Steric height variability in the Nordic Seas

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1 The variability of steric height in the Nordic Seas is analyzed on seasonal, interannual, and decadal timescales using a comprehensive data set of temperature and salinity observations for the second half of the twentieth century. Results from a regional Ocean General Circulation Model (OGCM) are used to assess the reliability of the averaging and the temporal interpolation of the inhomogeneous distributed observation data. The annual cycle explains only a minor part of the monthly variability for most of the region. The analysis on interannual to decadal timescales confined to the Norwegian Sea displays a clear rising trend starting at the end of the 1960s with particularly strong changes along the Barents Sea opening (6 to 7 cm). Moreover, a general freshening is found for the entire Norwegian Sea. In addition, a north-south dipole of the thermal component of the steric height variability is identified. This dipole elevates the general rising trend along the Barents Sea opening and reduces it in the southern Norwegian Sea. The bulk of the interannual variability in steric height is governed by variations in the local meridional wind stress that determine the relative distribution of Atlantic and Arctic waters in the Norwegian Sea. In addition, reduced heat loss to the atmosphere strongly correlates with the North Atlantic Oscillation winter index. This, in turn, may in particular explain the large steric height increase found at the Barents Sea opening.


1. Introduction

2 The Nordic Seas [Hurdle, 1986; Drange et al., 2005] is the common name for the Greenland, Iceland and Norwegian Seas (Figure 1). The region is bounded by the Arctic Ocean to the north, the deep North Atlantic Ocean to the south, and the shallow North Sea to the southeast. The inflow of warm and saline water from the Northeast Atlantic Ocean to the Nordic Seas occurs primarily along two northward extending branches that respectively pass through the Faroe-Shetland Channel and across the Iceland-Faroe Ridge [Hansen and Østerhus, 2000; Østerhus et al., 2005]. Occasionally, northeast Atlantic water are also reported to enter the Denmark Strait west of Iceland [Hansen and Østerhus, 2000; Østerhus et al., 2005]. Combined with the spreading of water of Arctic origin that enters the Nordic Seas through the Fram Strait distinct, horizontal and vertical density gradients are formed. Monitoring of this density field is highly relevant in the context of climate studies [Dickson et al., 2002; Orvik and Skagseth, 2005].

3 While variations in the sea surface height related to the surface geostrophic currents can be derived directly from satellite altimeter data, an assessment of the absolute value of the Mean Dynamic Topography (MDT) (and hence the total surface current) requires that the elevation of a hypothetical ocean at rest, i.e., the geoid, is subtracted from the altimetric mean sea surface height. The MDT usually differs from the geoid with magnitudes in the range from 0.1 to 1 m over distances around 100 km. On that spatial scale, however, existing geoid models, based on a combination of terrestrial, ship- and space-borne gravimetry, provide an accuracy not better than several tens of centimeters [Johannessen et al., 2003]. The precise calculation of the MDT using mean sea surface height and geoid information has therefore hitherto not been possible at this wavelength. In turn, uncertainties in transport estimation have penalized [Wunsch and Stammer, 1998].

4 This will shortly change with the launch of the European Space Agency (ESA) Gravity field and steady state Ocean Circulation Explorer (GOCE) satellite mission (planned for March–April 2008). The accurate (~1 cm) and high-resolution (~100 km) marine geoid aimed to be observed with GOCE will in combination with precise satellite altimetry enable new estimates to be made of the MDT. In combination with in situ data and ocean models this will, in turn, provide a much needed high-resolution...
window” on the ocean circulation at depth as elaborated by Johannessen et al. [2003] and Knudsen et al. [2006].

[5] The mean steric height is an essential part of the MDT. It is defined as the vertical distance between two surfaces of prescribed pressure and is thus determined by the density profile at a given location [see, e.g., Tomczak and Godfrey, 2003]. Following the geostrophic balance assumption the MDT can be expressed by the mean steric height, computed from the surface to a depth where the horizontal density gradients vanish, plus the additional barotropic contribution connected with the bottom pressure.

[6] The first attempt to estimate the different contributors to the annual sea level variability was made by Gill and Niiler [1973]. For the North Atlantic and North Pacific at large spatial scales, of order 1000 km, one of their main findings was that the local heat flux is the main contributor to the seasonal sea level variability and that advection is not particularly important. This is basically in accordance with recent works that also include altimeter data [Ferry et al., 2000; Stammer, 1997; Volkov and van Aken, 2003]. Mork and Skagseth [2005] found that the seasonal steric height variability in the Nordic Seas is mainly caused by heat flux changes and it contributes about 40% to the annual sea level variability. This contribution for the Nordic Seas is less than what is found for the North Atlantic. According to Gill and Niiler [1973], annual sea level variability between 15°N and 55°N is dominated by steric effects.

[7] Regional investigations of steric height variability beyond the seasonal cycle on interannual to decadal timescales are rather limited. Proshutinsky et al. [2004] used data from tide gauge stations and hydrodynamic models to estimate the sea level trend in the Russian sector of the Arctic Ocean for the period 1954–1989. They found a rising rate of $1.85 \times 10^{-3}$ m a$^{-1}$ (corrected for postglacial isostatic adjustment) and estimated a steric contribution of approximately $0.64 \times 10^{-3}$ m a$^{-1}$. From altimetric observations and concurrent oceanic surface geophysical fields Efthymiadis et al. [2002] found a rise of $\sim 0.5$ cm a$^{-1}$ in sea level of the subtropical northeast Atlantic Ocean as well as a temporal rise of $\sim 3$ cm during 1995. They state that most of the variations can be explained as the result of a steric (i.e., diabatic) response of the Atlantic Ocean.

[8] This paper investigates the spatial and temporal variability of the steric height of the Nordic Seas. The analyses are based on a unique and comprehensive data set of observed temperature and salinity profiles in the region, covering the second half of the twentieth century. The mean steric height and its seasonal, interannual and decadal variability are derived, together with an analysis of the underlying contributions from temperature and salinity. To the best of our knowledge this is the first analysis of this kind for the Nordic Seas.

[9] The paper is organized as follows. In section 2, a description of the observations is given together with a brief presentation of the main characteristics of the Ocean General Circulation Model (OGCM) used. Section 3 presents the mean steric height for the period 1950–1999 in addition to its seasonal cycle and the decadal variability. In section 4 the forcing mechanisms of interannual steric height variability in the Nordic Seas are examined by presenting a Principal Component Analysis (PCA). The key findings are
highlighted and discussed in section 5 followed by summary and conclusions in section 6.

