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Camille Lique, Michael Steele. Seasonal to decadal variability of Arctic Ocean heat content: A model-based analysis and implications for autonomous observing systems. *Journal of Geophysical Research. Oceans*, 2013, 118 (4), pp.1673-1695. <10.1002/jgrc.20127>. <hal-05564822>

HAL Id: hal-05564822

<https://hal.science/hal-05564822v1>

Submitted on 24 Mar 2026

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Seasonal to decadal variability of Arctic Ocean heat content: A model-based analysis and implications for autonomous observing systems

Camille Lique^{1,2} and Michael Steele³

Received 2 July 2012; revised 7 February 2013; accepted 12 February 2013; published 3 April 2013.

[1] A high-resolution global ocean/sea ice model is used to investigate the modes of Arctic Ocean heat content variability for the period 1968–2007. A rotated empirical orthogonal function analysis is performed on the monthly mean vertically integrated heat content to investigate the mechanisms governing its spatiotemporal variations. In the model, 28% of the heat content variability is driven by the seasonal and interannual fluctuations of the atmospheric heat flux in the seasonally ice free regions. The heat flux variability associated with Atlantic Water advected through Fram Strait drives 31% of the heat content variability. Changes of temperature and circulation drive Fram Strait heat transport variability, and these two effects project on different modes and thus drive heat content variations in different parts of the Eurasian Basin. A second branch of Atlantic Water is modified in the Barents Sea and the variations of the heat flux associated with the Barents Sea water branch penetrating the deep Arctic yield heat content variations in the Eurasian Basin. The effect of the Bering Strait heat flux variations remains limited to the Chukchi Sea. Autonomous observing system may be able to capture the Arctic heat content variability. Sea surface temperature satellite observations combined with temperature profiles of the top 800 m in the deep Arctic covered by sea ice are sufficient to capture most of the variability signal. The results emphasize the crucial need for measurements in the Eurasian Basin.

Citation: Lique, C., and M. Steele (2013), Seasonal to decadal variability of Arctic Ocean heat content: A model-based analysis and implications for autonomous observing systems, *J. Geophys. Res. Oceans*, 118, 1673–1695, doi:10.1002/jgrc.20127.

1. Introduction

[2] There is a general agreement among the different IPCC-class climate models that the polar regions warm more than the rest of the globe when subjected to increasing levels of greenhouse gases in the atmosphere [IPCC, 2007]. This so-called “polar amplification” of climate change is primary due to ice-albedo feedback associated with changes in snow and sea ice coverage [e.g., Manabe and Stouffer, 1980].

[3] The warming of the Arctic Ocean has been clearly visible over the last few decades [e.g., Polyakov *et al.*, 2007]. Observational studies report a substantial warming affecting the different components of the Arctic system: the atmosphere [Rigor *et al.*, 2000], the Atlantic layer within the Arctic Ocean [Polyakov *et al.*, 2005], the Pacific inflow

[Woodgate *et al.*, 2006], and the surface ocean [Steele *et al.*, 2008]. The main motivation for the recent renewed interest in temperature changes of Arctic Ocean water masses is that a warming ocean might have implications for the sea ice pack. In the Canadian Basin, Shimada *et al.* [2006] and Woodgate *et al.* [2010] report that the warming of the Pacific inflow may be in part responsible for the large recent sea ice retreat observed in the Chukchi Sea, while Steele *et al.* [2010, 2011] and Jackson *et al.* [2012] report a link between the warming of the surface layer in the Canadian Basin during summer and the sea ice melt in this region. On the Eurasian side of the Arctic Basin, Polyakov *et al.* [2010] suggest that a recent increase of the Atlantic Water (AW) temperature causes an increase of the vertical heat flux to the halocline and could in turn contribute to increased sea ice melt.

[4] It remains very difficult to separate Arctic Ocean warming related to the global warming trend from that due to natural variability, which may occur on decadal or longer timescales [e.g., Polyakov *et al.*, 2004]. This is especially true because studies are often based on short observational time series that do not allow a quantification of the Arctic heat content variability. Steele *et al.* [2008] use satellite sea surface temperature (SST) observation to show that the surface layer temperature in the open water areas exhibits large interannual variations superimposed on the

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2169-9275/13/10.1002/jgrc.20127

seasonal fluctuations. Using historical hydrographic data, Bourgain and Gascard [2012] suggest that a strengthening of the inflow of warm water through Bering Strait in 2004 [Woodgate *et al.*, 2010] has caused a significant warming of the Pacific Water layer in the Canadian Basin. Substantial variability and changes affecting the Atlantic Water (AW) temperature over the past decades have been reported in the literature. Increases in the volume transport or the temperature of the AW inflow through Fram Strait [Schauer *et al.*, 2002] or through the Barents Sea route [Karcher *et al.*, 2003] are responsible for strong warming events in the AW layer; these have been subsequently traced within the Nansen and Canadian basins [Quadfasel *et al.*, 1991; Polyakov *et al.*, 2004; Polyakov *et al.*, 2005; Dmitrenko *et al.*, 2008]. On top of these interannual fluctuations, Polyakov *et al.* [2004] analyze historical hydrographic dataset over the past 100 years and suggest that the AW temperature variations might be dominated by low-frequency oscillations over a 50–80 year timescale.

[5] In these studies, authors often examine a specific water mass in the Arctic Basin. No study has yet investigated the relative contributions of the different water masses' temperature variations to the Arctic heat content in a basin-scale perspective. Here we use outputs from a global ocean/sea ice coupled model to examine the heat content and budget in the Arctic Basin and their seasonal-to-decadal variability. A rotated empirical orthogonal function (REOF) analysis is used to investigate the mechanisms responsible for this variability. We also evaluate the capacity of current autonomous observing systems to capture this variability. Based on a better understanding of the processes and mechanisms at play for heat content variations, we aim here to provide insight on how the heat content variations could be better monitored in the future.

[6] The remainder of this paper is organized as follows. The model and the simulation used for the study are described briefly in section 2. In section 3, the simulation realism is assessed, as we examine the mean and variations of the Arctic Basin heat content and budget. The modes of variability of the Arctic heat content are investigated in section 4. Finally, the ability of current autonomous observing systems to capture these modes of variability is examined and discussed in section 5. A summary and conclusion are given in section 6.

2. Numerical Experiment Description and Design

[7] The global ORCA025 coupled ocean/sea ice model configuration developed by the DRAKKAR project [The Drakkar group, 2007] is used to perform the simulation. An overall description of the model and its numerical details is given in Barnier *et al.* [2006]. This model configuration uses a global tripolar grid with 1442×1021 grid points and 75 vertical levels. Vertical grid spacing is finer near the surface (1 m) and increases with depth to 200 m at the bottom. Horizontal resolution is 27.75 km at the equator, 13.8 km at 60°N, and 10 km in the Arctic Ocean. The ocean/sea ice code is based on the NEMO (Nucleus for European Modelling of the Ocean) framework version 3.1 [Madec, 2008]. It uses a partial step representation of the bottom topography and an energy-enstrophy conserving

momentum advection scheme [Penduff *et al.*, 2007]. Parameterizations include a Laplacian mixing of temperature and salinity along isopycnals, a horizontal biharmonic viscosity, and a turbulent kinetic energy closure scheme for vertical mixing [Blanke and Delecluse, 1993]. A parametrization of the solar vertical penetration modulation by the phytoplankton concentration (using a monthly climatology of the sea surface color derived from SeaWiFS satellite observations) has also been added because it improves the representation of the sea ice conditions [Lengaigne *et al.*, 2009]. The bathymetry is derived from the 1 min resolution Etopo1 bathymetry file from the NGDC (National Geophysical Data Center). The sea ice model is the LIM model, which is a dynamic-thermodynamic model specifically designed for climate studies [Timmermann *et al.*, 2005].

[8] Our experiment runs from 1958 to 2007 with no spin-up. Initialization uses January data from the Polar Science Center Hydrographic T/S Climatology (PHC) [Steele *et al.*, 2001]. The forcing dataset is a blend of data from various origins at different frequencies, as described in details by Brodeau *et al.* [2010]. Precipitation and radiation are from the CORE dataset [Large and Yeager, 2004] at monthly and daily frequency, respectively, based on satellite observations when available. A climatology of the same satellite dataset is used for the early years up to 1979 and 1984, respectively. Air temperature, humidity, and wind speed are six-hourly fields from the ECMWF reanalysis ERA40 (1958–2001) and the ECMWF Operational Analysis (2002–2007). Turbulent fluxes (wind stress, latent, and sensible heat flux) are estimated using the CORE bulk formulae [Large and Yeager, 2004]. River runoff rates are prescribed using the Dai and Trenberth [2002] climatological dataset. To avoid an excessive model drift, the top layer salinity is restored to the monthly mean PHC climatology. The coefficient (0.027 m/day) amounts to a decay time of 371 days for 10 m of water depth; under the ice cover, restoring is five times stronger. This choice of a stronger restoring under sea ice was made within the DRAKKAR consortium because of problems encountered with earlier versions of the forcing data, especially in the Weddell Sea. The analysis presented here focuses on monthly mean output over the 40 year period 1968–2007, excluding the first 10 years of the simulation to allow for some model adjustment.

3. Simulated Arctic Heat Content and Budget

[9] The goal of this section is to assess the model performances in the Arctic Ocean, as we examine the mean and the variations of the Arctic heat budget and content, and compare them against available observations. We define the Arctic Ocean as the area enclosed by the following transects across ocean straits: Bering Strait, the Barents Sea opening (BSO) between Norway and Svalbard Archipelago (following the 20°E meridian), Fram Strait, and Davis Strait (Figure 1). These limits were chosen to enclose a domain by sections constrained by observations [e.g., Tsubouchi *et al.*, 2012]. A similar setup of the model has been used previously for Arctic studies, and extended validation of the ocean and sea ice model components can be found in Lique *et al.* [2009], Lique *et al.* [2010], and Lique and Steele [2012].



Figure 1. Arctic Ocean and localization of the main place names used in the text. The four sections enclosing the domain used for the heat budget and REOF analyses are shown as thick lines. The dashed line encloses the domain defined as Barents and Kara seas used to investigate the fourth REOF mode. Bathymetry contours 500, 1000, 2000, 3000, 4000, and 5000 m are drawn with thin lines.

3.1. Mean State

[10] Figure 2 compares the mean heat content in the Arctic Basin computed from the simulation over the period 1968–2007 and the PHC climatology [Steele *et al.*, 2001]. Here the heat content (HC) is computed as follows:

$$HC(t, x, y) = \int_z \rho_0 C_p (T(t, x, y, z) - T_{\text{ref}}) dz,$$

where T is the temperature at a given point, T_{ref} a reference temperature (chosen as -0.23°C in this study, which is the mean model temperature of our domain over the period 1968–2007; the value amounts to -0.27°C when computed from the PHC climatology), ρ_0 the density of ocean water, and C_p the ocean specific heat (4000 J/kg.K). Here we integrate the heat content from the surface to the bottom of the water column. This implies that the results integrate the space variations of the water column depth. This definition of the heat content tends to reduce the possible bias when comparing the mean from the simulation and the climatology, as the mean value of the integrated heat content at one point is partly set by depth of the ocean at this location. A comparison of the heat content integrated only in the top 1000 m of the water column for the model and the PHC climatology is shown and analyzed in Lique *et al.* [2010]. This bias does not, however, contribute to the temporal variations of the heat content, which is the focus of our study.

[11] Overall, the simulated mean heat content is in good agreement with the heat content computed from the climatology (Figures 2a and 2b). Both show warm water in the North Atlantic entering the Arctic Basin through the eastern part of Fram Strait and the Barents Sea. Warm AW is primarily advected around the Arctic Basin following the continental slopes [Rudels *et al.*, 1999; Aksenov *et al.*, 2011], which causes the heat content maximum visible on the slopes around the basin. In the interior of the Eurasian Basin, the heat content reaches its minimum, due to the deep bathymetry in this region. The simulated heat content is larger than climatology in the deep basin. The difference between the mean temperature profiles computed from the model output and the climatology in the Canadian and Eurasian basins (Figure 2d) shows that this possible bias is due to a model overestimation of the AW temperature in the intermediate layer compared to the PHC climatology. This difference could be explained, at least partly, by the fact that the climatology is heavily weighted toward the 1970s and 1980s when the Arctic Ocean (and in particular its intermediate layer) was likely in a colder state [Polyakov *et al.*, 2004]. In fact, we found that the simulated maximum temperature in the AW layer exhibits a cold bias compared to recent mooring observations conducted after 2000 [Lique and Steele, 2012].

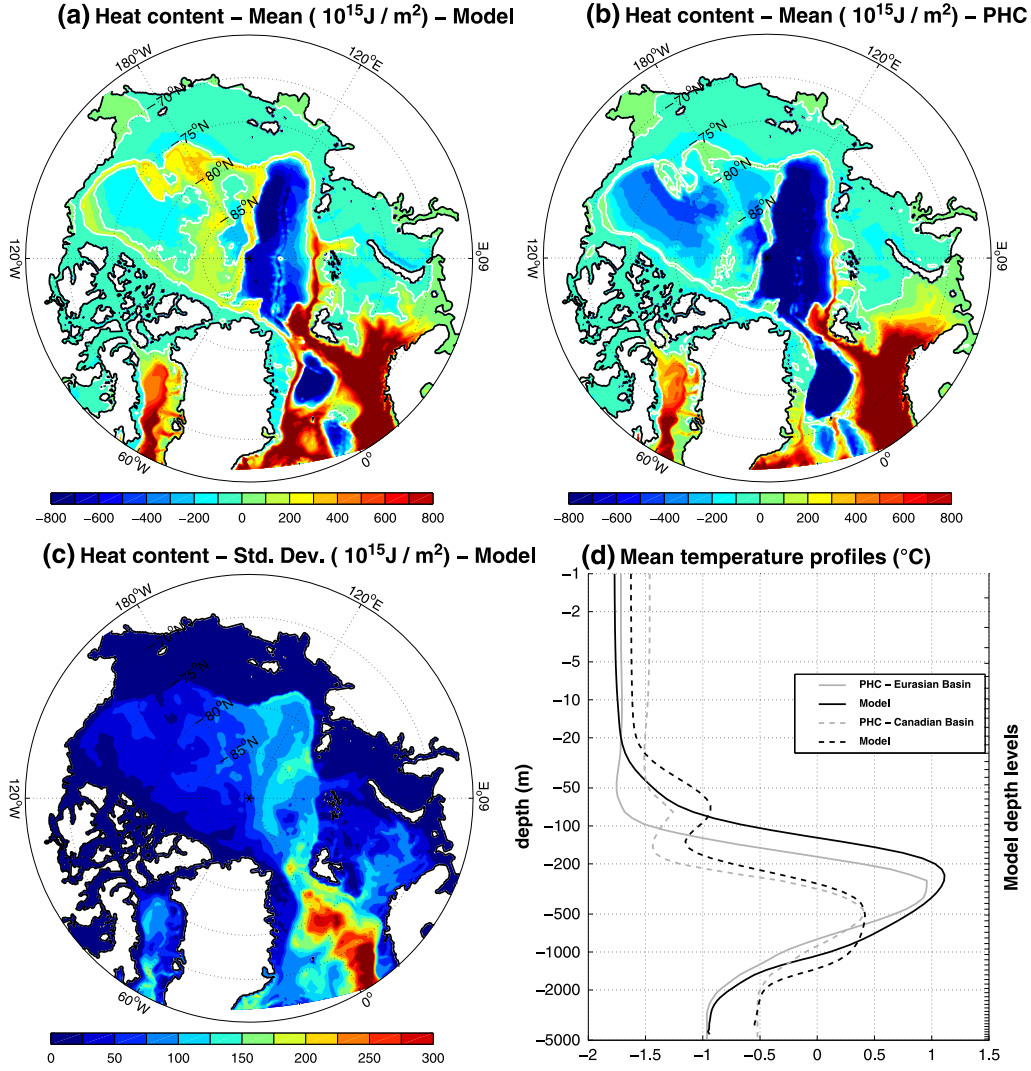


Figure 2. (a) Model mean and (b) PHC climatology heat content, computed for the whole water column, with a reference temperature of -0.23°C . The white line indicates the zero contour. (c) Standard deviation of the simulated heat content, computed from annual mean. (d) Mean temperature profiles for the Eurasian Basin (solid line) and the Canadian Basin (dashed line), for the simulation (black) and the PHC climatology (grey). The vertical scale is logarithmic. The model depth levels are shown on the right-hand vertical scale.

