

# Variability and trends in Laptev Sea ice outflow between 1992 - 2011

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**Abstract.** Variability and trends in seasonal and interannual ice area export out of the Laptev Sea between 1992 and 2011 are investigated using satellite-based sea ice drift and concentration data. We found an average total winter (October to May) ice area transport across the northern and eastern Laptev Sea boundaries (NB and EB) of  $3.48 \times 10^5$  km<sup>2</sup>. The average transport across the NB ( $2.87 \times 10^5$  km<sup>2</sup>) is thereby higher than across the EB ( $0.61 \times 10^5$  km<sup>2</sup>), with a less pronounced seasonal cycle. The total Laptev Sea ice area flux significantly increased over the last decades ( $0.85 \times 10^5$  km<sup>2</sup>/decade,  $p > 0.95$ ), dominated by increasing export through the EB ( $0.55 \times 10^5$  km<sup>2</sup>/decade,  $p > 0.90$ ), while the increase in export across the NB is smaller ( $0.3 \times 10^5$  km<sup>2</sup>/decade) and statistically not significant. The strong coupling between across-boundary SLP gradient and ice drift velocity indicates that monthly variations in ice area flux are primarily controlled by changes in geostrophic wind velocities, although the Laptev Sea ice circulation shows no clear relationship with large-scale atmospheric indices. Also there is no evidence of increasing wind velocities that could explain the overall positive trends in ice export. The increased transport rates are rather the consequence of a changing ice cover such as thinning and/or a decrease in concentration. The use of a back-propagation method revealed that most of the ice that is incorporated into the Transpolar Drift is formed during freeze-up and originates from the central and western part of the Laptev Sea, while the exchange with the East Siberian Sea is dominated by ice coming from the central and south-eastern Laptev Sea. Furthermore, our results imply that years of high ice export in late winter (February to May) have a thinning effect on the ice cover, which in turn preconditions the occurrence of negative sea ice extent anomalies in summer.

## 1 Introduction

In recent years, the summer Arctic sea ice extent and thickness have undergone dramatic changes. The total sea ice extent is declining at an annual rate of approximately 3 % per decade over the satellite record (1978 - present), and the summer sea ice decline seems to be accelerating (Kwok et al., 2009; Kwok and Rothrock, 2009; Comiso, 2010; Stroeve et al., 2011). Following Kwok and Rothrock (2009), the thickness of sea ice decreased by 64 % (1.6 m) between 2003 and 2008 compared to early submarine measurements that were made between 1958 and 1976. The rapid reduction in Arctic summer ice extent and thickness is assumed to result from anomalously high surface air temperatures (Stroeve et al., 2005) and changes in the large-scale atmospheric circulation (Meier et al., 2007).

Climate models agree that the sea ice extent and thickness will further decline through the 21st century in response to atmospheric greenhouse gas loading (Zhang and Walsh, 2006; Stroeve et al., 2012; Massonnet et al., 2012). Furthermore, ice drift and deformation increase and net ice growth rates decrease (Spren et al., 2011; Rampal et al., 2009). To determine associated changes in the Arctic sea ice volume requires consideration of changes in ice volume fluxes that appear at the major gates of the Arctic, such as Fram Strait. A positive trend in Fram Strait sea ice export was found first time by Smedsrud et al. (2011). The authors used geostrophic winds derived from reanalysis data to calculate the ice area export between Spitsbergen and Greenland and estimated it to be about 25 % larger than during the 1960's. Whether Fram Strait volume fluxes are decreasing, or if the reduction in sea ice thickness is capable of compensating for the observed increased ice export out of the Arctic is currently under discussion.

Understanding changes that appear at the major flux gates requires a closer look at the origin of the ice leaving the Arctic Ocean. It is assumed that a substantial part of the

70 Fram Strait export is formed in the Siberian shelf seas, con- 125  
 nected with Fram Strait via the Transpolar Drift. Among  
 the Siberian shelf seas the Laptev Sea is considered one of  
 the most significant regions of net ice production and export  
 (Zakharov, 1966a; Dethleff et al., 1998). It is very shallow  
 75 with water depths between 15 and 200 m (Timokhov, 1994)  
 and comprises an area of approximately  $500 \times 10^3 \text{ km}^2$ . The  
 Laptev Sea is located between the coast of Siberia, Severnaya 130  
 Zemlya and the New Siberian Islands (Fig. 1), and is ice cov-  
 ered from October to June (Bareiss and Goergen, 2005). The  
 80 ice cover can be divided into three regimes: the fast ice, the  
 pack ice, and flaw polynyas (Eicken et al., 2005). The freely  
 floating ice pack offshore the fast ice edge consists mainly  
 of ice formed during fall. According to systematic obser- 135  
 vations carried out by the Soviet Union since the 1930s, it  
 reaches a mean thickness of  $1.57 \text{ m} \pm 0.25 \text{ m}$  (Romanov,  
 1996). The pack ice drift is dominated by persistent offshore  
 winds leading to a continuous export of ice out of the Laptev  
 Sea into the basin and/or the East Siberian Sea (Timokhov,  
 1994; Rigor and Colony, 1997). 140

90 Rigor and Colony (1997) found by means of a combina-  
 tion of modeling results and observations that as much as  
 20 % of the ice transported through Fram Strait is produced  
 in the Laptev Sea, giving it a key role in the future fate of  
 the Arctic sea ice. The seasonal and interannual variability  
 95 of sea ice exchange with the surrounding seas was first ex-  
 amined by Zakharov (1966b, 1967). Following Alexandrov  
 et al. (2000), their estimates are based on average monthly  
 gradients of atmospheric pressure across the northern and  
 north-eastern Laptev Sea boundaries and revealed a mean  
 100 winter sea ice area transport of approximately  $3.3 \times 10^5$   
 $\text{km}^2$ . Alexandrov et al. (2000) investigated ice area fluxes by  
 means of a numerical model and found flow rates of  $4.83 \times$  150  
 $10^5 \text{ km}^2$  per winter. In addition a number of studies examine  
 sea ice circulation patterns and their linkage to atmospheric  
 and oceanic forcing on shorter time scales (e.g., Rigor and  
 Colony, 1997; Eicken et al., 1997; Haas and Eicken, 2001). 155

Given the importance of the Laptev Sea for the Arctic  
 Ocean sea ice budget, the aim of this paper is to update pre-  
 vious estimates of ice exchange between the Laptev Sea and  
 110 the surrounding seas. Since we are lacking of ice thickness  
 information, the investigation is limited to the determination  
 of area fluxes derived from state-of-the-art satellite ice motion  
 and ice concentration information. Below we examine how  
 the ice export in the Laptev Sea changed throughout the past  
 115 two decades and identify the governing mechanisms control-  
 ling ice exchange with the surrounding seas. Furthermore, 165  
 we investigate the role of atmospheric forcing for observed  
 trends in ice transport.

The structure of the paper is as follows. A description of  
 120 the satellite data used to derive ice fluxes along the Laptev  
 Sea boundaries is given in section 2. In addition, the ac-  
 curacy of the dataset is determined through a comparison  
 with high resolution satellite data and mooring observations.  
 In section 3, we analyze seasonal and interannual flux rates

across boundaries and identify the origin of sea ice leaving  
 the Laptev Sea in late winter. In section 4 we discuss re-  
 sults and relate fluctuations in winter ice export to changes in  
 atmospheric circulation patterns, as well as to sea ice anom-  
 alies in summer. Conclusions are drawn in section 5.

## 2 Dataset description and validation

The ice area flux out of the Laptev Sea into the Transpolar  
 Drift and East Siberian Sea is calculated based on ice drift ve-  
 locities and ice concentration at the northern boundary (NB)  
 and eastern boundary (EB) of the study region (Fig. 1). The  
 NB spans a length of 700 km and is positioned at  $81^\circ\text{N}$ , be-  
 tween Cape Arkticheskiy and  $140^\circ\text{E}$ . The EB with a length  
 of 460 km, connects the eastern end of the NB with Kotelnyy  
 ( $76.6^\circ\text{N}$ ,  $140^\circ\text{E}$ ). For easier comparison, our northern and  
 eastern boundaries are equivalent to those boundaries used  
 by Alexandrov et al. (2000).

### 2.1 Sea ice drift dataset

Sea ice drift can be estimated using various techniques be-  
 tween time-lag satellite images (e.g., Emery et al., 1991;  
 Kwok and Rothrock., 1999; Girard-Ardhuin and Ezraty,  
 2012). Ice drift information obtained from satellites has  
 been applied in various studies to examine Arctic-wide spa-  
 tial trends in sea ice drift velocity and area flux (e.g., Kwok,  
 2009; Spreen et al., 2011). In particular during winter  
 months, when the atmospheric moisture content is low and  
 surface melt processes are absent, the quality of the drift es-  
 timates is high. Restrictions may arise from the spatial reso-  
 lution of the sensors in near-shore regions characterized by a  
 complex coastline, extensive fast ice areas and polynyas.

The accuracy of ice drift data in the Laptev Sea was in-  
 vestigated by Rozman et al. (2011) through a comparison  
 with in-situ measurements. The best performance was found  
 in the ice drift product provided by the European Space  
 Agency (ESA) via the Center for Satellite Exploitation and  
 Research (CERSAT) at the Institut Francais de Recherche  
 pour d'Exploitation de la Mer (IFREMER), France. Hence,  
 in the following the IFREMER data set is used to calculate  
 winter fluxes at the NB and EB. The motion fields are avail-  
 able on an operational level from September 1 until the end  
 of May, covering the period from 1992 to present. They are  
 based on a combination of drift vectors estimated from scat-  
 terometer data (the National Aeronautics and Space Admin-  
 istration SeaWinds/QuikSCAT for 1992-2009 period, and  
 the European Advanced Scatterometer (ASCAT)/MetOp for  
 2009 - present) and radiometer data (the 85 GHz channel data  
 of Special Sensor Microwave Imager SSM/I on-board the  
 Defense Meteorological Satellite Program, available since  
 1992). The data are available on a polar stereographic grid  
 with a grid size of 62.5 km, using time lags of 3 days. Details

about processing and validation of these data can be found in  
175 Girard-Ardhuin and Ezraty (2012).

## 2.2 Accuracy of sea ice drift data

As an additional quality control we compare the IFREMER  
180 motion estimates with monthly ice drift information obtained  
from Environmental Satellite (ENVISAT) Synthetic Aper-  
ture Radar (SAR) images (Fig. 2). Ice drift information from  
ENVISAT SAR scenes can be easily extracted by identify-  
ing identical ice flows on consecutive images. In this study,  
185 monthly ice drift information from ENVISAT Wide Swath  
(WS) scenes were obtained in areas near the NB and EB  
from images covering the beginning and end of April 2004,  
December and January 2007, January to May and December  
2008, February and December 2009, and February to March  
2011. The ENVISAT C-band WS data is VV polarized and  
covers an area of approximately  $400 \times 400 \text{ km}^2$  with a spa-  
190 tial resolution of  $150 \times 150 \text{ m}^2$  (Krumpen et al., 2011b,a).  
Overall 12 monthly ice drift estimates are available. The  
comparison with the IFREMER dataset was done by interpo-  
lating IFREMER estimates to the locations of SAR ice drift  
retrieval. The agreement between IFREMER data and man-  
195 ually extracted ENVISAT ice drift information is high for  
both, the zonal ( $U$ ) and meridional ( $V$ ) ice drift components  
(correlation coefficient ( $R$ ) = 0.87 and 0.95, Fig. 2). The  
estimated standard deviation (SE) for the IFREMER ice drift  
velocity is 0.56 cm/s for  $V$  and 0.6 cm/s for the  $U$  drift com-  
200 ponents. For comparison, the uncertainty in ice drift velocity  
reported by Rozman et al. (2011) is around 1 cm/s for the  
Laptev Sea.