2. Methods

2.1. Observational Database

[10] The observational data used to examine and quantify the variability of the steric height originates from several sources. The collection of the data was done in the Nordic NISE (Norwegian Iceland Seas Experiment) project where the aim was to describe the ocean climate variability and its link to fishery resources in the Nordic Seas. The NISE data set is based on data from the ICES database (http://www.ices.dk/) together with data from the Arctic and Antarctic Research Institute (AARI) in Russia, Institute of Marine Research (IMR) in Norway, Marine Research Institute (MRI) in Iceland and the Faroese Fisheries Laboratory (FRS). The hydrographic database consists of measurements of pressure, temperature, salinity and oxygen. Data earlier than the mid 1970s as well as the AARI data have a coarse vertical resolution since they are from water samples. Nearly all the remaining profiles are high-resolution CTD casts. The AARI data are interpolated at 30 standard depths (0, 10, 20, 30, 50, 75, 100, 125, 150, 200, 250, 300, 400, 500, 600, 700, 800, 900, 1000, 1100, 1200, 1300, 1400, 1500, 1750, 2000, 2200, 2500, 3000, and 3500 m).

[11] In total, the data set used here consists of about 300,000 profiles. Despite the large amount of observations, the data are inhomogeneously distributed in space and time.

[12] The majority of the data is collected during the period 1950–1995, with highest numbers of profiles occurring in the years 1984 and 1989 (Figure 2). The monthly distribution shows most observations in June, with the number of hydrographic stations exceeding 55,000 (19% of the total), and fewest observations in December, with only about 10,000 stations (3% of the total). Regionally, most of the profiles are collected in the southern Norwegian Sea whereas no observations are available along the ice-covered east coast of Greenland.

[13] As noticed, two periods stand out with most data in the 1980s and least data in the 1970s (Figure 2). In addition, during the latter period the data is more unevenly distributed in space (not shown). For the winter season (DJFM) in the 1970s only very few observations are available while for the 1980s the region is essentially fully covered, except for the ice-covered areas.

[14] Owing to the spatial sparseness of the database the observations are averaged into $1^\circ$-latitude $\times$ $2^\circ$-longitude boxes for the area $60^\circ$N–$80^\circ$N, $30^\circ$W–$20^\circ$E. Further averaging and interpolation procedures depend on the timescales considered and are described in subsection 2.4.

2.2. Description of the OGCM

[15] To complement the in situ data, output from a regional set-up of the Nansen Center version of the Miami Isopycnic Coordinate Ocean Model (MICOM) [Bleck et al., 1992] has been used to derive information of the temporal and spatial variability of the DT in the North Atlantic/Nordic Seas region. The model covers the Atlantic Ocean between $30^\circ$N and $78^\circ$N. The regional model is one way nested into a global version of the same model and hence also initialized with interpolated model data from the global model. The global fields are read once a week and interpolated in time to specify the relaxation boundary conditions for the regional model at each time step.

[16] The global model is configured with a local horizontal orthogonal grid system with one pole over North America and the other over central Europe [Bentsen et al., 1999]. By using a global horizontal grid with a focus area, no artificial boundary conditions are needed, in contrast to the case of a limited area model. Therefore decadal and longer integration runs are feasible without the classical problem of prescribing the evolution of the essentially unknown lateral boundary conditions. For the simulations described here, the horizontal grid resolution varies between 40 and 270 km, with 40 km resolution in the North Atlantic/Nordic Seas region. The regional model runs on the same grid configuration but with doubled resolution, for example, 20 km for the North Atlantic/Nordic Seas region.

[17] In the vertical, both model versions have 26 layers of which the uppermost mixed layer (ML) has temporal and spatial varying density, and the 25 layers below have constant density. The vertically homogeneous ML utilizes the Gaspar [1988] bulk parameterization for the dissipation of turbulent kinetic energy, and has temperature, salinity and layer thickness as the prognostic variables. In the isopycnic layers below the ML, temperature and layer thickness are the prognostic variables, whereas salinity is diagnostically determined by means of the simplified equation of state of Friedrich and Levitus [1972]. The bathymetry is computed as the arithmetic mean value based on the ETOPO-5 database (Data Announcement 88-MGG-02, Digital relief of the Surface of the Earth, 1988, NOAA, National Geophysical Data Center, Boulder, Colorado, available at http://www.ngdc.noaa.gov/mgg/global/etopo5.html).

[18] The thermodynamic module incorporates freezing and melting of sea ice and snow covered sea ice [Drange and Simonsen, 1996], and is based on the thermodynamics of Semtner [1976], Parkinson and Washington [1979], and Fichefet and Gaspar [1988]. The dynamic part of the sea ice module is based on the viscous-plastic rheology of Hibler.
[1979], where sea ice is considered as a two-dimensional continuum. The dynamic ice module has been further modified by Harder [1996] to include description of sea ice roughness and the age of sea ice, and utilizing the advection scheme of Smolarkiewicz [1984]. For the ocean, the continuity, momentum and tracer equations are discretized on an Arakawa C-grid [Chang, 1977]. The diffusive velocities (diffusivities divided by the size of the grid cell) for layer interface diffusion, momentum dissipation, and tracer dispersion are 0.015 m s⁻¹, 0.010 m s⁻¹ and 0.005 m s⁻¹, respectively, yielding actual diffusivities of about 10³ m² s⁻¹. A flux corrected transport scheme [Zalesak, 1979; Smolarkiewicz and Clark, 1986] is used to advect the model layer thickness and the tracer quantities. The diapycnal mixing coefficient $K_d$ (m² s⁻¹) is parameterized according to Gargett [1984] expression $K_d = 3 \times 10^{-1}/N$, where $N = (g \rho^{-1} \partial \rho/\partial z)^{1/2}$ (s⁻¹) is the Brunt-Väisälä frequency (here $g$ (m s⁻²) is the gravity acceleration, $\rho$ (kg m⁻³) is the density and $z$ (m) is the depth). The numerical implementation of the diapycnal mixing follows the scheme of McDougall and Dewar [1998].

[19] The global model was initialized by the January Levitus and Boyer [1994] and Levitus et al. [1994] climatological temperature and salinity fields, respectively, a 2 m thick sea ice cover based on the climatological sea ice extent, and an ocean at rest. The model was then integrated for O(10 years) by applying the monthly mean NCEP/NCAR atmospheric forcing fields, and thereafter forced with daily NCEP/NCAR reanalysis [Kalnay et al., 1996] fields for the period 1948 to 1999. From the NCEP/NCAR reanalysis, wind stress, short-wave, long-wave, latent and sensible heat fluxes, precipitation, runoff and sea level pressure fields are used. The momentum, heat and fresh water fluxes are modified when the modeled surface state differs from the NCEP/NCAR reanalysis surface state by applying the Fairall et al. [1996] bulk parameterization scheme [Bentsen and Orkanger, 2000].

[20] In the regional model as well as during the spin-up phase of the global model, the ML temperature and salinity were relaxed toward the monthly mean climatological values of respectively Levitus and Boyer [1994] and Levitus et al. [1994]. The relaxation is carried out by applying fluxes of heat and salt proportional to the SST and SSS differences between model and climatology, respectively. The e-folding relaxation timescale was set to 30 days for a 50 m thick model ML, and it was increased linearly with ML thicknesses exceeding 50 m. No relaxation was applied in waters where sea ice is present in March in the Arctic and in September in the Antarctic to avoid relaxation toward temperature or salinity outliers in the poorly sampled polar waters. In addition, the mismatches between model and climatology were limited to $|\Delta SST| < 1.5^\circ C$ and $|\Delta SST| < 0.5$ in the computation of the relaxation fluxes. This, for example, avoids extreme fluxes in the vicinity of the western boundary currents which are not realistically separated from the coast with the current model resolution. Continental run-off is included by adding fresh water into the appropriate coastal grid cells.

