Alfred Wegener Institute for Polar and Marine Research (AWI) Bremerhaven

Palynologische Untersuchungen in der Karasee (Sibirische Arktis, Russland) zur Rekonstruktion der Paläo-Umweltbedingungen im Holozän

Holocene environmental history of the Kara Sea (Siberian Arctic, Russia) inferred from marine palynological records

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Dedicated to Katrin, Jonathan and Till

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2. Gutachter: Frau PD Dr. Karin A.F. Zonneveld

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Abstract

Two marine sediment cores and 13 surface sediment samples located along three N-S transects in the southern Kara Sea (Arctic Ocean, northern Siberia) were studied for their palynological content in order to reconstruct the Holocene paleoenvironmental evolution of the Kara Sea region. One sediment core (BP99-04) is located in the outer Yenisei estuary in the southwestern Kara Sea. The second core (BP00-38) is located in the outer Ob estuary. Both well-dated sediment cores show a continuous deposition with high sedimentation rates since ca. 9600 / 9400 cal. BP up to today and 600 years ago, respectively. The marine palynological analysis includes terrestrial palynomorphs (pollen and spores) as well as aquatic palynomorphs (organic-walled dinoflagellate cysts, chlorophycean algae, acritarchs, and organic benthic foraminifer linings).

The modern distribution of aquatic palynomorphs along the three transects revealed a distinct relationship to the respective sea-surface conditions. The pollen content of the nearshore surface sediment samples exhibited a higher proportion of non-arboreal pollen (such as herbaceous pollen) than the offshore surface samples.

The marine pollen records reflect the most favorable climatic conditions between ca. 9600 and 9000 cal. BP, which is related to the end of the Holocene thermal optimum (HTM) in the coastal area of the Kara Sea. The shrub tundra was displaced by the forest tundra with stronger contribution of tree birches and by spruce, which grew along the Yenisei river valley. As a consequence of both the rising sea and groundwater level due to permafrost degradation, floodplain and water rise mires grew along the lowlands. At ca. 7400, 5700 and 3800 cal. BP, the pollen spectra indicate a stepwise climate deterioration in the northernmost part of Siberia. At ca. 2000 cal. BP, the pollen assemblages reflect a displacement of the boreal forest by Arctic tundra communities and the establishment of modern conditions. The Ob pollen record reflects a more instable system during the last 1000 years that is related to human impact.

The aquatic palynomorph assemblages show, that the hydrographical evolution of the Kara Sea was strongly affected by the inundation history due to the post-glacial sea-level rise, by changes of river discharge and by the variable seasonal sea-ice cover. The occurrence of dinoflagellate cysts indicates the flooding of the Yenisei site at ca. 8900 cal. BP and of the Ob site at ca. 8500 cal. BP. Between ca. 8900 and ca. 7200 cal. BP, a marine thermal optimum was detected with a duration of about 2500 years. Fully marine/brackish water conditions and an enhanced stratification of the upper water column were reached since ca. 7400 to 7200 cal.

BP. Since ca. 3500 to 3300 cal. BP, conditions comparable to today with polar and cold surface water masses and extensive sea-ice formation was established.

Based on the time of flooding of both core locations and their modern water depth, we have calculated roughly a rate of sea-level rise for the southern Kara Sea that amounts between 14 to 28 mm/year. The eustatic sea-level rise was presumably slightly compensated by moderate isostatic uplift. The rebound rate was estimated about ca. 0,9 – 1,4 mm/yr at the Yenisei site and about ca. 1,1 – 1,5 mm/yr at the Ob site based on the comparison with the global sea-level curve according Fairbanks (1989).

The correlation of the local pollen stratigraphies (LPAZ) with the aquatic palynomorph stratigraphies (LAPAZ) and their respective paleoenvironmental implications reveals in principal synchronous trends, but also some differences between the onshore and offshore evolution. This is even made more clearly in the correlation of the LAPAZ with continental paleorecords in the adjacent coastal area. It reveals partly time-transgressive evolutions, which permits the tentative conclusion that the coastal climate and vegetation in relation to the hinterland have responded in different way to the marine transgression: While the coastal climate became less continental, the hinterland showed an amelioration, which occurred time-transgressive. Furthermore, it allows the estimation that the Holocene environmental evolution in the southern Kara Sea was influenced by local effects, which revealed in relation to the other circumarctic evolution some features on regional scale. However, large-scale trends such as the climate development in southern Siberia, the global atmospherical circulation patterns and the variations of the North Atlantic Current may have been mainly affected by environmental changes in the Kara Sea during the Holocene.

Kurzfassung

An zwei Sedimentkernen und weiteren 13 Oberflächen-Sedimentproben entlang eines N-S Transektes aus der südlichen Karasee (Arktischer Ozean, Nord-Sibirien) wurden marine palynologische Untersuchungen zur holozänen Umweltgeschichte der Karasee Region durchgeführt.

Der eine Sedimentkern (BP99-04) stammt vom äußeren Yenisei Ästuar in der südöstlichen Karasee während der zweite Kern (BP00-38) im äußeren Ob Ästuar in der südlichen Karasee genommen wurde. Beide gut datierten Sedimentkerne, die kontinuierliche Ablagerungsbedingungen mit hohen Sedimentationsraten aufweisen, ist die Zeit von 9600 cal BP bis heute (bzw. 9400 bis 600 cal. BP) aufgezeichnet. Die marine palynologische Analyse beinhaltet sowohl terrestrische Palynomorphen (Pollen und Sporen) als auch aquatische Palynomorphen (organisch-wandige Dinoflagellaten-Zysten, Grünalgen, Acritarchen und benthische Foraminiferen-Tapeten).

Die moderne Verbreitung der aquatischen Palynomorphen entlang der drei Transekte zeigt eine deutliche Beziehung zu den jeweils vorherrschenden Oberflächenwasser-Bedingungen. Der Pollen Inhalt der küstennahen Oberflächen-Sedimentproben zeigt einen höheren Anteil an Nicht-Baumpollen (z.B. Pollen von Kräutern) als die küstenfernen Oberflächen-Sedimentproben.

Die marinen Pollenprofile spiegeln die günstigsten klimatischen Bedingungen zwischen 9600 und 9000 cal. BP wider, die mit dem Ende des sogenannten „Holozänen Thermalen Optimum“ (HTM) in der Küstenregion der Karasee in Zusammenhang gebracht werden. Zu dieser Zeit wurde die Zwergstrauchtundra durch die Waldtundra verdrängt, die durch Baumbirken dominiert wurde und durch Fichten, die entlang des Yenisei Flusstales wuchsen.

Durch den Meeresspiegelanstieg und zum zweiten durch den Grundwasseranstieg als Folge der auftauenden Permafrostböden, bildeten sich im Bereich der Niederungen Überflutungs- und Versumpfungsmoore. Für ca. 7400, 5700 und 3800 cal. BP deuten die Pollenprofile für den nördlichsten Teil Sibiriens auf eine sich schrittweise vollziehende Klimaverschlechterung hin. Um 2000 cal. BP weisen die Pollenprofile auf ein Verdrängen des borealen Nadelwaldes durch arktische Tundrengesellschaften hin und auf die Etablierung des heutigen klimatischen Zustandes. Das Ob Pollenprofil spiegelt in den letzten 1000 Jahren ein instabileres System wider, das mit menschlichen Einflüssen („*human impact*“) erklärt wird.

Die aquatischen Palynomorphen zeigen, dass die Entwicklung der Oberflächenwassermassen in der Karasee in hohem Maße durch die Überflutungsgeschichte infolge des Meeresspiegelanstieges, durch Änderungen im Flusswasser-Eintrag und durch die Variabilität der Meer-Eisdecke beeinflusst wurde. Das Erscheinen der Dinoflagellaten-Zysten verrät den Zeitpunkt der Überflutung, die um ca. 8900 cal. BP am Yenisei Standort und um ca. 8500 cal. BP am Ob Standort stattfand. Für den Zeitraum von ca. 8900 bis 7200 cal. BP wurde ein marines thermales Optimum mit einer Dauer von ungefähr 2500 Jahren festgestellt. Vollständig marine Bedingungen bzw. Brackwasserbedingungen und eine erhöhte Schichtung der oberen Wassersäule stellten sich ab ca. 7400 bis 7200 cal. BP ein. Seit ca. 3500 bis 3300 cal. BP stellten sich mit den heutigen vergleichbare Umweltbedingungen ein, die gekennzeichnet sind durch polare und kalte Oberflächenwassermassen sowie durch eine erheblich erhöhte Meereisbildung.

Basierend auf den Überflutungs-Zeitpunkten der beiden Kern-Standorte und deren heutigen Wassertiefe wurde grob eine Meeresspiegel-Anstiegsrate für die südliche Karasee berechnet, die sich auf ungefähr 14 bis 28 mm/Jahr beläuft. Der eustatische Meeresspiegel-Anstieg wurde vermutlich durch moderate isostatische Landhebung schwach kompensiert. Die Ausgleichsrate wurde mit ungefähr 0,9 – 1,4 mm/Jahr für den Yenisei Standort und mit ungefähr. 1,1 – 1,5 mm/Jahr für den Ob Standort berechnet und basiert auf dem Vergleich mit der globalen Meeresspiegelkurve von Fairbanks (1989).

Die Korrelation der lokalen Pollen-Zonierungen (LPAZ = lokale Pollenzone) mit den lokalen aquatischen Palynomorphen-Zonierungen (LAPAZ = lokale aquatische Palynomorphenzone) und der jeweils abgeleiteten Umweltgeschichte offenbart grundsätzlich synchrone Trends, aber auch einige Unterschiede in der Entwicklung auf dem Kontinent und im Ozean. Dies wird noch deutlicher bei der Korrelation der LAPAZ mit den Paläoprofilen vom benachbarten Festland entlang der Karasee-Küste. Sie spiegelt eine zeitlich versetzte Entwicklung wider, die die vorläufige Schlussfolgerung erlaubt, dass das Küstenklima und die benachbarte Vegetation im Vergleich zum Hinterland in unterschiedlicher Art und Weise auf die marine Transgression reagiert hat: Während der kontinentale Charakter des Küstenklimas abnahm, zeigte das Hinterland eine Klimaverbesserung, die zeitversetzt stattfand. Weiterhin erlaubt es die Einschätzung, dass die holozäne Umweltgeschichte in der südlichen Karasee durch lokale Effekte verstärkt wurde, die sich im Vergleich zu den Entwicklung im übrigen zirkumarktischen Raum auf regionalem Niveau vollzog. Allerdings dürften hauptsächlich großräumige Trends wie zum Beispiel die Klima-Entwicklung in Süd-Sibirien, die globalen atmosphärischen Zirkulationsmuster, sowie Variationen des Nordatlantik-Stroms („NAC“) die Änderungen der Umweltbedingungen in der Karasee im Holozän ausgelöst haben.

1. Introduction

1.1 Scientific background

The Arctic Ocean is a semi-enclosed basin, which is surrounded by vast continental shelf seas (> 50 %, Jakobsson et al., 2003) and land areas that restrict the exchange of water masses with the Pacific and Atlantic Oceans to the shallow Bering Strait (ca. 50 m water depth), and the deep (>2500m) Fram Strait. The inflow of the North Atlantic Current (NAC) transporting relatively warm and saline waters to the north is counteracted by the outflow of cold, ice covered polar waters. A unique feature of the Arctic Ocean is the huge freshwater (ca. 3300 km^3, Aagard and Carmack, 1989) and sediment supply (contemporary ca. 250 Mt/yr, Holmes et al., 2002) by numerous rivers discharging into the shelf seas (Fig. 1-1). The pronounced freshwater supply strongly influences the hydrographic structure of the Arctic Ocean, and is exported to the south through the Fram Strait and the Canadian Arctic Archipelago (Aagard and Carmack, 1989; Prinsenberg and Hamilton, 2005). As a consequence, changes in river discharge may perturbate the thermohaline circulation (THC) and influence therefore the global atmospheric and oceanic circulation (e.g. Rahmstorf 1995; Broecker et al., 1997; Serreze et al., 2003; Rennermalm et al., 2006). Recently, several empirical studies and modeling experiments led to the conclusion that the North Atlantic is freshened by increasing river discharge in conjunction with global change (e.g. Serreze et al., 2000; Peterson et al., 2002; Wu et al., 2005). A decrease of the extent and the thickness of the Arctic pack ice has also been observed over the last decades (e.g. Vinnikov et al., 1999; Serreze et al., 2003), which is further attributed to global warming (e.g. Callaghan et al., 2004; ACIA, 2005).

The observed freshening and melting of sea ice may have further effects, of which sea-level rise (e.g. IPCC, 2007) and permafrost degradation (e.g. Frey et al., 2007) are only two of the most prominent examples. In the past, the areal extent of the Arctic Ocean was strongly influenced by sea level changes due to exposure and subsequent flooding of the vast shallow circumpolar continental shelf seas. During low sea level stands in glacial periods, in particular the Eurasian margin, which encompasses an area of 4628 x 10^3 km^2 (Jakobsson et al., 2002, redefined limits), was largely subaerially exposed. This led to substantial environmental changes during the glacial-interglacial cycle from periglacial to neritic marine conditions.

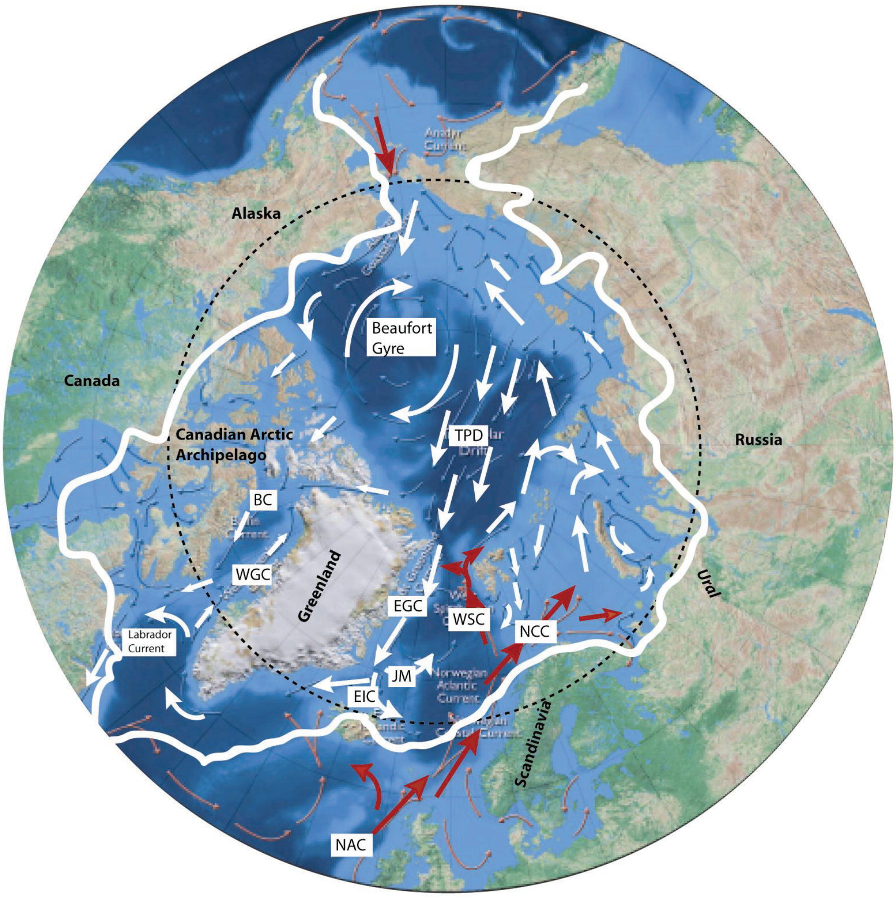


Fig. 1-1: Polar view of the Arctic Ocean (modified after AMAP, 1998) showing the main surface circulation patterns (TPD: Transpolar Drift; NAC: Norwegian Atlantic Current; NCC: North Cup Current; WSC: West Spitsbergen Current). The black arrows indicate warm, the white cold surface water masses, respectively. The black dashed line marks the Arctic Circle (66°N). The white line marks the 10° July isotherm approximately, which defines the border of the Arctic (AMAP, 1998).).

It is therefore of utmost importance to improve our knowledge of the environmental variability during our present interglacial in order to better understand the natural variability as well as underlying processes and mechanisms. In spite of a growing number of Late Quaternary marine as well as terrestrial paleorecords collected and analysed during the last decades by means of various methods, this youngest period of the geological history is comparatively little known in the northern high latitudes (see also Mayewski et al., 2004; de Vernal et al., 2005). This could be related to principal methodological problems of Quaternary research in the Arctic realm: suitable paleoenvironmental archives are scarce due to the low sediment accumulation rates on land as well as the marine seafloor, discontinuous sedimentary sequences (e.g. due to solifluction) onshore and poor preservation of marine calcareous and siliceous microfossils such as planktic foraminifers, coccolithophorids, diatoms and radiolarians offshore (e.g. Mudie and Shorts, 1985; Mudie et al., 2001).

1.2 Aims of this thesis

The problems in Arctic Quaternary research outlined above led to the idea to investigate sediment cores from the inner shelf seas, where comparatively high bioproductivity is given (see e.g. Makarevich et al., 2003) and where high sedimentation rates permit a better temporal resolution of the paleoenvironmental development. Therefore, two sediment cores were selected from the estuaries of the rivers Ob and Yenisei in the southern Kara Sea (Stein and Stepanets, 2000, 2001) that were collected within the German-Russian research project Siberian River Run-Off (SIRRO) (Stein et al., 2003, 2004, see also chapter 1-7).

The major objectives of this thesis are

1) to test the applicability of nearshore pollen records as tracers for climate, landscape and vegetation development in the coastal area of the southern Kara Sea and the adjacent hinterland during Holocene times,
2) to reconstruct changes in sea-surface conditions based on aquatic palynomorph assemblages (dinoflagellate cysts, chlorophycean algae, acritarchs), and
3) to compare pollen and aquatic palynomorph assemblages in order to characterize the timing of events in the marine and terrestrial realm (land-sea correlation, to find out possible simultaneous occurrences or/and discrepancies such as time-trangressive environmental changes).

This thesis is submitted in a "cumulative" form containing three independent manuscripts. In order to introduce into the topic of marine palynological research, notes on the possibilities and limitations of marine palynology as a tool for paleoenvironmental reconstruction are outlined and the research history is shortly described. Additionally, an overview of the Late Quaternary climate and vegetation research history with respect to western Siberia is given.

1.3 Marine palynology as a tool for paleoenvironmental reconstruction: An introduction to the research history

In this study, marine palynology considers both terrestrial (pollen and spores) and aquatic palynomorph (organic-walled dinoflagellate cysts (=dinocysts), acritarchs, chlorophycean algae) assemblages that were deposited in marine sediments.

Originally, marine palynology only encompassed pollen and spores from terrestrial plants and was mainly applied in the oil industry as a biostratigraphic tool in the middle of the 20th century (Hooghiemstra et al., 2006). A first overview of possibilities, limitations and problems of marine palynology was given by Groot and Groot (1966) who considered only marine pollen assemblages. Relatively early studies concentrated on pollen in shallow marine environments (Muller, 1959, Florer, 1973) and the transport and deposition of reworked pollen and spores in marine sediments (Stanley, 1966). However, the earlier studies on marine pollen assemblages were performed mainly on deep-sea records in the view of transport and deposition processes (e.g. Heusser and Balsam, 1977; Heusser and Shackleton, 1979).

At the beginning of the eighties of the last century, and in larger numbers since the nineties, marine pollen analysis focussed mainly on distribution patterns and the transport and deposition processes in marine environments, in particular offshore eastern North America (e.g. Mudie, 1982; Brush and deFries, 1981; Heusser, 1983; Short et al., 1989; Mudie and McCarthy, 1994; Traverse, 1994; Brush and Brush, 1994; Chmura and Eisma, 1995; Chmura et al., 1999). Also, the role of long-distance transport of pollen in northern high-latitudes was increasingly recognized (e.g. Kalugina et al., 1981; van der Knaap, 1987; Johansen and Hafsten, 1988).

Since the late eighties, one of the major areas for studying marine pollen was the southern Atlantic Ocean off west equatorial Africa (e.g. Hooghiemstra et al., 1986; Jahns et al., 1998; Dupont et al., 1998; Shi et al., 2001; Hooghiemstra et al., 2006; Dupont et al., 2007). Further studies focused on different parts of the Mediterranean Sea, where Zonneveld (1996) conducted one of the first direct land-sea correlations by using pollen and dinoflagellate cysts from record in the Adriatic Sea. Previously Grüger (1975) and Rossignol-Strick and Pastoret (1971) investigated in a very detailed manner marine pollen records from the Adriatic Sea. Marine pollen analysis were conducted in the Atlantic Ocean off the Iberian Peninsula (e.g. Sánchez Goñi et al., 1999; Turon et al., 2003; Desprat et al., 2003; Naughton et al., 2007), the south China Sea (e.g. Sun et al., 2003), southeastern Indonesian waters and offshore Western Australia (e.g. van der Kaars and de Deckker, 2003), southeastern Atlantic Ocean (e.g. Behling et al., 2002) and the Baltic Sea (e.g. Yu et al., 2005). A valuable review of previous studies dealing with marine palynology was given by Mudie and McCarthy (2006).

With respect to the Arctic realm, marine pollen records are relatively scarce. Apart from the studies conducted in the western part of the Arctic mainly by Canadian palynologists mentioned above (see also Levac and de Vernal, 1997; Levac, 2001) only few studies were performed along the Eurasian continental margin. In the Laptev Sea, Naidina and Bauch (1999, 2001) and Naidina (2006) examined the distribution of pollen in surface sediments and Holocene pollen records. In the Kara Sea, a singular marine record considered only selected pollen types in a non-radiocarbon dated sediment core (Kulikov and Khitrova, 1982). Otherwise, pollen was only presented as total concentration relative to other aquatic palynomorphs in bottom samples (Matthiessen, 1999; Matthiessen et al., 2000; Matthiessen and Kraus, 2001). Moreover, marine pollen records are absent in the Barents Sea except for an early palynological study of bottoms sediments in the White Sea Malyasova (1981) and a total pollen concentration curve in a Holocene record of the eastern Barents Sea (Voronina et al., 2001).

Marine palynology in the broader sense including dinoflagellate cysts

The application of dinoflagellate cysts as a proxy for Quaternary paleoenvironmental reconstructions is quite young, as the method was developed since the middle of the 20th century within the scope of hydrocarbon exploration (Matthiessen et al., 2005). A review of the history of academic research was described by Dale (1983) and Evitt (1985). More recently, there are several major reviews of dinoflagellate cysts as a tool for paleoenvironmental reconstructions and their distribution in recent sediments (e.g. Mudie and Harland, 1996; Mudie et al., 2001; de Vernal et al., 2001; Marret and Zonneveld, 2003; Pross et al., 2004; Matthiessen et al., 2005).

Probably, the earliest studies on dinoflagellate cysts in modern sediments and their environmental interpretation were performed by Erdtmann (1954) and Evitt (1961). Systematic work started in the sixties of the last century and advanced during the seventies focussing on taxonomy and terminology (e.g. Evitt et al., 1977; Williams et al., 1978), the distribution of modern dinoflagellate cysts (e.g. Wall, 1965; Wall and Dale, 1966, 1967, 1968; Reid, 1974), the relationship to their theca (e.g. Evitt and Davidson, 1964), and the dinoflagellate cyst assemblages in Quaternary sediments (e.g. Norris and McAndrews, 1970; Wall, 1971; Harland, 1973; Wall and Dale, 1974; Reid and Harland, 1977). First studies on the biology, cyst formation, cyst-theca relationships, life cycles and preservation were done by Dale (1976), Taylor (1980, 1987), Anderson et al. (1984, 1985) and Matsuoka (1988). In the last three decades, considerable progress has been made in the field of biology, ecology, taxonomy & terminology, and morphology of Quaternary dinocysts (Matthiessen et al., 2005 and references therein). Also, numerous studies on dinocysts' recent distribution were made in the low, middle and high latitudes of the World Ocean (Marret and Zonneveld, 2003; Rochon and Marret, 2004). Moreover, a large number of Quaternary dinocyst records were investigated and revealed an improved knowledge with respect to changes in sea-surface conditions (e.g. de Vernal and Hillaire-Marcel, 2006; Esper and Zonneveld, in press).

The North Atlantic Ocean, the Arctic Ocean, and adjacent basins are mostly considered together (e.g. de Vernal et al., 2001; de Vernal and Hillaire-Marcel, 2006) because they are linked through water mass exchange. Except for several earlier works (e.g. Harland, 1973, 1980), numerous studies, on modern dinocyst distribution in particular, were performed since the eighties of the last century (e.g. Miller et al., 1982; Scott et al., 1984; Mudie and Short, 1985; Aksu and Mudie, 1985; Mudie, 1992; de Vernal, 1994; Harland, 1994; Williams et al., 1995; Matthiessen, 1995; Levac and de Vernal, 1997; de Vernal et al., 2000; Levac et al., 2001; Matthiessen et al., 2001; Marret et al., 2004; de Vernal and Hillaire-Marcel, 2006, Fig. 1-2).

With respect to the shallow circumarctic shelf seas there are several studies on recent dinocyst distribution (Harland et al., 1980, 1982; Mudie 1992; Radi et al., 2001; Mudie and Rochon, 2001; Kunz-Pirrung, 1998, 1999, 2001a; Head et al., 2001; Voronina et al., 2001) except for the East Siberian and Kara Seas (Fig. 1-2). In contrast, Quaternary dinocyst records are relatively scarce: In the western part of the Arctic shelf seas, Harland et al. (1980) examined sub-recent dinocyst assemblages and de Vernal et al. (2005) reconstructed the variability of sea-ice cover during the Holocene in the Chukchi Sea. In the eastern part, there are only a few studies. In the Laptev Sea, a first dinocysts record was published by Kunz-Pirrung (2001b). Polyakova et al. (2005), Klyutvitkina and Bauch (2006) and Polyakova et al. (2006) reconstructed Holocene variability of sea-surface conditions on in total three sediment cores from the inner and outer shelf area. Finally, Voronina et al. (2001) performed dinocyst analysis on two sediment cores from the eastern Barents Sea. Up to now, no Quaternary dinocyst records from the Kara Sea do exist except for one record retrieved from the Eurasian continental margin in the northernmost part of the Kara Sea (Matthiessen et al., 2001).

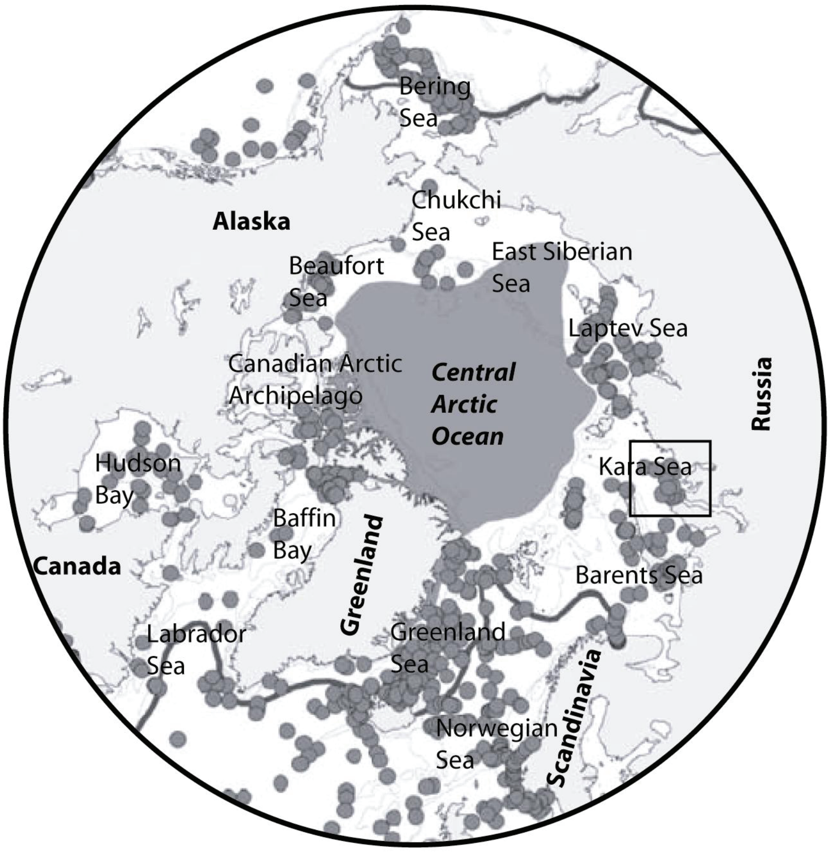


Fig. 1-2: Map of Arctic realm with topographical names mentioned in the text (modified according to de Vernal et al., 2005). Circles constitute approximately the 50°N latitude. Grey dyed area marks the minimum and the dark line the maximum sea ice cover extent respectively. The points indicate locations of dinocyst surface sediment samples used to develop the reference dinocyst database (de Vernal et al., 2005). The area under the perennial pack ice of the Arctic Ocean revealed barren dinocyst assemblages and samples are not included in the database (de Vernal et al., 2005). The rectangle shows the study area.

1.4 What are dinoflagellate cysts?

Dinoflagellate cysts are formed during the sexual reproduction of dinoflagellates, which produce during their life cycle so-called hypnozygotes (resting cysts) for a resting phase of variable duration (e.g. Rochon et al., 1999). These cysts consist of a protective organic wall containing sporopollenin-like material or dinosporin (Fensome et al., 1993 cited in Rochon et al., 1999). The formation of cysts is related to three possible functions: protection, propagation and dispersion (e.g. Dale, 1983).

In order to reveal the respective biological affinity incubation experiments were carried out (e.g. Wall and Dale, 1966, 1967, 1968). Approximately 15 % of the living dinoflagellates are known to produce such organic and calcareous-walled cysts (e.g. Head, 1996). However, cyst-theca relationships are still poorly known because the number of species that produce resting cysts (> 250 species) exceeds by far the number of species that have been observed to reproduce sexually (< 50 species) (Matthiessen et al., 2005 and therein cited references). It should be mentioned that not only organic-walled cysts but also calcareous and siliceous dinoflagellate cysts are produced (e.g. Zonneveld et al., 1999).

Dinoflagellates are eukaryotic, primarily unicellular organisms that inhabit freshwater as well as marine environments. Characteristic morphological features are the two flagella, which the majority of dinoflagellates have. This feature permits that dinoflagellates have a good motility allowing limited migrations through the water column. A second important morphological feature is that the cell wall is usually divided into cellulose plates within so-called amphiesmal vesicles, known as a theca. They have various nutritional strategies containing both autotrophic and heterotrophic as well as mixotrophic species. Together with diatoms and coccolithophores, dinoflagellates represent a major part of the eukaryotic primary production in the upper pelagic environment (e.g. Taylor, 1987).

Dinoflagellates are cosmopolitans occurring from low to high latitudes but the majority of species are found in tropical to temperate waters (Taylor, 1987) and species diversity is much higher in tropical than in polar waters. Related to the study area, approximately 90 species were found in the Kara Sea (Druzhkov and Makarevich, 1999).

The taxonomy of Quaternary dinocysts differs from the corresponding motile dinoflagellate, which is sometimes problematic and confusing. Historical reasons are mainly responsible for this dual taxonomic system. However, the motile name is now often used when new cyst-theca relationships are detected (Matthiessen et al., 2005). In some cases, the relationship to the adequate motile cell is not known, however molecular genetic studies might be able to fill this gap of knowledge (Matthiessen et al., 2005 and therein cited references). Usually, paleontologists assign the dinoflagellates and their cysts to the algae (Division Dinoflagellata), but in fact, systematical classification might be more difficult because biologists more often assign (heterotrophic) dinoflagellates to the group of protists because they are not primary producers in the strict sense.

Generally, dinocysts are used in marine paleoceanographical studies. However, it should be mentioned that it is known for a long time that dinocysts can also be applied as a tool in paleolimnological research, which has recently been rediscovered (e.g. Tardio et al., 2006).

1.5 Notes to transport and depositional processes influencing dinoflagellate cysts and pollen distribution in marine sediments

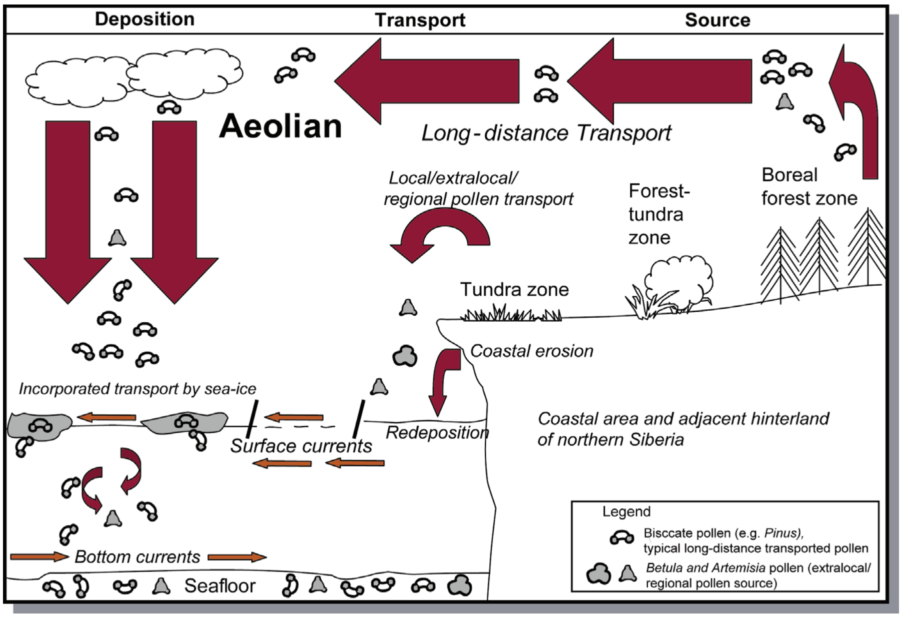
Sedimentation of dinoflagellates and their preservation as dinocysts

In general, several studies comparing dinoflagellate cells in water column with sediment data concluded that living and fossil communities differ considerable (Matthiessen et al., 2005 and references therein). Thus, the dinoflagellate cyst assemblages are selectively incorporated into the sediment because only a minority of dinoflagellates form cysts (e.g. Matthiessen et al., 2005). Moreover, the descent of cells and cysts through the water column is "dangerous" because of various consumers and due to physical degradation (e.g. currents, resuspension, redeposition, bioturbation) as well. In the sediment, the preservation of cysts depends on their walls' respective chemical composition (e.g. Dale, 1976). Zonneveld et al., (1997) grouped the dinocysts into four categories with respect to their sensitivity to oxygen availability in bottoms sediments: Accordingly, cysts formed by *Protoperidinium* species (*Islandinium minutum*, *I.? cezare*, *Echinidinium karaense*, *Brigantedinium* cyst taxa) are extremely sensitive to degradation. The dinocysts *Operculodinium centrocarpum, Impagedinium* spp. and *Spiniferites* cyst taxa are moderately sensitive. The dinocyst taxa *Impagidinium paradoxum* and *Nematosphaeropsis labyrinthus* are moderately resistant and finally following dinocysts are grouped to resistant species but are of minor importance for the study area: *Impagidinium aculeatum, I. patulum, Lingulodinium machaerophorum, Operculodinium israelianum and Polyshpaeridium zoharyi*.

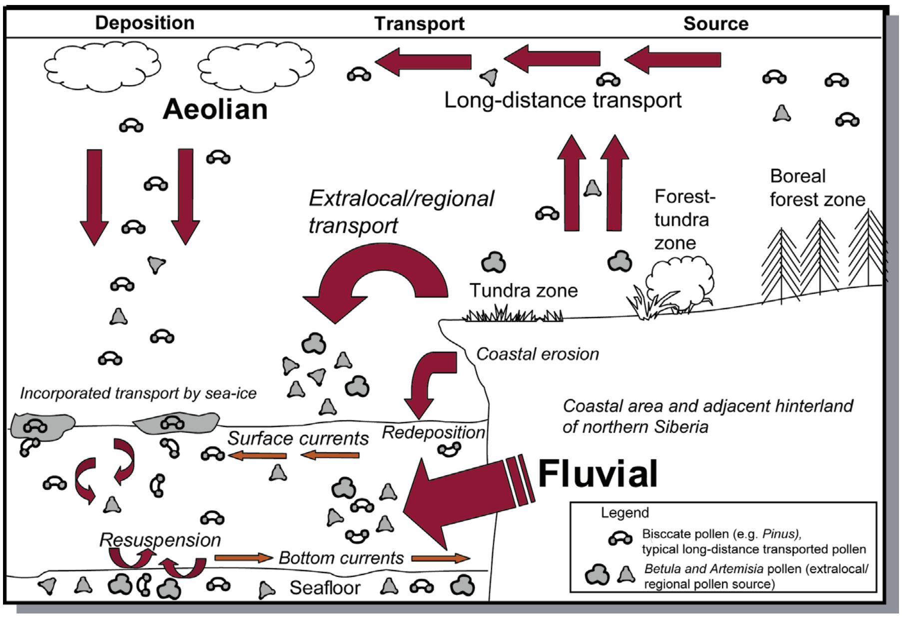
Pollen transport and deposition processes in marine(offshore) environment

The assemblages of marine pollen and their interpretation differ considerable from pollen records on the land because of the specific transport and deposition processes. Pollen grains are part of the suspended matter, which underlies similar transport and sedimentation processes. Fluvial pollen transport, selective pollen enrichment and degradation due to physical processes (e.g. currents, resuspension, redeposition, tidal flat, and ice-rafted transport) affect the pollen signal. However, problems of over representation and under representation, respectively, due to different pollen production of the plant producer and their preservation effects in a similar way both pollen assemblages on land and offshore. For example, north Siberian pollen assemblages usually show an under representation of *Larix* pollen in contrast to an over representation of *Pinus* pollen.

Several investigations were performed to study the pollen transport and deposition processes in different aquatic but not lacustrine environments in order to find out how far marine pollen records reflect the vegetation patterns on the adjacent land (e.g. offshore records more or less far away from the coast, nearshore records in the area of estuaries and deltas, records retrieved in river channels; e.g. Heusser and Balsam, 1977; Brush and de Fries, 1981; Mudie 1982; Mudie and Short, 1985; Chmura and Eisma, 1995; Chmura et al., 1999; Brush and Brush, 1994; Traverse, 1994; Mudie and McCarthy, 1994; Smirnov et al., 1996; Brown et al., 2007, for a review see Mudie and McCarthy, 2006). The majority of researchers chose locations enclosed by adjacent land-areas. In contrast, the pollen analysis in this study is performed on fine-grained terrigenous sediments retrieved in the shallow (24-32 m water depth) estuaries in front of the river mouths Ob and Yenisei in the southern Kara Sea shelf not far away from the present coastline (ca. 50-100 km, see also chapter 2, Fig. 2-1). With respect to the circumarctic realm, the works of Mudie (1982), Heusser (1983), Mudie and McCarthy (1995) are rather comparable to our locations since they were performed on samples from the submerged continental margin lying between the coastline and the shelf break (e.g Mudie, 1982). The overall implication is that marine pollen records show a relationship to the distance from the plant producers and to the respective vegetation biomes. Pollen and spore concentrations decrease with growing distance to the land (Mudie 1982; Heusser, 1983). Therefore, offshore pollen records reflect rather large-scale vegetational and climate as well as general paleoceanographic changes (e.g. changes in sediment flux). In contrast, nearshore pollen records containing a higher proportion of non-arboreal pollen could more or less accurately reflect changes on a regional scale, According to Mudie (1982), much of the fluvial pollen load from large rivers is deposited near the river mouths. The latter case can be assigned to the pollen records from the southern Kara Sea shelf. These nearshore, estuarine sediment cores are influenced strongly by fluvial pollen transport (Fig. 1-3 A and B).



A)



B)

Fig. 1-3 A and B: Schematized pollen transport and deposition patterns in the Arctic. A) Offshore situation (larger distance to the coast and to the pollen producers). Long-distance transported pollen (in particular bisaccate pollen) is over represented due to aeolian transport. B) The proportion of fluvial and extralocal/regional pollen transport and deposition increase proximal to the decrease of the distance to the coast / pollen producers. Non-arboreal pollen (also of herbaceous plants) and readily degraded pollen increases.

1.6 History of Late Quaternary climate and vegetation research with respect to western Siberia

In the past as well as at present, research on the Late Quaternary climate evolution is connected to the research on vegetation history on land. Besides other paleo proxies (in particular plant macrofossils used for macrofossils and dendrochronological analysis on wood remains), mainly palynological investigations performed on lakes sediments, peat-sections and other terrestrial deposits, are the most fruitful sources of data on Late Quaternary climate change (e.g. Faegri and Iversen, 1975; Khotinskiy, 1984).

During the cold war, Siberian Quaternary science was mainly restricted to the former Soviet Union. Russian scientific work only rarely reached the non-Russian world. Exceptions are the articles from Ilich Nejstadt (Neustadt), who worked in Moscow but published partly in German (e.g. Neustadt, 1959). His most famous work is the "History of forests and paleogeography of the USSR in the Holocene" (Neustadt, 1957). The German works of Burkhard Frenzel on the vegetation and climate history of northern Eurasia (Frenzel, 1960, 1968), represent a milestone. Further important sources were the early studies of Vladimir Grichuk, which were partly published in English (e.g. Grichuk, 1961). Vladimir Grichuk researched the floral history and climate evolution of the Russian plain in the Quaternary for more than a half century.