As stated earlier, the accuracy of passive microwave drift  
products may be reduced in near-shore areas. However,  
205 in section 3 we identify the origin of sea ice leaving the  
Laptev Sea in late winter by means of a backtracking ap-  
proach. Therefore the quality of the IFREMER data in shore  
areas is checked through a comparison with ice drift esti-  
mates taken from long-term moorings equipped with Acous-  
tic Doppler Current profilers (ADCP). The moorings were  
210 deployed between 2007 and 2009 near the fast ice edge in  
the south-eastern Laptev Sea (for exact position see Fig. 1).  
The sampling frequency of the device is 30 minutes with a  
beam width (footprint) of 5 m. An in-depth description of  
the mooring design and ADCP processing is given in Hoele-  
mann et al. (2011) and Janout et al. (2012). Prior compar-  
215 ison, the monthly mean ice drift was calculated from the ADCP  
data, and IFREMER drift data was interpolated to the moor-  
ing position. The correlation between IFREMER and ADCP  
data is slightly lower (Fig. 2,  $R = 0.84$  and  $0.9$  for  $U$  and  $V$ )  
with a higher standard deviation (0.72 cm/s for  $V$  and 0.83  
220 cm/s for  $U$ ). The limited number of samples included in this  
comparison and the differences in spatial and temporal sen-  
sor resolutions may impact the reliability of the comparison.  
225 In addition, the presence of fast ice, the complex coastline  
and the occurrence of polynyas are restricting factors. Nev-

ertheless, the high agreement shows that IFREMER data is  
capable of producing ice drift in near-coastal Laptev Sea ar-  
eas correctly.

## 2.3 Sea ice concentration

The ice concentration data used in this paper are also made  
available by IFREMER (Ezraty et al., 2007). The product is  
based on 85 GHz SSM/I brightness temperatures, using the  
ARTIST Sea Ice (ASI) algorithm developed at the University  
of Bremen (Spreen et al., 2008b). The use of the 85 GHz  
channels data enables to have sea ice concentration at 12.5  
km x 12.5 km pixel resolution.

## 2.4 Sea ice area flux estimates at the Laptev Sea bound- aries

Following Kwok (2009), the meridional and zonal ice area  
flux at the NB and EB respectively, is the integral of the  
product between the  $V$  and  $U$  component of the ice drift and  
ice concentration. Initially, ice motion vectors were interpo-  
lated to the 12.5 km ice concentration grid. In the following,  
a positive (negative) sign refers to an export out of (import  
into) the Laptev Sea. Transport (flux) rates are given in  $\text{km}^2$ .  
The sum of the meridional and zonal ice area flux across NB  
and EB is referred to as the total ice flux. After removing  
the seasonal cycle, trends were calculated in a least square  
sense, and significance at the 95 % confidence level ( $p$ ) was  
measured using the Student's t-test following Kwok (2009).  
Please see Kwok and Pang (2004) and Kwok (2009) for a  
more detailed method description and error analysis.

## 3 Results

We present ice flux estimates at the northern and eastern  
Laptev Sea boundaries computed from IFREMER ice drift  
information. In order to assess the relative consistency and  
to quantify the importance of Laptev Sea ice export for the  
total Arctic sea ice budget, the results are compared with flux  
estimates from NSIDC drift data (Fowler, 2003) and export  
rates through Fram Strait (Smedsrud et al., 2011). In the fol-  
lowing, we analyze the interannual and seasonal variability  
of meridional and zonal area fluxes through NB and EB, and  
relate the observed fluctuations to changes in sea level pres-  
sure (SLP) gradients across the boundaries. Finally, the ori-  
gin of the sea ice contributing to the annual export is exam-  
ined through a backtracking approach of sea ice leaving the  
Laptev Sea in early spring.

### 3.1 Laptev sea ice area flux

Figure 3 compares the cumulative winter (October to May)  
total ice area flux from 1992 - 2011 through the NB and  
EB calculated from IFREMER ice drift and concentration

data with estimates based on NSIDC drift data and IFREMER ice concentration information (see chapter 2.3). The ice drift data, provided by the NSIDC (Fowler, 2003) are computed from Advanced Very High Resolution Radiometer (AVHRR), SSM/I and IABP buoy data. Information is given at a higher spatial resolution (25 km grid) but is only available until 2010. Figure 3 shows that NSIDC flux estimates are lower than those from IFREMER ( $2.91 \times 10^5 \text{ km}^2 \pm 1.13 \times 10^5 \text{ km}^2$  vs.  $3.39 \times 10^5 \text{ km}^2 \pm 1.17 \times 10^5 \text{ km}^2$ ). A comparison of the NSIDC dataset with the SAR and ADCP derived drift information (see chapter 2.2) reveals that NSIDC drift vectors generally underestimate observed drift velocities (not shown here). This is in agreement with findings made by e.g. Schwegmann et al. (2011) in the Antarctic. According to the authors, 71 % of the NSIDC drift velocities are significantly lower than those observed by buoys. Nevertheless, the high correlation between NSIDC and IFREMER data ( $R = 0.82$ ) gives confidence about the relative consistency of the IFREMER data.

The interannual variability of the total ice transport is shown together with the zonal and meridional component in Figure 4. The net total winter transport is positive, but estimates show considerable interannual variability with the lowest rates occurring in winters of 1997/98, 1998/99 and 2003/04 and the highest export in the winters of 2004/05 and 2008/09. The average total winter ice flux amounts to  $3.48 \times 10^5 \text{ km}^2$  with a standard deviation ( $SE$ ) of  $1.2 \times 10^5 \text{ km}^2$ . The average transport across EB ( $0.61 \times 10^5 \text{ km}^2 \pm 0.97 \times 10^5 \text{ km}^2$ ) is generally lower than that across NB ( $2.87 \times 10^5 \text{ km}^2 \pm 0.78 \times 10^5 \text{ km}^2$ ). However, with respect to the length of our boundaries and mean flux, the zonal transport exhibits larger fluctuations, with years being characterized by a net import from the East Siberian Sea (2007/08, 2000/01, 2003/04 and 2007/08). Note that there is a statistically significant positive trend in the total Laptev Sea ice area flux across the boundaries of  $0.85 \times 10^5 \text{ km}^2/\text{decade}$  ( $p > 0.95$ ). The trend in zonal export across EB is smaller (and  $0.55 \times 10^5 \text{ km}^2/\text{decade}$ ) and significant only at the 90 % confidence level. The positive trend of the meridional export ( $0.3 \times 10^5 \text{ km}^2/\text{decade}$ ) across NB is not statistically significant. Our Laptev Sea ice export estimates amount to nearly 48 % of the winter (Fig. 4) and 41 % of the annual export rates through Fram Strait. Fram Strait area flux estimates were taken from Smedsrud et al. (2011). Different to our approach, the authors derived ice area transport rates from radar satellites and SLP differences. Note that no significant correlation was found between Fram Strait and Laptev Sea outflows ( $R = 0.32$ ).

The monthly averages of the total ice area flux (October to May) show a pronounced seasonal cycle with highest values between November and January and lowest flux rates in April and May (Fig. 5). The pronounced seasonal cycle is the consequence of a highly variable zonal transport across EB that attains its maximum in January and becomes negative in spring (April and May), corresponding to an ice import from

the New Siberian Sea. The seasonal cycle of the meridional transport across the NB is relatively constant throughout the season and does not show large monthly variations.

The normalized monthly anomalies in the total Laptev Sea ice flux rates across NB and EB (Fig. 6) were calculated by dividing the difference between monthly total ice flux and the 19-year monthly means by the 19-year monthly standard deviations. A trend analysis of the time series revealed an increase in ice transport rates during all months. However, trends are significant at 95 % confidence level only for January and March. Nevertheless, understanding the observed increase in total ice flux rates between 1992 and 2011, requires examining fluxes across the NB and EB separately. In Figure 7 the average monthly ice concentration, ice drift velocity, ice area flux and sea level pressure (SLP) gradients are shown with the corresponding trends across the NB and EB between 1992 and 2011. The ice drift velocity represents the boundary-perpendicular component ( $U$  and  $V$ ). The monthly mean SLP gradients across the boundaries provide a measure of the strength of the geostrophic wind component. The SLP gradients are the difference between the eastern and western end of the NB and northern and southern end of the EB. Gradients were calculated using monthly mean SLP data from the National Centers for Environmental Prediction (NCEP)/Department of Energy (DOE, Kalnay et al. (1996)). Southerly and westerly winds (positive gradients) results in ice export at the NB and EB, respectively, while negative gradients represent northerly and easterly winds and ice import across the NB and EB. Two representative examples of SLP distributions were chosen to highlight the location of atmospheric pressure systems and their impact during months with above and below average ice export (Fig. 8). A positive SLP gradient across the NB is favored by the location of a high pressure system over the East Siberian Sea, resulting in an anticyclonic circulation, i.e. increased offshore winds (Fig. 8 a). A displacement of the high pressure cell to the south introduces a westward component north of the New Siberian Islands and a positive SLP gradient across the EB. In contrast, the formation of a high pressure cell in the central Arctic suppresses ice export across boundaries (Fig. 8 b).

Figure 7 suggests that the overall positive trend in January and March total ice area flux (Fig. 6 and description in text) is the consequence of a significant increase in ice flux across the EB in January and NB in March, respectively. Interestingly during September and October, when the average ice concentration has undergone a dramatic decrease throughout the last decade, the ice export across the NB is increasing, rather than decreasing. The fact that negative trends in the ice concentration are not seen in the ice area flux rates is explained by the balance between an increase in ice drift velocities during September and October and the decrease in ice coverage. Changes in the ice drift velocity in all months are in turn reflected in changes in the SLP gradient. The agreement between variations in the SLP gradients and ice velocities is  $R = 0.92$  for the NB and slightly lower ( $R = 0.81$ ) for the EB.

The strong coupling between across-boundary SLP gradient and ice drift velocity apparent in each month indicates that monthly variations in ice area flux are primarily controlled by changes in the magnitude of the geostrophic winds. In particular along the NB, positive or negative trends in ice drift rates between November and May are the consequence of positive or negative trends in SLP gradients. However, along the EB, trends in SLP gradients and ice drift rates coincide less or are even of opposite sign (for example November and December). Variations in SLP gradients still explain year-to-year changes in monthly zonal ice transport rates, but do not account for observed trends in drift rates. Here, changes in the internal ice state and/or ocean forcing may be responsible for the observed trends in drift and ice area transport rates. This is discussed in more detail in section 4.

### 3.2 Laptev Sea ice pathways

To identify the sources and pathways of the sea ice, we back-tracked ice crossing the NB and EB at the end of April for a 6 month period between 1992 and 2011. The drift analysis is limited to the winter months (beginning November until the end of April), when temperatures are below freezing and ice production takes place in flaw polynyas along the extensive fast ice belt (Fig. 1).