[21] For the global model version we used here to nest the regional model, annually repeated heat and fresh water relaxation fluxes with weekly temporal resolution were added to the ML [Masson and Delecluse, 2001; Ferry and Reverdin, 2004]. In this way, temperature and salinity anomalies are free to evolve and propagate, whereas the mean thermodynamic state is kept fairly unchanged. It was found that such a procedure is of special importance for the SSS field, indicating problems with the forcing (either the prescribed NCEP/NCAR precipitation or run-off fields, or the computed evaporation field), or inherent model deficiencies in the representation of for instance horizontal advection processes or vertical mixing of the surface waters. The weekly heat and fresh water relaxation fluxes were diagnosed from a continuation of the spin-up integration. For this, daily forcing fields for the period 1974–1978 were applied. The period chosen as it represents a fairly neutral state of the North Atlantic Oscillation (NAO). Therefore the simulation presented here is the last element in a sequence of integrations: An O(10 years) spin-up with monthly mean forcing fields was followed by a full NCEP/NCAR cycle with SST and SSS relaxation toward climatology, succeeded by the period 1974–1978 for computation of the annual cycle of relaxation fluxes. Those relaxation fluxes were then prescribed in the integration of the global model used as relaxation boundary condition for the nested regional model presented in this paper.

2.3. Steric Height Computation

[22] The steric height $h$ is given by

$$h(p_1, p_2) = \frac{1}{g} \int_{p_1}^{p_2} \frac{1}{p} dp,$$

and describes the vertical distance between two surfaces of constant pressures $p_1$ and $p_2$ ($\rho$ is density). Usually, the steric height is referenced to a constant density $\rho_0$ (e.g., with salinity of 35 and temperature of 0°C) and the independent variable is, for practical reasons, changed from pressure to depth, so

$$h(z_1, z_2) \approx \int_{z_1}^{z_2} \left(1 - \frac{1}{\rho(z)} \right) \rho(z) dz = \int_{z_1}^{z_2} \left(\rho_0 - 1 \right) dz,$$

with the pressure-depth relation $z_1 = z(p_2)$, and $z_2 = z(p_1)$ obtained from a reference profile (see below). Further information on the concept and application of the steric height are given by, for example, Tomczak and Godfrey [2003]. To obtain the density dependent on temperature, salinity and pressure the Fofonoff and Millard [1983] equation of state is applied. The pressure $p$ is approximated from the water depth $D$ utilizing $p \approx 10,000$ Pa m⁻¹ $D$.

[23] Although the steric height at a given location is fully determined by the temperature and salinity profiles throughout the water column, a separation into these two components cannot be achieved owing to nonlinearities in the equation of state and from the definition of the steric height (equations (1) and (2)). However, for small deviations from a reference mean steric height, as regards the interannual to decadal anomalies addressed in section 3, an approximate separation is valid.

[24] Equation (2) can be rewritten as

$$h(z_1, z_2) = \int_{z_1}^{z_2} h_2 dz_2,$$
with \( h_z = \rho_0/\rho - 1 \). Since the integration limits are fixed, time averaging of \( h \) equals averaging the integrand. The anomaly \( h' \), defined as the deviation from the mean state \( \bar{h} = \int h dz \), can be written as

\[
h' = h - \bar{h} = \int h dz - \int \bar{h} dz = \int h dz - \int \bar{h} dz
\]

with integration limits left out, the bars indicating time averaging and defining \( h'_z = h_z - \bar{h}_z \), \( h_z' \) can then be approximated according to Taylor’s expansion formula by

\[
h'_z \approx \frac{\partial h'_z(T^*, S^*)}{\partial T}(T - T_0) + \frac{\partial h'_z(T^*, S^*)}{\partial S}(S - S_0)
\]

\[
= h'_z^T + h'_z^S
\]

(4)

where

\[
\frac{\partial h'_z}{\partial T} = -\frac{\rho_0}{\rho^2} \frac{\partial \rho}{\partial T}
\]

(5)

\[
\frac{\partial h'_z}{\partial S} = -\frac{\rho_0}{\rho^2} \frac{\partial \rho}{\partial S}
\]

(6)

\( T = T(z, t) \) and \( S = S(z, t) \) are observed temperature and salinity, respectively, as function of time \( t \) and depth \( z \), whereas \( T^* \) and \( S^* \) are typical values of temperature and salinity as described below.

[25] Since \( \bar{h}_z \) is not mapped one-to-one onto a salinity and temperature profile, two alternatives are tested for the vertical representation of the references \( T_0 \) and \( S_0 \). The first method uses the depth-dependent time average resulting from the mean profiles \( T_d(z) \) and \( S_d(z) \) for salinity and temperature, respectively, the second method uses scalar values obtained by vertical averaging of \( T_d(z) \) and \( S_d(z) \). Comparisons of the left and right-hand sides of equation (4) reveal that equation (4) is balanced within 20% of the total \( h'_z \) in almost all cases when using scalar reference values, while larger errors result with depth-dependent reference profiles. Thus scalar references are applied for the analyses in section 3.3.

[26] The partial derivatives of \( h_z \) have been evaluated at

\[
T^* = (T_0 + T)/2
\]

(7)

\[
S^* = (S_0 + S)/2.
\]

(8)

This choice ensures a good approximation of \( h'_z \) in equation (4).

[27] The thermal and haline components of an anomaly \( h' \) are finally given by

\[
h'_T = \int h'_z^T dt
\]

(9)

\[
h'_S = \int h'_z^S dt
\]

(10)

respectively.

[28] The steric height integrated from a reference depth \( H \) to the surface is directly linked to the baroclinic surface geostrophic current \( U_{bc} \) forced by the horizontal density gradients above \( H \), with the components \( u_{bc} \) and \( v_{bc} \), thus

\[
\frac{\partial h}{\partial y} = -\int \frac{\rho_0}{\rho^2} \frac{\partial \rho}{\partial y} dz \approx -\frac{1}{\rho_0} \int \frac{\partial \rho}{\partial y} dz = \frac{f}{g} u_{bc}
\]

(11)

\[
\frac{\partial h}{\partial x} = -\int \frac{\rho_0}{\rho^2} \frac{\partial \rho}{\partial x} dz \approx -\frac{1}{\rho_0} \int \frac{\partial \rho}{\partial x} dz = \frac{f}{g} v_{bc},
\]

(12)

using equation (2) and since \( 1/\rho^2 \approx 1/\rho_0^2 \). Ideally, the reference depth \( H \) is chosen deep enough to encompass the whole baroclinic signal, for example, \( \partial \rho/\partial x, \partial \rho/\partial y = 0 \) for \( z < H \). The total geostrophic surface current is then given by the sum of \( U_{bc} \) and the (depth-independent) barotropic current.

