Steric Sea Level Change in the Northern Seas

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ABSTRACT

Ocean temperature and salinity data over the period 1950–2000 in the Northern Seas, defined here as the North Atlantic Ocean (north of 50°N), North Pacific Ocean (north of 40°N), and Arctic Oceans, are combined to diagnose the steric (i.e., density) contribution to sea level variation. The individual contributions to steric height from temperature (thermosteric height) and salinity (halosteric height) are also analyzed. It is found that during 1950–2000, steric height rose over the study’s domain, mostly as a result of halosteric increases (i.e., freshening). Over a shorter time period (late 1960s to early 1990s) during which climate indices changed dramatically, steric height gradients near the Nordic Seas minimum were reduced by 18%–32%. It is speculated that this may be associated with a local slowing of both the Meridional Overturning Circulation and the southward flow through Fram Strait. However, steric height increases in the North Pacific Ocean during this time imply a possible acceleration of flow through the poorly measured Canadian Arctic. Evidence that the Great Salinity Anomaly of the late 1960s and 1970s had two distinct Arctic Ocean sources is also found: a late 1960s export of sea ice, and a delayed but more sustained 1970s export of liquid (ocean) freshwater. A simple calculation indicates that these Arctic Ocean freshwater sources were not sufficient to create the 1970s freshening observed in the North Atlantic Ocean.

1. Introduction

Global sea level rose by 1.5–2 cm decade\(^{-1}\) over the past century (Wadhams and Munk 2004). About 0.5 ± 0.2 cm decade\(^{-1}\) of this trend was from steric effects, that is, changes in density that affect the volume and thus the height. Most of the steric change was from warming, although freshening played a role at high latitudes (Antonov et al. 2002).

Levitus et al. (2005) examined the contributions from both temperature and salinity to linear steric height trends using interpolated, gridded hydrographic data averaged into 5-yr running means over 1955–98. They found steric height increases in the North Pacific subpolar gyre and in the Nordic Seas, mostly from freshening. The North Atlantic subpolar gyre also freshened, but a compensating cooling meant that the net steric height change was small. Curry and Mauritzen (2005), using a different interpolation technique, found a large increase in North Atlantic Ocean freshwater content during the Great Salinity Anomaly (GSA) of the 1970s. Freshening continued at a slower pace until the 1990s, when the trend reversed and freshwater content began to diminish.

In the Arctic Ocean, Swift et al. (2005) examined annual hydrographic fields collected by Russian scientists over the latter half of the twentieth century and found a salinification over the upper few hundred meters of the Arctic Ocean during the mid-1970s. Also, tide gauge data were analyzed by Proshutinsky et al. (2004) to determine that sea level rose along the Russian arctic coast from the 1950s to the 1980s. They then used a numerical sea ice–ocean model to determine that 35% of this rise, or 6.4 mm yr\(^{-1}\), was from steric effects. The model also indicated that sea level was falling in the central Arctic Ocean.

In this study, we use hydrographic data over the period 1950–2000 to reexamine steric sea level change in the Northern Seas, which we define here as the North Atlantic Ocean north of 50°N, the North Pacific Ocean north of 40°N, and the Arctic Ocean. We consider both the thermosteric and halosteric components of steric change. The combination of several data sources allows us to consider the large-scale exchange of water properties between ocean basins, as well as possible effects...
on global ocean circulation. In section 2, we present a
discussion of the data used in this study, followed in
section 3 by a review of the mean field of steric height
and its separate contributions from temperature and
salinity. We present basin-averaged time series in sec-
tion 4, including a detailed look at freshwater and heat
transfers between basins. In section 5 we discuss how
changes over the late 1960s to the early 1990s affected
the large-scale steric height field and speculate about
potential effects on the large-scale ocean circulation.
We close with a discussion of our results.