[12] A recent international effort has been made through observational programs (e.g., IPY, SHEBA, ASOF) to provide more accurate estimates for the different terms of the Arctic heat budget [e.g., *Serreze et al.*, 2007]. To assess the model realism, we compute the mean heat budget of the Arctic Ocean (Table 1). The heat budget for the ocean component (we do not consider here the sea ice pack) is computed as:

$$\frac{\partial HC_{tot}(t)}{\partial t} = Q_{adv} + Q_{surf},$$

where HC_{tot} is the heat content integrated over the Arctic Basin; Q_{surf} the heat flux received at the surface of the ocean, whether that surface lies under air, ice, or a combination of both (i.e., the sum of the net air–sea heat flux, the heat flux from solar radiation penetrating through sea ice, and the net ice–ocean heat flux); and Q_{adv} the total advective heat flux to the domain, computed as:

$$Q_{adv} = Q_{Bering} + Q_{Davis} + Q_{Fram} + Q_{BSO},$$

The heat flux through a section is computed as:

$$Q_{section} = \int_x \int_z \rho_0 C_p (T(t, x, z) - T_{ref}) \times V(t, x, z) dx dz,$$

where T and V are the temperature and velocity, respectively, at each grid point of the section. Note that the exchanges of heat through the sections by lateral diffusion are not considered here because they are very small compared to the advective exchanges (Table 1). Because there is a nonzero net volume flux across the different sections (Table 1), a problem arises for the computation of heat flux through a given gateway [*Schauer and Beszczynska-Möller*, 2009], and the contributions through the different gateways cannot be compared. The reader should thus keep in mind that the results presented here are dependent of the arbitrary choice of a reference temperature. In the present study, we have

Table 1. Average Arctic Ocean Heat Budget Over the Period 1968–2007^a

		Mean Volume Transport (Sv)	Heat flux (TW)	Standard Deviation (TW)	Correlation With $\frac{\partial HC}{\partial t}$
Bering Strait		1.2	5.7	2.0	−0.21
Davis Strait	total	−2.2	18.0	3.5	0
	inflow	1.2	17.4	3.9	−0.12
	outflow	−3.4	0.6	1.8	0.22
Fram Strait	total	−1.8	20.9	7.6	0.33
	inflow	5.6	30.7	9.1	0.27
	outflow	−7.4	−9.8	3.6	0
Barents Sea	total	2.5	60.5	6.1	0.43
Opening	inflow	3.8	72.1	7.5	0.39
	outflow	−1.3	−11.6	2.0	−0.16
Total Advective		−0.3	105.2	8.9	0.47
Surface heat flux			−106.2	10.6	0.21

^aThe reference temperature is -0.23°C , which is the mean temperature of the domain for this period. The mean volume transports through each section is also indicated in the first column. Means are calculated from monthly model output. Standard deviations and correlations with the heat content temporal derivative are calculated from annual means. Correlations indicated in bold front are significant above the 95% level, estimated from the effective degrees of freedom based on the integral timescales [Sciremammano, 1979]. The standard deviations of the different heat fluxes have to be compared with the standard deviation of the heat content temporal derivative (8.6 TW; see Figure 3). The sign convention is such that a source of heat into the Arctic Ocean is positive. Note that the sum of the budget terms, -1 TW, is of the same order of magnitude as the heat content change between January 1968 and December 2007 (2.3 TW). The imbalance is due to inaccuracies in the computation of the budget terms and the contribution of diffusion, which is not taken into account.

chosen the mean temperature of the Arctic Basin as a reference, following *Aagaard and Greisman* [1975]. We have also done the computation with different references (when needed) to allow direct comparison with observations.

[13] The simulated heat transports through the different sections match reasonably well the available observations. The simulated heat transport through Bering Strait agrees with the heat transport computed from mooring observations: 14.3 TW versus 10–20 TW in the study of *Woodgate et al.* [2010], with a reference temperature equal to -1.9°C in both calculations. Simulated net heat transport through Davis Strait (16 TW, relative to 0°C) is consistent with direct estimates of *Cuny et al.* [2005] (18 ± 17 TW) and *Curry et al.* [2011] (20 ± 9 TW) with the same reference. Across Fram Strait, the West Spitsbergen Current brings heat northward on the eastern side of the strait. The simulated value lies within the range of uncertainty of the different observational estimates. For instance, *Schauer et al.* [2004] find that the net heat transport varies from 16 to 40 TW (relative to -0.1°C), depending on the year considered. With the same reference, our mean value is about 20 TW. Across the BSO, our simulated heat transport (58.2 TW relative to 0°C) lies between the range of values from direct estimates by *Smedsrud et al.* [2010] (73 TW) or *Skagseth* [2008] (48 TW). The simulated surface heat flux is consistent with the estimate by *Serreze et al.* [2007] of 11 W/m^2 on the annual mean, which amounts to a total of -105.2 TW for our domain, compared to the model value of -106.2 TW.

[14] Note that all the terms accounting for the Arctic heat budget have been shown to be highly variable on seasonal-to-interannual time scales, and direct estimates made by different authors over different periods with different methods are often very different from one to another. Here we just aim at verifying that our model does not exhibit major flaw regarding the simulated heat content and budget, in order to give credibility to the results presented next.

3.2. Variability

[15] The standard deviations of the heat content, computed from annual means, show large values in the Nordic Seas (where the heat content mean is also the largest), along the slopes in the Eurasian Basin, and north of the Laptev and Kara seas (Figure 2c). In the Canadian Basin, the standard deviations remain small ($\sim 50 \times 10^{15} \text{ J/m}^2$). The pattern is coherent with the propagation of warm pulses around the Arctic within the boundary current along the basin slope [Polyakov et al., 2004].

[16] The time series of the different terms accounting for the Arctic heat budget, their standard deviations and correlations with the temporal derivative of the integrated heat content (Table 1) show that the surface flux tends to compensate the total advective flux and the two have similar standard deviations from the annual mean (~ 10 TW) (Figure 3). However, the advective term drives a larger part of the heat content interannual variability than the atmospheric heat flux, with a correlation between the total advective heat flux and the heat content temporal derivative of 0.47 (versus 0.21 for the surface heat flux). A different picture emerges when considering the seasonal cycle: the atmospheric heat flux and the heat content derivative have similar seasonal cycles (same amplitude and same timing), while the advective heat flux seasonal cycle is much smaller than that of the heat content derivative and they are not correlated (not shown). The main contribution to the mean and the interannual advective heat transport variability comes from the advection of AW through Fram Strait and the Barents Sea. The contribution through Bering Strait has a small mean value and a small standard deviation compared to the total advective heat transport. Although the mean net heat transport through Fram and Davis straits have similar amplitudes, only the Fram Strait contribution accounts for a significant part of the total heat advection variability, with a standard deviation twice as large as the Davis Strait

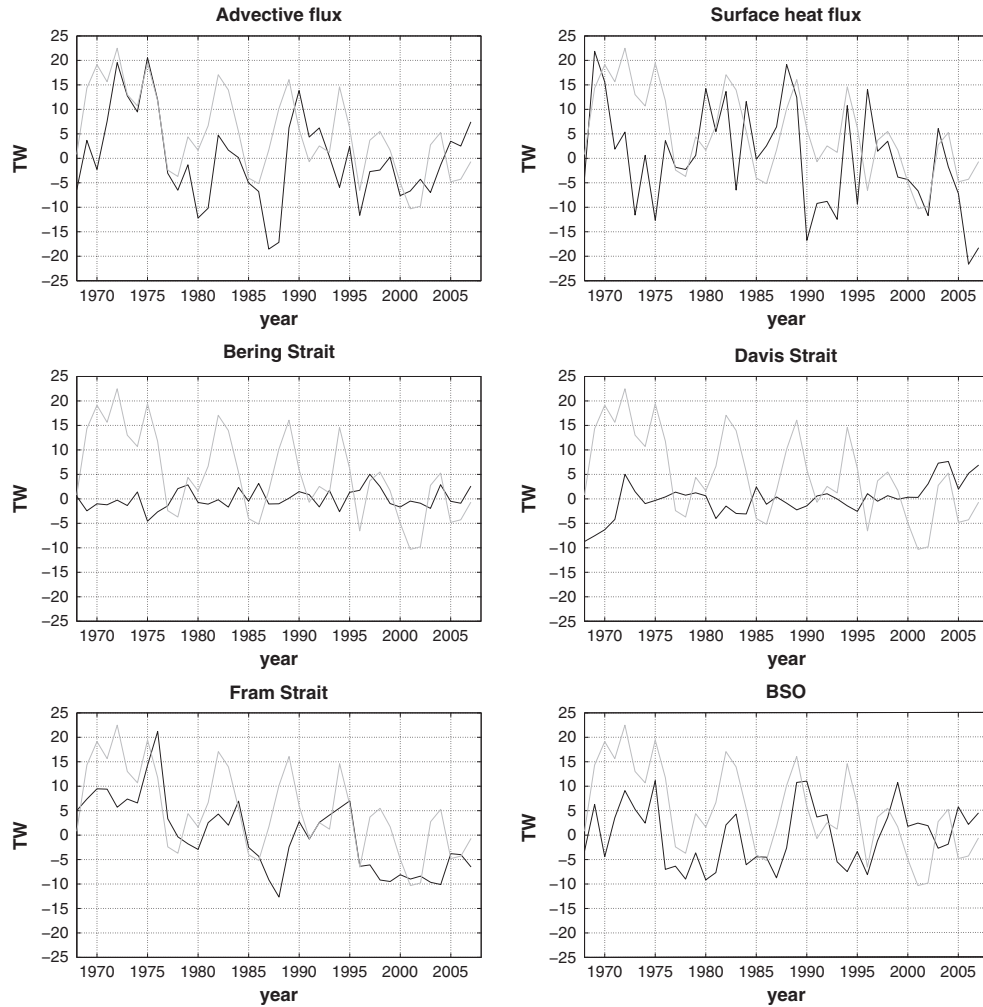


Figure 3. Time series of the annual anomalies of the different terms accounting for the Arctic heat budget. The time series of the heat content temporal derivative is also shown on each plot (gray line). The mean, standard deviation, and correlation with the heat content temporal derivative for each term are given in Table 1.

contribution, and a correlation with the heat content derivative of 0.64. The advective flux through the BSO accounts for more than one half of the domain total advective flux. Its standard deviation (6.1 TW) is also comparable with the standard deviation of the Fram Strait contribution (7.6 TW) or the heat content derivative (8.6 TW). The BSO heat transport drives a significant part of the total heat content variability, as shown by the high correlation with the heat content derivative (0.74).

[17] Model output shows that most of the Arctic heat content variability is driven by the advection of AW through Fram Strait and the Barents Sea. This is not surprising because most of the heat content mean and variations in the Arctic Basin are found in the intermediate layer of the water column where the water masses are much warmer [e.g., Polyakov *et al.*, 2004] and fed by AW. No significant correlations were found between the time series of the heat transports through Fram Strait and the BSO, suggesting that the variations of the heat inflow through the two sections may be driven by different mechanisms.

[18] Although the analysis presented above gives a coherent picture of the total Arctic heat content variability, it does

not provide any information on the spatial variations of the heat content within the Arctic Basin. A REOF analysis of the heat content is performed to identify the dominant modes of spatio-temporal variability of the heat content, and to determine the mechanisms responsible for them.

4. Arctic Ocean Heat Content Modes of Variability

4.1. Methods

[19] Here we provide details on the method used for our analysis. REOF decomposition is used widely in climate studies, and we do not aim to provide an overview of the mathematics underlying the method. The reader is referred for instance to von Storch and Zwiers [1999] for more details. The analysis is applied to monthly mean vertically integrated heat content. To focus on the seasonal-to-decadal variability, the time average (over 1968–2007) and the linear trend of the heat content time series at each grid point are removed prior to the calculation; this removes any potential effect of model drift on the results, although the results of

our analysis with or without the linear trends are very similar. When EOF decomposition is applied to an oceanic temperature dataset, the mean seasonal cycle is often removed from each grid point prior to the analysis [e.g., *Mo*, 2000; *Danielson et al.*, 2011]. This is because the ocean temperature at any location is correlated with the annual cycle of solar heating, and this creates artificial coherent variations with remote locations. Moreover, the seasonal cycle usually exhibits coherent variations with similar amplitude everywhere in the domain, and these variations would project on the first EOF/REOF mode and explain most of the variance. In the Arctic Ocean, the ocean surface is isolated from the atmosphere by the sea ice, except in spring and summer in the regions where the sea ice retreats. As a consequence, the seasonal cycle of temperature is not coherent over the basin and is highly variable from year to year at a given point, as discussed by *Lique and Steele* [2012], and is thus retained. The seasonal-to-decadal variations of the heat content can be examined, and the relative importance of the different time scales can be quantified. The same analysis has also been performed on the fields from which the mean seasonal cycle has been removed at each grid point, and the results are discussed here as necessary. Note that the results of the REOF (and especially the time series associated with some of the modes) can be sensitive to how the input fields are processed and in particular to the choice of retaining or not the seasonal cycle.

[20] REOF, rather than EOF, analysis is adopted because it has been demonstrated to be more capable at revealing spatial patterns [*Kawamura*, 1993]; it can extract two independent modes when they have similar spatial patterns. Examination of the EOF and the REOF analysis results (not shown) shows that the first three modes are identical. The fourth mode from the EOF projects on two distinct modes when the rotation is applied, although these two modes have different origins (section 4.4). In the following, we discuss only the results from the REOF decomposition, as the extracted modes from the REOF analysis are easier to interpret than the one from the EOF analysis. The reader should, however, keep in mind that the variances explained by the REOF modes are usually smaller than those by EOF modes [*von Storch and Zwiers*, 1999]. This justifies that modes of variability explaining only a small part of the total variance can still reach the 95% significance level and will thus be considered in the following.

[21] Varimax rotation is used to obtain the REOF modes [*Richman*, 1986]. The number of EOF modes selected for rotation (25) is determined with respect to the criterion established by *O'Lenic and Livezey* [1988]. Nonetheless, the results are little sensitive to the number of modes selected for the rotation.

[22] Analysis here is focused on the six leading REOF modes (Figure 4); they are significant above the 95% significance level, estimated from a Monte Carlo simulation on random data. The following modes that account for less than 2% of the total variance fall below the 95% significance level. Significance level for the REOF spatial patterns (indicated on Figure 4) corresponds to the significance of the correlation between the Principal Component (PC) time series of a given mode with the variations of the heat content at each grid point. The significance level for correlations is estimated from the effective degrees of freedom

based on the integral timescales [*Sciremammano*, 1979]. All the correlations discussed in the text are above the 95% significance level.