2.4. Sampling Uncertainty

[29] Owing to the inhomogeneous distribution of the hydrographic profiles in time and space care is needed when averaging and interpolating to predefined grids and time intervals. Special care has to be taken to ensure accurate description of the temporal variability on seasonal to decadal timescales since the bulk of the observations are taken in the summer season, and measurements in some regions are biased toward certain years or decades.

[30] The basic for calculation of mean and anomaly fields in section 3 are the steric heights derived from combined salinity and temperature profiles that reach a selected reference (depth) level. All temperature and salinity profiles are linearly interpolated to the standard depths as defined in section 2.1. Profiles with missing data for three standard depths or more are discarded.

[31] For calculation of the steric height from the interpolated profiles equation (2) is applied. The spatial resolution in our analysis is about 100 km and is determined by \( 1^\circ \)-latitude \( \times 2^\circ \)-longitude boxes. The mean steric height (see subsection 3.1) is obtained from binning all steric heights on the prescribed mesh. Bins with less than three profiles are discarded.

[32] The seasonal cycle of the steric height (see section 3.2) should ideally be investigated by processing a monthly climatology. As a consequence of the mentioned scarcity of the data (see Figure 2), this cannot be achieved. An essential part of the \( 1^\circ \times 2^\circ \) boxes contains very few or even no data in winter, especially for the months December and January. Thus, as an alternative approach, the harmonic function

\[
h_i(t) = a_i \cos(\omega t - \phi_i) + \bar{h}_i
\]

(13)

with

\[
\omega = 2\pi/365
\]

(14)

is fitted to the available data. Here \( h_i(t) \) describes the steric height of box \( i \) for day \( t \) of the year. Three free parameters have to be determined for every box (instead of 12 for a monthly climatology): The amplitude \( a_i \), the phase \( \phi_i \) (defined by the day of the maximum value of \( h_i \)), and the mean field \( \bar{h}_i \). The approach can only be viewed as an
approximation to the monthly mean climatology of $h_c$. Even if the harmonic function approach can fairly well describe the seasonal cycle for both temperature and salinity, this may not hold for the steric height, especially owing to the nonlinear dependence of density on temperature (see equation (1)).

[33] The essential aim here is to estimate the variability given by the seasonal cycle in relation to the overall longer-term variability, and to provide a rough estimate of the timing of the extremes. Therefore the following conditions have to be fulfilled for each box $i$: (1) less than 10 successive years without data; and (2) either observations for at least 5 months of the year (no matter which year) and not more than 4 successive months without data, or at least observations for 4 months of the year (no matter in which year) and not more than 3 months without data. The method is applied for the period 1950–1999.

[34] For the analysis of the interannual to decadal variability monthly anomalies in steric height and its thermal and haline components are calculated for each box. The anomaly in steric height is defined as deviation from the climatology. No anomaly is given for a box where no climatology is available. The thermal and haline components are determined following equations (9) and (10). Therefore reference temperatures $T_{0i}$ and salinities $S_{0i}$ are needed. For the temperatures monthly climatologies are calculated for each box and standard depth, using the same method as described for the steric height, followed by vertical averaging. For the salinity no significant seasonal variations are found. Thus the annual cycle was ignored, and $S_{0i}$ is defined as time and spatial mean for a given box.

[35] With the described method, anomalies in steric height including their thermal and haline components are calculated for every month with observed temperature and salinity profiles. Gaps in the time series for a box not exceeding five succeeding years are filled by linear interpolation, otherwise they remain unfilled. Boxes with gaps in the time series for the considered periods are discarded from the respective analysis.

[36] It is difficult to unambiguously estimate the errors of the gridded data sets from the observations themselves. The estimates are based on assumptions on the spatial and temporal distribution of the hydrographic profiles and their correlation with the spatial and temporal variability of the estimated fields. The error variance varies spatially with high values for regions with high spatial and temporal variability along the ocean margins and low values for the central deep basins far from coast lines. A bias in the estimated fields might be expected for regions with a strong seasonal cycle and uneven distribution of observations over the year but also for regions with strong spatial variability that cannot be resolved by the observations. Since the spatiotemporal covariances of the estimated fields are not known a priori and cannot be deduced from the observational data, because of sparseness in both time and space, ad hoc assumptions have to be applied or external data has to be taken into account.

[37] One source for external data is the application of a hydrodynamic model. This method is applied here. The condition to be fulfilled by the model is a realistic description of the spatiotemporal covariance of the fields considered. A realistic reproduction of the fields itself is, however, not imperative but corroborates the belief that the model describes accurately the statistics needed.

[38] From a long-term integration of the hydrodynamic model and the spatiotemporal distribution of the observed data error estimates could be calculated. However, since the error characteristics in a general sense are not the focus of this study, a sophisticated description of the error (co-)variances is not needed. Instead, error estimates for the fields analyzed in this study are provided by a more direct method. For this the simulated fields are extracted in time and space consistent with the observation-based fields. The subsample of the field is then treated identical to the observed data and thereafter compared to the full time-space field it was extracted from.

[39] Figure 3 shows such comparisons for the simulated mean steric height for the whole investigated period from 1950 to 1999, and for the anomaly for the decade 1970 to 1979. These are extreme cases with respect to the available data since the 1970s is the decade with the least amount of observed data of the four decades considered (see Figure 2). In addition the data distribution is inhomogeneous over the Nordic Seas area.

[40] Both examples reveal fairly good pattern and amplitude agreement comparing the full and the subsampled fields. Small-scale patterns of length scales shorter than 100 km are not resolved in the subsample because of the low spatial resolution. The subsampled mean steric height (Figure 3, top) is almost identical to the full sampled one, while the two representations of the steric height anomalies in the 1970s (Figure 3, bottom) reveal some structural differences on spatial scales of 100–200 km. However, the large-scale tripole anomalies in the Norwegian Sea, Lofoten Basin, and the Barents Sea opening are preserved in the subsample.

[41] A rough, quantitative estimate of the error in the mean field as well as the decadal anomalies is given by the standard error of the mean that is computed for each of the $1° \times 2°$ boxes as the standard deviation of the binned steric height (anomaly) estimates divided by the square root of the ensemble size. For the decadal steric height field, that is computed as simple average of the binned steric height estimates, this approach results in a reasonable error estimate. For the decadal anomalies, however, it is considered an upper bound error estimate since the more sophisticated approach, that includes reduction of the seasonal cycle and temporal interpolation, is not considered. For the mean field the error estimate is 0.6 cm. Dependent on region, the (asymmetric) distribution of the error estimate has a standard deviation of 0.7 cm. The error estimates for the decadal anomalies are slightly higher and depend on the decade with somewhat lower values in the 1980s and 1990s (0.7 cm) than in the 1960s and 1950s (0.9 cm). The regional variations decrease temporally with 1.2 cm in the 1960s, down to 0.5 cm in the 1990s. Note that the error estimates are small compared to the strength of the spatial signal in case of the mean steric height field. For the decadal anomalies, owing to the smaller signal, the signal-to-noise ratio increases. However, the anomalies (~2 cm) still remain significantly larger than the estimated uncertainties. The comparisons with the modeled fields and the quantitative error estimates strongly indicate that the applied averaging and interpolation methods are applicable for the
3. Observed Steric Height, 1950–1999

In the following we focus the analyses on the mean steric height and its variability on seasonal, interannual and decadal timescales for the second half of the twentieth century.