2. Datasets

Our main dataset is the World Ocean Database 2001
(WOD01; Conkright et al. 2002), consisting in part of
90,905 paired temperature and salinity profiles during
the years 1950–2000 over latitudes 50°–80°N in the At-
lantic Ocean and 40°–60°N in the Pacific Ocean. We
consider here only those profiles that extend to at least
1000-m depth. Baffin Bay is not included in our analy-
sis, owing to a particularly severe summer bias in the
small area enclosed by the 1000-m isobath (99.6% of
data there are from July–October). The data density
(Fig. 1) is best in the Nordic Seas and close to the North
Pacific Ocean coastlines. Within the Nordic Seas, the
North Atlantic subpolar gyre, and the North Pacific
Ocean (see Fig. 1), profile pairs were discarded if their
steric height (see section 3) exceeded three standard
deviations from the mean in each 200-km bin. This
eliminates less than 1% of the profile pairs.
In the central Arctic Ocean, WOD01 is too sparse for meaningful time series analysis, although it can be used to study trends over the Russian arctic shelves (Steele and Ermold 2004). Historical Russian temperature and salinity data are publicly available only as decadal averages from the 1950s through the 1980s (Timokhov and Tanis 1997, 1998; hereafter referred to by its common name, EWG for Environmental Working Group). However, gridded fields of annual average steric height from these same hydrographic data are available (Timokhov and Tanis 1997) on a roughly 182-km grid. These fields have been relatively underutilized by the scientific community to date. Unfortunately, steric heights were referenced to a relatively shallow 200-m depth. Using the decadal mean hydrography, we find that layer contains 71% of the Arctic Ocean mean 0–1000-m steric height. To better compare with the 0–1000-m steric height computed using WOD01 data, we have extended the Arctic Ocean mean steric height field discussed in section 3 to 1000 m by using the gridded Polar Science Center Hydrographic Climatology (PHC; Steele et al. 2001). Since PHC is a time-mean climatology, temporal variability in Arctic Ocean steric height in this study comes only from the upper 200 m.

3. Mean steric height in the Northern Seas

The Northern Seas play a vital role in two large-scale ocean circulation pathways: the Atlantic Ocean’s Meridional Overturning Circulation (MOC) and the Arctic Flow-Through (AFT). The first is well documented in the literature. The MOC represents the Atlantic branch of the “global ocean conveyor belt” (Broecker 1991), which at the surface brings relatively warm, salty water northward in the North Atlantic current. Along the way, its density increases as heat is lost to the atmosphere, until much of it sinks in the Nordic Seas and returns southward. The second circulation pathway is less well recognized. The AFT is the flow across the Arctic Ocean, usually assumed to be driven by high sea level in the North Pacific Ocean relative to the North Atlantic Ocean. While the volume flux is small [~0.8 Sv (1 Sv = 10⁶ m³ s⁻¹) mean flow through Bering Strait], the freshwater and chemical nutrient signals are significant (De Boer and Nof 2004; Jones et al. 1998; Wijffels et al. 1992; Woodgate et al. 2005). This pathway represents a shortcut, or “backdoor,” for the global ocean circulation between the North Pacific and North Atlantic Oceans. When this door closes (e.g., during an ice age), model simulations suggest that the MOC may be strongly affected (De Boer and Nof 2004; Goosse et al. 1997).

Both of these pathways are represented in Fig. 2, which shows sector zonal averages of steric height over the upper 1000 m of the ocean, using PHC (Steele et al. 2001). Steric height is also known as dynamic height (DH) because lateral gradients in DH drive surface ocean circulation via geostrophic balance. Dynamic

![Fig. 2. Sector zonal means of DH (black), thermosteric dynamic height (DHₜ, red), and halosteric dynamic height (DHₛ, green) over 0–1000-m depth, using the PHC database (Steele et al. 2001). The Atlantic sector is defined as 90°W to 100°E north of 60°N, and 90°W to 0° between 20° and 60°N, while the Pacific sector is defined as 100°E to 90°W from 20° to 90°N, excluding the western Caribbean Sea (inset). Vertical lines show ±1 standard deviation from the zonal sector means. Bering Strait is ~50 m deep. Arrows denote surface transport in the MOC and AFT.](http://journals.ametsoc.org/jcli/article-pdf/20/3/403/3940993/jcli4022_1.pdf)
height is the vertically integrated departure of density \( \rho \) at salinity \( S \), temperature \( T \), and pressure \( P \) from a standard reference value \( \rho_{\text{ref}} = \rho_{T_{\text{ref}},S_{\text{ref}},P} \), that is,

\[
DH = \int_{0m}^{1000m} \left( \frac{\rho_{\text{ref}} - \rho_{T_{\text{ref}},S_{\text{ref}},P}}{\rho_{\text{ref}}} \right) dz.
\]  

(1)

For calculating absolute mean steric height fields as in Fig. 2, we use \( S_{\text{ref}} = 36 \) and \( T_{\text{ref}} = 0^\circ \text{C} \), which produces mostly positive values for DH and its temperature and salinity components. On the other hand, for calculating anomaly time series (section 4) we use climatological mean fields of salinity and temperature as references in order to avoid numerical artifacts from differencing large, similar numbers. Figure 2 also shows thermosteric dynamic height (\( DH_T \)) and halosteric dynamic height (\( DH_S \)), that is,

\[
DH_T = \int_{0m}^{1000m} \left( \frac{\rho_{\text{ref}} - \rho_{T_{\text{ref}},S_{\text{ref}},P}}{\rho_{\text{ref}}} \right) dz
\]  

