Impact of flaw polynyas on the hydrography of the Laptev Sea

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Abstract

Based on hydrological data from 1979 to 1999 the average long-term salinity of the flaw polynya in the Eastern Laptev Sea is estimated. A new method to evaluate ice production based on hydrological rather than sea-ice observations is proposed. Average annual ice production in the polynya ranges between 3 and 4 m. The probability of convective mixing penetrating down to the seafloor is highest in the regions of the flaw polynya, but does not exceed 20% in the Eastern and 70% in the Western Laptev Sea. Conductivity–temperature–depth (CTD) measurements and observations of currents carried out in April–May 1999 allowed us to investigate the surface circulation along the margins of the Laptev Sea flaw polynya. The convective nature of the surface currents, with velocities measured as high as 62 cm/s, is discussed. Currents are most likely part of circulation cells, which arise as a result of brine rejection due to intensive ice formation in the polynya. It is shown that the spatial alignment of sea ice crystals in the marginal part of the polynya is most likely a consequence of the quasi-stationary cellular circulation.

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1. Introduction

Large, persistent areas of open water and young ice off the land-fast ice are referred to as flaw polynyas (World Meteorological Organization (WMO), 1970; Zakharov, 1997). The system of flaw polynyas on the Russian Arctic shelf, known as the Great Siberian Polynya, is an important component of the Arctic climate system. Extensive stretches of open water up to 200 km wide combined with extremely low air temperatures induce intensive ice formation and local increases in salinity of the water column during winter and early spring. The Great Siberian Polynya is a

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possible source of both saline shelf waters for the Arctic Ocean (Martin and Cavalieri, 1989; Cavalieri and Martin, 1994; Winsor and Björk, 2000) and Transpolar Drift ice (Eicken et al., 1997; Dethleff et al., 1998; Alexandrov et al., 2000). The total ice thickness formed in the Great Siberian Polynya during winter is estimated to range from 3.6 m ([Zakharov, 1966], central Laptev Sea) to 21 m ([Martin and Cavalieri, 1989], eastward from the Severnaya Zemlya Islands). Among western scientists, Martin and Cavalieri (1989) were the first to point out the importance of the Siberian coastal polynya in the Arctic climate system. Although flaw polynyas in the Russian Arctic have long been of interest to researchers, relevant studies are underrepresented in the literature; for example, in their comprehensive overview paper on physical processes and the environment of polynyas, Smith et al. (1990) include only one short paragraph on the Siberian Shelf Polynya.

Previous studies have described the Great Siberian Polynya as a latent heat polynya; offshore components of surface wind forcing may create open water and induce new ice formation in the absence of sensible heat input from below (Zakharov, 1966, 1997; Dethleff et al., 1998). However, based on studies performed from 1993 to 1998, Dmitrenko et al. (1999a,b) have suggested that among other factors, sensible heat transport impacts the location of the fast-ice edge.

In the Siberian Arctic seas, flaw polynyas are most distinct in the Laptev Sea (Fig. 1; Table 1). In comparison with the polynyas of the Barents, Kara and East Siberian Seas, they are the largest in size and occur most frequently (Zakharov, 1997). Zakharov (1966) is the most recent Russian publication describing the impact of the flaw polynyas on the hydrography of the Laptev Sea shelf. Since then, extensive data sets on Laptev Sea hydrography and ice conditions have been collected during a number of international and Russian expeditions in the area. Our understanding of the physics underlying the relevant oceanographic processes has been substantially improved, not least due to advances in measurement techniques, including the application of satellite remote sensing. Newly available data sets now provide an opportunity to evaluate the impact of the flaw polynyas on the hydrography of the Laptev Sea in more detail and to a higher degree of accuracy and, furthermore, allow us to expand on and possible revise earlier, mostly qualitative assessments and hypotheses.

In the first part of this contribution, we will discuss the impact of the flaw polynyas on the hydrography of

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Table 1

<table>
<thead>
<tr>
<th>Polynya</th>
<th>Recurrence, %</th>
<th>Width, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eastern Severnaya Zemlya</td>
<td>89</td>
<td>68</td>
</tr>
<tr>
<td>Northeastern Taimyr</td>
<td>100</td>
<td>80</td>
</tr>
<tr>
<td>Taimyr</td>
<td>83</td>
<td>63</td>
</tr>
<tr>
<td>Anabar Lena</td>
<td>96</td>
<td>86</td>
</tr>
<tr>
<td>Western New Siberian</td>
<td>67</td>
<td>84</td>
</tr>
<tr>
<td>New Siberian</td>
<td>85</td>
<td>82</td>
</tr>
</tbody>
</table>
the Laptev Sea, taking into account historical oceanographic data collected by the Arctic and Antarctic Research Institute, St. Petersburg, Russia (AARI). The average winter salinity distribution and its variability is estimated for the period of 1979–1999. The impact of the coastal polynya on surface salinity, along with variable river runoff and atmospheric forcing, is discussed and ice production rates are estimated based on the polynya hydrography. Ice production can be obtained from the salinity distribution, which reflects the amount of brine rejection in the polynya. The rate of salinity adjustment in response to ice formation is evaluated statistically from time series of winter salinity observations. The probability of winter convective mixing penetrating down to the seafloor is evaluated statistically from the annual time series of surface and bottom salinity. It will be shown that both the vertical density stratification, mainly controlled by river runoff, and the position of the main polynyas determine the degree of convection penetrating down to the seafloor.