3.1. Mean Steric Height; Reference Level

Ideally, the steric height should be computed by integration from the sea surface to the depth of horizontally homogeneous water. This should allow the retrieval of the full baroclinic signal associated with the slope of the steric height at the ocean surface. However, the number of profiles extending to such a reference depth is unevenly distributed in space and time. The mean steric heights have therefore been examined using reference depths of 500 m, 1000 m, and 1500 m (Figure 4).

To allow for a convenient comparison all three fields are offset by a constant yielding the same global minimum value (≈0 cm). In all cases, the mean steric height displays minima in the central Greenland and Iceland Seas and high values along the coasts of Norway and Greenland. The two minima have about the same value (0–2 cm) and are separated by a small elevation of a few centimeters north
of Jan Mayen. Significant larger values (20–25 cm) of steric height are found along the west coast of Norway increasing with reference depth, especially from 500 m to 1000 m. In comparison, the steric height is only about 10 cm along the coast of Greenland regardless of choice of reference depth. Clear maxima are found in the Lofoten Basin and along the Norwegian continental shelf break south of the Vøring Plateau.

A significant north-northeastward oriented slope (20–25 cm over 300–400 km) in the steric height is found across the Iceland-Faroe-Ridge (IFR) when profiles extending to 500 m depth are used (Figure 4, left). The average depth of this ridge is 500–600 m (see Figure 1). In turn, the use of a reference depth of 500 m is probably more appropriate for the analyses of the steric height field pattern across the IFR.

In the Nordic Seas the difference between the three steric height fields primarily reveals the effect of the spatial distribution of temperature and salinity between, particularly, 500 m and 1000 m. The most striking difference between the fields is the apparent increase in the steric height maximum in the Lofoten Basin with increasing reference depth. Apart from that the overall structure remains largely unchanged when the density structures between 1000 and 1500 m are included. This suggests that a major part of the baroclinic circulation can be explained by the temperature and salinity structures in the upper 1000 m. As the number of profiles to 1000 m is more favorable, a standard reference depth of 1000 m has consequently been used in the remaining analyses.

The large-scale baroclinic surface circulation pattern induced by the slope of the mean steric height field favors a general cyclonic flow in the Nordic Seas combined with weaker closed cyclonic circulation in the Greenland and Iceland Seas associated with the two local minima. A northward flow is depicted in the Norwegian Basin turning eastward at about 70°N. Part of this flow is entering the Barents Sea while the remaining flow turns northward toward the Fram Strait. Weaker southward flow is encountered along the east coast of Greenland and through the Denmark Strait. By representing the mean slope in the Norwegian Sea with a 20 cm steric height difference over a distance of 500 km it follows that the corresponding baroclinic surface current is about 3 cm s⁻¹ (with $U_{bc} = g/f \times \Delta h/L$, where $g = 9.81$ m s⁻² and $f = 1.4 \times 10^{-4}$ s⁻¹ the Coriolis parameter). In comparison, the stronger slope across the IFR yields a mean baroclinic surface current of 5–6 cm s⁻¹ steered along the ridge to the east-southeast. In summary the baroclinic circulation inverted from the mean steric height field reveals similarities in structure and magnitudes to direct surface current observations derived from Lagrangian Drifters [Jakobsen et al., 2003, Figure 5].

### 3.2. Seasonal Variability

The seasonal cycle of the steric height provides an indication of the main annual changes in the Nordic Seas hydrography. The amplitude $\alpha$ and the phase $\phi$ of the seasonal cycle as determined from equation (12) are shown in Figure 5. The mean steric height field $h_i$ is not shown explicitly, but it reproduces the field presented in the middle plot of Figure 4 reasonably well. Note also the apparent uneven data distribution resulting from the minimization of sampling uncertainty associated with undersampled data boxes (seen as white).

The seasonal cycle is generally stronger in the Lofoten Basin and along the Norwegian coast south of the Voring Plateau (about 5 cm) compared to the Greenland and Iceland Seas (about 2 cm) (Figure 5, left). This is expected, as the overall temporal variability is higher along the path of northward flowing Atlantic Water than for the other water masses occupying the basins. Maximum steric height is largely found in late September and early October. However, over the shelf break east of Greenland and west of the Voring Plateau it occurs already in August, while it appears first in early November at the southern Barents Sea opening and over the Mohns Ridge.

The general pattern in the amplitude of the seasonal cycle of the steric height is in agreement with Mork and Skagseth [2005] using the World Ocean Atlas [WOA01; Conkright et al., 2002] database, although their results are more smoothed. On the other hand, differences are found in the phase estimates. The NISE data gives a sea level
maximum 20 days later in the Greenland Sea and up to about 40 days earlier in the Iceland Sea compared to the WOA01 data. Since Mork and Skagseth [2005] also used a harmonic function to estimate the seasonal cycle most of these differences are explained by error estimates [see Mork and Skagseth, 2005], different interpolation methods, as well as the inherent differences in the data sets.

The relative importance of the seasonal variability is shown in Figure 5 (right). This is determined from the amount of variability in the monthly steric height that can be explained by the mean annual cycle. It turns out that in general only about 20% of the fluctuations in the monthly field is due to the seasonal variability, with some local exceptions.

The temperature and salinity contributions to the seasonal cycle of the steric height have also been examined. The results (not shown) suggest that the predominant impact is associated with the temperature variations. This is in agreement with other studies [e.g., Gill and Niler, 1973; Stammer, 1997; Mork and Skagseth, 2005]. The impact from the salinity variations, on the other hand, is weak, since the annual changes in salinity are small. In turn, the corresponding phases cannot be adequately determined with sufficient precision for most of the Nordic Seas.

### 3.3 Decadal Variability

The decadal-scale anomalies (relative to the mean from 1950–1999) of the steric height for the four decades from the 1960s to the 1990s are displayed in Figure 6, together with the individual contribution from the thermal and haline components.

The steric height has increased by as much as 8 cm over the 40 years period (Figure 6, left) in most of the central and northeastern part of the Nordic Seas. This region comprises the slope west of the Barents Sea opening, the Lofoten Basin and the eastern part of the Greenland Sea. The western part of the Greenland Sea is excluded from the analysis owing to sparse data coverage. The Iceland Plateau and the Voring Plateau both show a positive change of 3 cm in the 1990s compared to the three decades before. In the Norwegian Basin, no uniform trend is found. Overall, steric height has increased in the northern and central parts of the basin. North of the Faroe Islands on the contrary, a negative trend of roughly 3 cm is found when comparing the 1960s with the 1990s.