(2)

\[
DH_S = \int_{0m}^{1000m} \left( \frac{\rho_{\text{ref}} - \rho_{T_{\text{ref}},S_{\text{ref}},P}}{\rho_{\text{ref}}} \right) dz
\]  

(3)

such that \( DH_T \) (\( DH_S \)) is the contribution to dynamic height from temperature (salinity) alone, that is, \( DH = DH_T + DH_S \), modulo nonlinear terms in the equation of state that are generally less than 1\% (see also Landerer et al. 2007).

Thermosteric and halosteric dynamic heights can also be related to vertically integrated heat content (HC) and freshwater content (FWC) for small departures of temperature and salinity from their reference values, that is,

\[
DH_T \equiv \left( \frac{1}{\rho_{\text{ref}} c_p} \right) \left( \frac{\partial \rho}{\partial T} \right)_{T_{\text{ref}},S_{\text{ref}}} \int_{0m}^{1000m} \rho_{\text{ref}} c_p (T_{\text{ref}} - T) dz
\]  

(4)

\[
DH_S \equiv \left( \frac{S_{\text{ref}}}{\rho_{\text{ref}}} \right) \left( \frac{\partial \rho}{\partial S} \right)_{T_{\text{ref}},S_{\text{ref}}} \int_{0m}^{1000m} \left( \frac{S_{\text{ref}} - S}{S_{\text{ref}}} \right) dz,
\]  

(5)

where heat capacity \( c_p = 4218 \text{ J} \cdot \text{kg}^{-1} \cdot \text{K}^{-1} \), angle brackets ( ) denote the vertical average, and the integrals in Eqs. (4) and (5) are HC and FWC, respectively. In the arctic seas, the proportionality factor between \( DH_T \) and HC depends on \( \partial \rho / \partial T \), which varies strongly with temperature and thus with position. However, the proportionality factor between \( DH_S \) and FWC is essentially a constant \( \approx 0.028 \) because \( \partial \rho / \partial S \) is a very weak function of salinity. This allows easy conversion between \( DH_S \) and FWC, as discussed below.

Figure 2 shows that the Nordic Seas (60°-80°N) contain a DH minimum, that is, a gravitational “potential well” drawing surface water from both the south (the North Atlantic Ocean), and from the north (the North Pacific Ocean via the Arctic Ocean). Of course the large-scale ocean current and sea surface height fields in the ocean arise from a complex combination of both thermohaline and wind forcings. The meridional pressure gradients represented in Fig. 2 are associated with meridional ocean transports (e.g., Griesel and Maqueda 2006) via geostrophy and basin geometry such that they are expressed as relatively narrow currents, for example, the North Atlantic Current from the south, and the East Greenland Current from the north.

Dynamic height is influenced by both temperature and salinity in much of our domain. Exceptions are the subtropical North Atlantic Ocean, where temperature dominates, and the Arctic Ocean and far northern Pacific and Atlantic Oceans, where salinity dominates. Figure 2 also shows that the North Pacific Ocean contains more freshwater than the Arctic Ocean. Surface salinity in the Arctic Ocean is quite fresh, but this freshwater layer extends only to \( \sim 300 \text{ m} \), below which resides much saltier water of Atlantic Ocean origins. Another interesting feature in Fig. 2 is the steep change in all quantities across Bering Strait. Higher DH in the North Pacific Ocean is believed to drive year-round northward fluxes of volume, heat, and freshwater through Bering Strait, modulated by local winds (most recently, Woodgate et al. 2005).

Figure 3 shows the horizontal distribution of these quantities. A variety of gridded and ungridded data are used in this figure, described in detail in section 2. Only data from areas with depths of at least 1000 m are plotted. Figure 3 shows that the North Atlantic DH maximum in Fig. 2 actually contains a northwest-southeast gradient, along which the North Atlantic Current flows. A weaker east-west DH gradient is also evident in the Nordic Seas, driving flow into the Arctic Ocean. Note that strong slope currents along Norway and Greenland do not appear in this deepwater view.