Figure 9 a) and b) shows the ice drift pathways for a year with high (winter of 2008/09) and low (winter of 1998/99) ice flux rates across boundaries. For describing the variability in the pathways, we analyzed the length and the displacement of ice leaving the Laptev Sea in April through the center of the NB and EB. The trajectories for winter of 2008/09 and 1998/99 are shown in Figure 9 a) and b) as black lines. The lengths of the trajectories indicate the net displacement of an ice parcel throughout the 6 month period. The average length of the pathway passing the center of the NB between 1992 and 2011 is  $1046 \text{ km} \pm 188 \text{ km}$ . The net displacement of the ice passing through the center of the EB is on average 9 % shorter with a higher standard deviation ( $956 \text{ km} \pm 233 \text{ km}$ ). Maximum displacement of ice crossing the NB and EB were obtained in the winter of 2006/07 with 1550 km and 1407 km, respectively. Lowest rates occurred in the winter of 1998/99 with 610 km (EB) and 744 km (NB). On average, the ice leaving the Laptev Sea through the central NB originates from the western Laptev Sea at  $77.7^\circ \text{N}$  ( $\pm 1^\circ$ ),  $118^\circ \text{E}$  ( $\pm 8^\circ$ ). Quantities within parentheses are standard deviations. The origin of ice passing through the center of the EB is located further south-east at  $75.7^\circ \text{N}$  ( $\pm 1^\circ$ ),  $132^\circ \text{E}$  ( $\pm 11^\circ$ ). Hence, most of the ice that is incorporated into the Transpolar Drift originates from the central and western part of the Laptev Sea, while the exchange with the East Siberian Sea is primarily dominated by ice coming from the central and south-eastern Laptev Sea. Figure 9 c) presents the frequency distribution map of the origin of sea ice (positions of ice in November) calculated from pathways between 1992 and 2011. Because the origin of most of the

ice leaving the Laptev Sea by the end of April is situated in the central Laptev Sea in November and not in near-coastal zones, the contribution of polynyas to the winter ice area flux is rather small and limited to events that take place in the vicinity of the Laptev Sea boundaries (for example NS and NET polynyas; see Fig. 1). Hence, ice export during winter months is dominated by ice formed during freeze-up. Nevertheless, ice that is formed in polynyas occupies large portions of the Laptev Sea area at the end of the winter. If the polynya ice is then incorporated into the Transpolar Drift during summer months, or if it becomes subject of melting can not be answered. Too little information is available on the drift of Laptev Sea ice in summer.

### 4 Discussion of area flux estimates

Through a comparison with ENVISAT and ADCP ice drift information, we have shown that the accuracy of the IFREMER data in the Laptev Sea is high (see section 2). This is in agreement with findings of Rozman et al. (2011). Assuming that the error in the IFREMER ice drift estimates is normally distributed throughout the year, unbiased and uncorrelated, one can calculate the uncertainty in NB and EB ice flux based on the boundary length and the  $SE$  of the  $V$  and  $U$  component (Fig. 2), respectively. For the winter season (October to May) this results in an uncertainty in ice area flux of  $8.1 \times 10^4 \text{ km}^2$  and  $5.7 \times 10^4 \text{ km}^2$  for the NB and EB, respectively.

The ice exchange of the Laptev Sea with surrounding seas were first time investigated by Zakharov (1966a) and Zakharov (1967) using monthly gradients of atmospheric pressure for the period between 1937 and 1958. According to Alexandrov et al. (2000), the authors found the ice exchange between fall and spring to be within the range of our findings. However, the publication is available in Russian only which makes a direct comparison with our estimates difficult. Sea ice circulation in the Laptev Sea and ice exchange with the Arctic Ocean have been further studied by Alexandrov et al. (2000). The authors investigated ice exchange through the NB and EB based on a large-scale thermodynamic-dynamic sea ice model from 1979 to 1995. Following Alexandrov et al. (2000), the average winter (October to May) ice flux varies between  $2.51 \times 10^5 \text{ km}^2$  and  $7.32 \times 10^5 \text{ km}^2$  with a mean value of  $4.83 \times 10^5 \text{ km}^2$ . Due to the different study period, a quantitative comparison of our flux estimates with model estimates of Alexandrov et al. (2000) is not possible. However, their computations exceed our calculations by approximately 40 % ( $3.48 \times 10^5 \text{ km}^2$ ). A direct comparison of data from the three year overlap between the two time series (1992/93 - 1994/95) indicates an offset of 29 %. This number is consistent with the uncertainty in model computations of Alexandrov et al. (2000). Through a comparison of model results with satellite derived fluxes (SSM/I) for the winter of

1987/88 and 1994/95 the authors could show that their model calculations overestimate observations by as much as 24 %.

Overall, our analysis showed that the winter (October - May) total ice area flux out of the Laptev Sea has increased throughout the last two decades. The trend in total ice transport across boundaries is significant at the 95 % confidence level, while the trend in zonal flux across EB is significant only at 90 % *p*. The positive trend of the flux rates across NB is thereby smaller and statistically not significant. The comparison of SLP gradients with ice concentration and across-boundary ice drift showed that most of the interannual and seasonal variability in the Laptev Sea ice area flux can be linked to changes in geostrophic winds (compare Fig. 7). However, although variations in SLP gradients may explain large fractions of the year-to-year changes in monthly ice transport rates, they do not explain the observed long-term positive trends in area fluxes.

Figure 10 shows the daily anomalies of the pressure gradients for the past 19 years at NB and EB separately. Trends, which were calculated in a least square sense from these daily anomalies on the original data grids, are negative across both boundaries but not statistically significant. Along the EB the decrease is more pronounced and amounts to approximately -0.74 hPa/decade as compared to -0.68 hPa/decade at the NB. Note that the negative trends in the SLP gradients are also visible in the across-boundary component of the wind velocities (*U* and *V*). At the EB, the *U* component decreases by approximately -0.56 m/sec/decade, while the NB shows a reduction in *V* by -0.50 m/sec/decade (not shown here). Hence, we conclude that no evidence exists that the increase in ice area flux is favored by an increase in wind speed.

Our results are consistent with findings made by Spreen et al. (2011), who examined spatial trends in Arctic sea ice drift by means of SSM/I satellite data for the winter month of October through May between 1992 and 2009. They found a significant increase in ice drift velocity over large fractions of the Arctic Ocean and in the Laptev Sea. According to the authors, the increase in wind speed in the central Arctic explains a large fraction of the observed increase in drift velocities. However, in other regions, such as the Northern Barents, Kara, Laptev, and East Siberian Seas, wind speed trends are mainly negative and do not correlate with trends in ice motion.

Therefore, other mechanisms may be responsible for the observed changes in ice transport rates throughout the past two decades (in particular across the EB). Along the NB and EB, surface currents are believed to be primarily wind and ice driven. Following Spreen et al. (2011) we assume that the increasing ice drift speed and flux rates in the Laptev Sea may be rather associated with a change in the ice cover (thinning and/or decreasing concentration), caused by the rapid loss and thinning of thick multiyear ice (Haas et al., 2008; Kwok and Rothrock, 2009; Comiso, 2012). This is in agreement with Rampal et al. (2009) who found that changing thickness, deformation and compactness of the ice cover are ac-

companied by an increase in sea ice drift velocities measured by drift-buoys. Because of the absence of a positive trend in surface wind speed, they concluded that changes in the ice cover play a dominant role in explaining the observed trends in sea ice drift velocities.

#### 4.1 Linkage between ice area flux and large-scale atmospheric circulation

Previous studies have shown that changes in Laptev Sea ice drift patterns are linked to changes in the large-scale atmospheric circulation (e.g., Proshutinsky and Johnson, 1997; Alexandrov et al., 2000). Following e.g. Proshutinsky and Johnson (1997) or Deser et al. (2000) the Arctic atmospheric and oceanic circulation regime alternates between an intensified or suppressed cyclonic circulation in the Northern Hemisphere and the Arctic and a weakened or strengthened anticyclonic circulation (Beaufort Gyre). The decadal and multi-decadal variability in the atmospheric and oceanic circulation is summarized in the Arctic Oscillation (AO) index, which is defined as the leading principal component of the Northern Hemisphere SLP (Rigor et al., 2002). In this context, Proshutinsky and Johnson (1997) define the two arctic-wide observed wind-driven oceanic circulation patterns as the anticyclonic (low AO index) and cyclonic (high AO index) circulation regime, each of which persists from 5 to 7 years.

A negative coupling between AO index and winter ice export was observed by Proshutinsky and Johnson (1997) and Alexandrov et al. (2000) between the late 1970s and the early 1990s. According to Proshutinsky and Johnson (1997), during low AO index phases, ice leaving the Laptev Sea through the northern boundary is directly incorporated into the Transpolar Drift. This phase is characterized by an above average winter ice area flux through the northern boundary. In contrast, during high AO phases when the Transpolar Drift is shifted east towards the North American Arctic (Bareiss and Goergen, 2005), most of the sea ice is exported from the Laptev Sea to the East Siberian Sea across the eastern boundary. Alexandrov et al. (2000) who found that the magnitude and direction of Laptev Sea ice exchange with surrounding seas agrees well with the large-scale drift pattern during periods of prevailing anticyclonic or cyclonic circulation. The standardized detrended anomalies of the total Laptev Sea ice area flux (December - February) and the interannual variability of the winter AO index (December - February) are presented for the period from 1992 to 2010 in Figure 11. However, in contrast to Proshutinsky and Johnson (1997) and Alexandrov et al. (2000), we find the winter sea ice area flux anomalies in the Laptev Sea to be only weakly correlated with the AO-index ( $R = 0.24$ ).

Wu et al. (2006), Watanabe et al. (2006), and Wang et al. (2009) conclude that the influence of the winter Dipole Anomaly (DA) on the Arctic-wide sea ice motion is greater than that of the winter AO. The DA is defined by the second EOF (Empirical Orthogonal Function) mode of the SLP

north of 70°N. Following Wu et al. (2006), a positive phase of the DA favors a weakening of the Beaufort Gyre, increases sea ice export out of the Arctic basin through the Fram Strait and the northern Barents Sea, and enhances sea ice import from the Laptev Sea and the East Siberian Sea into the Arctic basin. During the negative phase of the DA, the opposite scenario occurs, i.e., the Beaufort gyre is strengthened, more sea ice remains in the western Arctic, and sea ice exports from the Laptev Sea and Kara Sea decreases. A comparison of the DA index phases with our ice area flux estimates reveals that there is indeed a correlation between ice export and the second EOF mode of SLP north of 70°N (Fig. 11). However, the correlation is negative ( $R = -0.62$ ), which stands in contrast to findings of Wu et al. (2006) and Watanabe et al. (2006).

The apparent weak response of ice area flux to AO and DA patterns indicates that no clear linkage exists between Laptev Sea ice circulation, and the large-scale atmospheric circulation. The magnitude of ice flux is rather determined by local atmospheric patterns, such as the occurrence of a high pressure system over the East Siberian Sea resulting in enhanced ice area flux. Consequently, earlier studies that linked other processes such as fast ice extent, polynya activity or freshwater distribution in the Laptev Sea to AO/DA patterns, may require reconsideration, in particular if the linked processes are partially dependent on the sea ice transport/circulation. Nevertheless, one must not forget that large-scale atmospheric patterns only explain a fraction of the SLP variability (Rigor et al., 2002) and that the linkage with ice conditions can be sensitive to the position of the SLP anomalies (Stroeve et al., 2011) that are not captured by EOF loading patterns.

#### 4.2 Linkage between ice area flux and summer ice conditions

The Laptev Sea ice extent is currently undergoing a significant decrease in summer (compare e.g. Fig. 9, sea ice concentration in September along the NB and EB). Following e.g. Bareiss and Goergen (2005) anomalies in summer sea ice extent are thought to be the consequence of I) synoptic-scale processes (e.g. cyclones) superimposed on the large-scale atmospheric circulation during summer, and II), to a lesser extent, from preconditioning processes in late winter and spring. The connection between shifts in the atmospheric circulation and in particular the role of cyclonicity for anomalies in summer sea ice concentration were previously discussed by Serreze et al. (1993); Serreze (1995); Maslanik et al. (1996) and Maslanik et al. (2000). In particular cyclones entering the Laptev Sea from the south-west enhance the northward ice transport and are associated with an inflow of anomalous warm air masses of above average air temperatures, which may contribute to sea ice melt. Persistent low sea level pressure and above average summer surface air temperatures summer may further promote early break-up, rapid melt and subsequent reductions in surface albedo as indicated

by Haas and Eicken (2001) and Serreze et al. (2003). A detailed discussion on potential linkages between summer ice coverage and atmospheric processes is given in Bareiss and Goergen (2005).