[23] Each mode is examined separately to investigate the physical mechanism responsible for them. Though the REOF analysis allows the data to be separated into “modes of variability,” these are primarily data modes and not necessarily physical modes. Some are easy to interpret, while others are more complex and required more investigation to be fully understood.

4.2. Mode 1: Atmospheric Heat Flux

[24] Figures 4 and 5 show the map and the PC time series, respectively, associated with the first REOF mode of heat content variability. This mode accounts for 28% of the total variance and has large values ($\sim 100 \times 10^{15} \text{ J/m}^2$) in the seasonally ice-free regions, in particular north of Bering Strait in the Chukchi Sea and in the Barents Sea. The PC has low-frequency variability (~ 10 years) superimposed on a large seasonal cycle. The spatial pattern of REOF1 mode is similar to the standard deviations of SST or the atmospheric heat flux, which vary mostly in the seasonal ice-free regions (not shown).

[25] On a seasonal time scale, PC1 is maximum in late summer (September–October), when the Barents and the Chukchi seas are ice free and exposed to the greatest solar heating, and is minimum in April, when the sea ice pack has its greatest extent and thickness, preventing exchange between the ocean surface (being at the freezing point) and the atmosphere. The correlations between the monthly PC1 and the mean SST or the surface heat flux over the Arctic domain are 0.82 (with a 1 month lag, the SST leads) and 0.81 (with a 2 month lag, the surface heat flux leads), respectively, and these correlations amount to 0.49 and 0.46 when the seasonal cycle of the time series is removed (Figure 5). These correlations suggest that the PC1 variations are driven by seasonal-to-interannual variations in the surface heat flux.

[26] When the heat content seasonal cycle is removed prior to the REOF analysis, the fourth REOF mode (which explains 9% of variance) projects on a mode localized in the Barents Sea with anomaly of the order of $80 \times 10^{15} \text{ J/m}^2$ (not shown). The PC of this mode is very similar to the PC1, from which the seasonal cycle has been eliminated by a running mean (Figure 5). This suggests that the interannual variations of the REOF1 mode occurs mainly in the Barents Sea. On an interannual time scale, REOF1 mode has warm periods in the mid-1970s and the early 1990s, which are also identified by *Carton et al.* [2011] in their examination of the variations of the AW temperature in the Nordic and Barents seas. However, *Carton et al.* [2011] also report a warm period beginning after 1999, but our PC1 does not, which indicates that this latter event may have a different origin than the two earlier events or a possible model deficiency.

4.3. Modes 2, 3, and 5: Fram Strait Heat Inflow

[27] The spatial patterns of the second, third, and fifth REOF modes of the heat content variability are shown in Figure 4, and their respective time series in Figures 6, 7, and 8. These three REOF modes are driven by the heat flux through Fram Strait. From 3 years of mooring observations, *Schauer et al.* [2004] report that the variability of the heat

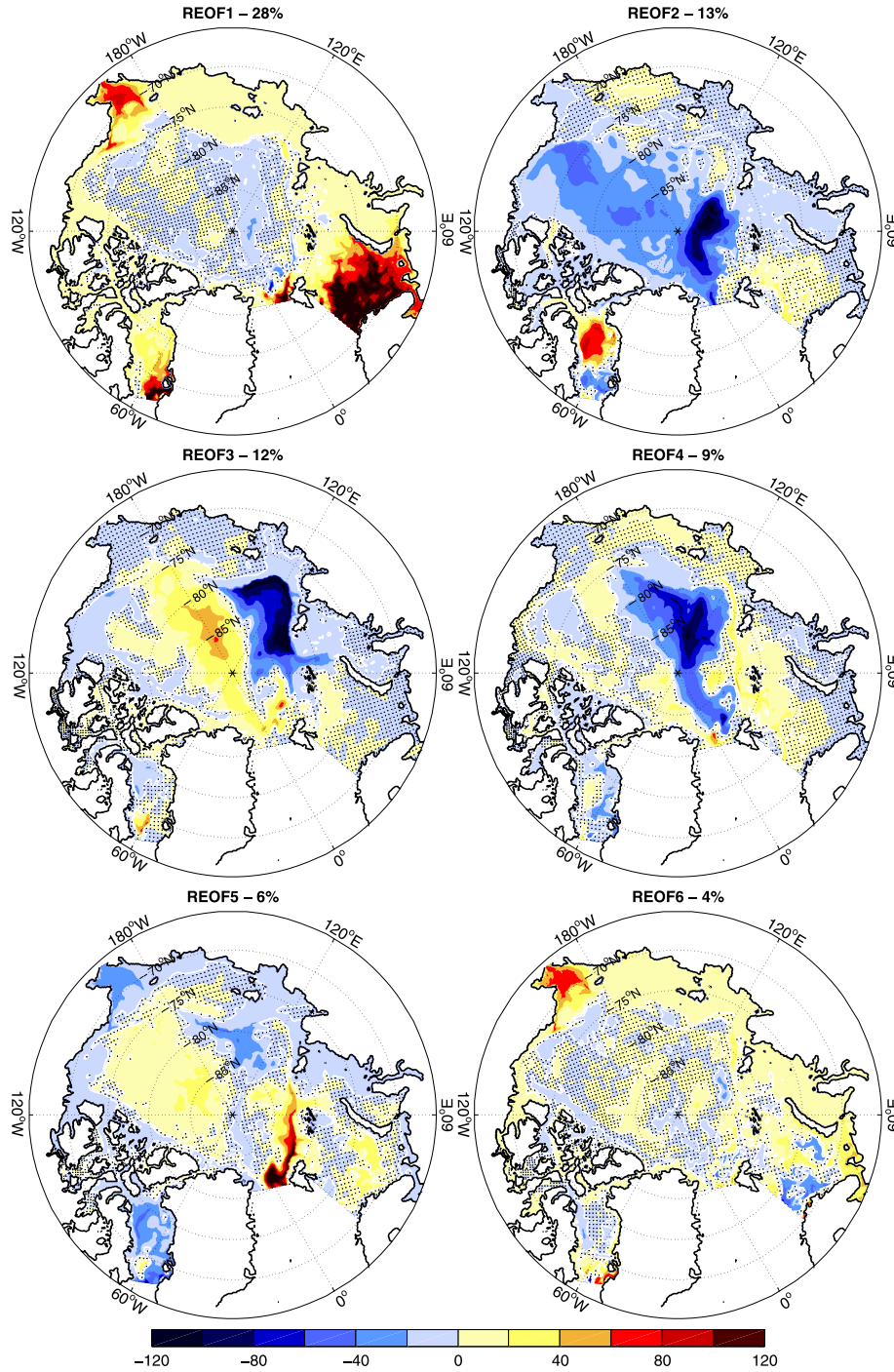


Figure 4. Spatial patterns (in 10^{15} J/m^2) of the leading six REOF modes of monthly mean heat content detrended anomalies over 1968–2007. The area where the significance falls under the 95% significance level is indicated with black dots. The percentage of variance for each mode is also indicated.

transport through Fram Strait is due equally to change in temperature and transport intensity. Here we distinguish the effects of variations of temperature and circulation on the Arctic heat content.

[28] The second and third modes account for similar parts of the variance (13% and 12%, respectively) and have patterns with similar amplitude in the Eurasian Basin. Their time series both exhibit very low-frequency variations with

little seasonal and interannual variability; this complicates the determination of their origin through statistical analyses, because their PC has only small degrees of freedom. These two modes explain a similar amount of variance and must be considered together because they might be driven by the same physical mechanism.

[29] REOF2 and REOF3 are related to different aspects of the circulation variability through Fram Strait. Using

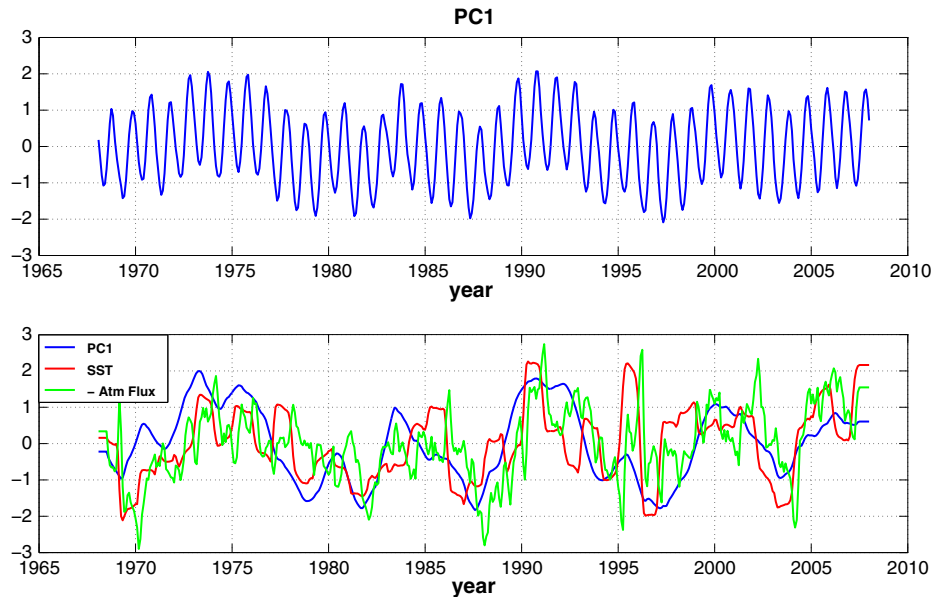


Figure 5. (top) Time series of PC1. (bottom) Time series of PC1, the mean Arctic SST, and the mean Arctic surface heat flux (positive to the surface). A running mean with a 12 month window has been applied to the time series to remove the seasonal variations, and the time series have then been normalized so that the corresponding series of monthly averages has a standard deviation of 1.

hydrographic data and floats to study the large-scale circulation through Fram Strait, *Quadfasel et al.* [1987] report that a part of the AW inflow through the strait penetrates the Arctic while another part recirculates southward. From hydrographic observations, *Bourke et al.* [1988] and *Manley* [1995] suggest that up to 50% of the AW inflow through Fram Strait could recirculate southward. The transport of AW through Fram Strait also has a large seasonal to decadal variability [*Schauer et al.*, 2008], although mechanisms responsible for these variations remain unclear. Using mooring observations supplemented by a high-resolution model, *Fieig et al.* [2010] report that the transports of the branch entering the Arctic Basin and the recirculating branch may exhibit low-frequency variations, and the variability of the two circulation features may be unrelated. Moreover, *Fieig et al.* [2010] show from model results and mooring data that the inflow and the outflow through Fram Strait are highly anti-correlated. Although the physical interpretation of these two modes is not straightforward and would require further investigation beyond the scope of this paper, we suggest that in our model, the second mode of the heat content variability is linked to the variations of the circulation pathway (i.e., change of the recirculation through Fram Strait), while the third mode is related to the intensity of the AW branch that penetrates northward to the Arctic Basin and propagates around the slope of the basin.

[30] The spatial pattern associated with REOF2 mode (Figure 4) shows a strong negative loading ($\sim -120 \times 10^{15} \text{ J/m}^2$) north of Fram Strait in the interior of the Nansen Basin, which extends over most of the Canadian Basin, and a positive loading ($\sim 60 \times 10^{15} \text{ J/m}^2$) in Baffin Bay. PC2 has a very low frequency (Figure 6), and the correlation between the PC2 fluctuations and the variations of the net volume transport through Fram Strait (with a 12 month running mean) amounts to -0.51 (Figure 6; the

net transport is negative, which explains the sign of the correlation). There is no significant correlation with the northward or southward components of the transport nor with their temperature. Figure 6 also shows that PC2 resembles the Low Frequency Oscillation (LFO) index; also, the correlation between the two time series is not significant due to the small number of degrees of freedom of the two time series. The Low Frequency Oscillation (LFO) is identified by *Polyakov and Johnson* [2000] as the low-pass filtered NAO index. *Polyakov and Johnson* [2000] discuss the influence of the LFO on the sea ice thickness and drift, while *Polyakov et al.* [2004] suggest that the LFO might drive variations of the intermediate AW layer temperature at 50–80 year time scales. We thus speculate that REOF2 mode is likely driven by the decadal-to-multidecadal variations of sea level pressure (SLP), although a longer integration of the model would be required to assess and fully understand the link between REOF2 and the LFO index.

[31] To investigate this physical mechanism likely responsible for REOF2 mode, the model ocean velocity fields, averaged between the surface and 100 m, are examined because variations of the SLP will primarily drive variations of the circulation in the surface layer. To focus on PC2 low-frequency variability, a running mean with a 12 month window is applied at each grid point. The linear trend is also removed at each grid point for consistency with the REOF analysis. Each component of the velocity field is then regressed on PC2. The regression pattern (Figure 6, bottom) suggests that when PC2 is positive, three circulation structures are intensified.

[32] First, the anticyclonic Beaufort Gyre is intensified in the surface layer of the Canadian Basin. This is also shown by the correlation between PC2 and the maximum sea surface height (SSH) detected in the Canadian Basin (Figure 6, top). SSH variations are linked to the Beaufort Gyre

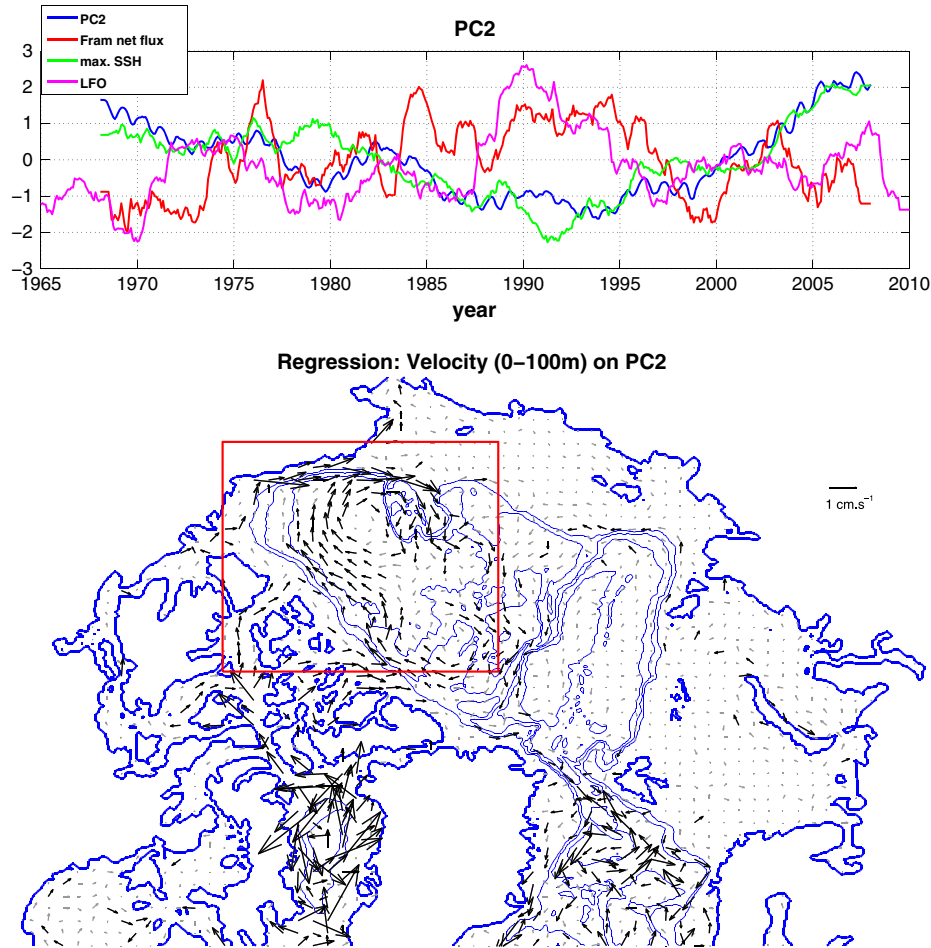


Figure 6. (top) Time series of PC2 (blue), the net transport through Fram Strait (red), the SSH index (green), and the low-frequency time series of the AO index (magenta) obtained with a 36 month window running mean. A running mean with a 12 month window has been applied to the Fram Strait net transport time series and the SSH index time series to remove the seasonal variations, and the time series has then been normalized so that the corresponding series of monthly averages has a standard deviation of 1. The SSH index corresponds to the maximum SSH detected within the red box indicated in the bottom figure. This is a proxy of the Beaufort Gyre intensity. (bottom) Detrended and smoothed (with a 12 month running mean) velocity averaged between the surface and 100 m projected on PC2. Black arrows highlight the vectors that are significant above the 95% confidence level. Blue line contours indicate the bathymetry with 1000 m isobath intervals. The field is shown on the native model grid, and one point every eight is displayed.

intensity, as predicted for instance by *Proshutinsky and Johnson* [1997], reporting that an increase of the Beaufort High over the Canadian Basin (occurring during an LFO negative phase, *Polyakov and Johnson* [2000]) yields an intensification of the Beaufort Gyre and an increase of the SSH in the interior of the gyre through Ekman pumping. This intensification of the anticyclonic circulation in the surface layer tends to decrease the intensity of the cyclonic circulation of the AW intermediate layer (not shown) and thus limit the penetration of AW from the Eurasian Basin. This circulation pattern is consistent with the circulation scheme of the AW intermediate layer proposed by *Karcher et al.* [2012] when the anticyclonic circulation in the surface layer is intensified. As warm AW preferentially recirculates in the Eurasian Basin along the Lomonosov Ridge (instead

of entering the Canadian Basin), it leads to a cold anomaly in the Canadian Basin (Figure 4).