The Eurasian Basin of the Arctic Ocean shares many properties with the inflowing waters from the Nordic Seas, that is, relatively salty and warm, with a low DH. On the other hand, the Canadian Basin of the Arctic Ocean contains a small DH maximum within the relatively fresh Beaufort Gyre. In the North Pacific Ocean, properties are more zonal than in the North Atlantic Ocean. Nonetheless, DH contours suggest a northwestward geostrophic flow south of Alaska, which contributes to northward flow through the Aleutian Islands and eventually through Bering Strait.

Figure 3d shows the average buoyancy frequency \( \langle N \rangle \) over 0-1000-m depth, that is,
where $g$ is gravitational acceleration and $\sigma_0^{Xm}$ is potential density at $X$ m depth, referenced to the surface. The northern seas are relatively poorly stratified, especially in the Nordic Seas and the Arctic Ocean below its ~300 m thick surface fresh layer. This allows for a strong barotropic signal, meaning that steric sea level gradients may be less important for diagnosing the surface geostrophic ocean circulation in these regions, relative to lower-latitude oceans. Note, however, that Proshutinsky et al. (2004) found strong correspondence in the arctic seas between steric and total sea level trends in a numerical model. Nonetheless, Fig. 3d serves to remind us that steric sea level is only part of the sea level puzzle in the arctic seas.

4. Regional time series

The following analysis focuses on change, and thus we compute steric anomalies relative to a long-term mean. For the WOD01 data, the anomaly is formed by subtracting at each profile position the interpolated
mean value ($T_{ref}$, $S_{ref}$, $P$) from the gridded World Ocean Atlas 2001 (WOA01; Boyer et al. 2002; Stephens et al. 2002). For the EWG data in the Arctic Ocean, we formed a record-mean DH field from the individual years and used this, in combination with PHC between 200 and 1000 m, as the mean.

a. The Arctic Ocean

Figure 4 shows dynamic height changes over 0–200 m depth in the central Arctic Ocean during 1950–90, averaged over the colored grid boxes in Fig. 3a (i.e., where the ocean is at least 1000 m deep). The trends are nearly identical if the shallower gray boxes in Fig. 3a are also included. Dynamic height generally decreased during 1950–90, almost certainly because of increasing salinity since the thermal contribution to density is small at cold temperatures (Figs. 2 and 3). (A more definitive analysis could be performed using annual temperature and salinity fields, but as discussed in section 2, these are not yet publicly available.) A linear least squares regression line through the entire time series has a slope of $-0.19 \pm 0.07$ cm decade$^{-1}$ (mean ±1.96 × standard deviation of the slope, i.e., 95% confidence), implying a total DH decrease of 0.76 ± 0.28 cm over 40 yr. This central arctic DH decrease confirms the modeling results of Proshutinsky et al. (2004) and is opposite to the freshening and DH rise observed in recent decades on the Siberian shelves (Proshutinsky et al. 2004; Steele and Ermold 2004). Arctic Ocean DH also demonstrates significant decadal variability. In fact, there seems to be an approximately 15-yr time scale during which DH slowly rises, then discharges over ~5 yr. (Unfortunately, the time series is too short for this to be rigorously tested.) Since Arctic Ocean DH is controlled by freshwater (DH ≅ DH$_{fw}$; see Fig. 2), the 15-yr rises represent a freshwater “charging” of the arctic “capacitor,” similar to the concept of a Beaufort Gyre flywheel described by Proshutinsky et al. (2002). One charging period during 1958–72 includes the start of the GSA in the late 1960s (Dickson et al. 1988), when sea ice is thought to have streamed out of the Arctic Ocean and melted in the North Atlantic Ocean. This could be a response to changing winds associated with the North Atlantic Oscillation or its close relative, the Arctic Oscillation (AO; Thompson and Wallace 1998), although this connection is not robust for all time periods (Koberle and Gerdes 2003; Tremblay 2001). A weak DH drop at this
time (marked “GSA-ice?” in Fig. 4) is consistent with the salinification expected from enhanced Arctic Ocean sea ice growth during the winter following enhanced GSA sea ice export. Model simulations by Rothrock and Zhang (2005) produce a single year of enhanced ice export in 1968, and a subsequent single year of enhanced Arctic Ocean ice growth in 1969. However, total sea ice volume does not completely recover to pre-GSA levels, owing mostly to slightly warmer surface air temperatures after the GSA.

