

Zonation of the Laptev Sea landfast ice cover and its importance in a frozen estuary

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Abstract

The interaction between river water and landfast sea ice was investigated through synthetic aperture radar (SAR) remote sensing, ice-growth modeling and ground-based ice-core and hydrographic studies in the Laptev Sea, Siberian Arctic, in 1996/97 and 1998/99. Ice-core data in conjunction with ice-growth and SAR backscatter modeling demonstrated that the contrasts in dielectric and microstructural properties between freshwater/brackish (salinity <1 ‰) and sea ice allow a mapping of the distribution of freshwater and brackish ice as influenced by Lena River discharge. This brackish zone (surface water salinities <5 psu) extended over 2000-3000 km² inshore of the 10-m isobath and exhibited distinct SAR backscatter coefficients and image texture. In the nearshore zone, bottomfast ice growth could be identified and tracked over the growth season. Occupying up to 250 km² along the Lena Delta, bottomfast ice was not as widespread as previously hypothesized, possibly due to ice being thinner by 10 to 20 % relative to the long-term mean. In SAR and ERS-2 scatterometer data Laptev Sea landfast

ice exhibits the lowest backscatter signatures of any ice type in the Arctic Ocean, due to the lack of major deformation features. Stable-isotope data show that the landfast ice is composed to about 62 % of river water, locking up 24 % of the total annual Lena and Yana discharge. From ice-growth/isotopic-fractionation modeling and ice-core analysis, time series of surface water salinity have been derived, indicating freshening of under-ice waters during winter and north-/northeastward spreading of the river plume with under-ice spreading rates of 1.0 to 2.7 cm s⁻¹. A river freshwater budget for the inner Laptev shelf indicates flushing times of a year or more with cross-shelf export of 627 km³ during winter of 1998/99. Based on these findings, the southeastern Laptev Sea can be considered an open, seasonally frozen estuary. This system contrasts with North American shelf environments with a different response to climate variability and change.

1. Introduction

1.1. Overview

Rivers assume an important role in the Eurasian and North American Arctic as sources of freshwater (Gordeev et al., 1996; Macdonald, 2000; Lammers et al., 2001) and dissolved or particulate matter discharged into the marginal seas (Gordeev et al., 1996; Rachold et al., 1996; Lobbes et al., 2000). Supply and dispersal of freshwater have a strong impact on the thermo-haline circulation and sea-ice regimes over the shelves and in the Arctic Basin (Nikiforov et al., 1980; Aagaard and Carmack, 1989; Macdonald et al., 1995; Harms et al., 2000) and represent an important constraint for the marine ecology of Arctic shelves (Power, 1997; Petryashov et al., 1999). In turn, sea-ice processes in the river delta environment affect winter and spring freshwater dispersal as well as coastal evolution and hence constitute an important component of land-ocean interaction (Nalimov, 1995; Reimnitz, 2002). Thus, water and sediment supply and alongshore transport in Arctic river deltas are strongly affected by the sea-ice zonation, such as the distribution of bottomfast ice along the 2-m topographic ramp (Reimnitz, 2002). Owing to the strong salinity gradients in the off-delta region, ice physical properties as well as ice-ocean interaction vary considerably across a zone of typically several tens to >100 kilometers width, encompassing the entire range of ice types, from freshwater to brackish to ordinary sea ice (Macdonald et al., 1999).

Previous studies that have considered the interaction between river water discharged onto the shelf and the role of the landfast ice cover in modifying the dispersal, mixing and retention of freshwater have mostly been confined to the Mackenzie River Delta in the Canadian Arctic (Dean et al., 1994; Macdonald et al., 1999) and to a lesser extent the Alaskan Arctic, where river-ice break-up processes have received considerable attention (Walker, 1973; Reimnitz 2000). The Siberian Arctic, with three rivers (Ob, Yenisey and Lena) contributing almost half of the total freshwater discharge into the Arctic Ocean (Gordeev et al., 1996) has received less attention from this perspective. In

contrast, the nature and timing of spring flooding, its impact on the ice cover as well as its importance for coastal dynamics and the development of numerical models of these processes has been addressed in more detail for Siberian than North American rivers (e.g., Nikiforov et al., 1980; Ivanov et al., 1990a,b).

In a synthesis of river-landfast ice interaction, Macdonald (2000) introduces the concept of a frozen estuary based on work off the Mackenzie Delta and laments the lack of comparable studies in the Siberian Arctic to help substantiate and broaden this concept. According to Macdonald, such a frozen estuary comprises an onshore, positive (from the perspective of freshwater influx) component and an offshore, negative component, where the development of a flaw lead or system of polynyas along the landfast ice edge results in substantial salt release into the water column (Macdonald, 2000; see also Fig. 1). Here, we examine one of the largest rivers draining into the Arctic Ocean, the Lena, and its interaction with the landfast ice cover of the southern Laptev Sea.

1.2. The Lena-Laptev shelf and aims of this study

The sea-ice of the southern Laptev Sea can be divided into several distinct components, including the nearshore bottomfast ice, the landfast ice comprising level and deformed areas and the offshore ice pack, often separated from the landfast ice by a flaw lead (Fig. 1; Ivanov and Nalimov, 1981; Reimnitz et al., 1994; Macdonald, 2000; Reimnitz, 2002).

The bottomfast ice is important as it helps maintain submarine permafrost in the nearshore area and thus controls coastal morphology. Reimnitz (2002) and Nalimov (1995) showed that these processes result in the formation of a 10 to 30 km wide, shallow bench at approximately 2 m depth offshore of the Lena and other Arctic deltas. The origins of this bathymetric feature are ill-understood, but the bottomfast ice cover is central to most of the likely explanations (Reimnitz, 2002).

The floating landfast ice covers much of the southern Laptev Sea and in places extends more than 200 km out from the coast (Timokhov, 1994). In contrast with landfast ice in the North American Arctic, which is typically grounded at water depths around 15 to 25 m by a line of shear ridges or stamukhi (Reimnitz et al., 1994; Shapiro and Barnes, 1991), limited evidence suggests that the Laptev Sea landfast ice cover lacks such features (Gudkovich et al., 1979; Gorbunov, 1979; Reimnitz et al., 1994; Dmitrenko et al., 1999). This is in line with the fact that the sea-ice regime in this area is mostly extensional, with little onshore ice motion during winter and spring (Timokhov, 1994; Rigor and Colony, 1997). Thus, the lateral extent of the landfast ice and hence the location of the flaw leads and polynyas is controlled by processes other than anchoring of the seaward margin (Dethleff et al., 1998; Dmitrenko et al., 1999), including high ocean heat fluxes at the ice edge due to localized entrainment of warmer water or other factors related to water depth and wave propagation.

In contrast with the Mackenzie river's substantial year-round flow due to drainage of larger lakes (AMAP, 1998), Lena discharge subsides in the winter (Ivanov and Piskun, 1999; Ye et al., 2003). While the presence of rough ice and stamukhi over the Mackenzie shelf may aid in the formation of a large under-ice "lake" (Macdonald et al., 1995), residence times of freshwater over the Mackenzie shelf are short (0.5-1 year, Macdonald, 2000). Tracer studies in the Eurasian Arctic suggest that residence time of surface water and river runoff over the central Siberian shelves may be on the order of 3.5 ± 2 years (Schlosser et al., 1994; Ekwurzel et al., 2001; Guay et al., 2001), greatly enhancing the potential for river-landfast ice exchange.

The Laptev Sea ice cover and in particular the landfast ice is also of importance in the context of sediment transport by sea ice. The eastern Laptev and western East Siberian Sea have been identified as key source regions of basinwide sediment export by sea ice (Pfirman et al., 1997; Dethleff et al., 2000; Eicken et al., 2000). It has been hypothesized that even single entrainment and export events are responsible for a

significant fraction of total sediment export from the shelf and constitute a major portion of the deep-basin sediment budget in the Arctic Ocean (Eicken et al., 2000). In this context, the extent and nature of the eastern Laptev Sea landfast ice is critical, however, since it occupies much of the potential sediment entrainment areas. The distribution of river water has been shown to affect sediment entrainment (Dmitrenko et al., 1998; Eicken et al., 2000) and it also seems to impact the formation and extent of the landfast ice, indicating a need for more detailed studies of potential linkages between these processes.

Given the size and inaccessibility of the study area, it is not surprising that few if any studies to date have considered river-sea ice-ocean interactions and their impact on, and control by, the zonation of the landfast ice cover in the Laptev Sea in more detail. Here, we propose that this problem is particularly suited for a study combining satellite remote sensing with ground-based measurements and modeling. Along these lines, we have studied backscatter signatures of different ice types in the Laptev Sea in Radarsat synthetic aperture radar (SAR) data for a number of years. SAR data are particularly valuable for remote studies of freshwater dispersal in the ice-ocean system owing to the strong contrasts in dielectric properties and hence backscatter signatures of freshwater, brackish and ordinary sea ice. This study aims to demonstrate their value in elucidating key features and processes in the context of the problems discussed above. Guided by the analysis of SAR imagery, a detailed field program was completed in spring of 1999 to validate the remote sensing data and derive the ice-growth history and contribution by river water to the ice mass balance from ice-core analysis (Fig. 2). Aided by hydrographic survey data and large-scale model simulations, our findings are synthesized and discussed in the context of the frozen-estuary concept.

