto-cyclic fluvial development. The majority belongs to the meandering river regime (Chapter 3.5). complexes.

**Sediment volume**

Sediment accumulation on the Kara Sea Shelf can be estimated from acoustic data assuming that the widely observed near sub bottom unconformity represents the Holocene transgressional surface. Occasionally, difficulties were faced when mapping the unconformity due to the presence of gas (Figs 4 and 5).

The lateral resolution of our sounding data proved not to be suitable for the calculation of contour plots. Although having a huge data set, the individual expeditions did

![Diagram of acoustic facies on the inner Kara Sea Shelf](image)
not cover the Kara Sea in an equivalent pattern producing areas, as for example the Yenisei Estuary with very high sounding density, while coverage of the shelf holds some gaps across which makes it difficult for automatic procedures to interpolate. Attempts to evaluate the sediment thickness by software utilising kringing algorithms (for e.g. Surfer and Ocean Data View) failed, because cruise tracks have a preferred and irregular spatial distribution.

Therefore, we used a different, simplified approach. Having defined the surface facies types A-C (Figs 4 and 5) we chose areas for each individual type where the spatial coverage of soundings is high, and estimated an average value of sediment thickness (Fig. 13, Table 3).

We refined our estimates for the Yenisei Estuary by a total of 800 line kilometres of new PARASOUND profiles (Dittmers and Schoster, 2004). Our calculations for Facies A, the Yenisei and Ob estuaries, $8.0 \times 10^{10}$ t and $14.1 \times 10^{10}$ t, respectively (Table 3), lie in the same range as described by Dittmers et al. (2003) with slightly lower values for the Yenisei Estuary.

For Facies B, covering the largest area with the least sediment thickness, the outcome sums up to $468$ km$^3$ sediment volume and a mass of $42.2 \times 10^{10}$ t (Table 3). The

Table 3. Sediment mass estimates for differentiated Facies A-C (Figs 3 and 4). For the calculation of the sediment mass an average porosity of 66 % has been used (Dittmers et al., 2003).

<table>
<thead>
<tr>
<th>Holocene sediment thickness (m)</th>
<th>A ESTUARY</th>
<th>B: SHELF</th>
<th>C GLACIAL</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>OB</td>
<td>Yenisei</td>
<td>west</td>
</tr>
<tr>
<td>soundings in sample area</td>
<td>534</td>
<td>157</td>
<td>554</td>
</tr>
<tr>
<td>Min:</td>
<td>2</td>
<td>2</td>
<td>0</td>
</tr>
<tr>
<td>Max:</td>
<td>21</td>
<td>23</td>
<td>15</td>
</tr>
<tr>
<td>Mean:</td>
<td>8</td>
<td>8</td>
<td>3</td>
</tr>
<tr>
<td>spatial extend of FACIES type (km$^2$)</td>
<td>19,697</td>
<td>11,532</td>
<td>15,8718</td>
</tr>
<tr>
<td>sediment Volume (km$^3$)</td>
<td>157</td>
<td>89</td>
<td>468</td>
</tr>
<tr>
<td>Netto sediment mass($10^{10}$ t)</td>
<td>14</td>
<td>8</td>
<td>42</td>
</tr>
</tbody>
</table>
Figure 13. Acoustic sediment penetration above post-glacial unconformity. Boxes indicate areas on the basis of which average thicknesses for each Facies type have been determined.
two areas of Facies C accumulate to nearly the same sediment volume, i.e. for the western facies 266 km$^3$ and 24.0 x 10$^{10}$ t respectively and 285 km$^3$ and 25.7 x 10$^{10}$ t (Table 3). In total the sediment mass adds up to 114 x 10$^{10}$ t. These values are in good agreement with the results of Stein and Fahl (2003). They based their estimates mainly on AMS $^{14}$C-dated sediment cores and in comparison to our data these estimates rather represent an underestimate. Stein and Fahl (2003) come out with a total of 123 x 10$^{10}$ t averaged for the Holocene, their data base extends as far as 82° N, while in this study we only considered data south of 78° N. Therefore, their values must be reduced by approximately 20 x 10$^{10}$ t.

Our data based on acoustic profiling rather overestimate the sediment mass especially for Facies C for the following reasons.

Sedimentation in Facies C is not interrupted by clear unconformities on acoustic profiles, thus it is not exactly clear which time interval is included by sediments overlying the unconformity. In Facies A and B the unconformity itself is of post-LGM age, but sedimentation in Facies C started earlier than on the adjacent Kara Sea Shelf, and the unconformity grades into a conformal surface, because water depth was sufficiently deep (Coe et al., 2003).

**Sediment sources**

It is beyond the scope of this work to account for all the factors, often impossible to attribute correctly, controlling the sediment input and output so we rather try to evaluate the sedimentary mass in an evolutionary sense. Besides the ambiguity of an absolute age model it is evident that sedimentation in Facies C occurred during sea level lowstand and possibly during the LGM as summarised in the conceptual model in Figure 14. This would account for the high sediment mass included by this facies type. The sediment making up the northern sections of Facies A (Figs 8b and c), closely resembles the conceptual model of a sequence by Vail et al. (1987) with most of the LST probably missing. These sediments are virtually absent in Facies A and B, where sediments by-passed and accumulated in Facies C. It is possible, though that portions of LST strata accumulated in small sinks and “survived” the sea level lowstand, but these are comparatively small, negligible values.

During transgression over the very shallow slope of the Kara Sea Shelf the in-situ sediments were subjected to several erosion processes related to the formation of the transgressional surface, as e.g. intense reworking due to wave-action and subsequent shore face ravinement formation (Thorne and Swift, 1991). Although, owing to the long subaerial exposure, large areas were probably permafrozen with abundant ice complexes, consisting of alluvial plains and river channels (Alekseev, 1997). High erosion
rates as well as coastal retreat rates increased 5 m per year in permafrozen soils due to thermokarst erosion by the formation of thermocirques (Rekant et al., 2005).

Typically, the upper 5-15 m may be removed during transgression (Walker, 1992), depending on the storm wave base and re-occurrence ratio of storms. Hine and Snyder (1985) found for the North Carolina shelf that (shoreface) erosion stripped off sediments and formed the ravinement surface. As known from the well-studied Eel margin, 50 % of the sediment above the unconformity are older reworked bedrock strata (Burger et al., 2001). We think it reasonable to account at least the same value for the Kara Sea Shelf with its gentle slope, exposing vast shelf areas to the action of waves and storms.

Most parts of the shelf are flooded since 8-9 ka BP (Figs 7b and 9) and since then terrestrial (fluvial) material has been dispersed on the shelf.

Dittmers et al. (2003) established the first parameter-based sediment budget for the region. It emerged that the estuaries effectively acted as “marginal filter” (Lisitzin, 1995) and filtered out about 90 % of the suspended and 30 % of the dissolved material, summing up to $25.5 \times 10^{10}$ t for the Ob and to $32.4 \times 10^{10}$ t for the Yenisei for the last 7,500 years, the time they assumed the estuaries to have been active under recent conditions. However the “real” volumes of sediment accumulated in the estuaries are lower, indicating weak material export (Dittmers et al., 2003). This material is thought to
accumulate in Facies B. By subtracting the sediment mass of the Facies A \((25.5+32.4)\times 10^{10}\) t – \(22.1 \times 10^{10}\) t the net export to Facies B is estimated to = \(35.8 \times 10^{10}\) t.

For the Holocene, taking the modern numbers, this would mean that during the first 2500 years another total of \(7.4 \times 10^{10}\) t of suspended and \(19.2 \times 10^{10}\) t of dissolved fluvial material by-passed Facies A. Assuming an estuarine setting in Facies B at the river ocean boundary, it would filter \(6.6 \times 10^{10}\) t of suspended and \(5.8 \times 10^{10}\) t of dissolved material while \(0.8 \times 10^{10}\) t of suspended and \(13.2 \times 10^{10}\) t of dissolved matter would have sedimented in Facies C. However, it is not exactly determinable how much of the material is deposited in Facies B and how much is passed on to Facies C. Furthermore it is difficult to assess how much material was delivered to Facies C during sea level lowstand and how much reworked Facies B material is incorporated into Facies C.

It is still unclear how much the coastal erosion effects the total sediment balance (e.g. Stein et al., 2004). There are values of Romankevich and Vetrov (2001) of \(109 \times 10^6\) t/y, which are definitely too high, because recent interpretation of satellite images of Rachold and Cherkashov (2004) indicate erosion rates in the order of \(30 \times 10^6\) t/y (Vasiliev et al., 2005).

5. Conclusion:
The Kara Sea can be subdivided into three Facies provinces:
• Facies province A: a high mud accumulation, fluvial to marine succession with laterally consistent reflectors. It is restricted to the modern estuaries up to 74° N.
• Facies province B: covers the shallow and flat central shelf to water depths up to 120 m; sediment thickness is variable and concentrates in depressions. Outside the depressions sediment penetration is minor. It is restricted to the shelf.
• Facies province C: with the highest relief energy associated and little sediment accumulation and high sediment thickness with draping reflectors in channel-levees complexes. It occurs north and westward of Facies B and generally below the 120 m isobath.
• Sediments above the regional unconformity are of Holocene age for Facies A and B, while in Facies C constant deep water conditions prevailed, with no distinct unconformity.
• Late Weichselian to Holocene strata consist of a threefold division, according to the concept of sequence stratigraphy. including
• The total post-LGM sediment mass through all three facies areas sums up to \(114 \times 10^{10}\) t.
• Constant deposition occurred continuely in Facies C, with higher sedimentation rates
during sea level lowstand when all material by-passed Facies A and B. Sediment accumulation progressively shifted towards the modern coastline and the recent estuary of Ob and Yenisei, making up Facies A.

- There is a significant component of reworked material in Facies B making up at least 50% of the sediment mass.

**Acknowledgements**

We are grateful to the crew of RV “Akademic Boris Petrov” for their support. Many thanks to Jens Matthiessen and Volker Rachold for their helpful advices. Special thanks to Frank Schoster and Mathias Kraus for many fruitful discussions. This study has been performed within the German-Russian research project “Siberian River Runoff (Sirro)”, financial support by the German Ministry of Education, science, Research and Technology (BMBF) and the Russian Foundation of Basic research is gratefully acknowledged.

Data are available at www.pangaea.de.
3.5 Late Weichselian Fluvial Evolution on the Southern Kara Sea Shelf, North Siberia
3.5 Late Weichselian Fluvial Evolution on the Southern Kara Sea Shelf, North Siberia

Dittmers, K.*, Niessen, F. and Stein, R.
Alfred Wegener Institute for Polar and Marine Research, Bremerhaven, Germany
Corresponding Author. kdittmers@awi-bremerhaven.de

Abstract

Glaciations had a profound impact on the global sea level and particularly on the Arctic environments. One of the key questions related to this topic is how the discharge of the Siberian Ob and Yenisei rivers did interact with a proximal ice sheet? In order to answer this question high-resolution (1-12 kHz), shallow-penetration seismic profiles were collected on the passive continental margin of the Kara Sea Shelf to study the paleo-drainage pattern of Ob and Yenisei rivers. Both rivers incised into the recent shelf, leaving filled and unfilled river channels and river canyons/valleys connecting to a complex paleo-drainage network.

These channels have been subaerially formed over a period of regressive phase of the global sea level during the Last Glacial Maximum. Beyond recent shelf depths of 120 m particle transport is manifested in submarine channel-levee complexes acting as conveyor for fluvially-derived fines. In the NE area uniform draping sediments are observed. Major morphology determining factors are a) sea level fluctuations and b) LGM ice sheet influence. Most individual channels show geometries typical for meandering rivers and appear to be an order of magnitude larger than recent channel profiles of gauge stations on land. The Yenisei paleo-channel dimensions are than the Ob examples and could be originated by additional water release during the melt down of LGM Putoran ice masses.

Asymmetrical submarine channel-levee complexes with Channel depths of 60 m and more developed, in some places bordered by glacially dominated morphology. Channels situated on the shelf above 120 m water depth exhibit no phases of ponding and or infill during sea level lowstand.