A much larger DH drop of about 1 cm ± 40 cm of freshwater is also evident during the mid-1970s, and is marked “GSA-ocn” in Fig. 4. This is the time that Swift et al. (2005) found marked Arctic Ocean salinification, attributed by them to increased ice growth after the 1968 GSA. However, the model simulation cited previously (Rothrock and Zhang 2005) suggests that enhanced ice growth from the GSA ice export event happened within one year and did not last through the 1970s. Thus another explanation must be found. The Arctic Ocean freshwater layer (0–200-m depth) has a residence time of ~10 yr (Serreze et al. 2006), with the bulk of it residing some distance away from Fram Strait in the Canadian Basin (Proshutinsky et al. 2002; Serreze et al. 2006). Thus, it seems probable that continued salinification through the 1970s was caused by a draining of liquid (ocean) freshwater from the Arctic Ocean, in response to winds associated with an increasing AO index. It is not clear whether the separation in Fig. 4 of GSA-ice and GSA-ocean by a slight DH increase during the early 1970s is real or an artifact of noisy data. The AO index does not monotonically increase over the late 1960s through the 1970s, which might delay liquid freshwater export relative to sea ice export, as winds shift annually. On the other hand, if the unusually large +1.3 cm annual DH anomaly in 1973 is eliminated, Arctic Ocean running-mean DH is nearly monotonic downward from the mid-1960s through the 1970s.

Thus we propose that the 1970s DH drop in Fig. 4 is a response to liquid (ocean) freshwater export from the Arctic Ocean (probably balanced to some extent below 300 m by increased salty Atlantic water inflow). Such a delayed response in ocean freshwater export has been noted in model simulations (Haak et al. 2003; Karcher et al. 2005). Our main idea here is that the AO-associated wind shift of the late 1960s forced an immediate Arctic Ocean sea ice export as well as a delayed and sustained liquid (ocean) freshwater export.

b. The North Atlantic Ocean

Figures 5 and 6 show DH, DH$_T$, and DH$_S$ trends in the North Atlantic Ocean during 1950–2000, divided into the subpolar gyre (Fig. 5: between 50°N and the Greenland–Iceland–Scotland Ridge) and the Nordic Seas (Fig. 6: north of this ridge to 80°N). Anomalies were computed by subtracting interpolated monthly mean WOA01 values from WOD01 profile data, which were then averaged into 200-km bins to avoid spatial bias. The spatial coverage within each region of annual mean 200-km bins (gray dots in the figures) can be poor, which is why for most of our analysis, we focus on the 5-yr running-mean binned data (green dots in the figures). Collecting data over 5-yr bins generally provides 90% or greater areal coverage. Thus no spatial interpolation or smoothing was performed.

The linear DH trend over 1950–2000 in the North Atlantic subpolar gyre is not significantly different from zero, although the Nordic Seas DH trend is a positive 0.56 ± 0.23 cm decade$^{-1}$. Further, there are obvious variations on decadal and longer time scales, especially in the temperature (DH$_T$) and salinity (DH$_S$) components. During the years 1965–90, both the Arctic Oscillation index and DH$_S$ rise in the subpolar and Nordic seas. This North Atlantic freshening was also found by Curry and Mauritzen (2005), although theirs is sustained (albeit at a slower rate) through the 1980s, while ours stops rising by 1980. The largest freshening occurs during the 1970s, starting in the subpolar gyre in 1968 and the Nordic Seas a few years later. This delay was proposed by Curry and Mauritzen (2005) to result from the GSA’s sea ice (and some resulting meltwater) flowing quickly down the East Greenland Current into the subpolar gyre, and only slowly diffusing eastward across the Nordic Seas.

The subpolar gyre freshened but also cooled during the GSA, possibly in response to cold Arctic water and sea ice (Fig. 5; see also Levitus et al. 2005). The cooling dominates the density change, creating a small net DH decrease of 1.40 ± 0.63 cm over 1965–90. Cooling is not so important to steric height in the Nordic Seas (Fig. 6), however, since density is controlled much more by salinity than by temperature in these cold waters. Thus DH rose in the Nordic Seas by 3.78 ± 0.78 cm over 1965–90 in response to GSA freshening.

Satellite altimeter observations since the early 1990s show rising sea surface heights in much of the North Atlantic subpolar gyre (Cabanes et al. 2001; Hakkinen and Rhines 2004; Levinson 2005), a result also seen in our DH data (Fig. 5). The altimeter data also indicate a weak 1990s sea surface height increase in the Nordic Seas (F. Mertz, CLS/DOS, 2006, personal communication), which is not seen in our DH time series (Fig. 6).

c. The northeast Pacific Ocean

Figure 7 shows dynamic height trends for the Northeast Pacific Ocean, defined as the region bounded by
the North American coast to the east, 160°E to the west, Bering Strait to the north, and a line roughly between the southern tip of Kamchatka, Russia, and northern California (see following section). This is the region containing the inflows to the Arctic Ocean and shows spatially uniform trends in dynamic height that are opposite to nearby regions in the Pacific Ocean (see section 5a).