2. Methods

2.1. Remote sensing

Information on the large-scale ice conditions, location of open water along the fast-ice margin and progression of freeze-up were obtained from Advanced Very High Resolution Radiometer (AVHRR) satellite data made available through the Geophysical Institute's High Resolution Picture Transmission (HRPT) receiving station and the NOAA Satellite Active Archive (SAA). Data were geolocated (including final navigation with the help of prominent landmarks) with errors typically <3 km, and the visible-range and near-infrared (IR) channels 1, 2 and 4 were radiometrically calibrated.

Detailed mapping of landfast ice structure and distribution was achieved through the analysis of Radarsat SAR data covering the entire ice growth season (November through June) in 1996/97 and 1998/99. We worked with the ScanSAR Wide B product as distributed by the Alaska Satellite Facility (ASF) at a nominal ground-projected pixel size of 100 m, subsampled to 500 and 1000 m, and covering an incidence-angle range from 20.0 to 46.6°. Radarsat operates at 5.3 GHz (C-band) at HH-polarization (i.e., the radar signals are transmitted and received at horizontal polarization). SAR data were radiometrically calibrated using ASF software tools. The spatial coverage and type of data product chosen for this study were constrained by the large size of the study area and its location at the very edge of the ASF station mask. Sub-region D (Fig. 3) exhibited a few pixels with anomalously low backscatter signatures affected by calibration error. We defined a cut-off backscatter coefficient of -35 dB and spatially interpolated low data values.

2.2. Field work

Data on ice properties in the study area had been obtained during earlier field programs, in particular in fall of 1995 (Eicken et al., 2000). A comprehensive sea-ice field program was carried out from mid-April to mid-May of 1999, with study sites

selected based on the analysis of SAR and AVHRR imagery obtained for 1996/97 and 1998/99. The program comprised coring and under-ice hydrographic measurements at 12 sites. At each site, an azimuthally oriented ice core (18 cm diameter) was obtained over the entire depth of the ice cover. Horizontal and vertical thick and thin sections were prepared in the field to determine the ice textural stratigraphy and investigate ice crystal alignment processes based on examination between crossed polarizers on a rotating stage (Dmitrenko et al., submitted). At regular intervals (typically every 10 to 20 cm), a 2 cm thick horizontal section was cut from the core on site, transferred to a plastic bag and melted overnight. These melted samples were then transferred into glass bottles for storage prior to conductivity and stable isotope measurements. Ice bulk salinity was determined from conductivity measurements with a YSI model 30 sonde (measurement error <0.02 or <1 ‰ of the bulk salinity, whichever is larger). Stable-isotope concentrations of H₂¹⁸O were measured at the Stable Isotope Laboratory, Dept. of Physics and Astronomy, University of Calgary, on a VG 903 mass spectrometer (carbon dioxide equilibration, measured against VSMOW) at a precision of better than 0.4 ‰. Data are reported in the δ¹⁸O notation, with

$$\delta^{18}\text{O}(\text{sample } s) = \left[\frac{({}^{18}\text{O}/{}^{16}\text{O})_s}{({}^{18}\text{O}/{}^{16}\text{O})_{\text{VSMOW}}} - 1 \right] \cdot 1000 \text{ ‰} \quad (1).$$

Hydrographic data of temperature, salinity and under-ice currents were obtained during the field expedition (for details see Dmitrenko et al., submitted) as well as during the late summer/early fall in the previous year, just prior to freeze-up (Dmitrenko et al., unpublished data).

2.3. Ice-growth and radar backscatter modeling

The growth and vertical property profiles of sea ice were computed with an ice-growth/salt-flux model as described by Eicken (1998). The model derives the ice growth rate dH/dt by solving the surface energy balance equation at the upper ice/snow-air

interface (Equation 2) and lower ice-water interface (Equation 3), with the conductive heat flux F_c determined by solving the heat-transfer equation (Eicken, 1998):

$$(1 - \alpha)F_r - I_0 + F_L \Downarrow - F_L \Uparrow + F_s + F_e + F_c + F_m = 0 \quad (2)$$

$$F_c + F_w + \rho_i L \frac{dH}{dt} = 0 \quad (3)$$

with the incoming solar shortwave flux F_r (and ice albedo α), the shortwave flux penetrating into the ice/water I_0 , the incoming longwave flux $F_L \Downarrow$, the outgoing longwave flux $F_L \Uparrow$, the turbulent sensible and latent heat fluxes F_s and F_e , the heat flux due to melting or freezing of ice at the surface F_m , the ocean heat flux F_w , ice latent heat of freezing L and ice density ρ_i . As the model integrations were terminated prior to the onset of surface melt, $F_m = 0$. Also, the ocean heat flux F_w well inside of the shelf break is assumed to be zero based on hydrographic data (Dmitrenko et al., 2002, submitted). The turbulent fluxes were derived from standard bulk-approach parameterizations as a function of air temperature, humidity and wind speed (Maykut, 1986). The model was forced with daily average meteorological data (air temperature, dew point, wind speed, total snow accumulation) as measured at the Tiksi Meteorological Station, obtained online from the National Climate Data Center (www.ncdc.noaa.gov). A climatology of these meteorological variables for the years 1966-1997 has been derived from data obtained for Tiksi from the German Weather Service. Downwelling longwave fluxes $F_L \Downarrow$ were derived from cloud climatology and air temperatures according to the parameterization by König-Langlo and Augstein (1994). Downwelling shortwave fluxes F_r were computed according to Zillman (1972), with a high-latitude snow/ice cover correction proposed by Shine (1984) and taking into account cloud cover as described by Laevastu (1960). The fraction of shortwave radiation penetrating into the ice and underlying water I_0 was determined as described by Maykut (1986). The balance of

fluxes at the upper and lower surfaces and the heat-transfer equation thus allow for the derivation of the surface ice temperature and the ice-growth rate.

Based on these ice growth rates derived, we were also able to compute the isotopic fractionation and bulk isotopic composition ($\delta^{18}\text{O}$) of sea ice layers. This follows an approach described in more detail in Eicken (1998). Specifically, the effective isotopic fractionation coefficient ϵ_{eff} , which describes by how much ice is isotopically heavier than the water it grew from, has been estimated according to the boundary-layer model described in the aforementioned paper (equation 22, boundary-layer thickness 0.5 mm). With ice thickness and ϵ_{eff} provided by the model, we can then derive the fraction of river water present underneath the ice cover at different times during the course of the ice-growth season.

SAR backscatter signatures have been simulated with an Integral Equation Model for surface scattering and an Independent Rayleigh Scattering Model for volume scattering (Fung, 1994). The ice cover is represented by three or four layers of different salinity and temperature (assumed to be constant within each layer), based on field measurements and ice-growth model simulations. Dielectric properties of these layers have been obtained by interpolating between empirical data for the complex dielectric permittivity as a function of brine volume fraction and temperature compiled by Hallikainen and Winebrenner (1992) for 1, 4, and 10 GHz. Ice surface and bottom roughness values are based on data for smooth, level first-year ice (Onstott, 1992), while the size of scatterers (gas and brine inclusions) is derived from our field observations and data compilations (Onstott, 1992). Data on distribution of brine and gas inclusions are also from field observations.

3. Results: Mapping the zonation of the landfast ice cover with synthetic aperture radar data

3.1. Spatial and temporal variability of SAR backscatter signatures

The major ice types comprising the landfast ice cover of the Laptev Sea can be recognized in the Radarsat SAR scene shown in Fig. 3. Backscatter signatures vary substantially across the scene, with a belt of high-backscatter ice 20 km wide lining the eastern coast of the Lena Delta, landward of the 10-m depth contour in Fig. 3. This coastal ice exhibits a narrow spectrum of backscatter coefficients (Fig. 4, sub-region B), in contrast with the sea ice further offshore (to the right in Fig. 3), which reveals a higher variability of low and high scattering intensities with a winter mean backscatter coefficient of $\sigma^{\circ} = -19.5 \pm 3.5$ dB (Fig. 4, sub-region A). Along the coast, the high-backscatter ice (winter mean $\sigma^{\circ} = -14.7 \pm 1.7$ dB) is mostly confined to the area adjacent to the main river channels (Trofimov and Bykov channels, Fig. 3), that account for between 80 and 90 % of the total Lena discharge (Rachold et al., 1996; Ivanov and Piskun, 1999). The transition to backscatter signatures typical of the offshore landfast ice appears to roughly follow the 10-m isobath, with a more gradual transition in the southern reaches of this ice type (Fig. 3). Neither the Tumat nor the Olenek channels branching towards the north and northwest from the main Lena channel exhibited comparable high-backscatter ice in the nearshore waters.

During the course of the cold season, the backscatter coefficients of both of the aforementioned ice types remain comparatively stable (Fig. 4). Shallow-water coastal ice in a semi-enclosed lagoon, with a winter mean σ° of -18.0 ± 2.5 dB, exhibits slightly larger variations as a function of time, with a significant drop in σ° at the end of the observation period (Fig. 4, sub-region C). This has been studied in more detail in sub-region E off the eastern Lena Delta (Fig. 5, see Fig. 3 for location), where a belt of low-backscatter ice in the nearshore zone is seen to increase in extent as the ice-growth season

progresses. One such area has been marked by an arrow in Figs. 5a-d. In the early stages of ice growth, these low-backscatter areas are isolated patches of few hundred meters width (Fig. 5a) that extend out to approximately 3 km offshore from an island in one of the channels by late May as ice melt begins. A similar development is apparent along the entire eastern part of the Lena Delta, with approximately three-quarters of the coastline affected. Zone D (Fig. 3) represents extremely smooth, saline landfast ice formed in the protected region of Tiksi Bay.