During the seventies of the last century, Quaternary research advanced and several important studies were still published in Russian language, of which only the most cited are mentioned here such as Velichko (1973), Kind (1974) and Khotinskiy (1977). During this time, most paleoecological studies were conducted based on bio- and climatostratigraphical correlation as radiocarbon dating was still an exception. Klimanov (1976) developed a technique for quantitative reconstruction of climate from pollen records, the so-called "information statitistical method" (IS-method, see also Klimanov, 1984). However, this method was rather suitable for forested areas farther south and only partly applicable in the northernmost part (see also Andreev et al., 2003).

Since the eighties, a growing number of publications appeared but still mainly in Russian and rarely on the northern part of western Russia such as Nikol'skaya (1982), Ukraintseva (1988) and Andreev et al. (1989). Mentionable is also the first overview study of the "recent pollen spectra and zonal vegetation in the western USSR" from Peterson (1983), who generated isopoll maps in order to illustrate the pollen-vegetation relationships in the

Soviet Union west of 100° E . But the most sustainable work was the first English synopsis on "Late Quaternary environments of the Soviet Union" edited by Velichko (1984). Therein, the overview work about Holocene climate and vegetation history by Khotinskiy (1984) was published. The presented classification and nomenclature of the Quaternary stratigraphy as well as the paleoenvironmental characterizations revealed some differences in comparison to the system according to Mangerud et al. (1974) (see also Fig. 1-4). Also, since the eighties, model simulations of Holocene paleoclimate and mechanisms (e.g. vegetation-atmosphere-ocean interactions) became more important including also Siberian data from paleorecords (COHMAP, 1988).

Since the nineties, a new era started, mainly as consequence of the perestroika, and several overview articles were published, and increasingly in English. Peterson (1993) provided a review about the "vegetational and climatic history about the western former Soviet Union". Based on a new data collection and until then published records, Velichko et al. (1997) summarized the climate and vegetation dynamics in the tundra and forest zone during the Late Glacial and Holocene in an article, which was published in the special issue "Quaternary of northern Eurasia: Late Pleistocene and Holocene landscapes, stratigraphie and environments" in the journal "Quaternary International". In this special issue, also the first high Arctic pollen record from the Sverdrup Island in the Kara Sea was published (Andreev et al., 1997). One of the major results in regard to the Kara Sea realm was that a climatic optimum occurred in the coastal and islands area during the early Holocene (Fig. 1-4). This outcome was supported by several other studies with respect to the Kara Sea region (Serebryanny et al., 1998; Serebryanny and Malyasova, 1998; Andreev et al., 1998). A further important work was published by Kremenetski et al. (1998), who reconstructed the north Eurasian Holocene tree-line history based on all until then available radiocarbon-dated macrofossils together with new data. Farther south at the southern border of the WSL, Blyakharchuk and Sulerzhitsky (1999) provided a detailed view of the Holocene vegetational and climatic changes in the forest zone of Western Siberia.

Increasingly, radiocarbon-dated paleorecords were collected, which considerably promoted the accuracy of the chronostratigraphical fundament. Also, combined analysis of plant macrofossils and pollen appeared, which were until then relatively rare (Koshkarova, 1995; Clayden et al., 1996; Peteet et al., 1998). They outcropped large differences between pollen deposition pattern and zonal occurrence of pollen producers such as larch trees (see also Kienast et al., 2001; Birks and Birks, 2000).

During the nineties, statistical methods were applied: Tarasov et al. (1998) assigned pollen spectra of the former Soviet Union to biomes for paleovegetation at 6000 years ago based on the plant-functional-type method (pft-method) (Prentice et al., 1996). Model simulation revealed potential vegetation maps and showed the potential role of vegetation feedbacks in the climate sensitivity of high-latitudes (TEMPO, 1996). A bioclimatic vegetation model was used to reconstruct the paleoclimate of Siberia (Monserud et al., 1998).

With the beginning of the 21st century, paleorecords were increasingly used for quantitative reconstructions (e.g. Andreev and Klimanov, 2000). Our knowledge about climate and vegetation evolution of northern Siberia improved particularly due to the lake records from the Taymyr Peninsula, which revealed continuous long-term sequences (Andreev et al., 2002, 2003 including modified pollen analysis previously published by Hahne and Melles, 1999, 2004). A milestone was the since then widely cited article from MacDonald et al. (2000), who studied the Holocene treeline history and climate change across northern Eurasia based on numerous new collected macrofossils. The comparison of terrestrial, ice core, and marine paleodata with simulations based on the "general circulation model" (GCM) captured an early Holocene warming across Eurasian Arctic (CAPE Project members, 2001). With regard to the Late glacial-Holocene transition, Velichko et al. (2002) reviewed the available paleorecords including an updated data set in the range of East Europe and Siberia. As a relatively young tool, dendrochronological reconstructions were conducted in order to calculate temperature variability during the Holocene (e.g. Hantemirov and Shiyatov, 2002).

Recently, within the scope of global change problematic, the WSL, which is the world's largest high-latitude wetland accounting for over 900,000 km^2 of peatland (Krementski et al., 2003), is studied more intensively because it is considered a major player as producer and sink in the atmospheric CH_4 and CO_2 fluctuations, respectively (e.g. Smith et al., 2004; MacDonald et al., 2006).

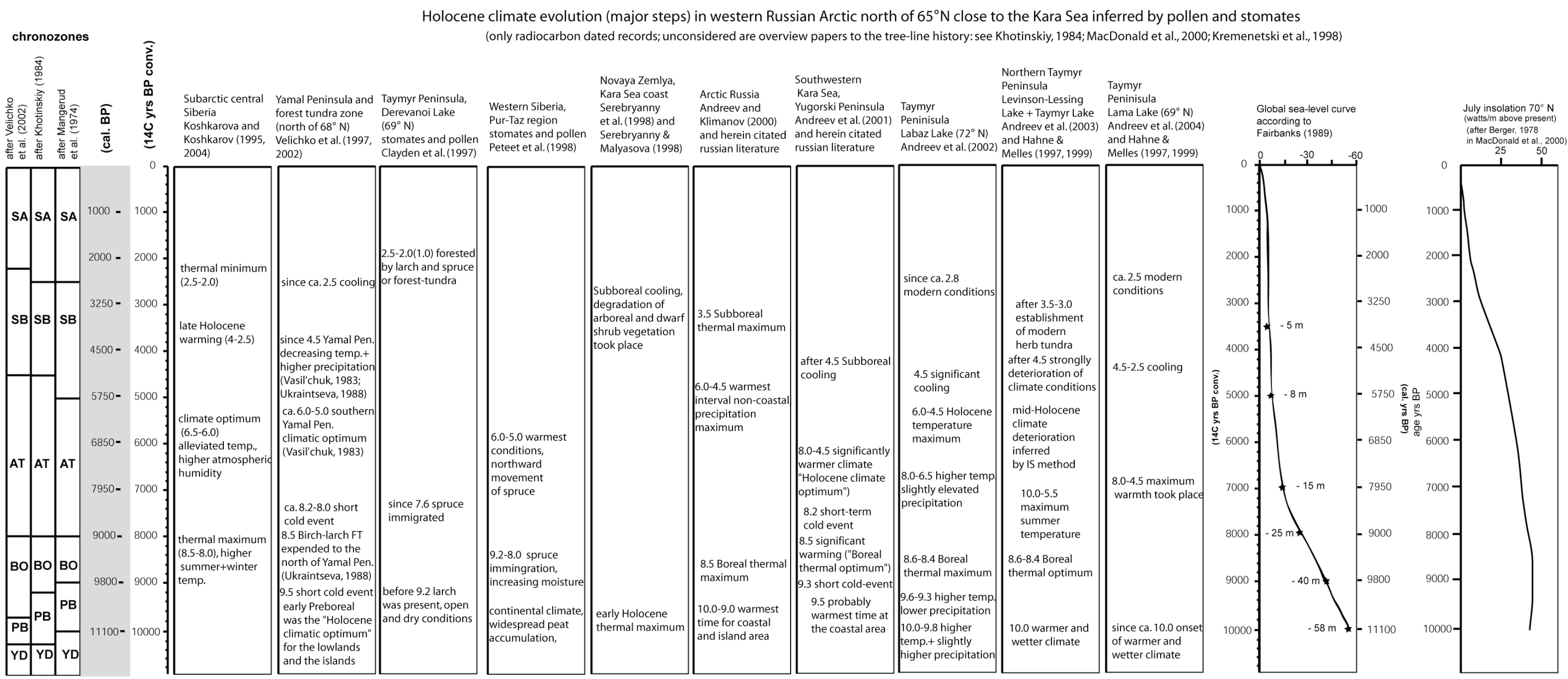


Fig. 1.4: Overview to the literature refer to the Holocene climate evolution (major steps) in the western Russian Arctic north of 65°N close to the Kara Sea inferred by pollen and stomata analysis (only radiocarbon dated records are considered, publications to the tree-line history are not included). On the left site the different chronostratigraphies are shown and the age in calibrated dates as well as the conventional radiocarbon dates.

Also west, on the top of the Ural Mountains and on the northeast European Russian/European Arctic Russia close to the Barents- and Kara Sea, paleoecological research was intensified in the last ten years mainly within the scope of Scandinavian-Russian research projects (e.g. Kaakinen and Eronen, 2000; Oksanen et al., 2001; Kultti et al., 2003; Sarmaja-Korjonen et al., 2003; Paus et al., 2003; Väliranta et al., 2003; Andreev et al., 2005; Jankovská t al. 2006; Väliranta et al., 2006; Wohlfarth et al., 2007). However, these studies can only partly be directly connected to our study area due to the biogeographical difference between the area west and east of the Ural Mountains, however, the climatic evolution in the hinterland across the coastline might be comparable.

1.7 Outline of the (paleo-) oceanographic research history of the Kara Sea

The development in the marine province considerable lacks studies on land, except for the most earliest efforts in the exploration and map presentation of the Eurasian Arctic shelf (e.g. Nansen, 1902; Seibold, 2001). This is related to the nuclear weapon tests on and near Novaya Zemlya during the middle of the last century bearing in mind the cold war and the dumping of radioactive waste by the former Soviet Union. A second reason might be the discovery of large gas and oil resources in the southern Kara Sea that lead to restrictions in studying Late Quaternary processes by means of geophysical properties. However, few earlier data about water mass characteristics in the Kara Sea were obtained (e.g. Johnson and Milligan, 1967; Hanzlick and Aagaard, 1980). A first English summary on Russian scientific studies about surface circulation patterns and Kara Sea hydrography was given by Palvlov et al. (1993).

Since the nineties, a new era began for the oceanographic research due to “perestroika”. Primary, contaminants and their pathways were studied within the range of first international projects (e.g. Joint Russian-Norwegian Expert Group, 1996). As a result of this and other endeavors to solve the increasing environmental problems, recent circulation and hydrographical patterns were studied in particular (e.g. Pavlov and Pfirman, 1995; Johnson et al., 1997; Harms and Karcher, 1999; Harms et al., 2000; McClimans et al., 2000).

During 1984 to 1993, several ship-based expeditions were undertaken focussing mainly on the exploration of the geotechnical conditions in the Kara Sea (Polyak et al., 2000 and references therein; Lisitzin and Vinogradov, 1995). Based on these expeditions, several publications appeared on the Late Quaternary evolution of the Kara Sea (e.g. Levitan et al., 1995; Lisitzin, 1995; Lisitzin et al., 1995; Polyak et al., 2000, 2002, 2003; Kuptsov and Lisitzin, 2003; Lisitzin and Kuptsov, 2003). From the beginning, the river discharge and its sediment flux was highlighted (e.g. Lisitzin, 1995; Gordeev et al., 1996). Holmes et al. (2002) summarized the available estimates of sediment fluxes discharged from the Yenisei and Ob rivers into the Kara Sea and towards the Arctic Ocean.

The German-Russian project Siberian river run-off (SIRRO)

In spite of proceedings in marine sciences in the Kara Sea, the knowledge was still poor at the end of the nineties. Thus, between 1997 and 2002, five expeditions with the Russian research vessel "Akademik Boris Petrov" were realized within the joint German-Russian research project "Siberian river run-off" (SIRRO) (Matthiessen and Stepanets, 1998, Stein and Stepanets, 2000, 2001, 2002; Schoster and Levitan, 2003). The main goals of this multidisciplinary project were to improve our knowledge about modern processes on discharge with respect to the biology, (bio-) geochemistry, oceanography and geology and about past processes with emphasis on the Late Quaternary evolution (e.g. Fütterer and Galimov, 2003).

The main outcomes of this project with respect to the environmental history are outlined as follows: The discharge of the Yenisei and Ob rivers shows a high variability since the Late Glacial (e.g. Stein et al., 2003; Polyakova and Stein, 2004). During the early Holocene (ca. 11-9 ka cal. BP), highest accumulation rates of bulk (siliclastic) and of total organic carbon were detected, which is explained by increased river discharge and/or coastal erosion (e.g. Stein et al., 2003, 2004a). The increased river discharge is related to the final decay of the Putoran Mountain ice-sheet, which fed its melt water to the Yenisei river, which is expressed in the maximum peak of magnetic susceptibility (e.g. Stein et al., 2003, 2004a; Dittmers et al., 2003) and with higher summer temperatures (e.g. Andreev and Klimanov, 2000; Andreev et al., 2002).

At that time, large areas of the Kara Sea shelf were still exposed and freshwater was discharged considerable farther north (see also Stein et al., 2003, 2004a). Rivers incised channels in the exposed shelf, which were submerged subsequently due to the sea-level rise and left a widespread submarine meandering system (e.g. Dittmers et al., 2007). River load, which includes large quantities of terrigenous sediments, was mainly accumulated in the area of the so-called "marginal filter" (Lisitzin, 1995). This depocenter, where fine-grained particles were accumulated due to flocculation and coagulation processes (Lisitzin, 1995), retreated gradually southward relative to the sea-level rise (e.g. Stein et al., 2003; Polyakova and Stein, 2004; Simstich et al., 2004, 2005). In this context, the gradually decrease of accumulation rates between 9 and 4 ka cal. BP is explained with this southward displacement. Except for short-term variations, river discharge continued to decrease during the last 2000 years, which is related to an enhanced effectiveness of the marginal filter (e.g. Polyakova and Stein, 2004; Fahl and Stein, 2007).

Based on a comprehensive data collection (for further details see Stein and Fahl, 2004b), an average Holocene budget of total organic carbon and total sediment relating to defined provinces and to specific time intervals was calculated (Stein and Fahl, 2004). Also, a mass balance for both rivers (Dittmers et al., 2003) and for the entire Kara Sea was estimated, revealing that during the late Holocene only 19 % of the of total sediment input was exported towards the central Arctic Ocean (Stein and Fahl, 2004).

Key results of the SIRRO project were entered in the Kara Sea chapter in a book of the organic carbon cycle in the Arctic Ocean (Stein and Fahl, 2004) and in a contribution (Stein et al., 2004) for the special issue (edited by THIEDE) in the journal "Quaternary Science Reviews" summarizing the results of the Europan Science foundation (ESF)-funded project "Quaternary environments of the Eurasian north" (QUEEN).

2. Study area

2.1 Physiographic settings

The Kara Sea (Fig. 2-1) represents one of the Eurasian shelf seas with an area of approximately 930 x 10^3 km^2, which is almost 10 % of the whole area of the Arctic Ocean (Jakobsson et al., 2002, redefined limits). Its most characteristical feature are the two large rivers Ob and Yenisei discharging together ca. 1050 km^3/yr onto the shelf and toward the central Arctic Ocean (Gordeev et al., 1996). The Yenisei river transports comparably less sediment onto the shelf in relation to the Ob river (4.7 and 15.5 Mt/yr, Holmes et al., 2002), because of the building of several dams since the middle of the 20^{th} century. Generally, river discharge is strongly influenced by seasonal and interannual variability (e.g. Gebhardt et al., 2004). Both rivers drain gigantic hinterland areas: The Yenisei river has a catchment area encompassing over 2.5 million km^2. The Ob river drains the West Siberian Lowland (WSL), representing one of the world's largest peatland complexes.

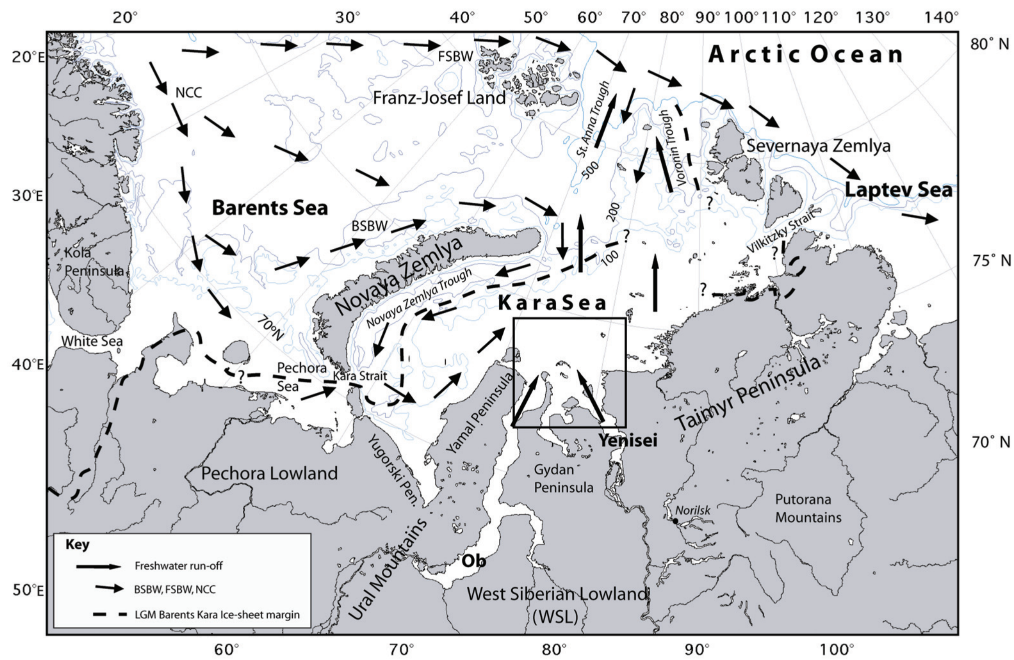


Fig. 2-1: Western Eurasian Arctic shelf seas and circulation patterns of surface water currents. The quadrate marks the study area, where the sediment cores were collected.

The mean water depth of the Kara Sea is relatively low with 114 m (Jakobsson, 2002), whereas the largest part in the southern and central Kara Sea is lower than 50 m (see also Jakobsson, 2002). The bottom topography is characterized by the deep Novaya Zemlya Trough extending along the east coast of Novaya Zemlya and the St. Anna and St. Voronin Troughs in the northernmost part at the continental slope. Otherwise, the flat central and southern area shows a hummocky relief as result of the glacial-interglacial sea-level changes (e.g. Fairbanks, 1989) and as result of the Pleistocene ice-sheet oscillations, which overran at least during the early and mid Pleistocene larger areas of the Kara Sea shelf (e.g. Svendsen et al., 2004). During sea-level low stands, Siberian rivers formed widespread channels in the exposed shelf, which became partly inactive during sea-level high stands (e.g. Dittmers, 2006). Moreover, periglacial/proglacial pedogenetical processes (such as solifluction) formed the past subaerial surface, which is now inundated.

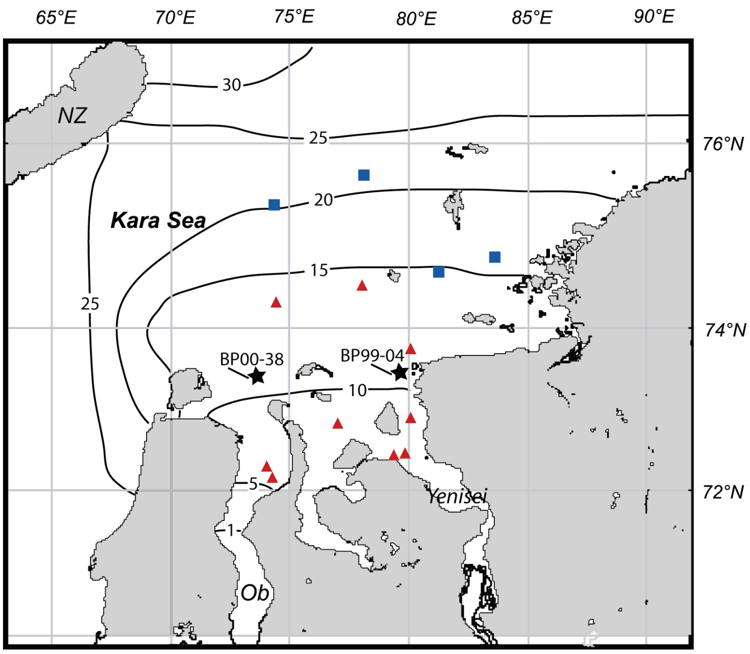


Fig. 2-2: Location of the two sediment cores BP99-04 and BP-99-04 and of the surface sediment cores. The black lines show the surface salinity distribution during summer (August-September) on average according to Pivovarov et al. (2003).

Observational data and models show that the river plumes contribute considerable to the hydrographic structure of the southern Kara Sea superimposed by seasonal and interannual variability (e.g. Pavlov and Pfirman, 1995; Pivovarov et al., 2003; Harms and Karcher, 1999, Fig. 2-2). The inflow of relatively warm and saline bottom water from the north along the (paleo-) channels to the south lead to a typical stratification of the water column with a strong hypnocline, which is temporally developed. Also the wind-driven inflow of Atlantic water through the Kara Strait in the southeastern part is enhanced during the winter when strong south to southwesterly winds prevail (Harms and Karcher, 1999). The main circulation patterns, the southward flow of the eastern Novaya Zemlya current and the northward Yamal current (Pavlov and Pfirman, 1995), are probably a seasonal feature (Harms and Karcher, 1999). The circulation patterns are still poorly known and more observational data in particular during spring are necessary (Karcher et al., 2003).

A further important feature, which is related to the freshwater supply, is the sea-ice formation, the sea-ice cover and export to the central Arctic Ocean, which substantially influence the biological and geological processes. Satellite images and images from aircrafts show a highly spatially and temporally variable sea-ice formation and sea-ice cover in the Kara Sea (e.g. Pfirman et al., 1995; Divine et al., 2004). Most characteristic is the extensive shore-fast ice along the coast and within the inner estuaries, and the open water polynya belts, strongly influenced by the prevailing wind direction and the variability of river discharge. The sea ice, which is exported toward the central Arctic Ocean and by the Transpolar Drift to the east coast of Greenland transports considerable amounts of incorporated particulate matter (e.g. Reimnitz et al., 1994; Nürnberg et al., 1994) and it is moreover suggested that phytoplankton and other biological components are transported in a similar way.

2.1 Modern climate and vegetation

The modern vegetation zones (without subdivisions) in the hinterland are shown in Fig. 2-3. As an aid in the interpretation of the pollen diagram, the distribution and ecology of important trees are outlined here according to Blyakharchuk and Sulerzhitsky (1999) with emphasis to the northwestern Siberia. According to them, forests with a high proportion of larch (*Larix sibirica* LEDEB.) are restricted to areas of permafrost including the north of western Siberia, where larch occurs in forest-tundra together with spruce. Spruce (*Picea obovata*) as larch tolerates permafrost but demands wetter climate and grows mostly along river valleys on well-drained river banks up to the northern boundary of the forest zone in Western Siberia. *Pinus sylvestris* grows only on unfrozen soils in sandy areas where it can be the only tree to dominate the forest. *Pinus sibirica* grows on both dry and wet unfrozen clayey and sandy-clayey soils, forming forests together with tree birch, *Abies* and *Picea*. Occasionally, it grows on peat should the actice layer of the permafrost be thick. *Betula pendula* forms zonal forests south of the taiga zone in Western Siberia. *Betula pubescens* (EHRH.) grows mostly in the birch forest-steppe in the south of Western Siberia and in forest-tundra in the north of Western Siberia; in the forest zone it also forms forests on eutrophic fens. *Betula nana* grows both in tundra and in mires in the forest zone. Abies sibirica (LEDEB.) is the most thermophilous among Siberian trees and has the narrowest ecological range, not growing on permafrost, peat, or wet soils.

Fig. 2-3: Vegetation zones (without subdivisions) in hinterland of the Kara Sea with respect to the West Siberian Lowland (WSL) (in Kremenetski et al., 2003 after Davydova and Rachkovskaya, 1990).

Further details on geobotanical features and vegetation zones with respect to north western Siberia consulted for this thesis refer mainly to Aleksandrova (1980, 1988), Atlas Arktiki (1985), Walter and Breckle (1994) and Schultz (1995). Further, the map of the floristic provinces of the "west Siberia group" according to CAVM Team (2003) were referred.

Finally, the online database of the Swedish Naturhistoriska Riksmuseet (http://linnaeus.nrm.se/flora/welcome.html) including mostly descriptions and maps on northwestern Siberian species was helpful.

The modern climate conditions in the Kara Sea region near the shore are influenced on a broad scale in the cold season by Atlantic air masses (Icelandic low pressure system) and in the warm season by the Siberian anticyclonic high pressure system (Atlas Arktiki, 1985; Pavlov and Pfirman, 1995). Across the boundary of the West Siberian lowland and the middle Siberian plateau, parallel to the course of the Yenisei river, the climate changes by gradually increasing Siberian anticyclone activity that expands seasonally westwards. The prevailing wind direction in the warm season is from northeast to northwest and in the cold season from south to southeast (Atlas Arktiki, 1985; Pavlov and Pfirman, 1995).

3. Material and methods

The two sediment cores BP99-04 and BP00-38 and the surface samples were retrieved during summer expeditions with the Russian research vessel "Akademik Boris Petrov" in 1999 and 2000 (Stein and Stepanets, 2000, 2001). Core BP99-04 is located in the northern Yenisei estuary (73°24,9´N, 79°40,5´E, 32 m water depth). The core recovery is 795 cm and covers the time period since 9400 to 600 cal. BP. The uppermost part was lost due to over penetration during coring resulting in a loss of sediments representing the past 600 years. Core BP00-38 is located in the outer Ob estuary on the slope of a filled river channel (73º11,8'N, 73º14,3'E, 20 (20 m water depth, Stein, 2001b). The core recovery is 652 cm representing the time since 9600 cal. BP. The 13 surface samples are situated between ca. 72°N and 76°N and ca. 74°E and 83°E and represent three N-S transects.

<u>*Laboratory treatment*</u>

Dinocysts are relatively sensitive to harsh chemicals. Thus, careful treatment has to be applied in order to avoid selective loss. Numerous former studies on dinocyst distribution could not easily be included into databases (see de Vernal et al., 2001) because of non-uniform laboratory procedure. In particular since the 1990s due to the collaborative agreements involving GEOTOP (University of Quebec) and the DGO (Département de Géologie et d'Océanographie, University of Bordeaux), the laboratory procedures were standardized predominantly.

The laboratory processing follows standard palynological procedures without using acetolysis (e.g. Rochon et al., 1999). After freeze-drying, samples were treated with cold hydrochloric (10%) and hydrofluoric acids (38-40%) to dissolve carbonates and silicates. Most of the fine organic matter was removed by wet sieving to enrich the particulate organic matter (POM) larger than 6 µm. *Lycopodium* spore tablets were added to calculate pollen and aquatic palynomorph concentrations (Stockmarr, 1971; Berglund and Jalska-Jasiewiczowa, 1986; Faegri and Iversen, 1989).

Palynological analysis includes terrestrial palynomorphs (pollen & spores) and selected aquatic palynomorphs (dinoflagellate cysts, freshwater chlorococcalean algae, acritarchs, and organic linings of benthic foraminifers) as well. They were counted under a Zeiss light microscope (Axioplan) using phase and differential interference contrasts at a magnification of 400x and 1000x.

The simultaneous analyses of terrestrial and aquatic palynomorphs in the same samples offer various advantages, in particular in the context of land-sea correlation. Usually, the analyses are conducted by two analysts ("pollen counter" and "dinocyst counter") and mostly based on two different samples from the same core, whose laboratory processing could differ considerably (e.g. pollen samples are treated by acetolysis).

Differing from this practice, herein, the analyses were performed by one analyst simultaneously on the same sample with equal laboratory processing. The main reason to provide the palynological analysis in this way, was

- to reduce differences in the results caused by the different laboratory treatment, which could influence the palynomorph assemblages (e.g. selective degradation), and
- to economize one complete sample collection and one counter.

Lack of application of the dinocyst assemblages for statistical analysis

Dinocyst assemblages are often used for further statistical analysis in order to obtain quantitative paleodata (e.g. de Vernal et al., 2001). Few recent studies were performed without quantitative reconstruction. This practice provide undoubtedly valuable data for modelling experiments about climate change in past and future. However, this application is based on sufficient number of counted dinocysts for each sample in order to reveal a reliable data set. In the circumarctic shelf seas such as the Canadian Arctic Archipelago (e.g. Mudie et al., 2001) or the Eurasian shelf seas (e.g. Kunz-Pirrung, 1998; Polyakova et al., 2005; de Vernal et al., 2005), dinocyst content is often relatively low and only semi-quantitative or/and qualitative reconstructions can be performed. Moreover, the used modern hydrographic data set for the dinocyst reference data base (de Vernal et al., 2001, 2005) does not cover adequately the inner and shallow shelf areas of the Kara Sea.

At first sight, the qualitative interpretation of the data appears to be less accurately in relation to the quantitative reconstructions. Nevertheless, just in less-known areas, qualitative dinocysts records give important evidence on hydrographical changes and can fill gaps due to the low abundance or absence of other microfossils such as foraminifers.

4. A Holocene marine pollen record from the northern Yenisei Estuary (southeastern Kara Sea, Siberia)

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4.1 Abstract

A 780 cm long sediment core from the northern Yenisei Estuary (southeastern Kara Sea) was analysed for pollen to reconstruct the Holocene vegetation and climate history of the coastal area of the Kara Sea region. The core shows a high and continuous deposition of sediments from 8900 yrs BP (9400 cal. BP) to ca. 600 yrs BP. A pronounced change of the lithology and the occurrence of marine to brackish water dinoflagellate cysts and molluscs indicate that the core location was reached by sea water at 8600 yrs BP (9200 cal. BP) when the global sea-level was approximately 30m below the present level. The depositional environment changed gradually from fluvial to estuarine conditions.

Favourable climatic conditions with higher mean temperature than at present and a widespread occurrence of spruce in boreal forests in the hinterland prevailed between 8900 and 7400 yrs BP (9400 to 8300 cal. BP). Between 7400 and 5000 yrs BP (8300 to 5700 cal. BP), relatively stable warm climatic conditions were established. Sedges dominated fens and peat bogs were widespread in the coastal lowlands indicating high water saturation and moist climate conditions. Since 5000 yrs BP (5700 cal. BP), and more pronounced since 3800 yrs BP (4200 cal. BP), long-distance transported pollen (mainly pollen of *Pinus sylvestris*) increased gradually and *Picea* pollen decreased reflecting the onset of climate cooling and the movement of the arctic tundra vegetation zone southward. A short-term warming event occurred between 4200 and 3800 yrs BP (4750 to 4200 cal. BP). The most pronounced change occurred at ca. 2200 yrs BP, when

Picea pollen decreased notably indicating the retreat of the spruce tree line. Additionally, the simultaneous increase of pollen taxa such as *Salix*, *Artemisia*, Ranunculaceae, and *Thalictrum* suggests a colder climate.

4.2 Introduction

The paleoenvironmental evolution of the Kara Sea region has attracted much attention in the past years because of its crucial location at the eastern margin of the late Weichselian Barents-Kara Ice Sheet. Particularly, the eastern extent of the ice sheet during the Last Glacial Maximum (LGM) is a matter of controversy (e.g. Grosswald and Hughes, 2002; Mangerud et al., 2002). The generally accepted reconstruction, based on extensive fieldwork on well-dated land sections, implies that the southeastern Kara Sea was not glaciated, possibly with the exception of the northern part of the Taimyr Peninsula (e.g. Astakhov et al., 1999; Manley et al., 2001; Mangerud et al., 2001, 2002; Alexanderson et al., 2002; Forman et al., 2002). Therefore, the southern Kara Sea and the adjacent hinterland are well-suited for detailed studies of late Weichselian and Holocene paleoenvironments.

Palynological and macrofossil analyses performed on a number of sections from peat bogs and lakes in the hinterland of the Kara Sea revealed distinct movements of vegetation zones and associated changes in the position of the Arctic tree line since the LGM (e.g. Khotinskiy, 1984; Koshkarova, 1995; Velichko et al., 1997; Peteet et al., 1998; Andreev et al., 1998, 2001, 2002; Andreev and Klimanov, 2000; Kremenetski et al., 1998a; Hahne and Melles, 1999; MacDonald et al., 2000). Moreover, reconstructions of Siberian paleoclimate using numerical methods and models have yielded quantitative information on temperature and precipitation (Peterson, 1993; TEMPO, 1996; Ganopolski et al., 1998; Monserud et al., 1998; Tarasov et al., 1999; Andreev and Klimanov, 2000; CAPE, 2001; Velichko et al., 2002). However, paleoclimate data from the coastal regions of the southern Kara Sea are sparse due to the lack of continuous late Weichselian and Holocene sequences (e.g. Andreev et al., 1998, 2001; Serebryanny et al., 1998; Serebryanny and Malyasova, 1998). Thus, the northernmost land sections covering most of the Holocene are located more than 400 km south and southeast of the shore line (Clayden et al., 1997; Jasinski et al., 1998; Andreev et al., 1998; Hahne and Melles, 1999; Andreev and Klimanov, 2000). Therefore, terrestrial paleoclimate gradients cannot be adequately resolved in the coastal region and compared with the marine paleoclimate evolution of the southeastern Kara Sea.

Marine pollen sequences from the inner Kara Sea may fill this gap because estuaries such as those of the Yenisei and Ob rivers are areas of high and continuous deposition of sediments since they were flooded due to the postglacial sea-level rise (e.g. Stein et al., 2003). Previous studies on Arctic and subarctic shelf sediments also illustrate the potential of pollen to reconstruct paleoclimate change in the adjacent coastal region and to link the terrestrial and marine paleoclimate records (e.g. Kulikov and Khitrova, 1982; Mudie, 1982; Mudie and Short, 1985; Hill et al., 1985; Mudie and McCarthy, 1994; Naidina and Bauch, 2001; Levac et al., 2001). Furthermore, offshore pollen spectra may provide supplementary information on large-scale patterns and processes occurring in the adjacent coastal area, whereas pollen records from terrestrial sites reflect rather the local and regional landscape development (Faegri and Iversen, 1989; Moore et al., 1991).

In this study, we present the first high-resolution, well-dated Holocene pollen record from the southern Kara Sea. We correlate this local pollen stratigraphy with published data from continental pollen sites and provide new information on the large-scale mid-Holocene warming and Late Holocene cooling trends, which were interrupted by short climatic events (e.g. Andreev and Klimanov, 2000), accompanied by vegetation zone shifts and tree-line movements in the hinterland.

4.3 Study area

4.3.1 Oceanography and Hydrography

The Kara Sea is one of the large Eurasian continental shelf seas covering an area of approximately 883,000 km^2 (Pavlov and Pfirman, 1995) (Fig. 4-1A). The bottom topography is much more variable than that of the other circum-arctic shelf seas. The western and northern parts are dissected by the deep (>400 m) Novaya Zemlya, St. Anna and Voronin troughs. The southeastern shelf is less than 50 m deep, with numerous shoals and islands north of the Ob and Yenisei estuaries and west of Severnaya Zemlya. Submarine channels extend from the estuaries to the Novaya Zemlya Trough (Johnson and Milligan, 1967).

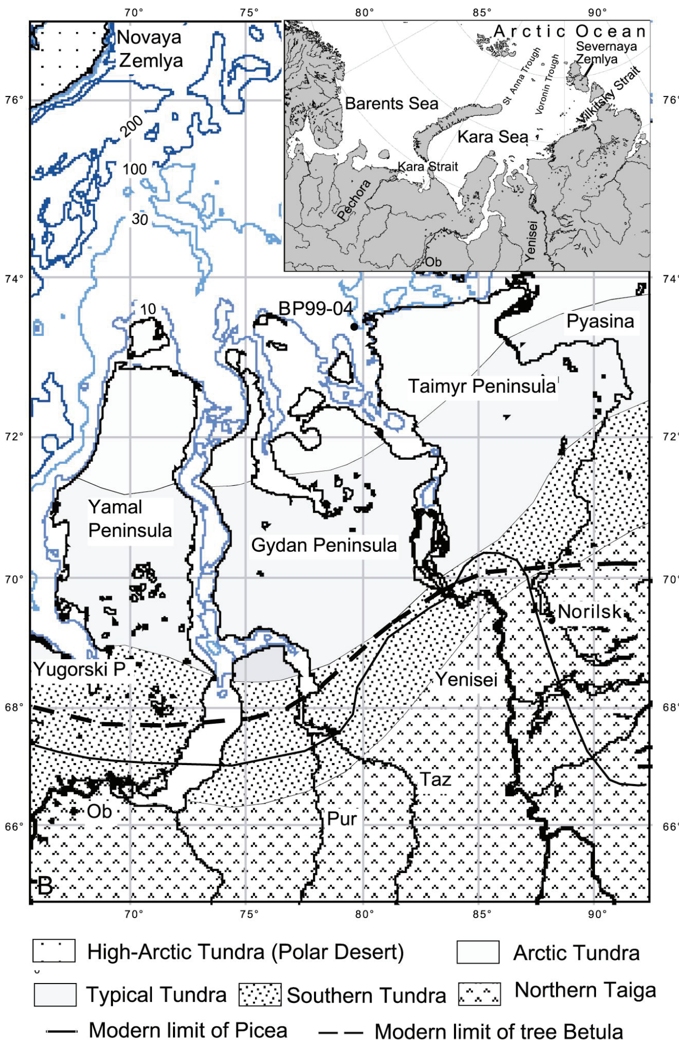

Fig. 4-1: (A) Overview map of northern Eurasia and the adjacent Siberian shelf seas with topographical data named in the text. (B) Detailed map of the southeastern Kara Sea with the core location of BP99-04. The vegetation zones and modern limits of *Picea obovata* and tree *Betula* in the coastal region and hinterland are shown (Atlas Arktiki, 1985; MacDonald et al., 2000; Kremenetski et al., 1998a).

The inner Kara shelf is semi-enclosed by Novaya Zemlya, the Severnaya Zemlya Archipelago, and the Siberian hinterland (Fig. 4-1A). The southern Kara Sea is connected with the Barents and Laptev seas through the Kara Strait and the Vilkitsky Strait, respectively. The St. Anna and Voronin troughs link the southern Kara Sea with the central Arctic Ocean. Freshwater discharge by the rivers Ob and Yenisei which drain about 2,580,000 km^2 strongly influence the hydrographical and depositional conditions in the southern Kara Sea (Pfirman et al., 1995; Gordeev et al., 1996; Polyak et al., 2002; Dittmers et al., 2003). The riverine suspended matter was mainly deposited in the outer estuaries during the Holocene (Lisitzin, 1995; Dittmers et al., 2003). Although various processes such as resuspension by storm activity, sea-ice formation and sea-ice gouging may lead to export of sediments from the depositional center (Pavlov and Pfirman, 1995), radiocarbon dating revealed that continuous Holocene sections may be retrieved at some locations (Stein et al., 2003).

4.3.2 Climate

The modern climate conditions in the Kara Sea region near the shore are influenced on a broad scale in the cold season by Atlantic air masses (Icelandic low pressure system) and in the warm season by the Siberian anticyclonic high pressure system (Atlas Arktiki, 1985; Pavlov and Pfirman, 1995). Across the boundary of the West Siberian lowland and the middle Siberian plateau, parallel to the course of the Yenisei river, the climate changes by gradually increasing Siberian anticyclone activity that expands seasonally westwards. The prevailing wind direction in the warm season is from northeast to northwest and in the cold season from south to southeast (Atlas Arktiki, 1985; Pavlov and Pfirman, 1995). The storm activity is relatively high (Pavlov and Pfirman, 1995).

The average air temperature shows a large range during the cold season (January) between –20 ˚C in the western part to –30 ˚C in the eastern part, and in the warm season (July) between ca. +6 ˚C in the coastal area and +10 ˚C to +12 ˚C in the hinterland (Atlas Arktiki, 1985). The annual precipitation on average ranges between 300 to 600 (800) mm depending on the topographical location. Permanent permafrost soils cause characteristic periglacial pedogenetical and geomorphogical conditions (e.g. IPA, 1998; Schultz, 1995).

4.3.3 Modern vegetation

The modern vegetation in the coastal area and the adjacent hinterland is characterized by a latitudinal succession from south to north (Fig. 4-1B). Boreal forests and forest tundra are replaced by subarctic and arctic tundra and finally by polar desert covering most islands (Aleksandrova, 1980, 1988; Atlas Arktiki, 1985; Walter and Breckle, 1994). Longitudinal changes are most pronounced across the eastern boundary of the West Siberian lowland and the middle Siberian plateau (Taimyr Peninsula, Putorana Plateau), characterized by an increase of plant species numbers towards the east. While the western part of the study area is influenced by European species, the eastern part is covered by typical Siberian species (Alexandrova, 1988; Kienast et al., 2001).