Results of the joint Russian–German expeditions carried out in the Eastern Laptev Sea in 1996 and 1999 are discussed in the second half of this paper. The rate of surface salinity increase from land-fast ice toward the polynya was determined through direct conductivity–temperature–depth (CTD) salinity measurements in spring of 1999. Preferred alignment of ice crystals in the ice cover along the fast-ice margin of the Laptev Sea coastal polynya was also observed and will be discussed in the context of circulation patterns forced by intensive brine rejection during ice formation.

2. Hydrography of the flaw polynya in the Laptev Sea: analysis of historical data

2.1. Data sets

We have used temperature and salinity records obtained in the Laptev Sea during the AARI “Sever” expeditions (1979–1990, 1992, and 1993), together with CTD-soundings performed during the Russian–German expeditions TRANSDRIFT IV (1996) and TRANSDRIFT VI (1999). The observation time varied from March until May. A total of 1685 hydrological stations was analyzed. The distribution of stations in the four principal sub-regions of the Laptev Sea is shown in Fig. 2.

Ten-day mean air temperature records from 1979 till 1993, for the Tiksi, Dunay and Kotelnyy meteorological stations were used to calculate the total amount of freezing degree-days. Station positions are shown in Fig. 3a.

The location of the fast-ice edge for February 11–20 each year from 1979 until 1999 (Fig. 3b) was obtained from AARI ice charts based on satellite images and airborne ice observations. In February the location of the fast-ice edge becomes nearly stationary and hardly changes at all till the end of May.

![Fig. 2. Distribution of oceanographic stations over the Laptev Sea. The panels (I)–(IV) represent the sub-regions marked by red squares in Fig. 3a.](image-url)
Fig. 3. Long-term average (1979–1999) salinity $S$ (a, bold solid line) and (b, dashed line) its variance $\sigma^2$ (psu$^2$) at the surface of the Laptev Sea in winter. Arrows on panel (a) are the directions of the summer surface circulation from (Dobrovolskiy and Zalozin, 1982). Thin solid line on panel (a) represents bottom topography. Dashed red line on panel (b) shows the average location of the fast-ice edge in mid-February. Grey solid lines correspond to locations of the fast-ice edge in mid-February from 1979 to 1999.
2.2. Salinity increase in the flaw polynya and ice production estimates

2.2.1. Approach

Intensive ice formation in the flaw polynya, accompanied by the rejection of brine, leads to an increase in the salinity of the surface water layer down to the depth of mixed layer density interface. The location of the flaw polynyas varies from year to year. The flaw polynyas themselves are not stationary, as they change their configuration and size depending on weather conditions resulting in a spatial variability of surface salinity during any given winter. Over longer interannual time scales, this spatial variability affects the variance of salinity. Local maxima in salinity variance are hence postulated to coincide with the longer-term mean location of the flaw polynya. This assumes that all other factors, such as interannual variability of thermohaline and wind-forced circulation, do not significantly contribute to interannual salinity variations.

Statistically the average surface salinity $S$ from a random-variable time series is determined as $S = S + \Delta s$, where $\Delta s$ is the confidence interval. For 20° of freedom at a 70% significance level $\Delta s$ corresponds to the root-mean-square salinity deviation (RMSD). Therefore, the average range of salinization, corresponding to the doubled RMSD value ($2\Delta s$), has been further considered to be the range of average salinization caused by ice formation in the polynya.

To reveal the impact of flaw polynyas on Laptev Sea hydrography, we studied the average winter surface salinity ($S$) and its variance ($\Delta s^2$) for the period from 1979 till 1999. The historical hydrographical data from 1685 stations were mainly obtained during the AARI airborne expeditions from March to May, i.e. conditions still representative of the winter regime. The average long-term salinity (Fig. 3a) was estimated by averaging the annual measurements, linearly interpolated to a regular grid with intervals of 0.5° of latitude and 2° of longitude. A 10 km search radius was employed in the interpolation procedure. The salinity variance $\Delta s^2$, presented in Fig. 3b, was calculated for each grid point using the interpolated value of yearly measured surface salinity. Given a rather uniform data coverage over most of the Laptev Sea, we assume that the obtained salinity field is representative of the actual conditions.

Due to intensive summer river runoff and ice melting, together with limited input of river water through the eastern Lena River channels during winter, the salinity of the southeastern Laptev Sea remains relatively low in winter. It increases northward from 12 to 14 practical salinity units (psu) near the Lena River delta to 28–30 psu offshore in the northeastern part of the sea. In the Western Laptev Sea salinity ranges between 30 and 32 psu (Fig. 3a). The variability of surface salinity is relatively low throughout the Laptev Sea, with the variance typically not exceeding 4–5 psu² (Fig. 3b). The highest $\Delta s^2$ was detected in the estuarine areas of the Yana, Olenek, and Buor-Haya Bays, and are attributed to interannual variations in winter river runoff.

The eastern part of the sea also exhibits high salinity variance, with the $\Delta s^2$ maximum coinciding with the average position of the fast-ice edge (Fig. 3b). This is also where according to Zakharov (1997) the Western New Siberian Polynya develops in March or later in the year (see Fig. 1 and Table 1). The northern boundary of the zone of increased $\Delta s^2$ ($\Delta s^2 > 3$) coincides approximately with the long-term average northern limit of the Western New Siberian Polynya (with a width of 75 km, Table 1). The southern boundary lies within the limits of interannual variation of the fast-ice edge (Fig. 3b).