With the onset of surface melt in late May, σ^0 undergoes substantial short-term variations for all ice types as a result of surface melt processes (Gogineni et al., 1992; Barber et al., 1995), which will not be considered in this study.

Given the substantial overlap in the histograms of σ^0 between the different ice types (Fig. 4), we investigated the utility of measures of texture in discriminating between different SAR backscatter signatures and ice types. Here, we derived texture parameters from the Neighbouring Grey Level Dependence Matrix (NGLDM) as described by Sun and Wee (1982). An element of the NGLDM $Q(k,s)$ gives the number of pixels with grey level k having s neighbours with similar grey levels. A pixel and its neighbour are said to have similar grey levels if the absolute grey level difference does not exceed a value a , ranging between 0 and 9 (Fig. 6). An analysis of the temporal variations of the texture parameters and their dependence on the similarity measure applied in the analysis indicates that the second-order moment M and the entropy T are particularly useful in discriminating between different ice types. They are defined as:

$$M = \frac{\sum_{k=1}^K \sum_{s=1}^S [Q(k,s)]^2}{\sum_{k=1}^K \sum_{s=1}^S Q(k,s)} \quad (4)$$

$$T = - \frac{\sum_{k=1}^K \sum_{s=1}^S Q(k,s) \log_2 [Q(k,s)]}{\sum_{k=1}^K \sum_{s=1}^S Q(k,s)}$$

where K is total number of grey levels in the image, and S the maximum number of neighbours (eight). As evident from Fig. 6, the combination of second moment and entropy allows for complete discrimination between different ice types, while at the same time indicating similarities between, e.g., the coastal ice off the Lena Delta and in semi-enclosed lagoons (sub-regions B and C in Fig. 3). These significant differences hold irrespective of the similarity condition employed in deriving measures of texture from the NGLDM. Discrimination based on SAR texture hence allows for a mapping of different landfast and coastal ice zones in the study area.

In explaining the regional distribution of backscatter signals, and in particular the decrease of σ^0 in nearshore areas, the magnitude and gradients of surface water salinity are of great importance, since they control the salinity and hence the dielectric properties of the ice cover (Hallikainen and Winebrenner, 1992). Field measurements carried out in October 1995 (Eicken et al., 2000) and November 1996 (Semiletov, unpubl. data) show that the distribution of high-backscatter ice generally coincides with that of freshwater or low-salinity brackish ice with salinities smaller than 1 ‰, characterized by layers of sub-mm to mm-size gas inclusions. The offshore sea ice beyond the 10-m isobath is more saline and lacks prominent gas inclusions. Freshwater discharge from the major river channels reduces surface water salinities to below 2 to 3 ‰ in the adjacent stretches of shallow water (<10 m deep). This has been validated for the diffuse transition from high- to low-backscatter ice north of sub-region D which corresponds to a distinct change in under-ice surface water salinity from 6.5 ‰ at point S1 to 2.5 ‰ at S2 to 0.9 ‰ at S3 as measured in November 1996 (Semiletov, unpubl. data; see Fig. 3 for locations of points). Furthermore, the highest backscatter ice in the river channels and in sub-region C has the lowest salinities with values mostly below 0.1 ‰.

3.2. Modelling of SAR backscatter signatures

In order to explain the variations in backscatter signatures for the different ice types, SAR backscatter coefficients have been derived from an Integral Equation Model for surface scattering and an Independent Rayleigh Scattering Model for volume scattering (Fung, 1994), with simulations carried out for conditions representative of winter (March 5, 1997) and late spring/early summer (May 29, 1997). Ice properties for both cases have been derived from the ice growth model forced with local meteorological data. Given the strong dependence of dielectric properties on ice salinity and temperature, a number of simulations have been completed for each case with a surface water salinity of 0.5, 5, and 25 ‰, representative of freshwater, brackish and seawater conditions. Brine volume profiles for the winter case are shown in Fig. 4 (right column). The surface scattering model indicates only minor differences in the strength of the surface backscatter of level ice derived from these different water bodies. Volume scattering, on the other hand, is strongly affected by the brine and gas volume fraction. The dependence on brine volume fraction and hence salinity of the ice S_i and the parent water S_w from which the ice grew is summarized in Fig. 7, which shows the backscatter coefficient σ^0 as a function of S_w . Ice property profiles for these simulations have been obtained from ice-growth modeling as described above. In accordance with observations, air inclusion sizes are set to 2 mm (sphere diameter) at a volume fraction of 3 % and brine inclusions to 0.35 mm in radius for the backscatter simulations, with brine volume fractions given by the ice-growth model as determined from ice temperature and salinity. The substantial reduction in σ^0 by 5 dB for ice grown from water of salinity 0.5 to 10 ‰ agrees with differences observed between the near-coastal high-backscatter ice (region B in Fig. 3) and the adjacent offshore ice. This reduction in σ^0 is attributed mostly to the decrease in penetration depth with increasing brine volume (Hallikainen and Winebrenner, 1992), reducing the volume-scattering contribution from gas and brine inclusions in the lower ice layers. Simulations for different pore diameters in winter sea ice demonstrate the

strong impact of scatterer size on σ° , which varies by roughly 12 dB for spherical pores with diameters ranging between 1 and 2.5 mm at incidence angles of 40 to 50°. These contrasts are diminished as the ice warms in spring. A number of uncertainties and the lack of comprehensive data sets to better constrain the simulation of the ice-air and ice-water interface pose limits on the interpretation of model results. The same holds for the simulation of backscatter evolution during spring and early summer, with surface scattering less affected by changes in ice temperature and salinity than changes in surface roughness due to ablation processes. However, the distinct reduction in σ° as surface water salinity increases above approximately 5 ‰ (corresponding to a bulk ice salinity of 0.9 ‰) is corroborated by the field measurements of surface water salinity shown in Fig. 3 and plotted in Fig. 7. It remains to be investigated, how changes in the morphology and density of gas inclusions, which also strongly depend on the bulk ice salinity, may amplify such backscatter contrasts.

Other features of interest in this study are the low-backscatter regions in the coastal areas of the Lena Delta (Fig. 5). We hypothesize that these features correspond to regions of grounded landfast ice and that the increase in areal extent corresponds to the thickening and expansion of the ice. This interpretation is commensurate with the local nearshore bathymetry and the maximum simulated ice thicknesses of around 2 m in this area. Furthermore, previous research in northern Alaska has demonstrated that a similar drop in the backscatter coefficient of freshwater lake ice can be explained by bottom-freezing of the ice cover (Weeks et al., 1977; Jeffries et al., 1994; Kozlenko and Jeffries, 2000). Based on backscatter model simulations, Wakabayashi et al. (1993a) concluded that the contrast between bottom-fast and floating lake ice was due to reflection at the ice-water interface and backscattering by tubular gas inclusions in the ice. While the Wakabayashi model shows good qualitative agreement with observations, a quantitative comparison between measured and predicted values of σ° for Arctic lake ice reveals substantial discrepancies, however (Wakabayashi et al., 1993b). This may be explained

by the fact that Wakabayashi and co-workers assumed a perfectly planar ice-water interface and argued that with Fresnel reflection from this interface and subsequent scattering by gas inclusions, energy would be directed back towards the SAR antenna.

Here, we have examined the contribution of a rough ice-water interface to the backscatter signal (see also Dierking et al., 1999). Volume scattering was assessed for a freshwater (0.5 ‰ salinity) ice cover with properties as specified above for the backscatter simulations. The bottom surface roughness due to small-scale undulations at the ice water interface was described by the rms height s , varying between 0.5 and 10 mm, and the correlation length l (exponentially correlated height distribution), varying between 30 and 90 mm, with the Fresnel reflection coefficient given by the local incidence angle. Since the radar wavelength in freshwater ice is reduced by a factor of almost 2 compared to the wavelength in air, the ice-water interface appears rougher to the radar even if the roughness scales are similar to those at the ice surface. Fig. 8 indicates that the magnitude of σ° is highly dependent on the amplitude and wavelength of the ice bottom roughness, with a difference in σ° of more than 10 dB for a 6-fold increase in s from 0.5 to 3 mm. By contrast, the backscatter signal from an ice cover frozen to the bottom is near-negligible due to the high ice content and similar dielectric properties of frozen ground (Wakabayashi et al., 1993b). Hence it appears likely, that the contrast between low-backscatter bottomfast ice and high-backscatter floating ice is critically dependent on the nature and roughness of the ice-water interface as well as the presence of gas inclusions that enhance forward scattering of the bottom-reflected signal. The lack of prominent tubular gas inclusions in the core obtained from the coastal ice (Core 8, Fig. 9) corresponds to an overall lower backscatter signature of the coastal ice as compared to lake ice. Finally, the transition between freshwater/brackish and sea ice (points S1 to S3 in Fig. 3) also confirms that it is processes in the lower ice layers that account for the high values of σ° as these drop substantially with increasing attenuation of the signal in more saline ice.