Keywords, Late Weichselian/Last Glacial Maximum, river run-off, sea level, shelf, channel incision, Kara Sea

1. Introduction

The exact spatial ice sheet extent during the late Weichselian = Late Glacial Maximum (LGM) glaciation is still under debate, particularly in the Siberian Arctic. Large scale ice domes as proposed by Grosswald (1980) are not realistic for the Russian sector of the Arctic. Recent outcomes of the QUEEN project ("Quaternary Environments of the European North") summarised in Svendsen et al. (2004) rather give evidence for relative small ice masses in the Siberian sector, with some ambiguity about the exact spatial extent of such an ice cover, especially on the NE mainland (Alexanderson et al., 2001).
Especially for the Kara Sea one of the key questions related to this subject is: how did the discharge of the Ob and Yenisei rivers interact with a proximal ice sheet? Glaciations have sustained impact on the discharge pattern of large streams, in particular in high latitudes, by I) influencing the global water budget and II) by the physical presence of large ice sheets.

I) In the first place the build-up of huge ice bodies/sheets “consumes” water from the ocean, resulting in a world-wide eustatic sea level fall (Fairbanks, 1989; Chappell et al., 1996). This has profound effect on the longitudinal profile of a river. The river has to adjust to a new slope gradient which can locally result in pronounced erosion and deposition and in a total reorganisation of the whole drainage pattern and direction (Allen, 1992; Schumm, 1993; Leopold, 1994).

II) The second mode of action is the physical interaction between ice sheets and adjacent river systems. River systems may either be deflected by ice sheets or get totally dammed. This has also been proposed by several workers for the middle Weichselian glaciation in Siberia, when the Ob and Yenisei rivers were thus re-routed southward into the Caspian Sea via the Aral Sea, instead of a northward drainage into the Kara Sea (Mangerud et al., 2003, 2004).

There were several major pro-glacial lakes in the Arctic realm like the late Weichselian Lake Agassiz (e.g. Clark et al., 2004) in front of the Laurentide ice sheet, the middle Weichselian Lake Komi in central Siberia (e.g. Mangerud et al., 2004) and a White Sea lake (Mangerud et al., 2001a; 2001b). Their degradation is often coupled with cataclysmic dam failures as documented in the Altai Mountains (Baker et al., 1993) possibly resulting in earth’s greatest flood.

**Glaciation history**

The general eastern boundary of the LGM Barents Sea ice sheet extent is now well-studied with the exception of some small uncertainties in the north-eastern area of the Kara Sea Shelf (Fig. 1; Svendsen et al., 1999, 2004; Alexanderson et al., 2001; Mangerud et al., 2002; Polyak et al., 2002; Stein et al., 2002). There is still an enigmatic, isolated record of an ice advance towards the Taimyr peninsula in the late Weichselian (Svendsen et al., 1999; Alexanderson et al., 2001), forming the LGM part of the North Taimyr ice-marginal zone (NTZ, Alexanderson et al., 2001). It is still under debate whether this Taimyr ice sheet was a locally isolated ice cap or whether it was linked to the Barents and Kara Sea-ice sheet. In case this ice sheet covered the northern Kara Sea Shelf, it might have blocked the major Siberian rivers during the Holocene as postulated by Polyak et al. (2002) and as proposed for the early Weichselian (Grosswald, 1980; Arkhipov et al., 1995; Mangerud et al., 2001a, 2001b).
Study Area

In the Arctic Ocean continental shelves make up more than 50 % of the entire area, a significantly larger proportion compared to other world oceans (Jakobsson, 2002). The Kara Sea Shelf (Fig. 1) occupies 10 % of the whole Arctic area. Being very shallow with an average water depth of about 50 m for the inner Kara Sea, this shelf area is in
particular sensitive to eustatic sea level changes between glacial and interglacial time periods. This probably resulted in a profound impact on the drainage pattern of modern river extensions on the Kara Sea Shelf.

**Modern river run-off**

Today the Ob and Yenisei rivers supply more than 30% of the total Arctic freshwater volume and some of the suspended sediment into the Arctic Ocean with a strong inter-annual variation (Gordeev et al., 1996; Holmes et al., 2002; Rachold et al., 2003). The Ob and Yenisei discharge 429 km$^3$y$^{-1}$ and 620 km$^3$y$^{-1}$ respectively, through their estuaries into the Kara Sea (Aagaard and Carmack, 1994).

Sediment fluxes have been investigated by Stein et al. (2002, 2003b, 2004). For a detailed summary of factors acting on riverine originated sediments and their distribution on the shelf see Stein et al. (2004), Dittmers et al. (2003) and citations therein. Water masses supplied by the two rivers pass the estuaries, whereas the main portion of the suspended and minor dissolved matter flocculates and is deposited here (Lisitzin, 1995; Dittmers et al., 2003). The remaining, by-passing matter accumulates on the Kara Sea Shelf, only a minor amount is transferred into the deep Arctic areas (Stein and Fahl, 2003). Research mainly focussed on sediment fluxes (Stein et al., 2002, 2003b, 2004; Dittmers et al., 2003) but little has been known about paleo-discharge.

**Evidence of ancient fluvial activity on the recent shelf**

During sea level lowstand fluvial activity rather resulted in widespread channel incision on the shelf, typical for several LGM shelf areas (Schumm, 1993) and as for example the well studied New Jersey continental margin (Evans et al., 2000). This process was probably enhanced during Termination I (Fairbanks, 1989) after the LGM with increasing river run-off.

On the Kara Sea Shelf widespread occurrence of channels documents active fluvial discharge during lowered sea level and subaerial exposure, in contrast to an ice-dammed lake scenario (Stein et al., 2002; Dittmers et al., 2003).

Nansen (1904) was the first to observe incisions on the sea-floor of the Kara Sea Shelf. The first extensive high-frequency sonar investigations of the bottom relief were carried out by Johnson and Milligan (1968) who published results of geomorphologic investigations of the Kara Sea and reported unfilled channels of unknown origin on the shelf. Later Russian workers constructed charts of Quaternary fluvial features including reconstructions of fluvial valleys, drainage networks and watersheds of rivers on the basis of widespread seismic surveys (Lastochkin, 1977; 1978) but did not present the basis of their data sets, leaving a tentative picture of visible fluvial features on the shelf so far. Sediment investigations of the south-western Kara Sea region show
that sediments consist of fluvially-derived material, whereas subsequent Holocene sediments show stronger marine influences (Polyak et al., 2000).


The main objective of this study is to determine the extent and character of fluvial features visible on the sea-floor on the inner shelf of the Kara Sea concentrating on the morphometric dimensions of identified channels. The fluvial geometries bear important paleo-hydrological implications (Sidorchuk and Borisova, 2000; Sidorchuk et al., 2001). This information can be used to develop a better understanding of the paleo-riverine discharge and, thus, sediment dispersal and dynamics on the southern Kara Sea Shelf in the past. We mainly interpret high-resolution acoustic reflection profiles (2-12 kHz) to identify the morphology of the Kara Sea-bottom between 70° N and 78° N (Fig. 1) and to trace and measure individual channels. Furthermore, the behaviour of Siberian river run-off during the late Quaternary glaciation (MIS 2) and subsequent transgression is reconstructed.

Dittmers et al. (2003) established the first regional seismostratigraphic framework concentrating on the Ob and Yenisei estuaries. The most prominent feature is an erosional boundary, formed during subaerial exposure of the shelf, subsequently conformably overlain by marine sediments dating back into the Holocene (Dittmers et al., 2003, Stein et al., 2003b). This boundary at the base of channel-fill occurred during the transgressive postglacial sea level rise (Chapter 3.4). Channel incision is favoured during phases of extreme relative sea level lowstand as in the middle and late Weichselian. Subsequent channel infilling occurred during sea level rise - the Holocene transgression - see Dittmers et al. (2003) for detailed discussions.

2. Data Base and Methods

2.1 Geophysical data collection/acquisition

Data were collected during four expeditions of RV “Akademik Boris Petrov” in 1999, 2000, 2001 and 2003 (Fig. 2) using several different echosounding systems summarised in Table 1. For echosounding surveys in 1999-2001 we used the hull-mounted ELAC echograph of RV “Akademik Boris Petrov” (ELAC echograph LAZ 72, Honeywell-Nautik, Kiel, Germany) operating at a frequency of 12 kHz. Since 2003, RV “Akademik Boris Petrov” is equipped with a parametric ATLAS PARASOUND system, operated at a frequency of 3.5 kHz (Table 1).
2.2 Morphometric Analysis

Fluvial channels depicted by echo-soundings were morphometrically analysed. Channels were characterised on the basis of their classical geometric features basically described by channels` depth and width e. g. Langbein and Leopold (1966), Leeder (1973) and Schumm (1977) (Fig. 3). The following features were measured: channel width, channel depth (average vertical distance from thalweg bottom to base of levee sediments) and water depth (from sea level to levee crest) (Fig. 3). Channels width and depth were measured below Holocene sediment infill to account for channel formation during incision. Holocene sediment thickness can easily be identified by the basal unconformity underlying the Holocene sediments (Fig. 4), separating Units I and II (Dittmers et al., 2003; Stein et al., 2004). Channel depth and width are synonymously used with bankful stage coinciding with bankful discharge defined by Rosgen (1994).

In our data, most relief energies of channel cross sections are underestimated because track-lines are rarely exactly perpendicular to channel thalwegs and orientation of cruise track to the morphology is random. However, determination of channel depth is reliable in all cases. Nevertheless, these morphological features may still function as a useful qualitative, descriptive parameter. Furthermore, it must be underlined that the majority of all cruise tracks is concentrated in shallower water depths while on the outer shelf data density is lower. Thus, more data on shallow water channels have been collected.

On the basis of echo-soundings, together with additional information from navigational charts, a detailed digital elevation model of the Kara Sea Shelf relief was compiled.
3. Results

3.1 Topography

In general, the Kara Sea Shelf is characterised by a gentle slope gradient ranging from an average of 1:1,351 in a representative section between 73°30' N 73°18' E and 78°00' N 73°18' E north of the recent Ob Bay to 1:4,166 along a more easterly section between 73°30' N 79°40' E and 78°00' N 79°40' E north of the Yenisei Estuary (Fig. 2). The topography remains relatively flat to a depth of the present 50 m contour line where a distinct change in slope gradient from an average of 1:6,000 to more than 1:1,100 is observed, defining a major shelf/slope break. The whole inner shelf has a smooth relief with elongated deeps and troughs aligned in northerly direction in some places. At the shelf/slope break the relief gets very rough with high relief intensity and elevation changes of more than 50 m within lateral distances of a few hundred metres. These features are related to the eastern boundary of the grounded LGM Barents/Kara Sea ice sheet (Stein et al., 2002; Polyak et al., 2002).
3.2 Channel geometry and morphological parameters

Channels have concave-up erosional bases with different levels of sedimentary infill. In the western area, associated with the LGM glaciation margin, there are typical levees which form the channel margins. They are characterised by their typical external wedge shape, with internal sub-parallel layered reflectors, drapes, and thinning away from the channel axis. The analysis of echosounding profiles generally reveals two different types of channels defined by their morphometric and morphogenetic character (Fig. 4):

Type I consists of filled or buried channels, characterised by a deflection of the prominent basal reflector representing the top of Unit II (Dittmers et al., 2003). These channels are filled with younger well-stratified sediments (Figs 4a and b). The majority of these channels is dominated by parallel reflectors of the channel infill (Fig. 4a). In some cases these sediments display asymmetric fill patterns (Fig. 4b). Channel incision depth rarely exceeds 15 m, and reaches an average of 10.4 m (Table 2). In some places the sea-bottom smoothly ponds these incisions by leaving not more than 1 metre of relief. Channel width is 650 m on average (Table 2). The majority of this

<table>
<thead>
<tr>
<th>channel type</th>
<th>Water depth (m)</th>
<th>bankful width (m)</th>
<th>bankful depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>type I Filled channel</td>
<td>Min: 22.0</td>
<td>370.0</td>
<td>4.0</td>
</tr>
<tr>
<td></td>
<td>Max: 50.0</td>
<td>9600.0</td>
<td>35.0</td>
</tr>
<tr>
<td></td>
<td>Mean: 35.5</td>
<td>650.0</td>
<td>10.4</td>
</tr>
<tr>
<td>type II OB</td>
<td>Min: 25.0</td>
<td>250.0</td>
<td>12.0</td>
</tr>
<tr>
<td></td>
<td>Max: 80.0</td>
<td>13500.0</td>
<td>55.0</td>
</tr>
<tr>
<td></td>
<td>Mean: 44.1</td>
<td>2767.8</td>
<td>22.2</td>
</tr>
<tr>
<td>type II Yenisei</td>
<td>Min: 14.4</td>
<td>300.0</td>
<td>4.4</td>
</tr>
<tr>
<td></td>
<td>Max: 122.0</td>
<td>30000.0</td>
<td>106.4</td>
</tr>
<tr>
<td></td>
<td>Mean: 56.8</td>
<td>5178.7</td>
<td>28.5</td>
</tr>
<tr>
<td>type II Channel Levee</td>
<td>Min: 71.0</td>
<td>1230.0</td>
<td>13.0</td>
</tr>
<tr>
<td></td>
<td>Max: 120.0</td>
<td>35200.0</td>
<td>100.0</td>
</tr>
<tr>
<td></td>
<td>Mean: 97.3</td>
<td>10344.0</td>
<td>49.7</td>
</tr>
</tbody>
</table>
Chapter 3.5

Figure 4. Differentiated channel types; for location see Fig 2. Acoustic profiles of different river channel types: a - d) 3.5 kHz Parasound (Dittmers and Schostek, 2004), e) 12 kHz ELAC (Niesse and Dittmers, 2002) echograms.
channel type clusters predominantly in areas with water depths of 36 m or less.