The preconditioning role of winter and spring sea ice processes on the summer and fall ice conditions has received little attention. For example, Deser et al. (2000) and Rigor et al. (2002) argued that low AO phases coincide with an increase in ice advection away from the fast ice edge, which results in an increase in production of new thin ice in coastal polynyas. The presence of extensive thin ice zones at the end of the winter may in turn favor ice retreat in summer and results in a delayed freeze-up in fall. Likewise, Alexandrov et al. (2000) found that during years where ice export out of the Laptev Sea was above average (e.g. winters of 1992/93 and 1994/95), a below average summer ice extent could be observed. This implies that negative sea ice anomalies in summer may be caused by enhanced northward ice advection during the previous winter.

The thinning effect of enhanced offshore ice advection on the sea ice cover has recently been observed in field. At the end of April 2012, a sequence of electromagnetic (EM) helicopter-borne ice thickness measurements (Haas and Eicken, 2001) were carried out over the WNS polynya north of the Lena Delta (Fig. 1). The winter of 2011/12 (not included in this study) was characterized by the highest northward advection rates observed since 1992. The continuous ice export away from the fast ice edge led to the development of an almost 200 km wide thin ice zone of less than 40 cm ice thickness. In contrast, EM measurements that were made at the end of April, 2008, show a substantially different ice regime north of the fast ice edge. A below average northward ice transport and longer periods of onshore-directed ice drift, formed heavily compacted ice of more than 1.5 m thick (Rabenstein et al., 2012). It stands to reason that the presence of extensive thin ice areas at the end of winter, along with continuous offshore transport during early spring and the intrusion of warm air masses from the south may accelerate ice retreat in summer. In the following, we will focus on the potential linkage that exists between presence/absence of extensive thin ice areas in late winter and the occurrence of low/high summer ice extent.

Thin ice areas can be derived directly from AMSR-E and SSM/I scenes, using the Polynya Signature Simulation Method (PSSM, Markus and Burns, 1995; Kern, 2009; Willmes et al., 2011). The method provides a classification of thin ice and open water (polynya area) and is based on the sensitivity of passive microwave polarization ratios to ice thickness in the range of 0 - 20 cm (Willmes et al., 2010). However, the coarse spatial resolution of passive microwave satellite data, in particular if applied to narrow polynyas or polynyas formed during extremely cold periods, generates errors through mixed signals at the fast and pack-ice edges, and results in underestimating the thin ice and open water area (Willmes et al., 2010; Krumpfen et al., 2011b).

Because offshore wind favors both, ice transport away from the coast and the development of thin ice in  
 705 flaw polynyas, a close relationship exists between across-  
 boundary ice export and estimated polynya area (Fig. 12).  
 The high agreement ( $R = 0.85$ ) allows us, to focus on ice ex-  
 port as in order to further understand the link between thin  
 ice areas and the subsequent summer ice anomaly. Shown in  
 710 Figure 12 are the normalized anomalies of ice export at the  
 NB and EB and polynya area (thin ice/open water) as derived  
 from the PSSM for the period between 1992 and 2009. The  
 PSSM was driven with the daily Advanced Microwave Scan-  
 ning Radiometer-Earth Observing System (AMSR-E)/Aqua  
 715 L2A Global Swath Spatially Resampled Brightness Temper-  
 atures data set (Ashcroft, 2008), available since 2002. Prior  
 to 2002 the polynya area is deducted from the open water  
 area in specific regions using a polynomial regression model  
 that describes the relation between polynya area (from PSSM  
 720 with AMSR-E) and SSM/I sea ice concentrations. The model  
 is derived from the data overlapping period (2002- 2008). A  
 detailed description of the methodology is given in Willmes  
 et al. (2011).

The comparison of ice area flux in late winter/early spring  
 725 with summer ice anomalies of the following summer reveals  
 a negative coupling that support our assumption (Fig. 13,  
 lower panel): There is evidence that an enhanced north-  
 ward/eastward transport of ice between February and May  
 can result in larger, less deformed, thin ice areas that may  
 730 more rapidly melt once temperatures rise above freezing.  
 The correlation coefficient between summer ice coverage and  
 winter ice area flux (February to May) is  $R = -0.65$ . The fig-  
 ure indicates that years of above the average ice export are  
 generally characterized by below the average summer ice ex-  
 735 tent and vice versa (e.g. 2004 and 2005). Including ice ad-  
 vection that took place before February results in a weaker  
 negative correlation coefficient. Taking into account early  
 summer ice export rates (e.g. June and July) is not possible,  
 owing to the limited temporal coverage of the IFREMER ice  
 740 drift dataset.

Synoptic scale processes such as cyclones occurring dur-  
 ing summer are believed to further accelerate ice retreat  
 and melting (see discussion above). For example, Maslanik  
 et al. (1996) report that an increase in cyclonicity in northern  
 745 Siberia since 1989 coincides with negative sea ice anomalies  
 in the Laptev Sea during the early 1990s. To confirm this  
 relationship, we derived a time series of cyclone events (re-  
 gion:  $100^\circ - 140^\circ$  E,  $72^\circ - 79^\circ$  N, Fig. 13, upper panel). The  
 750 tracking approach is based on the relative vorticity field in  
 the 850 hPa pressure level computed from 6-hourly NCEP  
 reanalysis data. For a detailed method description we refer  
 to Hoskins and Hodges (2002) and Hodges et al. (2011). The  
 cyclone activity in the Laptev Sea is characterized by a sea-  
 805 sonal cycle, with generally lower activity during winter than  
 during summer. The average number of cyclones entering  
 the Laptev Sea per month is approximately 3.9. The trend  
 755 in cyclone frequency is negative ( $-0.67$  cyclones/decade) but

not significant.

Following Rademacher (2009), cyclones entering the  
 Laptev Sea from the west, passing  $100^\circ$  and  $120^\circ$  E be-  
 tween  $74^\circ$  and  $79^\circ$  N, can be expected to be of greater im-  
 pact on offshore transport and polynya activity (e.g. T, AL  
 and WNS; Fig. 1). These cyclones cause westerly winds on  
 their southern side and south-westerly winds on their front  
 side. In addition, cyclones entering from the south between  
 $110^\circ - 135^\circ$  E and  $74^\circ - 78^\circ$  N enhance offshore transport  
 through southerly winds in the central Laptev Sea and south-  
 easterly winds over the New Siberian Islands. Nevertheless,  
 a comparison of northward and eastward directed summer  
 cyclone activity with the sea ice minimum in September re-  
 veals no evidence of a connection between synoptic scale  
 processes and the summer sea ice situation. Anomalies in  
 the number of easterly and northerly moving cyclones for  
 the period between July and September are given in Figure  
 13 (lower panel). This time period was chosen, because after  
 July the sea ice cover becomes vulnerable to cyclone activity  
 and wind stress (Bareiss and Goergen, 2005).

Our results imply that the summer sea ice extent is at least  
 partially controlled by the magnitude of ice area flux during  
 late winter/early spring. The observed positive trends in off-  
 shore directed ice transport (compare Fig. 3), likely being  
 the consequence of a change in the sea ice cover (e.g. thin-  
 ning/decrease in concentration), may act as a positive feed-  
 back, further amplifying an early ice retreat in summer. The  
 absence of a positive coupling between synoptic scale pro-  
 cesses and summer sea ice conditions is somewhat surpris-  
 ing and indicates that the contribution of cyclones has been  
 overestimated or is not fully understood. We conclude that  
 the physical connection between late winter/early spring ice  
 area flux, synoptic scale processes and ice extent in sum-  
 mer remains unclear. A more detailed analysis is needed that  
 separates thermodynamic effects from dynamic components,  
 which, however requires the application of a coupled sea ice  
 ocean model and is part of an upcoming study.

## 5 Conclusions

In this paper we investigate fluctuations in seasonal and in-  
 terannual sea ice area flux across Laptev Sea boundaries and  
 link observed trends to changes in atmospheric forcing mech-  
 anisms. The cumulative October to May total ice area flux  
 amounts to  $3.48 \times 10^5$  km<sup>2</sup>. For comparison, this is ap-  
 proximately 48 % of the winter, and 41 % of the annual  
 Fram Strait export. The average transport across the eastern  
 boundary ( $0.61 \times 10^5$  km<sup>2</sup>) is generally lower than across the  
 northern boundary ( $2.87 \times 10^5$  km<sup>2</sup>). Ice flux rates show a  
 pronounced seasonal cycle with the highest fluxes in Novem-  
 ber and January. By use of a back-propagation method, we  
 found that most of the ice that is incorporated into the Trans-  
 polar Drift originates from the central and western part of  
 the Laptev Sea, while the exchange with the East Siberian

810 Sea is primarily controlled by ice coming from the central  
and south-eastern Laptev Sea. The contribution of polynyas  
to the winter ice area flux is small and limited to events that  
take place in the vicinity of the boundaries. The total ice area  
flux out of the Laptev Sea is undergoing a statistically sig-  
815 nificant positive trend of  $0.85 \times 10^5 \text{ km}^2/\text{decade}$  between  
1992 and 2010. The trend in zonal ice area flux across EB  
is  $0.55 \times 10^5 \text{ km}^2/\text{decade}$ . The positive trend in the meridional  
transport across NB is  $0.3 \times 10^5 \text{ km}^2/\text{decade}$  but not  
820 statistically significant.

820 A large fraction of the year-to-year changes in monthly  
ice transport can be explained by variations in SLP gradi-  
825 ents across boundaries. However, there is no evidence of  
increasing wind velocities that could explain the overall posi-  
tive trends in ice export. Following Spreen et al. (2011), we  
therefore assume that changes in ice flux rates may be related  
to a change in ice cover such as thinning and/or a decrease in  
concentration. A comparison of ice export rates with AO and  
DA patterns indicates that no clear linkage exists between sea  
ice circulation in the Laptev Sea and large-scale atmospheric  
830 circulation. However, there is evidence that years of high ice  
export in late winter have a thinning effect on the ice cover,  
which in turn preconditions negative sea ice extent anomalies  
in summer. The observed positive trend in offshore directed  
ice flux may act as a positive feedback, further accelerating  
835 early ice retreat in the Laptev Sea. The missing relation-  
ship between synoptic and mesoscale processes and summer  
sea ice conditions is somewhat surprising and indicates that  
the contribution of cyclones has been overestimated or is not  
fully understood.

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## 860 References

Alexandrov, V., Martin, T., Kolatschek, J., Eicken, H., and  
M. Kreyscher, Makshtas, A. P.: Sea ice circulation in the Laptev

Sea and ice export to the Arctic Ocean: Results from satellite  
remote sensing and numerical modeling, *Journal of Geophys-  
ical Research*, 105, 17 143–17 159, doi:10.1029/2000JC900029,  
2000.

Ashcroft, P. and Wentz, F.: AMSR-E/Aqua L2A global swath  
spatially-resampled brightness temperatures V001, 2002-2008,  
Tech. rep., Digital media. Boulder, CO: National Snow and Ice  
Data Center., 2008.