[33] Second, the intensification of the Beaufort Gyre also tends to reduce the outflow through the Canadian Arctic Archipelago (CAA), as suggested by *Jahn et al.* [2010]. PC2 is well correlated with the net transport through the CAA ($r = 0.6$). When PC2 is positive, the export of cold water through the CAA decreases, as shown by the northward flow in the CAA in the regression map (Figure 6, bottom). Associated with an intensification of the cyclonic circulation in Baffin Bay, it yields a warm anomaly there, where relatively warm waters are also advected within the northward flowing West Greenland Current.

[34] Third, an intensification of the cyclonic gyre in the Greenland Sea south of Fram Strait likely tends to limit

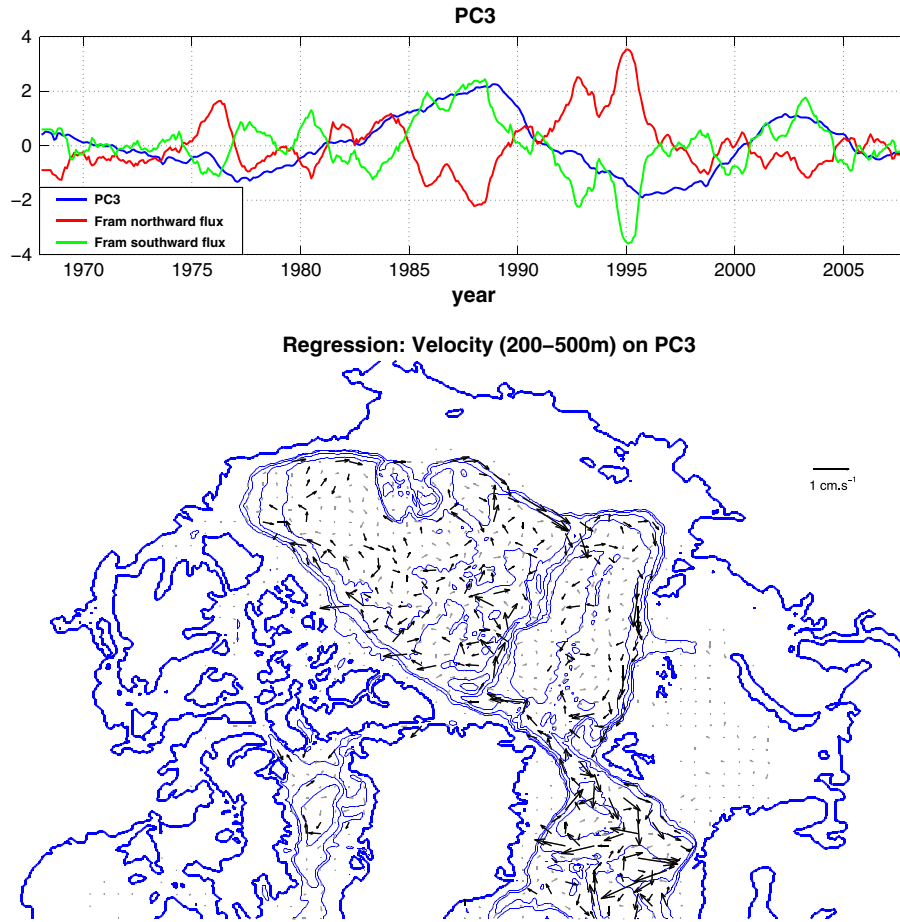


Figure 7. (top) Time series of PC3 (blue), and the northward (red) and southward (green) volume transports through Fram Strait. A running mean with a 12 month window has been applied to the transport time series to remove the seasonal variations, and the time series has then been normalized so that the corresponding series of monthly averages has a standard deviation of 1. (bottom) Detrended and smoothed (with a 12 month running mean) velocity averaged between 200 m and 500 m projected on PC3. Black arrows highlight the vectors that are significant above the 95% confidence level. Blue line contours indicate the bathymetry with 1000 m isobath intervals. The field is shown on the native model grid, and one point every eight is displayed.

the penetration of the AW inflow in the Eurasian Basin and thus lead to an intensification of the recirculation south of Fram Strait (Figure 6, bottom). When the recirculation is intensified, the inflow of warm water in the Eurasian Basin decreases, leading to a cold anomaly in the Eurasian Basin north of Fram Strait. This situation of intensified recirculation has been discussed in detail by *Lique et al.* [2011], as this dynamical anomaly was found to be responsible for a strong salinisation signal north of Fram Strait in 2007 and 2008 (note that PC2 reaches its maximum in 2007). *Polyakov et al.* [2004] suggest a link between the LFO and the intensity of both the inflow and the outflow through Fram Strait, although a mechanism underlying this relation would required further investigations to be fully understood.

[35] The spatial pattern associated with REOF3 mode (Figure 4) has a strong negative loading ($\sim -120 \times 10^{15} \text{ J/m}^2$) in the Nansen Basin, located north of the Laptev Sea, and a positive loading ($\sim 50 \times 10^{15} \text{ J/m}^2$) along the Canadian side of the Lomonosov Ridge, which extends up to the western part of Fram Strait. PC3 has a low-frequency

variability, with two positive phases in the late 1980s and early 2000s, and two negative phases in the late 1970s and 1990s (Figure 7). The maximum correlations we could find between the variations of PC3 and those of the northward and southward volume transport through Fram Strait (with a 12 month window running mean) are -0.58 and 0.66 with a 12 month lag, respectively (the two transport time series variations lead PC3 variations). As the main anomaly of this mode is in the Laptev Sea, the lag is consistent with the advection time from Fram Strait to the Laptev Sea, which is about one year in the model [*Lique and Steele*, 2012]. No correlation can be found however with the net transport. Note that the simulated southward and northward transport variations through Fram Strait are strongly correlated with a coefficient of -0.90 , which is consistent with the results of *Fieg et al.* [2010] based on model simulations and mooring observations. The forcing mechanism of this anticorrelation is, however, not straightforward to understand.

[36] The regression of the velocity in the AW layer (i.e., averaged between 200 m and 500 m) on PC3 does not results

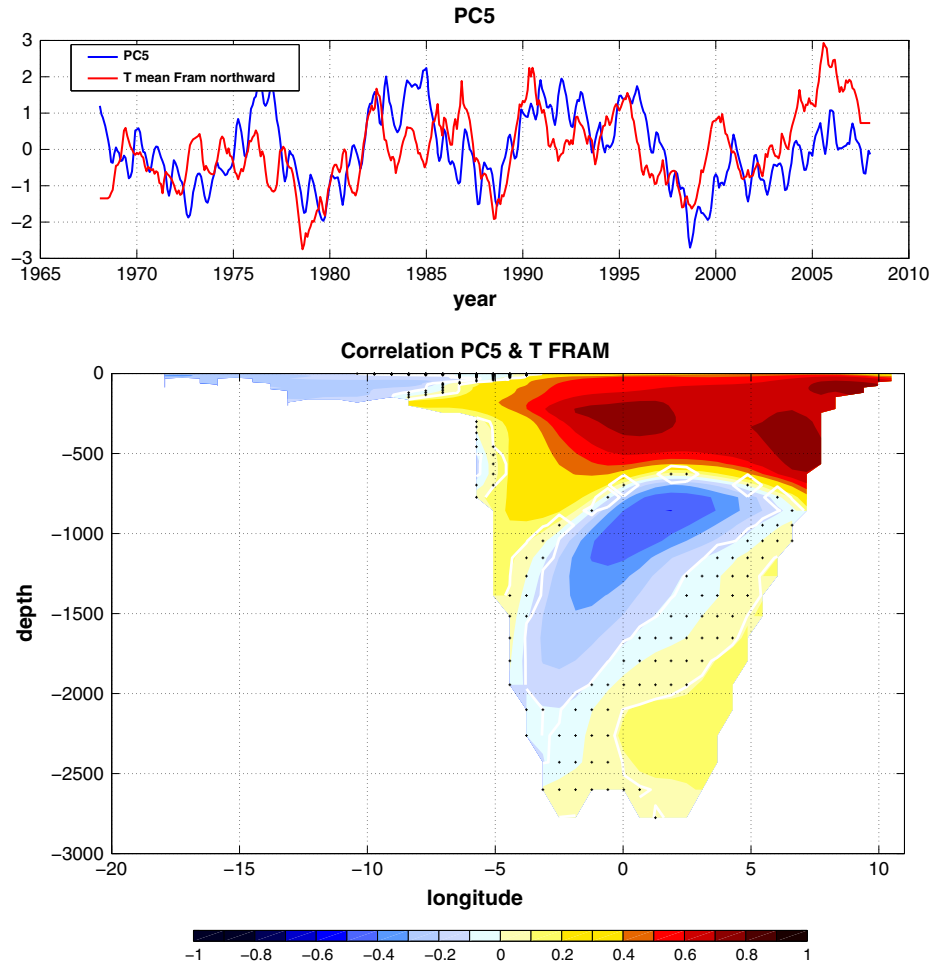


Figure 8. (top) Time series of PC5 (blue), and the mean temperature of the northward flux through Fram Strait (red). A running mean with a 12 month window has been applied to the temperature time series to remove the seasonal variations, and the time series has then been normalized so that the corresponding series of monthly averages has a standard deviation of 1. (bottom) Correlation between the monthly mean temperature through Fram Strait and PC5. The white line indicates the 95% significance level and the area where the significance falls under the 95% significance level is indicated with black dots. A running mean with a 12 month window has been applied to the temperature time series at each grid point.

in a clear pattern (Figure 7, bottom), probably due to the small degree of freedom of PC3. It suggests a link between PC3 and the intensity of the inflow and outflow branches through Fram Strait, as well as an intensification (meaning an acceleration when PC3 is negative) of the boundary current along the continental slope from Fram Strait to the Chukchi plateau. No significant correlation could be found between PC2 and the velocity, suggesting that the two modes have different drivers. From Figure 7 (bottom), we speculate the following scenario to explain the REOF3 mode: when PC3 is negative, the cyclonic boundary current at intermediate depth along the slopes of the Eurasian Basin increases, and the intensity of the inflow through Fram Strait increases. This brings more heat into the Eurasian Basin through the AW currents flowing along the continental slope and leads to a warm anomaly 1 year later in the Eurasian Basin, north of the Laptev Sea. The largest anomaly is found on the slope from St. Anna Trough up to the Lomonosov Ridge. Although the pattern of REOF3 shows a signal over St. Anna Trough and downstream in the Laptev Sea, we

could not find any correlation between PC3 and any time series corresponding to the St. Anna Trough inflow from the Barents Sea (i.e., with the inflow and net volume transport, the heat transport, the volume of dense water formed in the Barents Sea, or the temperature). This suggests that REOF3 mode is not driven by the AW Barents Sea Branch. Instead, we speculate that the signal found in St. Anna Trough is related to the propagation along the Nansen Basin slope of the Fram Strait Branch that enters and exits St. Anna Trough as it follows the bathymetry [Schauer *et al.*, 2002]. It is, however, surprising that the negative anomaly is limited downstream of St. Anna Trough and does not extend to Fram Strait. We speculate that a change of the AW inflow intensity might not influence too much the region north of Fram Strait, where the waters are always warm, but instead has its larger signature downstream where the quantity of heat advected depends on the intensity of the AW branch. Figure 7 (bottom) also suggests that when PC3 is positive, the intensity of the branch of AW entering the Canadian Basin north of the Laptev Sea decreases, causing a

continuation of the cold anomaly north of the Laptev Sea in the Canadian side. At the same time, Figure 7 (bottom) also suggests that the outflow of AW from the Canadian Basin to the Eurasian Basin in the Lincoln Sea north of the Greenland shelves is reduced during a positive phase of PC3, leading to the retention of warm AW in the Canadian Basin and thus a warm anomaly.

[37] Figure 7 (bottom) suggests the existence of a basin scale coherent mode for the circulation pattern (or at least in the Eurasian Basin), similar to the sub-monthly basin scale coherent mode of the Arctic bottom pressure described by *Peralta-Ferriz et al.* [2011]. They report that this mode of variability is driven by winds over the Nordic Sea that force a northward geostrophic slope current in the Arctic Basin. This basin scale coherent mode could explain the anti-correlation between the inflow and the outflow branch, although it requires further investigation, which is beyond the scope of this paper.

[38] REOF5 mode exhibits a large anomaly ($\sim 100 \times 10^{15} \text{ J/m}^2$) localized completely north of Fram Strait and along the continental slope in the Nansen Basin from Fram Strait up to St. Anna Trough. Its associated time series

has small seasonal variations superimposed on larger variations with a 10–15 years frequency. This mode of variability accounts for 6% of the total variance. It is related to the temperature variations of the AW inflow through Fram Strait. Large correlations between the PC5 fluctuations and the variations of the temperatures through Fram Strait are found in the AW inflow core, on the eastern side of the strait where temperatures are highest (Figure 8). The correlation between the variations of PC5 and those of the integrated temperature anomaly of the northward flow through Fram Strait is 0.48, and increases to 0.66 when a running mean with a 12-month window is applied to the two time series.