The individual thermal and haline contributions to the steric height anomaly (middle and right plots in Figure 6, respectively) suggest that the observed anomalies in steric height on decadal timescale are mainly governed by a basin-wide freshening of the upper 1000 m of the water column. The area-mean, haline increase of 3.5 cm is equivalent to a surplus of roughly 2000 km$^3$ of fresh water from the 1960s to the 1990s. Note that this estimate encompasses only the upper 1000 m of the water column, and roughly one third of the Nordic Seas area is excluded. For comparison, Curry and Mauritzen [2005] report ~4000 km$^3$ net freshwater storage increase for the entire Nordic Seas for the period 1965–1995 while Peterson et al. [2006] estimate for the same period ~3000 km$^3$ (their Figure 3c).

Changes in temperature have a dipole structure adding to steric height in the central and northeastern part of the Nordic Seas, but tending to counteract the freshening effect in the southern and southeastern part of the region (Figure 6, middle column). At its northern center of action along the Barents Sea opening, the temperature signal dominates the haline signal while at its southern center of action north of the Faroe-Shetland Ridge, the thermal component has roughly the same amplitude as the haline component and thus the signals cancel out each other. The apparent uniform contribution from salinity and the more dipole contribution from temperature on the evolution of the steric height strongly indicate different forcing mechanisms for the two anomalies.

Whereas variations in the air-sea heat flux will lead to changes in the temperature, changes in the salinity will be manifested through variations in evaporation minus precip-
itation as well as freshwater run-off from land. In addition, one has to account for differences in the water transformation processes of heat and salt in the Nordic Seas and Arctic Ocean. Finally, part of the anomalies found in the Nordic Seas likely originates elsewhere and have been advected to the region with the North Atlantic Current.

4. Leading Modes of Variability

[58] Figure 6 reveals that the hydrography in the upper 1000 m of the Nordic Seas has undergone considerable changes in the last four decades of the twentieth century. This is indicated by the interannual variability in the steric height field and its respective contribution from the thermal and haline components.

[59] The following analyses of the variability on interannual timescales are concentrated to the Norwegian Sea and the eastern part of the Iceland Plateau. These regions comprise the area that is occupied by Atlantic water, the Polar Front and some part of Arctic waters west of the front. Insufficient data coverage excludes analyses of the region toward Greenland. Different processes might change the hydrography in the upper 1000 m on interannual to decadal timescales, in particular through (1) changes of heat and salt fluxes via transport variability in the North Atlantic Current (NAC) across the GSR [Mork and Blindheim, 2000; Orvik and Skagseth, 2005; Hátún et al., 2005]; (2) changes in local air-sea heat and fresh water fluxes [Dickson et al., 2000; Furevik, 2001, 2000; Furevik and Nilsen, 2005]; and (3) redistribution of water masses through wind effects, such as a shift in the Polar front [Blindheim et al., 2000; Furevik et al., 2002].

[60] To reach further insight into the forcing of the interannual variability in the Nordic Seas hydrography the leading modes of variability of the steric height as well as the respective thermal and haline signal components are examined and compared (Figures 7–13). The Empirical Orthogonal Function (EOF) method is used to explore the spatial patterns, while their temporal evolutions are estimated from time series of the Principal Components (PCs) of the corresponding EOFs.

4.1. Steric Height Variability

[61] The spatial pattern of the leading mode of annual steric height variability (Figure 7) displays a basin-wide structure with higher values at the Barents Sea opening than over the Iceland Plateau and a regional maximum in the Lofoten Basin. In respect to the entire region the mode explains approximately 35% of the variance of the annual fields.

[62] In comparison more than 50% of the variability is explained in the Lofoten Basin and along the continental slope at the Barents Sea opening to the north. The same
holds for the western part of the Norwegian Basin, while somewhat lower values are found on the Icelandic Plateau. In the southeastern part of the Nordic Seas, on the other hand, the variance values drop below 10%.

The PC of the leading EOF describes a decrease from 1960 to 1968 followed by a strong increase until 1976 (Figure 8). After a short stagnation in the second half of the 1970s and a moderate decrease in the beginning 1980s it

Figure 9. Principal component analysis of the momentum flux at the atmosphere-ocean interface taken from NCEP/NCAR reanalysis data [Kalnay et al., 1996]. (left) Third EOF of the monthly mean vector field for the North Atlantic (22°N–83°N, 70°W–20°E) for the period 1948 to 2003 for the respective (top) zonal and (bottom) meridional components. (right) Explained variance (in percent) for the (top) zonal and (bottom) meridional components.

Figure 10. Principal component analysis of the haline component of the observed annual steric height anomaly for the period 1955–1997. (left) First EOF (in centimeters). (right) Explained variance (in percent).
increases further, reaching the absolute maximum in 1990. Thereafter it decreases to values of the second half of the 1970s. In association with these extremes in 1968 and 1990 the corresponding increase in steric height amounts to about 10 cm at the Barents Sea opening and 12 cm in the Lofoten Basin.

As seen from Figure 6, the variability in the haline component explains a significant part of the decadal variability in the steric height. This is also the case on the annual timescale. The PC (Figure 11) of the haline contribution to the steric height variability correlates strongly (0.80) with the leading mode of steric height variability (Figure 8). A large part of this correlation is made up by the strong decrease in salinity during the 1970s that, in turn, contributed to an increase in steric height. In the Lofoten Basin this increase exceeds 3 cm. This major increase in the 1970s is by far the most prominent signal in the PC and may, in fact, be characterized as a regime shift. The interannual variability both during 1955–1970 and 1979–1997 is clearly smaller than during this shift. Note, however, that temporal averaging applied to data gaps (see section 2.4) to some extent smooths the signal.

4.3. Thermal Component

The more localized changes in the spatial pattern and temporal evolution of the steric height (Figures 7–8) are to a large degree caused by the contribution from the temperature field (Figures 12 and 13). In contrast to the haline component no dominating mode of variability in the evolution of the temperature field is found on basis of a linear decomposition.

The leading mode (Figure 12, top) explains globally 22% of the annual variability but more than 50% of the variations in the eastern part of the Lofoten Basin. The pattern reveals an elongated structure that generally follows the pathway of the upper layer, northward flowing Norwegian Atlantic Current in the Norwegian Sea with maxima in the Lofoten Basin and around the Faroe Islands. In comparison, weak first and second EOF mode signals are found on the Iceland Plateau and north of the Lofoten Basin. The PC (Figure 13) shows a considerable correlation (0.75) with the leading time series for the haline component. This correlation is significant in a t-test on the 99% confidence level. Good agreement of the two time series is found especially on decadal timescales. Both PCs display a jump toward higher values in the second half of the 1970s. However, large discrepancies are found comparing interannual variability, in particular in the 1960s when the thermal component contains a strong minimum (1961–1962) followed by a maximum in 1968. During this period the PC for the haline component is comparably flat. Those differences
coincide with extremes in the hydrographic properties of the inflowing Atlantic water. Observations at the upper Faroe slope [Turrell et al., 1993] indicate temperatures about 0.7°C below the mean value in the second half of the twentieth century, and with salinities about 0.08 above the 50-year mean. In comparison, the temperature is very high (0.8°C above normal) while the salinity is 0.04 above normal during 1961/1962.