3.3. Zonation of Laptev Sea landfast ice: Determining the extent of bottom-fast, fresh/brackish and sea ice and assessing ice roughness

Based on an analysis of the SAR imagery acquired over the study area during the ice growth season 1996/97 and 1998/99 we can now determine the relative contribution of different ice types to the total mass of landfast ice in the southeastern Laptev Sea. The location of the stable landfast ice edge has been derived from time series of SAR imagery, backed up by consultation of thermal IR AVHRR scenes and validation during the field study in April and May of 1999 (Fig. 9). The total area of landfast ice (Table 1) amounts to an average of 158,600 km² for these two years, which accounts for 27 % of the total area of the Laptev Sea (based on boundaries as defined by Treshnikov, 1985). While the Radarsat SAR coverage did not allow monitoring of ice conditions west of approximately 120°E, the eastern portion of the Laptev Sea landfast ice cover accounts for approximately 75 % of the total landfast ice area based on long-term climatological data (Kotchetov et al., 1994). The differences between the two years in landfast ice extent are most distinct in the area northwest of Kotelnny Island where the 1997 ice edge comes to within 15 km of the coast, whereas the ice edge in 1999 is more than 50 km offshore.

A more detailed examination of the SAR imagery over the vast stretch of fast ice between the Lena Delta, Yana Bay and Kotelnny indicates that the ice cover is remarkably free of deformation features such as ridges or rubble ice. Most of the ridging is confined to Cape Buorkhaya where it appears to be associated with deformation in the early stages of ice formation. Similarly, some ridging is apparent between the 5 and 20 m isobaths east of the Lena Delta (Fig. 3). These findings are confirmed by the field program carried out in 1999 as well as field observations in a number of years off the southeastern Lena Delta and throughout the entire southeastern Laptev Sea by Semiletov et al. (unpublished data) and Dethleff et al. (1993). On several occasions, these groups

traversed wider parts of the ice in the study area with surface tracked vehicles and trucks and did not encounter substantial obstacles other than in the vicinity of Cape Buorkhaya.

The comparatively small degree of ridging of the ice cover is substantiated by the analysis of SAR backscatter signatures (Fig. 4). In fact, the mean value of σ^0 for sea ice of -19.5 dB is well below the lowest values of -17 to -18 dB reported for level, smooth first-year sea ice in the Beaufort Sea by Kwok and Cunningham (1994), who analyzed the radar signature characteristics of the winter ice cover in ERS-1 images. Given potential differences in incidence angle and dependence on along-track viewing geometry, we have assessed the large-scale roughness and deformation of the Laptev landfast-ice cover based on an analysis of backscatter coefficients derived from the European Remote Sensing Satellite 2 (ERS-2) C-band scatterometer at an incidence angle of 40° (Ezraty and Cavanie, 1999). Scatterometer data show a distinct dependence on ice roughness and deformation (Hallikainen, 1992; Long and Drinkwater, 1999). The frequency distribution of backscatter coefficients σ^0 derived from the gridded data set for all points covering the study area is shown in Fig. 10 for early April of 1997 and 1999. For comparison, the corresponding curves for the entire Arctic sea-ice region (i.e., the entire ice area within the Arctic Ocean and adjacent shelf seas with a total of 14,064 grid points) as well as the southeastern Beaufort Sea, covering the landfast ice area studied by Macdonald et al. (1995) are also shown. In both years, the Laptev landfast ice exhibits the lowest σ^0 values anywhere in the Arctic (mean values of -17.5 ± 1.78 dB in 1997 and -20.3 ± 1.96 dB in 1999, as compared to -13.9 ± 2.34 and -14.5 ± 2.53 , respectively, for the entire Arctic). Values are also lower than those in the southeastern Beaufort Sea (-16.6 ± 1.25 dB in 1997 and -18.7 ± 1.19 dB in 1999), indicating a smoother, less deformed sea ice cover.

The extent of freshwater and brackish ice derived from surface water with salinities below about 5 ‰ and ice salinity less than approximately 0.9 ‰ as outlined in the previous Section, has been determined based on the textural measures discussed

above. Despite interannual differences in the total extent (averaging at 2560 km²), the distribution pattern of this low-salinity ice is remarkably similar and correlates with the local bathymetry just inshore of the 10-m isobath (Figs. 3 and 9). Comparable high-backscatter low-salinity ice is absent from the northern and western Lena Delta, as corroborated by higher sea-ice salinities and lower riverine ice fractions (Fig. 9).

Bottomfast low-salinity ice, determined along the 100 km stretch of coastline in the eastern Lena Delta, is confined to a belt of mostly less than 1-2 km width that hugs most of the coastline, extending on average over 230 km² (Table 1).

4. Results: Landfast ice growth processes and the entrainment of river water into the ice cover

In order to determine the processes governing landfast ice growth and to assess the contribution of river water to the Laptev Sea landfast ice mass balance, a sea-ice coring program was completed at the locations shown in Figs. 2 and 9 in April and May of 1999. The results of these measurements are summarized in Table 2. The ice consisted almost exclusively of columnar ice (fibrous grains according to the Russian classification system, cf. Tyshko et al., 1997) indicating growth through quiet congelation of seawater at the ice-water interface and lack of frazil ice entrainment. The mean ice thickness amounts to 1.65 m, but this includes thin ice that accreted laterally along the northern and western margins of the landfast ice cover during the course of winter (see location of sites in high-backscatter ice shown in Fig. 11). The older, core area of the landfast ice is around 2 m thick and - as confirmed by the sequences of SAR images analysed over the course of the winter (Fig. 11) - completely stable.

Results from ice-growth modeling for the winter of 1996/97 and 1998/99 (Fig. 13) agree (within the limits of uncertainty in particular for snow accumulation) reasonably well with direct ice thickness measurements in the core section of the landfast

ice area (Fig. 9). The lower ice thickness for the two years studied as compared to climatology is corroborated by comparing our mean landfast ice thickness of 1.68 m to measurements carried out by Gudkovich et al. (1979 and unpublished data) yielding a value of 1.84 ± 0.21 m (with a mean snow depth of 0.08 ± 0.05 m) for 21 sites in the landfast ice of the southeastern Laptev Sea in April of 1976. Our observations are in line with below average landfast ice thickness observed in the Laptev Sea in 1996/97 and 1998/99 as compared to the long-term mean (1937-2000, Polyakov et al., 2003). Note, however, that the thickness of ice grown from freshwater (salinity of 1, all water salinities reported in practical salinity units, psu) exceeds that of brackish ice (parent water salinity of 15) by about 0.2 m or 10 %, solely due to the higher freezing point and higher thermal conductivity of freshwater ice (all other factors being equal).

The model simulations also allow us to derive the time of accretion for individual layers within the ice cores, with the onset of ice formation obtained from SAR, AVHRR and passive microwave satellite data in combination with ice-growth simulations for the thinner ice. Based on this assessment, we can derive an approximate date for the onset of azimuthal c-axis alignment in the core stratigraphy (Table 2).

The sea-ice salinity data (Table 2) reflect the impact of river water on ice growth. The average core salinity of 2.8 ‰ is lower than values typical of first-year ice by a factor of two to three (Cox and Weeks, 1988). The salinity profiles range between the typical C-shaped profile (Core 11, Fig. 12a) and a near-zero profile with somewhat higher values at the top (Core 8, Fig. 12a). Core samples were obtained before the onset of surface melt and there are no traces of meltwater infiltration or formation of superimposed ice from snow melt in any of the samples. The bulk salinity of the ice cover does provide some indication of the salinity of the parent water, and hence indirectly of the fraction of river water. However, in order to arrive at quantitative estimates we relied on the stable isotope composition and specifically the $\delta^{18}\text{O}$ of melted ice samples. Due to depletion of meteoric waters in the heavy isotope ^{18}O during

atmospheric Rayleigh distillation processes, precipitation in the Lena drainage basin and hence river runoff exhibit $\delta^{18}\text{O}$ values between approximately -18 and -21 ‰ (Schlosser et al., 1994; Eicken et al., 2000; Schlosser et al., 2000; Gibson, pers. comm.) as compared to 0.3 ‰ for the Atlantic inflow (Schlosser et al., 1994, Ekwurzel et al., 2001). Surface waters in the Laptev Sea are derived from mixing of these two endmembers and ideally fall onto a mixing line in a $\delta^{18}\text{O}$ -salinity diagram as shown in Fig. 14 for surface and river water sampled in the study area in fall of 1995. Here, the best estimate of $\delta^{18}\text{O}$ for the river water input has been taken as -19.5 ‰, based on data compiled by Schlosser et al. (2000) for Lena and Yana, and taking into account the relative contribution of the two to total discharge. Recent data from summer 2003 (J. J. Gibson, unpubl. data) indicate a mean $\delta^{18}\text{O}$ of -18.6 ‰ for Lena water. Hence, we have assessed the impact of seasonal and interannual isotopic variations and uncertainties by completing calculations for maximum and minimum endmember compositions of -18 and -21 ‰ as well. Given that the derived river water fraction can vary by more than 10% based on this uncertainty, it appears important to monitor the isotopic composition of Arctic rivers over longer time periods to aid in quantifying and understanding such isotopic variations.