Type II channels are made up of open, or partly filled, channels (Figs 4 c and d). There is some varying amount of sediment accumulation inside the channel axis. This type of channels has been subdivided on the basis of shelf morphology into a paleo-Ob River branch up to 79° E (Fig. 4) and a paleo-Yenisei River branch of channels east of 79° E (Figs 4c). These two types seem to represent and reflect two different morphologies mainly induced by different shelf/slope gradients (Fig. 2). In general, the gradient steepens to the north with increasing water depth, minimising channel width and increasing depth.

The paleo-Ob River channels occur at a mean water depth of 44 m with average bankful-depths of 22.2 m and an average width of almost 2.7 km (Table 2). Compared to the Ob system Yenisei paleo-channels show similar bankful-depths with an average of 28.5 m and distinctively wider developed morphologies averaging to 5.2 km. They cluster around a water depth of nearly 57 m (Table 2).

Nine Channel-levee complexes could be identified in the NW of the working area. In average they are found in 97.3 m water depth, are 10.3 km wide and 50 m deep.

In Figure 5a the relationship between channel-bankful-depth versus water depth is displayed. Type I channels display the lowest depth values. Type II data points scatter widely but show a general decrease in this ratio with increasing water depth, in particular below 50 m water depth. Ob River channels are generally distinguishable from Yenisei River channels by smaller widths. Channel depth in relation to water depth is shown in Figure 5 b. Type I channels are grouped below 50 m of water depth with low scattering and only few individual Channel depths exceeding 15 m. Type II channels show a wider scatter of Channel depth values with the difference that Ob River channels are dominant at higher water depths, displaying higher values than the Yenisei examples and are more scattered. In general, there is a pronounced drop of Channel depth in shallower than 50 m water depth.

The same trend is depicted in Figure 5c), which displays water depth versus channel width/ depth ratio (w/d-ratio). This ratio decreases significantly in more than 50 m of water depth. Type I channels are characterised by highest w/d-ratios, while the Type II channels of the Ob River form the other end-member, with low w/d-ratios. Yenisei River channels lie between both. Again the ratio diminishes distinctly with increasing water depth.

Channel-levee complexes are characterised by high bankful-depth values and relatively small w/d-ratios.
3.3 Spatial distribution of channels on the shelf

Evidence for fluvial incision is visible in the bathymetry especially at the 30 m isobath which is deeply entrenched in the northward extension of the modern river estuaries (North of 74° N; Figure 6). Downstream thalweg slope on the inner shelf, up to 120 m of water depth, gradients of Ob and Yenisei paleo-channels lie in the range of 1:1,351 and 1:4,166, respectively (Fig. 2). There is an apparent difference in the drainage patterns of the Ob and Yenisei rivers paleo-channels on the shelf (Fig. 6). The Yenisei paleo-river pathways display a more dendritic pattern, probably related to channel branch formation around several obstacles such as islands and topographic highs. There are more channel branches in the Yenisei River drainage system, generally trending north-eastward, with smaller tributaries orientated north-westward.

Clusters of paleo-channels have been mapped as river branches which extend across the entire shelf. According to the morphology and abundance of identified channels, paleo-river pathways have been defined (Fig. 6). Branch here describes the general path of channel(s) with incorporation of sea-

Figure 5. Channel morphometric parameters
a) Relationship of water depth (m) versus channel-bankful width (m)
b) Relationship of water depth (m) versus channel-bankful-depth (m)
c) Relationship of water depth (m) versus channel width/depth ratio
Figure 6. Channel distribution of identified channels, based on channel distribution and sea-floor morphology (Dittmers, 2004). A tentative distribution of major paleo-valleys has been inferred. Type I channel represented by filled triangles, Type II by circles. Paleo-channel pathways have been defined by morphology specified by the abundance of mapped channels.
floor morphology (Chapter 3.1). They are large scale depressions with all types of different channels (filled and unfilled), resulting from fluvial incision. Buried and filled (Type I) river channels are concentrated on the inner shelf and are very widespread above 30 m water depth (Fig. 6). Type II river channels show no distinct preferred allocation (Fig. 6) in dependence to water depth. Channel-levee complexes exclusively formed in water depths exceeding 70 m and are associated with the LGM ice margin westward of the Kara Sea Shelf.

4. Discussion

4.1 Isostasy

Little has been known about isostatic movement, the relaxation after crustal downbending from an ice sheet load, in the inner Kara Sea. In particular, there is still bias about the extent and especially the thickness and loading duration of such an ice sheet. However, these are important factors for the calculation of crustal movements.

Zeeberg et al. (2001) postulate a rebound rate of 1-2 mm/y from an estimated ice load of 1,000 m for Novaya Zemlya resulting in an uplift of 8-16 m for 8 ka but they present no estimates for the Kara Sea. Gataullin et al. (2001) report modest uplift following deglaciation from the south-eastern Pechora Sea, but this area is closer to the proposed centre of the Barents Sea-ice dome. Studies of raised beaches concentrating on Novaya Zemlya reported only moderate uplift of maximally 20 m during the Holocene (Forman et al., 1995, 2004; Zeeberg et al., 2001). This implies that isostatic rebound of the Kara Sea coast is less than eustatic rise and that consequently, the maximum thickness of any ice on the shelf and adjacent coastal plain is unlikely to have exceeded about 400 m (Lambeck et al., 1996; Svendsen et al., 2004). First isostatic rebound modelling results (Chapter 3.4) underline the field evidence.

A radiocarbon dated gravity core BP00-07/5 (632 cm core length) was obtained from the shallow Kara Sea Shelf (43 m water depth) at 80° E, 74° N approximately 150 km south of the islands (Fig. 2). Simstich et al. (2003) argue that linear extrapolation of 5 m uplift at Izvestia Islands for the last 5 ka backwards to 8 ka BP reveals a maximum isostatic depression of 8 m, here. In contrast, Svendsen et al. (2004) resume from the available sea level data that there was no transgression above the present sea level in the Kara Sea region. There is no evidence for raised beach lines on the Siberian mainland in our working area implying neglectable isostatic movement. The same conclusion is ascertained by Dittmers et al. (2003) by comparing dated sediment cores from the Yenisei Estuary with the Fairbanks (1989) sea level curve. Estimates of eustatic sea level during LGM indicate that sea level was approximately 120 m lower than today (Chappell and Shakleton, 1986; Fairbanks, 1989, Chappell et al., 1996).
Neglectable isostatic movement is also supported by first occurrence of dinoflagellates in core BP99-04/7 in correlation with Fairbanks’ (1989) sea level curve (Kraus et al., 2003; Stein et al., 2003b, 2004)

4.2 Fluvial style of Ob and Yenisei rivers and paleohydrol-ogy

In case the profile of a river does not coincide with the equilibrium profile, the river will re-adjust its gradient in order to reach equilibrium, by either erosion into its substratum, or the river will aggregate by accumulating sediment on the floodplain and the channel (Allen and Posamentier, 1993). When the river has attained its equilibrium profile, it is said to be at grade (Allen, 1992). When relative sea level exposes a shelf with a different slope than the equilibrium gradient of the river (Schumm, 1993), valley incision is commonly thought to occur at the knickpoint that develops at the topographic shoreline and shelf-slope break retreating upriver.

Channel classification

Leeder (1973) and many other researchers (e.g. Osborn and Stypula, 1987; Schumm and Khan, 1972) used w/d-ratio of stream channels as a classical parameter to characterise dimension and shape controlled by the hydraulic regime. Flood and Damuth (1987) were the first to apply fluvial geometries of subaerial formed meandering rivers for submarine channel levee systems of the Amazon fan successfully.

Fielding and Crane (1987) assembled data from numerous studies to differentiate styles of fluvial geometry. The w/d-ratio is suited here to discriminate between braided, meandering, and straight channels in particular. Although the relationship is not straightforward and there are other “degrees of freedom” (Morisawa, 1985) to be considered, as e. g. initial relief, lithology, climate, and vegetation (Schumm, 1977, 1981), it serves as a good first order discrimination and classification scheme. Our data plotted in this scheme illustrates the different fluvial styles of Type I and II channels and between Ob and Yenisei channels (Fig. 7). Interestingly, size relationships on the shelf represent a straight prolongation of recent channel proportions on land (after Chavlov, 1994), which is a typical feature of stream systems (Leopold and Maddock, 1953), indicating that they are related features. Generally, their morphologies display features an order of magnitude larger than the dimensions of recent fluvial cross sections on land.

Meandering rivers make up the majority of all channels under investigation. All channels on land and the majority of Type I and II channels represent this class (Fig. 7). Except for Type II channels and minor recent channels, most channels represent the “fully developed” meandering streams (Collinson, 1978). There is a slight difference between Type II Ob and Yenisei River channels. The Yenisei River channels cluster
Table 3. Discharge characteristics of several gauge stations in the West Siberian Lowland; data by courtesy of Aleksey Y. Sidorchuk (sidor@yas.geogr.msu.su)