River runoff is considerably reduced during winter-time. Hence, interannual variability of winter river discharge most likely only affects the area adjoining the river mouths, as evident in Fig. 3b. It is unlikely that the local zone of high $\Delta s^2$ values situated 300–400 km away from the river mouth is the result of river discharge interannual variability. Interannual variations in the surface circulation regime may be one cause of the surface salinity variance increase, especially in the vicinity of the land-fast-ice edge.

The large-scale salinity patterns are governed by the comparatively stable Lena discharge signal and do not vary as much from year to year. The wind-forced circulation is also subject to interannual variability, with the balance between the high sea-level pressure (SLP) center in the Western Arctic (the Siberian High) and the Icelandic Low mainly controlling atmospheric circulation over the Siberian Arctic (Johnson and Polyakov, 2001). During anticyclonic circulation
phases (Johnson and Polyakov, 2001), the Siberian High is well developed and the Icelandic Low is suppressed, with winds over the Laptev Sea tending towards the East Siberian Sea. During the cyclonic phases, the SLP high in the Western Arctic is weaker and the Icelandic Low is stronger, extending farther into the Barents and Kara seas. As a result, the prevailing wind over the Laptev Sea turns toward the Eurasian Basin of the Arctic Ocean (Johnson and Polyakov, 2001).

In order to evaluate the impact of atmospheric circulation on surface salinity, the National Center for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) Reanalysis data on SLP were analyzed. The winter (November–April) averaged vorticity index derived from monthly averaged SLP data has been applied to describe the atmospheric circulation over the Laptev Sea. The vorticity index, first employed to describe the circulation over the Arctic Ocean by Walsh et al. (1996), is the finite-differenced numerator of the Laplacian of SLP for the area within 550 km of 84°N and 125°E. The winter averaged vorticity index shows the intensity of wind vorticity in the Eurasian Low during winter (Fig. 4).

Although larger areas of the Eastern Laptev Sea are covered during winter by land-fast ice, the Lena River freshwater plume is nevertheless redistributed due to wind-forced sea level changes in the vicinity of the fast-ice edge (Dmitrenko et al., 2002). To examine the role of this factor, the correlation between interannual winter salinity variations and the winter averaged vorticity index was calculated for the same regular grid extended to the western East Siberian Sea. The calculation period covers all available winter surveys in the 1950s to 1990s for the Laptev and East Siberian seas.

Our results (Fig. 5) are in very good agreement with the basic conclusions of Nikiforov and Shpaiyher (1980), who postulated wind-driven redistribution of Siberian river runoff. Under the cyclonic regime (positive vorticity) the eastward diversion of Lena River water results in a negative salinity anomaly to

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Fig. 4. The time series of winter mean (November–April) vorticity index (mb) accounted according to Walsh et al. (1996). Bold curve corresponds to the polynomial smoothed data.

Fig. 5. Correlation between interannual winter surface salinity variations and winter averaged vorticity index for the period of 1950s to 1990s. Black bold curve corresponds to surface salinity RMSD maxima from (Polyakov et al., 2003). Thin solid line represents bottom topography.
the east of the Lena Delta and farther to the East Siberian Sea, and a positive anomaly to the north of the Lena Delta. Anticyclonic (negative) vorticity results in negative salinity anomalies northward from the Lena Delta due to freshwater advection toward the north, and a corresponding salinity increase eastward.

The relatively high correlation values (dark red color in Fig. 5) appear to delineate the areas in which interannual salinity variations are mainly wind-driven. In the Eastern Laptev Sea, this area is situated southward of the land-fast-ice edge in a sector associated with recurring polynyas and leads. Polyakov et al. (2003) published data on winter surface salinity variations for the same period of the 1950s to 1990s. A comparison of this data with the correlation field in Fig. 5 indicates that the area with high salinity variance is situated within the contour where the impact of atmospheric circulation on salinity variations is negligible (correlation coefficient ranges within ± 0.3). Therefore, we surmise that variability in the surface wind field is only of marginal importance for the generation of the maximum in salinity variance northward of the land-fast-ice edge.

A period of relatively uniform vorticity index values in the 1980s to 1990s (with mostly stable negative values from 1980 to 1995 as compared to earlier time periods, Fig. 4) also supports the assumption that salinity variations are driven primarily by ice production rates. In addition, the correlation of salinity and vorticity index suggests that the impact of winds on salinity variations is only of relevance inside the land-fast-ice edge.

Hence, the root-mean-square salinity deviation range \(2\Delta_s = 3.5–4.2 \text{ psu}\) is representative of long-term salinization, which is a consequence of ice formation in the Western New Siberian Polynya (Fig. 3b). This value exceeds the 2–2.5 psu suggested by Zakharov (1966) but is close to the measured values presented in the second part of this article. Estimates of 5.95–6.34 psu (Winsor and Björk, 2000) exceed our values probably due to the different approach of the latter authors. We did not consider advection of saline waters from the polynya and, hence, obtained a lower estimate of the possible salinity increase.

2.2.2. Method

The salinity increase in the polynya can be directly linked to the intensity of ice formation. Based on the \(\Delta_s\) values, the long-term average cumulative temperature below freezing (freezing degree-days, \(\theta\)), and on the average long-term temperature and salinity profiles in the region of the flaw polynya \([S, T = S, T(H)\), where \(H\) is the depth\], one can estimate the amount of ice produced in the polynya. In contrast with the method suggested here, which is based on the impact of ice formation on the surface salinity, most other estimates of ice formation are based on either direct field observations of ice growth in the polynyas (Kupetskiy, 1959; Zakharov, 1966), calculations of ice production rates as a function of meteorological conditions, supported by satellite observations (Martin and Cavalieri, 1989; Cavalieri and Martin, 1994; Dethleff et al., 1998; Winsor and Björk, 2000) or a combination of both (Drucker and Martin, 2003).