The land area adjacent to the Yenisei Estuary belongs in a geobotanical sense to the tundra region, which is subdivided into the ”Yamal-Gydan-West Taimyr subprovince” of the subarctic tundra and the ”Yamal-Gydan-Taimyr-Anabar subprovince” of the arctic tundra (Aleksandrova, 1980, 1988). Lichen and moss-lichen variants of shrub and herb-dwarf shrub tundras occur in the first subprovince. Various stages of polygonal mires and numerous lakes are characteristic landscape elements. Shrubs such as *Betula nana*, *Salix lanata*, *S. pulchra* and *Alnus fruticosa* are widespread. Typical low shrubs covering this area are *Vaccinium vitis-idaea*, *V. uliginosum* ssp. *microphyllum*, *Ledum decumbens*, *Rubus chamaemorus* and *Empetrum hermaphroditum*. Herbs and sedges such as *Dryas punctata*, *Cassiope tetragona*, *Eriophorum angustifolium*, *Carex ensifolia* ssp. *arctisibirica*, *C. stans* and *C. rariflora* are characteristic elements. Peat mosses (*Sphagnum* sp.) are abundant among the mosses, particularly in the polygonal mires. Brown mosses like *Drepanocladus revolvens*, *Hylocomnium splendens*, *Aulacomnium* sp., *Dicranum* sp., *Calliergon* sp. and *Polytrichum* sp. are the most characteristic species. *Cladonia* and *Cetraria* species are the most widespread lichens.

In the second subprovince, the vegetation cover is closed or almost closed. Polygonal mires are less frequent. The most characteristic floristic elements are arctic willows such as *Salix polaris* and *S. nummularia*. Furthermore, *Dryas punctata*, *D. octopetala*, *Luzula confusa*, *Carex ensifolia* ssp. *arctisibirica* and *C. stans* have a widespread distribution. Characteristic mosses are peat mosses (*Sphagnum* sp.), brown mosses of the genera *Aulacomnium*, *Hylocomnium*, *Ptilidium*, *Dicranum*, *Drepanocladus* and *Polytrichum*. Characteristic lichens are the genera *Cladonia*, *Cetraria* and *Dactylina*.

The polar desert occurs in the Kara Sea region mainly on the small islands and in the northern part of Novaya Zemlya and is characterized by a decreasing vegetation cover and the predominance of cryptogamous over angiosperms and lichens over mosses (Alexandrova, 1980, 1988).

4.4 Material and Methods

The gravity core BP99-04/07 was collected during the Kara Sea expedition of RV ”Akademik Boris Petrov” in 1999 (Stein and Stepanets, 2000). It is located in the northern Yenisei Estuary (73°24,9´N, 79°40,5´E, 32 m water depth), approximately 30 km from the shoreline (Fig. 4-1A und B), where high sedimentation rates prevailed during the Holocene (Dittmers et al., 2003). The core recovery was 795 cm, but the uppermost part was lost due to over-penetration during coring.

4.4.1 Lithology

The lithology of the core consists of relatively homogeneous bioturbated dark olive grey silty clays to clayey silts in the upper part, and bedded to bioturbated silty sands at the base (Fig. 4-2 and 4-3). Based on visual description, magnetic susceptibility data and acoustic profiles, Stein (2001) and Dittmers et al. (2003) distinguish two lithological units (Fig. 4-2). This description is refined by examination of x-radiographs, granulometric data and total organic carbon contents (TOC) (see also Stein et al., 2003).

Unit Ia (0 to 740 cm) consists of homogeneous bioturbated silty clays to clayey silts. Bivalves (e.g. *Portlandia* sp.) occur down to the base of the unit, but decrease in abundance from 440 cm core depth to the top of the core. Dinoflagellate cysts occur also in the Unit Ia. At 510 cm core depth, a piece of wood was found. The TOC content slightly decreases from 740 to 370 cm core depth and is almost constant (~1%) from 370 to 120 cm core depth (Fig. 4-3). In the uppermost part TOC contents increase notably. Unit Ia was deposited in an estuarine environment.

The contact of Unit Ia and Ib is characterized by a transition from bioturbated to bedded sediments. Unit Ib (740 to 780cm) consists of silty sands and shows a maximum of magnetic susceptibility (Dittmers et al., 2003; Stein et al., 2003). Bivalves and dinoflagellate cysts are absent in Unit Ib. Bioturbation is highly variable. Parallel bedding (partly laminae with <1mm thickness) to cross bedding and channel infill structures are common. Laminae are usually disrupted by bioturbation. Contacts between bioturbated and bedded layers are irregular showing erosional features. The basal part of the sediment core was probably deposited in a fluvial sedimentary environment.

4.4.2 Radiocarbon chronology

The age model is based on twelve AMS ^{14}C radiocarbon dates, which were performed on *Portlandia* sp. shells (Table 4.1; see also Stein et al., 2003). The ^{14}C ages were corrected for a reservoir effect of 440 yr (Mangerud and Gulliksen, 1975) and were calibrated into calendar years BP with the program CALIB 4.3 (Stuiver et al., 1998). There is one age reversal at 646 cm core depth which is located in a massive layer indicating rapid sedimentation. Therefore, it is not used to obtain the final age model. Ages for the core top and core base were extrapolated using the linear sedimentation rates of the next units.

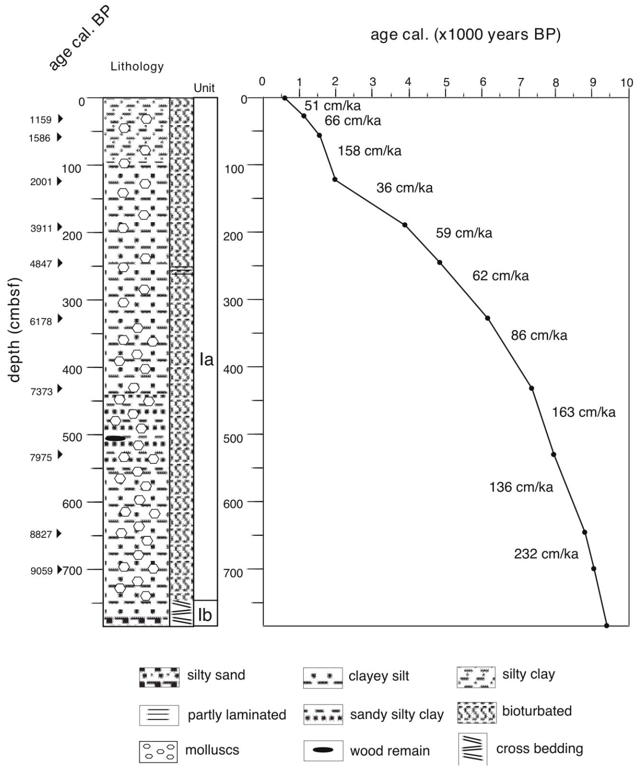

Fig. 4-2: Lithology, texture and linear sedimentation rates of sediment core BP99-04 (modified after Stein et al., 2003.).

Relatively high average sedimentation rates prevailed during the past ca. 9000 cal. yrs BP ranging from 36 cm/ka to 232 cm/ka (Fig. 4-2). The contact between Units Ia and Ib is dated at approximately 9200 cal. BP (8600 yrs BP).

4.4.3 Palynological analysis

In total, 52 samples were taken for palynological analysis at intervals of about 15 cm. The average temporal resolution ranges from 45 to 200 years, depending on the sampling interval and sedimentation rates.

The laboratory processing follows standard palynological procedures without using acetolysis (e.g. Rochon et al., 1999). After freeze-drying, samples were treated with cold hydrochloric (10%) and hydrofluoric acids (38-40%) to dissolve carbonates and silicates. Most of the fine organic matter was removed by wet sieving to enrich the particulate organic matter (POM) larger than 6 µm. *Lycopodium* spore tablets were added to calculate pollen concentrations (Stockmarr, 1971; Berglund and Jalska-Jasiewiczowa, 1986; Faegri and Iversen, 1989). Besides terrestrial palynomorphs, aquatic palynomorphs (cysts of dinoflagellates, chlorococcalean algae, acritarchs) were encountered. The results of the aquatic palynomorph studies will be published elsewhere. Reworked pre-Quaternary pollen were counted, but not assigned to taxa.

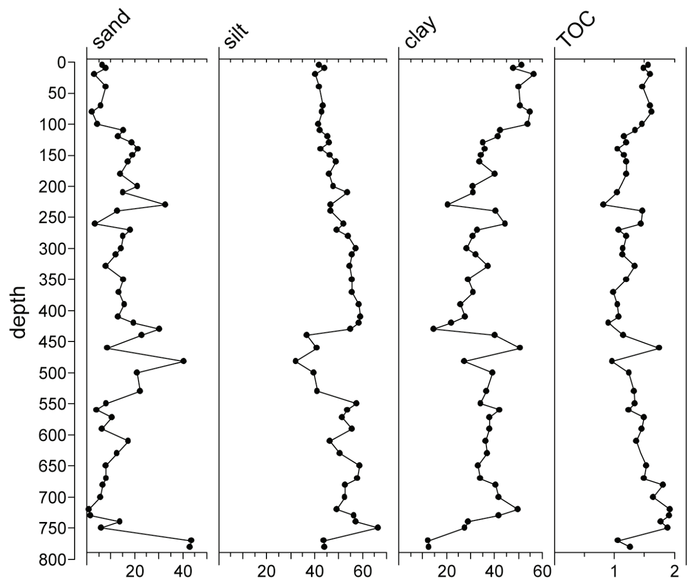

Fig. 4-3: Granulometric composition (%) and TOC content (%) of sediment core BP99-04.

Pollen were identified and counted under a light-microscope (Fa. ZEISS) using x400 and x1000 magnification. The nomenclature and taxonomy of pollen and spores basically follows Moore et al. (1991). The TILIA and TILIAGRAPH software was used to plot the pollen data (Grimm, 1991). In total, 53 pollen types were identified.

Because of the specific fluvial and aeolian transport and deposition processes in neritic environments (Brush and DeFries, 1981; Heusser, 1983; Brush and Brush, 1994; Traverse, 1994; Mudie and McCarthy, 1994), the standard pollen sum is on average much lower than in lacustrine records or peat sections (Faegri and Iversen, 1989; Berglund and Jalska-Jasiewiczowa, 1986). The pollen sum includes all arboreal (AP) and non-arboreal pollen (NAP) excluding the spores (Fig. 4-4). For discussion of the regional vegetation changes, typical long-distance transported pollen is excluded from the pollen sum and is separately shown in Figure 4-5.

The complete data set can be retrieved from the PANGAEA information system at the Alfred Wegener Institute for Polar and Marine Research, Bremerhaven (http://www.pangaea.de).

Table 4-1: AMS ^{14}C datings of sediment core BP99-04 performed on shells of *Portlandia* sp. (Stein et al., 2003). Reservoir correction after Mangerud and Gulliksen (1975). The interpolated age (*) was calculated for a depth midway between the dated levels at 350.0 and 700.0 cm core depth.

Core depth (cm)	^{14}C Age (BP)	Reservoir corrected age (-440 yr BP)	Calendar age (cal yrs BP)	Laboratory numbers
29.0	1630±20	1190±20	1159	KIA-12781
57.0	2070±25	1630±25	1586	KIA-12782
122.5	2450±30	1990±30	2001	KIA-10239
191.0	3980±30	3540±30	3911	KIA-10238
246.0	4695±30	4255±30	4847	KIA-10237
329.0	5800±40	5360±40	6178	KIA-10236
420.0	6855±35	6415±35	7234	KIA-10235
432.0	6890±45	6450±45	7373	KIA-10234
530.0	7585±35	7145±35	7975	KIA-10233
632.0	8345±50			KIA-10232
646.0		7887*	8827*	
658.5	8310±40			KIA-10231
700.0	8725±40	8285±40	9059	KIA-10230

4.5 Results

All analysed samples contain pollen in sufficient amounts (100-700 grains) to calculate percentage abundances (Fig. 4-4). Pollen concentrations show a maximum at the base of the core and are relatively constant from ca. 670 cm core depth to the top of the core although sediment grain sizes are relatively variable.

4.5.1 Pollen stratigraphy

Six local pollen assemblage zones (LPAZ) were defined and named after the location of the core in the Yenisei Estuary (Ye). The standard pollen diagram was plotted versus depth (cf. Berglund and Ralska-Jasiewiczowa, 1986; Moore et al., 1991).

The <u>LPAZ Ye-I</u> (780 – 674 cm) is characterized by maximum abundance of *Picea* pollen (>40%) in core BP 99-04, relatively high values (>40%) of *Pinus* Diploxylon pollen type, a notable decrease of *Betula* pollen, low values of Cyperaceae pollen (<20%), and the maximum of Polypodiales spores. Reworked pollen is rare.

Total pollen concentrations increase from the base of the core to a maximum of >24,000 grains/g dry sediment.

At the base of LPAZ Ye-II (674 – 574 cm), the pollen spectra show a notable change. The arboreal pollen spectra are marked by a change in dominance from *Picea* pollen to *Pinus* pollen. The *Pinus* Haploxylon type increase in abundance and is continuously present in zone Ye-II. *Salix* pollen is more abundant at the base and the top of the zone. *Betula* and *Alnus* pollen decrease from a maximum at the base to the top of this zone. The non-arboreal pollen spectra are characterized by a sharp increase of Cypera ceae pollen and a peak of Poaceae, *Artemisia* and Ericaceae pollen in the upper part. Furthermore, *Thalictrum* pollen is present. An increase of reworked pollen is notable.

LPAZ Ye-III (574 – 298 cm) is characterized by relatively low percentages of *Betula* pollen and *Pinus* Diploxylon pollen type. *Salix* pollen is only sporadically present in some samples. *Abies* pollen shows a closed curve reaching approximately 5%. Cyperaceae pollen is abundant (around 40%). The appearance of *Typha latifolia*, *T. angustifolia* and *Polemonium* pollen is notable. Reworked pollen shows a sharp increase.

LPAZ Ye-IV (298 – 123 cm) is characterized by a distinct increase of the *Pinus* Diploxylon pollen type and *Betula* pollen. The *Picea* pollen curve shows fluctuations, but decrease notably in the upper part of this zone. The non-arboreal pollen spectra are marked by a slight decrease of Cyperaceae pollen, and an increase of *Artemisia* and *Thalictrum* pollen. Furthermore, pollen of *Anthemis* type and Gentianaceae pollen is present. Reworked pollen decreases in zone LPAZ Ye-V.

LPAZ Ye-V (123 – 0 cm) is defined by a minimum of *Picea* pollen (< 5%), and by a sharp increase of the *Pinus* Diploxylon pollen type up to ca. 80% in the lower part of this zone. Furthermore, *Betula* pollen increases slightly. The increase is more distinct, when *Pinus* Diploxylon pollen type is exluded from the pollen sum (see Fig. 4-5). *Pinus* Haploxylon pollen type decreases. *Salix*, *Artemisia*, Ranunculaceae and *Thalictrum* pollen are continuously present with low amounts. Reworked pollen decreases to a minimum in core BP99-04. Pollen concentrations increase slightly.

4.6 Discussion

4.6.1 Depositional processes

In marine systems, various factors influence pollen transport and deposition (e.g. Brush and DeFries, 1981; Heusser, 1983; Mudie and McCarthy, 1994; Traverse, 1994; Naidina and Bauch, 1999). The distance of the core location from the source of pollen, the atmospheric and the ocean circulation patterns are the major processes affecting the composition of marine pollen spectra. Thus, these spectra may reflect both transport processes and changes of the vegetation cover in the hinterland.

Certain pollen taxa may be selectively enriched in marine assemblages with increasing distance from the source. Particularly, arboreal bisaccate pollen (e.g. *Abies*, *Pinus, Picea*) are an important component of the long-distance transported pollen (exotic pollen), whereas non-arboreal pollen such as herbs decrease their abundances rapidly further offshore (e.g. van der Knapp, 1987; Mudie and McCarthy, 1994). Thus, the AP:NAP ratio may change significantly until no realistic signal of the vegetation cover is preserved in the spectra.

Principally, it is assumed, that near-shore pollen records may reflect accurately the vegetation pattern in the coastal zone (e.g. Mudie and McCarthy, 1994). However, long-distance atmospheric transport may considerably influence pollen spectra in the Arctic region because of an extremely sparse vegetation cover or low pollen production (Gajewski et al., 1995; Andreev et al., 1997; Birks and Birks, 2000). Transport distances of up to 3000 km for pine and spruce were recorded for the Canadian Arctic under specific weather conditions (Campbell et al., 1999). Similarly, airborne sampling revealed significant amounts of exotic pollen over the Zevernaya Zemlya Archipelago (Kalugina et al., 1981) and Svalbard (Johansen and Hafsten, 1988). Exotic pollen in snow and firn samples from Franz-Josef Land and Severnaya Zemlya are sometimes dominated by *Pinus* pollen (Andreev et al., 1997; Bourgeois, 2000). In an ice core from Vavilov Ice Cap (79°27´N) on the Severnaya Zemlya Archipelago which is located in the polar desert, *Pinus* and *Picea* are the common components in the Holocene spectra (Andreev et al., 1997).

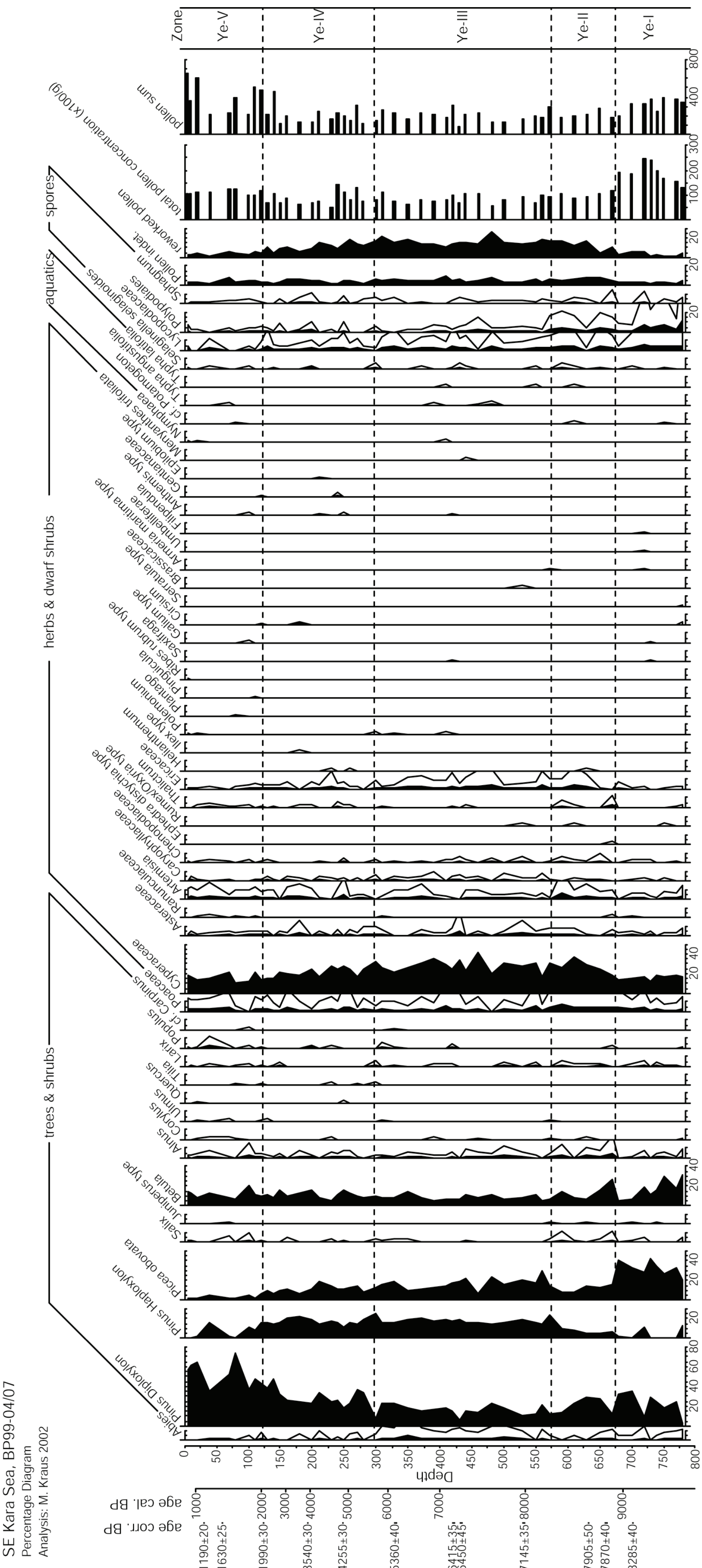

Fig. 4-4: Standard pollen percentage diagram of the marine sediment core BP99-04. The pollen sum includes AP and NAP (excl. spores). Low values are shown in 5x exaggeration.

Modern and Holocene pollen spectra from Severnaya Zemlya and other remote areas such as the Svalbard Archipelago and Jan Mayen also contained exotic pollen (Kalugina et al., 1981; van der Knaap, 1987,1988). Conifer pollen which are not present in the vegetation of the adjacent coastal area reach more than 90% in recent sediments from the Laptev Sea (Naidina and Bauch, 1999). Holocene sediments from the Kara and Laptev seas may comprise up to 40% of *Pinus* pollen (Kulikov and Khitrova, 1982; Naidina and Bauch, 2001).

Although core BP99-04 is located 400 to 500 km north of the modern limits of spruce, larch and tree birches, and 700-800 km north of the tree line of pine (Kremenetski et al., 1998a; Peteet et al., 1998; MacDonald et al., 2000), both the core top sample and modern pollen spectra from the adjacent tundra contained exotic tree pollen (cf. Fig. 4-4 and 4-5; Clayden et al., 1996; Tarasov et al., 1998). This illustrates that, even in the tundra, deposition of exotic pollen overprints the spectra, and that tree pollen may occur far beyond their modern limit (e.g. Gervais and MacDonald, 2001). Therefore, the marine pollen record of core BP99-04 comprises both a signal of long-distance transport and regional coastal vegetation cover.

The depositional environment is further influenced by sedimentary processes, which can obscure the real vegetation patterns (e.g. Brush and DeFries, 1981; Naidina and Bauch, 1999). Sediments and pollen may be supplied from the hinterland and the coast by erosion and river run-off. Resuspension of bottom sediments in the Yenisei Estuary by currents and by storms may lead then to deposition outside the respective vegetation zones. Pollen of core BP99/04 was deposited in a fluvial to estuarine environment. Distinct changes in the composition of pollen spectra, however, do not correlate with any change in grain-size composition (Fig. 4-3), suggesting that the pollen spectra primarily reflect real changes in the vegetation zones and subordinate the effects of the sedimentary processes.

4.6.2 Pollen stratigraphical correlation of long-distance transported pollen and implications for tree line history

The long-distance transported pollen *Pinus, Picea*, and *Abies* (and broad-leaved forest trees such as *Ulmus*, *Quercus* and *Tilia*) were excluded from the pollen sum in order to discuss the Holocene vegetation history of the adjacent coastal area (Fig. 4-5). Macrofossil analyses suggest that these trees did not expand further north than 71°N

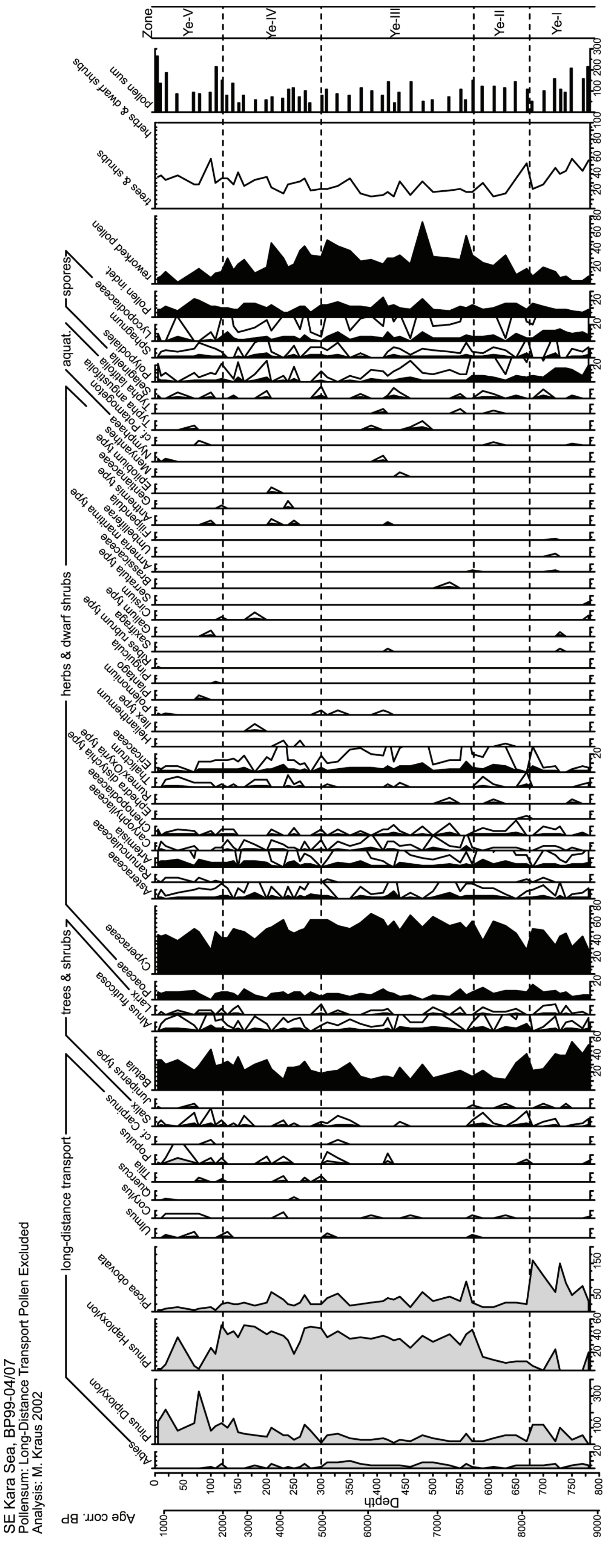

Fig. 4-5: Pollen percentage diagram of the marine sediment core BP99-04. Long-distance transported pollen are excluded from the pollen sum. Low values are shown in 5x exaggeration.

during the Holocene (Kremenetski et al., 1998a; MacDonald et al., 2000) and therefore never reached the coast immediately adjacent to the core location. Radiocarbon ages are used to correlate the marine pollen record with previous studies on land sections, which were all discussed with respect to radiocarbon ages.

The signal of long-distance transported pollen may be an excellent tool for correlation of terrestrial and marine sequences. Their temporal distribution may provide age control of sediments in cases where radiocarbon dates are unavailable (e.g. Levac and de Vernal, 1997). Furthermore, application of radiocarbon ages of organic remains may pose some problems because different material, such as macrofossils, may reveal considerable offsets in ages (Wohlfarth et al., 1998; Kilian et al., 2002; Andreev et al., 2004.). Arboreal pollen spectra and in particular long-distance transported pollen could fill this gap. However, a regional pollen stratigraphy which is a prerequisite for stratigraphic correlation is still missing in northern Siberia.

The terrestrial sites closest to core BP99-04 are located on the Taimyr Peninsula around Norilsk, in the Pur-Taz region, on the Yugorski Peninsula (Clayden et al., 1997; Hahne and Melles, 1997, 1999; Peteet et al., 1998; Andreev et al., 1998, 2001, 2004) and in the Pechora basin (Kaakinen and Eronen, 2000). The exotic pollen *Pinus* and *Picea obovata* may be excellent stratigraphic markers for correlation of marine and terrestrial Holocene pollen records along a north-south transect supporting the notion that wind-transport may cause the almost coeval changes at various sites.

Pollen of *Picea obovata* increased initially in the Pur-Taz region around 9150 yrs BP (Peteet et al., 1998) and in Lama Lake around 9200 yrs BP (Andreev et al., 2004). Both the Derevanoi Lake site (Clayden et al., 1997) and the Yenisei core BP99-04 did not recover sediments older than 9000 yrs BP but show consistent presence of *P. obovata* from the core base. The expansion of *Picea* to sites north of the modern tree line occurred between 9000 and 8000 yrs BP (MacDonald et al., 2000), which is well reflected in the high *Picea* percentages in the zone Ye-I between 8900 and 8100 yrs BP. Possibly, the high percentages of *Picea* in the marine pollen record at this time indicate the highest density of spruce in the hinterland. In the coastal zone of the Laptev Sea, *Picea* pollen occurred slightly later around 8500 yrs BP (Pisaric et al., 2001). In contrast, a pollen record from the Pechora basin showed a distinct increase of *Picea* around 9000 yrs BP (Kaakinen and Eronen, 2000). The continuous retreat of *Picea obovata* during the middle and late Holocene is observed in many pollen records but marked changes in abundance were apparently time-transgressive. Thus, the sharp decline in late Holocene occurred at 3800 yrs BP and more pronounced at 2000 yrs BP in core BP 99-04 (Figs. 4-4 and 4-5), at 3500 yrs BP in the coastal Laptev Sea (Pisaric et al., 2001), at 4000 yrs BP in Lama Lake (Andreev et al., 2004), at 2500 yrs BP in the Yenisei section (Andreev and Klimanov, 2000), at ca. 3800 yrs BP (4200 cal. BP) in the peat section of the Salym-Yugan mire in the boreal West Siberia (Pitkänen et al., 2002) and at 5500 yrs BP in the Pechora basin (Kaakinen and Eronen, 2000). Generally, MacDonald et al. (2000) state, that the retreat of *Picea* occurred between 4000 and 3000 yrs BP.

Windblown pollen of *Pinus* Diploxylon type (i.e. *Pinus sylvestris*) shows an early Holocene double peak between 8900 to 8100 yrs BP corresponding with the *Pinus* pollen curve from the Entarny record situated further south and southwest in the taiga zone (peak at ca. 8200 yrs BP) (Velichko et al., 2002). Generally, the distribution of *Pinus sylvestris* in the early Holocene is not known in detail. Scots pine grew in the early Holocene beyond its present limit along the Yenisei river at 8000 yrs BP and retreated gradually to its present northern limit in late Holocene (Kremenetski et al., 1998b; Clayden et al., 1997). A substantial increase of *Pinus Diploxylon* pollen occurs during the last 5500 yrs and particularly since 2000 yrs BP (Fig. 4-4 and 4-5) correlating with an initial increase of *Pinus* at ca. 4500 yrs BP which became more pronounced at 2500 yrs BP in the Lama Lake diagram (Hahne and Melles, 1997, 1999; Andreev et al., 2004). Pitkänen et al. (2002) assess that *Pinus sylvestris* became the most abundant tree species after ca. 3900 yrs BP in the Salym-Yugan Mire area in boreal West Siberia. Finally, the Entarny diagram shows a noticeable increase with a double peak after ca. 2000 yrs BP (Velichko et al., 2002).

Pollen of *Pinus* Haploxylon type (*P. sibirica* and *P. pumila*) is abundant from ca. 8000 to ca. 2000 yrs BP with the most pronounced increase after 7500 yrs BP. Because of the biogeographical distribution (Kremenetski et al., 1998b) of both species, we assume, that the pollen in our record mainly belong to *P. sibirica*. The distinct increase at ca. 7500 yrs BP indicates that *P. sibirica* spread out later than other coniferous trees in the hinterland. The main period of spread and population growth occurred between 8000 and 4000 yrs BP (Kremenetski et al., 1998b). Peteet et al. (1998) found no macrofossils of *P. sibirica* in the Pur-Taz region, but registered a slight increase in windblown pollen between 8000 to 4500 yrs BP. In contrast to the notable late Holocene increase of windblown pollen of *Pinus sylvestris,* pollen of *P. sibirica* shows a sharp decrease after 2000 yrs BP apart from a solitary peak around 1400 yrs BP.

Pollen of *Abies* occurs sporadically in the whole sediment core, but is more abundant between 7150 to 5000 yrs BP with a small maximum at 5600 yrs BP. Blyakharchuk and Sulerzhitsky (1999) also observed maximum abundances in a pollen record from the Bugristoye bog (situated in the southeastern part of the West Siberian Plain in the Tomsk province) between 6500 to 5500 yrs BP when *Picea* and *Pinus sylvestris* pollen decreased. After 2000 yrs BP, *Abies* pollen disappeared almost completely. Thus, the distribution of *Abies* pollen in core BP99-04 reflects the northward advance of the fir tree line during the middle Holocene in the southern hinterland and its retreat after 2000 yrs BP (Peteet et al., 1998). However, the tree line history of fir in Siberia and their advance is little known to allow a more accurate correlation.

Macrofossil samples from *Larix* across northern Siberia revealed that it was probably the dominant tree genus in the northern taiga zone between 8000 and 4000 yrs BP (Koshkarova, 1995; Peteet et al., 1998; MacDonald et al., 2000). In the Yenisei pollen record, pollen of *Larix* is rare because of a general under-representation in marine pollen records (Janssen, 1984; Peteet et al., 1998; Pisaric et al., 2001).

4.6.3 Paleoenvironmental reconstruction

The extrapolated age of the core base (ca. 8900 yrs BP/9400 cal. BP) indicates that the vegetation development and tree line history cannot be reconstructed for the earliest part of the Holocene. The sedimentary texture of the silty sands in the lower part of zone Ye-I from 780 to 745 cm core depth (Fig. 4-2), and the absence of marine dinoflagellate cysts show that the core was located in a fluvial depositional environment. The bioturbated muds, the presence of brackish water molluscs and the continuous increase of marine dinoflagellate cysts from 740 cm core depth show that the location was reached by rising sea-level since ca. 8600 yrs BP (9200 cal. BP). It was gradually flooded until 8100 yrs BP (8900 cal. BP), that is indicated by the distinct increase of the concentration of marine palynomorphs from 670 cm core depth. This is in good agreement with the global sea-level (Fairbanks, 1989), which was approximately 30 m below the present level at that time. Without considering compaction, the contact of Units Ia and Ib at ca. 36 m below seafloor was at ca. 6 m water depth at that time.

Pollen Zone Ye-I (ca. 8900 to 8100 yrs BP)

The high content of *Betula* pollen (probably mainly tree birch) and Polypodiales spores in the lower part of this zone and the absence or low abundance of typical cold indicators such as *Salix*, *Thalictrum*, Ranunculaceae and *Dryas* pollen indicate favourable climatic conditions at least from ca. 8900 yrs BP (9400 cal. BP). Obviously, tree birch dominated forests in the hinterland preceeded spruce dominated forest communities, whereas tree birch probably never occurred in the adjacent coastal area. Dated macrofossils from northern Eurasia conform with this interpretation and indicate that the northern limit of tree *Betula* was approximately at 72°N in the Gydan and western Taimyr Peninsulas (MacDonald et al., 2000; Forman et al., 2002). The discontinuous increase of long-distance transported pollen (mainly *Picea*, *Pinus* Diploxylon type) up to the top of zone Ye-I indicates a northern extent of the boreal forest, which was dominated by larch (Peteet et al., 1998; MacDonald et al., 2000; Pisaric et al., 2001) but also characterized by the occurrence of spruce. The highest values of *Picea* pollen between ca. 8900 to 8100 yrs BP demonstrate the widespread occurrence of this tree corresponding well with the northward expansion of spruce in the early Holocene as documented by stomate analyses (Clayden et al., 1997; Kremenetski et al., 1998a; MacDonald et al., 2000; Pisaric et al., 2001). Since the region was not covered by the Barents-Kara ice sheet during the LGM (Mangerud et al., 2002), *Picea* could relatively rapidly expand northward from their glacial refuge.

Furthermore, the highest pollen concentration and high sedimentation rate of more than 230 cm/ka (Fig. 4-2) may indicate high pollen productivity and high river-input from the Yenisei river, giving evidence for favourable large-scale climatic conditions such as higher mean summer temperature than today in the area. This is supported by a coeval maximum of pollen concentrations at about 9000 yrs BP in the Lama Lake pollen record from Taimyr Peninsula (Andreev et al., 2004). This period corresponds well with a so-called Boreal thermal optimum, that was described from many sites in Northern Eurasia (Velichko et al., 1997; Andreev and Klimanov, 2000; Andreev et al., 2002).

The discontinuous decrease of *Betula* pollen and the synchronous increase of Poaceae in LPAZ Ye-I (Fig. 4-5) as well as the occurrence of Ranunculaceae pollen probably indicate a short-term cooling between 720 to 680 cm core depth from ca. 8400 to 8100 yrs BP (9100 to 9000 cal. BP). This event is almost synchronous with a cooling event at ca. 8300 to 8000 yrs. BP, which was recorded from the whole northeastern

European Russian Arctic (Khotinskiy, 1984; Velichko et al., 1997) and the Pur-Taz area and the Yugorski Peninsula in the hinterland of Yenisei Estuary (Peetet et al., 1998; Andreev et al., 2000, 2001). However, the sample interval is too low to unequivocally resolve this event in core BP99-04. Possibly, there is a relationship between this cooling event and the sharp decrease of *Picea* at the base of zone Ye-II because of a temporary degradation of spruce forests at their northern limits (cf. Khotinskiy, 1984).

Pollen zone Ye-II (ca. 8100 to 7400 yrs BP)

After this temporary deterioration, the arboreal pollen spectra reflect a stronger differentiation of the boreal forest than before. Alder and willow shrubs were more common. The presence of *Pinus* Haploxylon pollen type (*Pinus sibirica* and *P. pumila*) indicates long-distance transport from further south and east, respectively (*P. pumila*) and probably a northward migration of *P. sibirica* (Kremenetski et al., 1997, 1998b). However, the northern limit of *P. sibirica* during the mid-Holocene is not well known. Peteet et al. (1998) found a slight increase of *P. sibirica* from about 8000 to 4500 yrs BP, but did not find any macrofossils which might indicate local occurrence. The occurrence of *Corylus* pollen reflects the movement of southern broad-leaved forests northward. Large-scale favourable climate conditions were established (e.g. Khotinskiy, 1984; Velichko et al., 1997, Andreev et al., 2001). Generally, the advance of the tree line in early Holocene is related to the increase of continentality due to the lower sea-level (cf. CAPE, 2001), which was approximately 30 to 25 m below the present level (Fairbanks, 1989). The coastline was a few tens of kilometers north of its current position and extensive regions of the Kara Sea shelf area were still exposed. Additionally, increased summer insolation at high northern latitudes in early Holocene could enhanced the effect of increased summer temperatures (e.g. Ritchie et al., 1983).

The increase of Cyperaceae pollen particularly in this zone (see also Fig. 4-5) indicates moister conditions than before and reflects the development of wetlands such as fens and peat bogs. Usually, Cyperaceae pollen is under-represented in marine sediments (Mudie, 1982), whereas *Sphagnum* spores are over-represented (Heusser, 1983). This relationship is reflected e.g. in Holocene pollen spectra from the Laptev Sea (Naidina and Bauch, 2001). In contrast, *Sphagnum* spores are continuously rare in the Yenisei pollen record (Fig. 4-4). Furthermore, Mudie (1982) observed a seaward increase of *Sphagnum* spores in recent sediments from the Labrador Sea and explained this pattern with offshore aeolian rather than fluvial transport. Thus, the combination of high abundances of Cyperaceae pollen and low abundances of *Sphagnum* spores can be interpreted either as product of specific pollen transport processes and selective deposition in estuarine environments, or may also indicate the predominant genesis of type of mires. The second interpretation is more likely: Due to the gradual sea-level rise, the river was piled up and bursted its banks. As a further consequence the drainage was reduced and the water table rose. Fixed nutrients were released and were now available for plants. Relatively eutrophic open-water mineral wetlands emerged, enabling the establisment of grasses-sedges phytocoenoses. Sedges, particularly species of *Carex* are important peat producers, and minerogenous mires such as water rise mires and flood mires could develop (e.g. Joosten and Clarke, 2001).

In contrast, *Sphagnum* mosses are usually characteristic floristic elements of (oligotrophic-mesotrophic) ombrogenous mires such as ”aapa” mires (Velichko et al., 1998). With respect to this geobotanical background, widespread sedge-fens and sedge-peat bogs probably occurred in lowlands and river valleys in the coastal area in the early to mid Holocene. In any case, peat accumulation was widespread and played an important role in this area. Probably, polygonal mires, characteristic of continuous permafrost, were absent and appeared subsequently in the late Holocene (Vardy et al., 1997; Jasinski et al., 1998; Peteet et al., 1998; Oksanen et al., 2001). In the hinterland, typical mires of boreal forests such as ”aapa” mires could have occurred in the boreal forest zone. However, the prevailing mire type in the low-lying coastal areas might have been water rise and flood mires, which are formed mainly by sedges.

The distinct increase of reworked pollen (pre-Quaternary pollen) in zone Ye-II and its continuous higher abundances mainly in zones Ye-III and Ye-IV is attributed to increased coastal erosion due to the sea-level rise. The constant high values of reworked pollen during the transgression shows that coastal erosion plays an important role in the Kara Sea for the sediment budget (cf. Rachold et al., 2000; Dittmers et al., 2003).

Pollen zone Ye-III (7400 yrs BP to 5000 yrs BP)

The pollen spectra of zone Ye-III reflect increased differentiation of the vegetation communities and the further advance of the forest zone northward (Khotinskiy, 1984), which is indicated by the closed curve of *Abies* pollen, continuous high percentages of *Pinus* Haploxylon pollen type and *Picea* pollen, and the rare occurrence of *Salix* pollen. In conjunction with the highest abundance of Cyperaceae pollen, these pollen spectra indicate high water saturation in the coastal region and adjacent areas accompanied by an adequate precipitation rate and warm conditions. The appearance of *Typha latifolia* and *T. angustifolia* might support this interpretation (Andreev et al., 2001). The high abundance of Ericaceae pollen is attributed to dwarf shrubs of boreal forest as well as heath of wetlands. However, both ecological groups indicate wet conditions in the landscape. The mid-Holocene interval spans the so-called ”Holocene climatic optimum” which was the warmest period in the Siberian Arctic during the Holocene (e.g. Velichko et al., 1997; Andreev and Klimanov, 2000; Andreev et al., 2002).