<table>
<thead>
<tr>
<th>Station/Location</th>
<th>Mean Annual Discharge (m³/s)</th>
<th>Bankful Discharge (m³/s)</th>
<th>Width (m)</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kubyshev</td>
<td>20.4</td>
<td>184</td>
<td>76.8</td>
<td>4.4</td>
</tr>
<tr>
<td>Demianka Lytkevskie</td>
<td>145</td>
<td>953</td>
<td>138</td>
<td>6.6</td>
</tr>
<tr>
<td>Konda Chantyria</td>
<td>76.1</td>
<td>102</td>
<td>83.3</td>
<td>2.6</td>
</tr>
<tr>
<td>Konda Urai</td>
<td>123</td>
<td>235</td>
<td>123</td>
<td>3.2</td>
</tr>
<tr>
<td>Konda with Bolchary</td>
<td>286</td>
<td>551</td>
<td>262</td>
<td>3.3</td>
</tr>
<tr>
<td>Konda with Altai</td>
<td>327</td>
<td>424</td>
<td>208</td>
<td>3.4</td>
</tr>
<tr>
<td>Amnya with Kazym</td>
<td>61.6</td>
<td>52.7</td>
<td>97.2</td>
<td>1.2</td>
</tr>
<tr>
<td>Sev Sosva with Nyazamvol</td>
<td>90.1</td>
<td>369</td>
<td>130</td>
<td>3.2</td>
</tr>
<tr>
<td>Sev Sosva Kimskauli</td>
<td>269</td>
<td>468</td>
<td>169</td>
<td>5.1</td>
</tr>
<tr>
<td>Sev Sosva with Sarynya</td>
<td>652</td>
<td>780</td>
<td>679</td>
<td>4.4</td>
</tr>
<tr>
<td>Sev Sosva Sigm</td>
<td>750</td>
<td>3190</td>
<td>497</td>
<td>6.7</td>
</tr>
<tr>
<td>Lyapin with Saran Paul</td>
<td>248</td>
<td>1360</td>
<td>520</td>
<td>5.5</td>
</tr>
<tr>
<td>Sob Kharp</td>
<td>30.1</td>
<td>150</td>
<td>63</td>
<td>2</td>
</tr>
<tr>
<td>Polui with Polui</td>
<td>132</td>
<td>364</td>
<td>133</td>
<td>4.6</td>
</tr>
<tr>
<td>Nadym Nadym</td>
<td>446</td>
<td>192</td>
<td>160</td>
<td>1.9</td>
</tr>
<tr>
<td>Pur Uurony</td>
<td>801</td>
<td>4470</td>
<td>1300</td>
<td>4</td>
</tr>
<tr>
<td>Pur Samburg</td>
<td>899</td>
<td>4610</td>
<td>1240</td>
<td>4.9</td>
</tr>
<tr>
<td>Pyakua Tarkoakeice</td>
<td>298</td>
<td>2740</td>
<td>5.6</td>
<td>4.4</td>
</tr>
<tr>
<td>Basin Sidorovsk</td>
<td>1070</td>
<td>3750</td>
<td>625</td>
<td>8.1</td>
</tr>
<tr>
<td>Ob Barnaular</td>
<td>1470</td>
<td>3960</td>
<td>338</td>
<td>9.1</td>
</tr>
<tr>
<td>Ob Stone Calculus</td>
<td>1560</td>
<td>4960</td>
<td>703</td>
<td>6.8</td>
</tr>
<tr>
<td>Ob Mogochaiko</td>
<td>4110</td>
<td>11700</td>
<td>1260</td>
<td>7.7</td>
</tr>
<tr>
<td>Ob Kolpashevo</td>
<td>4260</td>
<td>11400</td>
<td>835</td>
<td>9.6</td>
</tr>
<tr>
<td>Ob with Belogore</td>
<td>10300</td>
<td>21800</td>
<td>1300</td>
<td>16</td>
</tr>
<tr>
<td>Ob Salekhard</td>
<td>12300</td>
<td>25900</td>
<td>2240</td>
<td>13.1</td>
</tr>
<tr>
<td>Berd Masyanino</td>
<td>20.1</td>
<td>132</td>
<td>107</td>
<td>1.6</td>
</tr>
<tr>
<td>Berd Iskitim</td>
<td>45.8</td>
<td>300</td>
<td>110</td>
<td>2.3</td>
</tr>
<tr>
<td>Tom Meschedurechenski</td>
<td>168</td>
<td>1660</td>
<td>334</td>
<td>4.2</td>
</tr>
<tr>
<td>Miras Ust</td>
<td>67.1</td>
<td>616</td>
<td>103</td>
<td>2.9</td>
</tr>
<tr>
<td>Mundaybash Mundaybash</td>
<td>22.6</td>
<td>281</td>
<td>69.8</td>
<td>1.8</td>
</tr>
<tr>
<td>Top Ters Aspen Pleso</td>
<td>47.6</td>
<td>481</td>
<td>80</td>
<td>2.6</td>
</tr>
<tr>
<td>Comp Ters Monashica</td>
<td>82.2</td>
<td>587</td>
<td>142</td>
<td>1.9</td>
</tr>
<tr>
<td>Taidon Medvezhika</td>
<td>47</td>
<td>454</td>
<td>100</td>
<td>2.4</td>
</tr>
<tr>
<td>Chulym with Balbahts</td>
<td>102</td>
<td>599</td>
<td>113</td>
<td>3.9</td>
</tr>
<tr>
<td>Chulym with Zryanskoie</td>
<td>500</td>
<td>1530</td>
<td>467</td>
<td>3.7</td>
</tr>
<tr>
<td>Chulym with Sergeevo</td>
<td>737</td>
<td>2760</td>
<td>594</td>
<td>4.6</td>
</tr>
<tr>
<td>Chulym Kommunarka</td>
<td>794</td>
<td>2900</td>
<td>508</td>
<td>5.5</td>
</tr>
<tr>
<td>Urup with Izzytcheva</td>
<td>31.9</td>
<td>131</td>
<td>92.7</td>
<td>1.4</td>
</tr>
<tr>
<td>Kiyacuecue Kiyacue Marininsk</td>
<td>148</td>
<td>719</td>
<td>162</td>
<td>3.7</td>
</tr>
<tr>
<td>Kiyacuecue Kiyacue Okunevko</td>
<td>177</td>
<td>781</td>
<td>192</td>
<td>4.5</td>
</tr>
<tr>
<td>Ket Maximkin Yar</td>
<td>243</td>
<td>793</td>
<td>208</td>
<td>4.6</td>
</tr>
<tr>
<td>Orlovka Druzhny</td>
<td>63.5</td>
<td>281</td>
<td>128</td>
<td>2.9</td>
</tr>
<tr>
<td>Chuzik Osovo</td>
<td>30.5</td>
<td>225</td>
<td>240</td>
<td>2.2</td>
</tr>
<tr>
<td>Vaslugan</td>
<td>164</td>
<td>557</td>
<td>162</td>
<td>5.6</td>
</tr>
<tr>
<td>Vaslugan Naunak</td>
<td>381</td>
<td>1020</td>
<td>244</td>
<td>7</td>
</tr>
<tr>
<td>Tym with Varzhi</td>
<td>74.3</td>
<td>567</td>
<td>128</td>
<td>5</td>
</tr>
<tr>
<td>Tym with Napas</td>
<td>182</td>
<td>612</td>
<td>165</td>
<td>3.8</td>
</tr>
<tr>
<td>Vakh with Lubchinskoe</td>
<td>504</td>
<td>1150</td>
<td>245</td>
<td>7.8</td>
</tr>
<tr>
<td>Tromwell Ermakovo</td>
<td>102</td>
<td>362</td>
<td>244</td>
<td>2.1</td>
</tr>
<tr>
<td>Agan with Var</td>
<td>127</td>
<td>630</td>
<td>117</td>
<td>5.9</td>
</tr>
<tr>
<td>Pim Pim</td>
<td>68.2</td>
<td>217</td>
<td>168</td>
<td>1.8</td>
</tr>
<tr>
<td>Lyamin decides Goshikofo</td>
<td>81.4</td>
<td>423</td>
<td>111</td>
<td>4.4</td>
</tr>
<tr>
<td>Bollugan with Ugut</td>
<td>143</td>
<td>402</td>
<td>159</td>
<td>4.4</td>
</tr>
<tr>
<td>Tobol with Zvergolovskoe Zaboloch</td>
<td>26.2</td>
<td>172</td>
<td>126</td>
<td>3.2</td>
</tr>
<tr>
<td>Tobol Mound</td>
<td>45</td>
<td>364</td>
<td>124</td>
<td>4.6</td>
</tr>
<tr>
<td>Tobol with Lipovskoe</td>
<td>809</td>
<td>3500</td>
<td>411</td>
<td>8.3</td>
</tr>
<tr>
<td>Istet Kataisk</td>
<td>23.9</td>
<td>300</td>
<td>249</td>
<td>3.8</td>
</tr>
<tr>
<td>Istet with Mekhonskoie</td>
<td>64.8</td>
<td>109</td>
<td>175</td>
<td>2.9</td>
</tr>
<tr>
<td>Istet with iastskoe</td>
<td>68</td>
<td>299</td>
<td>150</td>
<td>4.2</td>
</tr>
<tr>
<td>Mias with Novely</td>
<td>7.1</td>
<td>50.1</td>
<td>33.9</td>
<td>1.4</td>
</tr>
<tr>
<td>Mias with Sosnovskoe</td>
<td>12.4</td>
<td>105</td>
<td>36.9</td>
<td>2.5</td>
</tr>
<tr>
<td>Mias with Karahelskoie</td>
<td>14.7</td>
<td>200</td>
<td>115</td>
<td>2.5</td>
</tr>
<tr>
<td>Tura Turinsk</td>
<td>109</td>
<td>500</td>
<td>149</td>
<td>5.2</td>
</tr>
<tr>
<td>Tura Tyumen</td>
<td>178</td>
<td>1180</td>
<td>230</td>
<td>5.9</td>
</tr>
<tr>
<td>Tavida Tavda</td>
<td>462</td>
<td>1580</td>
<td>244</td>
<td>7.8</td>
</tr>
</tbody>
</table>
Late Weichselian Fluvial Evolution on the Southern Kara Sea Shelf, North Siberia