4.4. Mode 4: Barents Sea Branch

[39] The spatial pattern (Figure 4) and the time series (Figure 9) of the REOF4 mode of the heat content variability, which accounts for 9% of the total variance, project on a negative structure in the central Arctic, with a minimum of $\sim -100 \times 10^{15} \text{ J/m}^2$ along the Eurasian side of the Lomonosov Ridge. PC4 exhibits mainly a low-frequency variability with a periodicity of ~ 15 years. The physical interpretation of the mode is not straightforward. Although

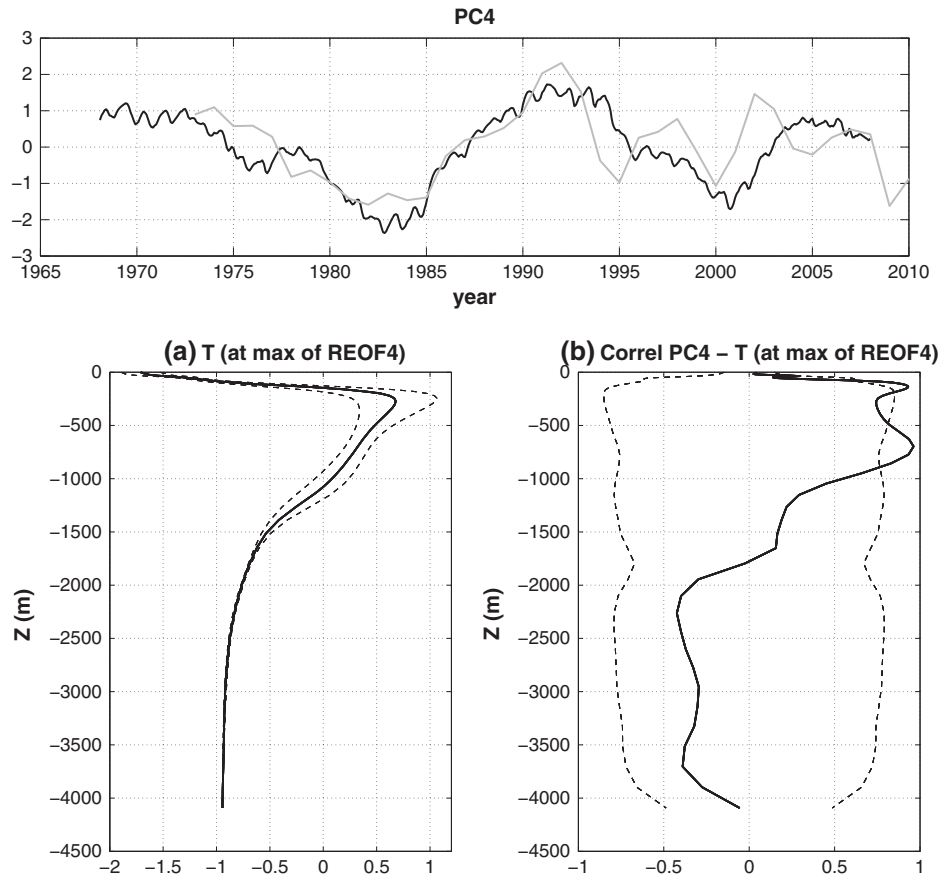


Figure 9. (top) Time series of PC4 (black), and the annual cumulative volume of waters with a density heavier than 1027.9 kg/m^3 in the Barents and the Kara seas with a 5 year lag (gray). The second time series has then been normalized so that the corresponding series of monthly averages has a standard deviation of 1. (bottom) Mean temperature profile (a) and correlation between PC4 and the variations of the temperature at each model vertical level (b) at the location of the REOF4 maximum (85.2°N and 135.4°E). The dotted lines indicate the standard deviation of the temperature at each model vertical level computed from monthly means (a) and the 95% significance level (b).

the pattern extends to north of Fram Strait, there is no significant correlation between PC4 and any volume transport, heat transport, or temperature time series associated with the inflow or outflow branches through Fram Strait. This suggests that this mode is not driven by the exchange through Fram Strait. Figure 9 (bottom) shows the correlation between PC4 and the temperature variations at each model vertical level, at the location of the REOF4 maximum anomaly. Comparison with the mean temperature profile at this location reveals that PC4 is not correlated with the temperature variations close to the depth of the temperature maximum. From hydrographic data, *Schauer et al.* [2002] report that the lower part of the AW layer originates from the Barents Sea, as the Barents Sea branch flowing out of St. Anna Trough tends to flow under the warmer Fram Strait branch. The high correlations found between 500 and 900 m suggests that REOF4 mode may originate from the Barents Sea.

[40] *Häkkinen and Geiger* [2000] performed an EOF analysis of the barotropic streamfunction in an ocean/sea ice model. They report that the second mode of variability, which explains 9% of the variance, contains two gyres with opposing cyclonicity in the Eurasian and Canadian basins (the intensity of the gyre in the Eurasian Basin is twice larger as the second one), and the associated time series has a 14 years cycle. They suggest that this mode of variability is driven by the variability of the dense water formation in the Barents Sea, associated with the variability of oceanic and atmospheric heat fluxes.

[41] Here the main structure of REOF4 mode looks very similar to the main anomaly of transport reported by *Häkkinen and Geiger* [2000] in the Eurasian Basin. The interannual variations of PC4 are well correlated with the annual cumulative volume of waters with a density heavier than 1027.9 kg/m^3 in the Barents and the Kara seas ($r = 0.66$, with a 5 year lag, the volume variations lead; see Figure 9). This density corresponds to that of the water masses flowing out of St. Anna Trough [*Schauer et al.*, 2002]. In the model, water masses with density heavier than 1027.9 kg/m^3 are found, on average, below 500 m in the Eurasian Basin, i.e., below the Fram Strait branch. We thus speculate that REOF4 mode is driven by the variations of the Barents Sea branch intensity that exits St. Anna Trough and flows along the slope of the Eurasian Basin. This is consistent with results reported in *Karcher et al.* [2003], who suggest that change of the inflowing water through the Barents Sea contributes to a temperature anomaly within the AW layer in the Arctic Basin.

[42] *Yang* [2005] and *Karcher et al.* [2007] report that the inflow of dense water from the Barents Sea is the main source of high potential vorticity for the Eurasian Basin and controls the cyclonic circulation of the intermediate layer in this basin. This may explain the remote effect of the dense water inflow through St. Anna Trough that yields heat content anomalies along the Lomonosov Ridge. In fact, a map of the regression of the velocity field averaged between 300 m and 800 m on PC4 shows an intensification of the cyclonic circulation around the Eurasian Basin (not shown). Correlation analysis reveals no simple relation between PC4 and the heat transport through BSO nor with the surface heat flux over the Barents Sea, although the dense water formation in the Barents Sea results from a delicate interplay between the

quantity of heat advected through the BSO and the quantity of heat release to the atmosphere, which is modulated by the position of the sea ice edge [e.g., *Aksenov et al.*, 2010]. Further analysis is required to fully understand the mechanisms responsible for the variations of the dense water export from the Barents Sea, which might also be modulated by atmospheric and oceanic conditions responsible for periods of accumulation of dense water on the Barents shelves, as suggested by *Häkkinen and Geiger* [2000].

4.5. Mode 6: Bering Strait Heat Inflow

[43] The spatial pattern (Figure 4) and the time series (Figure 10) of the REOF6 mode of the heat content variability, which accounts for 4% of the total variance, are associated with a local signal north of Bering Strait, in the Chukchi Sea, with an amplitude of $\sim 60 \times 10^{15} \text{ J/m}^2$, as well as a signal with opposite sign in the Barents Sea with an amplitude of $\sim 40 \times 10^{15} \text{ J/m}^2$. PC6 has a large seasonal cycle, with a maximum in August–September and a minimum in December–January. On top of these seasonal variations, PC6 also has important lower frequency variations, correlated with the mean temperature through Bering Strait ($r = 0.53$), but not with the heat transport through the strait at any lag or lead. This suggests that the temperature variations of the Pacific inflow are the main driver for the sixth leading mode of the heat content variability.

[44] The fluctuations of PC6 are also correlated significantly with the Pacific Decadal Oscillation (PDO) Index ($r = 0.59$); the PDO Index is the leading principal component of north Pacific monthly sea surface temperature variability (<http://jisao.washington.edu/pdo/>). *Danielson et al.* [2011] report large correlations between the variations of surface and subsurface temperatures in the Bering Sea and the PDO Index. This shows the influence on the ocean temperature of the large-scale fields of atmosphere-ocean heat fluxes. Computing the correlation between PC6 and SST variations at each grid point (not shown) yields higher correlations upstream from Bering Strait than downstream. This suggests that temperature variations in the Bering Strait control the temperature of the Bering Strait heat inflow and lead to an anomaly of heat content north of the strait in the Chukchi Sea. Although the amplitude of the loading pattern associated with REOF6 mode remains small, these anomalies occurs in the surface layer and might thus have significant implications for the sea ice pack [*Shimada et al.*, 2006; *Woodgate et al.*, 2010].

[45] Surprisingly, an anomaly associated with REOF6 can also be found in the Barents Sea. *Stepanov et al.* [2012] find a negative correlation between ENSO events and Barents Sea water temperatures in the top 200 m, and suggest that it could be due to atmospheric teleconnections [*Stepanov et al.*, 2012]. In the North Pacific, PDO and the ENSO influence sea surface temperatures, sea level pressure, and surface winds in similar ways [*Gershunov and Barnett*, 1998]. Thus, the PDO and ENSO might have similar teleconnections with the Barents Sea, which could explain the signal in the Barents for the REOF6 mode. More investigation would be required to fully understand this possible teleconnection between the PDO and the ocean temperature in the Barents Sea.

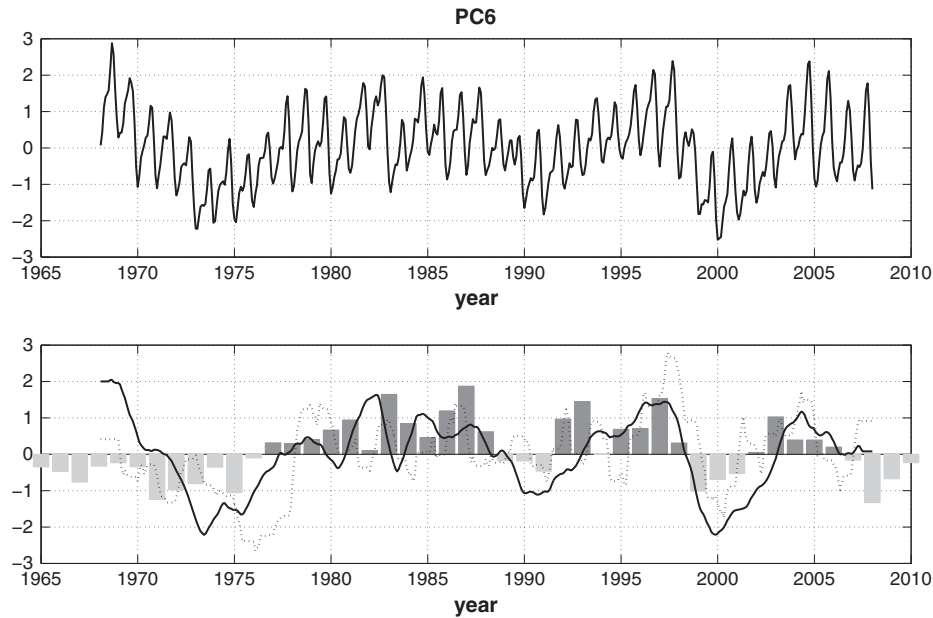


Figure 10. (top) Time series of PC6. (bottom) Time series of PC6 (black line), the mean temperature of the Bering Strait inflow (black dotted line), and the annual PDO index (gray). A running mean with a 12 month window has been applied to the temperature time series to remove the seasonal variations, and the time series have then been normalized so that the corresponding series of monthly averages has a standard deviation of 1.

5. Capability of Autonomous Observing Systems to Measure Heat Content Variability

[46] Here we estimate the capability of the observations conducted in the Arctic Basin to capture the spatial and temporal variability of the Arctic heat content and budget. The focus is only on the autonomous observing systems. The goal is not to provide a model-observation comparison; problems arising from observational data processing or subsampling problems are not addressed here, and the hypotheses made are highly idealized. For instance, we consider that the moorings sample the temperature on 75 levels, from the surface to the bottom, while in reality, only a few depths are sampled. Thus, we call the moorings considered here “virtual moorings,” and we assume that the virtual moorings provide an estimation of the heat content at a given grid point. Similarly, only a few moorings are deployed across the sections enclosing the Arctic Basin, while we use here temperature and velocity at each grid point along the straits to compute the heat transports. The consequences of this observational subsampling are discussed for instance by *Fieg et al.* [2010].

5.1. Mooring Observations

[47] International efforts have accelerated recently to monitor the exchanges between the Arctic Ocean and the surrounding regions. Moorings have been deployed along the four sections enclosing the basin, with the exception of the northern part of the BSO section. *Tsubouchi et al.* [2012] use data from these moorings as well as quasi synoptic hydrographic measurements to estimate the heat flux through the different sections. Other studies focus on the estimate of the heat flux through one specific section [see the recent synthesis by *Beszczynska-Möller et al.*, 2011].

Neglecting the inaccuracies between the different estimates and the observational approximations and errors, one can imagine that all the terms accounting for the advective component of the heat content budget may be estimated directly from observations, and atmospheric reanalysis provides a quantification of the surface heat flux over the Arctic domain. Thus a direct estimate of the total Arctic heat content temporal derivative could be obtained.

[48] However, the estimate of the heat transport through the sections does not provide information on the spatial variations of the heat content within the Arctic Basin, though changes of the heat content spatial distribution may have large consequences for the sea ice pack. Here the correlations are computed between the temporal variations of the heat content integrated along each section (as measured by virtual moorings located at every model grid point along the different sections) and the temporal variations of the vertically integrated heat content at each model grid point (Figure 11). The same calculations were performed with the heat transport through the sections (instead of the mean heat content) yielding similar patterns (not shown).