Pollen zone Ye-IV (ca. 5000 yrs BP to 2200 yrs BP)

The notable increase of windblown pollen of *Pinus sylvestris*, the decrease of *Abies* and *Picea* pollen and Cyperaceae pollen, the occurrence of *Thalictrum* pollen and the increase of *Artemisia* pollen indicate the onset of a substantial change in the marine pollen record and a trend to cooler climatic conditions.

This general trend is interrupted from ca. 4200 yrs BP to 3800 yrs BP shown by a *Picea* pollen peak, a *Betula* minimum, gaps in the *Artemisia* and *Thalictrum* pollen curves and a notable Ericaceae and Cyperaceae pollen peak. This warming event in the marine record is tentatively correlated with the lower part of the Lama Lake pollen zone VII at ca. 4200 yrs BP (Andreev et al., 2004). Khotinskiy (1984) described a ”middle subboreal warming” from 4100 to 3200 yrs BP and emphasized that broad-leaved forest advanced in the southern and middle taiga in this warm interval. This is reflected in the simultaneous presence of *Quercus* and *Tilia* pollen in the marine record.

Picea pollen shows in the Yenisei pollen record a pronounced decrease after ca. 3800 yrs BP indicating a retreat of the spruce tree line (MacDonald et al., 2000; Pisaric et al., 2001; Pitkänen et al., 2002; Andreev et al., 2004) At the same time, *Artemisia* pollen is constantly abundant indicating the advance of the tundra zone southward. The lowest sedimentation rate with less than 40 cm/ka between 3500 and 2000 yrs BP (Fig. 4-2; Stein et al., 2003) and the concomitant decrease of reworked pollen suggest that erosion in the hinterland and at the coast was strongly reduced. This was probably due to permafrost aggradation that commenced e.g. in the Labaz Lake area on the Taimyr Peninsula most pronounced after ca. 2900 yrs BP (Kienel et al., 1999) and at the Pechora Sea coast after 3100 yrs BP. (Oksanen et al., 2001) and the termination of the postglacial sea-level rise around 5000 yrs BP (Fairbanks, 1989).

Pollen zone Ye-V (ca. 2200 yrs BP to ca. 600 yrs BP)

The most pronounced environmental change occurred at the transition of zone Ye-IV to V. A strong cooling trend is indicated by a decrease of *Picea* pollen and *Pinus* Haploxylon pollen type, the occurrence of *Salix* pollen, the higher abundance of Poaceae and *Artemisia* pollen, the presence of Ranunculaceae and *Thalictrum* pollen and the decrease of Ericaceae pollen. The increase of windblown *Pinus* Diploxylon pollen type results from general impoverishment of vegetation cover.

The relatively high abundance of Cyperaceae pollen can be interpreted as widespread occurrence of sedges in vegetation communities, indicating continuing peat accumulation. Probably, peat accumulation was reduced but did not stop (see also Vardy et al., 1997). Due to the cooling trend, other genetic types of mires, particularly polygonal mires developed. Moreover, Cyperaceae pollen may reflect *Carex* species growing on wet mineral subsoils (e.g. *Carex ensifolia* ssp. *arctisibirica*) in the tundra vegetation zones (Aleksandrova, 1980, 1988).

Indirectly, the retreat of boreal forests in the hinterland indicates the expansion of continuous permafrost soils. A fundamental change in environmental conditions occurred which influenced both the marine and continental environments caused by complex interaction of the vegetation, atmosphere and ocean (e.g. TEMPO, 1996; Ganopolski et al., 1998; Peteet et al., 1998; MacDonald et al., 2000; CAPE, 2001).

4.7 Conclusions

The basal age of the marine sediment core is ca. 8900 yrs BP (9400 cal. BP), derived from the age model and supported by correlation of marine pollen stratigraphy with continental pollen zonation. At this time, the core was located in a fluvial environment on the exposed Kara Sea shelf. At 8600 yrs BP (9200 cal. BP), the seawater reached the core location due to the postglacial sea-level rise and a typical neritic-estuarine environment developed. High river discharge caused highest sedimentation rates in the Yenisei Estuary. The marine pollen spectra show a northward advance of boreal forest in the hinterland indicating warm and relatively moist environmental conditions. *Betula* dominated forests preceded *Picea* dominated forest communities. The role of larch is under-represented in this record caused by specific marine depositional environment. High values of *Picea* pollen indicate the northern occurrence of spruce woods at this time and their relatively high abundance in boreal forest. The highest pollen concentrations suggest high bioproductivity.

At ca. 8400 to 8100 yrs BP (9100 to 9000 cal. BP), a short-term climate deterioration overprinted the early Holocene favourable climate. Possibly, this cooling event involved a temporary degradation of spruce wood.

After ca. 8100 yrs BP to ca. 6400 yrs BP (9000 to 7300 cal. BP), the marine pollen spectra indicate the establishment of long-term favourable climate conditions in the coastal Kara Sea region and adjacent areas. The arboreal pollen spectra reflect a stronger differentiation of the boreal forest. In the lowlands of Yenisei and Ob rivers and their catchment area, sedge fen and peat bogs dominated water rise and flood mires asreconstructed mainly from the high abundance of Cyperaceae pollen. The distinct increase of reworked pollen is related to stronger coastal erosion due to the sea-level rise.

After ca. 6400 yrs BP (ca. 7300 cal. BP) favourable climatic conditions still prevailed but the changes in marine pollen spectra, in particular those of the arboreal pollen, suggest the onset of a long-term climate cooling. After a warming event from ca. 4300 to 3800 yrs BP (4900 to 4200 cal. BP), a pronounced retreat of boreal forest and the southward spread of tundra vegetation zone occurred. After ca. 2200 yrs BP unfavourable climatic conditions and the modern vegetation zones were established in the coastal area and in the hinterland of the Kara Sea region.

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5. Holocene variability of sea-surface conditions in the inner Kara Sea (Arctic Ocean) based on dinoflagellate cysts

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Abstract

On a well-dated sediment core from the outer Yenisei estuary (southeastern Kara Sea, Arctic Ocean, Russia), qualitative reconstructions of hydrographical changes since 9200 cal. BP were performed based on aquatic palynomorph assemblages (organic-walled dinoflagellate cysts, acritarchs, chlorophycean algae, benthic foraminifer linings). Fluvial/estuarine conditions prevailed until 8900 cal. BP, when sea level reached the site. A shallow water environment with brackish/marine conditions and an enhanced stratification of the upper water column existed since ca. 7200 cal. BP.

A local thermal optimum was detected with highest sea-surface temperatures (SST) and a minimum of seasonal sea-ice cover between 9200 and 7200 cal. BP, which was succeeded by a long-term cooling with distinct steps at 6400 and 4500 cal. BP. Since ca. 3300 cal. BP modern conditions with cold and polar water masses and extensive sea-ice cover were established.

The direct correlation of the aquatic palynomorph assemblages with pollen assemblages on the same sediment core (Kraus et al., 2003) revealed a comparable paleoenvironmental evolution both on land and in the sea. However, aquatic palynomorphs reflect a distinct termination of the thermal optimum in contrast to the marine pollen record, which show rather gradual climate deterioration with distinct steps at ca. 5700 and 3800 cal.

BP. The establishment of sea-surface conditions comparable to today preceded the establishment of climate conditions similar as today since ca. 2000 cal. BP. Our results support the previously finding of an early Holocene thermal optimum in the coastal and island area of the inner Kara Sea, which is based on only few paleorecords in this area.

Keywords: *Holocene, marine palynology, dinoflagellate cysts, pollen, Kara Sea, Arctic Ocean, paleoceanography, land-sea-correlation*

Introduction

The Eurasian shelf seas encompass more than 50% of the Arctic Ocean (Jakobsson et al., 2003) and play an important role in the hydrological cycle of the Arctic Ocean (Serreze et al., 2003.). The supply of freshwater by the large Siberian rivers is essential to maintain the halocline structure of the Arctic Ocean, and may influence the global thermohaline circulation (THC) (e.g. Rahmstorf, 1995; Broecker, 1997). In this context, a currrent increase of river discharge is discussed in relation to a freshening of the Arctic basin (e.g. Peterson et al., 2002; Simstich et al., 2005), which may lead to a reduction of the THC by its effect on the sea water density (Broecker, 1997, Rennermalm et al., 2006).

Since two of the largest rivers of the world, the Ob and Yenisei, discharge more than 50 % of the whole Eurasian Arctic freshwater into the Kara Sea which is transported to the Arctic Ocean (e.g. Holmes et al., 2002), an improved understanding of the paleoenvironmental evolution in the Kara Sea which is the third largest Arctic marginal sea (Jakobsson, 2002) is of crucial importance for the whole Arctic Ocean system. Furthermore, the Kara Sea forms an important link between the northern North Atlantic, Barents Sea and the eastern Eurasian shelf seas.

Therefore, a number of studies were conducted on marine sediments from the northern Kara Sea, foccussing on the St. Anna Trough (Hald et al., 1999; Polyak et al., 1997; Boucsein et al., 2002) and adjacent regions (e.g. Lubinski et al., 2001; Kleiber et al., 2001; Voronina et al., 2001). Since the middle of the nineties of the last century, also the southern Kara Sea attracted more attention by expeditions with the Russian research vessel "Dmitriy

Mendeleev" (e.g. Lisitzin and Vinogradov, 1995). New insights were given into the riverine variability, which changed considerable during the Holocene, the sedimentation history and last deglaciation (e.g Lisitzin, 1995; Levitan et al., 1995; Kuptsov and Lisitzin, 2003; Lisitzin and Kuptsov, 2003; Polyak et al., 2000, 2002, 2003).

Nevertheless, comparable little was known about the Late Quaternary environmental evolution of the southern Kara Sea when the German-Russian project "Siberian River Run-Off" (SIRRO) was initiated in 1997. Since then, various studies, which emanate from this project, were published and provided a better understanding of the variability of river discharge and of sea-surface salinity (e.g. Stein et al., 2003, 2004a, Polyakova and Stein, 2004; Fahl and Stein, 2007), of changes in bottom water hydrography (Simstich et al., 2004, 2005), of the glaciation and depositional history, and Holocene sediment flux (e.g. Stein et al., 2004a; Dittmers et al., 2003, in press), and of the climate and vegetational history (Kraus et al., 2003).

These studies provided a highly dynamic view of a land-ocean system that changed rapidly from a dry shelf to a brackish-marine water environment, strongly influenced by sea-level rise, river discharge variability superimposed by seasonal and interannual variability.

Polyakova and Stein (2004) provided a reconstruction of sea-surface conditions based on diatom assemblages and organic carbon accumulations rates on the same sediment core. However, in particular changes of sea-surface temperatures and sea-ice cover and the regional inundation history caused by the postglacial sea-level rise are still less known. Also the potential occurrence and timing of the (marine) Holocene thermal optimum that has been studied e.g in the subarctic northwest Atlantic region and the western Arctic Ocean (0 - 180°W) (e.g. Kaufman et al., 2004; de Vernal. et al., 2005, 2006; Kaplan and Wolfe, 2006), has not been reconstructed in detail for the Kara Sea region. Hence, our current knowledge is based only on the marine paleorecords mentioned above and on a few land-based records, which are mostly characterized by low resolution, discontinuous sequences and few radiocarbon data (e.g. Velichko et al., 1997; Serebryanny and Malyasova, 1998; Serebryanny et al., 1998; et al., 1998, 2001, 2003). It is of advantage, that we can compare our results with previously published data from the same sediment core, in order to complement missing parameters such as sea-surface temperature and to evaluate the paleoevironmental implications with a further independent paleoproxy.

Aquatic palynomorphs, and in particular organic-walled dinoflagellate cysts (=dinocysts), are considered as a suitable proxy to study these variability in sea-surface conditions (e.g. Rochon et al., 1999; Mudie et al., 2001). In contrast to the Barents Sea (Voronina et al., 2001) and the Laptev Sea (Kunz-Pirrung, 1998, Kunz-Pirrung, 2001, Polyakova et al., 2005; Klyuvitkina and Bauch, 2006), there exist only a few data about modern dinocyst assemblages in the Kara Sea (Matthiessen, 1999; Matthiessen et al., 2000; Matthiessen and Kraus, 2001; Head et al., 2001).

The main purpose of this study was

- to infer the timing and progress of the inundation history,
- to reconstruct changes in sea-surface conditions,
- to correlate the aquatic palynomorph-inferred environmental evolution with the pollen-inferred reconstruction (Kraus et al., 2003) and
- to compare our results with terrestrial paleorecords.

Environmental setting

Modern Oceanography and Hydrography

The most characteristic features are the two Siberian rivers Yenisei and Ob, which discharge together about 1480 km^3 y^{-1} into the Kara Sea (Holmes et al., 2002). This freshwater supply influences considerable the hydrographic structure of the southern Kara Sea which are shown by observational data and models and contributes to a strong stratification of the upper water column (e.g. Burenkov and Vasil'kov, 1995; Pavlov and Pfirman, 1995; Harms and Karcher, 1999; Pivovarov et al., 2003; Harms et al., 2000; McClimans et al., 2000). Both rivers form an extensive estuarine system with brackish waters.

The central and inner part of the Kara Sea is relatively even with an average depth of less than 50 m (Jakobsson et al., 2002). Important bathymetric features are widespread (paleo-) river channels, which were formed mainly during sea-level low stands (Dittmers et al., in press). These channels act as pathways for suspended matter, which is deposited within the so-called "marginal filter" (Lisitzin, 1995), and/or is transported along the channels towards the central Arctic Ocean (e.g. Stein et al., 2001). Marine bottom water counteracts this freshwater and sediment outflow by an inflow of cold and saline water along the channels from the north to the south. In addition, Atlantic water enters the Kara Strait from the west, between Novaya Zemlya and the Pechora Lowland, and Nowaya Zemlya and Franz-Josef-Land (Harms and Karcher, 1999; Schauer et al., 2002). The outflow and inflow of marine water masses are strongly influenced by seasonal and interannual variability.

The estuaries are main centers of sea-ice formation influencing considerable sedimentological and biological processes (e.g. due to the prolonged duration of sea-ice cover, development of polynyas, erosional processes, transport of incorporated material) (e.g. Nürnberg et al., 1994; Reimnitz et al., 1994; Rachold et al., 2000; Divine et al, 2004). The annual duration of sea-ice is 9-10 months on average. Sea-ice is exported to the central Arctic Ocean and is driven by the Transpolar Ice Drift towards the Fram Strait (Nürnberg et al., 1994; Pfirman et al., 1997).

The modern climate in the Kara Sea region is influenced on a broad scale in the cold season by Atlantic air masses (Icelandic low pressure system) and in the warm season by the Siberian anticyclonic high pressure system (Atlas Arktiki, 1985; Pavlov and Pfirman, 1995). The prevailing wind direction in the warm season is from northeast to northwest and in the cold season from south to southeast (Atlas Arktiki, 1985; Pavlov and Pfirman, 1995). The storm activity is relatively high (Pavlov and Pfirman, 1995). For further detail see Kraus et al., (2003).

Material and Methods

Sediment sampling

The gravity core BP99-04/07 was collected during the Kara Sea expedition of RV "Akademik Boris Petrov" in 1999 (Stein and Stepanets, 2000). It is located in the northern Yenisei estuary (73°24,9´N, 79°40,5´E, 32 m water depth), approximately 30 km from the shoreline (Fig. 5-1), where high sedimentation rates prevailed during the Holocene (Dittmers et al., 2003). The core recovery was 795 cm, but the uppermost part was lost due to overpenetration during coring resulting in a loss of sediments representing the past 600 years.

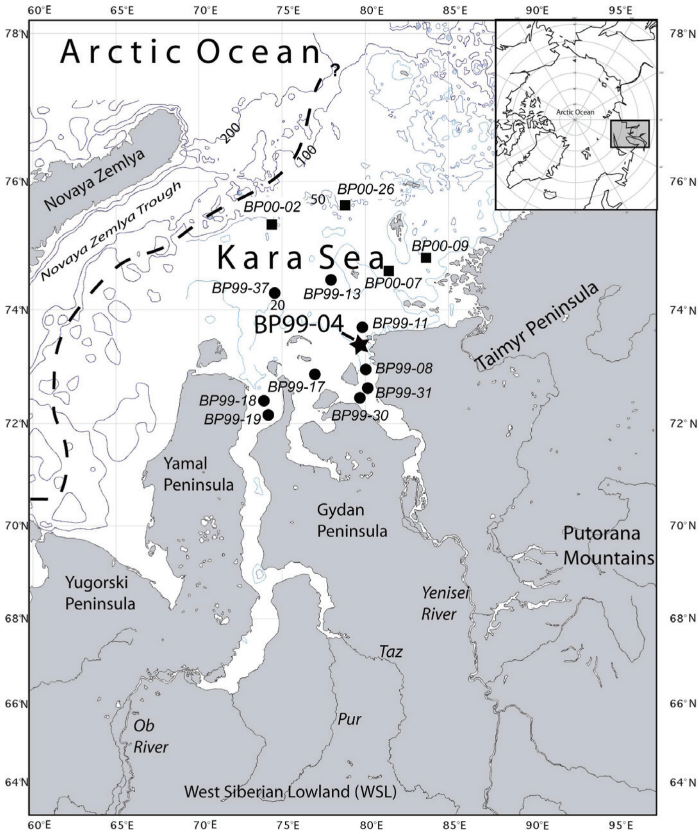


Fig. 5-1: Location of the sediment core BP99-04 (73°24,9´N, 79°40,5´E, 32 m water depth, southwestern Kara Sea) and of the surface sediment samples. Dashed line illustrates the margin of the LGM Barents-Kara ice sheet according to Svendsen et al. (2004). 20, 50, 100 and 200 m isobaths are plotted.

The multicorer surface sediment samples (BP99-08, BP99-11, BP99-13, BP99-17, BP99-18, BP99-19, BP99-30, BP99-31, BP99-37; BP00-02, BP00-07, BP00-09, BP00-26) were collected at 13 stations during the Kara Sea expedition of RV ”Akademik Boris Petrov” in 1999 (Stein and Stepanets, 2000) and 2000 (Stein and Stepanets, 2001) within the range of ca. 72°N and 76°N and ca. 74°E and 83°E, respectively to characterize the modern assemblages with respect to sea-surface conditions (Fig. 5-1). The locations of the stations can be retrieved from the data bank PANGAEA (http://www.pangaea.de) maintained by the Alfred Wegener Institute for Polar and Marine Research at Bremerhaven (Germany).

Lithology and sediment composition

The lithology of the sediment core consists of relatively homogeneous bioturbated dark olive grey silty clays to clayey silts in the upper part and partly laminated silty sands at the base

(Stein, 2001, Stein et al., 2003; see also Fig. 5-2). Stein (2001) and Dittmers et al. (2003) differentiated the sediment core BP99-04 into two lithological subunits, the bioturbated upper subunit Ia and the partly laminated subunit Ib at the basis of this core. The TOC content, which is relatively high, but typical for estuarine sediments, slightly decreases from 740 to 370 cm core depth and is almost constant (~1%) from 370 to 120 cm core depth. Only at 450 cm depth and between 260 to 240 cm depth, TOC show in each case a peak with occur coeval with a clay peak (Stein et al., 2003).

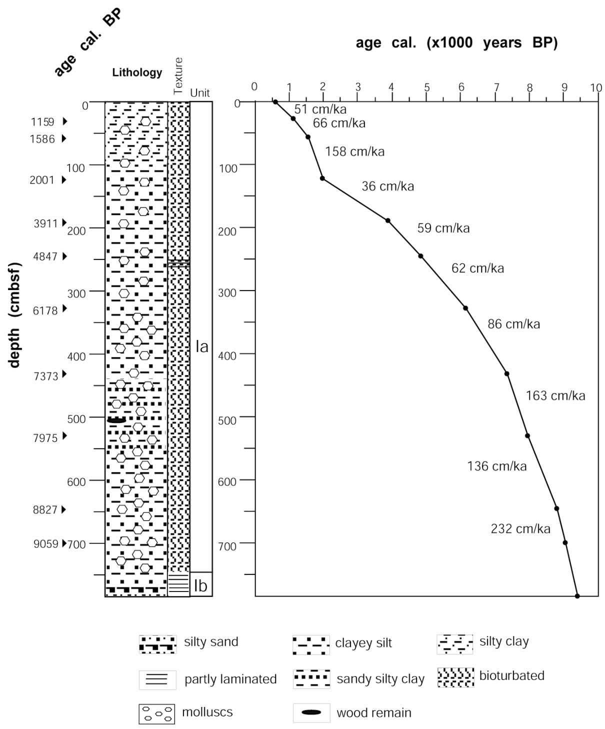


Fig. 5-2: Lithology, texture and linear sedimentation rates of sediment core BP99-04 (after Stein, 2000)

Radiocarbon chronology

The age model was previously published by Stein et al. (2003) and Kraus et al. (2003) and is based on twelve AMS ^{14}C radiocarbon dates on *Portlandia* sp. shells (Table 5-1). The ^{14}C ages were corrected for a reservoir effect of 440 yr (Mangerud and Gulliksen, 1975) and were calibrated into calendar years BP with the program CALIB 4.3 (Stuiver et al., 1998). There is one age reversal at 646 cm core depth that is located in a massive layer indicating rapid sedimentation. Therefore, it is not used to obtain the final age model. The lowermost radiocarbon age is 9059 cal. BP (Table 5-1). Ages for the core top and core base were extrapolated using the linear sedimentation rates of the next units.

Relatively high average sedimentation rates prevailed during the past ca. 9000 years ranging from 36 cm/ka to 232 cm/ka (Fig. 5-2). The contact between Units Ia and Ib is dated at approximately 9200 cal. BP.

Table 5-1: AMS ^{14}C dates from core BP99-04 according to Stein et al. (2003).

Core depth (cm)	^{14}C Age (BP)	Reservoir corrected age (-440 yr BP)	Calendar age (cal yrs BP)	Laboratory numbers
29.0	1630±20	1190±20	1159	KIA-12781
57.0	2070±25	1630±25	1586	KIA-12782
122.5	2450±30	1990±30	2001	KIA-10239
191.0	3980±30	3540±30	3911	KIA-10238
246.0	4695±30	4255±30	4847	KIA-10237
329.0	5800±40	5360±40	6178	KIA-10236
420.0	6855±35	6415±35	7234	KIA-10235
432.0	6890±45	6450±45	7373	KIA-10234
530.0	7585±35	7145±35	7975	KIA-10233
632.0	8345±50			KIA-10232
646.0		7887*	8827*	
658.5	8310±40			KIA-10231
700.0	8725±40	8285±40	9059	KIA-10230

Palynological analysis

In total, 52 samples were taken for palynological analysis at intervals of about 10-20 cm representing an average submillennial temporal resolution depending on the sampling interval and sedimentation rates. For the modern distribution of aquatic palynomorph assemblages, 13 multicorer cores from 0-1 cm were analysed for their palynological content. Standard palynological processing methods were carried out but no acetolysis was applied (Matthiessen, 1995; Rochon et al., 1999; Kraus et al., 2003). Because of abundant coarse particulate organic matter such as plant remains, samples were additionally sieved at 120µm mesh sizes.

Selected aquatic palynomorphs (dinoflagellate cysts, freshwater chlorophycean algae, acritarchs, and organic linings of benthic foraminifers) were counted under a Zeiss light microscope (Axioplan) using phase and differential interference contrasts at a magnification of 400x and 1000x. The dinocysts content was relatively low due to the influence of the frehswater plume. It was intended to count at least a minimum of 100 cysts but it could not be reached in all samples. Dinocyst sums < 50 were not shown as percentage values.

For calculation of percentages, concentrations and palynomorph accumulation rates and construction of the aquatic palynomorphs diagrams, the TILIA, TILIAGRAPH and TGVIEW software was used (Grimm, 1991, 2004). The complete data set can be retrieved from the data bank PANGAEA (http://www.pangaea.de) maintained by the Alfred Wegener Institute for Polar and Marine Research at Bremerhaven in Germany.

Unfortunately, the application of the modern analogue technique with the dinocyst reference data base (de Vernal et al., 2001; de Vernal et al., 2005), which was carried out in order to reveal a quantitative reconstruction of sea-surface conditions, did not provide useful data since the used modern hydrographic data set for the dinocyst reference data base does not cover adequately the inner and shallow shelf areas of the Kara Sea.

In total, 13 dinoflagellate cyst taxa were identified (Table 5-2). The calculation of the percentage diagram is based on the sum of all cysts counted (redeposited dinocysts are excluded) (Fig. 5-4). Concentrations and dinocyst accumulation rate were calculated for selected species (Fig. 5-5 and 5-6).

Nomenclature and taxonomy of the dinoflagellate cysts (dinocysts for brevity) follows basically Rochon et al. (1999) (Table 5-2). The determination of *Islandinium minutum* and related morphotypes follows Head et al. (2001). This species group of Protoperidiniaceae is difficult to identify due to their morphological variability, and thus not always, differentiation was possible or certain. Undeterminable specimens were assigned to *Islandinium* spp. indet. The determination of the morphotype *Polykrikos*? sp. sensu Kunz-Pirrung, 1998 follows the description and the photographic documentation by Kunz-Pirrung (1998, 2001). Cysts of *Operculodinium centrocarpum* with short processes which were found in this record, were assigned to the morphotype *O. centrocarpum* - short processes (Rochon et al., 1999). The majority of spherical brown protoperidinioid cysts could not be identified to species level, and were assigned to the species group *Brigantedinium* spp. indet.

Chlorophycean algae of the genera *Pediastrum* and *Botryococcus* were determined after Matthiessen and Brenner (1996) and Kunz-Pirrung (1998). The majority of the counted *Pediastrum* coenobia belongs to the species *Pediastrum boyranum.* There are some morphological varieties of *P. boryanum* in the samples, but they were not further distinguished and all varieties were assigned to *P. boryanum* agg. Only in four samples *P. kawraiskyi* was identified, which were assigned to the total algae concentration sum. All *Botryococcus* specimens were assigned to *B.* cf. *braunii* (see Matthiessen and Brenner, 1996).

Table 5-2: Detected dinoflagellate cysts in the sediment core BP99-04 and their affiliation to the respective biological taxon (thecate name).

Dinoflagellate cysts (paleontological name)	Biological affinity (thecate name)
Islandinium minutum Head, Harland & Matthiessen, 2001	*Unknown, probably Protoperidinium* sp. indet.
Islandinium? Cezare Head, Harland & Matthiessen, 2001	*Unknown, probably Protoperidinium* sp. indet.
Echinidinium karaense Head, Harland & Matthiessen, 2001	*Unknown, probably Protoperidinium* sp. indet.
Islandinium spp.	*Unknown, probably Protoperidinium* sp. indet.
Brigantedinium spp. Reid 1977	*Protoperidinium* sp. indet.
Brigantedinium simplex Wall 1965 ex Lentin & Williams	*Protoperidinium conicoides* (Paulsen) Balech
Brigantedinium cariacoense Wall 1965 ex Lentin & Williams	*Protoperidinium avellana* (Meunier) Balech
Brigantedinium denticulatum	*Protoperidinium* sp. indet.
Operculodinium centrocarpum sensu Wall & Dale 1966	*Protoceratium reticulatum* (Claparède & Lachmann) Bütschli
Operculodinium centrocarpum - short processes (sensu Wall & Dale 1966 - short processes, in Rochon et al., 1999)	*Protoceratium reticulatum* (Claparède & Lachmann) Bütschli
Polykrikos? sensu Kunz-Pirrung, 1998	probably *Polykrikos schwartzii* (Bütschli)
Spiniferites elongatus Reid 1974	*Gonyaulax elongata* (Reid) Ellegaard, Daugbjerg, Rochon & Lewis
Spiniferites ramosus (Ehrenberg) Mantell	*Gonyaulax* cf. *spinifera* (Claparède & Lachmann) Diesing
Rottnestia amphicavata Dobell & Norris	Unknown, probably *Gonyaulax spinifera*

The acritarchs *Halodinium* spp. and *Radiosperma corbiferum* were determined according to Kunz-Pirrung (1998). Besides the other mentioned aquatic palynomorphs, also organic linings of benthic foraminifers were counted.

<u>*Environmental significance of aquatic palynomorphs*</u>

In the following, the preferences and distribution patterns with emphasis to Eurasian shelf seas are shortly outlined. For further details about dinocysts and their respective significance to sea-surface conditions see Rochon et al. (1999), de Vernal et al. (2001), Head et al. (2001), Kunz-Pirrung (1999, 2001) and Matthiessen et al. (2005).

Dinocysts

The cyst assemblages are dominated by the cyst group of protoperidinioid cysts *I. minutum, I.? cezare* and *Echinidinium karaense*. This group is restricted to cold polar to subpolar water masses of the high latitudes (Head et al., 2001). High abundances are related to extensive seasonal sea-ice cover. Whereas *I. minutum* is specified to polar and north-temperate, the distribution of *I.? cezare* and *E. karaense* appear to be more restricted to polar environments (Head et al., 2001). *I. minutum* is a euryhaline species and prefer salinities ranging from about 10 to 35 psu (Kunz-Pirrung, 1998; Rochon et al., 1999; Head et al., 2001; de Vernal, 2001). The occurrence of *Polykrikos*? sp. sensu Kunz-Pirrung (1998) is associated with colder surface water temperatures ranging from –1.3 to 2.2 ºC with salinities ranging between 17 and 26 psu (Kunz-Pirrung, 2001).

In contrast to the latter cysts, the both taxa *Spiniferites elongatus* and *Rottnestia amphicavata* are restricted to milder conditions (Levac et al., 2001, de Vernal et al., 2001) and *Operculodinium centrocarpum* is associated with warmer sea-surface conditions (e.g. Voronina et al., 2001; de Vernal et al., 2001; Matthiessen et al., 2001).

The dinocyst *Brigantedinium* spp. indet. is a cosmopolitan and opportunistic species group, especially in epicontinental environments, but does not show any preferences with regard to temparature or salinity (de Vernal et al., 2001). However, the vegetative stage of this species is heterotrophic, and its elevated occurrence may therefore reflect higher nutrient contents and higher productivity..

Acritarchs

The distribution pattern of the acritrach *Radiosperma corbiferum* shows a certain affinity to higher salinities between 8 and 18 psu on the Laptev Sea shelf (Kunz-Pirrung, 1998). According to Matthiessen (1995), *Halodinium* spp. prefers probably lower salinities. In the Laptev Sea, this acritarch shows distinct preferences to low salinity values as well (Kunz-Pirrung, 1998).

Benthic foraminifer linings

Organic linings of benthic foraminifers were used as indicator for marine deposition (e.g Stancliffe, 1989; Batten, 1996; de Vernal., 1992; Kunz-Pirrung, 1998; Levac et al., 2001).

Chlorophycean algae

The chlorophycean algae *Pediastrum* spp and *Botryococcus* cf. *braunii* are used commonly as indicator for variability in river discharge (Matthiessen et al., 2000). Besides a hydrographical significance, changes in algae concentrations show also changes in the depositional environment, because these freshwater algae are dispersed mainly with the fine-grained sediments supplied by the rivers (Matthiessen et al., 2000).

Results

Modern distribution of aquatic palynomorphs in surface sediment samples

The composition of modern aquatic palynomorphs in surface sediment samples (dinocysts, chlorophycean algae, acritarchs, and organic benthic foraminifer linings) of the three transects along the surface salinity gradient from south to north revealed a distinct relationship to the respective surface water masses (Fig. 5-3). In the southern part within the range of the inner estuaries, chlorophycean algae (*Pediastrum* spp. and *Botryococcus* cf. *braunii*) represent almost completely the assemblages (> 80%). North of 73°N, the proportion of the chlorophycean algae decreases distinctly, but remains between ca. 25 to 40 % showing, that the river plume is still perceptible in the northern part of the Kara Sea at 75°N. The herein presented results coincide well with former studies on the distribution of aquatic palynomorphs in the Kara Sea, which documented a relationship to the respective surface water conditions (Matthiessen, 1999; Matthiessen and Kraus, 2001; Head et al., 2001) and with the distribution in the Laptev Sea (Kunz-Pirrung, 1998, 1999, 2001).

The dinocyst sums of the surface sediment samples north of 73°N are relatively low ranging between 33 and 187 counted dinocysts. Only three samples have more than 100 counted dinocysts, although it was intended to reach more than 100 cysts. Therefore, no statistical analysis was carried out, because commonly a minimum of 100 counted dinocysts is required. The dinocyst concentrations range between 49 and 1499 cysts/g dry sediment. The concentrations of chlorophycean algae amount to 100 to 3480 algae/g dry sediment.

The dinocyst assemblages show a quite low diversity (< 15 taxa). The protoperidinioid dinocyst *Islandinum minutum* predominates the surface sediment samples (>57 to 82%) together with the *Brigantedinium* spp. (6-16%). The complete data set can be retrieved from the data bank PANGAEA (http://www.pangaea.de) maintained by the Alfred Wegener Institute for Polar and Marine Research at Bremerhaven (Germany).

Due to the overall predominance of the protoperidinioid dinocyst *Islandinum minutum* and of related taxa, we assume, that selective degradation plays a minor role, because this species are considered as extremely sensitive to degradation (Zonneveld et al., 1997, 2001).

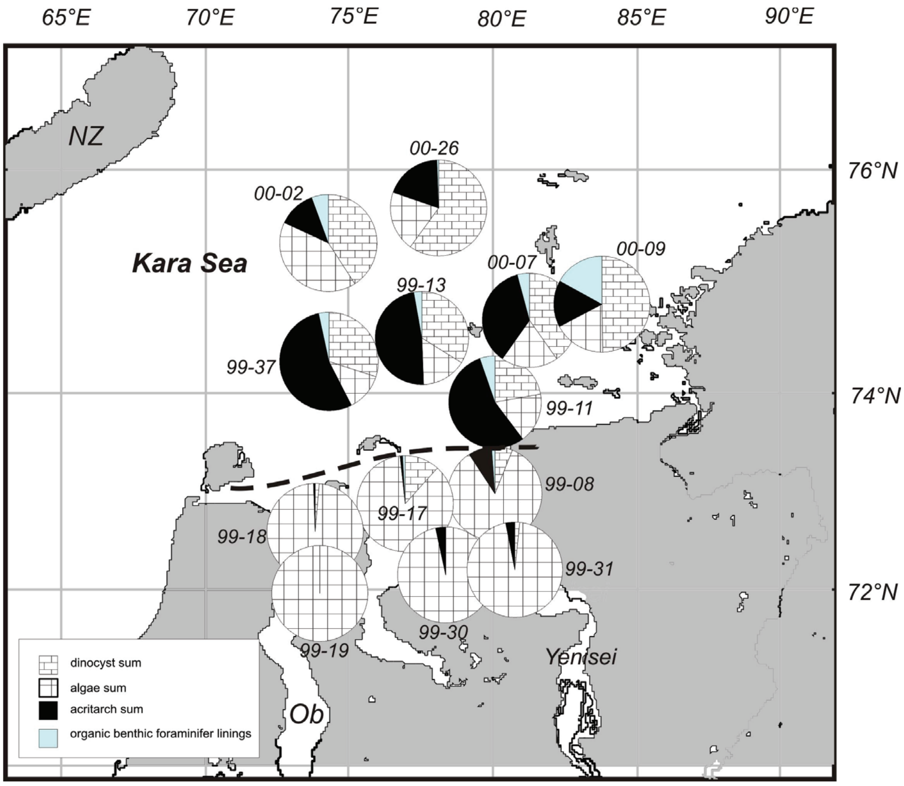


Fig. 5-3: Composition of aquatic palynomorph assemblages in surface sediment samples in the southern and central part of the Kara Sea. Three transects in N-S direction are shown. black=acritarchs; blue (grey)=dinocysts; shaded= chlorophycean algae; white=benthic foraminifer linings. The dashed line at ca. 73° N marks a notably change of all three transects dividing the assemblages in two distribution zones. For further details see text.

Aquatic palynomorph stratigraphy

Local aquatic palynomorph assemblage zones (LAPAZ) for the Yenisei estuary (Ye) are defined after visual inspection of the percentage diagram (Fig. 5-4), the concentration diagram (Fig. 5-5) and the accumulation rates of individual taxa (Fig. 5-6). Additionally, these zones are named after the respective predominant species for a first palaeocological characterization. In all samples, reworked dinocysts were found in small quantities with minimum values about 50 and maximum values of approximately 1000 cysts/g (at 310 cm and 720 cm core depth). In the upper part of the core, after ca. 4750 cal. BP, reworked dinocysts are more rare than downcore.

<u>*LAPAZ Ye-1 (780-665 cm; ca. 9400 - 8900 cal. BP; Pediastrum-Botryococcus zone)*</u> is characterized by the dominance of chlorophycean algae. *Pediastrum* spp. and *Botryococcus*

cf. *braunii* concentrations increase sharply from low levels at the bottom to high values up to ca. 6000 *Pediastrum* coenobia/g at 700 cm core depth. The accumulation rates of *Pediastrum* ssp. and *Botryococcus* cf. *braunii* increase sharply at the bottom to a maximum in the second part of this zone (Fig. 5-6). Dinocysts occur sporadically only in very small numbers (between 0 and 15 cysts per sample). In the bottom part between 780 and 740 cm, dinocysts are almost completely absent and do only slightly increase in the upper part. Benthic foraminifer linings are absent with the exception of two samples, where very low numbers were counted. *Halodinium* ssp. is also nearly absent. The acritarch *Radiosperma corbiferum* occurs sporadically in low values.

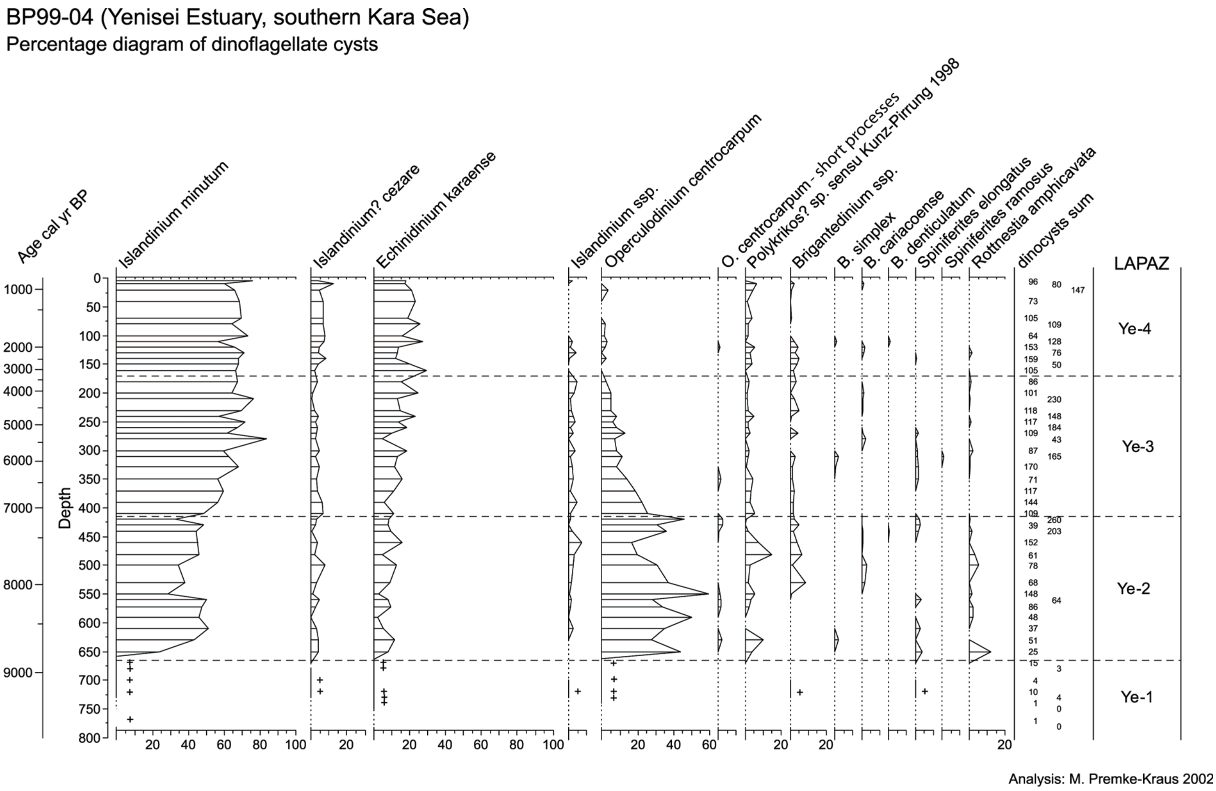


Fig. 5-4: Standard percentage diagram of the dinocyst assemblages of the sediment core BP99-04. The cross (+) indicates the presence of dinocysts within a sample, but of very low numbers (dinocyst sum < 25), which is not calculated as relative abundances.

<u>LAPAZ Ye-2 (665-415 cm; ca. 8900 – 7200 cal. BP; *Operculodinium centrocarpum zone*</u>) is marked by a gradual increase of dinocysts from very low numbers in zone Ye-1 up to higher values in the upper part of this zone (Fig. 5-4). The total dinocyst concentration increases up to a maximum of about approximately 4500 cysts/g at 430 cm depth (Fig. 5-5). The

Operculodinium centrocarpum percentage curve oscillates, but shows the highest percentage values up to 60% in this zone and the highest accumulation rate (Fig. 5-6). *Polykrikos*? sp. sensu Kunz-Pirrung 1998 has a small maximum in percentage values. *Rottnestia amphicavata* is most abundant in this zone. Also *Brigantedinium* spp.indet. shows a small maximum of the percentage curce in the second part of this zone. Benthic foraminifer linings were found more frequently. Also *Halodinium* ssp. occurs more frequently with two peaks in the uppermost part of this zone. Concentrations of the chlorophycean algae persist on higher level, but are less abundant above 550 cm. At the boundary between zone Ye-2 and Ye-3, the dinocyst accumulation rate shows a maximum peak of about > 250 cysts/cm^2/year. In contrast, accumulation rates *Pediastrum* ssp. and *Botryococcus* cf. *braunii* decreases gradually.