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Bankful Discharge (m$^3$/s)</th>
<th>Mean Annual Discharge (m$^3$/s)</th>
<th>Width (m)</th>
<th>Depth (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kuibyshev</td>
<td>20,4</td>
<td>184</td>
<td>76,8</td>
<td>4,4</td>
</tr>
<tr>
<td>Demianka Lymkoevskie</td>
<td>145</td>
<td>953</td>
<td>138</td>
<td>6,8</td>
</tr>
<tr>
<td>Konda Chantyria</td>
<td>76,1</td>
<td>102</td>
<td>83,3</td>
<td>2,6</td>
</tr>
<tr>
<td>Konda Urai</td>
<td>123</td>
<td>235</td>
<td>123</td>
<td>3,2</td>
</tr>
<tr>
<td>Konda with Bolchary</td>
<td>286</td>
<td>551</td>
<td>262</td>
<td>3,3</td>
</tr>
<tr>
<td>Konda with Altai</td>
<td>327</td>
<td>424</td>
<td>208</td>
<td>3,4</td>
</tr>
<tr>
<td>Amnya with Kazym</td>
<td>61,6</td>
<td>52,7</td>
<td>97,2</td>
<td>1,2</td>
</tr>
<tr>
<td>Sev Sosva with Nyaximvol</td>
<td>90,1</td>
<td>369</td>
<td>130</td>
<td>3,2</td>
</tr>
<tr>
<td>Sev Sosva Kimkiasui</td>
<td>269</td>
<td>468</td>
<td>169</td>
<td>5,1</td>
</tr>
<tr>
<td>Sev Sosva with Sartynia</td>
<td>652</td>
<td>780</td>
<td>679</td>
<td>4</td>
</tr>
<tr>
<td>Sev Sosva Igrim</td>
<td>750</td>
<td>3190</td>
<td>497</td>
<td>6,7</td>
</tr>
<tr>
<td>Lyapin with Saran Paul</td>
<td>248</td>
<td>1360</td>
<td>520</td>
<td>5,5</td>
</tr>
<tr>
<td>Sob Kharp</td>
<td>30,1</td>
<td>150</td>
<td>63</td>
<td>2</td>
</tr>
<tr>
<td>Polui with Polui</td>
<td>132</td>
<td>364</td>
<td>133</td>
<td>4,6</td>
</tr>
<tr>
<td>Nadym Nadym</td>
<td>446</td>
<td>192</td>
<td>160</td>
<td>1,9</td>
</tr>
<tr>
<td>Pur Urengoi</td>
<td>801</td>
<td>4470</td>
<td>1300</td>
<td>4</td>
</tr>
<tr>
<td>Pur Samburg</td>
<td>899</td>
<td>4610</td>
<td>1240</td>
<td>4,9</td>
</tr>
<tr>
<td>Pyaku Tarkocakes</td>
<td>298</td>
<td>2740</td>
<td>5,6</td>
<td>4,4</td>
</tr>
<tr>
<td>Basin Sidorovsk</td>
<td>1070</td>
<td>3750</td>
<td>625</td>
<td>8,1</td>
</tr>
<tr>
<td>Ob Barnaul</td>
<td>1470</td>
<td>3960</td>
<td>338</td>
<td>9,1</td>
</tr>
<tr>
<td>Ob Stone Calculuson</td>
<td>1560</td>
<td>4960</td>
<td>703</td>
<td>6,8</td>
</tr>
<tr>
<td>Ob Mogochino</td>
<td>4110</td>
<td>11700</td>
<td>1260</td>
<td>7,7</td>
</tr>
<tr>
<td>Ob Kolpashevo</td>
<td>4260</td>
<td>11400</td>
<td>835</td>
<td>9,6</td>
</tr>
<tr>
<td>Ob with Belogore</td>
<td>10300</td>
<td>21800</td>
<td>1300</td>
<td>16</td>
</tr>
<tr>
<td>Ob Salekhard</td>
<td>12300</td>
<td>25900</td>
<td>2240</td>
<td>13,1</td>
</tr>
<tr>
<td>Berd Maslyanino</td>
<td>20,1</td>
<td>132</td>
<td>107</td>
<td>1,6</td>
</tr>
<tr>
<td>Berd Iskitim</td>
<td>45,8</td>
<td>300</td>
<td>110</td>
<td>2,3</td>
</tr>
<tr>
<td>Tom Mezhdurechensk</td>
<td>168</td>
<td>1660</td>
<td>334</td>
<td>4,2</td>
</tr>
<tr>
<td>Mras Ust</td>
<td>67,1</td>
<td>616</td>
<td>103</td>
<td>2,9</td>
</tr>
<tr>
<td>Mundybash Mundybash</td>
<td>22,6</td>
<td>281</td>
<td>69,8</td>
<td>1,8</td>
</tr>
<tr>
<td>Top Ters Aspen Pleso</td>
<td>47,6</td>
<td>481</td>
<td>80</td>
<td>2,6</td>
</tr>
<tr>
<td>Comp Ters Monashka</td>
<td>82,2</td>
<td>587</td>
<td>142</td>
<td>1,9</td>
</tr>
<tr>
<td>Taidon Medvezhka</td>
<td>47</td>
<td>454</td>
<td>100</td>
<td>2,4</td>
</tr>
<tr>
<td>Chulym with Balakhta</td>
<td>102</td>
<td>599</td>
<td>113</td>
<td>3,9</td>
</tr>
<tr>
<td>Chulym with Zyryanskoe</td>
<td>550</td>
<td>1530</td>
<td>467</td>
<td>3,7</td>
</tr>
<tr>
<td>Chulym with Sergeevo</td>
<td>737</td>
<td>2760</td>
<td>594</td>
<td>4,6</td>
</tr>
<tr>
<td>Chulym Kommunarka</td>
<td>794</td>
<td>2900</td>
<td>508</td>
<td>5,5</td>
</tr>
<tr>
<td>Uriup with Izyndaeva</td>
<td>31,9</td>
<td>131</td>
<td>92,7</td>
<td>1,4</td>
</tr>
<tr>
<td>Kiyacuecue Kiyacue Mariinsk</td>
<td>148</td>
<td>719</td>
<td>162</td>
<td>3,7</td>
</tr>
<tr>
<td>Kiyacuecue Kiyacue Okuneevo</td>
<td>177</td>
<td>781</td>
<td>192</td>
<td>4,5</td>
</tr>
<tr>
<td>Ket Maximkin Yar</td>
<td>243</td>
<td>793</td>
<td>208</td>
<td>4,6</td>
</tr>
<tr>
<td>Orlovka Druzhnyi</td>
<td>63,5</td>
<td>281</td>
<td>128</td>
<td>2,9</td>
</tr>
<tr>
<td>Chuzik Osipovo</td>
<td>30,5</td>
<td>225</td>
<td>240</td>
<td>2,2</td>
</tr>
<tr>
<td>Vasiugan Naunak</td>
<td>381</td>
<td>1020</td>
<td>244</td>
<td>7</td>
</tr>
<tr>
<td>Tym with Vanzhil</td>
<td>74,3</td>
<td>567</td>
<td>128</td>
<td>5</td>
</tr>
<tr>
<td>Tym with Napas</td>
<td>182</td>
<td>612</td>
<td>195</td>
<td>3,8</td>
</tr>
<tr>
<td>Vakh with Lobchinsko</td>
<td>504</td>
<td>1150</td>
<td>245</td>
<td>7,8</td>
</tr>
<tr>
<td>Trom with Ermakovo</td>
<td>102</td>
<td>362</td>
<td>244</td>
<td>2,1</td>
</tr>
<tr>
<td>Agan with Var</td>
<td>127</td>
<td>630</td>
<td>117</td>
<td>5,9</td>
</tr>
<tr>
<td>Pim Pim</td>
<td>68,2</td>
<td>217</td>
<td>168</td>
<td>1,8</td>
</tr>
<tr>
<td>Lyamin fact Gorshkovo</td>
<td>81,4</td>
<td>423</td>
<td>111</td>
<td>4,4</td>
</tr>
<tr>
<td>Bol Iugan with Ugut</td>
<td>143</td>
<td>402</td>
<td>159</td>
<td>4,4</td>
</tr>
<tr>
<td>Tobol with Zverinogolovskoe Zaboloch</td>
<td>26,2</td>
<td>172</td>
<td>126</td>
<td>2,2</td>
</tr>
<tr>
<td>Tobol Mound</td>
<td>45</td>
<td>364</td>
<td>124</td>
<td>4,6</td>
</tr>
<tr>
<td>Tobol with Lipovskoe</td>
<td>809</td>
<td>3500</td>
<td>411</td>
<td>8,3</td>
</tr>
<tr>
<td>Iset Kataisk</td>
<td>23,9</td>
<td>300</td>
<td>249</td>
<td>3,8</td>
</tr>
<tr>
<td>Iset with Mekhonskoe</td>
<td>64,8</td>
<td>109</td>
<td>175</td>
<td>2,9</td>
</tr>
<tr>
<td>Iset with Isetskoe</td>
<td>68</td>
<td>299</td>
<td>150</td>
<td>4,2</td>
</tr>
<tr>
<td>Miass with Newly</td>
<td>7,1</td>
<td>50,1</td>
<td>33,9</td>
<td>1,4</td>
</tr>
<tr>
<td>Miass with Sosnovskoe</td>
<td>12,4</td>
<td>105</td>
<td>36,9</td>
<td>2,5</td>
</tr>
<tr>
<td>Miass with Karachelskoe</td>
<td>14,7</td>
<td>200</td>
<td>115</td>
<td>2,5</td>
</tr>
<tr>
<td>Tura Turinsk</td>
<td>109</td>
<td>500</td>
<td>149</td>
<td>5,2</td>
</tr>
<tr>
<td>Tura Tyumen</td>
<td>178</td>
<td>1180</td>
<td>230</td>
<td>5,9</td>
</tr>
<tr>
<td>Tavda Tavda</td>
<td>462</td>
<td>1580</td>
<td>244</td>
<td>7,8</td>
</tr>
</tbody>
</table>

Figure 7. Bankful width and depth of channels under investigation plotted in classification scheme according to Fielding and Crane (1987), recent dimensions of Ob and Yenisei rivers at Salekhard, Igarka, respectively; note double logarithmic scale. 1) refers to upper boundary for meandering rivers; 2) the limit of Collinson's (1975) "fully developed" meandering streams; 3) shows the bounding line for deposits of meandering rivers and represents deposits of laterally unrestricted rivers, such as braided rivers. Small filled circles stand for channel dimensions of recent gauge station of west Siberian lowland (see Table 2).
shelf physiography, whereas Type I channels probably represent a different mode of channel genesis. No channels plot in the "braided river field", but the majority of the "fully developed" meandering streams (Collinson, 1978). The Yenisei River channels cluster closely between line 1 and line 2 (Fig. 7), characterised by high w/d-ratios and some of them come close to the "braided river field", probably resulting from a lower general slope gradient. In contrast, the Ob River channels exclusively plot to the left of line 2 and closer to line 1 (Fig. 7). The Ob and Yenisei river geometries reflect different shelf physiography/gradient, whereas Type I channels probably represent a different mode of channel genesis. No channels plot in the "braided river field", but the majority of Type I plot between line 1 and 2 in the true meandering river field (Fig. 7). Channel-levee complexes are characterised by relatively high thalweg depth values and bankful widths, but depict the general trend and rarely fall into the straight channel field. It should be emphasised that they are underwater formations with different physical processes behind especially the density contrast between air and water and sediment-saturated water to sea water is highly different. But interestingly their dimensions, and morphometric behaviour is similar to subaerially formed meandering river systems (Clark and Pickering, 1996; Flood and Damuth, 1987).

The filled channels are consequently interpreted as small tributaries, existing for short distances, quickly feeding into more stable, longer lived channels that are more deeply incised, and self-adjusting to the shelf gradient (Schumm, 1993). This is reflected in higher w/d-ratios. This type could be classified as immature as it is not adapted to the equilibrium profile, because adjustment of base-level to a steady state has not been achieved. In contrast, meandering channels act as relatively stable sediment transport agents by-passing suspended material after incision (Schumm, 1981). The spatial distribution of the differentiated channel types is shown in Figure 6.

Examples of Yenisei channels are shown in Figure 8 a) to c) as typical meandering channels of the middle Kara Sea Shelf.
Submarine channel-levees are concentrated along the LGM ice margin (Figs 9 and 10). They are indicative of underwater formation as described by Flood and Damuth (1987) from the Amazon fan. Water depth of channels exceeds 80 m in all cases.

**Paleo-hydrology**

By analysing hydraulic geometry of streams, Leopold and Maddock (1953) and Matthes (1956), later refined by Dury (1964, 1965, 1967), found that in both, normal and flood conditions, stream width, depth and velocity increase as simple power functions of discharge. From records of streams and rivers in central and southwest USA, they established that the depth of a channel increases nearly as the square root of discharge, regardless whether it is a small mountainous river or a large stream (Dury, 1967). The same rules for channel width, but with a different local factor.

Rivers often act as recorders of sudden catastrophic events rather than of “normal” discharge conditions prevailing most of the year (Schumm, 1977, 1981). Channel forming stages and incision events, for example, occur during bankfull discharge, when stream power increases dramatically (Dury, 1986). In the Arctic realm these events usually coincide with the snow melt originated spring flood. Thus, all geometries might reflect only one event in an extreme situation. In our working area the quantification of pre-Holocene fluvial discharge is hard to determine from the maximum discharge channel (geometry) forming stage, as it rather reflects a momentary annually re-occurring event.

Although difficult to quantify it is evident from Figure 7 that channels on the shelf, thus, represent higher discharge events than their recent analogues on land (Fig. 7). There are several arguments for higher peak discharge that could explain the greater geometries of the channels under examination. These could be possibly related to a hinterland blocking of the Ob and Yenisei rivers as proposed by Yamskikh (1996, 1998). He reports several fluvial terraces, dating into the late Weichselian, in the upstream Ob and Yenisei rivers interpreted as polycyclic short time blocking events, in the order of months, due to ice damming in narrower parts of the river bed, combined with catastrophic release events.

Another explanation are “macro meanders”, fluvial features an order of magnitude larger than recent fluvial dimensions, formed during the LGM/postglacial transition until 14 ka BP (Sidorchuk et al., 2001). They relate these features to a very punctuated river run-off during glaciation; although the mean precipitation was lower than today, an increased discharge may have been caused by the existence of permafrost, draining all surface waters immediately towards the river.

Furthermore, river run-off volume is reduced by peat accumulation since 14 ka BP with a distinct peak during 11.5 - 9 ka BP in the West Siberian Lowlands (Kremenetski
Figure 9. Parasound run at 3.5 kHz Examples of fluvial facies neighbouring to LGM margin (Fig. 2 for location). Profiles were recorded with the 12 kHz ELAC system and Figure10 with ELAC 12 kHz. Glacially overprinted terrain is characterised by typical hummocky relief, related to glacial deformation. Channel levee facies is characterised by overdeepened channel with wide adjacent floodplains, made up of thick sediments. Note abrupt change in facies; there is no indication for glacial facies underneath fluvial facies. Moraines possibly represent terminal moraines of LGM.
Late Weichselian Fluvial Evolution on the Southern Kara Sea Shelf, North Siberia

Figure 9. Parasound run at 3.5 kHz Examples of fluvial facies neighbouring to LGM margin (Fig. 2 for location). Profiles were recorded with the 12 kHz ELAC system and Figure 10 with ELAC 12 kHz. Glacially overprinted terrain is characterised by typical hummocky relief, ... is no indication for glacial facies underneath fluvial facies. Moraines possibly represent terminal moraines of LGM.

Figure 10. ELAC 12 kHz examples of channel-levee facies neighbouring glaciated area.
et al., 2003; Smith et al., 2004). These accumulations have lead to the formation of the worlds largest peat lands possibly containing 50 % of the total worldwide peat (Smith et al., 2004). Today, this large area acts as a buffer for the whole discharge dynamics and store huge amounts of water. Another important factor for the sudden release of huge water masses is the whole post-glacial termination process, manifested for example in sediment cores by a distinct susceptibility spike after deglaciation (Kleiber et al., 2001; Dittmers et al., 2003). The peak is explained by enhanced decay of continental (Putoran) ice sheets and increased fluvial discharge rates due to climatic warming (Kleiber and Niessen, 2000; Dittmers et al., 2003). This incident does not only mark the melt down of the Circum-Arctic and Putoran ice sheets, but also the degradation of permafrost, enhanced precipitation, and the decay of mountain glaciations in the whole drainage area (Velichko, 2002; Svendsen et al., 2004). All these processes boosted up the amount of water released and drained towards the Arctic Ocean. The above arguments lead to the picture that peak discharge during late Weichselian and Termination I has been even more pronounced and enhanced, compared to today, despite of a smaller total net water discharge during glacial times.