The basic goal is to estimate the rate of ice production in the polynya, which corresponds to the surface salinity variation of \(2\Delta_s\). Zubov (1963) assumed convective adjustment to the input of salt by ice formation. His simplest convective model, obtained with the assumption that the salinity of ice is zero, related the ice growth \(I\) with the salinity \(S\) of the convectively mixed layer:

\[
I_k = \sum_{i=1}^{k} \frac{1.1H_i(S_i - S_{i-1})}{S_{i-1}}
\]

\[
S_k = f(\rho_k, T)
\]

where the index \(i\) indicates the depth level in the vertical profile \((i = 1\) is at the sea surface\); \(H_i\) and \(S_i\) are the depth and salinity, respectively, at level \(i\); \(H_k\), \(S_k\), and \(\rho_k\) are the depth, salinity, and density, respectively, of the mixed layer; \(I_k\) is the ice thickness formed when mixing penetrates down to the level \(k\); and function \(f\) is the equation of state for seawater solved numerically for salinity. The coefficient 1.1 is the density ratio of ice and water.

To increase the accuracy of the estimates, both the winter and summer average long-term temperature and salinity profiles were used. The profiles were calculated from data obtained during summer (mainly August–September) and winter (mainly March–May) hydrographic surveys in the Laptev Sea between 1979 and 1999 (Fig. 2) by averaging the annual measurements, interpolated onto a regular grid. An example of the average long-term vertical temperature and sal-
Fig. 6. The long-term average vertical temperature and salinity profiles in the Western New Siberian Polynya and estimates of the ice thickness $I_k$ required to explain mixing of the upper water layer down to depth $H_k$; a, b—76° N, 132°30’ E; c, d—75° N, 130° E; a, c—summer profiles; b, d—winter profiles. Green and red curves show the vertical average multi-annual salinity $S(H)$ and temperature $T(H)$ distributions respectively. Blue curve is the ice thickness derived from Eqs. (1) (2) using $S(H)$ and $T(H)$. Estimates of $I_k$ and $H_k$ are also represented in Table 2.
An empirical relationship exists between the sea ice thickness in the Arctic and the freezing degree-days $\theta$ (Zubov, 1963):

$$ I = -25 + \left[ (25 + I_{ini})^2 - 8\theta \right]^{1/2} \quad (3) $$

Here, $I_{ini}$ is the initial ice thickness at the start of the calculation ($t=0$); the freezing degree-days $\theta$ are given by the cumulative temperature below freezing:

$$ \theta = \int_0^t (T_f - T_a) dt \quad (4) $$

where $T_f$ is the freezing point of water; $T_a$ is the air temperature; and $t$ is time. The long-term average for the freezing degree-days $\theta$ has been derived from the data of meteorological observations at Tiksi, Dunai Island, and Kotelnyi Island stations from 1979 to 1994. For further convenience $\theta$ is approximated by polynomials:

$$ \theta = -52.931650 + 2.269147d - 0.384070d^2 $$

$$ + 0.001622d^3 - 0.000002d^4 \quad (5) $$

$$ \theta = -56.047127 - 0.065949d - 0.347992d^2 $$

$$ + 0.001505d^3 - 0.000002d^4 \quad (6) $$

$$ \theta = -48.037763 - 0.649094d - 0.358000d^2 $$

$$ + 0.001572d^3 - 0.000002d^4 \quad (7) $$

Here $d$ is the ordinal number of the day starting from the average fall freeze-up date (October 1). The Eqs. (5)–(7) are for the meteorological stations Tiksi, Dunai Island, and Kotelnyi Island, respectively. The initial ice thickness $I_{ini}$ on October 1 was set to zero.

The second 10-day period in April was considered as the mid-point of the winter hydrographic observations. The ice thickness $I_o$ for this time was estimated from Eqs. (3)–(7).

Convective cooling without ice formation precedes the freezing phase, accompanied by increase in salinity due to brine rejection. Therefore, the average summer profiles $S,T=S,T(H)$ were recalculated according to Eq. (2) in order to adjust the temperature of the upper layer (up to the seasonal pycnocline) to the freezing temperature ($T=T_f$). Furthermore, we have estimated the salinity $S_o$ and depth $H_o$ of the upper convective mixed layer at ice thickness $I_o$ from adjusted summer long-term average temperature and salinity profiles (Fig. 6a and c; Table 2) based on Eqs. (1) (2). The salinity increase from $S_o$ up to $S_k$ caused by ice formation in the polynya was determined as $2A_s$. The total ice production in the polynya $I_k$ was derived from Eqs. (1) (2) according to:

$$ I_k = I(S_o + 2A_s) \quad (8) $$

Finally the polynya ice production $I_A$ is described by:

$$ I_A = I_k - I_o. \quad (9) $$

Formally, $I_A$ is the ice thickness needed to increase salinity $S_o$ of the upper convective mixed layer with thickness $H_o$ by $2A_s$. As this takes place, the thickness of the convective mixed layer grows up to the value $H_k$.