[49] Overall, the influence of the temperature variations through the different sections remains localized to the region of the section, with the exception of Davis Strait. Through Bering Strait, the fluctuations of the inflow temperature impact significantly the heat content mostly in the Chukchi Sea north of the strait. The pattern is similar to the pattern of REOF6 mode; the heat content averaged along the strait is correlated with PC6. Although the Pacific inflow is warmer than the Arctic water, the heat is quickly released to the atmosphere. Though the effect may remain localized, the Pacific inflow temperature increase may cause increased sea ice melt in the Canadian Basin [*Woodgate et al.*, 2010], because the Pacific Water circulates at or near surface and

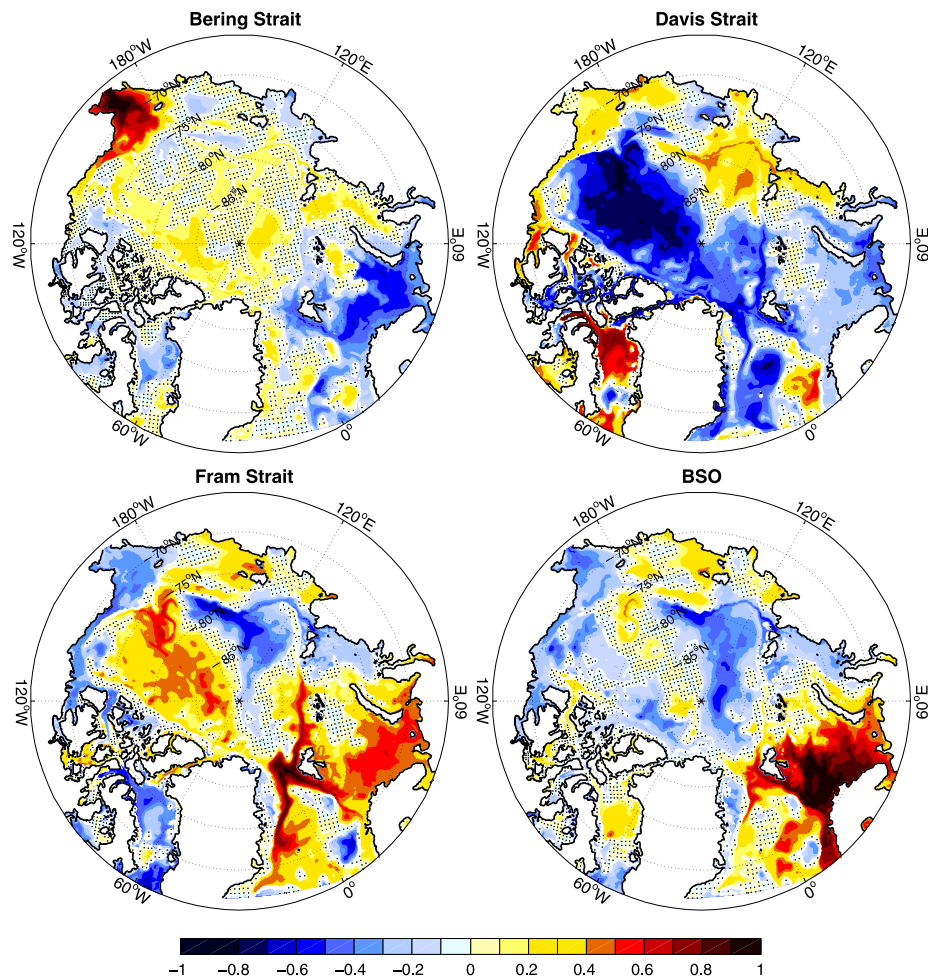


Figure 11. Correlation between the heat content integrated along each section and the heat content at each model grid point. The area where the significance falls under the 95% significance level is indicated with black dots. A 12 month window running mean was applied to each time series before computing the correlations.

is in close contact with the sea ice pack. Because the temperature variations of the Bering Strait inflow are driving the REOF6 mode, which exhibits a loading pattern in the Barents Sea, correlations with the heat content variations are found in the Barents Sea. A large amount of heat penetrates the Arctic domain through the BSO (Table 1), but most of the heat is released quickly to the atmosphere in the Barents Sea. This explains the large correlation found in the Barents Sea.

[50] The heat content variations along Fram Strait drive heat content variations along the continental slope in the Eurasian Basin, where AW is primarily advected [Aksenov *et al.*, 2011], as well as along the East Greenland Current in the recirculating branch. The pattern is similar to that of REOF5 mode, which is coherent with the correlation between PC5 and the inflow mean temperature. Though the heat transport through Davis Strait is not correlated with the total heat content derivative (Table 1), there are significant anti-correlations with the heat content through the strait in a large part of the basin, and in particular in the Canadian Basin, as well as significant correlations in the Baffin Bay. The pattern is somewhat similar to the spatial distribution

of REOF2 mode. The temperature variations through Davis Strait may be only driving variations of the Baffin Bay heat content through the West Greenland Current flowing northward along Greenland and recirculating in the Baffin Bay. As shown by REOF2 mode, the heat content variations in the Baffin Bay are anti-correlated with the heat content variations in the Canadian Basin, leading to an anti-correlation between the heat content variations in the Canadian Basin and the temperature of the exchange through Davis Strait. Contrary to the other sections, the correlations of the heat content with the mean Davis Strait temperature remain small at each grid point along Davis Strait, due to the fact that the anomalies of the heat content on the east and west sides of the strait tend to cancel out each other.

[51] In addition to the moorings deployed across the sections enclosing the Arctic domain, several moorings have been deployed within the basin at various locations. Here we consider the same set of moorings as used in Lique and Steele [2012], namely, moorings from the NABOS (Nansen and Amundsen Basins Observational System) [Polyakov *et al.*, 2003], NPEO (North Pole Environmental Observatory) [Morison *et al.* 2002], and BGOS

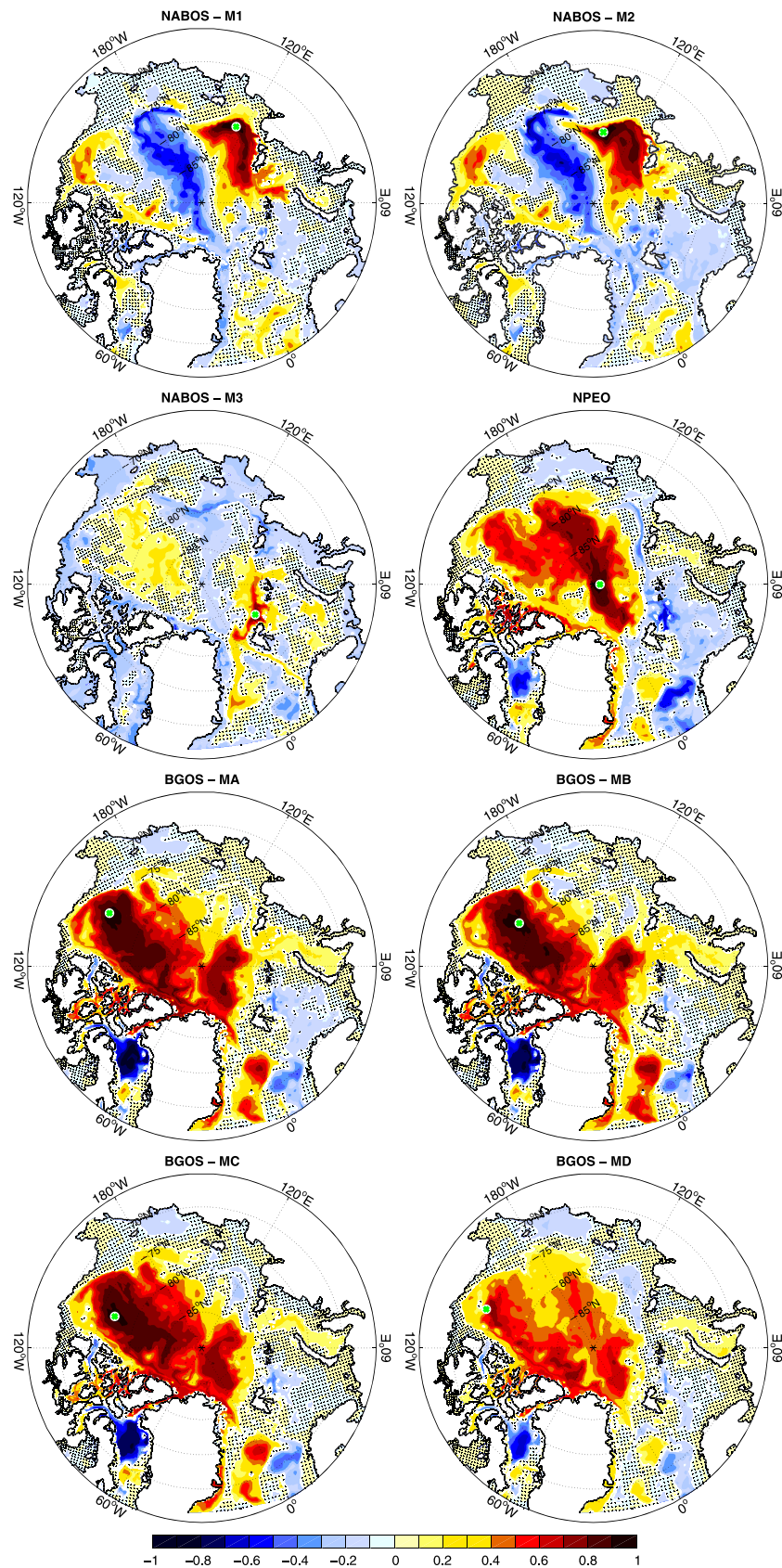


Figure 12. Correlation between heat content at the different mooring locations and the heat content at each model grid point. The area where the significance falls under the 95% significance level is indicated with black dots. The position of the mooring considered is indicated with a green star. A 12 month window running mean was applied to each time series before computing the correlations.

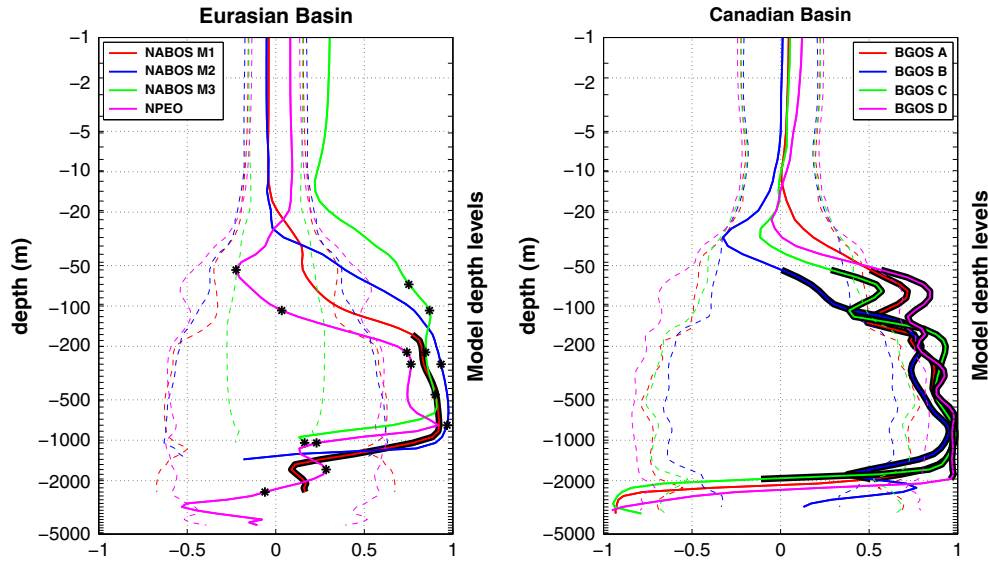


Figure 13. Correlation between the mean temperature of the water column and the temperature at each model depth level, for virtual moorings in the Eurasian (left) and Canadian (right) basins. The thin dotted lines indicate the 95% significance level. The location of the mooring can be seen on Figure 12. The vertical scale is logarithmic, and the model depth levels are shown on the right side vertical scale. The depths where the real moorings are sampling temperature with instruments at fixed depth (NPEO, NABOS M2 and M3) are indicated with a black star. For the M2 and M3 NABOS mooring, average depth of the instruments is shown because they have been deployed at different depths over the years. For profiling moorings (BGOS and NABOS M1), the sampled part of the water column is indicated with a black line.

(Beaufort Gyre Observing System) [Proshutinsky *et al.*, 2009] projects. More details about the instruments and the realism of the simulation regarding the temperature at these locations are given in Lique and Steele [2012]. Correlations are calculated for the vertically integrated heat content at each model grid point, and the vertically integrated heat content at the virtual mooring locations (Figure 12).

[52] Except for the NABOS M3 mooring, all the virtual moorings capture the variations of heat content over a large part of the basin (Figure 12). The four virtual moorings in the Beaufort Gyre (similar to the BGOS moorings) as well as the NPEO virtual mooring close to the North Pole capture the second leading mode of the heat content variability (REOF2 mode), and thus, we find large correlation (over 0.7) in the Canadian Basin and a part of the Eurasian Basin north of Fram Strait. Note that the BGOS virtual mooring MD, situated closer to the continental slope, captures a smaller part of the REOF2 mode and thus the correlations of the heat content at this specific location with the heat content at each grid point in the basin are lower. The NPEO virtual mooring captures the fourth leading mode in addition to the second mode, which explains the anticorrelations along the continental slope in the Eurasian Basin. In the Eurasian Basin, the NABOS virtual moorings M1 and M2 capture the third mode of variability, and thus, Figure 12 shows a dipole, with large correlations north of the Laptev Sea in the Eurasian Basin, and large anticorrelations along the transpolar drift pathway. Regarding the NABOS virtual mooring M3 located just downstream from Fram Strait, the correlations are only significant along the slope in the Eurasian Basin. This virtual mooring (similarly to the heat content variations along Fram Strait; Figure 11), only captures the

REOF5 mode, whose signal is localized downstream from Fram Strait.

[53] We consider here that the temperature is sampled at each of the 75 grid model vertical levels. In reality, instruments are only deployed at a few fixed depths. To get a sense of the optimal instruments deployments depths to capture the variations of the vertically integrated heat content at each location, correlations are computed between the heat content variations (integrated over the whole water column) and the variations of the temperature at each model vertical level, for each of our virtual moorings (Figure 13). Comparing the results with the mean temperature profiles (Figure 2d) reveals that the highest correlation is found in the intermediate AW layer, below the depth of the temperature maximum (~ 300 m in the Eurasian Basin and ~ 500 m in the Canadian Basin), although high correlations are found everywhere in the AW layer. With the exception of the NPEO mooring, all the moorings might capture the variations of the vertically integrated heat content at these locations with a reasonable accuracy.

5.2. Observations by Profiling Floats and Satellites

[54] In addition to the mooring observations discussed previously, we also consider the capability of two other types of autonomous observing systems to capture the variations of the heat content: Ice-Tethered Profilers (ITP) [Toole *et al.*, 2011] and SST satellite observations [e.g., Steele *et al.*, 2008]. The satellite observations of SST provides measurement of the very thin surface layer. Under sea ice, the SST equals the freezing point temperature, so the SST data provide information only on the seasonal ice cover zone. The ITP program

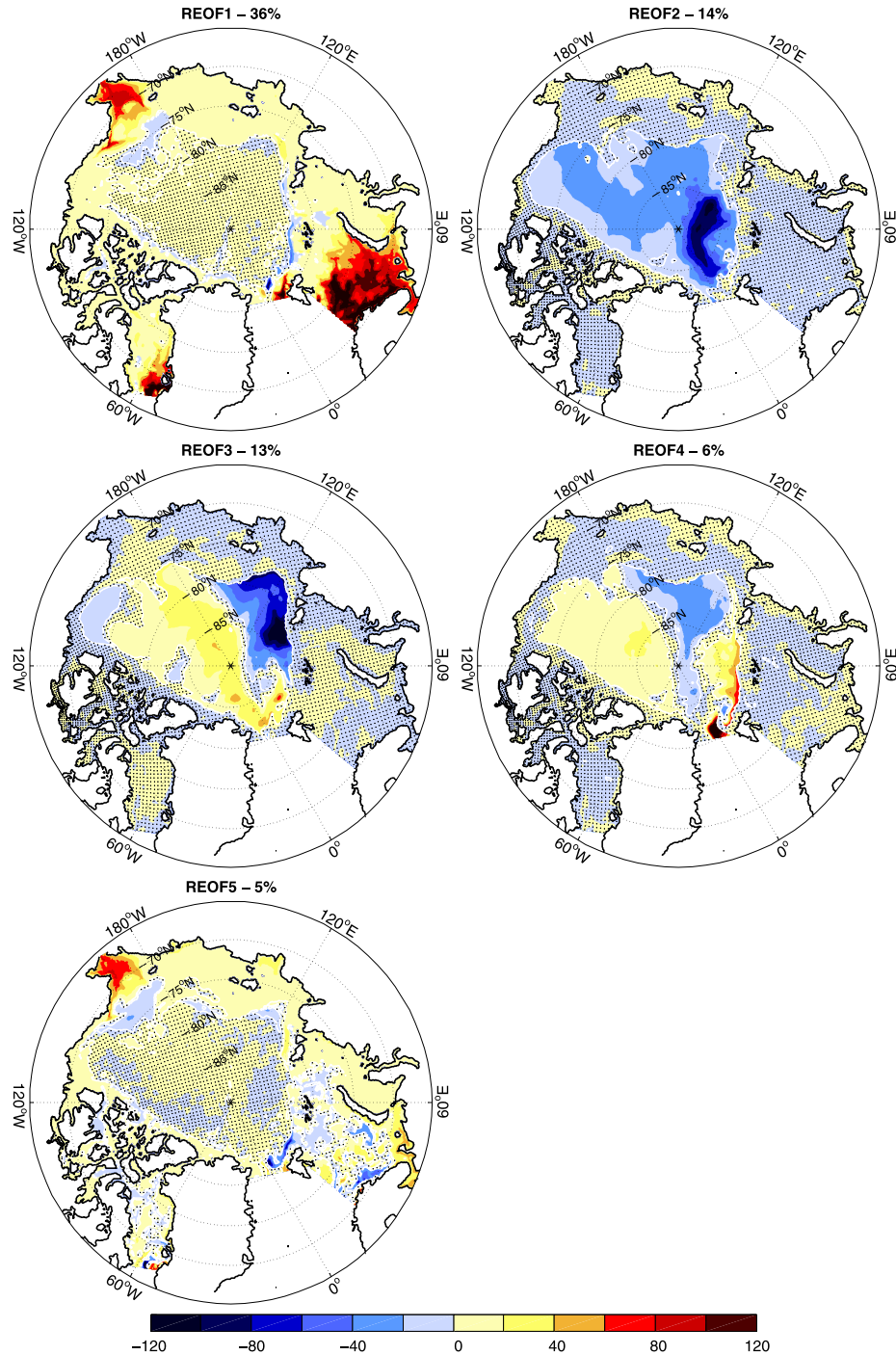


Figure 14. Same as Figure 4, but the REOF analysis is computed on the fields that mimic an objective analysis based on ITPs and the SST observations. Only the first five modes are shown because they achieve the 95% significance level, according to a Monte Carlo test.