**Difference between Ob and Yensiei paleo-channels**

To estimate the discharge of both quantitatively, we calculated the relationship of channel depth and discharge volume of recent distributary channels in order to establish a characteristic basin discharge (data provided by Sidorchuk, pers. comm., 2004). As outlined earlier the depth of the paleo-channels is always determined correctly. On the basis of this relationship we took the average depth of Yenisei and Ob paleo-channels (Table 2) and calculated both the annual discharge, 15,089 m$^3$/s and 8,494 m$^3$/s, respectively, and the bankful discharge, 20,225 m$^3$/s and 12,271 m$^3$/s, respectively (Fig. 11). As expected from Figure 7 the paleo-examples resulted from higher (peak) discharge, but furthermore the Yenisei channels indicate significant higher discharge volumes, almost twice as high as the Ob examples.

Spielhagen et al. (2005) studied sediment core PS 2456 on the Laptev Sea shelf and describe oxygen-isotope fluctuations in foraminifera caused by regional freshwater release from ice masses. Their study reveals a major excursion from 13 ka BP to 12.2 ka BP with the distinct lack of foraminifera in this interval, due to a salinity drop below critical limit. They interprete this finding with the massive partly ice-dammed river-water release of the Lena river. This event lasted for 350-400 years and they postulate a catastrophic water release of 100, 000 km$^3$ and resulting in a salinity drop of 1.2 ‰. We think this event marks the deglaciation of the Hinterland feeding the drainage area, as documented by the abundant susceptibility spike of Kleiber and Niessen (2000, 2001),
which is documented in the Kara Sea as well (Chapter 3.3 and 3.4). We would like to consider the possibility that the Putoran Mountains contributed to the paleo-discharge of the Yenisei river and could be responsible for the larger fluvial dimensions of Yenisei channels and larger paleo discharge values.

Today 80 % of the water volume drained from the Putoran Mountains flows into the Yenisei, i.e. 430 km$^3$/y, while approximately 100 km$^3$/y drain into the Kathanga and Anabar (Rachold et al. in Stein and MacDonald, 2004). We determined the portion of the Putoran Mountains that lie above 500 m with the GIS tool Global Mapper based on the IBCAO data set: 285,015 km$^2$. Assuming that 300 m of ice accumulated summing up to a volume of 85,205 km$^3$, a number in correspondence with Velichko et al.`s (1997) estimate of 80,000 km$^3$ (Table 4). We think it plausible to assume that this additional melt-water volume boosted up the Yenisei paleo-discharge and is responsible

<table>
<thead>
<tr>
<th>Ice thickness (m)</th>
<th>Total Ice volume (km$^3$)</th>
<th>Drainage area (km$^2$)</th>
<th>Water volume draining into the Yenisei (km$^3$)</th>
<th>Average annual discharge during 200 years (km$^3$/y)</th>
<th>Average annual discharge during 400 years (km$^3$/y)</th>
<th>Average annual discharge during 1000 years (km$^3$/y)</th>
</tr>
</thead>
<tbody>
<tr>
<td>100</td>
<td>28502</td>
<td>285015</td>
<td>17101</td>
<td>86</td>
<td>100</td>
<td>114</td>
</tr>
<tr>
<td>200</td>
<td>57003</td>
<td>285015</td>
<td>34202</td>
<td>171</td>
<td>200</td>
<td>228</td>
</tr>
<tr>
<td>300</td>
<td>85505</td>
<td>285015</td>
<td>51303</td>
<td>257</td>
<td>299</td>
<td>342</td>
</tr>
<tr>
<td>400</td>
<td>114006</td>
<td>285015</td>
<td>68404</td>
<td>342</td>
<td>399</td>
<td>456</td>
</tr>
<tr>
<td>500</td>
<td>142508</td>
<td>285015</td>
<td>85505</td>
<td>428</td>
<td>499</td>
<td>570</td>
</tr>
</tbody>
</table>

Table 4. Discharge estimates based on different ice thicknesses of the LGM Putoran ice cap are shown in the first row (hatched box denotes the most realistic) and annual discharge volumes additionally released into the Yensiei in the last columns for different discharge periods. Today approximately 80 % of the total Putoran water Volume is drained into the Yenisei (Rachold et al. In Stein and MacDonald 2004). The shaded boxes indicate values that match into the estimated discharge difference of Ob and Yenisei...
for the larger extensions of Yenisei paleo-channels on the shelf. Table 4 summarises the ice volume in relation to different ice thickness and discharge portions draining into the Yenisei.

It is remarkable that a sudden decay of a relatively thin ice cap (300 to 400 m, red box in Table 4), could account for the additional water volume released over a time period of several hundred years. Spielhagen et al. (2005) date their “freshwater event” after 13 ka falling into the younger Dryas. It is known from pollen studies that temperatures in middle Siberia had been higher than today (Hubberten et al., 2004). A change from dry hot to humid climate occurred since 12.5 ka BP, which could mark the triggering of the ice decay. We propose a relatively sudden warming event that led to a rapid (self accelerating) non-linear meltdown of the Putoran/Siberian ice masses. It is obvious that a complete decay of the whole ice mass could not have occurred in this short interval. However, the process could have been enhanced by additional hinterland melting in the upper reaches of the river system in the Altai mountains for example. The same processes could have happened in the Laptev Sea and boosted up the riverine water discharge.

4.3 Development/evolution of channel geometry

Channel depth and width in dependence to water depth (Table 1, Figure 5) increase and decrease, respectively, down slope as observed for natural rivers (Schumm, 1977). This confirms subaerial formation and no underwater setting as for e.g. the well-studied Amazon fan where these attributes behave inversely (Flood and Damuth, 1987). Schumm (1993) reports that channels rather incise into steeper parts of the slope than deposit material. This accounts for the progressive decrease of channel width and vice versa, increasing channel depth (Fig. 5c) with increasing water depth, which yields a steeper gradient in our working area (Fig. 2). Considering the distribution of channels versus water depth (Fig. 5a) in the light of changing sea level (Chappell et al. 1996; Fig. 12) the Kara Sea comprises some information about channel incision. In Figure 12 channel formation and infill is drafted according to their sequence stratigraphic interpretation introduced by Dittmers et al. (2003). Channel incision ceases as soon as sea level starts to rise giving (accommodation) space for transgressive sediments. Channel-fill pattern is dependent on sediment supply and relative position of the channel on the shelf during transgression. Filled channels concentrate in water depths below 50 m, probably representing an area of most intense reworking during transgression. On the other hand, terrestrial sediment supply might have abated during passage over the shelf, only reaching shallow water depths. As shown by Dittmers et al. (2003) most of the fluvially-derived material is stored in the Ob and Yenisei estu-
Figure 13 shows a longitudinal profile of the whole drainage area. The slope increases continuously from south to north and as a consequence channels incised into the shelf in contrast to an "unincised" scenario (Schumm, 1993; Sommerfield et al., 1995) where the shelf gradient is lower than in the Hinterland. As slope increases stream power, it increases the tendency for channels to incise (Schumm, 1993). During subsequent transgression, these areas of increased accommodation space are preferred sedimentation areas.

4.4 Paleo-environmental implications

It is still under debate whether there was a pro-glacial basin in front of the LGM ice sheet resulting from blocked Siberian rivers or not (e.g. Mangerud et al., 2001; Polyak et al., 2002; Stein et al., 2002). From our data there is no evidence for such a lake, a deeper/still-water sedimentation on the shelf such as draping sediments in morphological depths as proposed by Polyak et al. (2002). In contrast underneath the recent
water depth of 100 m another facies is encountered at the NW and NE corner of the working area, representing subaqueous formations. Stratigraphically they represent the upper part of a "lowstand fan" (Piper and Normark, 2001) of a "lowstand wedge" which must have been formed during sea level lowstand (Posamentier and Vail, 1988; Posamentier et al., 1992).

Channels on the shelf shallower than 100 m recent water depth

Most of the channels under investigation plot in the fully developed "meandering river" field (Fig. 7), meaning they are quite mature in terms of fluvial development. Although an absolute chronology is missing, the nearly 700 km northward extent of fluvial features on the shelf north of the Ob river, the good conservation of incised channels and their maturity, argue for a long lasting fluvial-dominated history of the Kara Sea Shelf. The findings of Mangerud et al. (1999) and Svendsen et al. (1999) and the widespread occurrence of the “markhida line” on the Siberian mainland indicate a middle Weichselian glaciation having originated in the north at approximately 60 ka BP. This former glaciation would have annihilated the most prominent geomorphological features by overriding the whole southern Kara Sea Shelf, leaving behind a typical rough ground moraine surface as visible in Figure 9 and 10. Into this newly formed topography subaerially exposed rivers started to incise, creating the regional basal unconformity (Dittmers et al., 2003) (Fig. 12), conveying sediments across the shelf towards its edge.

Figure 13. Schematic section through the Kara Sea Shelf and adjacent hinterland of the West Siberian lowland; principle slopes of morphological segments are indicated at the top; amplitude of sea level fluctuations during the Weichselian shown in the left; SL1= recent; SL2= LGM.
Deeper water (<120 m) facies
NW area

Channel-levees show a delineation of the thalweg route by the LGM ice sheet, and in some places, individual channels directly border to end moraines (Figs 9 and 10) and topographic highs which acted as dams. The marine extensions of the Ob branch clearly postdate the underlying glacial morphology. Thus, here the time of channel formation could not have been initiated before the LGM (approximately 20 ka BP). There are several examples where the areas occupied by LGM ice with its severely deformed topography (Figs 10 a and b) lie deeper than the adjacent channel-overbank complexes. This fact leads us to the conclusion that there still was a certain amount of ice that hindered the channel to enter the lower terrain and channel-levee formation occurred during the physical existence of ice bodies (Chapter 3.4). In classical glaciomarine environments sediment normally accumulates adjacent to the ice in depressions, whereas morainic facies is found in the higher ground (e.g. Dowdeswell et al., 1998; Syvitski et al., 1996; Vorren et al., 1998). According to the Fairbanks (1989) sea level curve and given the fact that isostasy only plays a minor role in our working area, a average water depth of ca. 100 m of the channel-levee complexes would give a maximum age of 14 ka BP (Fig. 14), as they represent underwater formations. Although our evidence bears no absolute chronology, the relative timing implies a synchronous or later formation of channels adjacent to the LGM ice border. It is also possible that large dead ice blocks were separated from the ice sheet during its decay, with the channels flowing around them. The ice blocks could have survived the major part of the postglacial transgression, especially when subjected to sediment entrainment during stranding on the shelf. The water column necessary to elevate an ice package can be estimated by sea-ice = 900 kg m$^{-3}$ water = 1,028 kg m$^{-3}$ densities (Fricker and Padman, 2002) meaning that the water depth needs to exceed ice thickness by approximately 114%. Channel-levee complexes have been identified at an average water depth of ca. 100 m. Some 120 m of sea water column are necessary to establish a hydrodynamic equilibrium and higher water levels are necessary to elevate an ice body with minimum thickness of 100 m; thus a relatively thin body of ice could have survived the major part of the post-glacial transgression. Probably all processes interacted and submarine channels were deflected by the remnants of the LGM ice sheet, but during its decay, there were some “break-throughs” and the flow pattern reorganised itself, finding new pathways, typical for glacial environment (Elverhoi et al., 1998, 1989). However, an exact (side scan sonar) survey of ice marginal channel-levee complexes with a very high spatial resolution could reveal the exact timing.

The Amazon fan has become a reference example for passive margin channel-levee systems and acts as a good depositional analogue for channel-levee formation.
in our working area. Mikkelsen (1997) report sedimentation differing in three orders of magnitude rates of 5 cm/ka during interglacial times with only pelagic background sedimentation and 5,000 cm/ka during glacial times when sediments are channelled to the deep sea. Typical “lifespans” of individual channel-levee systems vary between 1-3 ka (Piper et al., 1997). Once a submarine levee route has been established it is relatively stable, and major adjustments in the self-stabilizing channels pathway are unlikely, ignoring the surrounding morphology. Avulsion and abandonment of channel-levee complexes is triggered by changing sea levels, climatic controlled sediment supply and auto-cyclic processes (Flood et al., 1991): It is likely for our specific environment sea level fluctuations combined with formation and disintegration of (dead) ice caused channel fluctuation and migration.