The salinity and isotopic composition of sea ice grown from these waters is shifted towards lower salinities and higher $\delta^{18}\text{O}$ values due to salt segregation and rejection and isotopic fractionation, respectively (Eicken, 1998; Macdonald et al., 1999). Fractionation coefficients at typical sea-ice growth velocities range between 1.5 and 2.5 ‰, with an equilibrium value (zero growth velocity) of around 2.7 ‰ (Eicken, 1998; Macdonald et al., 1999). Thus, it is possible to derive the fraction of river water present in a sea-ice volume from the isotope mass balance according to:

$$f_{\text{riv}} = \frac{\left[(\delta^{18}\text{O})_{\text{ice}} - \epsilon - (\delta^{18}\text{O})_{\text{Atl}} \right]}{\left[(\delta^{18}\text{O})_{\text{riv}} - (\delta^{18}\text{O})_{\text{Atl}} \right]} \quad (5)$$

where $(\delta^{18}\text{O})_{\text{ice,Atl,riv}}$ is the isotopic composition of ice, Atlantic water and river water and ϵ_{eff} the effective sea-ice fractionation coefficient. The fractionation coefficient has been

derived from the ice-growth/fractionation model as described in Section 2.3 and is 2.05 ‰ for the average ice growth rate. It ranges between 1.90 and 2.24 ‰ throughout most of the ice cover (i.e., 0.1 m below the top and 0.1 m above the bottom of the ice). In general, the fractionation coefficient increases with decreasing growth velocity, but on scales much smaller than the down-core trends in isotopic composition shown in Figure 12b. For those cores where a growth history could be established (all those shown in Fig. 12 and an additional two), fractionation coefficients have been calculated for each 2 cm core sample segment. A comparison between these data sets and the bulk fractionation coefficient of 2.05 ‰ indicates a difference of 0.01 in derived river water fractions between these two approaches. Also, it needs to be pointed out that based on bathymetric and oceanographic mooring data (Dmitrenko et al., 2002 and submitted) there is no evidence of reduced exchange or formation of semi-enclosed reservoirs that could have affected isotopic composition (Gibson and Prowse, 1999). Cores obtained well within the landfast ice show consistent isotope and salinity profiles and the examples shown here are typical of these. The ice samples obtained at the margin of the landfast ice in the polynya zone exhibit more variability, mostly due to the different growth and accretion history that affects the outermost edge of the landfast ice. This is also apparent from significant differences in ice thickness (Fig. 9) and contrasting radar backscatter signatures (Fig. 11).

The mean fraction of river water f_{riv} derived for the ice cores sampled in 1999 ranges between 57 and 91 % and averages 63 % for the best estimate river water composition, with values of 59 and 68 % for the respective minimum and maximum estimate of river water composition. Based on a linear relationship between mean core salinity and river water fraction ($f_{riv} = 0.915 - 0.080 S$, $r^2 = 0.879$) we have also derived river water fractions for two cores for which no isotopes were measured, with the best estimate average being 62 % for all samples.

For an average landfast ice thickness of 1.68 m and integrating the stable-isotope measurements over the sampling area (which accounts for roughly half the total landfast ice area in the study region, Table 1), the total amount of river water held in the ice cover is 85 km³ in 1999. Approximately 4 km³ of river water are part of the freshwater ice zone bordering on the eastern Lena Delta (Table 1). With a conservative estimate of 30 % river water present in the remainder of the landfast ice in the area covered by satellite imagery (Table 1) at least 126 km³, or roughly 24 %, of the total annual river discharge by Lena and Yana are held in the landfast ice cover. As will be shown later, the total fraction is likely higher and may amount to as much as a third or half of the total annual discharge.

Based on our ice-growth simulations, we derive a time-depth scale for the cores collected at different sites. This approach is similar to that taken by Macdonald et al. (1995, 1999), although in our case we can employ a more sophisticated ice-growth and isotopic fractionation model, coupled with remote-sensing data that provide constraints on onset of ice formation in the region. With a time scale derived for each of the cores (8, 1, 3, 11) along a transect from the Lena Delta to the landfast-ice edge (Fig. 9), we determine temporal changes in the surface water composition, namely the fraction of river water f_{riv} and the salinity of the surface parent water mass S_p that the ice grew from. The latter is given as

$$S_p = (1 - f_{riv}) S_{Ash} \quad (6)$$

where the salinity of Arctic Ocean surface and halocline water S_{Ash} is taken as 34.4 (Ekwurzel et al., 2001). The time series for these cores (Fig. 15) reveals a distinct contrast between the interior landfast ice and nearshore delta locations (Cores 1 and 8) as compared to the locations further towards the landfast-ice edge (Cores 3 and 11). At the upper delta front, parent water mass salinities drop substantially and tend towards zero after the onset of ice formation (since not all core samples could be retrieved for isotope measurements, the data for later in the season are lacking). At location 1, surface water salinities are comparatively stable until about day 410 (February 14, 1999), when they

start to drop continuously to values below 10. The site located further towards the landfast-ice edge (Core 3) exhibits a more variable surface water composition, with a less distinct drop in surface water salinity commencing around day 400 (February 4, 1999). After a brief initial drop, the site located directly at the landfast ice edge shows an overall increase in surface water salinity with time. A detailed comparison between under-ice currents obtained from moorings (at coring sites 1 and 11) and downcore ice crystal c-axis alignment patterns has demonstrated that crystal alignment is a useful proxy for direction and persistence of under-ice currents (Dmitrenko et al., submitted; see also Weeks and Gow, 1978 and Tyshko et al., 1997). Thus, for all sites shown in Fig. 15, the changes in surface water salinity are preceded by the onset of persistent, uni-directional under-ice currents as determined from downcore ice crystal c-axis alignment. This spatial pattern is commensurate with an under-ice dispersal of fall and winter freshwater discharge, which is estimated at approximately 80 km^3 for the months of October through March based on data for the 1990s (Ye et al., 2003). Similar downcore profiles were observed by Macdonald et al. (1995), though with a more pronounced contrast between the fraction of river water entrained into the ice before and after the passage of an under-ice plume.

The time series of surface water salinity derived from ice-core profiles allow for a calculation of the under-ice freshwater spreading rate. Following Macdonald et al. (1995), two different approaches have been taken. First, the spreading rate of the low-salinity frontal zone has been determined from the timing of the drop in salinity at core sites 1 (day 410) and 8 (287), relative to the onset of ice formation during freeze-up (day 279), and the distance of the coring site from the mouth of the main Trofimov channel. This yields a spreading rate of 2.7 cm s^{-1} for the location close to the delta (core 8) and 1.0 cm s^{-1} further offshore at coring site 1. A second approach involves the determination of the freshening rate at the two core locations through linear regression of the time series after the onset of freshening (-0.071 and $-0.051 \text{ psu day}^{-1}$ at site 8 and 1, with the

regression model explaining 72 and 97 % of the observed variance, respectively). Based on the spatial gradient of the surface salinity contours determined underneath the fast ice during the field trip (Fig. 15), these freshening rates translate into under-ice freshwater spreading rates of 1.3 and 1.0 cm s⁻¹ for cores 8 and 1, respectively. Given the potential sources of error in this approach, values derived with the two different methods correspond reasonably well. Furthermore, they agree well with surface layer velocities of around 2 cm s⁻¹ measured at the mooring sites (Dmitrenko et al., submitted).

5. Discussion

5.1. Zonation of Laptev landfast sea ice

The combination of remote-sensing and ground-based data has provided a clear picture of the zonation of the landfast sea-ice cover in the eastern Laptev Sea, both in terms of the distribution of the key landfast ice features (bottomfast ice, freshwater/brackish ice, major deformation features, landfast ice edge, ice stratigraphy; Fig. 2) as well as with respect to the regionally varying contribution of river freshwater to the ice mass balance. Overall low SAR backscatter signatures and a general lack of prominent deformation features in the SAR scenes in conjunction with our on-ground ice observations confirm the general lack of highly deformed, grounded ice that is characteristic of the western Arctic shelves (Reimnitz et al., 1978; Shapiro and Barnes, 1991; Macdonald, 2000). This is underscored by ERS-2 scatterometer data which indicate that Laptev landfast ice has the lowest σ^0 values for sea ice anywhere in the Arctic. A number of studies have shown a clear dependence of the backscatter signal on ice roughness and ridging (Sun et al., 1992; Hallikainen, 1992; Haas et al., 1999). In the absence of surface melt processes or flooding in the Laptev Sea in April, the low σ^0 values can only be explained by the lack of roughness or deformation features.

A few larger ridge systems were identified in SAR data along the 10-m isobath off the eastern Lena Delta and running northwest-southeast between the Lena Delta and Cape Buorkhaya. While more massive, stamukha-type ridges have been reported for the southeastern Laptev Sea (Gorbunov, 1979), these appear to be quite rare and are uncharacteristic of a mostly level ice cover that has undergone little deformation prior to becoming landlocked. The same holds true for the seaward edge of the landfast ice which revealed few deformation features both in the satellite imagery and during the ground-based observations, with no evidence for any extensive deep, grounded ridges in the vicinity of the sites visited. While not observed by us, some grounded ridges appear to be associated with shoals, such as a grounded ridge observed near our sampling site 11 (Fig. 9) by Reimnitz, Dethleff and co-workers (Reimnitz, personal communication). This general lack of deformation features is commensurate with the gradual process of landfast ice accretion through attachment of larger, undeformed young ice sheets, as observed in SAR imagery (Fig. 11) and reflected in the thickness gradations apparent in these younger accretion zones (Cores 2, 10, 11, Fig. 9). This lack of heavily ridged ice is an expression of a largely extensional sea-ice regime dominated by export of sea ice from the Laptev Sea into the Arctic Ocean (Timokhov, 1994; Kotchetov et al., 1994; Rigor and Colony, 1997) and allows for comparatively free circulation and exchange underneath the landfast ice. The latter is reflected in both ice-core data and under-ice current measurements (Dmitrenko et al., 1998 and submitted).