LAPAZ Ye-3 (415-170 cm; ca. 7200 – 3300 cal. BP; Islandinium minutum-Operculodinium centrocarpum zone) shows a sharp decrease of both percentage and concentration values of *O. centrocarpum*. *Islandinium minutum* is the most abundant dinoflagellate cyst and reaches the maximum in this zone. Percentage values of the dinocyst *Echinidium karaense* increase slightly. Concentration of benthic foraminifer linings has a maximum in this zone and the acritarch *Halodinium* spp. is most abundant in this zone. Concentrations of *Pediastrum* spp. and *Botryococcus* cf. *braunii* have a lower level as before. *Radiosperma corbiferum* assemblages show oscillations as before.

LAPAZ Ye-4 (170-0 cm; ca. 3300 – 600 cal. BP; Islandinium minutum-Echinidium karaense zone) is characterized by the lowest percentage and concentration values of *O. centrocarpum* cysts, whereas *O. centrocarpum* were not found continuously. The three protoperidinioid dinocysts *I. minutum*, *I. ? cezare* and *E. karaense* constantly predominate with >90%. However, the period between ca. 2000 and 1700 cal. BP shows a certain irregularity: The dinocyst species group *Islandinium minutum* and *Echinidium karaense* show a peak in concentration and accumulation rate (Fig. 5-6), the benthic foraminifer linings show only in accumulation rate a peak but not in concentration and the acritarchs have low values in concentration but a small peak in accumulation rate. Afterwards, concentration and accumulation rates of both cysts decrease sharp up to the uppermost sample. The concentrations and accumulation rates of the chlorophycean algae show minimum values downcore. Concentration and accumulation rate of benthic foraminifer linings decreases notably. Abundances of *Halodinium* spp. and *Radiosperma corbiferum* are relatively less abundant in this zone.

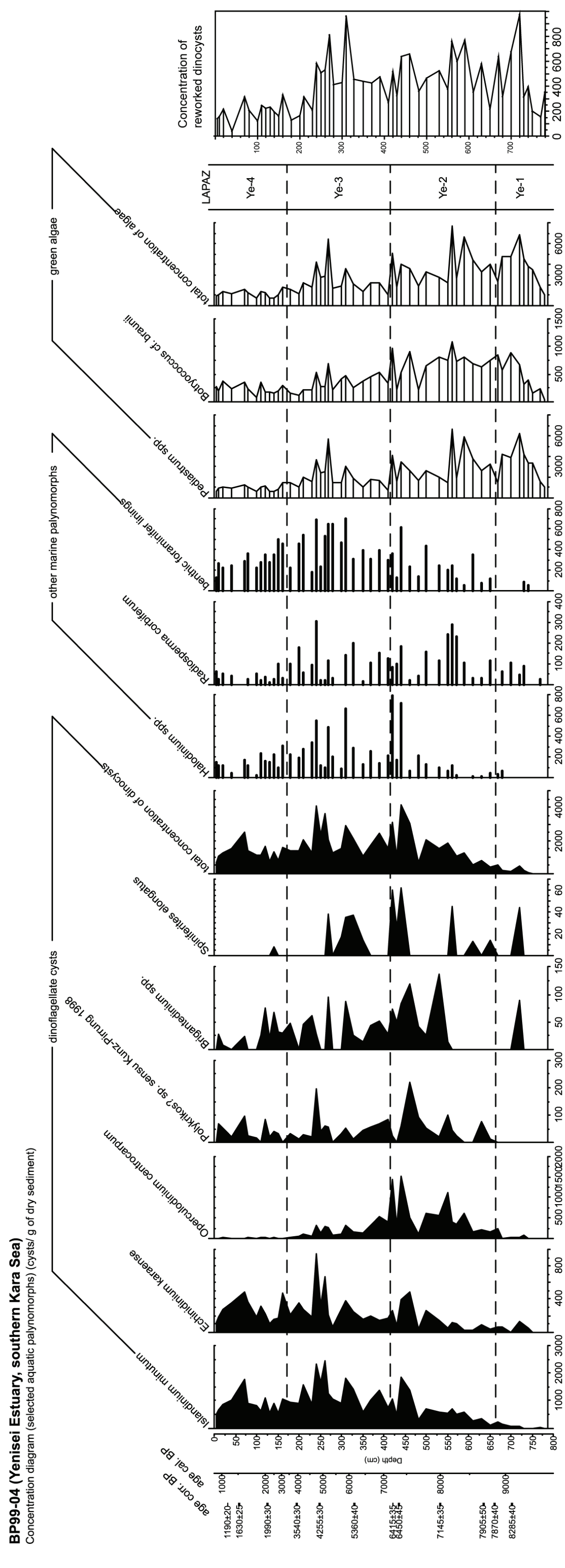


Fig. 5.5: Concentration diagram of selected aquatic palynomorphs of the sediment core BP99-04 and the concentration of reworked dinocysts (rightmost).

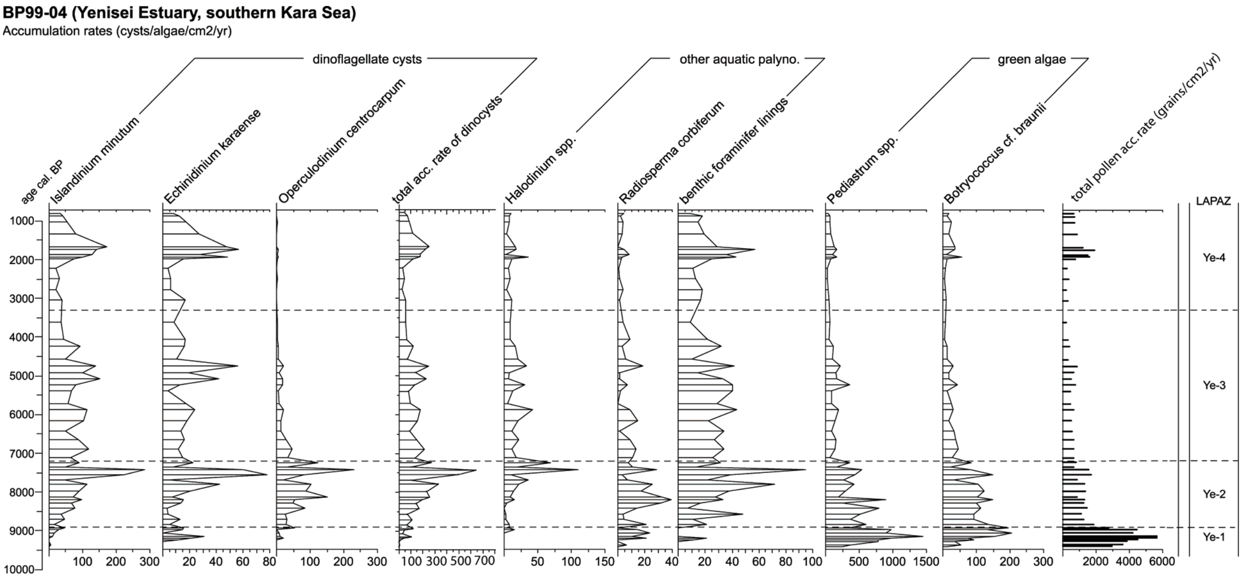


Fig. 5-6: Accumulation rate diagram of selected dinoflagellate cysts, chlorophycean algae, and acritarchs and total pollen accumulation rate versus calibrated ages.

Implications to the Holocene local inundation and hydrographic history

The earliest phase in the study area reflects a highly dynamic system and a rapid change of environmental conditions during approximately 500 years, which can be subdivided in two steps between 9400 and 8900 cal. BP: First phase (9400 – 9200 cal. BP): pre-transgressive system, fluvial environment. Second phase (9200 – 8900 cal. BP): fluvial system, growing influenced by sea-level rise, but not yet flooded by transgression. Since 8900 cal. BP (see LAPAZ Ye-2) it commenced the third phase: flooding by sea-level rise, river-water influenced marine/brackish neritic system. The chronology is slightly revised from Polyakova and Stein (2004), Dittmers et al. (in press) and Kraus et al. (2003). Therein, the point of flooding is set at ca. 9200 cal. BP. In the following, we appoint the flooding time by transgression at ca. 8900 cal. BP.

LAPAZ Ye-1 (780-665 cm; ca. 9400 - 8900 cal. BP; Pediastrum-Botryococcus zone):

First phase (9400 – 9200 cal. BP, pre-transgressive system, fluvial environment): We interprete the maximum peak of chlorophycean algae accumulation rates as an indicator for high freshwater discharge at our site. Indicators for marine conditions are almost absent. We suggest the following scenario for the environmental conditions before 9200 cal. BP: The site was not yet flooded by sea-level rise, but was quite close to the Yenisei river mouth. Fluvial/estuarine conditions prevailed in the river channel, where our sediment core is located surrounded by a still largely exposed inner shelf areas at this time (Dittmers et al., 2003, in press).

<u>Second phase (9200 – 8900 cal. BP, fluvial system, increasingly influenced by sea-level rise, but not yet flooded by transgression)</u>: Between 9200 and ca. 8900 cal. BP, marine palynomorphs occur sporadically in very low abundances reflecting growing influence of sea water. However, we do not believe, that the very sparse occurrence of marine palynomorphs reflects an brackish environment due flooding by the sea. Rather, we interprete this as a fluvial/estuarine system with growing influence of marine waters. We explain the occurrence of sparse marine palynomorphs as a result of sporadic storm events. Such extreme weather events might have supplied marine water upstream.

Interestingly, also chlorophycean algae are relatively sparse in the first two of the bottom samples. There are two possible reasons for the almost absence of chlorophycean algae at the bottom. On one hand, low concentrations could result from the coarser sediment in the lowermost part (subunit Ib), which consists of silty sand in contrast to the fine-grained silty-clay to clayey silts above 740 cm. Matthiessen et al. (2000) showed a relationship between low concentrations of chlorophycean algae and coarse sediments. Therefore, the relatively low concentration of chlorophycean algae in the lowermost part results rather from a depositional than from a hydrographical change. On the other hand, the low concentrations could be a result from a decreased stream velocity, when environmental conditions changed from a fluvial to estuarine milieu. Hence, the sharp increase of chlorophycean algae reflects possibly an autochthonous algae bloom at the core site, whereas above in the second phase, hydrographical conditions changed towards a brackish/marine water regime with low productivity of autochthonous chlorophycean algae and an increase of allochthonous supply of chlorophycean algae. According to Jankovská and Komárek (2000), *Pediastrum* settlement is more characteristic for standing water body as for a water regime with higher velocity. Both explanations, depositional change and hydrographical change, are imaginable, and maybe both factors contributed to this situation.

LAPAZ Ye-2 (665-415 cm; ca. 8900 – 7200 cal. BP; Operculodinium centrocarpum zone):

Third phase (since 8900 cal. BP, flooding by sea-level rise, river-water influenced marine/brackish neritic system): Marine palynomorphs occur continuously and with increasing concentration. We interprete the transition of the zones Ye-1 to Ye-2 as the time of the flooding. The water depth is still low, so that marine water inflow is mixed with river water to a decreasing degree. The decline of chlorophycean algae concentration is interpreted as a consequence of the growing distance to the river mouth and less as a change in productivity and not directly as a change in salinity. The salinity increase of the more or less unstratified water column is rather reflected in the occurrence of marine dinocysts and of the benthic foraminifer linings.

The relatively high percentage values of *Operculodinium centrocarpum* reflect increased sea-surface temperature (SST) prevailing until 3200 cal. BP, but decreasing after 7200 cal. BP. This is supported by the higher abundances of the dinocysts *Rottnestia amphicavata* and *Spiniferites elongatus*, which indicate warmer temperatures (Levac et al., 2001, de Vernal et al., 2001). Remarkably, percentage values of *O. centrocarpum* show a distinct depression between ca. 8100 and 7500 cal. BP within this phase of higher SST, whereas *Polykrikos*? sp. sensu Kunz-Pirrung (1998) have a maximum peak. We suggest, that this constellation could possibly indicate a short-term cold event.

The total dinocyst concentration peak (> 4000 cysts/g of dry sediment) at 450 cm depth (ca. 7550 cal. BP) is related to a TOC peak and also a clay peak at 460 cm depth (Kraus et al., 2003). Thus, we infer, that the peak in dinocyst concentration reflect rather depositional change as high productivity.

LAPAZ Ye-3 (415-170 cm; ca. 7200 – 3300 cal. BP; Islandinium minutum-Operculodinium centrocarpum zone):

After the establishment of the brackish/marine Yenisei estuary, the concentration curves of chlorophycean algae and of the total dinocyst concentration are quite even, which is interpreted as enhanced stratification of the upper water column and relatively stabile surface water masses after 7200 cal. BP. We suppose, that this enhanced stratification could reflect a stabilization of the sea-level rise, which took place at ca. 7000 cal. BP (Fairbanks, 1989; Zeeberg et al., 2001). However, the aquatic palynomorph assemblages in this zone does not reflect the high stand of sea-level rise, which is dated at about 5000 cal. BP (Bauch et al., 2001; Stein et al., 2003; Dittmers, 2006).

The gradual decrease of *O. centrocarpum* indicate the decrease of SST after 7200 cal. BP and a stepwise cooling at ca. 6400 cal. BP and at 4500 cal. BP (Fig. 5-4, Fig. 5-6). The predominance of *I. minutum* and *Echinidium karaense* is related to an increase of sea-ice formation.

LAPAZ Ye-4 (170-0 cm; ca. 3300 – 600 cal. BP; Islandinium minutum-Echinidium karaense zone):

After the stepwise decline of *O. centrocarpum* in zone Ye-3, this indicator for warmer SST was found only sporadically in low numbers since 3300 cal. BP. We interprete this as the onset of sea-surface conditions comparable as today with polar water masses and extensive sea-ice cover. The modern dinocyst assemblages in surface sediment samples in the Kara Sea show similar small values < 10 %. This is supported by the absence of *Rottnestia amphicavata* and *Spiniferites elongatus*, which prefer also warmer SST.

Discussion

Sea-level rise chronology and early Holocene hydrographic changes

The time of the flooding at ca. 8900 cal. BP is similar to that inferred from diatoms at about 9300 cal. BP for the same core (Polyakova and Stein, 2004). This small difference results probably from our rather conservative interpretation of the lowermost zone LAPAZ Ye-1, where only very low dinocysts were found. Therefore, we do not interprete these assemblages as the time of flooding, but still at the transition of zone Ye-1 to Ye-2 at ca 8900 cal. BP, when dinocyst assemblages increased notably.

Highest accumulation rates of total sediment and total organic carbon were ascertained near the base of the sediment core at > 8800 cal. BP (Stein et al., 2003; Stein et al., 2004a;

Stein and Fahl, 2004b), and are explained by enhanced river discharge, corresponding well with our interpretation of the pre-transgressive, fluvial/estuarine phase. At the same time, high magnetic susceptibility values were interpreted as result of the final decay of the LGM ice-sheet on the Putoran Mountains (Stein et al., 2003, 2004; Stein and Fahl, 2004). Dittmers et al. (in press) proposed, that this large amount of melt water including high suspended matter, could be mainly responsible for the larger fluvial dimension of the Yenisei river channel and larger paleo-river discharge, besides other promotive factors such as degradation of permafrost soils. By means of detailed examinations of numerous seismic profiles in conjunction with geophysical properties of well-dated sediment cores, Dittmers et al. (in press) described the pre-transgressive phase on the Kara Sea shelf as a high dynamic system of incised meandering rivers.

Polyak et al. (2003) reconstructed a dramatic decline in river discharge prior ca. 9000 cal. BP very close to our site, which support our interpretation that fluvial to estuarine conditions with high river discharge still prevailed between 9400 to 9200 cal BP.

Polyakova and Stein (2004) calculated an increase in sea-surface salinity (SSS) up to 11-13 from 7500 to 6000 cal. BP, which is in good agreement with our results reflecting also enhanced SSS since 7200 cal. BP, when the upper water column became more stratified due to a certain stabilization of the sea-level rise and a weaker signal of the freshwater plume at our site. Interestingly, also bottom water reflects a diminishing influence of fluvial water between 7000 and 5000 cal. BP showed by a $\delta^{18}O$ shift of stable isotopes in ostracods (Simstich et al., 2004, 2005).

According to our results, the thermal optimum was between 8900 to 7200 cal. BP at our site. However, the concentrations of *O. centrocarpum* are quite high in relation to *I. minutum* in this zone. Thus, we believe, that SST may have been increased considerably. We do not know, when higher SST occurred in the Kara Sea region in areas, which were earlier reached by sea-level rise, because the late glacial-Holocene transition is not recorded at our site. But higher SST probably persisted already before 9200 cal. BP, because assemblages of *O. centrocarpum* are high by initiation of the percentage curve. This assumption is supported by an overall productivity rise of foraminifera west of Yamal Peninsula between 10,000 to 9500 ^{14}C ka (ca. 11,100-10,500 cal. BP) which is related to a significant amelioration of sea-surface conditions (Polyak et al., 2002).

We suppose, that the higher SST emerged probably by an enhanced and/or warmer inflow of Atlantic water, which flew presumably along the (paleo-) river channels from the continental slope through the troughs to the south (Lubinski et al., 2001). We suggest, that local effects (e.g. higher summer insolation) (Berger, 1978) could have strengthened the warmer SST as a positive feedback.

In general, our results about the sea-level chronology and the hydrographical changes in the Kara Sea agree very well with the well-studied Laptev Sea shelf (e.g. Bauch et al., 2001). Bauch et al. (2001) and Klyuvitkina and Bauch (2006) specified the flooding of the inner Laptev Sea shelf at ca. 8900 cal. BP for the 31 m and 32 m isobath, respectively, which coincides very well with our interpretation. A diatom-inferred salinity record by Bauch and Polyakova (2003) from the inner Laptev Sea shelf revealed the most drastic hydrographical change at ca. 7400 cal. BP, and afterwards marine diatoms became the dominant group. This correspond also well with our reconstruction of elevated SSS in connection with an enhanced stratification of the upper water column.

Moreover, the Laptev Sea data set also reflects a pronounced change in palynomorph assemblages between 10,700 and 9200 cal BP and from 8900 to 7400 cal. BP which is related to an enhanced and/or warmer influx of Atlantic water to the outer and inner shelf areas (Polyakova et al., 2005; Klyuvitkina and Bauch, 2006). The latter one shows an exact coincidence to our suggested thermal optimum of SSTin the Kara Sea.

Mid to late Holocene cooling

The gradual decrease of *O. centrocarpum* and of *S. elongatus* and *R. amphicava* is interpreted as a stepwise cooling of SST at ca. 7200, 6400 and 4500 cal. BP and the onset of sea-surface conditions comparable to today since ca. 3300 cal. BP. The most pronounced change in our dinocyst record occurs at the transition of zones Ye-2 to Ye-3 at ca. 7200 cal. BP, when percentage values and accumulation rates of *O. centrocarpum* decrease after the HTM. Between 7200 and 6400 cal. BP, higher SST prevailed, but notably reduced. Since ca. 6400 cal. BP to 3300 cal. BP, dinocyst assemblages reflect only a weak signal of elevated SST. This decline of these dinocysts which are associated with warmer sea-surface conditions and conversely the increase of protoperidinioid dinocysts (*Islandinium minutum, Echinidium karaense*) which are considered as indicators for polar and cold surface water masses, is interpreted as the increase of seasonal sea-ice formation and a longer duration of sea-ice cover since 6400 cal. BP. At ca. 3300 cal. BP, modern conditions of sea-ice formation and cover was reached.

The period between ca. 2000 and 1700 cal. BP shows a certain short-term event, since the accumulation rates of *I. minutum, E. karaense* and further of benthic formaninifer linings show a distinct peak. A relationship to a lithological change can be excluded, because in such a case, also the other palynomorphs should show the same feature, but only the total pollen accumulation rate show a small peak. Moreover, the granulometric composition and organic matter content do not show such a change (Kraus et al., 2003; Stein et al., 2003). We have no firm explanation for this situation. According to Stein et al. (2003), this period is characterized by an increase of accumulation rates of total sediments and TOC, which is related to a temporary increase of river discharge within a phase of a reduced freshwater run-off over the last ca. 2000 years (Stein et al., 2003, 2004; Stein and Fahl, 2004; see also Polyakova and Stein, 2004). This interpretation is supported by Polyak et al. (2002), who showed an increase of river discharge only at the surface of the water column at ca. 1500 years ago.

Maybe, this could have caused also the fluctuations in our palynomorph assemblages, although the chlorophycean algae do not show a peak, which we would expect at first sight. However, enhanced freshwater discharge does not induce obligatory a peak in chlorophycean algae assemblages. Indeed, enhanced freshwater supply is commonly related to an increase of chlorophycean algae assemblages (as part of the suspended matter flux) (e.g. Matthiessen et al., 2000). But we do not know surely, if this is also valid for phases with extensive sea-ice formation and long duration of sea-ice cover, respectively. We would suppose, that under such conditions the productivity might be reduced due to harsh conditions and the decrease of light under the sea-ice. However, this is only an assumption, which has to be verified by in-situ experiments.

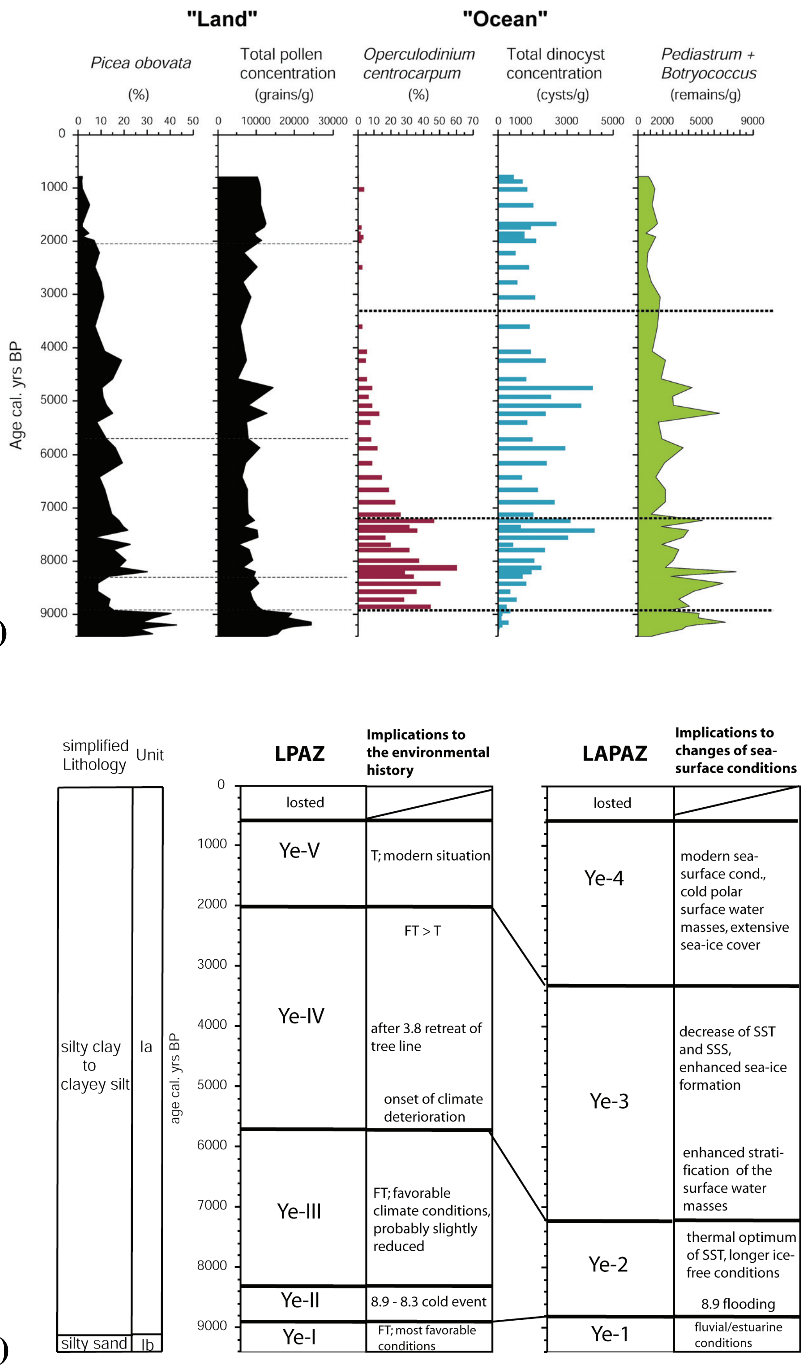
The low assemblages of reworked dinocysts after ca. 4750 cal. BP (see Fig. 5-5) correlate obviously with an increase of the fine-grained silty-clay granulometric composition in the upper part of the sediment core (Stein et al., 2003, 2004; Stein and Fahl, 2004). This coincides also with a decrease of reworked pollen since ca. 5700 cal. BP (Kraus et al., 2003). Reworked palynomorphs are commonly related to enhanced resuspension and coastal erosion (e.g. Mudie, 1992; Matthiessen et al., 2000). However, the decrease of reworked dinocysts in relation to the increase of fine-grained particles, reflect rather the enhancement of the effectiveness of the marginal filter (Lisitzin, 1995) in connection with a movement of the depocenter to our site at this time. This feature was discussed by Polyakova and Stein (2004) in order to explain a speficial discrepancy between sedimentological findings (Stein et al., 2003, 2004; Stein and Fahl, 2004) and the decrease of salinity indicated by diatom assemblages.

Land-sea-correlation

Studies about direct onshore-offshore correlation (analysis of terrestrial and aquatic palynomorphs at the same sediment core) in shelf seas in the northern high-latitudes are rare (e.g. Levac and de Vernal, 1997; Levac, 2001; Levac, 2003) in opposite to a more frequent application in the south Atlantic (e.g. Hooghiemstra et al., 2006). Land-sea correlations could provide similarities or / and discrepancies in the development of the atmospheric, vegetational and the oceanic evolution (e.g. Mudie and McCarthy, 2006).

Generally, the comparison between our results with continental paleorecords in the coastal area and adjacent hinterland is for several reasons hampered: First of all, pollen records in the adjacent land area are relatively scarce compared to the strong gradients along the climatic and vegetational latitudinal and longitudinal changes. Secondly, a regional stratigraphical framework fails since paleorecords are often discontinuous or/and have a low temporal resolution. Third, the comparison of marine pollen records with land-based pollen stratigraphies is hampered due to the unequal pollen deposition producing different pollen spectra. For further details see Kraus et al., (2003).

In the following, three levels of land-sea correlations are discussed: (1) A short summary of the results published by Kraus et al. (2003), which compared the marine pollen record BP99-04 with continental pollen records, (2) a comparison of herein presented LAPAZ with continental pollen records, and (3) a direct correlation between LAPAZ and LPAZ (published in Kraus et al., 2003, see also Fig. 5-7a and b).



a)

b)

Fig. 5-7a) and b): (a) Direct land-sea correlation of selected indicator palynomorphs in core BP99-04 versus calibrated ages. The pollen curve of *Picea* and the total pollen concentration is arranged according to Kraus et al. (2003). **(b)** Land-sea correlation showing pollen and aquatic palynomorph zones with respective paleoenvironmental interpretation. Abbreviations: FT: Forest Tundra; T: Tundra

<u>Comparison LPAZ with land-based pollen records</u>

The marine pollen record indicates that the earliest Holocene (ca. 9400 – 9000 cal. BP) was the warmest phase and marks probably the termination of the Holocene thermal maximum

(HTM) in the coastal Kara Sea. This is similar to previous studies from continental site, which separate between the warmest time for sites from modern coastal and islands between 10,000 to 9000 yr BP (ca. 11,100 – 9800 cal. BP) and for non-coastal areas between 6000 to 4500 yr BP (ca. 6850 – 5000 cal. BP) (e.g. Velichko et al., 1997; Serebryanny and Malyasova, 1998; Serebryanny et al., 1998; Andreev and Klimanov, 2000; Andreev et al., 2001). After 9000 cal BP, except for a short-term cold event from 8900 to 8300 cal. BP, favorable climate conditions prevailed until ca. 5700 cal. BP. Pollen records in the adjacent land area reflect overall at ca. 4500 conv. yr BP (ca. 5200 cal. BP), which represents the Atlantic-Subboreal boundary according to Khotinskiy (1984) and Velichko et al. (2002), a strong climate deterioration (e.g. Velichko et al., 1997; Serebryanny et al., 1998; Andreev et al., 2003), which is comparable to our results. At ca. 3800 cal. BP, our marine pollen record reflects the onset of forest retreat and the displacement of the forest tundra by Arctic tundra communities in the coastal area, which correspond to the overall evidence from radiocarbon-dated macrofossils indicating a retreat to its modern position between 4000 and 3000 yr BP (ca. 4500 – 3250 cal. BP) (e.g. MacDonald et al., 2000). At ca. 2000 cal BP. the establishment of climate conditions comparable to today is reflected in our marine pollen record, which also agrees with the continental pollen sites, which show between ca. 3500 to 2800 –2500 conv. yr BP modern conditions (e.g. Velichko et al., 1997; Andreev et al., 2003).

<u>Comparison LAPAZ with land-based records</u>

Land-based paleorecords of northern Siberia reflect usually a “Holocene climate optimum” for the modern coastal and islands between 10,000 to 9000 yr BP (ca. 11,100 – 9800 cal. BP) and for non-coastal areas between 6000 to 4500 yr BP (ca. 6850 – 5000 cal. BP) (references cited above). Andreev et al. (2001) reconstructed a longer duration of “Holocene climate optimum” prevailing until 4500 yr BP based on a pollen record from the Yugorski Peninsula (southeastern Kara Sea) except for short-term events. A further comparable study based on pollen data from the Levinson-Lessing Lake and Taymyr Lake (northern Taymyr Peninsula) revealed a temperature maximum between 10,000 and 5500 yr BP (ca. 11,100 – 6200 cal. BP) (Andreev et al., 2003). Our dinocyst-based marine thermal optimum (8900 – 7200 cal. BP) does only partly overlap with these reconstructions. Obviously, the early atmospherical temperature maximum occurred during the period of sea-level lowstand probably forced by local effects. Accompanied with the flooding of the shelf, the presumably continental character of the climate changed gradually to wetter and milder conditions with subsequent decreasing summer temperatures (e.g. Vasil'chuk et al., 2001; Hantemirov and Shiyatov, 2002; Andreev et al., 2001; 2003). Thus, our dinocyst-inferred thermal optimum lags the atmospherical / continental thermal optimum compared with the land areas close to the past coastline. However, it preceded the thermal optimum in the hinterland, which occurred time-transgressive. Following scenario is suggested: The flooding of the shelf lead to a cooling in the coastal environment, but drove the delayed thermal optimum in the hinterland due to warmer SST from 8900 to 7200 cal. BP. Otherwise, the ending of the marine thermal optimum preceded the climate deterioration in the hinterland. It allows the estimation that the Holocene environmental evolution in the southern Kara Sea was influenced by local effects, which revealed in relation to the other circumarctic evolution some features on regional scale. However, large-scale trends such as the climate development in southern Siberia, the global atmospherical circulation patterns and the variations of the North Atlantic Current may have been mainly affected by environmental changes in the Kara Sea during the Holocene.

Direct correlation between LAPAZ with LPAZ

The direct correlation of the local aquatic palynomorph assemblages with the local pollen assemblages' support our interpretation mentioned above. The pollen assemblages does not reflect such clear termination onshore, but rather a gradual climate deterioration with steps at 5700 and 3800 cal. BP. A short-term fluctuation in the marine pollen record between 8900 and 8300 cal. BP was interpreted as cold event. In opposite, aquatic palynomorphs do not show at this time such a deterioration, but slightly later, between ca. 8100 and 7500 cal. BP within a phase of high SST a cooler period (Fig. 5-7 a and b). Aquatic palynomorph assemblages indicate the establishment of modern sea-surface conditions at ca. 3300 cal. BP, which preceded the onset of climate conditions similar as today reflected by the marine pollen record at ca. 2000 cal. BP.

Conclusions

A well-dated sediment core located in the outer Yenisei estuary in the southeastern Kara Sea was analysed for aquatic palynomorph content and reveals a dynamic land-ocean system since

9400 cal. BP. The main results of the hydrographical changes and of the land-ocean correlation are summarised below:

- The relative sea-level rise chronology derived from the aquatic palynomorph record proceded in two steps between 9400 and 8900 cal. BP: The first phase (9400 – 9200 cal. BP) reflects a pre-transgressive system and fluvial to estuarine conditions. The second phase (9200 – 8900 cal. BP) reveals a fluvial/estuarine system with growing influence of the sea-level rise, but the site was not yet flooded by transgression.
- The flooding by sea-level rise occurred at ca. 8900 cal. BP and a river water influenced marine/brackish neritic system was established.
- Highest sea-surface temperatures (SST) prevailed until 7200 cal. BP. During this period of marine thermal optimum, sea-ice cover was considerable reduced compared with today.
- Since 7200 cal. BP, the upper water column became more stratified which is related to the decrease of freshwater plume influence at the core site due to the southward retreat of Yenisei river mouth and with the stabilization of the global sea-level rise.
- A stepwise cooling of SST took place at ca. 6400 and at ca. 4500 cal. BP. After 3300 cal. BP, sea-surface conditions like today took place with enhanced seasonal sea-ice formation and longer duration of sea-ice cover.
- In general, our sea-level chronology and the detected hydrographical changes show a good concordance to the results of studies, which emanated from the SIRRO project, in particular to the studies on the same sediment core (e.g. Polyakova and Stein, 2004) and to other paleoceanographic studies in the Kara Sea.
- Our results is in good agreement to the well-known Laptev Sea sea-level chronology and hydrographic evolution
- The land-sea correlation revealed a time-transgressive progress of onshore and offshore evolution of the Holocene thermal optimum. We speculate that the flooding of the shelf led to a cooling in the coastal environment, but drove the delayed thermal optimum in the hinterland due to warmer SST from 8900 to 7200 cal. BP.
- The time-transgressive progress of onshore and offshore evolution permits the tentatively conclusion, that Holocene environmental evolution in the southern Kara Sea was strongly influenced by local effects and the land-sea correlation revealed some features on regional scale.

- The direct correlation of the local aquatic palynomorph assemblages with the local pollen assemblages supports our interpretation of a time-transgressive development.

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6. Holocene climate and hydrographic history of the southern Kara Sea shelf (Arctic Ocean): Evidence from marine palynological records

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Abstract

Based on two marine nearshore pollen and aquatic palynomorph records (organic-walled dinoflagellate cysts, chlorophycean algae, acritarchs, organic-walled benthic foraminifer linings) from the estuaries of Ob and Yenisei in the southern Kara Sea (Arctic Ocean, Russia), we reconstruct the regional environmental history since 9600 cal. BP. Fluvial conditions prevailed until the proto-Ob and Yenisei estuaries were inundated by the sea-level rise at ca. 8900 and 8500-8200 cal. BP, respectively. A dinocyst-inferred marine thermal optimum occurred between 8900 and 7400-7200 cal. BP and is connected to an enhanced and/or warmer Atlantic water inflow from the north to the inner Kara Sea. At ca. 7400 cal. BP, aquatic palynomorphs reflect an enhanced stratification of the upper water column and sea-surface temperatures and sea-surface salinities decrease gradually. At ca. 3500-3300 cal. BP, modern sea-surface conditions were established with enhanced sea-ice formation and longer duration of sea-ice cover. The Ob record shows the last ca. 1000 years distinct oscillations, which are related to human impact.

The pollen assemblages reflect the most favorable conditions in the coastal area up to ca. 8900 cal. BP, possibly up to 7400 cal. BP. The late Holocene climate deterioration is well reflected in both marine pollen records. At ca. 3800 cal. BP, the retreat of boreal forest is indicated by the increase of herbaceous pollen such *Artemisia* and Poaceae pollen and the decrease of *Picea* pollen. Since ca. 2000 cal. BP, modern conditions were established.

The comparison with other circumarctic paleorecords revealed that the Karas Sea thermal optimum of surface water masses, which had a duration of about 2500 years, corresponds rather with the thermal optimum in the northern Norwegian Sea than with that in the eastern Barents Sea, but was probably slightly delayed. In opposite, the Laptev Sea shows a very similar evolution.

Keywords: Holocene, Kara Sea, Arctic Ocean, marine palynology, pollen, dinoflagellate cysts, paleoceanography, Holocene thermal optimum

Introduction

The Late Quaternary evolution of the Arctic was highly influenced by the build and decay of ice sheets and related mass exchanges, which led to high sea level fluctuations (e.g. Lambeck and Chappell, 2001). In view of the late glacial-Holocene transition and our present interglacial, the different patterns of ice sheets and their decay contributed considerably to large spatio-temporal differences in the evolution of the sub-arctic and circumarctic environment (e.g. Kaufman et al., 2004; de Vernal et al., 2006; Kaplan and Wolfe, 2006). The comparison of paleodata with model simulations showed a broad agreement in general (e.g. CAPE, 2001). However, they outcropped also gaps of knowledges with regard to the Eurasian Arctic part despite of an increasing number of paleorecords on land and sea:

In the northern Kara Sea, paleoceanographic studies at the continental margin considered the Late Weichselian deglaciation history of the St. Anna Trough (Polyak et al., 1997), changes of the Atlantic water inflow since the late glacial (Hald et al. 1999; Lubinski et al., 2001) and changes in supply of terrigenous organic carbon during the late Quaternary (Stein et al., 2001; Boucsein et al., 2002).

In the southern (inner) Kara Sea, a profound data set were collected by several ship-based expeditions by Russian research vessels, which proceeded in two phases, one from 1984 to 1993 (e.g. Lisitzin and Vinogradov, 1995) and the second phase between 1997 and 2002 within the joint German-Russian research project “Siberian River Run-Off” (SIRRO) (Matthiessen and Stepanets, 1997; Stein and Stepanets, 2000, 2001, 2002; Schoster and

Levitan, 2003). Numerous related studies in the context of environmental changes showed that the paleoceanography and hydrography was strongly affected by changes in river discharge and sea-level rise (e.g. Lisitzin, 1995; Levitan et al., 1995; Matthiessen et al., 1999; Polyak et al., 2000, 2002; Stein et al., 2002; Kuptsov and Lisitzin, 2003; Lisitzin and Kuptsov, 2003; Polyak et al., 2003; Stein et al., 2003, Dittmers et al., 2003; Simstich et al., 2004; Stein et al., 2004; Stein and Fahl, 2004, Polyakova and Stein, 2004, Simstich et al., 2005, Fahl and Stein, 2007, Dittmers et al., in press, Premke-Kraus et al., subm.).

On the adjacent land area, the climate and vegetation development is based mainly on few pollen-based paleorecords (e.g. Velichko et al., 1997; Serebryanny and Malyasova, 1998; Serebryanny et al., 1998; Andreev and Klimanov, 2000; Andreev et al., 2001, 2003), which revealed an early Holocene thermal optimum in the coastal and island area of the Kara sea. Also a marine pollen record from the southeastern Kara Sea supported these findings (Kraus et al., 2003). The Late Quaternary ice sheet history was profound reviewed by Svendsen et al. (2004).

Polyakova and Stein (2004) and Premke-Kraus et al. (subm.) provided a very detailed reconstruction of Holocene hydrographical changes based on a multiproxy, well-dated sediment core from the southeastern Kara Sea. Based on this data, a land-sea correlation revealed time-transgressive evolution of the Holocene thermal optimum onshore and offshore, which is still tentatively (Premke-Kraus et al., subm.).

It is of crucial importance to reveal a better spatial resolution about the paleoenvironmental changes in the southern Kara sea realm. Therefore, a second well-dated sediment core from the Ob estuary was additionally investigated by its content of terrestrial (pollen and spores) and aquatic palynomorphs (dinoflagellate cysts, acritarchs, chlorophycean algae, and organic benthic foraminifer linings). The main purpose of this study was

- to reconstruct hydrographical changes in the southern Kara Sea on a regional scale,
- to compare the inundation history in the Ob estuary with that in the Yenisei estuary to yield a regional rate of sea-level rise,
- to compare the environmental evolution of both Siberian rivers,
- to compare the paleoenvironmental evolution in the southern Kara Sea with other circumarctic shelf seas, in order to assess the timing of the Holocene thermal optimum both on land and in the sea,

Physiographic settings

The Kara Sea is one of the Eurasian Arctic shelf seas, which encompass more than 50 % of the Arctic Ocean (Jakobsson et al., 2003). It is considered as a nucleation area for past and present changes in the Arctic system, because of the two large Siberian rivers Yenisei and Ob, with discharge more than 50 % of the whole circumarctic freshwater into the Kara Sea and the Arctic Ocean (e.g. Holmes et al., 2002) and therefore influence profoundly the Arctic circulation (e.g. Aagaard and Carmack, 1989). The freshwater carries large amounts of suspended matter, which is deposited mainly due to flocculation/aggregation processes in the estuaries (Lisitzin, 1995; Stein et al., 2003, 2004).