Interestingly the channel-levee complexes are located at transition from the shelf to basin areas as for example the Novaja Zemelya trough at the shelf/slope breaks with relatively steep gradients. Direct downslope gradient (perpendicular to the ice margin) would have resulted in canyon entrenchment, typical for the upper fan (Bouma, 2000). In contrast sediment transport was oriented parallel to the shelf morphology, resulting in a shallower thalweg gradient enabling the upward growing channel-levee complexes. According to Nelson and Kulm (1973) channel-levee systems occur preferentially between slop gradients of 1:1,000 and 1:4,000. For the Kara Sea these values are found at the shelf/slope break where all of the identified channel-levee systems are located. Average sediment thickness at thalweg is 15 m, which is in good agreement to the average life span of 2 ka Documented channel-levee complexes are not likely to have been connected or formed at the same time, although channel thalweg depths could indicate a progressive northward deepening (Fig. 14). For the Kara Sea channel-levee complexes, we favour an analogon to the Amazon fan model where the documented channels on the shelf act as feeder for the channel-levee complexes formed as under water extensions of the river mouth estuarine point sources, characterised by locally high sedimentation rates. With progressive transgression these point sources migrated southward towards shallower water depths. According to this hypothesis the northernmost channel-levee system is the oldest and to the south, with shallower water depths, channels are younging.

Deeper water (<120 m) facies
NE area

The sediment accumulation in the NE corner of the working area characterised by ponded muds (Fig. 14), with high sediment thicknesses (Fig. 15). No LGM ice or glaciation traces are visible for the LGM, no major undercurrent sub aquatic channel-
Figure 14. Paleo-geographic map of Kara Sea area and fluvial run-off routes. Grey shaded area corresponds to recent 120 m isobath, and proposed sea level during LGM (Fairbanks 1989). Dashed lines indicate time transgressive position of the paleo-shoreline during the Holocene transgression according to modern bathymetry. Filled dots indicate major channel-levee systems, numbers represent recent water depth.
Figure 15. Example of high-resolution CHIRP profile of 2001 cruise (Stein et al., 2002; Niessen and Dittmers, 2002) and physical properties (Dittmers and Niessen, 2002) of dated sediment core BP01-39/2 (Stein et al., 2004).
levee complex, no ice margin setting, smooth relief. The sediment core BP01-39 (106 m water depth) shows continuous marine conditions (Stein et al., 2004) with high accumulation rates up to 10,000 BP in a core at a water depth from approximately 110 m. This indicates forced sedimentation during sea level lowstand and enhanced delivery proximal to the source, while (sea level) highstand sedimentation is much more diminished, manifested in metres of sediment deposited during 10,000 years. To our understanding, this facies is a still-water deposit formed in depressions, possibly adjacent to the ice sheet, rather than relicts of a reservoir lake expanding over the whole shelf.

Alternatively, these accumulations could be normal pro-deltaic marine sediments that accumulated during a period of low sea level stand as reported by Kleiber et al. (2001) from the eastern Vilkitsky Strait, interpreted as a delta which was possibly fed by fluvial input through the Kara Sea and no local point source. We think that the examples at the edge of the shallow shelf, the channel-levee facies in the NW area and the draping sediment accumulation facies in the NE area document the same in two different primarily morphology-controlled areas. The NW area is characterised by channelled sediment flow, while the NE is characterised by lower slope gradients and suspended sediment spread out into a wide area.

Evidence of Spielhagen et al. (2004) who reviewed the sediment composition of the central Arctic Ocean in the light of ice sheet records, supports our idea of a constant (channelised) northward sediment flux. Major deglaciation events are normally represented by ice-rafted debris (IRD) rich layers associated with elevated smectite concentrations indicative of a source in the eastern Kara Sea and western Laptev Sea area (Nuernberg et al., 1994; Stein and Korolev, 1994; Wahsner et al., 1999). In contrast, indications of melt-water events in the central Arctic Ocean during the last glacial Termination I are much smaller than those during Terminations II and III at 130 and 50 ka, respectively (Knies and Vogt, 2003; Spielhagen et al. 2004 and references therein). This excludes a large LGM ice-dammed lake scenario with its typical self accelerating fresh water outburst, as reported for Lake Agassi (Magny and Bégeot, 2004). Constant fluvial run-off, therefore, seems reasonable for the Kara Sea at that time.

5. Conclusions
There is evidence that there was a long lasting fluvial activity on the Kara Sea Shelf. The analysis of numerous channels on the Kara Sea Shelf led to the conclusion that they were formed during the last global sea level regression phase, having its peak during the LGM. There is a distinct difference between the Ob and Yenisei branch/system resulting from the different slope gradient of their paleo-drainage paths. Both systems display filled channels, preferable found in shallow water depths, and open,
or partly filled channels. We think, that our data support only a brief glacial surge across the northern Kara Sea during the LGM (Alexanderson et al., 2002) with no evidence of a large territory occupied by an ice-dammed lake for the LGM and subsequent times. Morphometric analysis of channel dimensions showed that they predominantly belong to the meandering river regime. These findings argue for a time interval of subaerial exposure and channel formation sufficiently long enough for rivers to incise and to adjust under equilibrium conditions. The channels dimensions are larger than the modern dimensions of gauge stations of Ob and Yenisei rivers, probably formed under colder climate conditions during LGM resulting in a distinct peakedness of river run-off.

Acknowledgements

We gratefully thank the crew and the Captain of RV “Akademic Petrov” for their support and cooperation. Aleksey Sidorchuk is much obliged for providing quickly and uncomplicated modern gauge data of Ob and Yenisei rivers, with river dimensions and flow velocity. Special thanks to Jens Matthiessen and Frank Schoster for many fruitful discussions and for their helpful comments.

This study has been performed within the German-Russian research project “Siberian River Run-off (SIRRO)”, Financial support by the German Ministry of Education, science, Research and Technology (BMBF) and the Russian Foundation of Basic research is gratefully acknowledged.
Part C

Synthesis
Conclusions

Key results

• How is the fluvially-derived material dispersed on the (recent) shelf: pathways of sediment? (CHAPTER 3.2, partly CHAPTER 3.3)

Silt size surface sediments exhibit a widespread bimodal grain-size distribution, indicative of at least two transport processes acting on grain-size populations in the mud fraction. We favour seasonality with different conditions affecting the sediment distribution. In the summer the whole system is controlled by strong riverine water and sediment influx. Terrestrially-derived sediment settles out of the river plume, being influenced by the prevailing surface currents, mainly induced by winds. After the particles made their way through the water column they flow density-induced along the bottom relief.

The winter situation is characterised by little direct riverine sediment and freshwater input but due to the high current energy, ice transport, and ice erosion, rather marks a period of intense sediment resuspension and redistribution. Coastal, nearshore, and estuarine areas are effectively sheltered by the fast-ice from vigour wave induced currents and finer grain-size populations are dominant.

Terrestrially-derived material is transported along the Kara Sea Shelf bottom relief. Magnetic susceptibility enlightens the pathways of signal and dispersion following the morphology predetermined by ancient incised channels with the established paleo-river “drainage network”, probably related to morphology-controlled density-driven sediment distribution for example by a bottom nepheloid layer. The exact transport processes could not be identified. The majority of sediment is trapped in the estuaries of Ob and Yenisei as visible on acoustic profiles.

• The establishment of a Holocene Sediment budget

Estuaries

The total post LGM sediment mass through all three facies areas sums up to $114 \times 10^{10}$ t. These numbers are in good agreement with results of Stein and Fahl (2003) averaging $123 \times 10^{10}$ t for the Holocene. Our calculations for Facies A, the Ob and Yenisei estuaries, $8.0 \times 10^{10}$ t and $14.1 \times 10^{10}$ t, respectively, lie in the same range as described by Dittmers et al. (2003) with slightly lower values for the Yenisei Estuary.

Shelf

For Facies B, covering the largest area with least sediment sediment mass of $42.2 \times 10^{10}$ t is estimated. Facies C accumulates a sediment volume of $49.7 \times 10^{10}$ t.

Deposition occurred constantly in Facies C, with higher sedimentation rates during
sea level lowstand when all material by-passed Facies A and B. Sediment accumulation progressively shifted towards the recent estuaries of Ob and Yenisei (Facies A) with elevating sea level. There is a significant component of reworked material in Facies B making up at least 50% of the sediment mass reducing the net sediment mass to some $24.85 \times 10^{10}$ t. It appears as if the estuaries effectively hindered Holocene sediments from entering the Kara Sea. The shelf has been supplied by sediments that “escaped” the estuaries and mostly by coastal erosion, besides the major reworked component. Facies C was built up during sea level lowstand.

**What was the late Weichselian to Holocene evolution of the inner Kara Sea Shelf (CHAPTER 3.3, 3.4 and 3.5)**

The sediments in the Ob and Yenisei estuaries can be divided into two major acoustic and lithological Units (I and II). Unit II forms the pre-Holocene basement with little acoustic penetration. During the last sea level lowstand, the erosional surface was formed on top of this unit including several incised river channels. The younger Unit I can be divided into three Subunits, Ia to Ic. The lowermost Subunits Ic and Ib show fluvial seismic features comparable to typical subaerial formed channel-levee complexes representing a fine-grained meandering river depositional environment overlain by a transgressive system tract.

We correlated these units with strata on the shelf, establishing a regional chronostratigraphic framework:

Regressive and lowstand system tracts (RST and LST) before 15 ka BP: During the sea level lowstand the present shelf areas were transformed into alluvial plains and valleys due to fluvial incision and erosion acting mostly as non-deposition sediment bypassing zones with sediment accumulation beyond the shelf edge. Formation of sequence bounding unconformity on top of Unit II.

Transgressive system tract (TST) 15 ka BP- 5 ka BP: Most sediments on the shelf (< 120 m water depth) comprising this time interval are of fluvial origin. Shortly after the LGM, shoreline transgressed over the shelf area involving a sea level rise of more than 80 m. Accommodation space was rapidly increased resulting in a deepening upward sequence. The transgression formed a shore face ravinement surface defining the boundary between subunit Ic to subunit Ib, dated to 8.5 ka BP in the Yenisei area. In Facies B Holocene sediments contain a reworked component and a recent one above the transgressive surface and maximum flooding surface at ca. 8 ka BP marking the highest rate of available accommodation space between sea and sediment surface and relatively deepest water conditions during sea level rise.

Highstand system tract (HST) since 5 ka BP: Since approximately 6 ka BP sea level has approached a relatively stable position. The deposition of draping HST started at
least 5 ka BP, implying weak currents in relatively deep water conditions comparable to the recent estuarine environment.

- **What was the fate of the large west Siberian rivers (Ob and Yenisei) during last glacial maximum sea level lowstand, and how far did they extend onto the shelf? (CHAPTER 3.4 and 3.5)**

  The analysis of numerous channels on the Kara Sea Shelf led to the conclusion that they were formed during the last global sea level regression phase. Morphometric analyses of channel dimensions showed that they predominantly belong to the meandering river regime. Dimensions of channels on the shelf are larger than recent analogues of Ob and Yenisei rivers. These findings argue for a time interval of subaerial exposure and channel formation sufficiently long enough for rivers to incise and to adjust under equilibrium conditions. Furthermore river run-off pattern was different than today with minor total run-off but more spiked main flooding events, producing larger geometries.

  The interface between LGM morainic morphology with under water extensions of the channels implies a deflection of channel-levee systems by the ice sheet and furthermore a constant drainage to the north. This finding invalidates the proposed link between the Barents Sea-Novaya Zemlya ice sheet and the Taimyr Peninsula glaciation: with the caveat that an exact dating is missing.

  There is no geomorphic or sedimentological expression of a large reservoir storage. Our data support only a brief glacial surge across the northern Kara Sea during the LGM (Alexanderson et al., 2002) with no evidence of a large territory occupied by an ice-dammed lake for the LGM and subsequent times. Therefore, we conclude that the lake could only have endured a couple of hundred years - a time span not sufficient enough traceable in the geological record on our data basis.
5. Synthesis

This chapter discusses the results of the thesis relating the main topics to unsolved problems in this particular working area. The discussion of the critical facts specifies the focus of future research. An open question and controversially discussed subject is the question whether there was an LGM ice-dammed lake or not (see Mangerud et al., 2004 and Svendsen et al., 2004, for comprehensive summaries). Another open question is: how was the drainage of Ob and Yenisei rivers influenced by the proximal ice sheet? This issue has been addressed previously in this thesis Chapters 3.4 and 3.5 and to a minor extent in Chapter 3.3, but here we want to tackle the topic in a more speculative way and show where the future work has to focus on.