In the North American Arctic, substantial deformation and grounding of pressure ridges exerts important controls on under-ice circulation. In the case of the Mackenzie Delta massive pressure ridges may be key in retaining an under-ice freshwater pool of Mackenzie winter discharge and reducing under-ice spreading rates (Macdonald et al., 1995). These contrasts in ice morphology between the Laptev and Beaufort landfast ice are confirmed by both SAR and scatterometer data (Fig. 10). Thus, and as discussed by Dmitrenko et al. (1999), the landfast ice extent in the Laptev Sea is not controlled by ice

deformation and the grounding position of the deepest pressure ridge keels. Rather, it appears to be closely linked to the dispersal of river freshwater prior to fall freeze-up through its impact on thermohaline circulation and stabilization of an ice cover. Hence, substantial interannual variability and differences in landfast ice extent are observed even in areas where the local topography does play a role in anchoring the ice cover, such as apparent in the differences between ice edge positions in 1997 and 1999 north of the Lena Delta and west of Kotelnny Island (Fig. 9).

Relying on the gradients in ice dielectric properties and their impact on the radar backscatter signal as a mapping tool in delineating the extent of brackish, freshwater and bottomfast ice appears to be highly promising. Thus, the extent of the low-salinity lens (<5) off the Lena Delta appears remarkably constant in the two years studied here, extending to approximately the 10 m isobath in both cases. While ridging along this depth contour may contribute to the containment of discharge, the stable-isotope data from a core collected at the margin of this feature and hydrographic measurements suggest that the extent of this zone is mostly controlled by the balance between winter river discharge and offshore dispersion of this water. As a result, this zone retains a sizeable fraction (4 km^3) of the total winter (October to March) discharge of around 80 km^3 .

A fascinating aspect of this steady supply of freshwater throughout the ice growth season is the formation of low-salinity ice in the nearshore zone of the Lena Delta that is near-transparent at radar wavelengths. As shown in Section 3 and Fig. 5, it allows the mapping of the seasonal bottomfast sea ice along the coastline and its interannual variability. Bottom freezing along the coast is important because it greatly enhances the total annual heat flux out of the seafloor sediments, providing a potential mechanism for the development and sustenance of sub-sea permafrost and ice-bonding of sediments, both of which are of importance in the context of delta build-up and coastal erosion (Nalimov, 1995; Reimnitz, 2002). Reimnitz hypothesized that the extent of bottomfast

sea ice and its impact on under-ice hydraulics in a micro-tidal ocean plays a crucial role in maintaining the unusually broad 2-m ramp that typically surrounds Arctic deltas. A remarkable outcome of this study that needs to be examined in more detail is the lack of a broader belt of bottom-fast ice as described by Reimnitz (2002) and others. Rather, we find relatively small, kilometer sized patches of bottomfast ice that grow in extent with the progression of the ice-growth season but only rarely coalesce into a broader belt. At present, it is unclear whether this may in part be due to methodological problems in deriving bottomfast ice extent. However, the ice-growth simulations for the winters of 1996/97 and 1998/99 suggest that thinner, and hence less expansive bottomfast ice may be partially responsible for this observation. Thus, both of these years exhibit ice thicknesses that are lower than the maxima derived from a climatology simulation by as much as a few decimeters (Fig. 13). Such changes in ice thicknesses can have drastic effects on the extent of bottomfast ice, since the depth of the 2-m ramp appears to be controlled by an interplay of hydraulic and sea-ice processes (Reimnitz, 2002). Variability and potentially reduced thickness of landfast ice in this region (see also Polyakov et al., 2003), may hence have resulted in reduced bottomfast ice extent, possibly altering the thermal regime of nearshore, ice-bonded sediments with potential consequences for delta morphology and coastal erosion.

5.2. The contribution of river water to the landfast ice mass balance: Large-scale freshwater dispersal and budget

The stable-isotope ice-core data demonstrate that retention of river water by landfast ice is an important process in the context of both the cross-shelf transfer of freshwater as well as for the sea-ice mass balance. Thus, river water contributes roughly two thirds to the total landfast ice mass in the study area and may also exert an influence on the ice mass balance due to its impact on the surface water freezing point and thermal

ice properties (see Section 4, Fig. 13). At the same time, based on conservative estimates, 126 km³ or 24 % of the total annual discharge of Lena (486 km³ at Stolb at the apex of the delta for the time period 1976 to 1994, data obtained from the R-ArcticNet Data Base, a regional, electronic, hydrographic data network for the Arctic region, www.r-arcticnet.sr.unh.edu) and Yana (32 km³ at Ubileynaya for the time period 1972 to 1994) into the Laptev Sea are seasonally locked up in the ice cover. This number compares with roughly 12 km³ (or 16 % of total winter flow) of Mackenzie River freshwater locked up in the landfast ice over the Canadian Beaufort Shelf (Macdonald et al., 1995). Likely, the river water fraction derived for the Laptev landfast ice is underestimating the total amount present. First, we may have overestimated surface seawater salinities for the Gulf of Buorkhaya, since work by Letolle et al. (1993) indicates that this region may trap a disproportionate amount of river water over longer periods of time. Second, with isotopically heavier (on average by about 2 ‰ in $\delta^{18}\text{O}$) sea-ice meltwater contributing to the freshwater reservoir of the upper water column, the refreezing of this ice meltwater (incl. its river component) will result in an underestimation of the river water fraction in the outer reaches of the landfast ice cover.

Apart from direct entrainment, the vast extent of the eastern Laptev and western East Siberian Sea landfast ice cover also decouples the dispersing river plume from wind forcing and hence prevents export of freshwater into the coastal polynyas and the Arctic surface layer and upper halocline. Moorings deployed underneath the landfast ice in 1998/99 (Dmitrenko et al., 2002 and submitted) and hydrodynamic modeling (Pavlov and Pavlov, 1999) indicate that under-ice currents above the pycnocline (maintained by river discharge) are setting towards the North and North-East, allowing for a transfer of river water into the New Siberian Islands Archipelago and the East Siberian Sea. A significantly narrower landfast ice belt, such as in the western and central Laptev Sea, or reductions in fast ice duration would promote offshelf transport of the surface layer (Rigor and Colony, 1997).

Based on measurements of the salinity fields at the end of summer in 1998 and the end of winter in 1999 (Fig. 16), and integrating the other data from this study, a crude, first estimate of the riverine freshwater budget of the inner Laptev shelf can be made for the winter season (Fig. 17). Lacking stable isotope data to derive the river water fraction, we have assumed that water salinities below 30 psu contained a river water fraction in proportion to the reduction in salinity. Based on this, we arrived at an estimate of river water export out of the inner shelf (defined here as inside the landfast ice edge) of 627 km^3 during the winter of 1998/99 (October to May). The total freshwater stock of 1489 km^3 at the end of summer in 1998 is most likely overestimating the amount of river water present on the inner shelf for a number of reasons. First, hydrographic stations are somewhat clustered downstream of the main Lena plume (Fig. 16). Second, freshwater influx into the study region from the west, i.e., mainly the Kara Sea inflow (which is poorly understood) and Khatanga discharge may contribute to total freshwater flux. Finally, comparison with climatological salinity data (Fig. 18) suggests that 1998 was characterized by anomalously fresh shelf waters. If the freshwater stock of summer and winter 1998/99 were to scale with surface salinity, then climatological data shown in Fig. 18 indicate that the 1998/99 freshwater stock is 2.4 times higher than the long-term average. Corrected for the long-term average, winter export of river water would amount to 242 km^3 and summer export to 244 km^3 .

In summary, this first approximation of a Laptev inner shelf freshwater budget indicates that flushing times of the inner shelf are on the order of a year or more, with substantial retention due to incorporation of freshwater into the ice. These processes appear to be an important factor in explaining the long residence times of river water over the Siberian shelves (3.5 ± 2 years, Schlosser et al., 1994; Ekwurzel et al., 2001; Guay et al., 2001). At the same time, the fast-ice regime also helps convey river discharge into the East Siberian Sea, contributing to recent surface-layer salinization observed in the Eurasian Arctic during the 1990s (Johnson and Polyakov, 2001).

Considering the width and shallow depth of the Laptev shelf, the landfast ice-river water interaction is not just contingent upon the wintertime dispersal and entrainment of river water at the base of the ice sheet but also depends on the surface salinity field at the onset of ice growth. This is underscored by the time series of surface water salinity derived from ice-core data (Fig. 15) which indicates low surface salinities even at the onset of the ice growth season and further decline as river water is advected during the course of winter with the prevailing northward flow. Hydrographic surveys and previous work have demonstrated the persistent impact of river discharge on the surface salinity well into the New Siberian Island archipelago (Eicken et al., 2000; Dmitrenko et al., submitted). The surface salinity for the late summer of 1998 as obtained during the Transdrift V cruise is shown in Fig. 16. As discussed in detail by Dmitrenko et al. (1999), the surface salinity field and hence the dispersal of the Lena discharge signal is typically confined to the eastern stretches of the Laptev Sea, with the summer surface wind forcing determining the orientation and location of salinity contour lines. Comparing the hydrographic data with surface water salinities derived from the isotopic composition of the uppermost two ice samples in each core (as outlined in Section 4) yields good correspondence between ice core data and the large-scale, long-term salinity distribution patterns (Fig. 18). The significant deviations in the central part of the landfast ice cover, where the core data underestimate the river water fraction, are attributed to the completion of the survey before the onset of fall wind and thermohaline mixing. A further source of error is our inability to distinguish between freshwater of riverine origin as opposed to sea-ice melt, owing to our lack of stable-isotope data from the water column. However, based on typical sea-ice melt and advection patterns in the Laptev Sea (Rigor and Colony, 1997), this latter factor is deemed to be of only minor importance.