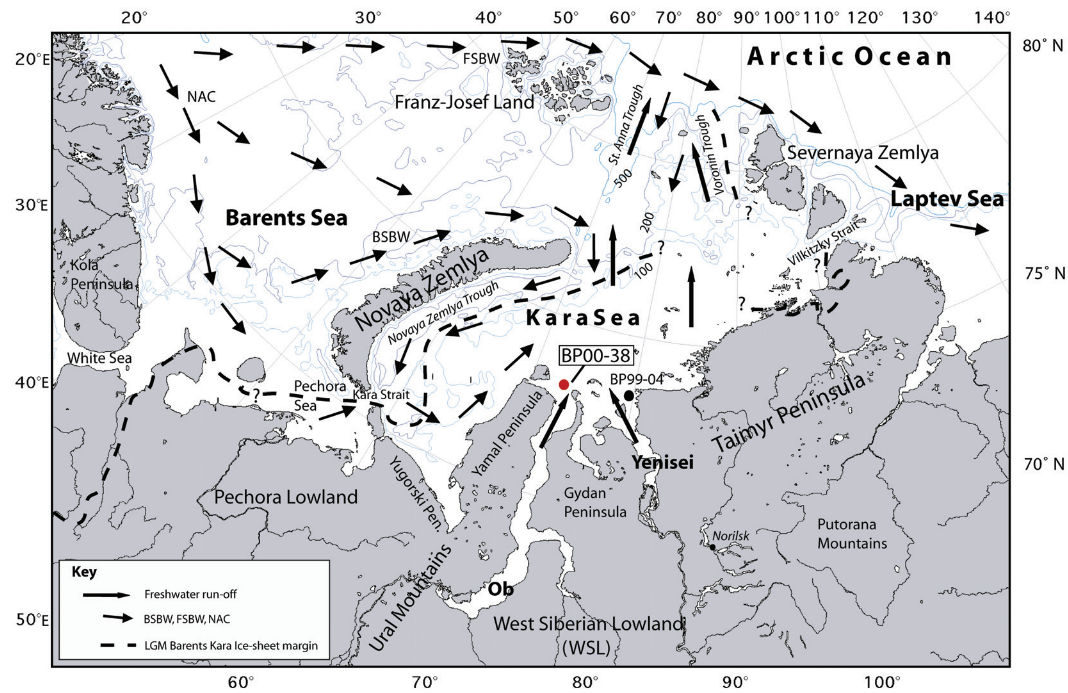

Fig. 6-1: Overview map of the Kara Sea, of the adjacent shelf seas and the hinterland and location of the sediment core BP00-38 (Ob estuary) and mentioned BP99-04 (outer Yenisei estuary). Also shown are schematically main surface water currents and the limit of the LGM Barents Kara Ice sheet (according to Svendsen et al., 2004). Abbreviations: NAC=North Atlantic Current, BSBW=Barents Sea Branch Water, FSBW= Fram Strait Branch water (according to Lubisnki et al., 2001).

The inner Kara Sea is a relatively shallow semi-closed basin (Fig. 6-1). That is connected with the Barents Sea through the Kara Strait and with the Laptev Sea through the Vilkitsky

Strait. The northern open part is directly connected to the Arctic Ocean through the deep (>400 m) St. Anna and Voronin troughs.

The cyclonic surface circulation transports relatively warmer water with the eastern Novaya Zemlya current into the southwestern Kara Sea (Fig. 1) and turns, after mixing with riverine waters to the northeast as Yamal current (Pavlov and Pfirman, 1995; McClimans et al., 2000; Pivovarov et al. 2003). However, this circulation pattern is strongly influenced by seasonality and wind (Harms and Karcher, 1999).

The water column in the southern Kara Sea is strongly stratified due to the poor vertical mixing of the thin surface layer of river water with the underlying cold and saline bottom sea-water superimposed by seasonal and interannual variability in both salinity and temperature.

A further important aspect is sea ice formation in the inner river estuaries (e.g. Reimnitz et al., 1994; Nürnberg et al., 1994, Divine et al., 2004), which cover more than eight months per year the Kara Sea and influencing the hydrographic structure (Divine et al., 2004).

Material and Methods

Sediment sampling and processing

The gravity core BP00-38 was collected during the Kara Sea expedition of RV "Akademik Boris Petrov" in 2000 (Stein and Stepanets, 2001). Core BP00-38 is located in the outer Ob estuary on the slope of a filled river channel (73º11,8'N, 73º14,3'E, 20 (20)m water depth) (Stein, 2001).

Lithology and sediment composition

Generally, the sediments consist of fine-grained bioturbated muds (Fig. 6-2A). Dittmers et al. (2003) and Dittmers (2006) differentiated two lithological subunits, the bioturbated upper subunit Ia and the partly laminated subunit Ib at the base of this core based on visual description, magnetic susceptibility and acoustic profiles (see also Dittmers et al., 2003). This description is refined by examination of granulometric data (Fig. 6-2B). Sandy clayey silt to clayey silty sand in the lowermost part from 652 cm to 630 cm core depth (Unit Ib) are replaced by very homogeneous bioturbated silty clay to clayey silt muds (Unit Ia). Thin silty sandy layers occur at 132 and 352 cm core depth. Bivalves (mainly *Portlandia* sp.) occur down to 533 cm core depth.

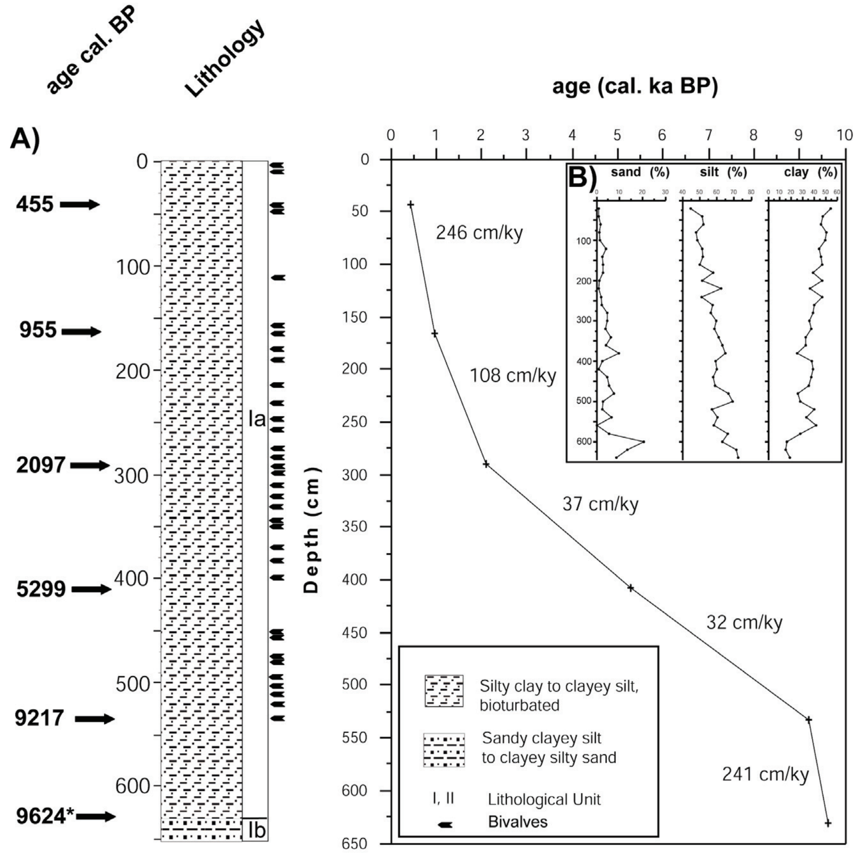

Fig. 6-2 A) and B): Lithology, lithological units (according to Dittmers, 2006), linear sedimentation rates (A) and granulometric composition (B) of sediment core BP00-38 (modified after Stein, 2001). Asterisk: The lowermost age is revealed by magnetostratigraphical correlation with parallel core BP01-83.

Radiocarbon chronology

The chronostratigraphy of core BP00-38 is based on five AMS ^{14}C radiocarbon dates, which were performed on *Portlandia* sp. shells (Table 6-1) at the Leibniz-Laboratory for Radiometric Dating and Stable Isotope Research (University Kiel) in Germany. A correction of 440 yr has been applied to account for the air-sea reservoir difference (Mangerud and Gulliksen, 1975) and radiocarbon ages were calibrated using the software CALIB 4.3 (Stuiver et al., 1998). Additionally, the magnetostratigraphical correlation with a parallel core BP01-83 (73º11,8'N, 73º14,4'E, 24m water depth) (Stein and Stepanets, 2002; Fig. 6-3), allowed to assess an age of ca. 9600 cal. BP for 631 cm core depth. A bivalve (*Portlandia arctica*) from 660 cm core depth was dated at 9624 cal. BP. The basal age of BP00-38 was defined by extrapolation of the linear sedimentation rate (SR) downcore. Our calculated basal age of ca. 9600 cal. BP is slightly younger than the estimated age of about 9800 cal. BP for the boundary of subunits Ia and Ib based on correlation with core BP00-26/4 according to Stein et al. (2003).

Tab. 6-1: AMS ^{14}C datings of sediment core BP00-38 performed on shells of *Portlandia* sp. Reservoir correction after Mangerud and Gulliksen (1975). The date at 631.0 was retrieved by the magnetostratigraphical correlation with the parallel core BP01-83 (Stein and Stepanets, 2002).

Core depth (cm)	^{14}C Age (BP)	Reservoir corrected age (-440 yr BP)	Calendar age (cal yrs BP)	Laboratory numbers
43.5	845±25	405±25	455	KIA-14052
166.5	1465±30	1025±30	955	KIA-14053
290.0	2490±30	2050±30	2097	KIA-14054
408.0	5005±35	4565±35	5299	KIA-14055
533.0	8830±70	8390±70	9217	KIA-14056
631.0	9060+45/-40	8620+45/-40	9624	KIA-19568*

Unfortunately, the data about the water depths of BP00-38 (20 m water depth, Stein and Stepanets, 2001) and BP01-83 (24 m, Stein and Stepanets, 2002) are not the same, although they were retrieved at the same position. The calculation of sea-level rise rates based on the comparison of the time of flooding of two sediment cores in the inner Kara Sea (see below) provides the most realistic scenario, if a water depth of 24 m is taken for the calculation, and allows the conclusion that the exact water depth of BP00-38 is most likely 24 m.

Relatively high SR prevailed during the early Holocene (241 cm/ky) and during the late Holocene ranging between 108 to almost 250 cm/ky. For the time interval between 9200 to 2100 cal., the SR is lower ranging between 32 and 37 cm/ky differing to the relatively high SR in the Yenisei core (Stein et al., 2003).

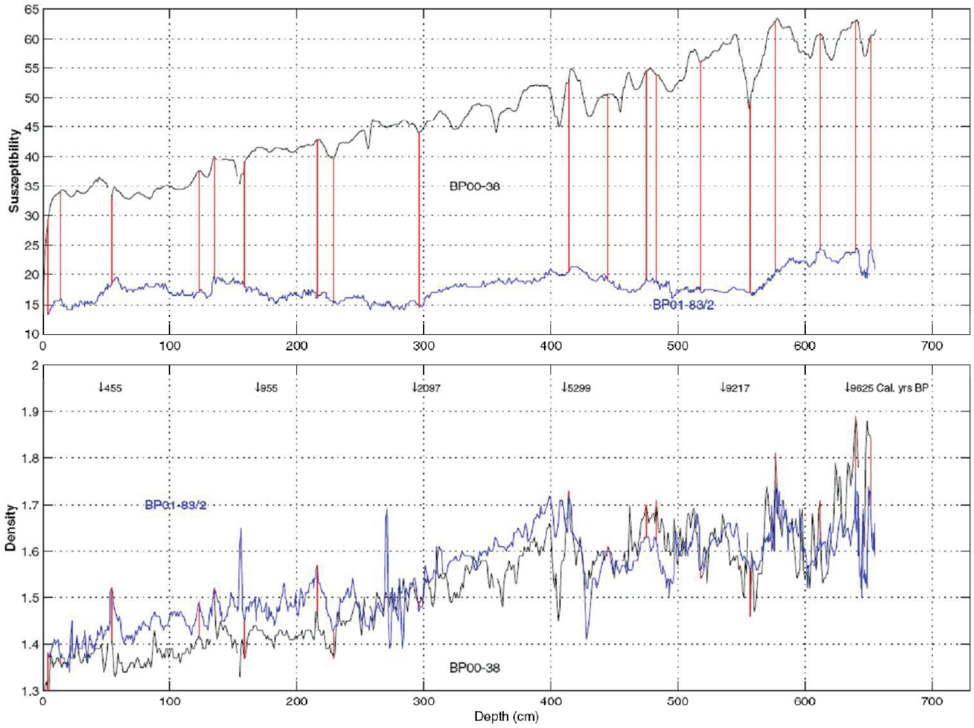

Fig. 6-3: Magnetostratigraphical correlation between the parallel cores BP00-38 and BP01-83 (for further details see text).

Palynological analysis

Totally, 47 samples were taken for palynological investigations at intervals of about 10-20 cm resulting in submillennial scale resolution depending on the sampling interval and sedimentation rates. Standard palynological processing methods were applied (Rochon et al., 1999; Kraus et al., 2003). Because of abundant coarse particulate organic matter such as plant remains, samples were sieved at 120µm mesh sizes, additionally. Terrestrial (pollen and spores) and aquatic palynomorphs (dinocysts, freshwater algae, acritarchs) as well as organic linings of benthic foraminifers were counted. No acetolysis was carried out. For calculation and construction of the pollen and dinocysts diagrams, the TILIA, TILIAGRAPH and TGVIEW software was used (Grimm, 1991, 2004). The complete data set can be retrieved from the PANGAEA information system at the Alfred Wegener Institute for Polar and Marine Research at Bremerhaven in Germany (http://www.pangaea.de).

Local pollen assemblage zones (LPAZ) and the local aquatic palynomorph assemblage zones (LAPAZ) were defined after visual inspection of the percentage diagram with consideration of the concentration diagrams.

Pollen and spores

All pollen grains were counted and determined because of the absence of other complete marine pollen diagrams in the Kara Sea except for the Yenisei pollen record (see Kraus et al., 2003). The first marine pollen diagram from the Kara Sea (Kulikov and Khitrova, 1982) presents only selected pollen from an undated sediment core. Another marine pollen record is from adjacent Laptev Sea (Naidina and Bauch, 2001; Naidina, 2006). Thus, little is known about the application of offshore pollen records along the Eurasian continental margin. We will test the suitability of this proxy for Holocene climate evolution and vegetation history in the coastal area of the Siberian Arctic.

Generally, Siberian Arctic pollen records are underrepresented by *Larix* pollen as an indicator for tree-line shifts (e.g. Clayden et al., 1996). It is of major relevance to distinguish *Betula* pollen in order to separate dwarf birch (*B. nana*) and tree birch (*B. pendula* and *B. pubescens*), but commonly determination are hindered by the wide range of size and morphological variability. Thus, the classical ratio of non-arboreal (NAP) and boreal (AP) pollen is of limited use as indicator for the tree-line history. In contrast to the Yenisei pollen record (see Kraus et al., 2003), it was attempted to differentiate *Betula* pollen in tree birches (*Betula* Section Nanae) and dwarf birches (*B.* Section Nanae) allowing a better interpretation of the vegetation development.

Onshore pollen records in this area are scarce due to discontinuous sequences (permafrost processes), relatively low temporal resolution as consequence of the low bioproductivity and limited radiocarbon dates. Offshore and nearshore pollen records, respectively, could fill this gap, because generally it is assumed that nearshore pollen records may reflect accurately the vegetation pattern in the adjacent coastal hinterland (e.g. Mudie and McCarthy, 1994).

However, core locations have to be chosen carefully with respect to the pollen source, the depositional environment (e.g. tidal, resuspension and redeposition and preservation), and the respective atmospherical circulation patterns (e.g. offshore winds). For further details on factors influencing nearshore pollen records see Kraus et al. (2003).

In our case, we decided to choose a sediment core in front of the Ob estuary, because of the high sedimentation rates and a absolute chronological framework on the basis of numerous bivalves, which are suitable for radiocarbon dating.

The nomenclature and taxonomy of pollen and spores follows basically Moore et al. (1991) with reference to Bobrov et al. (1983). The determination of the Betulaceae and Corylaceae pollen complex (includes Betula, Corylus, Alnus pollen) is difficult under the light microscope (for further details see e.g. Blackmore et al., 2003). Nevertheless, the differentiation between tree Betula pollen (includes mainly *B. pendula, B. pubenscens*) and dwarf *Betula* pollen (*B. nana*) is in view of the vegetation zone shifts in the Arctic tundra essential. In contrast to Kraus et al., (2003), herein, it was attempt to differentiate between *Betula* Section Albae (includes mainly pollen of *Betula pubescens* (Ehrh.) and *B. pendula* (Roth.)) and *Betula* Section Nanae (includes mainly pollen of *B. nana)*, which is the most common practice in palynological research for Eurasian birches (Peteet et al., 1998). All uncertain pollen of Betulaceae/Corylaceae pollen complex was assigned to *Betula* spec. *Pinus* pollen was distinguished between *Pinus* Diploxylon type (mainly pollen of *P. sylvestris*) and *Pinus* Haploxylon type (mainly pollen of *P. sibirica* and further *P. pumila*) (see Kremenetski et al, 1998). This group is also in some cases problematical because of variation and preservation.

Several pollen types were not plotted in the pollen diagram, if their age was not clear (due to the preservation state, sometimes reworked and Quaternary pollen grains could not be distinguished certainly), or if they are very sporadic or determination was not clear (cf. *Cyperus*, cf. *Potamogeton*, cf. *Callitriche*, cf. *Valerianella*, *Limonium vulgare* type, *Drosera*, *Ilex* type, *Galium* type, *Myricaria* type, *Veronica*, type, *Illecebrum verticillatum/Herniaria*). Also reworked pollen and spores grains or such belonging to the group above were not plotted singular in the diagram (cf. *Taxodium*, cf. *Ostrya* type, *Juglans*, cf. *Cornus suecica* type, cf. *Nyssa*, *Carya*, *Cedrus*, cf. *Woodsia* type, *Pterocarya*). In general, pollen preservation was quite bad, expressed in a qualitative term, which is expressed in th erelatively high abundances of reworked pollen.

The pollen content of the samples was relatively low. It was attempted to count at least 500 pollen grains but this sum was not achieved in all samples. The total pollen sum, including *Pinus* pollen, is shown in Fig. 6-4A. The pollen record is shown by standard percentage diagram (Fig. 6-4 A and B) and selected pollen as concentration diagram (Fig. 6-5), respectively. Long-distance transported *Pinus* pollen were excluded from the pollen sum in order to express more distinct changes of the extralocal / regional pollen signal.

Dinocysts

Nomenclature and taxonomy of the dinocysts (dinocysts for brevity) follows basically Rochon et al. (1999) and Head et al. (2001) and Premke-Kraus et al (subm.) (Table 6-2). The determination of the other aquatic palynomorphs (chlorophycean algae, acritarchs, organic benthic foraminifer linings) follows Premke-Kraus et al. (subm.).

Tab. 6-2: Detected dinoflagellate cysts in the sediment core BP00-38 and their affiliation to the respective biological taxon.

Dinoflagellate cysts (paleontological name)	**Biological affinity (thecate name)**
Brigantedinium cariacoense Wall 1965 ex Lentin & Williams	*Protoperidinium avellana* (Meunier) Balech
Brigantedinium simplex Wall 1965 ex Lentin & Williams	*Protoperidinium conicoides* (Paulsen) Balech
Brigantedinium spp. Indet. Reid 1977	*Protoperidinium* spp.
Echinidinium karaense Head, Harland & Matthiessen, 2001	Protoperidiniaceae
Islandinium minutum Head, Harland & Matthiessen, 2001	Protoperidiniaceae
Islandinium spp. indet.	Protoperidiniaceae
Islandinium? cezare Head, Harland & Matthiessen, 2001	Protoperidiniaceae
Nematosphaeropsis labyrinthus (Ostenfeld 1903) Reid 1974	*Gonyaulax spinifera* (Claprede & Lachmann) Diesing 1866, nach Wall & Dale 1968
Operculodinium centrocarpum - short processes (sensu Wall & Dale 1966 - short processes, in Rochon et al., 1999)	*Protoceratium reticulatum* (Claparède & Lachmann) Bütschli
Operculodinium centrocarpum sensu Wall & Dale 1966	*Protoceratium reticulatum* (Claparède & Lachmann) Bütschli
Polykrikos? sp. sensu Kunz-Pirrung 1998	probably *Polykrikos schwartzii* (Bütschli)
Rottnestia amphicavata Dobell & Norris	Unknown, probably *Gonyaulax spinifera*
Spiniferites elongatus Reid 1974	*Gonyaulax elongata* (Reid) Ellegaard, Daugbjerg, Rochon & Lewis
Spiniferites spp. indet.	*Gonyaulax* cf. *spinifera* (Claparède & Lachmann) Diesing

Some samples contain ancient reworked dinocysts (mainly *Wetzeliella*), and they were distinguished form Quaternary dinocysts based on the diagenetic alteration of the membran. Their presence indicates erosion and redeposition of pre-Quaternary sediments.

In total, 11 dinocyst taxa were identified (Table 6-2). The calculation of the percentage diagram is based on the sum of all cysts counted (Fig. 6-6). Selected dinocysts were presented additionally as concentration (Fig. 6-7).

The dinocyst sum is low in the bottom samples and in the uppermost samples, respectively, but if possible, a minimum of 100 cysts per sample have been identified and counted under a Zeiss light microscope (Axioplan) using phase and differential interference contrasts at a magnification of 400x and 1000x.

The quantitative reconstruction of the sea-surface conditions by means of the dinocyst reference data base (de Vernal et al., 2001; de Vernal et al., 2005) did not provide useful data since the used modern hydrographic data set for the dinocyst reference data base does not cover adequately the inner and shallow shelf areas of the Kara Sea.

The modern distribution patterns of dinocyst and chlorophycean algae assemblages in surface sediments show a relationship to the respective hydrographic conditions (salinity, sea-ice distribution, influence of river water inflow) (e.g. Matthiessen, 1999; Kunz-Pirrung, 1999, 2001; Matthiessen et al., 2000; Matthiessen and Kraus, 2001). Notes to the modern distribution and their respective paleoenvironmental significance of the aquatic palynomorphs with regard to the Eurasian shelf seas is described in detail in Premke-Kraus et al., (subm.).

Due to the overall predominance of the protoperidinioid dinocyst *Islandinum minutum* and of related taxa, we assume that selective degradation plays a minor role, because this species are considered as extremely sensitive to degradation (Zonneveld et al., 1997).

Results

Pollen stratigraphy

Five LPAZ were defined and named after the core location (Ob). The pollen sum is composed of arboreal pollen (AP) and non-arboreal pollen (NAP). *Pinus* pollen types are dominantly over-represented and are not included in the pollen sum. Pteridophyta are also excluded. The pollen types are grouped as over-represented / long-distance transported pollen, trees and shrubs, herbs, dwarf shrubs and aquatics and finally spores. A classic grouping on an ecological basis is not conducted because of the absence of a local / extralocal pollen deposition in marine environments.

The *LPAZ Ob-I (640–535 cm; ca. 9600-9100 cal. BP)* is characterized by the predominance of AP at the base decreasing from 60% to <50% (see AP: NAP ratio). *Betula* tree pollen (*Betula* Section Albae) is common and pollen of dwarf birch (*B*. Section Nanae, includes mainly pollen of *Betula nana*) decreases from a notable peak at the base of this zone and has a second maximum in the upper part of this zone. Poaceae pollen, *Artemisia* and *Pinus* Diploxylon pollen show lowest values in this zone. *Equisetum* spores have a distinct maximum and spores of *Lycopodium* spec. are most abundant. Only few herb pollen occur in this zone. Total pollen concentrations have a maximum of >15,000 grains/g dry sediment at the base of this zone, decreasing upcores.

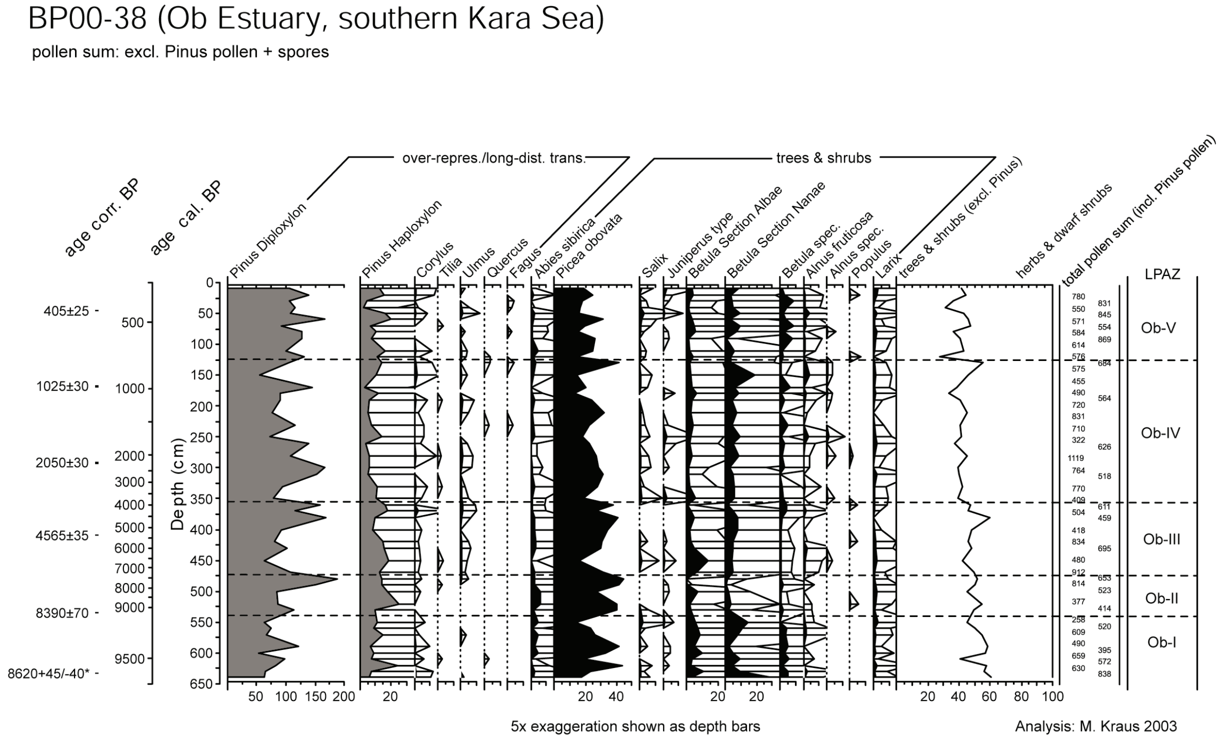

Fig. 6-4 A) and B): Pollen percentage diagram of Ob estuary sediment core BP00-38. A) Pollen percentage diagram of trees & shrubs, over-represented / long-distance transported pollen and AP/NAP ratio. Pollen of over represented / long-distance transported and spores are not included in the pollen sum. B) Continued pollen percentage diagram showing non-arboreal pollen and spores. On the right site is display the total pollen sum (including *Pinus* pollen). Asterisk: The lowermost age is revealed by magnetostratigraphical correlation with parallel core BP01-83 (see below).

<u>*LPAZ Ob-II (535-475 cm; ca. 9100-7400 cal. BP*</u>) is characterized by a distinct minimum of pollen of *Betula nana* and *Betula* tree pollen. *Salix* pollen is nearly absent in this zone. Long-distance transported bisaccate pollen of *Pinus* Diploxylon, *Pinus* Haploxylon and *Abies* pollen have a distinct maximum in this zone.

In <u>*LPAZ Ob-III (475-355 cm; ca. 7400-3800 cal. BP*</u>), pollen of *Betula* trees shows a maximum peak at the basis of this zone. Poaceae pollen and *Artemisia* pollen increase continuously. Numerous herbaceous pollen types occur for the first time in this zone. *Ulmus* pollen is more abundant. Concentration of Ericaceae pollen shows high values at the base of this zone, decreasing upcore. Total pollen concentration is still quite high.

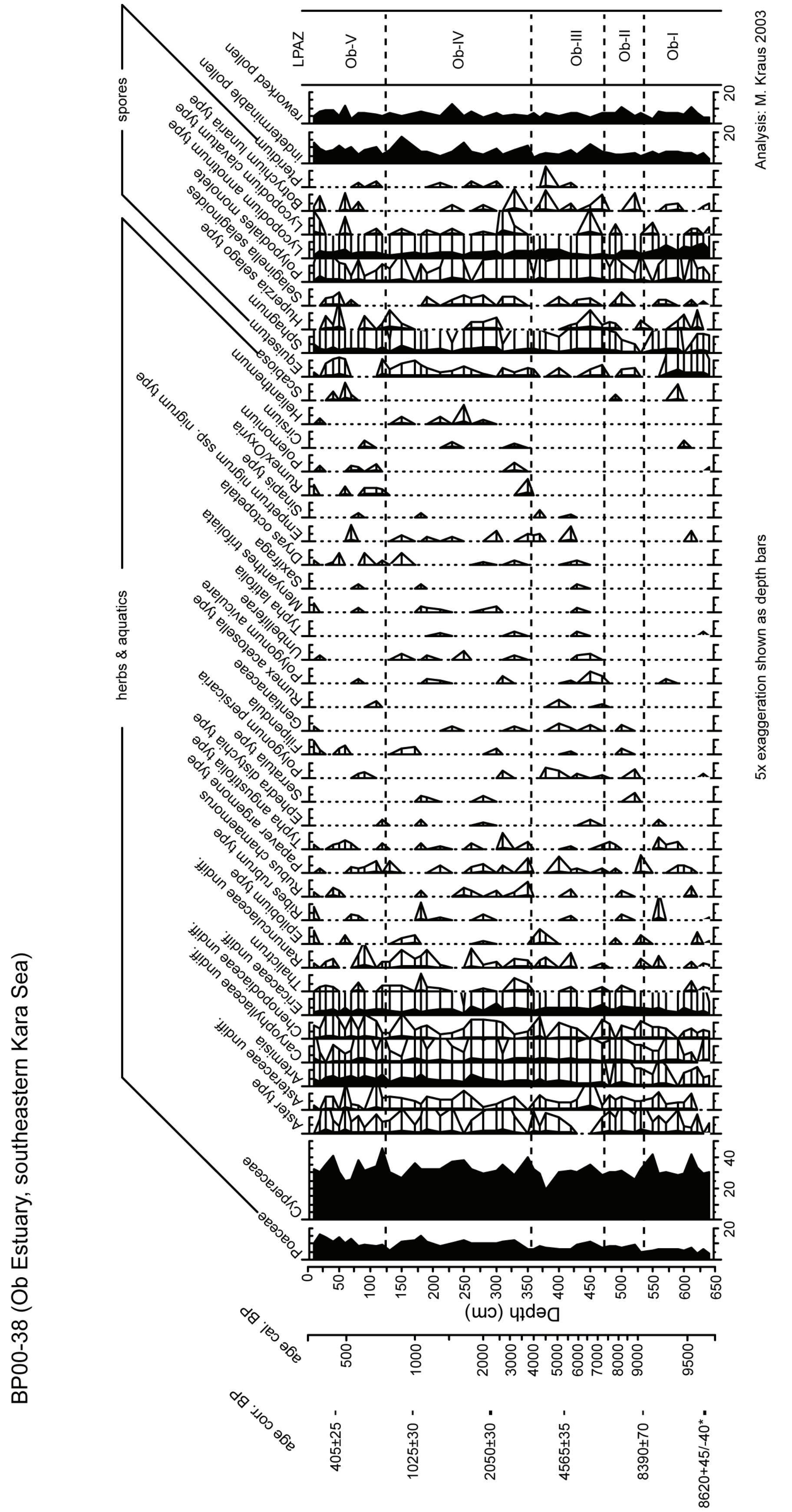

Fig. 6-4 B) continued

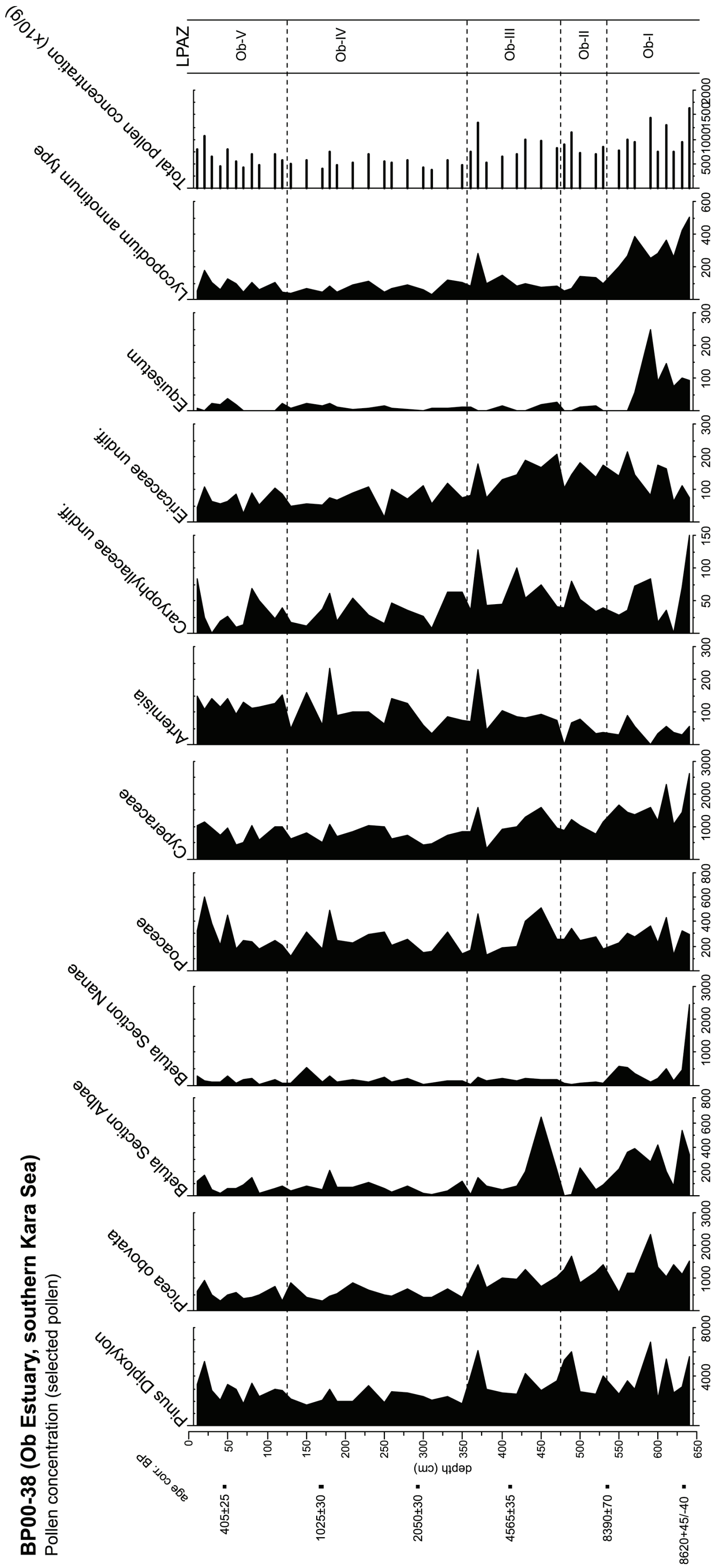

Fig. 6-5: Pollen concentration diagram of selected pollen and spores of Ob sediment core BP00-38. Asterisk: The lowermost age is revealed by magnetostratigraphical correlation with parallel core BP01-83 (for further details see text).

LPAZ Ob-IV (355-125 cm; ca. 3800-800 cal. BP) is marked by the minimum of total pollen concentrations. Pollen concentration of *Pinus* Diploxylon and *Picea* pollen shows lower values as in zone Ob-III.

LPAZ Ob-V (125-0 cm; ca. 800-recent time) is defined by an increase of the herb group caused mainly by noticeable amount of *Artemisia* pollen (see also concentration diagram) and increase of Poaceae pollen. Pollen of *Dryas octopetala, Polemonium, Rumex/Oxyria* are more frequent. *Picea* pollen percentages show lowest values. *Pinus* Diploxylon pollen is most dominant. The group of spores shows higher values in this zone. The total pollen concentration increases notably up to the uppermost part. NAP is predominant in the AP: NAP ratio.

Aquatic palynomorph stratigraphy

Five local aquatic palynomorph assemblage zones (LAPAZ) were defined after visual inspection of the dinocyst percentage diagram (Fig. 6-6) and of the aquatic palynomorphs concentration diagram. For a first paleoecological characterisation, LAPAZ are named after the respective predominant species.

LAPAZ Ob-1 (640-510cm; ca. 9600-8500 cal. BP; Operculodinium centrocarpum-Spiniferites elongatus zone) is characterized by very low dinocyst sums (<25 cysts) and very low dinocyst concentrations (<500 cysts/g). *O. centrocarpum* cysts and *S. elongatus* are the predominant cysts, but occur in very small percentages (Fig. 6-6). The acritarchs *Halodinium* spp. and *R. corbiferum* as well as benthic foraminifer linings occur sporadically in very small percentages. *Pediastrum* spp. shows a maximum in this zone and *B*. cf. *braunii* is also abundant.

LAPAZ Ob-2 (510-475 cm; ca. 8500-8200 to7400 cal. BP; O. centrocarpum zone) differ from zones Ob-1 and Ob-3 by a distinct peak of *O. centrocarpum* percentage values. Also *S. elongatus* and *Polykrikos*? sensu Kunz-Pirrung (1998) show a small peak in percentages.

Dinocyst sum and total dinocyst concentration is distinctly elevated in relation to zone Ob-1, but is still lower than in zone Ob-3.

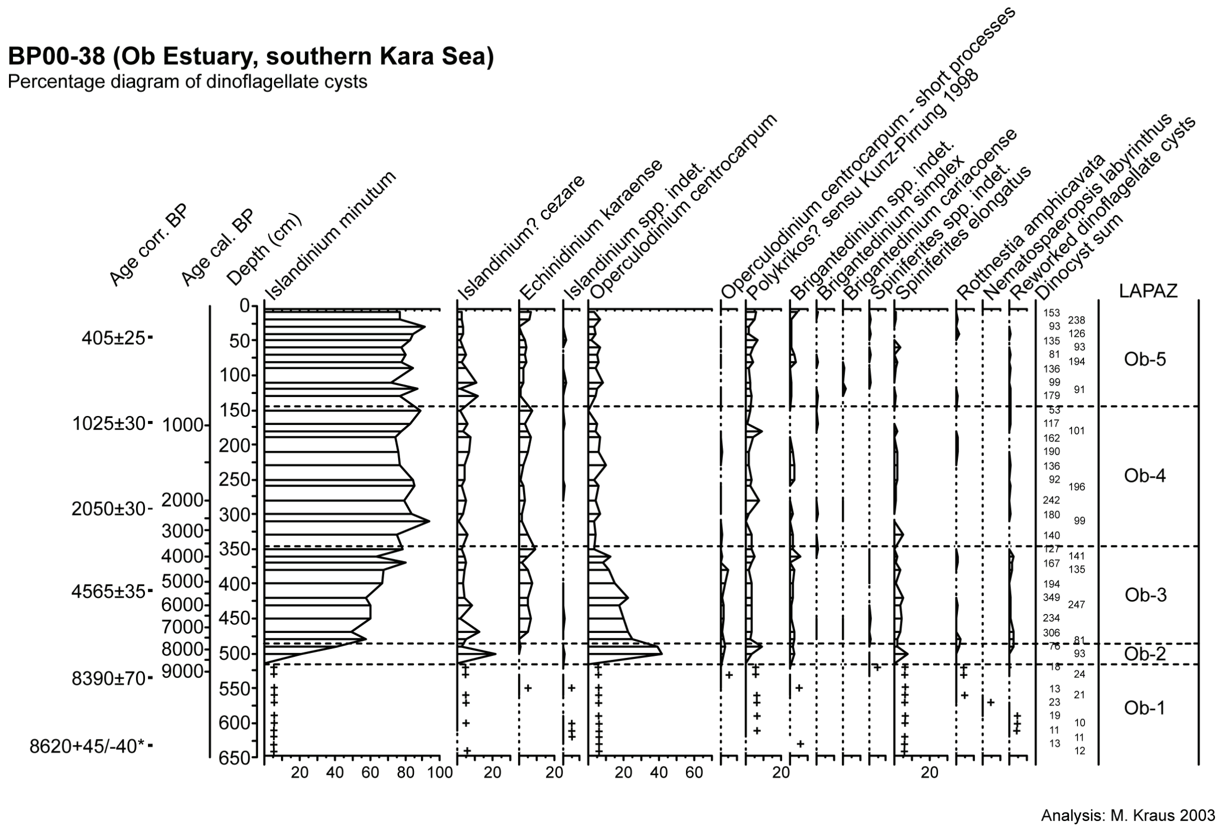

Fig. 6-6: Standard dinocyst percentage diagram of the Ob sediment core BP00-38. The cross (+) indicates the presence of dinocysts within a sample, but of very low numbers (dinocyst sum < 25), which is not suitabel for the calculation as relative abundances. Asterisk: The lowermost age is revealed by magnetostratigraphical correlation with parallel core BP01-83.

<u>LAPAZ Ob-3 (475-340cm; ca. 7400-3500 cal. BP; *Islandinium minutum-O. centrocarpum* zone)</u> is marked by rapidly increasing total dinocyst concentrations up to a maximum of approximately 4500 cysts/g at 450 and 370 cm core depth. Percentages of *I. minutum* increase sharply from the beginning of this zone. Percentage values of *O. centrocarpum* cysts are also still elevated, but decrease gradually. Cysts of *Echinidium karaense* occur for the first time in this zone. *S. elongatus* and *O. centrocarpum* – short processes show a small amount in percentage values. Linings of benthic foraminifers have a maximum in the upper part of this zone and *Halodinium* spp. occurs more abundant in this zone. Chlorophycean algae are furthermore abundant, but decrease slightly except for a peak in the uppermost part of this zone.