Bareiss, J. and Goergen, K.: Spatial and temporal variability of sea  
ice in the Laptev Sea: Analysis and review of satellite passiv-  
microwave data and model results, 1997 to 2002, *Global and  
Planetary Change*, 48, 28–54, doi:10.1016/j.gloplacha.2004.12.  
004, 2005.

Comiso, J. C.: Variability and Trends of the Arctic Sea Ice Cover:  
Sea Ice, vol. 2, Wiley-Blackwell, New York, US, 2010.

Comiso, J. C.: Large Decadal Decline of the Arctic Multiyear  
Ice Cover, *Journal of Climate*, 25, 11761193, doi:10.1175/  
JCLI-D-11-00113.1, 2012.

Deser, C., Walsh, J., and Timlin, M.: Arctic sea ice variability in  
the context of recent atmospheric circulation trends, *Journal of  
Climate*, 13, 617–633, doi:10.1175/1520-0442(2000)013<0617:  
ASIVIT>2.0.CO;2, 2000.

Dethleff, D., Loewe, P., and Kline, E.: The Laptev Sea flaw  
lead - Detailed investigation on ice formation and export during  
1991/1992 winter season, *Cold Region Science and Technology*,  
27, 225–243, 1998.

Eicken, H., Reimnitz, E., Alexandrov, V., Martin, T., Kassens, H.,  
and Viehoff, T.: Sea-ice processes in the Laptev Sea and their  
importance for sediment export, *Continental Shelf Research*,  
17, 205–233, doi:10.1016/S0278-4343(96)00024-6, <http://www.sciencedirect.com/science/article/B6VBJ-3SVHM6V-R/2/4ad7018e301f16fbdf4e8202cce319e>, 1997.

Eicken, H., Dmitrenko, I. A., Tyshko, K., Darovskikh, A., Dierk-  
ing, W., Blahak, U., Groves, J., and Kassens, H.: Zonation  
of the Laptev Sea landfast ice cover and its importance in a  
frozen estuary, *Global and Planetary Change*, 48, 55–83, doi:  
10.1016/j.gloplacha.2004.12.005, 2005.

Emery, W. J., W., F. C., J., H., and Preller, R. H.: Fram strait satellite  
image-derived ice motions, *Journal of Geophysical Research*, 96,  
4751–4768, doi:10.1029/90JC02273, 1991.

Ezraty, R., Girard-Arduin, F., Pioll, J. F., Kaleschke, L., and  
Heygster, G.: Arctic and Antarctic Sea Ice Concentration  
and Arctic Sea Ice Drift Estimated from Special Sensor Mi-  
crowave Data, Dpartement dOcanographie Physique et Spa-  
tiale, IFREMER, Brest, France and University of Bremen,  
Germany, 2.1 edn., [ftp://ftp.ifremer.fr/ifremer/cersat/products/  
gridded/psi-drift/documentation/ssmi.pdf](ftp://ftp.ifremer.fr/ifremer/cersat/products/gridded/psi-drift/documentation/ssmi.pdf), 2007.

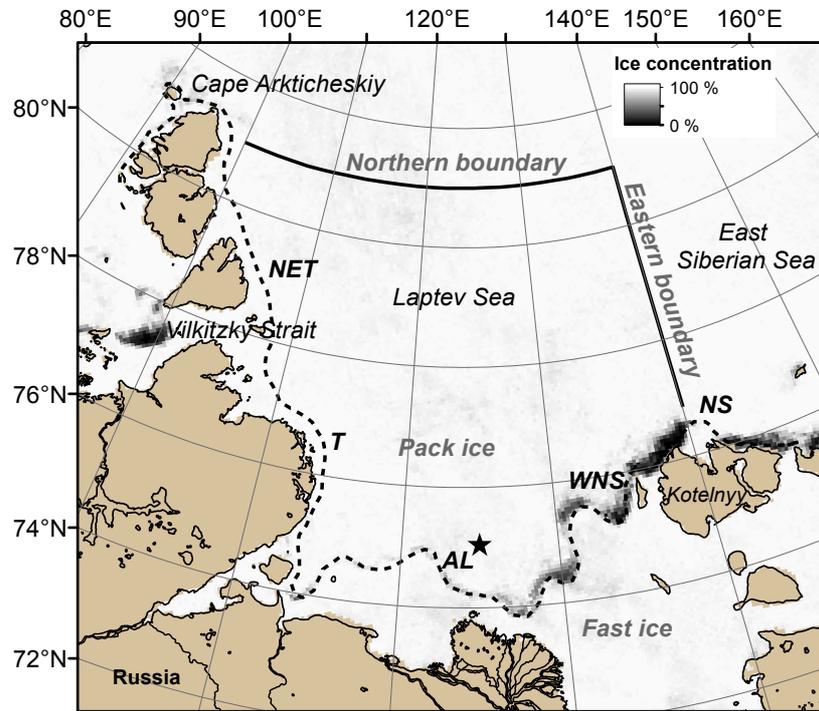
Fowler, C.: Polar Pathfinder Daily 25 km EASE-Grid Sea Ice Mo-  
tion Vectors, Tech. rep., National Snow and Ice Data Center.,  
Boulder, Colorado, USA, 2003.

Girard-Arduin, F. and Ezraty, R.: Enhanced Arctic sea ice drift  
estimation merging radiometer and scatterometer data, *IEEE  
Transaction in Geoscience and Remote Sensing*, Recent ad-  
vances in C-band scatterometry Special Issue, 50, 2639–2648,  
doi:10.1109/TGRS.2012.2184124, 2012.

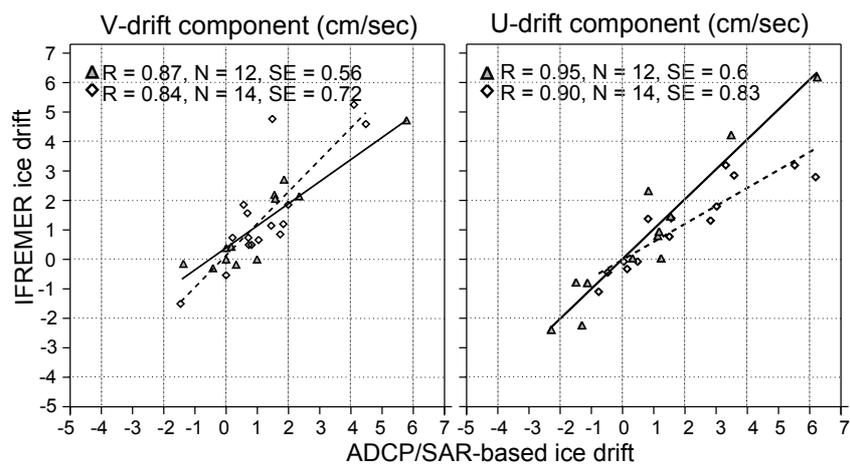
Haas, C. and Eicken, H.: Interannual variability of summer sea ice  
thickness in the Siberian and central Arctic under differnt atmo-  
spheric circulation regimes, *Journal of Geophysical Research*,  
106, 4449–4462, doi:10.1029/1999JC000088, 2001.

- Haas, C., Pfaffling, A., Hendricks, S., Rabenstein, L., L., E. J., and Rigor, I.: Reduced ice thickness in Arctic Transpolar Drift favors rapid ice retreat, *Geophysical Research Letters*, 35, doi:10.1029/2008GL034457, 2008.
- Hodges, K. I., Lee, R. W., and Bengtsson, L.: A Comparison of Extratropical Cyclones in Recent Reanalyses ERA-Interim, NASA MERRA, NCEP CFSR, and JRA-25, *Journal of Climate*, 24, 4888–4906, doi:10.3402/tellusa.v64i0.17196, 2011.
- Hoelemann, J. A., Kirillov, S., Klagge, T., Novikhin, A., Kassens, H., and Timokhov, L.: Near-bottom water warming in the Laptev Sea in response to atmospheric and sea ice conditions in 2007, *Polar Research*, 30, doi:10.3402/polar.v30i0.6425, 2011.
- Hoskins, B. J. and Hodges, K. I.: New perspectives on the Northern Hemisphere winter storm tracks, *Journal of Atmospheric Science*, 59, 1041–1061, doi:10.1175/1520-0469(2002)059<1041:NPTNH>2.0.CO;2, 2002.
- Janout, M. A., Hölemann, J. A., and Krumpfen, T.: Cross-shelf transport of warm and saline water in response to sea ice drift on the Laptev Sea shelf, *Journal of Geophysical Research*, p. in press, doi:10.1029/2011JC007731, inpress, 2012.
- Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J., Zhu, Z., Leetmaa, A., Reynolds, R., Chelliah, M., Ebisuzaki, W., Higgins, W., Janowiak, J., Ropelewski, C., Wang, J., Jenne, R., and Joseph, D.: The NCEP/NCAR 40-year reanalysis project, *Bulletin of the American Meteorological Society*, 44, 437–470, doi:10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2, 1996.
- Kern, S.: Wintertime Antarctic coastal polynya area: 1992–2008, *Geophysical Research Letters*, 36, 1–5, doi:10.1029/2009GL038062, 2009.
- Krumpfen, T., Hoelemann, J. A., Willmes, S., Maqueda, M. A. M., Busche, T., Dmitrenko, I. A., Gerdes, R., Haas, C., Hendricks, S., Kirillov, S., Rabenstein, L., and Schroeder, D.: Sea ice production and water mass modification in the eastern Laptev Sea, *Journal of Geophysical Research*, 116, doi:10.1029/2010JC006545, 2011a.
- Krumpfen, T., Willmes, S., Morales Maqueda, M. A., Gerdes, R., Haas, C., Hoelemann, J., and Schroeder, D.: Evaluation of a polynya flux model by means of thermal infrared satellite observations, *Annals of Glaciology*, 52, 52–60, doi:10.3189/172756411795931615, 2011b.
- Kwok, R.: Outflow of Arctic Ocean Sea Ice into the Greenland and Barents Seas: 1979–2007, *Journal of Climate*, 22, 2438–2457, doi:10.1175/2008JCLI2819.1, 2009.
- Kwok, R. and Pang, S. S.: Fram Strait sea ice outflow, *Journal of Geophysical Research*, 109, 1–14, doi:10.1029/2003JC001785., 2004.
- Kwok, R. and Rothrock, D. A.: Variability of Fram Strait ice flux and North Atlantic Oscillation, *Journal of Geophysical Research*, 104, 5177–5189, doi:10.1029/1998JC900103, 1999.
- Kwok, R. and Rothrock, D. A.: Decline in Arctic sea ice thickness from submarine and ICESat records: 1958–2008, *Geophysical Research Letters*, 36, 1–5, doi:10.1029/2009GL039035, 2009.
- Kwok, R., Cunningham, G. F., Wensnahan, M., Rigor, I., Zwally, H. J., and Yi, D.: Thinning and volume loss of the Arctic Ocean sea ice cover: 2003–2008., *Journal of Geophysical Research*, 114, doi:10.1029/2009JC005312, 2009.
- Markus, T. and Burns, B. A.: A method to estimate subpixel-scale coastal polynyas with satellite passive microwave data, *Journal of Geophysical Research*, 100, 4473–4488, doi:10.1029/94JC02278, 1995.
- Maslanik, J., Serreze, M., and Barry, R.: Recent decreases in Arctic summer ice cover and linkages to atmospheric circulation anomalies, *Geophysical Research Letters*, 23, 1677–1680, doi:10.1029/96GL01426, 1996.
- Maslanik, J., Lynch, A., Serreze, M., and Wu, W.: A case study of regional climate anomalies in the Arctic: performance requirements for a coupled model, *Journal of Climate*, 13, 3834–3841, doi:10.1175/1520-0442(2000)013<0383:ACSORC>2.0.CO;2, 2000.
- Massonnet, F., Fichefet, T., Goosse, H., Bitz, C. M., Philippon-Berthier, G., Holland, M. M., and Barriat, P. Y.: Constraining projections of summer Arctic sea ice, *The Cryosphere*, 6, 1383–1394, doi:10.5194/tc-6-1383-2012, 2012.
- Meier, W. N., Stroeve, J., and Fetterer, F.: Whither Arctic sea ice? A clear signal of decline regionally, seasonally and extending beyond the satellite record, *Annals of Glaciology*, 46, 428–434, doi:10.3189/172756407782871170, 2007.
- Proshutinsky, A. Y. and Johnson, M. A.: Two circulation regimes of the wind-driven Arctic Ocean, *Journal of Geophysical Research*, 102, 12 493–12 514, doi:10.1029/97JC00738, 1997.
- Rabenstein, L., Krumpfen, T., Hendricks, S., Haas, C., Hoelemann, J., and Koeberle, C.: Thickness distribution and volume production of up to 4 months old polynya sea ice in the Laptev Sea, Manuscript prepared for *The Cryosphere*, 2012.
- Rademacher, C.: Tracking of synoptic and mesoscale systems in the Siberian Arctic and their impact on the Laptev Sea polynya, Master's thesis, Meteorologisches Institut, Bonn University, Bonn, Germany, 2009.
- Rampal, P., Weiss, J., and Marsan, D.: Positive trend in the mean speed and deformation rate of Arctic sea ice, *Journal of Geophysical Research*, 114, doi:10.1029/2008JC005066., 2009.
- Rigor, I. G. and Colony, R. L.: Sea-ice production and transport of pollutants in the Laptev Sea, 1997–1993, *Science of the Total Environment*, 202, 89–110, doi:10.1016/S0048-9697(97)00107-1., 1997.
- Rigor, I. G., Wallace, J., and Colony, R.: Response of Sea Ice to the Arctic Oscillation, *Journal of Climate*, 15, 2648–2663, doi:10.1175/1520-0442(2002)015<2648:ROSITT>2.0.CO;2, 2002.
- Romanov, I. P.: Atlas of Ice and Snow of the Arctic Basin and the Siberian Shelf Seas, Backbone Publishing Company, Fair Lawn, US, 1996.
- Rozman, P., Hoelemann, J., Krumpfen, T., Gerdes, R., Koeberle, C., Lavergne, T., and Adams, S.: Validating satellite derived and modelled sea-ice drift in the Laptev Sea with in situ measurements from the winter of 2007/08, *Polar Research*, 30, doi:10.3402/polar.v30i0.7218, 2011.
- Schwegmann, S., Haas, C., Fowler, C., and Gerdes, R.: A comparison of satellite-derived sea ice motion with drifting buoy data in the Weddell Sea, *Annals of Glaciology*, 52, 103–110, doi:10.3189/172756411795931813, 2011.
- Serreze, M.: Climatological aspects of cyclone development and decay in the Arctic, *Atmosphere-Ocean*, 33, 1–23, doi:10.1080/07055900.1995.9649522, 1995.
- Serreze, M., Box, J., Barry, R., and Walsh, J.: Characteristics of Arctic synoptic activity, 1952–1989, *Meteorology and Atmospheric Physics*, 51, 147–164, doi:10.1007/BF01030491, 1993.
- Serreze, M., Maslanik, J., Scambos, T., Fetterer, F., Stroeve, J., Knowles, K., Fowler, C., Drobot, S., Barry, R., and T.M., H.:

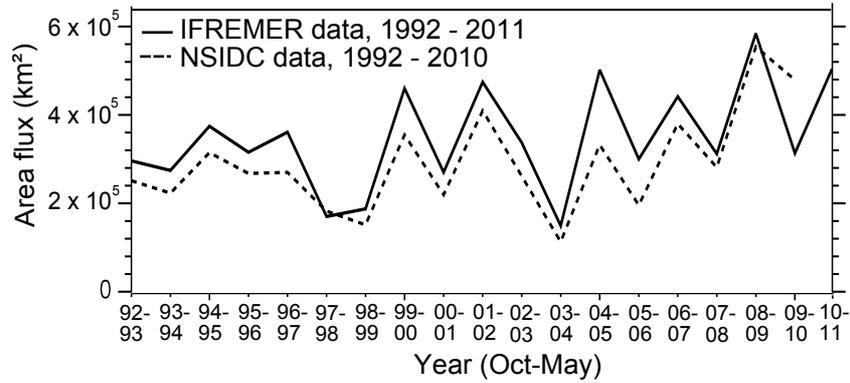
- 1040 A record minimum arctic sea ice extent and area in 2002.,  
Geophysical Research Letters, 30, doi:10.1029/2002GL016406,<sup>1100</sup>  
2003, 2003.
- Smedsrud, L. H., Sirevaag, A., Kloster, K., Sorteberg, A., and Sand-  
ven, S.: Recent wind driven high sea ice area export in the Fram  
1045 Strait contributes to Arctic sea ice decline, *The Cryosphere*, 5,  
821829, doi:10.5194/tc-5-821-2011, 2011.
- 1105 Spreen, G., Kaleschke, L., and Heygster, G.: Sea Ice Remote Sens-  
ing Using AMSR-E 89 GHz Channels, *Journal of Geophysical  
Research*, 113, 1–14, doi:10.1029/2005JC003384, 2008a.
- 1050 Spreen, G., Kaleschke, L., and Heygster, G.: Sea ice remote sens-  
ing using AMSR-E 89 GHz channels, *Journal of Geophysical  
Research*, 113, doi:10.1029/2005JC003384., 2008b.
- Spreen, G., Kwok, R., and Menemenlis, D.: Trends in Arctic sea ice  
drift and tole of wind forcing: 1992–2009, *Geophysical Research  
1055 Letters*, 38, 1–14, doi:10.1029/2011GL048970, 2011.
- Stroeve, J. C., Serreze, M. C., Fetterer, F., Arbetter, T., Meier, W.,  
Maslanik, J., and Knowles, K.: Tracking the Arctic’s shrinking  
ice cover: Another extreme September minimum in 2004, *Geo-  
physical Research Letters*, 32, 1–4, doi:10.1029/2004GL021810,  
1060 2005.
- Stroeve, J. C., Maslanik, J., Serreze, M. C., Rigor, I., Meier, W.,  
and Fowler, C.: Sea ice response to an extreme negative phase  
of the Arctic Oscillation during winter 2009/2010, *Geophysical  
Research Letters*, 38, doi:10.1029/2010GL045662, 2011.
- 1065 Stroeve, J. C., Kattsov, V., Barrett, A., Serreze, M., Pavlova, T.,  
Holland, M., and Meier, W. N.: Trends in Arctic sea ice extent  
from CMIP5, CMIP3 and observations., *Geophysical Research  
Letters*, 39, doi:10.1029/2012GL052676., 2012.
- Timokhov, L. A.: Regional characteristics of the Laptev and the  
1070 East Siberian seas: climate, topography, ice phases, thermohaline  
regime, circulation, *Berichte zur Polarforschung*, 114, 15–32,  
<http://epic.awi.de/26322/1/BerPolarforsch1994144.pdf>, 1994.
- Wang, J., Zhang, J., Watanabe, E., Ikeda, M., Mizobata, K., Walsh,  
J. E., Bai, X., and W., B.: Is the Dipole Anomaly a major driver  
1075 to record lows in Arctic summer sea ice extent?, *Geophysical  
Research Letters*, 36, doi:10.1029/2008GL036706, 2009.
- Watanabe, E., Wang, J., Sumi, A., and Hasumi, H.: Arctic dipole  
anomaly and its contribution to sea ice export from the Arctic  
Ocean in the 20th century, *Geophysical Research Letters*, 33, doi:  
10.1029/2006GL028112, 2006.
- 1080 Willmes, S., Krumpfen, T., Adams, S., Rabenstein, L., Haas, C.,  
Hoelemann, J., Hendricks, S., and Heinemann, G.: Cross-  
Validation of polynya monitoring methods from multi-sensor  
satellite and airborne data: A case study from the Laptev Sea,  
1085 *Canadian Journal of Remote Sensing*, 36, 196–210, doi:10.5589/  
m10-012, 2010.
- Willmes, S., Adams, S., Schroeder, D., and Heinemann, G.: Spa-  
tiotemporal variability of polynya dynamics and ice production  
in the Laptev Sea between the winters of 1979/80 and 2007/08,  
1090 *Polar Research*, Thematic cluster, climate drivers of the North:  
The Laptev Sea system (in press), doi:10.3402/polar.v30i0.5971,  
2011.
- Wu, B., Wang, J., and Walsh, J. E.: Dipole Anomaly in the winter  
Arctic atmosphere and its association with Arctic sea ice motion,  
1095 *Journal of Climate*, 18, 210–225, doi:10.1175/JCLI3619.1, 2006.
- Zakharov, V. F.: The role of flaw leads off the edge of fast ice in  
the hydrological and ice regime of the Laptev Sea, *Oceanology*,  
6, 815–821, 1966a.
- Zakharov, V. F.: Some features of the estimated ice drift in the  
Laptev Sea (in Russian), *Problems of the Arctic and Antarctic*,  
23, 45–46, 1966b.
- Zakharov, V. F.: Change of the Laptev Sea ice cover in relation with  
the oscillations of the baric field in the Arctic (in Russian), *Tr.  
Arkt. Antarkt. Inst.*, 257, 52–72, 1967.
- Zhang, X. and Walsh, J. E.: Towards a seasonally ice-covered Arc-  
tic Ocea: Scenarios from the IPCC AR4 simulations, *Journal of  
Climate*, 19, 1730–1747, doi:10.1175/JCLI3767.1, 2006.



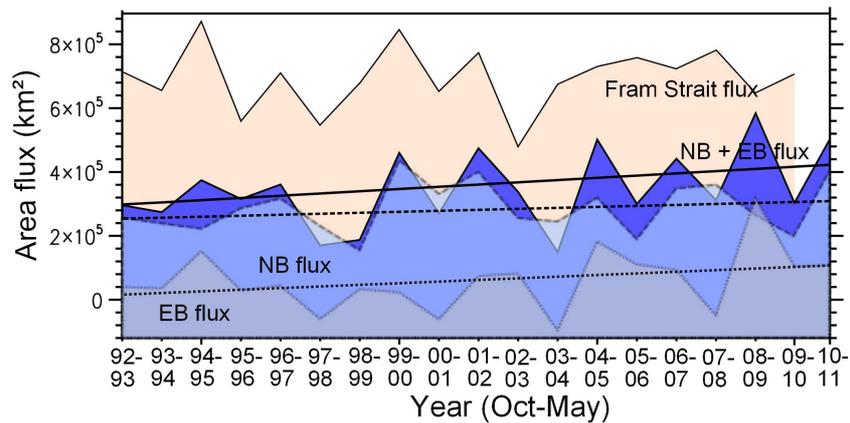
**Fig. 1.** The geographical location of the Laptev Sea and the northern and eastern boundaries (solid black lines) on which meridional and zonal ice area flux estimates are based. The dashed line represents the mean fast ice edge location. Between pack ice and fast ice edge, flaw polynyas are formed: The New Siberian polynya (NS), the Western New Siberian polynya (WNS), the Anabar-Lena polynya (AL), the Taymyr polynya and the North-Eastern Taymyr (NET) polynya. Color coding corresponds to the sea ice concentration as obtained from Advanced Microwave Scanning Radiometer (AMSR-E) on May 7, 2008 (source: IUP Bremen, Spreen et al. (2008a)). The position of the moorings used for satellite ice motion data validation are indicated by black stars.