The river water signal laid down in the ice cores later in the season, is in good agreement with the under-ice hydrography. In particular, our estimates of the under-ice freshwater spreading rates between 1.0 and 2.7 cm s⁻¹ (Fig. 17) correspond closely with

the advection velocity of the river plume of around 2 cm s^{-1} measured by under-ice current-meter moorings (Dmitrenko et al., 2002 and submitted). The large-scale underice circulation pattern, with northward and northeastward advection of freshwater in accordance with geostrophic circulation (Pavlov and Pavlov, 1999) is also confirmed by the azimuthal alignment of ice crystal c-axes observed over much of the landfast ice area (Table 2; Dmitrenko et al., submitted). Only the sites immediately adjacent to the flaw polynya exhibit more complex circulation patterns due to under-ice flow into the polynya zone. This is also indicated by the fact that the a core obtained from the landfast ice immediately adjacent to the polynya shows a surface water salinization at the rate of $0.027 \text{ psu day}^{-1}$ during the latter half of the ice-growth season. While river water appears to be lost into the polynya region through entrainment, hydrographic data and under-ice moorings suggest that only a minor portion of the total annual discharge is affected. Nevertheless, with approximately 3-4 m of ice produced annually over the polynya (Zakharov, 1966; Dmitrenko et al., submitted), even the diversion of 1 % of the total annual Lena discharge into the polynya zone (estimated at 5 % of the total fast ice area) would still yield a freshwater layer that amounts to 15-20 % of the total net ice growth.

These findings contrast substantially with the only other similar series of studies that we are aware of, conducted over the Mackenzie shelf by Macdonald and co-workers. First, over the Beaufort Sea shelf fall mixing appears to be vigorous enough to restore surface water salinities (and hence sea ice composition) to those characteristic of Arctic surface waters (Weingartner et al., 1998; Macdonald, 2000). Second, under-ice spreading velocities are up to an order of magnitude smaller (0.2 cm s^{-1}) than those in the Laptev Sea and even the highest velocities found directly along the Mackenzie shelf coastline (1.3 cm s^{-1}) are comparatively small. This latter contrast appears directly linked to the offshore sea-ice deformation regime over the Mackenzie shelf that results in a complex ice topography with rubble and ridged ice aligning itself parallel to the coast (Reimnitz et al., 1978; Shapiro and Barnes, 1991; Macdonald, 2000). Lower spreading rates and in

particular the strong contrast in alongshore and offshore spreading velocities that appear to be important for the retention of freshwater on the inner Mackenzie shelf for most of the winter can hence be explained by enhanced drag as well as the retention or diversion of under-ice freshwater plumes. This contrast with the comparatively level, smooth ice cover of the Laptev Sea, as borne out by the analysis of ERS-2 scatterometer data (Fig. 10), is a key factor in explaining the substantial differences in spreading rates.

5.3. The Southeastern Laptev Sea as a frozen estuary and its significance in the context of climate variability and change

Based on the evidence presented in this paper and summarized in Fig. 1, the southeastern Laptev Sea with its landfast ice cover can be considered a frozen estuary (as discussed for the inner Mackenzie shelf by Macdonald, 2000). A major fraction of the landfast ice consists of river runoff, the coastal ice in the Lena Delta region is almost exclusively composed of river water and the relatively undeformed nature of the landfast ice cover as a whole allows free spreading and under-ice circulation of river discharge throughout the year. Moreover, a substantial fraction (close to a quarter) of the annual Lena and Yana discharge is locked up in the landfast ice for up to 9 months out of the year. A unique and important characteristic of the Laptev Sea from the perspective of an estuary is its size and lack of confining features such as ice ridges, coastal topography or a narrow shelf. In contrast with the North American Arctic and most of the other Siberian rivers, the southeastern Laptev shelf can thus be considered an "open" estuary that is subject to a wide range of processes controlling the transfer of freshwater across the shelf. A most remarkable aspect of this system is its differentiation into distinct zones (see Section 3, Fig. 1, Fig. 9) and the comparatively small interannual variability found in the extent and characteristics of these zones as examined in this and other studies (Dmitrenko et al., 1998, 1999, submitted). This is exemplified by the consistency of the

linear salinity gradient found along the mean flow path of freshwater entering the Laptev Sea at the Lena Delta and exiting through the New Siberian Island archipelago and the northeastern Laptev Sea (Fig. 18). Even with substantial interannual variations in circulation, discharge and sea-ice patterns (Dmitrenko et al., 1998; Haas and Eicken, 2001; Ye et al., 2003), the meridional salinity gradient is quite consistent. While the nature of the mixing and dispersion processes of the north- and northeastward surface flow is currently not clear, the combination of divergence, transfer across the halocline and horizontal advection (both in the form of water and ice) appears effective in maintaining a steady state, relative to the late summer/early-fall snapshot of available hydrographic data sets, between the dispersal of river discharge and the entrainment of more saline offshore Arctic water into the surface layer.

Most of the interannual variability in the distribution patterns and dispersal of river water is explained by the surface atmospheric pressure field and its impact on summer freshwater transport in the surface layer as well as its influence on under-ice diversion of freshwater (Dmitrenko et al. 1998; Johnson and Polyakov, 2001). Considering the substantial freshwater residence time and the importance of vertical and lateral mixing, it is not clear to what an extent the total amount of freshwater discharged annually determines the spatial and temporal variability of the surface salinity field and entrainment of river water into the ice cover.

The open nature of the Laptev frozen estuary has important ramifications in the context of its response to climate variability and change. Thus, freshwater retention and comparatively long residence times as a result of ice entrainment and decoupling of surface ocean and atmosphere in winter can help reduce interannual variability as it allows for dissipative mixing and dispersal, effectively smoothing out initial variations in the open-water surface salinity field. Furthermore, once the ice becomes landfast, the importance of wind forcing in driving flow and freshwater mixing is greatly diminished (Dmitrenko et al., 2002 and submitted). Hence, landfast ice helps preserve the under-ice

halocline and has important consequences for the ocean heat flow to the bottom of the ice cover as well as the potential entrainment of sedimentary particles into the sea ice (Dmitrenko et al., 1998; Dethleff et al., 2000; Eicken et al., 2000). A simple probabilistic model indicates a less than 10 % probability of convection down to the seafloor in the entire southeastern Laptev Sea, as compared to a probability of 70 % in the western Laptev polynya (Dmitrenko et al., submitted). This actually helps stabilize the status quo of a strongly stratified surface with a stable surface salinity gradient and substantial entrainment of river water into the landfast ice cover. Ultimately, these processes aid in the diversion of freshwater away from the Laptev Sea polynya, an important site of dense brine formation (Zakharov et al., 1966; Dethleff et al., 1998, Dmitrenko et al., submitted). In the absence of penetrative mixing and only moderate wind stirring during fall freeze-up, such a frozen estuary can be quite stable with little interannual variability over periods of years. Based on modeling work by Johnson and Polyakov (2001), it appears as if the 1990s are in fact a decade characterized by a highly stable frozen-estuary regime. In contrast, a more dynamic ice cover, enhanced thermohaline and wind mixing as well as changes in shelf circulation can substantially reduce the amount of freshwater retained within the landfast ice system. Presently it is not clear whether the Laptev Sea would sustain such a regime, given the magnitude of river runoff and the annual persistence of landfast ice. However, the possibility of such a transition into a regime where freshwater is rapidly dissipated and flushed out without segregation into the surface layer and interaction with the ice cover would have important consequences for the production of dense water over the shelf and the ice mass balance of the southeastern Laptev Sea.

6. Conclusions

Based on radar remote sensing and sea-ice and surface hydrography field data, we have delineated the major units within the Laptev Sea estuarine/landfast-ice system. The

strong dependence of SAR backscatter signatures (both in magnitude and texture) allows for a discrimination between freshwater/brackish and ordinary sea ice. While further modeling and field studies are required to validate and improve the method, it may be of substantial use in studying the interannual and spatial variability in freshwater discharge and dispersal onto a seasonally frozen shelf. In the Laptev Sea, the amount and distribution of sea ice with salinities less than approximately 1 ‰ (grown from seawater with salinities of around 5 or less) did not vary substantially between 1996/97 and 1998/99 and were mostly following the 10-m isobath on the shelf off the eastern Lena Delta. However, with pan-Arctic wide-swath SAR data coverage, the approach outlined here in combining ice-growth/salt-flux models and analysis of SAR backscatter signatures should allow for studies of the freshwater dispersal and river-plume/ice-interaction processes in the entire circum-Arctic. Such studies could contribute to improving models of freshwater transfer and mixing over the shelves, an important gap in current understanding of Arctic freshwater cycling (Harms et al., 2000; Lammers et al., 2001).