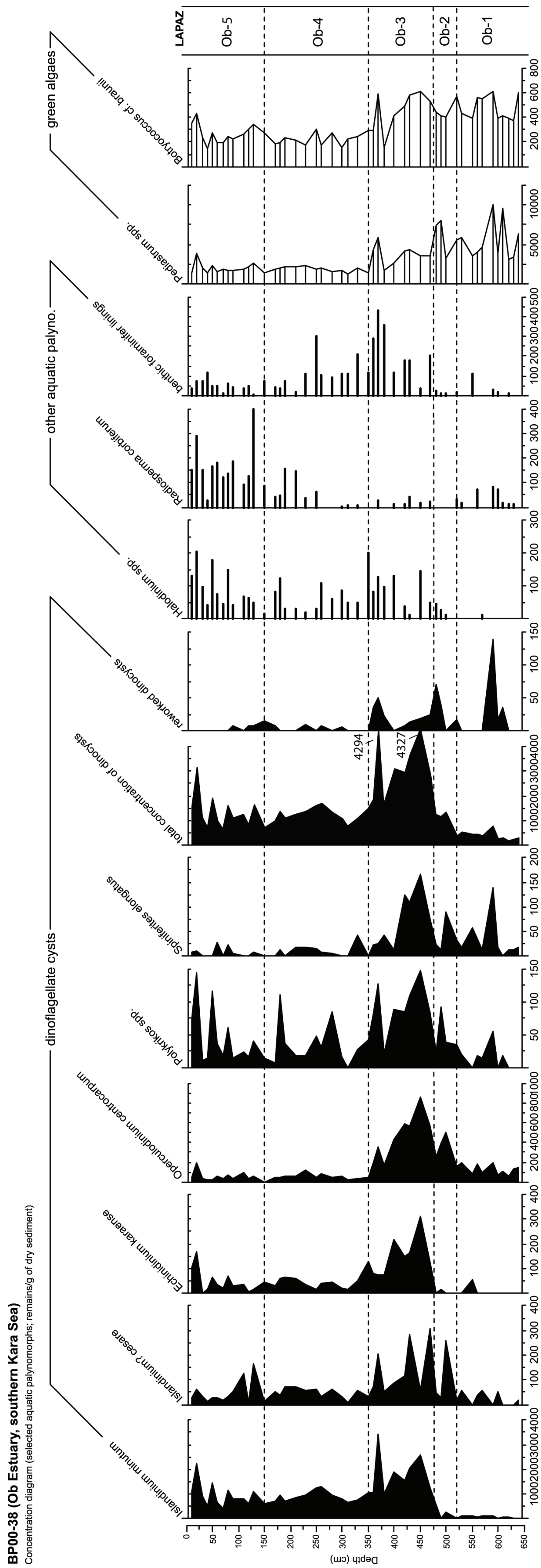

Fig. 6-7: Concentration diagram of selected aquatic palynomorphs of the Ob sediment core BP00-38.

LAPAZ Ob-4 (340-140cm; ca. 3500-850 cal. BP; Islandinium minutum zone I) is characterized by low and equal percentage values of *O. centrocarpum* cyst (± 10%) and high percentage values of *I. minutum* cyst, respectively. The total dinocyst concentration is relatively low and also equal, whereas there is a slight amount at 260 cm. Cysts of *Spiniferites elongatus* occurs only sporadically. In the second part of this zone, the acritarch *R. corbiferum* is for the first time more abundant. Chlorophycean algae show minimum values.

LAPAZ Ob-5 (140-0cm; ca. 850 cal. BP-recent; Islandinium minutum zone II) is characterized by the beginning of noticeable oscillations of the dinocyst and of *Halodinium* spp. concentration curve as well after a stable interval in the zone Ob-4. Percentage values of *I. minutum* remains at the same high level, whereas percentage values of *O. centrocarpum* cyst is persistently low. *R. corbiferum* show maximum values in this zone. Chlorophycean algae show slightly elevated concentration values.

Discussion

Implications of the marine Ob pollen record for the terrestrial environmental history of Kara Sea region

Principally, the analysis of the Ob nearshore pollen record revealed the predominance of fluvial and windblown pollen deposition. Long-distance transported pollen such as bisaccate pollen of *Pinus* Diploxylon and *P.* Haploxylon are selectively enriched due to their larger size and are therefore over-represented in the pollen diagram. Also *Picea obovata* pollen is selectively enriched due to the large grain size. However, we decided to include spruce pollen in the pollen sum, because it grew not far from the modern coastline during Holocene (e.g. Kremenetski et al., 1998; MacDonald et al., 2000).

Since the specific pollen deposition is influenced by increasing water depth due to sea-level rise, fluvial deposition (resuspension, redeposition) and selective pollen enrichment and transport, we consider the Ob pollen record as a mixed signal of large-scale atmospheric patterns and extralocal/regional vegetation shifts (tree-line movements, vegetation zone shifts, pattern of landscape history such as peatland and permafrost initiation in the hinterland). Moreover, as a consequence of the specific depositional conditions of nearshore pollen records, the direct correlation with onshore pollen records is difficult.

LPAZ Ob-I (640–535 cm; ca. 9600-9100 cal. BP):

The depositional conditions were still fluvial/estuarine in this zone indicated by the nearly absence of dinocysts (see below). The water depth was shallow and the inner Kara shelf close to the river channel was still exposed. Maximum of total pollen concentration reflects high influx of extralocal pollen source and higher productivity.

The pollen assemblages in this zone indicate that the most favorable climate conditions probably occurred in the coastal area at ca. 9600 cal. BP. A *Betula nana* pollen peak preceded higher percentage values of tree *Betula* pollen suggesting that the shrub tundra was displaced by the forest tundra with stronger contribution of tree birch and larch. The sparse herbaceous pollen suggests a relatively low species diversity at this time. Vasil'chuk et al. (2001) found remains of autochthonous tree birches that grew up to the Seyaha river area (nearly 70 °N). According to them, the first (tree birch-larch) forest arrived at about 9000 yrs BP (ca. 9800 cal. BP) in this area, which is in good agreement with our results. Forman et al. (2002) described a fossil *Betula* horizon in the Marresale area (western Yamal coast) dated between 8900 and 7800 yrs BP (ca. 9700 – 8800 cal. BP) and even a remain of tree birch from 110 km north of Kharasavey river mouth yielded an age of ca. 8800 yrs BP (ca. 9600 cal. BP).

We interprete the high amount of *Equisetum* spores (Fig. 6-4B and 6.5) as an indicator for paludification of minero-eutrophic peat-bogs. This was caused by the rising sea-level and groundwater table as consequence of permafrost degradation due to higher summer insolation (e.g. Berger, 1978). Possibly, this large-scale floodplains (see also Mann et al., 2002) are associated with minerotrophic fens as major producer of CH_4 that occurred between 11 to 8 ka in the circumarctic peatland complex (MacDonald et al., 2006).

Between ca. 9300 to 9100 cal. BP, at the end of the boreal period, the distinct peak of *Betula* Section *Nanae* pollen and the decrease of *Picea* pollen might indicate a short-lived cold event.

LPAZ Ob-II (535-475 cm; ca. 9100-7400 cal. BP):

Between 8500 and 8200 cal. BP, the site was flooded by the sea indicated by the increase of dinocyst concentration (see below). This substantial change in the ambient environment, which led also to the ablation of paleosoils, is reflected by the decrease of pollen concentration due to the diminishing of extralocal pollen rain.

Within the extralocal / regional pollen source, the minimum of *Betula nana* pollen indicates that dwarf birch might have been of minor importance at this time in the forest tundra, which was mainly composed of tree birch-spruce-larch dominated communities in the hinterland (e.g Andreev et al., 2001). According to Kremenetski et al. (1998), *Picea* grew in the region of Ob river mouth by around 9500 yrs BP (ca. 10,500 cal. BP). *Salix* pollen, which is an indicator for cold environment, is almost absent. Concentration of Ericaceae pollen is quite high. Peteet et al. (1998) suggest that increase of ericaceous species may indicate a transition from geogenous mires to ombrogenetic mires (terminology follows Joosten & Clarke, 2002) meaning formation of raised-bogs above the water table. Prentice et al. (1996) associates Ericaceae pollen with sufficient precipitation to establish forest vegetation.

Long-distance transported pollen such as pollen of *Pinus* Diploxylon, *P*. Haploxylon and *Abies sibirica* have a maximum in this zone suggesting a prevailing high pressure anticyclonic system. This tree has the narrowest ecological range among all Western Siberia trees and demands warmer and moist conditions (Blyakharchuk and Sulerzhitsky, 1999). Thus, winter must have been milder and wetter than before.

LPAZ Ob-III (475-355 cm; ca. 7400-3800 cal. BP):

The pollen spectrum reflects slightly more favorable climate conditions in the coastal Kara Sea region indicated by the weakly elevated *Betula nana* pollen percentages and by the first occurrence or increased abundances of some herbaceous pollen, indicating colder conditions and a more open landscape such as *Artemisia* pollen, *Saxifraga* and *Rumex acetosella* type. High pollen concentration of Caryophyllaceae indicates possibly also more unfavorable conditions, because this pollen type includes numerous genera (e.g. *Silene, Dianthus, Cerastium*), which are associated with grasslands, steppes and/or colder environments. In contrast, pollen concentration of Ericaeceae decreases notably.

The decrease of long-distance transported bisaccate pollen, indicating rather regional to large-scale changes, might be connected with pollen diagrams from farther south. Pollen records from Ulagan high-mountain plateau in the elevated Altai mountains in southern Siberia show also a decrease of *Abies* and *Picea* at about 7400 cal. BP (Blyakharchuk et al., 2004). In contrast, long-distance transported pollen of broad-leaved trees such as *Ulmus* indicates advanced climate amelioration in the hinterland.

Several herbaceous pollen types (*Empetrum nigrum* ssp. *nigrum* type, *Saxifraga*, *Dryas octopetala*, *Sinapis* type, *Menyanthes trifoliata*, Umbelliferae, *Polygonum aviculare*, *Rumex acetosella* type, *Artemisia*) appears for the first time or are more abundant indicating increased species richness and more structural diversity in the landscape. Pollen of *Menyanthes trifoliata* represents an indicator for shallow and nutrient water regime.

Our results are in good agreement with the dendrochronological results from the southern part of the Yamal peninsula from Hantemirov and Shiyatov (1999, 2002), which assessed a considerable decrease of the density of forest in Yamal at about 5400 BC (ca. 7350 cal. BP).

LPAZ Ob-IV (355-125 cm; 3800-800 cal. BP):

At 3800 cal. BP the most distinct change occurred reflecting a substantial cooling accompanied by the gradually retreat of the northern tree-line boundary. Most notable is the increase of non-arboreal pollen (*Poaceae, Artemisia)* and the decrease of *Picea* pollen. Total pollen concentration is the lowest. Also herbaceous pollen types such as *Helianthemum*, *Rubus chamaemorus* and Ranunculaceae pollen indicate a climate deterioration. This coincides well with other studies. According to MacDonald et al. (2000), decline of forest occurred between 4000 and 3000 yrs BP (ca. 4500 to 3250 cal. BP) and the treeline retreated to its modern position. Hantemirov and Shiyatov (2002) arrived at the same conclusion that, at approximately 1700 BC (ca. 3650 ca. BP), the polar boundary of open forests moved to the south. According to Pitkänen et al. (2002), *Picea* pollen decreased about 4000 yrs BP at Salym-Yugan mire area (60° N).

The colder temperature is probably responsible for a distinct change in the sedimentation rate (SR). After ca. 2100 cal. BP, SR increased from ca. 40 cm/ky to ca. 110 cm/ky (Fig. 6-2A). After ca. 1000 cal. BP, it is still higher (ca. 250 cm/ky). We explain it as follows: the climate was warmer and wetter, water was retained through the deepening of the active layer due to permafrost degradation (e.g. Forman et al., 2002), by the accumulation of peat and soil and finally by the spread out of boreal forest. Possibly, river discharge decreased and stream velocity slowed (Sidorchuk et al., 2001). With the beginning of climate deterioration, peatland formation slowed and came to rest in parts of the West Siberian Lowland (Blyakharchuk and Sulerzhitsky, 1999; Kremenetski et al., 2003). Continuous permafrost spread to the south (e.g. Kondratjeva et al., 1993). Hence, drainage of surface water increased to the Kara Sea and the main depocenter was displaced farther north in the Ob estuary probably due to a higher stream velocity.

LPAZ Ob-V (*125-0 cm; ca. 800 cal. BP-recent time*):

The pollen spectrum in this zone reflects the onward opening of landscape in the adjacent area accompanied with increased importance of grasses (Poaceae pollen) and cold resistant herbs in particular of *Artemisia* pollen.

Implications to the Holocene hydrographical history based on aquatic palynomorphs

LAPAZ Ob-1 (640-510cm; ca. 9600-8500 cal. BP; Operculodinium centrocarpum-Spiniferites elongatus zone):

We suppose that the very low numbers of dinocysts and of benthic foraminifer linings indicate an estuarine environment, which was sporadically flooded by marine waters caused by extreme weather events such as storms. The ingressing marine water masses were relatively warm as indicated by the relatively high concentration of *O. centrocarpum* in relation to *I. minutum* (Fig. 6-7). The maximum of chlorophycean algae concentration, the relatively high sedimentation rates, the lithological subunit Ib and the sandy clayey silt layer (Fig. 6-2B) are further indicators arguing for a pre-transgressive system strongly influenced by river discharge.

LAPAZ Ob-2 (510-475 cm; ca. 8500-8200 to 7400 cal. BP; O. centrocarpum zone):

The sharp increase of dinocyst assemblages indicates a flooding of our site between 8500 and 8200 cal. BP due to the global sea-level rise. The freshwater signal is still high indicated by continuing high concentrations of chlorophycean algae. The high *O. centrocarpum* percentages and the higher abundances of *S. elongatus* are related to the thermal maximum of sea-surface temperatures (SST) and a longer duration of seasonal ice-free conditions as today in this zone between 8500-8200 to 7400 cal. BP.

<u>*LAPAZ Ob-3 (475-340cm; ca. 7400-3500 cal. BP; Islandinium minutum-O. centrocarpum zone*</u>):

This zone reflects the establishment of full brackish/marine conditions in the surface water, preceded by a long period of high freshwater supply. The amount of dinocyst concentration might be primarily a result of the low sedimentation rates in this sequence (Fig. 6-2A) and does not directly show higher productivity. The gradual decrease of *O. centrocarpum* percentages after ca. 7400 cal. BP reflects the end of the thermal maximum and the approach of modern-like conditions at the transition of the zone Ob-3 to Ob-4 at ca. 3500 cal. BP.

Our results coincide well with salinity reconstructions based on diatom assemblages suggesting an increase at ca. 7500 cal. BP (Polyakova and Stein, 2004; Premke-Kraus et al., subm.). In a comparable record from the inner Laptev Sea shelf from the Yana paleo-valley (27 m water depth), modern-like environmental conditions were reached about 7400 cal. BP (Bauch et al., 2001, Polyakova et al., 2005) in agreement with our results from the inner Kara Sea shelf.

<u>*LAPAZ Ob-4 (340-140cm; ca. 3500-900 cal.BP; Islandinium minutum zone I*</u>):

After a highly dynamic environment with sharp and rapid changes, aquatic palynomorphs indicate uniform and stable surface water masses since 3500 cal. BP. Constant concentrations of chlorophycean algae and of dinocysts prevailed during this zone. We suppose that this onset of stable conditions reflect the sea-level high-stand at this site. The dinocyst percentages indicate the establishment of modern SST with enhanced seasonal sea-ice formation.

<u>*LAPAZ Ob-5 (140-0cm; ca. 900 cal. BP-recent; Islandinium minutum zone II*</u>):

The last approximately 1000 years are well resolved in a centennial scale due to the relatively high sedimentation rates of ca. 250 cm/ky (Fig. 6-2) and reveal noticeable oscillations. There are no substantial differences in comparison with the zone Ob-4, thus polar water masses and extensive sea-ice formation persisted, but the oscillations reflect a more changeable environment. We suggest that human land-use activities in the hinterland (e.g. settlements, agriculture, deforestiation) are the major reasons.

The high sedimentation rates, which increased already at ca. 2100 cal. BP from ca. 40cm/ky to 110 cm/ky, are interpreted as a change in the depositional environment. In conjunction with the climate deterioration related to forest retreat, aggradation of permafrost, lesser accumulation of peat, increase of sea-ice cover, possibly the water drained off superficial and river discharge increased. Consequently, the stream velocity increased slightly and the main depocenter moved slightly northward to our core location.

Palynostratigraphical correlation of the Ob and Yenisei estuary records

Kraus et al. (2003) and Premke-Kraus et al. (subm.) presented a first continuous Holocene pollen and aquatic palynomorph record from the Yenisei estuary in the Kara Sea. We discuss our new results from the Ob core and compare them with the Yenisei record in order to reveal general patterns and trends on a regional scale, which can used for a regional framework of the Holocene environmental evolution of the Kara Sea. Therefore, we connected the local palynostratigraphy of both records (Fig. 6-8) and summarize the results in Fig. 6-9. The distance between both cores is only about 170 km (Fig. 6-1).

The sedimentation history of both locations differ considerably, with exception of the early Holocene period, where both cores show high sedimentation rates (SR) up to ca. 8000 cal. BP (see Premke-Kraus et al., subm., this study Fig. 6-2). In the Yenisei record, the mid-Holocene has a high temporal resolution due to prevailing high SL. In contrast, SR in the Ob core is only approximately 30 cm/ky. Stein et al. (2003, 2004) explained changes in SR as a consequence of the southward retreat of the river mouths due to sea-level rise associated with a weakening of the so-called “marginal filter” (Lisitzin, 1995). This depocenter, where fine-grained suspended matter flocculate and settle to the seafloor, moved to the south at the Yenisei site between ca. 9000 and 4000 cal. BP (Stein et al., 2003, 2004). Obviously, the Ob site was affected for a shorter time by this marginal filter in the early Holocene. Otherwise, since ca. 2100 cal. BP, SR increases notably at the Ob site reflecting possibly a shift of the marginal filter northward. We explain this by the different river geometries of Yenisei and Ob (for further details Dittmers et al., in press). As a consequence of the different time resolution, palynostratigraphical correlation is difficult.

Comparison of the Ob with the Yenisei marine pollen record

Generally, pollen spectra of both nearshore records are similar. As expected, long-distance transported pollen such as bisaccate pollen of *Pinus* Diploxylon and *P.* Haploxylon predominate. However, the *Picea* pollen percentage curve of the Ob record (Fig. 6-9) does not reflect so clearly the retreat of boreal forest as the *Picea* pollen curve of the Yenisei pollen profile.

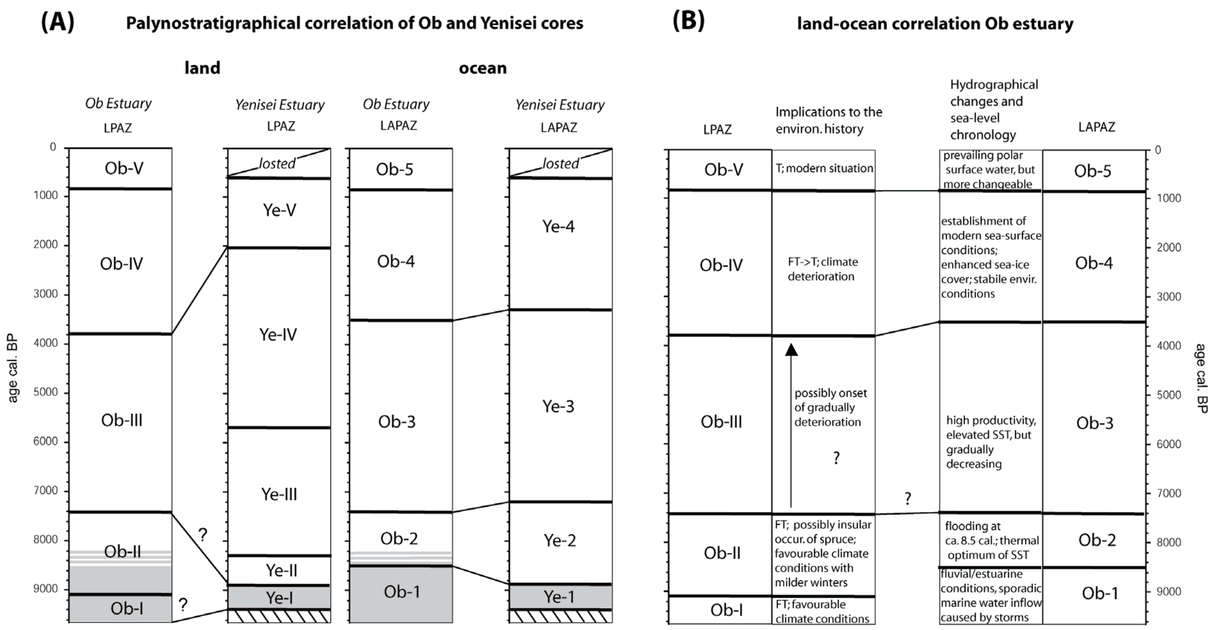

Fig. 6-8 A) and B): (A) Palynostratigraphical correlation between Yenisei estuary record BP99-04 (Kraus et al., 2003; Premke-Kraus et al., subm.) and Ob estuary record vs. calibrated ages and (B) land-ocean correlation of LPAZ with LAPAZ vs calibrated ages of core BP00-38 (Ob estuary). Abbreviations: FT=Forest Tundra; T=Tundra. Grey shadded boxes mark flooding of core site.

We infer that the *Picea* pollen curve in the Yenisei record does reflect an extralocal pollen source whereas the pollen source of the Ob *Picea* pollen curve is farther away and represents regional / long-distance transported pollen. The Yenisei river flows mainly through forested areas in contrast to the Ob river (Peterson, 1983). Thus, *Picea* pollen of the Ob record is less suitable to reconstruct advance and retreat of the boreal forest in the hinterland.

Within the non-arboreal pollen, Poaceae and Cyperaceae pollen are most abundant downcores. Poaceae pollen shows higher percentage values in the Ob pollen record. Furthermore, the Ob pollen record shows a higher diversity of non arboreal pollen, in particular of herbs. Changes in NAP: AP ratio, Poaceae, Ranunculaceae, *Artemisia*, *Thalictrum*, *Dryas,* *Helianthemum* pollen, various spores (in particular *Equisetum*, *Lycopodium* types) and total pollen concentration are major indicators showing vegetation development and climate evolution in the coastal Kara Sea region.

The palynostratigraphical correlation shows (Fig. 6-8) some similarities and also larger discrepancies: LPAZ Ye-I (9400-8900 cal. BP) and Ob-I (9600-9100 cal. BP) show similar high total pollen concentrations (from ca. 15,000 (Ob) to 25,000 (Ye) grains/g). *Betula* pollen shows fluctuations, but is dominant. Spores (*Lycopodium annotinum* type, *Equisetum* spores, Polypodiales) are also high abundant. The AP: NAP ratio reflects a predominance of AP. In contrast, the high amount of *Picea* pollen in Ye-I is not reflected in Ob-I. The minimum of Poaceae pollen in Ob-I does not show Ye-I. We explain these discrepancies by different catchment areas of the rivers throughout the Holocene. The Yenisei river drains mainly forested areas, whereas the Ob river drains the West Siberian Lowland, which consists mainly of wetlands and more open landscape vegetation communities.

LPAZ Ob-II (9100-7400 cal. BP) and Ye-II (8900-8300 cal. BP) overlap only partly. Whereas in Ob LPAZ-II, flooding occurred within this zone at ca. 8200 cal. BP, Ye-II encompasses already brackish/marine environment. Zonation of Ye-II was defined after a distinct depression of *Picea* pollen, which reflect a cold event related probably to the prominent 8200 cal. BP cold event (see e.g. Barber et al., 1999). The Ob pollen record reflects also a short-term cold event, but earlier between 9300 to 9100 cal. BP in the zone Ob-I. We interpret the zone Ob-II rather as phase with higher temperature.

We conclude that the most favorable conditions in the coastal area of the Kara Sea realm are reflected in the Yenisei pollen record in zone Ye-I, whereas zone Ye-II reflects a cold event. In opposite, the zonation of the Ob pollen record contains also the zone Ob-II. The respective pollen assemblages reflect directly or indirectly higher temperatures, the advance of boreal forest to the north far of its modern range, the presence of forest tundra communities in the coastal area, the extralocal pollen deposition due to the exposed shelf and probably evidences of the paludification of geogenous fen flood and water rise mires (terminology according to Joosten and Clarke, 2002) in the river and coastal lowlands. What we can not derive from these pollen assemblages is the magnitude of precipitation and the temperature profile during summer and winter.

The subsequent pollen zones are more difficult to connect because of the different temporal resolution of the cores. The LPAZ Ob-III (7400-3800 cal. BP) encompasses the zones Ye-III (8300-5700 cal. BP) and Ye-IV (5700-2000 cal. BP). The variations of pollen percentage values between zone Ye-III and Ye-IV are relatively low and reflect mainly the gradually decrease of *Picea* pollen curce, which is associated with favorable climate conditions. The change in pollen assemblages at the transition to zone Ob-IV at 3800 cal. BP indicating climate deterioration and forest retreat is not reflected in the Yenisei pollen record. Only *Piceae* pollen percentage curve show a noticeable decrease after ca. 4200 cal. BP in zone Ye-IV. Otherwise, the distinct change in the Yenisei record at the transition to zone Ye-V, indicating climate deterioration and forest retreat at 2000 cal. BP is not reflected in the Ob-IV. The zone Ye-V is connected with Ob-IV in part. Zone Ob-V is not recorded in the Yenisei record due to the loss of the uppermost sediment by overpenetration during coring.

We conclude that both the nearshore Kara Sea pollen records have relatively rich pollen contents and show downcores similar changes of pollen assemblages. The connecting of the local pollen assemblage zones does works partly and palynostratigraphical correlation could be an important help in poorly dated sediment cores as a further paleoproxy. Major trends of Holocene environmental evolution are reflected. Atmospherical changes are probably even better reflected, because the marine pollen record is quasi the "extract" of large-scale changes of biosphere and atmosphere. However, (Holocene) interglacial variability is less distinct reflected. Unfortunately, we could not test the benefit within interglacial-glacial cycle, because the Lateglacial-Holocene transition is not recorded in our investigated records.

<u>*Comparison of Ob and Yenisei aquatic palynomorph assemblages*</u>

The aquatic palynomorph assemblages in the Yenisei and Ob estuary record are very similar (Fig. 6-8). The same spectra were found with similar proportions of individual species. The local aquatic palynomorph zones (LAPAZ) Ye-1 (9400-8900 cal. BP) and Ob-1 (9600-8500 cal. BP) reflect both pre-transgressive, fluvial to estuarine conditions with small occurrence of dinocysts and can be connected well. LAPAZ Ye-2 (8900-7200 cal. BP) and Ob-2 (8500-8200 cal. BP) reflect time-trangressive the flooding, because the Ob sediment core is situated in shallower water depth. Also zones Ye-3 (7200-3300 cal. BP) and Ob-3 (7400-3500 cal. BP) can be well connected reflecting the end of the thermal maximum and the gradually decrease of SST. At the Yenisei site, full marine conditions and an enhanced stratification of the upper water column were reached at ca. 7200 cal. BP and at the Ob site at ca. 7400 cal. BP. Finally, zones Ye-4 (3300-600 cal. BP) and Ob-4 (3500-900 cal. BP) show both the onset of modern-like conditions. Zone Ob-5 (900 cal. BP- recent) is not recovered in the Yenisei record.

Both percentage dinocyst records are almost the same, except of the different resolution during mid-Holocene. Local stratigraphies of aquatic palynomorphs could be transferred smoothly to a regional level. The hydrographical evolution of both sites is similar except for the time-transgressive flooding.

Land-ocean correlation

Premke-Kraus et al. (subm.) discussed in detail a land-sea correlation based on the Yenisei record. They showed that in general, terrestrial and aquatic palynomorphs record similar trends in the environmental evolution both on land and in the sea. However, the correlation also revealed that the aquatic palynomorphs reflect the termination of the thermal maximum more distinct than the marine pollen record. Furthermore, the cooling of the SST preceded probably the climate deterioration.

These general features are supported by land-sea correlation of the Ob site (Fig. 6-9). We showed that aquatic palynomorphs reflect a distinct thermal maximum prevailing until to 7400 cal. BP. Pollen assemblages indicate also probably most favorable climate conditions in the coastal realm, but do not show the termination in such a clear way as the aquatic palynomorphs. At 350 cm depth at about 3700 cal. BP, marine pollen record, particularly in the concentration diagram (Fig. 6-5), and dinocyst assemblages show synchronously the most distinct change. The marine signal reflects colder SST as well as pollen indicates colder temperatures accompanied with the gradually retreat of tree-line boundary. An abrupt change in sedimentology is negligible, because grain-size composition and the lithology does not show this (Fig. 6-2A and B).

Notes to the "Holocene thermal maximum" in the Kara Sea region

For the coastal lowlands and islands in northwestern Siberia, including the Kara Sea, the early Holocene (11,100-9800 cal. BP) is associated with the "Holocene climatic optimum" (Velichko et al., 1997) or the "thermal maximum" (Serebryanny et al., 1998, Serebryanny and Malyasova, 1998). Also Andreev and Klimanov (2000) and Andreev et al. (2001) appoint this period as the warmest time during Holocene for coastal and islands area. It is doubtless that there were higher summer temperatures, probably the highest in this area.

Several local factors may have caused these favorable conditions superimposed by global atmosphere, oceanic circulation patterns and earth's orbital variations, which are only shortly mentioned here because of their complex climate forcing mechanisms (see e.g. Mayewski et al., 2004). Overall, in the early Holocene, no local effects of an adjacent final stage of LGM ice sheet influence the early Holocene climate evolution, because the complete decay of LGM ice sheet remains in the area of Barents Sea and Novaya Zemlya occurred until ca. >11,000 cal. BP (e.g. Zeeberg et al., 2001). At this time, large shelf areas were subaerial and the coastline was located as much as 150 km north of its modern location. The exposed shelf and the shift of boreal forest to the north would have decreased the albedo effect leading to enhanced temperatures (e.g Bonan et al., 1992). Increased summer insolation (Berger, 1978; Fig. 6-9) and warmer Atlantic air masses into western Russia (e.g. Harrison et al., 1996) may have played a major role for the more favorable climatic conditions in this time period. These processes are strongly influenced by positive feedbacks, which are shown by atmosphere-ocean-vegetation models (e.g. TEMPO, 1996; Ganopolski et al., 1998).

Unfortunately, the late glacial-Holocene transition and the beginning of the Holocene are not recorded in our sediment cores. What we can say is that the following period recorded in our sites since 9600 cal. BP is the most favorable upcores and persisted probably until 8900 cal. BP. Possibly, the warmest phase is already alleviated in our record. At 8900 and 8200 cal. BP, the area of the subaerial shelf shrunk considerable due to the flooding of our sites situating ca. <100 km afar from the present coastline. The dinocyst assemblages show that the inflowing marine water was already warm and the following period from 8900 and 8200 cal. BP to ca. 7200 cal. BP, sea-surface temperature was the highest upcores. As consequence of this, climate was presumably less continental and cooler with enhanced precipitation. The marine transgression influenced also the climate evolution in the hinterland.

In opposite to *one* thermal maximum in the coastal area, there exist a lot of "optima" in the non-coastal area such as the Boreal thermal maximum at ca. 9400 cal. BP (Andreev and Klimanov, 2000, Andreev et al., 2002), the Boreal thermal optimum (Andreev et al., 2001, 2003), the thermal maximum at 9400 to 9000 cal. BP (Koshkarova and Koshkarov, 2004), the climate optimum from ca. 6900 to 5800 cal. BP (Vasil'chuk, 1983 in Velichko et al., 1997 and the Holocene temperature maximum from 6900 to ca. 5000 cal. BP (Andreev et al., 2002). In addition, the so-called “Holocene climate optimum” in the mid-Holocene, which is used most of all differ due to higher humidity from the early Holocene coastal thermal maximum. It would be helpful, if *one* “Holocene thermal maximum” (HTM) following the western Arctic (Kaufman et al., 2004) would be defined.

The review of Kaufman et al. (2004) considering the spatio-temporal patterns of the HTM shows that the warming was time-transgressive across the western Arctic. Generally, Alaska and northwest Canada (including the central and eastern Beringia) experienced the HTM between 11,000 and 9000 cal. BP, about 4000 years prior to the HTM in northeast Canada, which is related to the final decay of the Laurentide Ice Sheet (LIS) (Kaufman et al., 2004). In the area of Greenland and Iceland, which is a relatively well-studied region, the HTM was delayed between 9000 and 5000 cal. BP. Kaplan and Wolfe (2006), who considered additionally paleorecords from the northeastern Atlantic sector, including the Nordic Seas and Scandinavia, come to the same conclusion, but revealed also some anomalous records. In this region, the HTM occurred earlier approximately congruent with the maximum of summer insolation between 11,000 to 9000 years ago and were less affected by the decay of the Laurentide Ice Sheet than the Baffin Bay and Labrador Sea.

Compared to this comprehensive data sets, our coastal Kara Sea HTM tend to correspond with paleorecords from the northeastern part of the North Atlantic region. However, it seems to be that our study area is less affected by the impacts of the final decay of the LIS as regions in the North Atlantic sector. Moreover, local effects might have been compensated partly large-scale events during the phase, when large areas of the Kara Sea shelf were exposed, and had a positive feedback for the coastal climate. Therefore, possibly, coastal Kara Sea climate in early Holocene responded less to large-scale climate events, such the 8.2 ka cold event, which is evident in the Barents Sea records (Duplessy et al., 2001; Sarnthein et al., 2003), is probably not or only weakly reflected in our pollen and dinocyst records.

Evidences for isostatic uplift?

The flooding of the Yenisei site by the global sea-level rise is well reflected by the occurrence of marine palynomorphs at about 8900 cal. BP (670 cm core depth) (see Premke-Kraus et al., subm.). The comparison of the well-constrained stratigraphic framework of core BP99-04 with the global sea-level curve of Fairbanks (1989) reveals a certain discrepancy. Global seal-level was at this time approximately 25 to 30 m lower than the present sea-level (Fig. 6-9). According to the flooding of the Yenisei site, the sea-level was at that time approximately 39 to 40 m below the present sea-level (32 m water depth + 6,7 m core depth). The discrepancy of about 9 to 14 m may be caused by isostatic uplift.

Regarding to the Ob site, the sharp increase of dinocysts between 510 and 500 cm core depth indicate the flooding of the Ob site ca. 8500-8200 cal. BP (24 m water depth + 500-510 cm core depth= ca. 29 m below seafloor). According to Fairbanks (1989), the global sea-level reached approximately the 17 to 20 m paleo-isobath at this time (Fig. 6-9). The discrepancy amounts about 9 to 12 m.

The dimension could be realistic: The rebound rate amounts approximately 0,9 – 1,4 mm/yr at the Yenisei site and ca. 1,1 – 1,5 mm/yr at the Ob site. That means 9-14 m for 9000 years and 9-12 m for ca. 8000 years. Considering the possible compaction of the sediment core especially in lower part (the organic-rich, fine-grained sediment cores from Yenisei estuary have relatively high water content), we suggest that the rate is approximately < 1mm/yr. But the Ob site have slightly elevated uplift rates as the Yenisei site, because, this site is ca. 170 km closer to the LGM Barents Kara ice margin.

In comparison, Zeeberg et al. (2001) calculate a rebound rate of 1-2 mm/yr from an estimated ice load of 1000 m for Novaya Zemlya resulting in an uplift of 8-16 m for 8 ka (see also Dittmers et al., in press). By rates of relative sea-level rise ranging between 5.4 to 13.3 mm/yr (see Bauch et al., 2001), this would imply that isostatic uplift of the southwestern Kara Sea is less than eustatic rise and do not leave pronounced raised beach lines along the southern Kara Sea coast. Manley et al. (2001) did not find evidence of marine deposits indicative for postglacial isostatic emergence at the Cape Shpindler area. Nevertheless, isostatic rebounds in our working area is still under debate and it requires further detailed investigations. In view of the difficulties in this still well-known key question, our result give only vague evidences about possible isostatic rebound.

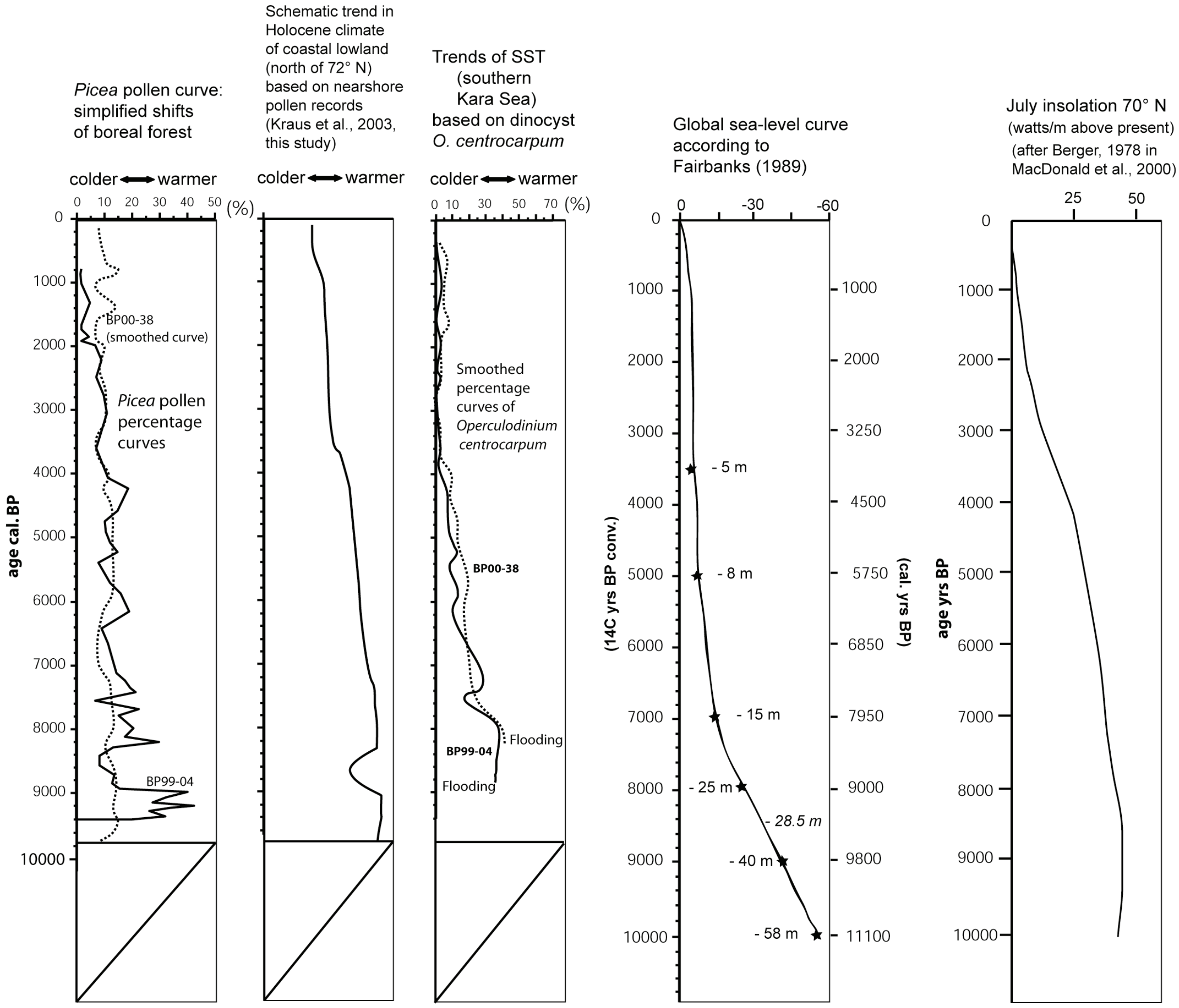

Fig. 6-9: Synthesis diagram of Holocene environmental history of the southern Kara Sea based on marine pollen records (Kraus et al., 2003; this study), aquatic palynomorph assemblages (Premke-Kraus et al., subm., this study) in relation to global sea-level curve (Fairbanks, 1989) and summer insolation at 70° N (redrawned after MacDonald et al., 2000 according toBerger, 1978) vs. calibrated ages. Dotted base line=BP00-38. Pollen sum = all pollen included, but excl. spores. Trends in climate history and vegetation zones shifts are simplified shown by *Picea* pollen curves (percentage curve of core BP00-38 is hown by smoothed curve). Changes in sea-surface temperatures (SST) are shown by the dinocyst *Operculodinium centrocarpum* smoothed percentage curves.

For the northern Kara Sea and Novaya Zemlya, isostatic rebound is detected by Forman et al. (1999). For the southern part of the Kara Sea, there is still a great insecurity about the isostasy movements in this area. Dittmers et al. (in press) discussed this issue and concluded that isostatic movement is negligible. Also Polyak et al. (2002) suggested that Ob and Yenisei estuaries were likely not affected by significant glacio-isostatic uplift.