Fig. 6a and c, and Table 2 display the $S_o$, $S_k$, $I_o$, $I_k$, $H_o$, $H_k$, $I_A$ values calculated from the average summer multiannual temperature and salinity profiles for the Western New Siberian Polynya.

Average winter temperature and salinity profiles have mostly been obtained between April 11 and 20, and have hence already adjusted to the ice formation that had occurred prior to this date ($I_o=0$). Eq. (8) is applicable for winter profiles only, with $S_o$ specified

### Table 2

<table>
<thead>
<tr>
<th>Position</th>
<th>Season</th>
<th>$S_o$ (psu)</th>
<th>$S_k$ (psu)</th>
<th>$I_o$ (m)</th>
<th>$I_k$ (m)</th>
<th>$H_o$ (m)</th>
<th>$H_k$ (m)</th>
<th>$I_A$ (m)</th>
<th>Panel in Fig. 6</th>
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</thead>
<tbody>
<tr>
<td>76°N 132°30′E</td>
<td>summer</td>
<td>24.49</td>
<td>28.43</td>
<td>1.74</td>
<td>3.95</td>
<td>11.8</td>
<td>17.3</td>
<td>2.21</td>
<td>a</td>
</tr>
<tr>
<td>76°N 132°30′E</td>
<td>winter</td>
<td>26.66</td>
<td>30.60</td>
<td>0</td>
<td>2.11</td>
<td>0</td>
<td>20.5</td>
<td>2.11</td>
<td>b</td>
</tr>
<tr>
<td>75°N 130°E</td>
<td>summer</td>
<td>21.06</td>
<td>25.28</td>
<td>1.72</td>
<td>3.91</td>
<td>9.8</td>
<td>13.1</td>
<td>2.19</td>
<td>c</td>
</tr>
<tr>
<td>75°N 130°E</td>
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<td>1.32</td>
<td>0</td>
<td>12.6</td>
<td>1.32</td>
<td>d</td>
</tr>
</tbody>
</table>
as the average winter salinity at the sea surface. $2\Delta s$ is taken to be the salinity increase due to ice formation in the polynya. It hence corresponds to an increase of the convective mixing depth up to the value of $H_k$ (Fig. 6b and d; Table 2).

2.2.3. Results

Our calculations, performed separately for summer and winter data sets, indicate that the total ice production $I_k$ in the Western New Siberian Polynya varies from 3 to 4 m depending on the data set, and the geographical location of the individual points (Table 2). In the northwestern part of the Western New Siberian Polynya the ice production estimates derived from summer and winter profiles coincide quite well, deviating by only a few centimeters (Fig. 6a and b; Table 2). In the southeastern part of the polynya the differences are greater and can reach up to 0.9 m (Fig. 6c and d; Table 2). The substantial impact of winter river discharge in the southeastern sector of the polynya may partly explain these discrepancies. The hydrography-based estimates of total ice production in the polynya of 3 to 4 m are close to Zakharov’s (1966) estimate of 3.6 m based on in-situ ice observations and Zubov’s freezing degree-day model (Eq. (3)). Winsor and Björk (2000) have estimated Western New Siberian Polynya ice production as 4 km$^3$. At 3.4 km mean polynya width and a length of 83 km (Winsor and Björk, 2000) this corresponds to 14.2 m of ice production, which appears to be too high. The Winsor and Björk (2000) estimate is also contradicted by the hydrographic observations. If such amounts of ice were formed, convective mixing would penetrate down to the seafloor, which is not what has been observed in the field (see Section 2.3; Figs. 6b, d and 7).

The magnitude of $2\Delta s$ calculated from the long-term average data is smaller than the actual total salinization due to advection of saline water from the polynya, salt exchange with the underlying layers, and ice salinity exceeding the assumption of zero psu. With actual new ice salinities ranging around 4 psu, total ice production based on the hydrographic data should be considered a lower bound. Following Zubov (1963), the difference between 0 and 4 psu ice salinities were assessed for the ice production estimates shown in Table 2, indicating that actual ice production may be higher by as much as 15–17%, with a maximum estimate of around 5 m. The observations discussed in the second part of this article furthermore indicate that vertical exchange processes in the flaw polynya also have to be taken into account.

Fig. 7. Probability (%) of convective mixing penetrating down to the seafloor in the Laptev Sea. Dashed black line shows the average location of the fast-ice edge in mid-February.
2.3. Convection penetration down to the seafloor

The question of whether convective mixing in the polynyas can penetrate to the seafloor is an important, unsettled issue that is of prime importance for the analysis of dense water formation and seafloor sediment dynamics, in particular the potential resuspension of sediments due to convective action. The winter sea surface and the bottom salinity fields (for the time period from 1979 to 1999) were used to determine the probability of convective mixing penetrating down to the seafloor. In the calculations, a statistical normal distribution was assumed for the salinity at each location in all years. Penetration of convective mixing down to the sea floor results in an equalization of the surface and bottom salinity. Such a mixing event has been assigned probability \( P \). Mathematically, \( P \) can be expressed as the integral of the surface and bottom salinity probability density functions \( f_{\text{surface}} \) and \( f_{\text{bottom}} \) between the point of intersection \( S_0 \) (\( f_{\text{surface}}(S_0) = f_{\text{bottom}}(S_0) \)) and the maximum and minimum salinities, respectively:

\[
P = \int_{S_0}^{S_{\text{max}}} f_{\text{surface}} \, dS_{\text{surface}} + \int_{S_0}^{S_{\text{min}}} f_{\text{bottom}} \, dS_{\text{bottom}}
\]