SAR can also provide valuable data on the distribution of bottom-fast ice at various stages of the ice-growth season in brackish or freshwater coastal waters. Such data are crucial in understanding thermal constraints on erosion or accretion in Arctic deltas and estuaries (Are, 1998; Reimnitz, 2002). A potentially important finding of this study is the comparatively small areal extent of bottomfast ice at the end of winter in 1997 and 1999, which is in distinct contrast with a broad bottomfast ice belt hypothesized by Reimnitz (2002) and believed to be of importance in maintaining the nearshore morphology (2-m ramp) of the Lena and other Arctic deltas. While tidal fluctuations as well as other factors may play a role, our ice growth simulations also point to lower ice thicknesses in the 1990s compared to the long-term climatology as important contributing factors.

The field work carried out in 1999 confirmed and extended the findings from satellite data analysis, insofar as it could be shown that the landfast ice cover of the Laptev Sea is receiving substantial (around 60 %) contributions to its mass from river water as it disperses through the Laptev Sea. Ice growth modeling also confirmed our interpretation of ice thickness gradations found in the southern parts of the landfast ice as being due to differences in the amount of river water entrained. Thus, differences in freezing point and thermal properties of freshwater as compared to sea ice are hypothesized to account for a 0.2 m difference in ice thickness between the coastal freshwater ice and offshore brackish ice (Figs. 9 and 13). While the modeling is subject to uncertainties, the detection of differential ice growth requires comparable surface meteorological forcing and snow deposition for the sites that are being compared. This latter assumption is not unreasonable for the comparatively small area, with different snow depositional histories accounting for the largest potential error. Changes in the amount of freshwater entrainment may thus be reflected in interannual or large-scale spatial thickness variations of undeformed landfast ice. Given that many coastal measurement sites in the Kara and Laptev Seas are located within the realm of influence of river plumes, it may have to be investigated to what an extent recently reported longer-term changes in Siberian landfast ice thickness (Polyakov et al., 2003) might in part be driven by variations in river discharge (Lammers et al., 2001; Ye et al., 2003).

The entrainment of river water into the landfast ice cover is also of importance from the perspective of cross-shelf freshwater transfer and large-scale mixing processes. In 1999, roughly one quarter of the total river discharge was locked up in the southeastern Laptev Sea landfast ice, and possibly as much as another 10 to 20 % may have been entrained into the landfast ice of the western East Siberian Sea. At the same time, the landfast ice cover substantially reduces under-ice mixing and in the presence of a distinct halocline, it appears that even the limited salt rejection from growing landfast ice is highly unlikely to penetrate to deeper shelf water. Thus, freshwater entrainment and

decoupling from the atmosphere work in concert to substantially enhance residence times of freshwater over the broad Siberian shelves. The increased river discharge onto the shelf (Ye et al., 2003) and the absence of sufficient wind and tidal mixing during the past decade, help maintain the status quo of a highly stratified large-scale estuarine system that is frozen for around nine months out of the year. Not least due to the ramifications of reducing the buoyancy flux into the coastal polynya environment, a brine-production site of pan-Arctic importance (Cavaliere and Martin, 1994; Winsor and Björk, 2000), climate variability and change can have a substantial, non-linear impact on this landfast-ice/estuary/polynya system.

In discussing interannual variability, however, it needs to be pointed out that at least in the 1990s, the large-scale transfer of river water and, by implication, its interaction with the ice cover, appear to have been remarkably consistent (Fig. 18). Analyses of surface salinity fields have established the close correspondence between the landfast ice extent and river-water dispersal (Dmitrenko et al., 1999). Here, we hypothesize that the combination of a number of processes aiding dispersal as well as the critical role of the landfast ice discussed above help in maintaining such a relatively stable pattern. One important conclusion from these findings concerning an open, frozen estuary is that so far, the landfast-ice system has proven to be quite resilient in light of recent changes in the large-scale offshore sea-ice regime (Serreze et al., 2000; Brigham, 2000). However, the current distribution of river water over the Laptev shelf as well as its cross-shelf transfer and interaction appear to be the product of a dynamic equilibrium and further research will have to evaluate its sensitivity to change.

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Figure captions

Fig. 1: Schematic summary of the southern Laptev Sea ice cover and river processes (S_w salinity of surface seawater, psu; S_i bulk sea-ice salinity; the order-of-magnitude, approximate width of the different zones is also indicated).

Fig. 2: Map of the study area; the rectangular box delineates the coverage of the SAR scene shown in Fig. 3.

Fig. 3: Radarsat SAR scene for March 5, 1997 showing Eastern Lena Delta and adjacent landfast ice cover (for location of scene see map in Fig. 2). SAR incidence angle varies roughly between 40 and 45° across the scene. Also shown are sampling locations S1 through S3 visited in November of 1996 and subregions A to E analysed in more detail. The 10-m depth contour has been digitized from sparse depth soundings published in Russian hydrographic charts.

Fig. 4: Time series (left) and frequency distributions (center, March 5, 1997) of backscatter coefficients σ^0 and ice brine-volume fractions from model simulations for March 5, 1997 (right).

Fig. 5: Extent of low-backscatter ice off the eastern Lena Delta during the course of the ice-growth season (sub-region E, see Fig. 3 for location). Each scene is 12.5 by 75 km in size and has been contrast-enhanced. The areal extent of the low-backscatter ice has been derived from low-pass filtered images segmented based on mean backscatter coefficients and is shown at the bottom. A prominent region of such changes in backscatter signal is highlighted by an arrow.

Fig. 6: Results of texture analysis, depicting the second moment and entropy derived from the NGLDM. For each subregion, the NGLDM parameters have been calculated for maximum allowable differences between adjacent pixels ranging from 0 to 9 grey values. These differences correspond to parameter a (see Section 3.1) and the points linked by a line represent data for each sub-region A through D, with a ranging between 0 and 9. The circle at the end of these lines indicates the point with $a = 0$ (i.e., with the similarity criterion requiring that a pixel and its neighbour show exactly the same grey value).

Fig. 7: Backscatter coefficient σ^0 of level sea ice as a function of the salinity of the parent water mass for summer (late May) and winter (early March) conditions as based on ice growth modeling (radar incidence angle 50°). The solid dots are based on data obtained

from field measurements of surface water salinity in fall of 1996, with the bars indicating the standard deviation of the SAR scene sub-area.

Fig. 8: Simulated scattering contributions of the ice-water interface layer for winter freshwater ice. Bottom roughness varies as specified by the rms heights and the correlation length l .

Fig. 9: Location of landfast ice edge and boundary of freshwater/brackish ice off Lena Delta in 1997 (grey, thick line) and 1999 (black, thick line). Sampling locations of 1999 field sampling campaign along with the fraction of river water (in %), total ice thickness (in m) and core number are also shown. An asterisk indicates that river water fraction in the ice has been derived from ice core salinities. Further details are given in the text and in Table 1. A thin line joins cores C8, C1, C3 and C11 along a transect from the Lena Delta to the landfast ice edge as discussed in more detail in the text.

Fig. 10: Frequency distribution of backscatter coefficients σ^0 for ERS-2 C-band scatterometer data (40° incidence angle) covering the landfast ice of the southeastern Laptev and southeastern Beaufort Sea as well as for the entire Arctic (A: April 7-13, 1997; B: April 5-11, 1999; note scale change in vertical axis).

Fig. 11: Radarsat SAR scenes from early (A: December 26, 1998) and late (B: May 5, 1999) in the ice-growth season. White dots indicate locations of ice core sampling points.

Fig. 12: Vertical profiles of sea-ice salinity (a) and $\delta^{18}\text{O}$ (b) for samples taken in 1999. Locations of coring sites shown in Fig. 9. The arrows mark the onset of azimuthal crystal c-axis alignment as determined from oriented core sections (Dmitrenko et al., submitted), indicating the onset of a directionally stable under-ice current.

Fig. 13: Time series of ice thickness from ice-growth simulations for 1996/97, 1998/99 and model simulations forced with a climatological time series of meteorological data for the period 1966-97. Simulations for 1998/99 include those representative of thick landfast ice, freshwater ice (Core 8) and two additional sites (3 and 11) along the transect shown in Fig. 9.

Fig. 14: Summary plot of $\delta^{18}\text{O}$ vs. salinity for ice samples collected in 1999 (dots). For comparison, ice and water samples obtained in fall 1995 in the same region (Eicken et al., 2000) are also shown. The mixing line for Lena water and the Atlantic inflow are shown,

with the best-estimate Lena $\delta^{18}\text{O}$ (-19.5‰) represented by a solid line and the maximum and minimum estimates (-18 and -21‰) shown as dashed lines.

Fig. 15: Time series of parent water salinity as derived from ice-growth/isotopic-fractionation modeling and ice core data for four locations along a transect from the Lena Delta to the landfast ice edge (Fig. 9). Arrows indicate the onset of ice crystal c-axis azimuthal alignment (a measure of persistent, unidirectional currents) as derived from the core textural data.

Fig. 16: Surface salinity field in the study area obtained during the Transdrift V cruise in late summer of 1998 (A) and from under-ice measurements in May of 1999 (B).

Fig. 17: River freshwater (RFW) budget and fluxes for the Laptev Sea inner shelf (inside the landfast ice edge) for winter of 1998/99, shown in italics. Also indicated are the under-ice freshwater spreading rates derived from cores 8 and 1 (for details see text).

Fig. 18: Surface water salinity obtained from hydrographic transects (1989-1998) completed in the study area between 127 and 135°E , plotted as a function of latitude starting at the river mouth. Also shown are salinities derived from the $\delta^{18}\text{O}$ of the uppermost two ice segments in each core (see Section 4).