Sea-level rise rates

Sea-level changes are primarily caused by the growth and decay of ice sheets (e.g. Lambeck and Chappel, 2001). In the Kara Sea region, the complete glaciation of the western part directly east of Novaya Zemlya, occurred ca. 10,000 years ago (Polyak et al., 2000). In the Barents Sea realm and northwestern Kara Sea at the continental margin, the timing of final deglaciation was approximately 13 to 10,000 years ago (e.g. Polyak et al., 1997; Forman et al., 1999; Zeeberg et al., 2001; Svendsen et al., 2004). Thus, the local effect of early Holocene remains of the LGM ice sheet on eustatic sea-level rise in view of exchange of mass between ice sheet and ocean is probably negligible. Thus global sea-level rise is locally counteracted due to glacio-isostatic uplift.

With respect to this, we calculate roughly relative sea-level rise rates for the southern Kara Sea, assuming that compaction of both cores and isostatic rebound is similar (see above). Between the flooding of the Yenisei and the Ob site are ca. 400 to 700 years. The sea-level rose in this time interval ca. 14-15 m (10-11 m) that yields between 14 to 28 mm/year (difference between 39-40 m below present sea-level at 8.9 ka cal. BP at the Yenisei site and 17-20 m below present sea-level at 8500-8200 cal. BP at the Ob site). However, the sea-level rise is alleviated by the counteracting isostatic uplift and the water column might have presumably remained more or less equal since ca. 7400 cal. BP, when aquatic palynomorphs show certain stabilization (see Premke-Kraus et al., subm.). At that time, global sea-level rise show a certain stabilization (Fairbanks, 1989; Zeeberg et al., 2001).

In comparison, Bauch et al. (2001) calculated sea-level rise rates between 5.4 to 13.3 mm/year for the Laptev Sea shelf. Lambeck and Chappell (2001) specified based on several areas a period of rapid and sustained sea-level rise for the time between 11.5 to 8.0 years ago approaching 15mm/year.

Both, the calculation of isostatic uplift and relative sea-level rise rates are in the southern Kara Sea are generally complex processes, which are difficult to resolve in adequate spatial-temporal resolution. Our results could only give suggestions about isostatic uplift and relative sea-level rise rates, which have to be improved by further investigations onshore as well as offshore.

Atlantic water inflow to the southern Kara Sea and marine thermal optimum

We suggest that the thermal optimum prevailing between ca. 8900 to 7200 cal. BP. in the Kara Sea was caused by relatively warm water inflow, which was probably fed by the North Cape Current (NCC) and the West Spitsbergen Current (WSC), the northernmost extension of the Norwegian Atlantic Current (NAC). Atlantic water, which was warmer than present, flew from the northern Kara Sea to the south initiating higher sea-surface temperature in the southern Kara Sea and possibly local effects amplified cumulatively this inflow. Several studies support this suggestion:

According to Hald et al. (1999), an increased heat transport occurred into the St. Anna Trough since 9500 yr BP (ca. 10,500 cal. BP). This warmer water masses represents Atlantic water, which sunk due to the increase of density by the cooling as subsurface water along the Eurasian margin and feeds between ca. 100 and 500 m depth the St. Anna Trough in the northern Kara Sea (see also Hald et al., 1999). Evidence about still earlier higher SST was revealed by Matthiessen et al. (2001), who found high abundances of the dinocyst *Operculodinium centrocarpum* prior to ca. 11,100 cal. BP.

Lubinski et al. (2001) reconstructed very cold subsurface waters and bottom waters in the southern St. Anna Trough from ca. 9000 - 7500 ^{14}C (ca. 9800 – 8500 cal. BP), whereas surface conditions at least in the northwestern portion of the Barents Sea were similar or warmer than present (Lubinski et al., 2001). Further on, they reconstructed relatively warm SST resulting from Fram Strait Branch Water (FSBW) returned to the St. Anna Trough during ca. 7500-6000 ^{14}C interval (ca. 8500 – 6900 cal. BP) in the northern Kara Sea (St. Anna Trough), preceded by a stronger influence of colder Barents Sea Branch Water (BSBW). At 7000 ka ^{14}C (ca. 8000 cal. BP), FSBW have a maximum in St. Anna Trough.

In contrast, the data set in the southern Kara Sea is very scarce. Polyak et al. (2002) reconstructed an overall productivity rise of foraminifera west of Yamal Peninsula between 10,000 to 9500 ^{14}C ka (ca. 11,100-10,500 cal. BP) reflecting significant amelioration of sea-surface conditions. However, the bottom age at 10,100 ^{14}C ka is problematic because it is performed on terrestrial plant detritus, which can be redeposited in younger sediments. According to Polyakova and Stein (2004), the sea-surface salinity was between 7500 and 6000 cal. BP well above modern levels, which is also related to the Atlantic water inflow from the northern Kara Sea.

Summarized, the referred studies tend to reflect the inflow of warmer Atlantic water masses towards the Eurasian continental margin since ca. 10,500 cal BP (Hald et al., 1999), but the strongest influence was probably from ca. 8500-8000 to 6900 cal. BP with a maximum at ca. 7000 cal. BP in the northern Kara Sea (Lubinski et al., 2001).

Remarkably, a river-influenced surface water regime and warmer than present SST prevailed coevally until 7200 cal. BP. Probably, inflowing marine water was mixed as long as the water depth was very shallow and a stratification of the upper water column was still not developed.

Local effect could have strengthened positively the inflow of warm Atlantic water: (1) mixing of water cold, dense bottom water with warmer river plume water, (2) highest summer insolation at 70º N, which is decreasing gradually after 9000 cal. BP (Berger, 1978), (3) most favorable climatic conditions ("Holocene thermal optimum") prevailing until 7400 cal. BP in the Kara Sea realm (see also Premke-Kraus et al., subm.).

The second option that heat transport to the southern Kara Sea came from the southwest through the Kara Strait, which is today an important passage of Atlantic water (e.g. Harms and Karcher, 1999), can be excluded, because at this time, there existed only the deeper incision between the south of Novaya Zemlya and the Vaigach Island at about 100 m water depth (Gataullin et al., 2001). But the water flew along the inner east Novaya Zemlya trough to the north and could not directly flow from west to east, as today (see Harms and Karcher, 1999) since the shallow shelf areas were largely still exposed and not flooded by eustatic sea-level rise. Today, there is a relatively weak but consistent throughflow of coastal water from the Pechora Sea / Barents Sea to the southern Kara Sea (McClimans et al, 2000).

Comparison with other circumarctic marine records with respect on the thermal optimum

An overall comparison of sea-surface condition reconstructions for the central and northwest North Atlantic based on dinocyst records and isotopic analyses revealed larger regional differences (de Vernal and Hillaire-Marcel, 2006), however there is a tendency to an early to mid Holocene thermal optimum except for the eastern Canadian margin, where no significant SST maximum could assessed.

We expected a priori a similar progress as in the adjacent Barents Sea. However, the relatively early onset of the thermal optimum, which had a long duration of about 2500 years, corresponds rather with studies from the northern Norwegian Sea than with the Barents Sea paleoceanographic situation. Despite of numerous studies particularly focussing on the Barents Sea, it is revealed on closer inspection some differences in the respective detected thermal optimum. Duplessy et al. (2005) exhibited the thermal optimum from 7800 to 6800 cal. BP revealed from an intermediate water mass. Voronina et al. (2001) reconstructed warmer sea-surface conditions between ca. 8000 and 5000 cal. BP in the southern Barents Sea based on dinoflagellate cyst. However, de Vernal and Hillaire-Marcel (2006) concluded based on the the same data set that no optimum of SST occurred during early to mid Holocene in the southern Barents Sea, suggesting that the North Atlantic Current (NAC) flow as subsurface stream to the north.

For the northernmost Norwegian Sea, Sarnthein et al. (2003) reconstructed a very early Holocene thermal optimum between 10,700 to 7700 cal. BP based on SST. Consequently, Duplessy et al. (2005) concluded that the duration of the thermal optimum in the eastern Barents Sea was very short (about 1000 years). Temperature of the warm Atlantic water current, which fed directly the Nordic seas was suddenly reduced by 6700 cal. BP (Birks and Koç, 2002; Sarnthein et al., 2003) and did not reach any more the Barents Sea (Duplessy et al., 2005).

The paleoceanographic evolution of East Siberian and Chukchi seas are relatively less known compared to the Barents Sea region and the northern Norwegian seas, except for the Laptev Sea. Regarding to the latter, our thermal optimum correspond well with higher SST in the Laptev Sea between ca. 10,700 and 9200 cal BP and from 8900 to 7400 cal. BP derived from diatom and dinocyst records (Polyakova et al., 2005; Klyuvitkina and Bauch, 2006). The relatively early onset of higher SST therein supports the assumption that the Atlantic water inflow carried already earlier heat to the Kara Sea as recorded in our sites.

In opposite, the Eurasian shelf seas closer to the Bering Strait and to the Pacific Ocean reflect a Holocene paleoceanographic evolution on a more regional scale (Radi et al., 2001; de Vernal et al., 2005) and a considerable anomalous development related to our study area. Recently, de Vernal et al. (2005) studied the hydrographical changes in the Chukchi Sea based on two dinocyst records and revealed a very early thermal optimum with minimum sea-ice cover prior to 12,000 cal. BP. According to them, the sea surface conditions do not reflect an early-to-mid-Holocene thermal optimum comparable to many high-latitude records, but a subsurface thermal optimum around 8000 cal. BP, which is attributed to the enhanced inflow of (warmer) Atlantic water. This anomalous stratification of the upper water masses in the Chukchi Sea revealed the difficulties of arctic wide reconstruction of paleoceanographic evolution and demonstrates the need for more comprehensive studies in sensitive locations. Moreover, it shows also limitations of quantitative reconstruction, because the visual inspection of the dinocyst record does not show in our opinion clearly the same results that were derived from quantitative reconstructions.

Conclusions

Based on two well-dated sediment cores located in the outer estuaries of Yenisei and Ob rivers (southern Kara Sea), from which we presented herein the results of marine palynological analysis of the latter one, we give an improved finding about the Holocene environmental evolution of the southern Kara Sea since 9600 cal. BP. The main results are summarized below:

- The most favorable climatic conditions in the coastal area of the Kara Sea region are reflected in both marine pollen records from ca. 9600 to 8900 cal. BP. After ca. 8900 pollen records reflect slightly alleviated favorable conditions in the coastal area of the Kara Sea region.
- The marine pollen records reflect a gradually climate deterioration with more distinct steps at ca. 3800 and 2000 cal. BP. At 3800 cal. BP, the retreat of the boreal forest is refelcetd and probably at . 2000 cal. BP, modern conditions were estbalished.
- The inundation as consequence of the sea-level rise occurred time-transgressive (8900 at the Yenisei site and 8500-8200 cal. BP at the Ob site) and give evidence about the rate, which was roughly calculated with 14mm/year.
- The eustatic sea-level rise was slightly compensated by isostatic uplift, which was roughly calculated with of ca. 0,9 to 1,5 mm/year.

- In both records, a thermal optimum is reflected (at the Yenisei site between 8900 - 7400 cal. BP and at the Ob site between 8500-8200 to 7400 cal. BP) with a relatively long duration of about 2500 years.
- We connected the thermal optimum of SST with an inflow of Atlantic water from the north to the south and reject a direct contribution of a heat transport at this time through the Kara Strait.
- Both aquatic palynomorph records reflect the establishment of an enhanced stratification of the upper water column at ca. 7400- 7200 cal. BP.
- After 3300 cal. BP, sea-surface conditions comparable to today took place with enhanced seasonal sea-ice formation and longer duration of sea-ice cover.
- Major trends of Holocene vegetational evolution are reflected in both marine pollen records.
- The general features of the land-sea correlation according to Premke-Kraus et al. (subm.) are support by the Ob record.
- The Kara Sea marine thermal optimum corresponds rather with the development and duration in the northern Norwegian Sea than with the Barents Sea.

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7. Synthesis

7.1 Marine pollen assemblages in the Kara Sea

Characteristics of pollen assemblages

For the first time, complete marine pollen records were analysed from two comparable sediment cores in the Kara Sea. The most characteristic features of the sediment cores as well as the surface sediment samples are the following (Fig. 7-1): arboreal, bisaccate pollen types in particular of *Pinus* Diploxylon type (represents mainly pollen of *Pinus sylvestris*) predominate both pollen spectra. These pollen grains are related to long-distance transport and were associated with depositional changes and changes in atmospherical circulation rather than with the real advance and retreat, respectively of this tree. Also pollen of *Pinus* Haploxylon type (representing mainly pollen of *Pinus sibirica* further *P. pumila*) is over represented in the downcore assemblages, maybe except for the zones I. In opposite, pollen of *Picea obovata* is related rather to real movements of this tree in the hinterland, which advanced nearly to the present coast. The Yenisei pollen record does reflect the distribution of spruce more accurately as the Ob record, which is interpreted with the catchment area of the Yenisei river and with the occurrence of spruce along the river valley, respectively. The *Picea* pollen curve of the Yenisei record is interpreted as a suitable indicator for coastal climate evolution.

Besides the over representation of bisaccate pollen, non-arboreal pollen comprises a relatively large proportion. Cyperaceae pollen is the most abundant pollen type downcore. In contrast to the selective enrichment of bisaccate pollen, this pollen type is interpreted as an indicator for the predominant role of sedges in the landscape. High abundances of sedges during the Holocene were commonly interpreted as being an indicator for wetland areas. However, a few sedges occur also in upplands in northern Siberia and are assigned to steppe indicators (e.g. Kienast et al., 2001).

A further abundant pollen type is Poaceae (Gramineae) pollen. Poaceae pollen is an important indicator for open landscapes related to colder conditions. The relative high pollen diversity of non-arboreal pollen of both marine pollen records argue for the determination and counting of all pollen types because in particular herbaceous pollen such as *Artemisia* pollen is a major indicator for colder climatic conditions and for open landscapes.

As expected, different spore types are also relatively abundant. However, *Sphagnum* spores, which are major peat producers and important indicators for the paludifications of ombrogenous mires, have relatively small abundances downcore. In the Ob record, *Sphagnum* spores are slightly more abundant than in the Yenisei record, which is related to the catchment area of the Ob river draining the West Siberian Lowland, one of the largest peatbogs worldwide. Nevertheless, *Sphagnum* spores in the nearshore Kara Sea shelf records play only a minor role. Also Bryales spores were only scarcely found in the Kara Sea shelf records, which is surprising, because Naidina and Bauch (2001) found relatively high abundances in a Laptev Sea shelf record.

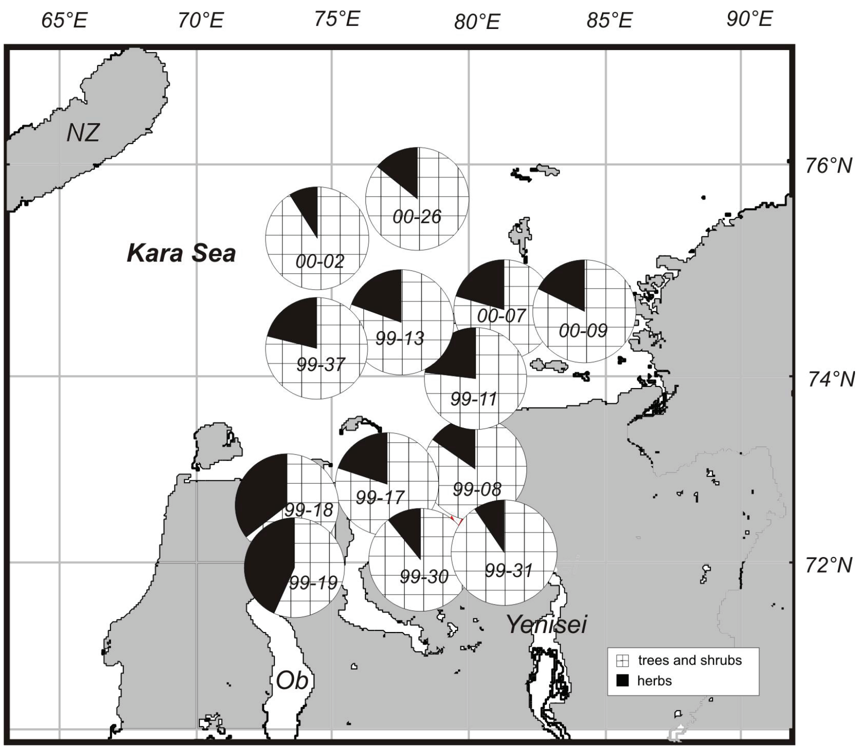

Fig. 7-1: Tree & shrub pollen (including pollen of *Pinus* Diploxylon type, *P.* Haploxylon type, *Picea obovata, Betula* Section *Nanae and B.* Section *Albae* pollen and other arboreal pollen) vs. non-arboreal pollen (NAP) in 13 surface sediment samples in the Kara Sea.

<u>*Processes of pollen deposition*</u>

Both studied nearshore Kara Sea shelf pollen records reflect a mixed signal of large-scale and extralocal/regional changes in vegetation and landscape evolution. During the sea level lowstand, when fluvial to estuarine sedimentation prevailed (LPAZ I), pollen spectra reflect the environmental evolution of the adjacent areas. In this period, the surrounding area was subaerially exposed and pollen accumulated within the incised river channels. Besides fluvial deposition, pollen spectra reflect also local/extralocal pollen rain. After the inundation due to sea-level rise, the water depth increased, river mouths moved to the south and estuarine to marine depositional conditions were established. The distance to the coast line and to potential pollen producers increased and the local/extralocal pollen signal was reduced. Therefore, pollen spectra of the LPAZ II to V reflect rather large-scale changes such as the northward and southward shift, respectively, of boreal forest, of the permafrost belts, probably paludification of peatlands and coastal climate evolution. Besides the paleoenvironmental significance, also changes in river discharge and deposition were reflected, because pollen grains represent a part of suspended matter, which is accumulated together with other fine-grained particles in relation to the prevailing local sedimentation circumstances.

7.2 Aquatic palynomorph assemblages in the Kara Sea

The composition of modern aquatic palynomorphs in surface sediment samples (dinocysts, chlorophycean algae, acritarchs, and organic benthic foraminifer linings) of three transects along the surface salinity gradient from south to the north revealed a distinct relationship to the respective surface water masses. In the southern part within the range of the inner estuaries, chlorophycean algae (*Pediastrum* spp. and *Botryococcus* cf. *braunii*) represent almost completely the assemblages (> 80%). North of 73°N, the proportion of the chlorophycean algae decreases distinctly, but remains between ca. 25 to 40 % showing, that river plumes is still perceptible in the northern part of the Kara Sea at 75°N.

The dinocyst sums are relatively low ranging between 0 and 187 in the surface samples and between 0 and 349 in the sediment cores, respectively. Within the samples with a dinocyst sum of > 25, the dinocyst assemblages show a quite low diversity (< 15 taxa) and overall relatively low dinocysts concentrations occur between ca. 50 to 1500 cysts/g in the surface samples and between 0 and 4327 cysts/g dry sediment in the sediment cores, respectively. The protoperidinioid dinocyst *Islandinum minutum* predominates the surface sediment samples (Fig. 7-2). Also the downcore dinocyst assemblages are predominated by *I. minutum* after full marine conditions were established. Protoperidinioid dinocysts are extremely sensitive to degradation due to oxygen availability in bottoms sediments (e.g. Zonneveld et al., 1997). Because of the high abundances of *I. minutum*, selective degradation of dinocysts might play a minor role at the investigated shelf sediments.

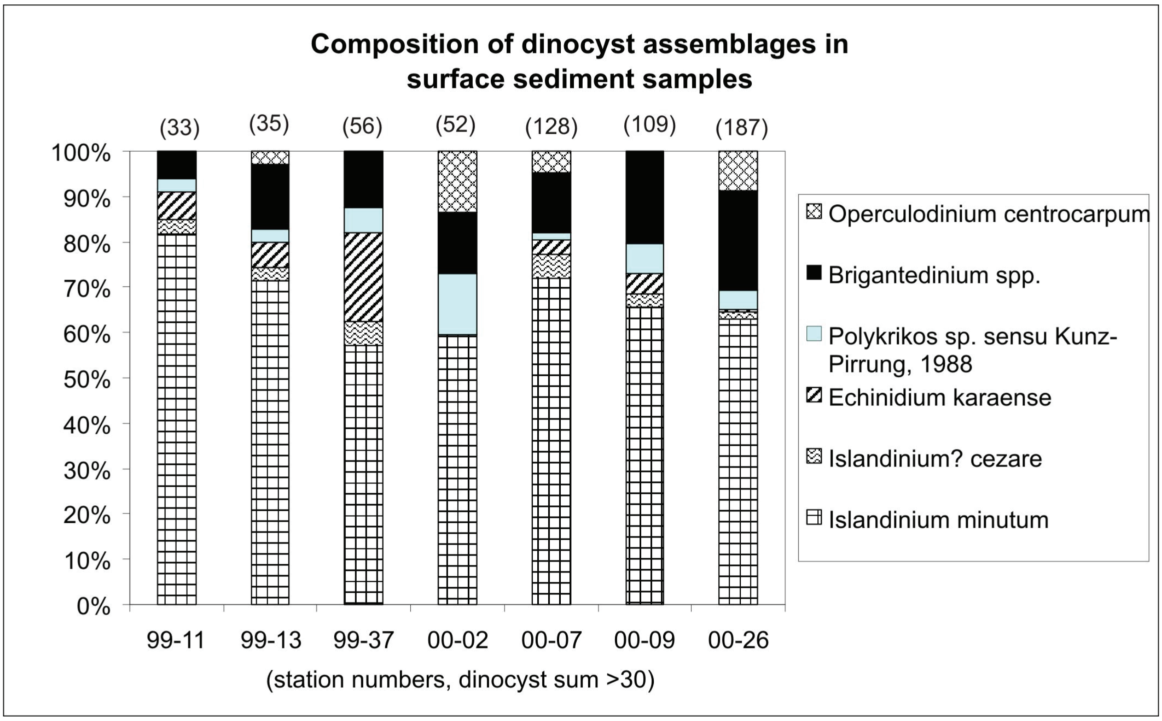

Fig. 7-2: Composition of dinocyst assemblages in surface sediment samples in the Kara Sea. Only samples are shown, which have a dinocyst sum > 30 (bracket numbers show dinocyst sum). To reveal more clarity, *Brigantedinium simplex* and *B. cariacoense* were subsumed to *Brigantedinium* spp. and the found of one morphotype *Operculodinium centrocarpum* – short processes is subsumed to *O. centrocarpum*. *Rottnestia amphicavata* and *Nematosphaeropsis labyrinthus* was found once only and is also not shown.

7.3 Definition of local assemblage zones and palynostratigraphical correlation

The local pollen and aquatic palynomorph zones were defined after visual inspection of the percentage diagrams with consideration of the concentration diagrams. In order to test this subjective definition, additional zones were automatically constructed by means of the CONISS software (Grimm, 1991) for the Ob pollen percentage diagram. The constrained cluster analysis revealed a similar zonation as the subjective zonation after visual inspection and supports the subjective zonation criteria (Fig. 7-3).

Pollenstratigraphical correlation

The establishment of a regional Kara Sea pollen zonation based on the correlation of both marine local pollen assemblage zones (LPAZ) is quite difficult and not unequivocal (Fig. 7-4). In the Yenisei record, arboreal bisaccate pollen, in particular the *Picea* pollen curve is supposed to be the most suitable indicator for tree-line movements and climate evolution, which is explained by the occurrence of spruce along the Yenisei river valley and the catchment area different to the Ob hinterland. In contrast, in the Ob record, the non-arboreal pollen, in particular Poaceae, Ericaceae, *Artemisia*, *Helianthemum*, *Dryas* and other herbaceous pollen reflect with a better accuracy the Holocene environmental evolution, which is also interpreted as a result of the catchment area (West Siberian Lowland). The direct correlation is also hampered due to the different temporal resolution of both sediment cores. According to the age model of core BP00-38, the period between ca. 9000 to 3000 cal. BP has a relatively low resolution, whereas the last two thousands years show a high resolution.

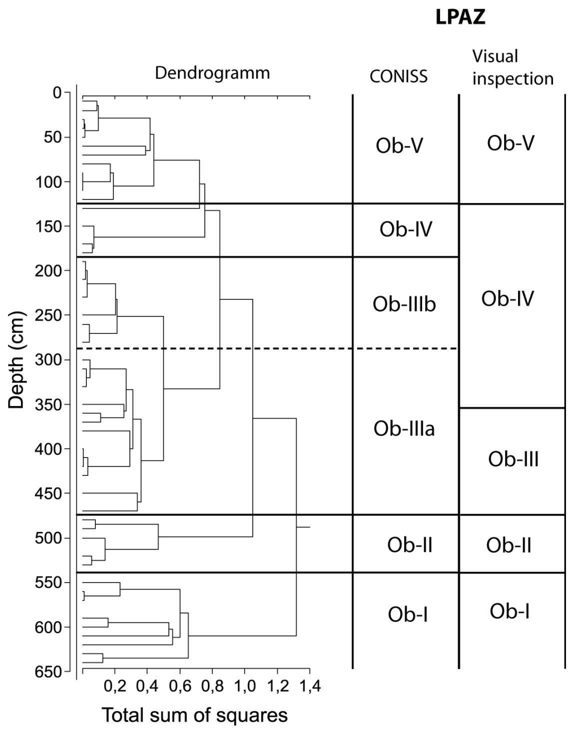

Fig. 7-3: Comparison of two different arrangements of local pollen assemblage zones (LPAZ) defined after automatically construction of constrained cluster analysis (left column) by means of the CONISS software (Grimm, 1991) and after visual inspection (right column). According to the latter method, zonation of the palynomorph assemblages was defined.

Due to differences of marine pollen assemblages relative to pollen assemblages on the adjacent land derived from lacustrine, peat or other sections, the palynostratigraphical correlation is hampered between the Kara Sea shelf records and the continental pollen records. Generally, regional correlation of local pollen assemblage zones in the adjacent coastal area of northern Siberia is difficult since paleorecords are still scarce in relation to the huge area and large north-south and east-west gradients, respectively and a majority of the continental pollen records are discontinuous and/or have a relatively poorly temporal resolution due to the low sedimentation rates.

Correlation of aquatic palynomorph assemblage zones

The correlation of the local aquatic palynomorph assemblage zones (LAPAZ) allows to propose a regional aquatic palynomorph assemblage zone (RAPAZ) for the Kara Sea (Fig. 7-4). They have a preliminary status as a better spatial resolution is required in order to evaluate this biostratrigraphical classification. Five zones are proposed and entitled with the abbreviation KS (Kara Sea) and the appendix “offshore” in order to separate from a potential regional biostratigraphical zonation for the continental hinterland of the Kara Sea. The zonation “KS offshore-1” (Kara Sea offshore zone) characterizes the pre-transgressive period, when still fluvial to estuarine conditions prevailed. The zone KS offshore-2 marks marine conditions, the thermal optimum of the surface water masses and seasonal ice-free conditions over a longer period as today. The zone KS offhsore-3 is associated with the enhanced stratification of the upper water and the gradually decrease of sea-surface temperature (SST) and sea-surface salinity (SSS), respectively. Zone KS offshore-4 represents the transition to modern cold polar surface water masses and extensive sea-ice formation. Finally, zone KS offhsore-5 encompasses approximately the last 1000 years and is characterized by certain oscillations caused probably by human impact.

7.4 Implications for the paleoenvironmental history

Climate, vegetation and landscape evolution revealed by the marine pollen records

The pollen stratigraphical correlation implies certain difficulties as mentioned before. Nevertheless, the similarities do outbalance the discrepancies and both marine pollen records reflect the same trend in the environmental evolution (Fig. 7-4). First of all, both marine pollen records indicate, that the earliest Holocene (ca. 9600 – 9000 cal. BP) was the warmest phase and marks probably the termination of the Holocene thermal maximum (HTM) in the coastal Kara Sea. The shrub tundra was displaced by the forest tundra with a stronger contribution of tree birch. The sparse herbaceous pollen suggests relatively low species diversity at this time. As a consequence of the rising sea level and groundwater table due to permafrost degradation, floodplains grew along the lowlands.

Except for short-term cold events, favorable climate conditions prevailed subsequently since 9000 cal. BP, but a gradually reduction is indicated. This led at first to better conditions for tree growth such as fir and spruce due to milder winters. However, at ca. 7400, 5700 and 3800 cal. BP, the stepwise increases of herbaceous pollen such *Artemisia* and Poaceae pollen and the decrease of *Picea* pollen, respectively indicate climate deterioration. More pronounced, at ca. 2000 cal. BP, pollen assemblages reflect a displacement of the boreal forest by Arctic tundra communities and the establishment of modern conditions. The Ob core, where the approximately last 1000 years are recorded, shows an increase of oscillations, which are interpreted as a more instable system due to human influence.

Hydrographical changes and paleoceanography inferred by aquatic palynomorphs downcore variations

The aquatic palynomorph assemblages are strongly affected by the inundation of the Kara Sea shelf, by the river discharge influencing the surface water masses during the entire Holocene and by the variability of seasonal sea-ice cover, which have been changed through Holocene time (see below, Fig. 7-4). Chlorophycean algae predominate the aquatic palynomorph assemblages, in particular in the period of sea-level low stand.

The sharp increase of dinocysts reflects the flooding of the Yenisei site at ca. 8900 cal. BP and of the Ob site at ca. 8500 cal. BP preceded by a phase of weak marine influence prevailing over approximately 700 to 1000 years. During this relatively long period, fluvial changed to estuarine conditions, but marine influence remained weak. This situation is related to sporadic storm events, which brought saline water into the still fluvial to estuarine freshwater dominated system. A second scenario is speculative: Glacial relics such as Pleistocene moraines could be responsible, that river water was retained and marine water inflow occurred delayed. After a certain sea level stand was reached, marine water flew over this potential barrier and led to the rapid increase of salinity. Also a relationship with isostatic uplift and the interplay with the eustatic rise is discussed. In this context, isostatic uplift rates are roughly calculated of about approximately 0,9 – 1,4 mm/yr at the Yenisei site and ca. 1,1 – 1,5 mm/yr at the Ob site based on the flooding timing of the both core locations related to the global sea-level rise curve according to Fairbanks (1989).

After the inundation, river discharge influenced further the surface water masses in the southern Kara Sea. Between ca. 8900 cal. BP and ca. 7200 cal. BP, a thermal optimum was detected with a duration of about 2500 years. During this time, high accumulation rates of dinocysts are related to higher productivity and seasonal ice-free conditions prevailed for a longer period in the year as today.

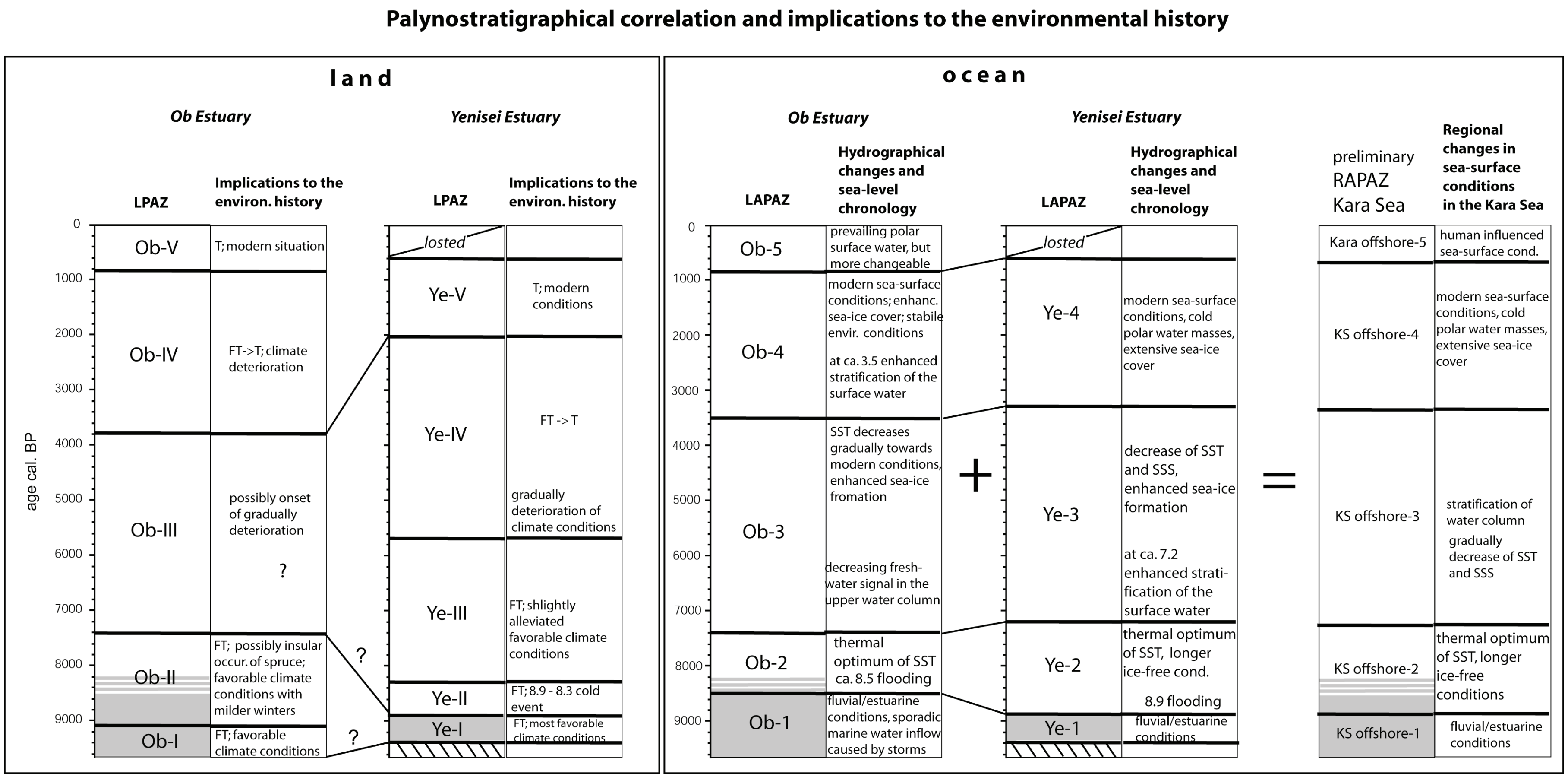

Fig. 7.4: Palynostratigraphical correlation of the sediment cores BP99-04 (Yenisei estuary) and BP00-38 (Ob estuary) and implications to the environmental history of the adjacent hinterland. The grey shaded zone marks the pre-transgressive phase during the sea-level lowstand. On the left site: stratigraphical correlation of LPAZ and the reconstructed onshore development. On the right site: stratigraphical correlation of LAPAZ and the reconstructed hydrographical changes. Rightmost: Preliminary regional aquatic palynomorph assemblage zones (RAPAZ) for the Kara Sea. Dates in calibrated ages.

The thermal optimum is explained by the inflow of warmer Atlantic water masses, which were transported by the North Atlantic Current (NAC) to the Barents Sea and along the Eurasian continental margin. From there, a branch flew mainly through the St. Anna Trough along the paleo river channels into the southern Kara Sea. In addition, local effects (e.g. higher insolation) are hold responsible for the amplified SST. Full marine conditions and a certain stratification of the upper water column were reached at ca. 7400 to 7200 cal. BP. The onset of modern-like conditions with polar and cold surface water masses and extensive sea-ice formation were established since ca. 3500 to 3300 cal. BP. The Ob record reflects the last ca. 1000 years with high temporal resolution and shows distinct oscillations, which are related to human impact.

7.5 Comparison with other circumarctic paleorecords with respect on thermal optimum

Pollen-inferred thermal optimum

In the coastal lowlands and islands in northwestern Siberia, including the Kara Sea, the "Holocene climatic optimum" occurred between 11,100 and 9,800 cal. BP (Velichko et al., 1997). Andreev and Klimanov (2000) and Andreev et al. (2001) appointed this period as the warmest time during Holocene for coastal and islands area (see also chapter 1, Fig. 1-5). Unfortunately, the Lateglacial-Holocene transition and the beginning of the Holocene are not recorded in both sediment cores. But since 9600 cal. BP, the pollen assemblages reflect the most favorable climatic conditions persisting until 8900 cal. BP. Our coastal Kara Sea Holocene thermal optimum (HTM) tends to correspond with paleorecords from the northeastern part of the North Atlantic region. However, it seems to be that our study area is less affected by the impacts of the final decay of the LIS as regions in the North Atlantic sector. Moreover, local effects (e.g. alleviated albedo effect due to exposed shelf and seasonal ice-free conditions, early settlements of boreal forest ca. 150 km north of its modern position) did presumably compensate and overprinted partly large-scale events during this phase, when large areas of the Kara Sea shelf were still exposed. Therefore, possibly, the 8.2 ka cold event, which is evident in the Barents Sea records (Duplessy et al., 2001; Sarnthein et al., 2003), is probably not or only weakly reflected in our pollen records.

Aquatic palynomorph-inferred (marine) thermal optimum

Our marine thermal optimum from ca. 8900 cal. BP to 7200 cal. BP agrees well with paleoceanographic studies from the northern Kara Sea, which exhibited the strongest influence probably from ca. 8500 to 6900 cal. BP with a maximum at ca. 7000 cal. BP in the northern Kara Sea (Lubinski et al., 2001).

Compared to other circumarctic reconstructions of sea-surface conditions, the Karas Sea thermal optimum of surface water masses, which had a duration of about 2500 years, corresponds rather with the thermal optimum in the northern Norwegian Sea than with that in the eastern Barents Sea, but was probably slightly delayed. In the Barents Sea probably no thermal optimum developed at the surface or was only weak (Voronina et al. 2001; de Vernal and Hillaire-Marcel, 2006). A much shorter thermal optimum was reconstructed for the intermediate water mass layer (Duplessy et al., 2005). In contrast, a very early Holocene thermal optimum with a duration of about 3000 years between 10,700 to 7,700 cal. BP was reconstructed based on SST for the northernmost Norwegian Sea (Sarnthein et al., 2003).

Regarding to paleorecords farther east, the Laptev Sea shows a similar evolution, with enhanced Atlantic water inflow between ca. 10,700 and 7400 cal. BP and warmer SST than today reconstructed based on dinocysts (Polyakova et al., 2005; Klyuvitkina and Bauch, 2006). This relatively early onset of higher SST supports the assumption, that the Atlantic water inflow carried already earlier heat to the Kara Sea as recorded in our sites.

In the opposite, a recent study from the Chukchi Sea does not reflect an early-to-mid-Holocene thermal optimum of the sea surface conditions comparable to ours. However, a subsurface thermal optimum was reconstructed for around 8000 cal. BP (de Vernal et al., 2005). Also, the records from the eastern Canadian Arctic margin (Baffin Bay, Labrador Sea, Nova Scotia shelf) revealed no or only a short and weak thermal optimum (e.g. Levac and de Vernal, 1997; Levac et al., 2001, de Vernal and Hillaire-Marcel, 2006).

7.6 Land-sea correlation

In general, the marine pollen records indicate the same Holocene climate and vegetation evolution as the land-based pollen records from the coastline and the adjacent hinterland reflecting an early Holocene thermal optimum, the occurrence of forest tundra at the coastal area, and the late Holocene displacement by an Artic tundra.

Both marine pollen and aquatic palynomorph records indicate an early Holocene coeval thermal optimum. However, the onset of the thermal optima onshore and offshore is not reflected in the sediment cores and it is not clear, whether they occurred time-transgressive. However, the aquatic palynomorphs show a distinct termination of the thermal optimum in contrast to the pollen records, which show rather a gradually climate deterioration after ca. 7400 cal. BP with steps at 5700 and 3800 cal. BP. Possibly, the marine thermal optimum was shorter. Also the onset of environmental conditions comparable to today occurred time-transgressive. Aquatic palynomorph assemblages indicate the establishment of modern sea-surface conditions at ca. 3300 cal. BP in contrast to the pollen assemblages, which did show this until ca. 2000 cal. BP. Thus, modern sea-surface conditions were earlier established than modern conditions onshore.

8. Outlook

The accomplishment of this study presented some difficulties as no comparable marine palynological analyses were conducted before in the Kara Sea. Thus, in a certain way, the success of this work was open with regard to the availability of this method. However, the application of the marine palynological analysis showed that in general this proxy is a suitable tracer for the reconstruction of the paeloenvironmental evolution and there is a large potential.

The herein presented data set respresents a first test. For a better spatio-temporal resolution, further investigations are necessary.

Concretely, following proposals are made for a future research:

- The box of the surface sediment samples' data set has to be enlarged northwards and westwards to reveal distribution patterns of palynomorphs for the whole Kara Sea.
- It should be tested whether the compilation of modern hydrographic data set with the enlarged surface sediment sample set by use of the best analogue method (e.g. Guiot and Goeury, 1996; de Vernal et al., 2001) leads to suitable quantitative data.
- If further marine pollen records are studied, sediment cores from the inner Ob estuary should be investigated as they show the highest proportion of non-arboreal pollen and have therefore a high potential to reconstruct the extralocal/regional vegetation development in the coastal area of the Kara Sea in more detail.
- In order to study more detailed the spatio-temporal progress of the marine (and continental) Holocene thermal optimum, sediment cores, which reflect also the late Glacial-Holocene transition should be analysed for their pollen and aquatic palynomorph content. If the glacial-interglacial cycle is studied, marine pollen records have a large potential to reflect very accurately these changes.
- Sediment cores north of 73°N should particularly be analysed for their aquatic palynomorph content in order to obtain further dinocyst records, which can possibly be used for quantitative reconstruction.
- A modern pollen map for the adjacent hinterland would be helpfully to facilitate the comparison of marine pollen data with pollen records on land (and also helpful for palynological studies on land).
- Water samples along the water column should be analysed for dinocysts in order to study the regional sedimentation processes of dinocysts.

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