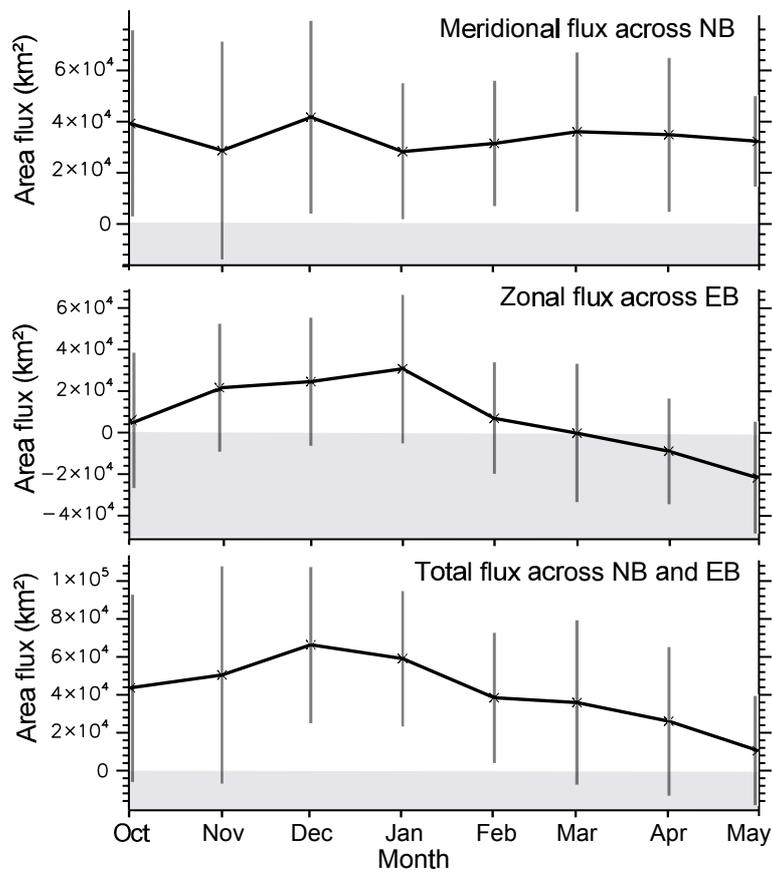
**Fig. 2.** Comparison between monthly mean IFREMER ice motion estimates and ENVISAT SAR (triangles) and ADCP based (diamonds) monthly mean ice drift. The  $U$  and  $V$  drift components are shown in the left and right panel, respectively. The solid and dashed line show the linear fit between IFREMER data and SAR and ADCP estimates. SAR motion information is based on a selection of scenes taken between 2006 and 2010 (December to April), while ADCP ice drift data originate from 3 year-round moorings (2007 - 2009) deployed in the vicinity of the fast ice edge. In addition, the correlation coefficient ( $R$ ), number of samples ( $N$ ) and standard deviation ( $SE$ ) is given.



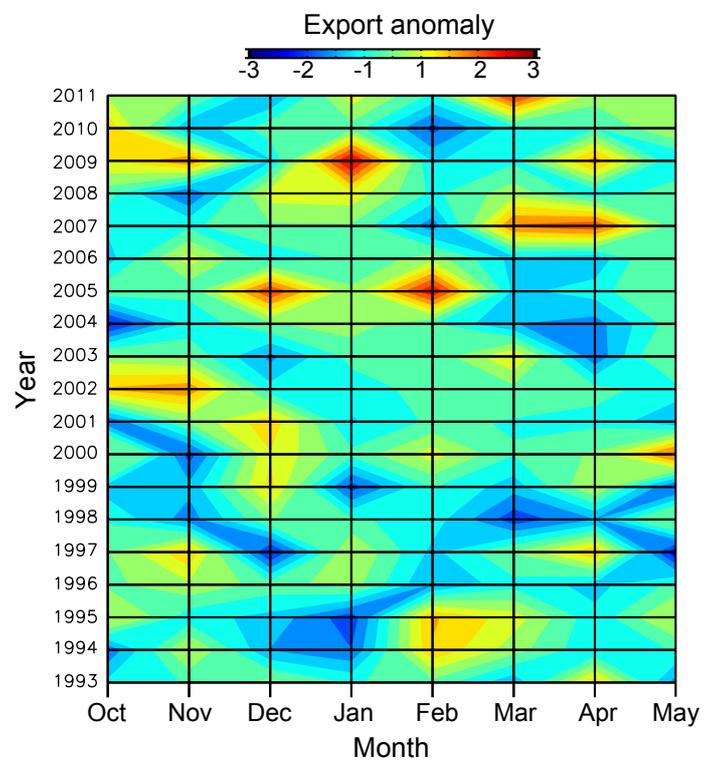
**Fig. 3.** Comparison between cumulative winter (October to May) IFREMER (solid line) and NSIDC (dashed line) total ice area flux estimates (given in 10<sup>5</sup> km<sup>2</sup>) through the northern and eastern boundaries.



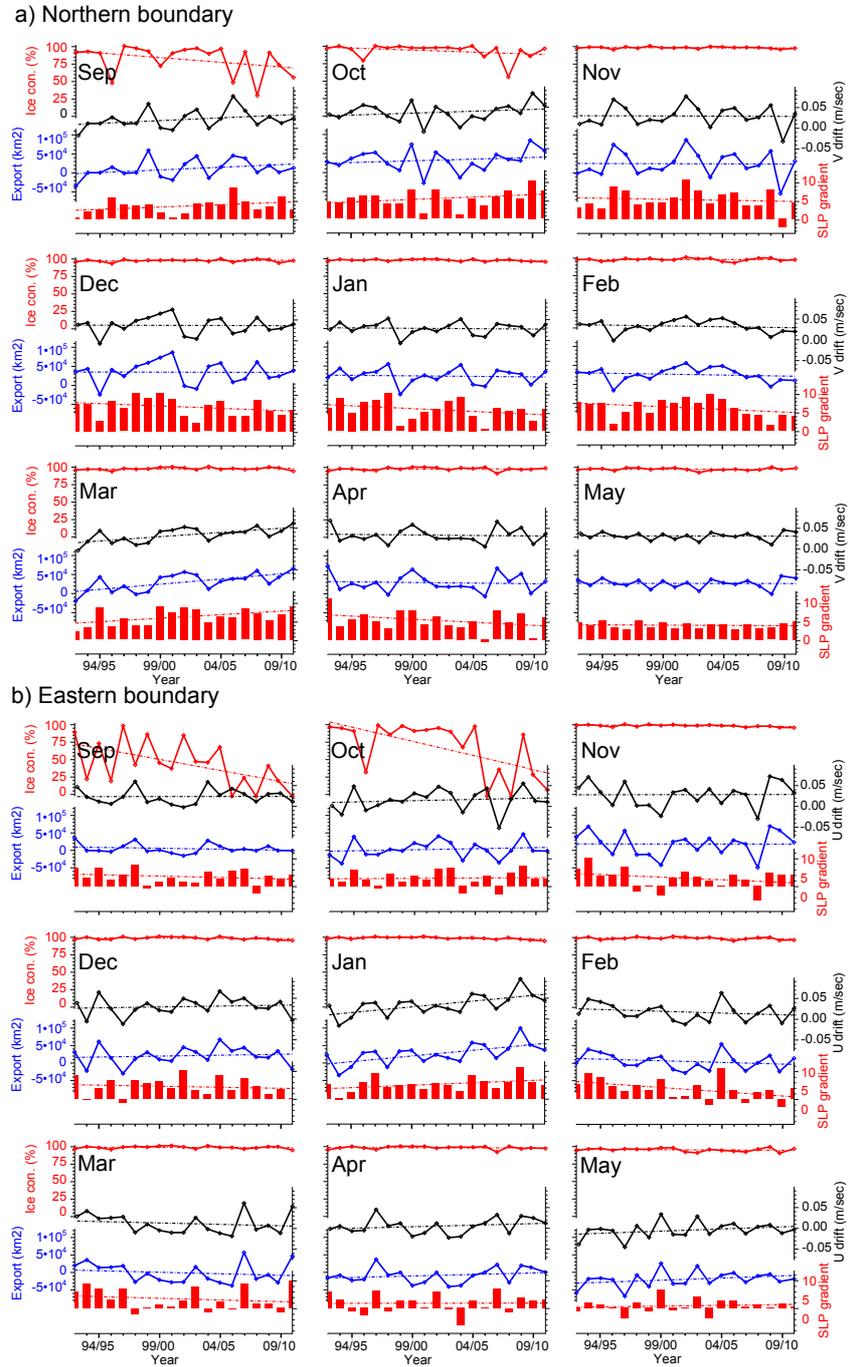
**Fig. 4.** Cumulative winter (October to May) ice area flux across the northern boundary (NB flux: dashed line, light blue filling), eastern boundary (EB flux: dotted line, grey filling) and northern and eastern boundaries (NB + EB flux: solid line, blue filling) between 1992 and 2011. Trend lines for total ice area flux (solid trend line,  $0.85 \times 10^5$  km<sup>2</sup>/decade), zonal flux across EB (dotted trend line,  $0.55 \times 10^5$  km<sup>2</sup>/decade) and meridional flux across NB (dashed trend line,  $0.3 \times 10^5$  km<sup>2</sup>/decade) are plotted on top. For quantitative comparison, the Fram Strait flux (solid line, beige filling) as calculated by Smedsrud et al. (2011) is shown.



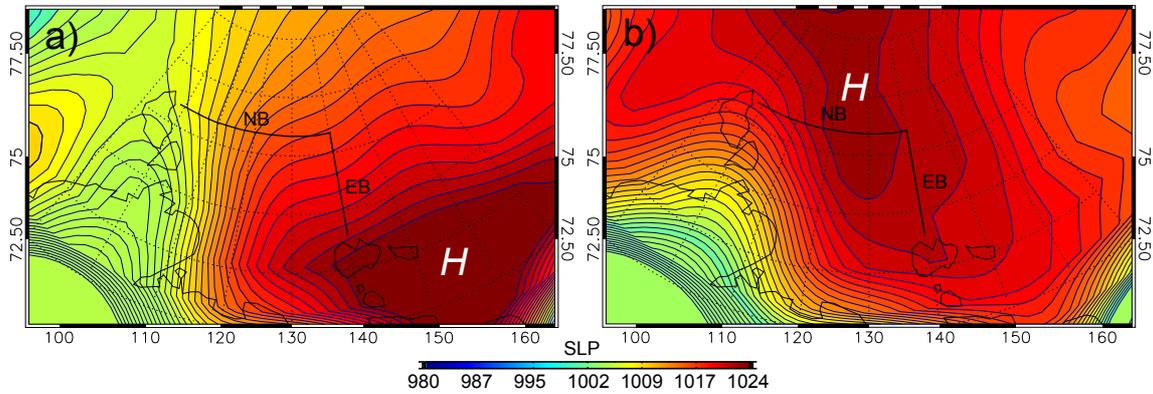
**Fig. 5.** Seasonal cycle of average monthly meridional ice area flux across northern boundary (upper panel), zonal flux across eastern boundary (center panel) and total flux (lower panel, given in  $10^4 \text{ km}^2$ ) for the period between 1992 and 2011. Bars denote  $SE$ . Grey shaded areas correspond to ice inflow (negative flux).