[72] The second mode of the thermal component (Figure 12, bottom) explains globally 13% of the variability. Positive values are found at the Barents Sea opening while negative values are displayed in the southern part of the Nordic Seas east of Iceland. The PC of the second thermal component of steric height variability is highly correlated to both the winter (DJFM) NAO index [Hurrell, 1995] as well as the leading mode of the annual North Atlantic total heat flux (see Figure 13). The correlation to the NAO winter index is very high for the period after 1972 (0.89) while vanishing correlation is found from 1965 to 1971 (r = −0.09).

[73] The total annual heat flux data is taken from NCEP/NCAR reanalysis for the Atlantic region (22°N–80°N, 70°W–20°E) for the period 1952 to 2001. The EOF of the leading mode (not shown here) is similar to the pattern found by Furevik and Nilsen [2005, Figure 9d] when regressing total heat flux onto the NAO index for approximately the same area and period. A correlation of 0.77 of the PC to the 2nd PC of the thermal contribution to the steric height variability and 0.86 to the winter NAO index when both the NAO index and the heat flux are filtered with a running 5-year mean.

[74] A reasonable correlation is also found to the cyclone counts and intensity north of 60° N in winter [see Serreze et al., 2000, Figure 9]. They reported maxima in the first half

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**Figure 12.** Principal component analysis of the thermal component to the observed annual steric height variability for the period 1955–1997. (left) (top) First and (bottom) second EOFs (in centimeters). (right) Explained variances for the (top) first and (bottom) second EOFs (in percent).
A comprehensive data set of temperature and salinity observations has been used to analyze the spatial and temporal variability of the steric height in the Nordic Seas for the second half of the twentieth century. The mean seasonal cycle as well as variations on interannual to decadal timescales have been considered.

The upper 1000 m comprise the main part of spatial variability of steric height in the Nordic Seas (Figure 4). Using a hydrographical atlas for the Nordic Seas based on the NISE data the steric height shows mainly the spatial variability of the temperature signal for the whole Nordic Seas except for a narrow (few hundred kilometers) band along the Norwegian and Greenland coast lines where low surface salinity results in a high steric height. Characteristic temperatures (salinities) vary from around 0°C (34.9) for the Arctic water in the Greenland Sea to 8°C (35.2) for the inflowing Atlantic water over the GSR. The corresponding density variations are found utilizing the equation of state [Fofonoff and Millard, 1983] using constant salinity (temperature) of 35 (4°C) and constant pressure of $p = 2 \times 10^6$ Pa (corresponding to 200 m water depth). The thermal contribution to the spatial density variations reaches 0.87 kg m$^{-3}$ while the haline component is less than one third of that (0.24 kg m$^{-3}$). Thus the high (low) steric heights in the Norwegian Sea (Iceland and Greenland Seas) reflect the high (low) temperature of the water masses originating from Atlantic (Polar) sources [see, e.g., Hopkins, 1991; Hansen and Østerhus, 2000].

Spatial steric height variability below 1500 m was not considered here. Moreover barotropic currents that cause additional slopes in the MDT are not included in the steric height field. However, from observations [Skagseth and Orvik, 2002; Fahrbach et al., 2001] and theoretical considerations [Nøst and Isachsen, 2003] strong barotropic flow can be expected along the margins of the basins and over the continental shelf breaks. This, in turn, could imply impact of the barotropic component to the shape of the MDT.

On the other hand, the MDT for the region between Greenland and the UK obtained from gravimetric observations, altimetric MSSH and a composite MDT from hydrodynamic modeling [Hipkin and Hunegnaw, 2006] confirm the general structure and the magnitude of the spatial variations found in the mean steric height. The similarity of these fields suggests that the contribution from the barotropic currents and temperature and salinity variations for depths below 1000 m have negligible impact on the MDT. The mean steric height obtained for the upper 1000 m is therefore considered to be a reasonable estimate of the MDT. Note however that for the accurate determination of the barotropic currents very precise knowledge of the MDT is needed. A difference in the sea level of $\Delta \zeta = 1$ cm, due to a slope in the sea surface, corresponds to roughly 2 Sv barotropic transport $U$ (with $U_{bt} = g/f \Delta \zeta \times H$, $f = 1.4 \times 10^{-4}$ s$^{-1}$, $g = 9.8$ m s$^{-2}$ and $H = 2000$ m). A baroclinic transport of the same magnitude will, on the other hand, be related to a strong current that is confined to the upper several 100 m of the water column. To achieve this the slope in the surface topography for a given length scale has to be several times the slope related to the barotropic current depending on the characteristic depth of the baroclinic flow.

A strong connection between the temporal variability in both the steric height field and the Dynamic Topography (DT) is expected for periods longer than the synoptic scale of wind field variability. According to results from the regional OGCM a major part of the monthly-scale variability in DT is related to fluctuations in the steric height (Figure 14). Furthermore, for all regions with strong variability in DT, the correlation to the steric height is high. For interannual and longer periods advection plays an important role in altering the hydrographic properties. It can therefore be expected that for those timescales the part of variability in DT explained by variations in the steric height will increase.

The mean seasonal cycle of the steric height (Figure 5) is found to be generally small in the order of a few centimeters and is not significantly altering the structure of the mean field. Larger variability is found on decadal timescales (Figure 6). The different regions show a rising trend in steric height of 3 to 8 cm from the 1960s to the 1990s. Moreover, variability on interannual timescales are even larger. The leading mode of annual steric height variability (Figures 7 and 8) indicates strong variability in the Lofoten Basin with variability in the order of 3 cm and a 12 cm higher steric height value in 1990 compared to 1968. While the mode explains 60% of the variability in the Lofoten area (Figure 7, right) it has to be noted, however,
that other modes of variability have to be taken into account that might diminish the overall variability and trend.

[s1] The range of temporal steric height variability (5–10 cm) at interannual to decadal timescales is significantly less than the spatial variability (25 cm). It is therefore not strong enough to destroy the overall structure of the mean steric height field. On the other hand, when the DT is composed of an anomaly added to a reference MDT, as done in data assimilation to reference altimetric sea level anomaly, the reference period for the MDT and the anomaly has to be consistent to avoid significant errors due to the strong decadal trend in the DT [Siegismund, 2006].

[s2] The general rise in the large-scale steric height pattern is not changing the baroclinic part of the circulation internally. However, an interesting finding is that the region north of the Faroes displays a trend with high values in the 1960s and 1970s and low values in the 1990s. This might have caused a reduction of the Faroe Current in the 1990s compared to the 1960s and 1970s. This cannot be validated owing to lack of direct current observations.

[s3] While spatial variations in temperature is the main factor for the mean steric height signature of the Nordic Seas, the interannual to decadal variability is dominated by the variation in salinity. The freshwater surplus of the Norwegian and Iceland Seas could either be dominated by local air-sea freshwater flux or advected to the area. Dickson et al. [2000] estimate a 15 cm difference in annual precipitation comparing composites of years with high and low NAO winter (DIF) indices, while evaporation is hardly changing [Furevik and Nilsen, 2005; Hurrell, 1995]. With the density of freshwater roughly 1000 kg m\(^{-3}\) and \(\rho_0 = 1027\) kg m\(^{-3}\), 15 cm of fresh water results in 0.4 cm rise in steric height (equation (2)). This number is small compared to the overall variability found. From that we conclude that effects of evaporation and precipitation play a minor role for the interannual to decadal steric height variability in the Norwegian Sea.