The probability \( P \) calculated for the Laptev Sea is shown in Fig. 7. In general, it reflects the vertical stratification characteristics of the water column. Thus, low values of \( P \) (0 to 10%) in the Eastern Laptev Sea and in the vicinity of the river mouths reflect strong stratification due to river runoff. In the east, only the Western New Siberian Polynya exhibits values of \( P \) as high as 20%, with the southern boundary of this zone located within the boundaries of the interannually varying fast-ice edge (Fig. 3b). Due to weak vertical density stratification in the Western Laptev Sea, the probability of convection penetrating to the seafloor is 2–4 times higher and may exceed 70%. The zone of the highest convection probability is located in the area of the Anabar Lena Polynya. Winter surveys of the “Sever” expedition have repeatedly recorded the presence of cold, saline bottom waters formed in this Anabar Lena sector of the Laptev Sea coastal polynya (Churun and Timokhov, 1995).

3. Circulation patterns in the Laptev Sea flaw polynya: analysis of the data from the 1999 winter expedition

3.1. Measurements and data sets

3.1.1. CTD, current and meteorological observations

Oceanographic investigations in the region of the Anabar Lena and Western New Siberian polynyas in the Laptev Sea were carried out in April–May 1999 with the help of helicopters within the framework of a joint Russian–German project “Laptev Sea System 2000”. The study area of the TRANSDRIFT VI helicopter expedition is shown in Fig. 1. In order to study oceanographic processes in the flaw polynya, oceanographic stations TI9908, 7/22, 18/19, 14, 02/06/23, and 24 were occupied directly at or close to the fast-ice edge (Fig. 8a). At stations TI9902/23, 7, and 8, hydrographic measurements were conducted along transects of 500–600 m length, perpendicular to the fast-ice edge and extending over the fast ice, open water, and young ice. Oceanographic data were obtained with a ME (Meerestechnik Electronik) OTS-3 CTD sonde (manufactured by Meerestechnik Electronik GmbH, Germany) under the fast ice and with a Seacat SBE 19 CTD (manufactured by Sea-Bird Electronics, Inc., USA) deployed from a rubber boat in open water, or directly under thin new ice. Currents were measured with an acoustic Doppler current meter (3D-ACM, FSI, USA).

Routine current observations were only carried out from stationary ice, with the instrument lowered through a hole 22 cm in diameter. Starting at the ice bottom, measurements were taken for 3–4 min each at 2–3 m vertical spacing down to the seafloor. At each depth level, data were averaged from 6 to 8 sets of 30-s mean measurements. Current observations were computed at 6 stations (TI9918, 19, 20, 22, 23, and 24, Fig. 8a). Stations TI9918 and 19 were 2.5 km apart from each other, the first on 108 cm thick fast ice, the other on grey young ice 24 cm thick at a distance of 5 m from open water. TI9922 was located 100 m and TI9924 6 km away from the fast-ice edge. Station TI9920 was located on the fast ice at a distance of about 120 km from the flaw polynya.

The mean daily air temperature at the Tiksi and Kotelny meteorological stations from October 1, 1998 to May 10, 1999 was used to estimate the
freezing degree-days based on Zubov’s (1963) empirical relationship (Eqs. (3) and (4)).

3.1.2. Ice crystal analysis

At stations TI19901, 2, 3, 4, 7, 8, 9, 11, 16, 17, 18, and 24 (Fig. 8a), ice cores were obtained close to the hydrographic measurement site for an analysis of crystal microstructure and orientation. Azimuthally oriented cores were cut into segments of about 20 to 30 cm length. From each segment, one or two vertical and one horizontal thin section (less than 1 mm thick, 9 × 12 cm wide and high) were prepared. From an inspection of thin sections between crossed polarizing filters, we derived the texture, size, and the azimuthal orientation of the crystal basal planes. While the horizontal sections indicated whether crystals were azimuthally aligned, the vertical sections, obtained between two adjacent horizontal sections helped identify the depth at which alignment set in.

3.2. Evolution of the flaw polynya in the Eastern Laptev Sea during Late winter 1998–1999

As indicated by satellite imagery (Radarsat Scanning Synthetic Aperture Radar [ScanSAR]),

Advanced Very High Resolution Radiometer data in the infra-red range and passive microwave data) the system of flaw polynyas along the Laptev Sea fast-ice edge was not very well developed or was absent prior to the mid-March during the winter of 1998–1999. Open water occurred only in a chain of separate, narrow flaw leads until the second half of March, as exemplified by the Radarsat ScanSAR image of March 21 (Fig. 9a), in which a flaw polynya is completely absent. Kotelny Island meteorological data indicate that southeasterly winds increased up to 16 m/s after March 25, continuing up until March 29 (Fig. 10a). Later, winds decreased to 5–7 m/s, but their direction remained unchanged until April 7. The opening of a polynya (reaching up to 10 km in width) and new ice formation (over a zone 20–23 km wide) are clearly evident in the satellite scenes of April 5 (Fig. 9b). On April 12 southeasterly winds of up to 12–16 m/s picked up again, continuing until April 17 (Fig. 10a). This induced further widening of the polynya area, and by April 25 its width had increased to 90 km in some locations (Fig. 9c). With new ice thickening in this region, another opening event occurred from May 4 to 7, with southerly winds blowing at up to 16 m/s, creating an area of open
water and new ice up to 35 km wide (Fig. 9d). It was these repeated, strong southeasterly winds that fostered the opening of the polynya and widespread formation of young ice between late March and early May, during a period of overall low air temperatures and subsequently high ice formation rates (Fig. 10b).