Landforms and features diagnostic for an ice-dammed lake after Clarke et al. (2004) and references therein are:

1. incised outlet channels, showing fluvial erosion, often characterised by rapid water release due to catastrophic (cataclysmic) dam failure (Jökulhaups; Björnsson, 2003)
2. cliffs eroded along the paleo-beach line by lake waves; these should be visible in topography and be documented in the sedimentary record as beach gravel and sand deposits
3. deltaic deposits at the places where rivers enter the lake basin
4. fine-grained lacustrine sedimentation in the central lake, rich in ice rafted debris (IRD)
5. intensive ice gouging on the lake bottom in the relatively shallow waters (Dyke, 2003; Josenhans and Zevenhuizen, 1990)

Possible traces that such a lake could have left behind and evidence from our data:

1. Although we report the widespread abundance of channels on the shelf (Chapter 3.5), no particular lake outlet channel could be found in our working area. It should be characterised by deep incision due to upper flow regime during cataclysm with progressive incision during water release (e.g. Baker et al., 1973, 1993). These channels occur not only at the lake outlets, but are also found in the central lake basin, as reported by Josenhans and Zevenhuizen (1990). Deeper water (<100 m) channels found in the working area, are growth structures, typical for channel-levee complexes (Flood and Damuth, 1989), rather than incised outlet channels.
2. In our soundings and the related elevation model no pronounced beach cliff or paleo-shoreline is visible on the shelf that could have been formed by a lake. Possibly this large scale depression could be revealed by future research including high-resolution side scan sonar surveys.
3. The predecessors of Ob and Yenisei rivers should have had their expression in an ice-dammed lake, where they would have dumped their sediment load. These deltas must have been massive because the sedimentation regime shifts rapidly from moving water to still water and the rivers release a large amount of their (suspension) load, although their net suspended freight volume would be smaller
than today because of the dryer climate. Their typical external form makes it difficult to oversee these features, and it seems not very likely that they have been completely removed by erosion during transgression, but no evidence for deltaic features could be found within our working area.

4. It is difficult to differentiate between typical marine muds encountered in our working area and sediments originated in a still water-lake environment. But these sediments should be expressed as draping units in acoustic data, because their sedimentation is not current-controlled. There are no widespread draping sediments.
of greater lateral extent visible in our working area.

5. Chapter 3.4 describes a preferred occurrence of ice gouges at approximately 50 m water depth. It is theoretically possible that they were formed in a pro-glacial lake setting, but it remains questionable whether they survived the subsequent reworking processes accompanied with the transgression. Given the fact that they are not covered by sediments of significant thickness and their good state of preservation, it is more likely that they are of recent origin, or have been formed during the transgression.

We conclude that there is little morphological/sedimentological evidence for the existence of an ice-dammed lake in the Kara Sea. However it is still possible that this lake existed and that all its "whitenesses" has been erased by the subsequent transgression, as proposed by Polyak et al. (2002). Nevertheless, even a short lived lake should have been resulted in a catastrophic freshwater release during disintegration of the ice dam.

For further discussion we first estimate the volume of such a lake (Table 5.1; Figure, 5.1), assuming that the river run-off was only the half of the recent value during glacial times due to the glacial climatic conditions, we come to an annual run-off for the Ob and Yenisei rivers of 220 km$^3$ and 310 km$^3$, respectively (after Gordeev et al., 1996; Rachold et al., 2003), summing up to 530 km$^3$. It would take only some 60 years to fill up the estimated pro-glacial lake volume of 27,313 km$^3$; Lake Agassiz measured 9,500 km$^3$ (Clarke et al., 2004).

The water would have been rapidly banked up in front of the ice sheet, therefore it would have been very likely that drainage occurred during a sudden catastrophic event with typical channel base incision during flood release (Baker et al., 1993; Clarke et al., 2004).

Outburst floods, or islandic “Jökulhaups”, as described by (Björnsson, 2003), are documented by the presence of kilometre wide, deeply excavated channels (Josenhans

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Area within ice limit (km$^2$)</th>
<th>Volume (km$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>120</td>
<td>457, 123</td>
<td>11036.9</td>
</tr>
<tr>
<td>100</td>
<td>365, 149</td>
<td>3321</td>
</tr>
<tr>
<td>80</td>
<td>331, 939</td>
<td>2323.1</td>
</tr>
<tr>
<td>60</td>
<td>302, 900</td>
<td>473.4</td>
</tr>
<tr>
<td>50</td>
<td>295, 010</td>
<td>5920.3</td>
</tr>
<tr>
<td>40</td>
<td>176, 605</td>
<td>1414.5</td>
</tr>
<tr>
<td>20</td>
<td>141, 243</td>
<td>2824.9</td>
</tr>
</tbody>
</table>

27314.1
and Zevenhuizen, 1990). The outburst event of lake Agassiz was triggered by northward drainage of the lake (Clarke et al., 2004); catastrophic floodings have a profound impact on global climate as the example of lake Agassiz, that probably stopped the north Atlantic current during the 8.2 ka BP cooling event (Clark et al., 2001).

In the Siberian analogy, these water masses should have been at least recorded in the central Arctic Ocean by IRD peaks or isotopic excursions related to the massive freshwater signal, if not in a deterioration in global climate. Furthermore, Mangerud et al. (2004) report that modern modelling of thermohaline circulation (Stoecker and Wright, 1991) indicate that freshwater discharge of 0.1 Sv (1 x 10^6 m^3/s) influences the oceanic circulation. The meltwater flux of the lake Agassiz outburst is estimated to 4-5 Sv (Clarke et al., 2004) which is the same range as the LGM Kara See lake would have

---

### Figure 5.2

Relation of peak discharge to the product of dam height (m) and times lake volume (m^3) after Baker et al. (1993). Data for lake Komi after Mangerud et al. (2004); Lake Agassiz’ volume taken from Clarke et al. (2004).
reached utilising Baker’s et al. (1993) equations (Fig. 5.2).

However, data from the central Arctic Ocean show no traces of major freshwater signals (Norgard Petersen et al., 2004; Spielhagen et al., 2004). Norgard Petersen et al. (1998, 2003) report low O$^{18}$ values during LGM highest value in MIS2, IRD maxima before LGM 23-28 ka and only restricted ice proximal meltwater input. Central Arctic ocean shows very low IRD with a peak at 17 ka BP and few icebergs in the central Arctic (Darby et al., 2002).

Therefore, we conclude that the lake only could have endured a couple of hundred years - a time span not sufficient to be handed down in the geological record on our data basis.

**Future research perspective**

Based on the aforementioned discussion the following topics are of special interest. An exact survey of ice marginal channel levee complexes by means of geo-acoustics with a very high spatial resolution; combined with sedimentological studies emphasising on the discrimination of marine-fluvial environments: isotopic investigations, palynological and pollen analysis.

there a two areas of high interest:

1. examination of the continental margin north of the Kara Sea Shelf as proposed by Stein et al. in VERITAS in order to investigate sediment by-passing the shelf during sea level lowstand, accumulating in a lowstand wedge/fan

2. close the gap between the Vilkitzky straight and the inner Kara Sea.

In both areas, sediment budget determinations and depositional history reconstructions based on high-resolution acoustic stratigraphy would shed new light on the reconstruction of the LGM paleoenvironment. It would be important to close the gap in data coverage between the Vilkitzky straight (Weiel, 1998; Kleiber and Niessen, 2000) as well as to investigate the continental margin to the North of the Kara Sea Shelf, to trace sediments/fluxes during sea level lowstand and to evaluate sediment masses. These investigations could also lift the last secrets concerning the last glacial maximum and the controversy about a pro-glacial lake.

The employment of a Vibro corer in areas of little Holocene depositions, as well in glacially overprinted terrain would be another important step to improve the regional stratigraphic framework.

In the light of present global change it is of major interest of course to predict future trends. Having established and characterised most processes relevant for the recent
system, the frame has been established, especially for high-resolution studies in the future. The fluvial dominated Siberian part of the Arctic will be strongly influenced by global change. Labat et al. (2004) propose a change in the global hydrological cycle with 4% higher run off for every 1°C of warming. Furthermore, Manabe et al. (2004) propose that climate change will have 15% stronger effect in higher latitudes. Peterson et al. (2002) predict the average annual discharge of freshwater from the six largest Eurasian rivers to the Arctic to increase by nearly 10% during this century. The increase of discharge is coupled to the NAO (North Atlantic Oscillation) and increased surface temperatures (Peterson et al., 2002). The Intergovernmental Panel on Climate Change (IPCC, 2001) estimates a winter warming of more than 40% greater than the global mean, for high latitude regions by the end of this century.

High-resolution studies of the sedimentary history will play a key role in the prediction of these enormous changes. A proposal (DI 119/2-1) for funding has been submitted to the DFG in order to investigate the Holocene in high resolution (Deutsche Forschungs Gemeinschaft; “Rekonstruktion holozäner Umweltbedingungen in der KaraSee” RUKS).
Part D

Acknowledgements

Literature
Acknowledgements

First of all I want to thank Prof. Dieter Fütterer for his enthusiastic leadership in the “SIRRO” project, continuing after his retirement.

I would like to thank my supervisors Prof. Ruediger Stein, who initiated this project including this thesis and gave steady impulse to its forthcoming. He supported me whenever possible and enabled me several attendances on conferences. He always had an open ear for questions. For the scientific support I thank Frank Niessen and his intelligent witty advices. Despite his busy schedule he was able to meet with me and always had a good idea how to solve problems. Both gave me fair amount of freedom, while still stopping me from running into dead ends. I thank Prof. R. Henrich for his agreement on overtaking the second review of this thesis.

Working with my fellows at the Alfred Wegener Institute for Polar and Marine Research in Bremerhaven was both profit and pleasure. First of all I want to thank for their excellent technical help: Rita Fröhlking, Beate Hollmann, Walter Luttmer and Michael Seebeck. Among those I would especially thank is the fabulous “Arctic container” crew: Daniel Birgel, Claudia Didié, Kirsten Fahl, Christian Hass, Jens Hefter, Matthias Kraus, Kristoph Kierdorf, Uwe Langrock, Jens Matthiessen, Michael Pirrung, Christoph Schäfer, Frank Schoster, Tatjana Steinke, Petra Weller and Daniel Winkelmann. Dominik Weiel being no active member anymore is thanked for many fruitful discussion. Christoph Vogt, as former member is also thanked for discussions and advices. Michael Pirrung is thanked for the excellent introduction and sharing of knowledge of the measurement of physical properties and especially magnetic susceptibility. Christain Hass gave me valuable advices in the Silt analysis and insight into heavy metal of the 80’s. A very, very special thank you is dedicated to Jens Mathiessen, who was never tired to read, advice and motivate. Thanks to all crew members of Boris Petrov and cruise participants for their excellent cooperation during three expeditions.

I enjoyed cooperating with the german members of SIRRO in particular Ingo Fetzer, Johannes Simstich, Ingo Harms and Catalina Gebhardt.

I want to thank all my friends in and outside Bremerhaven for all the good times we have been wasted. Frank Schoster for being much more than my flatmate during the expeditions but also for friendship and sharing his music taste. The member of “Erosion” who kept me rocking even at bad times. Christian Müller for many hours not only at the Weser dyke, for good advices not only technically!!!!

Thanks to my sisters and parents for everything and support in any situation. Last but not least I am very grateful for Inken and her endurance, understanding, and support and to make me and my manuscripts digestible LUH.


Fielding, C.R. and Crane, R.C., 1987. An application of statistical modelling to the prediction of


Herdendorf, C.E., 1985. Inventory of the morphometric and limnologic characteristics of the large lakes of the world. Inventory of the morphometric and limnologic characteristics of the large lakes of the world, 17. State University technical bulletin, Ohio, 78 pp.


Lambeck, K., 1996. Limits on the areal extent of the Barents Sea ice sheet in late Weichselian time; Impact of glaciations on basin evolution; data and models from the Norwegian margin and adjacent areas. Elsevier, Amsterdam, Netherlands, pp. 41-51.


Lambeck, K., Yokoyama, Y. and Purcell, A., 2002. Into and out of the last glacial maximum; sea level change during oxygen isotope stages 3 and 2, Ice sheets and sea level of the last glacial maximum. Pergamon, Oxford, United Kingdom, pp. 343-360.


Posamentier, H.W., Jervey, M.T. and Vail, P.R., 1988. Eustatic controls on clastic deposition; I,
Literature

Conceptual framework, Sea level changes; an integrated approach. SEPM (Society for Sedimentary Geology), Tulsa, OK, United States, pp. 109-124.


St. Lawrence, the Mackenzie, the Yukon, the Arctic Alaskan, the Arctic Basin rivers in the Soviet Union and the Yenisei. In: E.T. Degens, S. Kempe and J. Rickey (Editors), Biogeochemistry of Major World Rivers. SCOPE, pp. 75-104.