[s4] We rather expect advection to dominate the interannual to decadal variability in steric height. First, the dominance of the haline over the thermal component of steric height variability points to advective processes. Second, the leading mode of the haline component of annual steric height anomaly (Figures 10 and 11) explains the predominant part of the variability, particularly 70% of the variability in that part of the Nordic Seas that is occupied by Atlantic Water. The dominance of the leading mode is expected since the considered timescales of one to several years are in the order or above the transit time of Atlantic water through the Norwegian Sea [see, e.g., Gao et al., 2005; Orre et al., 2007]. Signals both in the mass transport itself or in the water properties transported are thus connected on these timescales.

[s5] Strong positive freshwater source anomalies in the Arctic Ocean in the second part of the twentieth century [Peterson et al., 2006] suggest a freshening of the Arctic water mass as the main reason for the salinity decrease found in the Nordic Seas. However, Peterson et al. [2006] conclude from their findings that flushing of freshwater accumulated in the Arctic into the Nordic Seas and the Subpolar Basins depend on the atmospheric circulation patterns. Furthermore, B2000 suggest that the spreading of the Arctic water in the Norwegian Sea is connected to the zonal extent of Atlantic Water.

[s6] From our analysis we conclude that the dominant process for the freshening of the Nordic Seas from the 1960s to the 1990s is an eastward shift in the Polar Front connected to the local wind forcing. We found a strong correlation between the leading mode of variability in steric height and the zonal extent of Atlantic Water; low steric height in relation to westward extension of the Atlantic Water and vice versa. In addition the leading mode of steric height variability is better correlated to the north-south component of local wind stress variability than with the NAO index (Figure 8). This indicates fluctuations in the local Ekman transport as initial cause for interannual to decadal steric height variability rather than a remote control.

[s7] As is expected for advection/mixing processes the leading modes of the thermal and haline contributions to the steric height variability are clearly correlated (Figures 10–13). The time series (Figures 11 and 13) reveal, however,
some remarkable differences, especially in 1968 with a significant extreme in the thermal component which was encountered with an extreme of opposite sign in the haline component. Another, but less significant discrepancy took place in 1961/1962. Both differences can be explained by observed anomalies in the temperature and salinity of the Atlantic inflow over the Iceland-Scotland Ridge. In conclusion, the leading mode of salinity and temperature contribution to the steric height variability contains both the result of local redistribution of water masses from Atlantic and Arctic origin as well as variations in the properties of the Atlantic inflow. Since the variations in salinity and temperature of opposite phase cause additive contributions to the steric height it is especially sensitive to the 1968 event. Consequently, the very strong global minimum in the leading PC of steric height variability is found in 1968.

[88] For the thermal component of the steric height variability the ocean-atmosphere heat flux adds to the advective contribution. Furevik [2000] analyzed a positive temperature anomaly in the Fugløya-Bjørnøya section at the southern part of the opening of the Barents Sea for the years 1990–1992. He suggested reduced heat flux to the atmosphere in the Nordic Seas region, and increased effective advection speed of the North Atlantic Current (NAC), as two likely explanations for the signal. From our analysis we conclude that the 1990s on average were a period with high temperatures along the whole Barents Sea opening and west of Svalbard. The temperature signal in the 1990s as seen from the thermal component of the steric height anomaly (Figure 6) is positive along the path of the NAC north of the Lofoten Islands. However, the signal switches upstream and remains negative up to the inflow region along the Iceland-Faroe-Shetland Ridge. Consequently, it cannot be explained by advection within the NAC.

[89] In periods with a high NAO index as in the second half of the 1980s and the first half of the 1990s a reduced heat flux is expected for the eastern part of the Nordic Seas [Furevik and Nilsen, 2005; Blindheim et al., 2000]. The reduction in the heat flux might be connected to both a narrowing and acceleration of the NAC that limit the area and exposure time of warm water at the surface [Blindheim et al., 2000; Furevik, 2001]. In addition, changes in the local atmospheric boundary layer conditions may also affect the heat flux [Furevik and Nilsen, 2005]. Our steric height and temperature increase encountered at the Barents Sea opening is supported by this explanation of a local reduction of the heat flux. On the other hand, the second mode of variability in the thermal component of the steric height anomaly is strongly correlated to the winter NAO index and to the total heat flux variability of the North Atlantic (Figure 13). This reproduces the dipole structure of the decadal anomalies (Figure 6) and displays very high anomalies for the first half of the 1990s. However, part of the variability in the thermal contribution to the annual steric height variability is not explained by the leading two modes.

6. Conclusion

[90] The steric height variability in the Nordic Seas has been examined at seasonal, interannual and decadal scales using in situ and modeled data for the period 1950 to 1999. The main findings are as follows.

[91] 1. The bulk of the spatial variations in hydrography of the Nordic Seas occurs within the upper 1000 m of the water column.

[92] 2. The spatial temperature variations determine the pattern of the mean steric height field with secondary contribution from the salinity. This is valid for both the Atlantic Water in the eastern and the Polar Waters in the western part of the Nordic Seas. Salinity dominates only in regions with very fresh water along the coasts of Norway and Greenland.

[93] 3. The range of spatial change in the steric height reaches 25 cm, and maintains a general shape that favors the cyclonic circulation pattern in the Nordic Seas with two additional internal cells around the minima in the Greenland and Iceland Seas.

[94] 4. The seasonal cycle in steric height varies in amplitude from ~2 cm in the Greenland and Iceland Seas to higher values along the Norwegian coast with maximum west of the Lofoten Islands (~5 cm). The seasonality is mainly caused by the corresponding variations in the thermal expansion and is quite low in comparison to the interannual to decadal variability (~10 cm).

[95] 5. The interannual to decadal variability in steric height is caused by a general freshening of the Nordic Seas while the contribution from temperature variations consists mainly of a north-south dipole resulting in an enhanced increase of steric height at the entrance to the Barents Sea (due to warming) and a reduced increase of steric height in the Norwegian Basin (due to cooling).

[96] 6. The bulk of the variability of the steric height in the Norwegian Sea is connected to local meridional wind stress variations and correlates with the zonal extension of Atlantic Water in the area. Reduced heat flux to the atmosphere in a high NAO index regime is one major cause for the increasing temperatures from the 1960s to the 1990s toward the Barents Sea opening.

[97] These findings suggest that reliable knowledge of the comparatively high spatial variability of the steric height and MDT on regional scales within the Nordic Seas will benefit from the precise estimation of the geoid to be derived from the GOCE mission with planned launch in March–April 2008. However, when the absolute DT is constructed from the sea level anomaly and the MDT, as in data assimilation, the reference period for the MDT and the anomaly has to be consistent to avoid significant aliasing due to the strong decadal trend in the steric height [Siegismund, 2006].

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References


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