3.3. Hydrography of the flaw polynya and the adjacent fast ice

The impact of the flaw polynya on the salinity distribution in the Eastern Laptev Sea has been discussed in Section 2.2. According to Zakharov (1966), the surface salinity is higher by about 2 to 2.5
psu (up to a maximum differential of 6 psu) in the polynya region. Our analysis has shown that the long-term mean increase of salinity in polynyas of the Eastern Laptev Sea varies between 3.5 and 4.2 psu. A system of meso-scale currents should develop in association with the polynya due to such local salinity variations. Information about currents in the Laptev Sea in winter is restricted to a few, episodic observations that are inadequate to support conclusions about the specifics of the winter surface water circulation, in particular in the narrow polynya zone. The descriptions provided in a number of summary articles are based more on general geographical knowledge than on systematic observations (Dobrovolskiy and Zalogin, 1982; Pavlov et al., 1996). Thus, the alongshore southward current from the northwest has been described as turning farther to the northeast, contributing to the cyclonic surface circulation (Fig. 3a). The main features of the thermohaline and wind-driven surface circulation can be obtained from hydrodynamic models [for example, Pavlov and Pavlov (1999)]. These models indicate that the circulation is mostly controlled by the large-scale surface salinity distribution (Fig. 3a), with winter current velocities generally below 5–8 cm/s. The geostrophic currents in the area northward of the Lena River Delta are directed to the north, dominating the offshore winter circulation regime in this area. The local salinity increase in the polynyas and its impact on surface flow has not been taken into account in any of the studies referred to above, however.

Salinity increases due to ice formation in the polynya measured along short transects across the fast-ice edge varied between 1 psu (stations TI9902, Fig. 11a, and TI9908) and 5 psu (station TI9923, Fig. 11b). At the stations closest to the fast-ice edge (100 m to 2 km distance) sub-ice currents were recorded in the surface water layer down to the pycnocline, flowing towards the open water of the flaw polynya (stations TI9918, TI9923a, and TI9924, Figs. 8b and 11b). The velocity of these currents varied from 10 to 24 cm/s. In the underlying layer, current direction (except at station TI9918) changed by 90–180°. A maximum current velocity of 62 cm/s, directed to the northeast (26°), was measured in the sub-ice layer directly at the fast-ice edge at station TI9923. At a depth of 4.6 m, directly underneath the pycnocline, the current was setting in almost the opposite direction (239°) at a velocity of 26 cm/s (Fig. 11b). The direction of 26° was normally oriented to the local position of the fast-ice edge. During measurements at station TI9923 the weather was relatively calm, with winds below 5 m/s from the northeast. A similar situation was observed at the fast-ice edge at St. TI9922, where a current of 5.9 cm/s was flowing northwestward (320°) towards open water at a depth of 3.8 m. In the pycnocline at 8.5 m depth, a near-opposite current (115°) with a velocity of 5.8 cm/s was registered. Below the thin new ice, the sub-ice currents were homogeneous from the surface to the pycnocline depth, and opposite to the sub-ice currents below the fast-ice edge (stations TI9919 and TI9923d, Fig. 11b). Thus, at the fast-ice edge, two jets constitute the current system in the polynya, with water in the sub-ice layer directed towards the polynya and flow beneath the pycnocline in the opposite direction.

A bottom-moored upward-looking Acoustic Doppler Current Profiler (ADCP) has been deployed in September 1998 for a 1-year period in the eastern part of...
of the Laptev Sea (Fig. 8a). During April–May 1999 the mooring was situated in the vicinity of the fast-ice edge near station TI2499 (Fig. 8a). The ADCP observations have shown that the semidiurnal frequency band of currents appears to be dominated by tides (Dmitrenko et al., 2002). During the ice-covered period, a strong baroclinic internal tide with maximum current of 20 cm/s at the density interface was recorded. In general, however, the maximum tidal current (lunar semidiurnal tide M₂) did not exceed 5–6 cm/s (Dmitrenko et al., 2002) during the time interval of the polynya’s existence. The tidal energy was rather uniformly distributed with depth. Therefore the distribution of currents observed at station TI9923 could not be explained by the baroclinicity of the internal tide.

3.4. Ice texture, crystal orientation and circulation patterns in the flaw polynya and the adjacent fast ice

It has been shown in several studies that consistent sub-ice currents determine the azimuthal crystal alignment in a sea-ice cover (Weeks and Gow, 1978, 1980; Stander and Michel, 1989; Strakhov, 1989). The dependence of crystal alignment on the current regime, hypothesized based on field measurements, has been verified in laboratory experiments (Cherepanov and Strakhov, 1989). Hence, the data on crystal alignment in fast-ice cores collected as part of this study can provide insight into the under-ice currents during the period of ice growth.