(<http://www.whoi.edu/website/itp/overview>) aims at becoming the equivalent of the Argo program in the Arctic region. From the current global Argo array, von Schuckmann *et al.* [2009] have built monthly gridded global temperature and salinity fields. One can imagine that a sufficient deployment of ITPs could allow a similar analysis for the Arctic region; however, the ITP observations would have some

limitations. Since 2004, ITPs perform CTD profiles from 7 m to 760 m (or shallower in some cases), meaning that they do not penetrate the regions in the Arctic shallower than 800 m. Moreover, ITPs are usually deployed on multi-year ice floes. Although ITPs are now designed to be able to float in the open water regions, this kind of instrument may have trouble exiting and re-entering the sea ice pack, due to

overrafted sea ice and other stresses. Thus here, we consider that ITPs only return observations under the sea ice pack, where the sea ice concentration is higher than 50%.

[55] Here we aim to mimic the fields that could be obtained from an objective analysis using observations from ITPs and satellite SST. By comparing the results of the REOF analysis based on this objective analysis-like fields with the results presented in section 4, we will be able to provide insight on the capability of these observations to capture the heat content variations. The fields are built with the following method: The three-dimensional model temperature fields are kept where ITPs and satellite SST provide observations. Elsewhere, the three-dimensional model temperature fields are replaced by mean monthly fields built over the period 1968–2007. To mimic the ITP observations, we keep the three-dimensional varying temperature fields under sea ice (where concentration is higher than 50%), from 7 m to 760 m and where the bathymetry is deeper than 800 m (and we also exclude Baffin Bay because the CAA is too shallow to allow a connection). To mimic the SST observations, we keep the first layer model varying temperature everywhere in the domain. Under sea ice, SST equals the freezing point temperature. Then the time-varying vertically integrated heat content is computed, and a REOF analysis is performed.

[56] Results are shown in Figure 14. This time, only the first five modes reach the 95% significance level based on a Monte Carlo test. Overall, the modes of variability computed here are similar to the ones reported in section 4 [same pattern, same PC (not shown), and same variance explained]. The exception is the fourth mode that does not appear here. This is expected because this mode is related to the variations of the dense water formed in the Barents Sea. Because ITP observations are restricted to the upper 800 m, the variability of the dense waters is not captured here. The first mode here explains a larger part of the variance (36% versus 28% for the fully varying fields), and the low-frequency variability of PC1 is somewhat attenuated. In the Chukchi and Barents seas, the only interannual variations in the objective analysis-like fields come from the SST fields, because these regions are shallower than 800 m. In the Barents Sea, it is expected that the SST variations are a good proxy for the heat content (from 0 to 300 m) variations [Chepurin and Carton, 2012]. The attenuation of the low-frequency variations of PC1 here suggests that the true REOF1 mode is also driven by subsurface variability of temperature in the Chukchi Sea, and that the surface and subsurface temperatures may vary in a coherent way. Thus the heat content anomalies only due to SST variations are smaller than the heat content anomalies due to the full depth temperature variations.

[57] The results of the REOF analysis performed here suggest that ITPs-like observations (only under sea ice, in the regions deeper than 800 m and in the top 800 m) supplemented by SST data in the open water regions may be sufficient to capture most of the spatial and temporal variability of the heat content. However, ITPs have been almost exclusively deployed in the Beaufort Gyre and in the Canadian Basin [Toole *et al.*, 2011]. Due to the Arctic circulation, they do not often enter the Eurasian Basin. We thus do the same exercise, considering only ITP observations in the Canadian side of the Arctic (i.e. for longitude between

0°W and 180°W) and applying climatological fields in the Eurasian Basin and everywhere the sea ice concentration is under 50% as well as in the shallow regions. This time, only the first and sixth modes of the initial REOF analysis are captured; they explain 64% and 10% of the variance, respectively. Note that these two modes are also captured by a REOF analysis of the SST fields.

[58] These results underline the crucial need for measurements in the Eurasian Basin. Contrary to the freshwater content, for which most of the variability occurs in the Beaufort Gyre and in the Canadian Basin [Proshutinsky *et al.*, 2009], most of the heat content variability occurs on the Eurasian side of the basin, and autonomous observing systems need to be deployed in this region to better quantify and understand the variability of the heat content.

6. Summary and Conclusion

[59] The spatiotemporal variability of the Arctic Ocean heat content and budget is a key component of the present climate, though it remains poorly understood. In particular, recent studies underline the possible significant contribution of the ocean temperature variations to the recent decline of the sea ice extent and thickness [Woodgate *et al.*, 2010; Steele *et al.*, 2010; Polyakov *et al.*, 2010].

[60] The DRAKKAR high-resolution global ocean/sea ice model is used to investigate the heat budget of the Arctic Ocean and analyze the variations of the heat content within the basin. A correlation analysis shows that the seasonal cycle of the Arctic total heat content is driven by the seasonal cycle of the surface heat flux, while the heat content interannual variability is driven mostly by the advective heat flux to the Arctic Ocean, and in particular by the variability of the heat transport associated with the AW inflow through Fram Strait and the Barents Sea.

[61] A REOF analysis is then performed on the modeled monthly mean vertically integrated heat content to investigate the mechanisms governing its spatiotemporal variations. The linear trend has been removed prior to the analysis. A total of 28% of the heat content variability is driven by the seasonal and interannual fluctuations of the atmospheric heat flux in the seasonally ice-free regions, i.e., in the Chukchi and Barents seas and Baffin Bay. The variability of the heat transport associated with the advection of AW through Fram Strait drives a large part of the heat content variability (31% in total) and projects on three distinct modes. The second (13% of the total variance) and third (12%) modes of variability are driven by the variations of the volume transport through Fram Strait, due to changes of both the intensity of the recirculation through the strait and of the velocity of the northward penetrating branch. The effect of the variations of the Fram Strait inflow temperature remains localized north of Fram Strait on the continental slope. These three modes of variability explain most of the heat content variability on the continental side of the Eurasian Basin. The remaining part of the variability in the Eurasian Basin is likely driven by the export through St. Anna Trough of dense water masses formed in the Barents Sea. The last significant mode of variability explains only 4% of the variance; it is driven by the variations of the Bering Strait inflow temperature. Although the spatial pattern associated to this mode of variability remains

very localized in the Chukchi Sea, the effect of the heat content variations in this region may have consequences for the sea ice pack, because the Pacific Water in the surface layer is in direct contact with the sea ice pack or the atmosphere.

[62] While REOF analysis is a powerful tool to better understand the variability of the Arctic heat content, there are two important caveats. First, it is based on an assumption of a stationary time series. All modes of variability exhibit a large interannual-to-interdecadal variability, and the 40 year period on which we have performed the analysis might be too short to completely capture some of these modes. The presence of such low-frequency variations, and in particular for the modes driven by the AW inflow variability, underlines the difficulty to distinguish long-term trends from natural variability. Caution should be taken when drawing conclusions about the warming of the Arctic Ocean [e.g., Polyakov *et al.*, 2005; Polyakov *et al.*, 2010]; warming over a given period might be one phase of the natural variability instead of the fingerprint of global warming.

[63] Second, the method is based on model simulations that may have flaws. Attempts are made to evaluate the model realism regarding the heat content and budget. All the terms of the heat budget can be estimated from direct observations (albeit the large uncertainties on the estimate) [e.g., Tsubouchi *et al.*, 2012], but direct observations of the heat content (i.e., temperature profiles) remain sparse in time and space in the Arctic Basin and are likely insufficient to quantify the spatiotemporal variations of the heat content. Attempts are also made to examine the capability of the current autonomous observing system to capture the variability of the Arctic heat content. Satellite observations of SST in the seasonally ice-free regions coupled with temperature profiles of the top 800 m in the deep Arctic covered by sea ice are sufficient to capture most of the heat content variability. However, at the present time, autonomous profilers such as ITPs are very rarely deployed on the Eurasian side of the Arctic Basin, although temperature profiles in this area are crucial to capture the heat content variability. Moorings such as those deployed through the NABOS project may be good alternative but would need to be sustained over a long time period.

[64] This study of the variability of the vertically integrated heat content is only the first step toward a full understanding of the Arctic heat content variability, which also occurs on the vertical dimension. The next step of this analysis would be to investigate the vertical structure of the heat content in the basin and the variability of the vertical heat flux between the different layers. For example, it remains unclear whether warming of the AW layer leads to an increase of the vertical heat flux to the halocline and thus an increase in sea ice melt, as suggested by Polyakov *et al.* [2010]. In opposition to that, Bourgain and Gascard [2011] present evidence that the cold halocline may have remained stable from 1997 to 2008 despite a significant warming of the AW layer.

[65] **Acknowledgments.** This study uses numerical experiments carried out within the European DRAKKAR project. The simulation has been run at the IDRIS CNRS-GENCI computer centre in Orsay, France, by J. M. Molines. We sincerely thank Cecilia Peralta-Ferriz and Jennifer Jackson for helpful comments and suggestions on an earlier version of the manuscript. Greg Holloway, three anonymous reviewers, and Andrey Proshutinsky provided very valuable comments that helped to improve the results and the

manuscript. C. Lique is supported by JISAO and the Program on Climate Change of the University of Washington. M. Steele is supported by NSF's office of polar programs, NASA's cryosphere and biogeosciences programs, and ONR's Arctic and Global Prediction Program.

References

- Aagaard, K., and P. Greisman (1975), Toward new mass and heat budgets for the Arctic Ocean, *J. Geophys. Res.*, **80**(27), 3,821–3,827.
- Aksenov, Y., S. Bacon, A. C. Coward, and A. J. G. Nurser (2010), The North Atlantic inflow to the Arctic Ocean: High-resolution model study, *J. Mar. Syst.*, **79**, 1–22, doi:10.1016/j.jmarsys.2009.05.003.
- Aksenov, Y., V. V. Ivanov, A. J. G. Nurser, S. Bacon, I. V. Polyakov, A. C. Coward, A. C. Naveira-Garabato, and A. Beszczynska-Moeller (2011), The Arctic Circumpolar Boundary Current, *J. Geophys. Res.*, **116**, C09,017, doi:10.1029/2010JC006637.
- Barnier, B., et al. (2006), Impact of partial steps and momentum advection schemes in a global ocean circulation model at eddy permitting resolution, *Ocean Dynam.*, **56**, 543–567, doi:10.1007/s10236-006-0082-1.
- Beszczynska-Möller, A., R. Woodgate, C. Lee, H. Melling, and M. Karcher (2011), A synthesis of exchanges through the main oceanic gateways to the Arctic Ocean, *Oceanography*, **24** (3), 82–99, doi:10.5670/oceanog.2011.59.
- Blanke, B., and P. Delecluse (1993), Variability of the Tropical Atlantic Ocean Simulated by a General Circulation Model with Two Different Mixed-Layer Physics, *J. Phys. Oceanogr.*, **23**, 1363–1388.
- Bourgain, P., and J. C. Gascard (2011), The Arctic Ocean halocline and its interannual variability from 1997 to 2008, *Deep-Sea Res. Pt. I*, **58**, 745–756, doi:10.1016/j.dsr.2011.05.001.
- Bourgain, P., and J. C. Gascard (2012), The Atlantic and summer Pacific waters variability in the Arctic Ocean from 1997 to 2008, *Geophys. Res. Lett.*, **39**, L05603, doi:10.1029/2012GL051045.
- Bourke, R. H., A. M. Weigel, and R. G. Paquette (1988), The westward turning branch of the West Spitsbergen Current, *J. Geophys. Res.*, **93**, 14,065–14,077, doi:10.1029/JC093iC11p14065.
- Brodeau, L., B. Barnier, T. Penduff, A. M. Treguier, and S. Gulev (2010), An ERA40-based atmospheric forcing for global ocean circulation models, *Ocean Model.*, **31**, 88–104, doi:10.1016/j.ocemod.2009.10.005.
- Carton, J. A., G. A. Chepurin, J. Reagan, and S. Häkkinen (2011), Interannual to decadal variability of Atlantic Water in the Nordic and adjacent seas, *J. Geophys. Res.*, **116**(C15), C11035, doi:10.1029/2011JC007102.
- Chepurin, G. A., and J. A. Carton (2012), Subarctic and Arctic sea surface temperature and its relation to ocean heat content 1982–2010, *J. Geophys. Res.*, **117**, C06019, doi:10.1029/2011JC007770.
- Cuny, J., P. B. Rhines, and R. Kwok (2005), Davis Strait volume, freshwater and heat fluxes, *Deep-Sea Res. Pt. I*, **52**, 519–542, doi:10.1016/j.dsr.2004.10.006.
- Curry, B., C. M. Lee, and B. Petrie (2011), Volume, Freshwater, and Heat Fluxes through Davis Strait, 2004–05, *J. Phys. Oceanogr.*, **41**, 429–436, doi:10.1175/2010JPO4536.1.
- Dai, A., and K. Trenberth (2002), Estimates of freshwater discharge from continents: latitudinal and seasonal variations, *J. Hydrometeorol.*, **3**, 660–687, doi:10.1175/1525-7541(2002)003<0660:EOFDFC>2.0.CO;2.
- Danielson, S., E. Curchitser, K. Hedstrom, T. Weingartner, and P. Staben (2011), On ocean and sea ice modes of variability in the Bering Sea, *J. Geophys. Res.*, **116**(C15), C12034, doi:10.1029/2011JC007389.
- Dmitrenko, I. A., I. V. Polyakov, S. A. Kirillov, L. A. Timokhov, I. E. Frolov, V. T. Sokolov, H. L. Simmons, V. V. Ivanov, and D. Walsh (2008), Toward a warmer Arctic Ocean: Spreading of the early 21st century Atlantic Water warm anomaly along the Eurasian Basin margins, *J. Geophys. Res.*, **113**(C12), C05,023, doi:10.1029/2007JC004158.
- Fieg, K., R. Gerdes, E. Fahrbach, A. Beszczynska-Möller, and U. Schauer (2010), Simulation of oceanic volume transports through Fram Strait 1995–2005, *Ocean Dynam.*, **60**, 491–502, doi:10.1007/s10236-010-0263-9.
- Gershunov, A., and T. P. Barnett (1998), Interdecadal Modulation of ENSO Teleconnections, *B. Am. Meteorol. Soc.*, **79**, 2715–2726, doi:10.1175/1520-0477(1998)079<2715:IMOET>2.0.CO;2.
- Häkkinen, S., and C. A. Geiger (2000), Simulated low-frequency modes of circulation in the Arctic Ocean, *J. Geophys. Res.*, **105**, 6549–6564, doi:10.1029/2000JC900003.
- IPCC (2007), *Climate change 2007 - The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, p. 996, Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
- Jackson, J. M., W. J. Williams, and E. C. Carmack (2012), Winter sea-ice melt in the Canada Basin, Arctic Ocean, *Geophys. Res. Lett.*, **39**, 3603, doi:10.1029/2011GL050219.