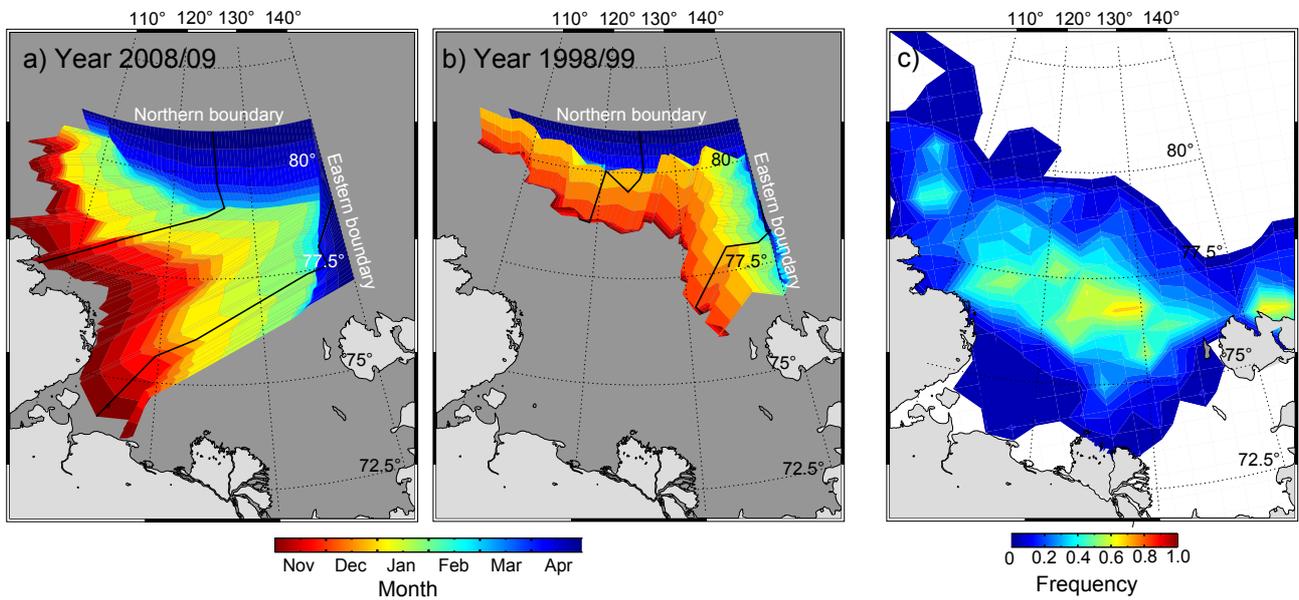
**Fig. 6.** Normalized monthly total ice area flux anomalies between October and May (1992 - 2011).



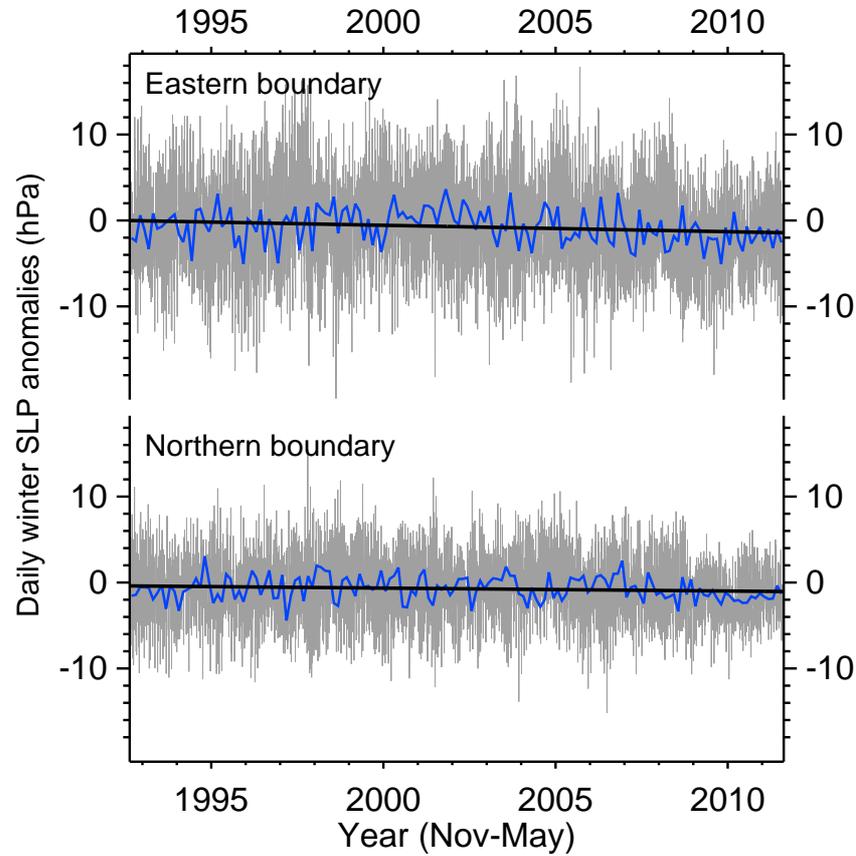
**Fig. 7.** Time series and trends of average monthly ice concentration, ice drift, ice area flux and sea level pressure gradient across northern boundary (a) and eastern boundary (b) between 1992 and 2011. The average monthly ice concentration is represented by the red line (upper left axis, given in %). The cross-boundary monthly ice drift ( $V$  and  $U$  component for a) and b) given in cm/sec, black line) and the monthly ice area flux (given in  $10^4$  km<sup>2</sup>, blue line) refer to the upper right and lower left axis, respectively. The cross-boundary sea level pressure gradient (given in hPa, lower right axis) is indicated by the red bars.



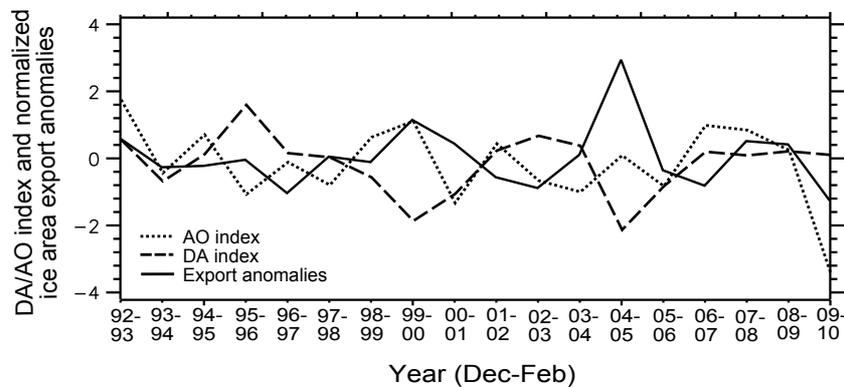
**Fig. 8.** Sea level pressure fields (mb) for a month of comparatively high (a: March, 2007) and low ice area flux rates (b: April, 1998) and across the northern and eastern boundary.



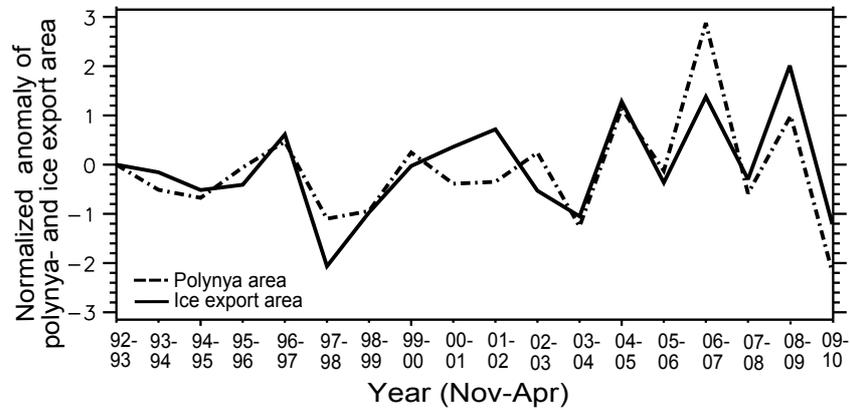
**Fig. 9.** Origin and drift of sea ice leaving the NB and EB at the end of April as obtained by the back-propagation of ice parcels over a 6-month period (see text for a detailed description). a) and b) show ice drift patterns for a winter with high (2008/09) and low (1998/99) ice transportation rates across boundaries. The color represents the position of the sea ice at a specific month before exiting the northern or eastern boundary. The trajectory of sea ice leaving the Laptev Sea through the center of the NB and EB are represented by black lines. c) shows the frequency of origins (position in November) of sea ice between 1992 and 2011.



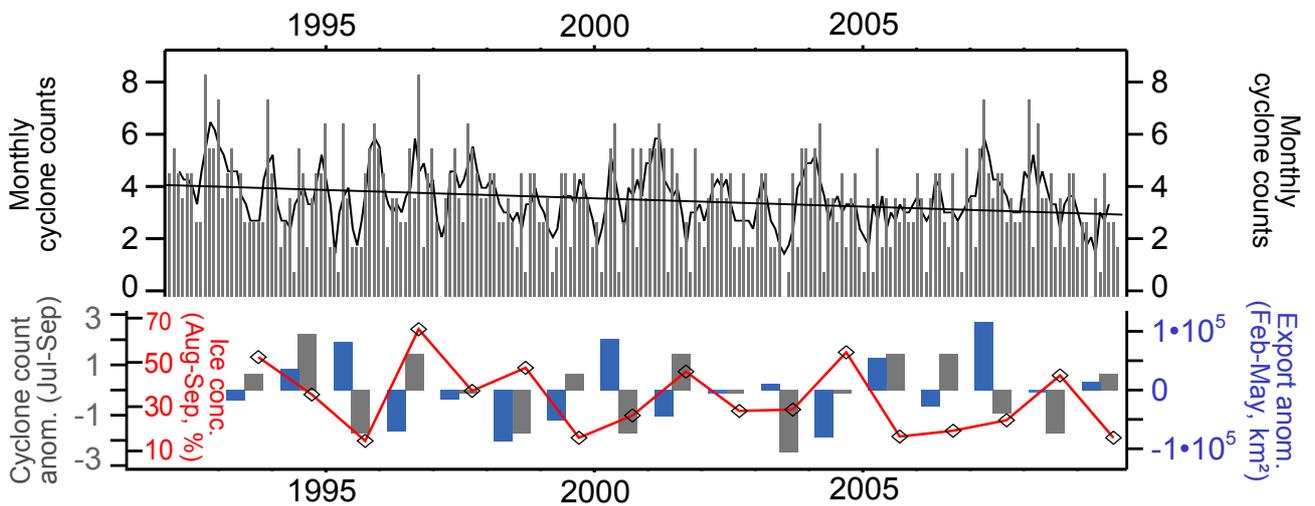
**Fig. 10.** Time series of Sea Level Pressure (SLP) gradient anomalies (October to May) across the eastern and northern boundaries. The blue line represents the monthly data. The trend line is calculated in a least square sense from the daily anomalies.



**Fig. 11.** Time series of winter (December - February) Dipole Anomaly (DA) and Arctic Oscillation (AO) index vs. normalized monthly anomalies of detrended total ice area flux out of the Laptev Sea from 1992 to 2010.



**Fig. 12.** Normalized anomalies of accumulated polynya area (thin ice and open water, dashed dotted line) and total ice area flux (solid line) between November and April from 1992 until 2009.



**Fig. 13.** The upper panel shows the frequency of cyclone counts per month in the Laptev Sea for the period between 1992 and 2009. The black lines represent the 3 month running mean and the trend. The lower panel presents anomalies in cyclone counts entering the Laptev Sea from the south and west between July and September (grey bar, left axis). The red line shows the average sea ice concentration for August - September (second left axis). Blue bars refer to anomalies in ice area flux out of the Laptev Sea across the NB and EB between February and May (given in km<sup>2</sup>, right axis).