Thin-section analysis revealed an azimuthal alignment of ice crystals at stations TI9901, 2, 4, 8, 16, 18, and 23 (Figs. 8b and 12). At stations TI9901, 2, 4, 8, 16, and 18, crystal C-axes were generally trending in a southeasterly to northwesterly direction, whereas at station TI9924 the alignment was south to north. Based on the stratigraphic analysis of horizontal and vertical thin sections, we were able to determine the approximate level of the onset of crystal alignment (as exemplified in Fig. 12 for station TI9902). To convert the depth of crystal alignment onset to an actual date, we computed the change in ice thickness with time based on the empirical relationship expressed in Eqs. (3) and (4), forced by daily temperature data from the meteorological stations on Dunay Island (Eq. (11)) and Kotelnuy Island (Eq. (12)) during the winter of
1998–1999. The age of ice \((d)\) at depth level was then derived based on the freezing degree-days \(\theta\) according to the polynomial approximations:

\[
\theta = -21.028120 - 5.433169d - 0.220287d^2 \\
+ 0.000554d^3 - 0.0000002d^4
\]

\(\theta = -20.366962 - 7.805000d - 0.164232d^2 \\
+ 0.000223d^3 + 0.0000008d^4
\)

The dates of crystal alignment onset that have been determined based on this method are listed in Table 3. Note that a more detailed analysis with a multi-layer ice-growth model taking into account radiative and other forcing (including measured snow depth) yields only slightly different dates (Eicken et al., this volume).

At all stations except TI9903, the crystal alignment in the ice cores confirms the general geostrophic surface circulation regime. Both the direction of alignment and the surface current measurements at stations TI9918, 22, and 23 correlate well with the surface circulation obtained from hydrodynamic modeling by Pavlov and Pavlov (1999). However, as is evident in Table 3, crystal alignment at the fast-ice margin stations TI9902 (Fig. 12), 8, and 18 only sets in at the end of March. At the stations close to the flaw polynya (TI9903 and 24) the onset of alignment occurred in mid- or late February. At those locations separated by a greater distance from the flaw polynya (TI9901 in fast ice and TI9904 in drifting ice), alignment commenced much earlier in the year. Hence, crystal alignment in the lower part of relatively thick fast ice (TI9903, 09, 10, and 11) is controlled by different physical forces than alignment in young ice formed adjacent to the flaw polynya. The onset of alignment along the fast-ice margin (Table 3, March 24–26) coincides with the intensification of south-easterly winds on March 25. The latter induced expansion of the flaw polynya (Figs. 9a and b, 10a). This suggests that crystal alignment near the fast-ice edge is due to the establishment of persistent sub-ice currents oriented towards open water in the flaw polynya. In the remaining cases studied here, alignment is most likely controlled by the geostrophic or tidal circulation patterns.

Persistent sub-ice surface currents that are oriented towards open water adjacent to an ice cover have been observed repeatedly in leads in the Arctic ice cover during winter (Smith et al., 1990; Morison et al., 1992). Smith (1973) was the first to explain that these
currents are the result of circulation cells induced by an increase in surface layer salinity due to ice formation. Thus, the surface circulation is driven by salt rejection from the growing ice sheet, resulting in off-ice flow of the upper water layers and a counter-current developing below the quasi-homogeneous sub-ice layer that sets towards the ice cover (Fig. 13). Velocities of such currents have been observed to vary between 2 and 12 cm/s (Morison et al., 1992). Smith and Morison (1993) derived the circulation pattern under leads and their impact on the hydrography of the Beaufort Sea. For the case of free convection the resulting current velocity was estimated at 8 cm/s, with convective cells approximately 1000 m in lateral extent.

The general character of the circulation and the density structure measured along the short transects at the fast-ice margin corresponds closely to those calculated by Smith and Morison (1993) for the case of free convection forced by salt rejection during ice growth. Hence, the surface currents are most likely due to salinization of the surface water caused by ice formation in the flaw polynya, resulting in the circulation patterns and crystal alignment discussed above (see also Fig. 13).

4. Conclusions

1. The persistence of a quasi-stationary system of flaw polynya is one of the main factors responsible for the variability of the surface hydrography in the Eastern Laptev Sea. The salinity increase in the upper water layer caused by ice production in the Western New Siberian Polynya by mid-April corresponded closely to the double root-mean-square deviation from the average long-term surface salinity ($2\sigma_s$) in the Laptev Sea in winter, amounting to 3.5–4.2 psu. These estimates were confirmed by salinity measurements in the polynya in April 1999.

2. A new method to evaluate the long-term average annual ice production in the flaw polynya has been described, based on hydrographic observations rather than sea-ice data. The extent and depth of convective mixing during winter has been calculated based on long-term average temperature and salinity profiles (winter and summer). The total ice growth required to increase the long-term average surface salinity by $2\sigma_s$ has been taken to correspond to the ice production in the flaw polynya. The value of 3–4 m total ice production by mid-April agrees well with estimates based on direct ice observations but is much smaller than results of model calculations based on satellite data.

3. The probability of winter convective mixing penetrating down to the seafloor has been linked to the distribution of river runoff in the Eastern Laptev Sea and the location of the main flaw polynyas. In the Eastern Laptev Sea penetration of convection down to the seafloor is most probable in the Western New Siberian Polynya (20%), whereas in the Western Laptev Sea the probability is highest in the Anabar Lena Polynya (70%).

4. Preferred alignment of ice crystals in the ice cover along the fast-ice margin of the Anabar Lena and Western New Siberian Polynya was identified during ice studies in the spring of 1999. This alignment is most likely due to the development of quasi-stationary circulation cells resulting from convergent flow in the polynya, forced by the salinization of surface waters as a result of ice growth. The surface currents are directed towards the open water of the polynya and can reach velocities as high as 62 cm/s. In the underlying water layer, a countercurrent is established, thereby promoting the turbulent exchange of heat and mass with the underlying ocean.

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