- Jahn, A., B. Tremblay, L. A. Mysak, and R. Newton (2010), Effect of the large-scale atmospheric circulation on the variability of the Arctic Ocean freshwater export, *Clim. Dynam.*, **34**, 201–222, doi:10.1007/s00382-009-0558-z.
- Karcher, M., F. Kauker, R. Gerdes, E. Hunke, and J. Zhang (2007), On the dynamics of Atlantic Water circulation in the Arctic Ocean, *J. Geophys. Res.*, **112**(C11), 4, doi:10.1029/2006JC003630.
- Karcher, M. J., R. Gerdes, F. Kauker, and C. Köberle (2003), Arctic warming: Evolution and spreading of the 1990s warm event in the Nordic seas and the Arctic Ocean, *J. Geophys. Res.*, **108**, 3034, doi:10.1029/2001JC001265.
- Karcher, M. J., J. N. Smith, F. Kauker, R. Gerdes, and W. M. S. Jr (2012), Recent changes of Arctic Ocean circulation revealed by 129 Iodine observations and modelling, *J. Geophys. Res.*, **117**(C16), 8007, doi:10.1029/2011JC007513.
- Kawamura, R. (1993), A rotated eof analysis of global sea surface temperature variability with interannual and interdecadal scales, *J. Phys. Oceanogr.*, **24**, 707–715.
- Large, W., and S. Yeager (2004), Diurnal to decadal global forcing for ocean and sea-ice models: the datasets and flux climatologies, *NCAR technical note NCAR/TN-460+STR*, CGD division of the National Center for Atmospheric Research, available on the GFDL CORE web site.
- Lengaigne, M., G. Madec, L. Bopp, C. Menkes, O. Aumont, and P. Cadule (2009), Bio-physical feedbacks in the Arctic Ocean using an Earth system model, *Geophys. Res. Lett.*, **36**, L21,602, doi:10.1029/2009GL040145.
- Lique, C., and M. Steele (2012), Where can we find a seasonal cycle of the Atlantic water temperature within the Arctic Basin? *J. Geophys. Res.*, **117**(C16), C03026, doi:10.1029/2011JC007612.
- Lique, C., A. M. Treguier, M. Scheinert, and T. Penduff (2009), A model-based study of ice and freshwater transport variability along both sides of Greenland, *Clim. Dynam.*, **33**, 685–705, doi:10.1007/s0038200805107.
- Lique, C., A. M. Treguier, B. Blanke, and N. Grima (2010), On the origins of water masses exported along both sides of Greenland: A Lagrangian Model Analysis, *J. Geophys. Res.*, **115**(C05019), 5019, doi:10.1029/2009JC005316.
- Lique, C., G. Garric, A.-M. Treguier, B. Barnier, N. Ferry, C.-E. Testut, and F. Girard-Arduin (2011), Evolution of the Arctic Ocean Salinity, 2007–08: Contrast between the Canadian and the Eurasian Basins, *J. Climate*, **24**, 1705–1717, doi:10.1175/2010JCLI3762.1.
- Madec, G., (2008), NEMO Ocean engine, *Note du pôle modélisation 27*, Institut Pierre-Simon Laplace (IPSL).
- Manabe, S., and R. J. Stouffer (1980), Sensitivity of a global climate model to an increase of CO₂ concentration in the atmosphere, *J. Geophys. Res.*, **85**, 5529–5554, doi:10.1029/JC085iC10p05529.
- Manley, T. O. (1995), Branching of Atlantic Water within the Greenland-Spitsbergen Passage: An estimate of recirculation, *J. Geophys. Res.*, **100**, 20,627–20,634, doi:10.1029/95JC01251.
- Mo, K. C. (2000), Relationships between Low-Frequency Variability in the Southern Hemisphere and Sea Surface Temperature Anomalies, *J. Climate*, **13**, 3599–3610, doi:10.1175/1520-0442(2000)013<3599:RBLFVI>2.0.CO;2.
- Morison, J., et al. (2002), North Pole Environmental Observatory delivers early results, *EOS Trans. AGU*, **83**, 357–361.
- O'Lenic, E. A., and R. E. Livezey (1988), Practical Considerations in the Use of Rotated Principal Component Analysis (RPCA) in Diagnostic Studies of Upper-Air Height Fields, *Mon. Weather Rev.*, **116**, 1682, doi:10.1175/1520-0493(1988)116<1682:PCITUO>2.0.CO;2.
- Penduff, T., J. Le Sommer, B. Barnier, A.-M. Treguier, J.-M. Molines, and G. Madec (2007), Influence of numerical schemes on current-topography interactions in 1/4 global Ocean simulations, *Ocean Sci.*, **3**, 509–524, doi:10.5194/os-3-509-2007.
- Peralta-Ferriz, C., J. H. Morison, J. M. Wallace, and J. Zhang (2011), A basin-coherent mode of sub-monthly variability in Arctic Ocean bottom pressure, *Geophys. Res. Lett.*, **38**, L14606, doi:10.1029/2011GL048142.
- Polyakov, I., D. Walsh, I. Dmitrenko, R. Colony, J. Hutchings, L. Timokhov, M. Johnson, and E. Carmack (2003), A long-term circulation and water mass monitoring program for the Arctic Ocean, *EOS Trans. AGU*, **84**, 281–285, doi:10.1029/2003EO300001.
- Polyakov, I., et al. (2007), Observational Program Tracks Arctic Ocean Transition to a Warmer State, *Eos Trans. AGU*, **88**, 398–399, doi:10.1029/2007EO400002.
- Polyakov, I. V., and M. A. Johnson (2000), Arctic decadal and interdecadal variability, *Geophys. Res. Lett.*, **27**, 4097–4100, doi:10.1029/2000GL011909.
- Polyakov, I. V., G. V. Alekseev, L. A. Timokhov, U. S. Bhatt, R. L. Colony, H. L. Simmons, D. Walsh, J. E. Walsh, and V. F. Zakharov (2004), Variability of the Intermediate Atlantic Water of the Arctic Ocean over the Last 100 Years, *J. Climate*, **17**, 4485–4497, doi:10.1175/JCLI-3224.1.
- Polyakov, I. V., et al. (2005), One more step toward a warmer Arctic, *Geophys. Res. Lett.*, **32**, 17605, doi:10.1029/2005GL023740.
- Polyakov, I. V., et al. (2010), Arctic Ocean Warming Contributes to Reduced Polar Ice Cap, *J. Phys. Oceanogr.*, **40**, 2743–2756, doi:10.1175/2010JP04339.1.
- Proshutinsky, A., R. Krishfield, M.-L. Timmermans, J. Toole, E. Carmack, F. McLaughlin, W. J. Williams, S. Zimmermann, M. Itoh, and K. Shimada (2009), Beaufort Gyre freshwater reservoir: State and variability from observations, *J. Geophys. Res.*, **114**(C00A10), doi:10.1029/2008JC005104.
- Proshutinsky, A. Y., and M. A. Johnson (1997), Two circulation regimes of the wind-driven Arctic Ocean, *J. Geophys. Res.*, **102**, 12,493–12,514.
- Quadfasel, D., J.-C. Gascard, and K.-P. Koltermann (1987), Large-scale oceanography in Fram Strait during the 1984 marginal ice zone experiment, *J. Geophys. Res.*, **92**, 6719–6728, doi:10.1029/JC092iC07p06719.
- Quadfasel, D., A. Sy, D. Wells, and A. Tunik (1991), Warming in the Arctic, *Nature*, **350**, 385, doi:10.1038/350385a0.
- Richman, M. B. (1986), Rotation of principal components, *International J. Climatol.*, **6**, 293–335, doi:10.1002/joc.3370060305.
- Rigor, I. G., R. L. Colony, and S. Martin (2000), Variations in Surface Air Temperature Observations in the Arctic, 1979–97, *J. Climate*, **13**, 896–914.
- Rudels, B., H. J. Friedrich, and D. Quadfasel (1999), The Arctic Circumpolar Boundary Current, *Deep Sea Res. Pt. II: Topical Studies in Oceanography*, **46**, 1023–1062, doi:10.1016/S0967-0645(99)00015-6.
- Schauer, U., and A. Beszczynska-Möller (2009), Problems with estimation and interpretation of oceanic heat transport - conceptual remarks for the case of Fram Strait in the Arctic Ocean, *Ocean Sci.*, **5**, 487–494, doi:10.5194/os-5-487-2009.
- Schauer, U., H. Loeng, B. Rudels, V. K. Ozhigin, and W. Dieck (2002), Atlantic Water flow through the Barents and Kara Seas, *Deep Sea Res. Pt. I*, **49**, 2281–2298, doi:10.1016/S0967-0637(02)00125-5.
- Schauer, U., E. Fahrbach, S. Osterhus, and G. Rohardt (2004), Arctic warming through the Fram Strait: Oceanic heat transport from 3 years of measurements, *J. Geophys. Res.*, **109**, 6026, doi:10.1029/2003JC001823.
- Schauer, U., A. Beszczynska-Möller, W. Walczowski, E. Fahrbach, J. Piechura, and E. Hansen (2008), Variation of measured heat flow through the Fram Strait between 1997 and 2006, in *Arctic-Subarctic Ocean Fluxes*, edited by Dickson, R. R., J. Meincke, and P. Rhines, pp. 385–404, Springer, Dordrecht.
- Sciremammano, F. (1979), A suggestion for the presentation of correlations and their significance levels, *J. Phys. Oceanogr.*, **9**, 1273–1276.
- Serreze, M. C., A. P. Barrett, A. G. Slater, M. Steele, J. Zhang, and K. E. Trenberth (2007), The large-scale energy budget of the Arctic, *J. Geophys. Res.*, **112**(D11), D11122, doi:10.1029/2006JD008230.
- Shimada, K., T. Kamoshida, M. Itoh, S. Nishino, E. Carmack, F. McLaughlin, S. Zimmermann, and A. Proshutinsky (2006), Pacific Ocean inflow: Influence on catastrophic reduction of sea ice cover in the Arctic Ocean, *Geophys. Res. Lett.*, **33**, L08605, doi:10.1029/2005GL025624.
- Skagseth, Ø. (2008), Recirculation of Atlantic Water in the western Barents Sea, *Geophys. Res. Lett.*, **35**, L11606, doi:10.1029/2008GL033785.
- Smedsrud, L. H., R. Ingvaldsen, J. E. Ø. Nilsen, and Ø. Skagseth (2010), Heat in the Barents Sea: transport, storage, and surface fluxes, *Ocean Sci.*, **6**, 219–234, doi:10.5194/os-6-219-2010.
- Steele, M., R. Morley, and W. Ermold (2001), PHC: a global ocean hydrography with a high quality Arctic Ocean, *J. Climate*, **14**, 2079–2087, doi:10.1175/1520-0442(2001)014<2079:PAGOHW>2.0.CO;2.
- Steele, M., W. Ermold, and J. Zhang (2008), Arctic Ocean surface warming trends over the past 100 years, *Geophys. Res. Lett.*, **35**(L02614), doi:10.1029/2007GL031651.
- Steele, M., J. Zhang, and W. Ermold (2010), Mechanisms of summertime upper Arctic Ocean warming and the effect on sea ice melt, *J. Geophys. Res.*, **115**(C14), C11004, doi:10.1029/2009JC005849.
- Steele, M., W. Ermold, and J. Zhang (2011), Modeling the formation and fate of the near-surface temperature maximum in the Canadian Basin of the Arctic Ocean, *J. Geophys. Res.*, **116**(C15), C11015, doi:10.1029/2010JC006803.
- Stepanov, V. N., H. Zuo, and K. Haines (2012), The link between the Barents Sea and ENSO events simulated by NEMO model, *Ocean Sci.*, **8**, 971–982, ISSN 1812-0792, doi:10.5194/os-8-971-2012.
- The Drakkar group (2007), Eddy-permitting Ocean circulation hindcasts of past decades, *CLIVAR Exchanges*, **42**, 8–10.
- Timmermann, R., H. Goosse, G. Madec, T. Fichefet, C. Etche, and V. Duliere (2005), On the representation of high latitude processes in the ORCA-LIM global coupled sea ice-ocean model, *Ocean Model.*, **8**, 175–201, doi:10.1016/j.ocemod.2003.12.009.

- Toole, J., R. Krishfield, M.-L. Timmermans, and A. Proshutinsky (2011), The Ice-Tethered Profiler: Argo of the Arctic, *Oceanography*, 24(3), 126–135, doi: 10.5670/oceanog.2011.64.
- Tsubouchi, T., S. Bacon, A. C. Naveira Garabato, Y. Aksenov, S. W. Laxon, E. Fahrbach, A. Beszczynska-Möller, E. Hansen, C. M. Lee, and R. B. Ingvaldsen (2012), The Arctic Ocean in summer: A quasi-synoptic inverse estimate of boundary fluxes and water mass transformation, *J. Geophys. Res.*, 117 (C16), C01024, doi:10.1029/2011JC007174.
- von Schuckmann, K., F. Gaillard, and P.-Y. Le Traon (2009), Global hydrographic variability patterns during 2003-2008, *J. Geophys. Res.*, 114(C13), C09007, doi:10.1029/2008JC005237.
- von Storch, H., and F. Zwiers (1999), *Statistical Analysis in Climate Research*, 484 pp., Cambridge University Press, Cambridge.
- Woodgate, R. A., K. Aagaard, and T. J. Weingartner (2006), Interannual changes in the Bering Strait fluxes of volume, heat and fresh-water between 1991 and 2004, *Geophys. Res. Lett.*, 33, L15609, doi: 10.1029/2006GL026931.
- Woodgate, R. A., T. Weingartner, and R. Lindsay (2010), The 2007 Bering Strait oceanic heat flux and anomalous Arctic sea-ice retreat, *Geophys. Res. Lett.*, 37, L01602, doi:10.1029/2009GL041621.
- Yang, J. (2005), The Arctic and Subarctic Ocean Flux of Potential Vorticity and the Arctic Ocean Circulation*, *J. Phys. Oceanogr.*, 35, 2387, doi: 10.1175/JPO2819.1.