The northeast Pacific Ocean shows a positive DH trend over the entire record of 1.01 ± 0.21 cm decade⁻¹, with contributions from both warming and freshening (see also Levitus et al. 2005). A major climate index for the North Pacific Ocean is the Pacific decadal oscillation (PDO), which is the first principle component in an empirical orthogonal function formed using sea surface temperature fields (Mantua et al. 1997). A major transition occurred in the mid-1970s from a negative to positive index, which brought warmer water to the North Pacific Ocean. Before 1960 the PDO index was weak and variable, while since 1990 the dominant climate forcing is best described by the second principle component (Bond et al. 2003), known as the Victoria Pattern (VP). This pattern is anticorrelated with warmer northeast Pacific Ocean temperatures, that is, a negative index implies a warming ocean. The net result of these climate forcings should be a DH rise from the mid-1960s through the mid-1990s, which is what Fig. 7 shows. Freshening also occurs during this period, although not as strongly. The net result is a DH rise of 4.93 ± 0.80 cm over 1970–95. Note that
the Victoria Pattern reversal in the late 1990s should result in a cooling, which would imply a DH fall if salinity changes are small. In fact, altimetric sea surface height does show a decrease of ~1 cm in this region over 1993–2004 (Levinson 2005).

d. Domain-average changes

Table 1 shows a summary of regional steric height changes over two time periods, as described in sections 4a–c. Most height changes are consistent in sign and even magnitude for both time periods. The major exception is the North Atlantic subpolar gyre, which has a significant DH decrease over 1965–90 but not over 1950–2000.

Table 1 also shows the area-weighted total steric height changes, with and without including the northeast Pacific Ocean. These total height changes indicate that DH increased overall when including the northeast Pacific Ocean, mostly as a result of a DH_S (i.e., freshwater) increase. This could be from increased net precipitation less evaporation at high latitudes in response to an accelerated hydrologic cycle (Curry et al. 2003), from increased net terrestrial and/or sea ice melting (Antonov et al. 2002; Wadhams and Munk 2004), or from net ocean freshwater transport into these high latitudes. While local DH_T changes may be large, the overall effect when including the Northeast Pacific Ocean is small, consistent with its small influence on density at cold temperature.

On the other hand, total steric height change is insignificant when considering only the Arctic and North Atlantic Oceans. In this domain, net freshening (DH_S increase) is essentially balanced by net cooling (DH_T decrease). This is consistent with sea ice import into the North Atlantic Ocean, which cools and freshens as a result. Thus the Arctic/Atlantic domain experiences net changes...
cooling and net freshening of the ocean (but not necessarily of the total ice plus ocean system).

5. Trends over the late 1960s through the early 1990s

a. Gridded fields

Figure 8 shows how trends of dynamic height and its temperature/salinity components $DH_T$ and $DH_S$ vary spatially during the mid-1960s through early 1990s. This is the period when climate indices in all regions strongly varied (Figs. 5, 6, and 7). Although some noise is evident, spatial trends are remarkably coherent over the broad regions discussed in the previous sections: the central Arctic Ocean, the Nordic Seas, the North Atlantic subpolar gyre, and the northeast Pacific Ocean. The Arctic Ocean shows particularly smooth trends because the EWG data are only available as gridded, smoothed fields (section 3). In other regions, hydrographic data from WOD01 have been averaged into 200 km $\times$ 200 km bins (but not smoothed). Dynamic height trends from the Nordic Seas extend with the same sign into the Eurasian Basin of the Arctic Ocean. There is also some indication of northeast Pacific Ocean trends extending into the Chukchi Sea.

The drop during 1965–90 in Arctic Ocean $DH \cong DH_S$ occurs mainly in the eastern Eurasian Basin and southern Makarov Basin, extending across the North Pole toward Greenland. This is the region that showed significant freshwater changes during the AO index rise in the early 1990s (e.g., Morison et al. 2000). Freshwater decreases as the Atlantic/Pacific front sweeps around cyclonically, bringing a more salty Atlantic influence to the central Arctic Ocean. Karcher et al. (2005) show in a model study how this might force freshwater out of the Arctic Ocean and into the North Atlantic Ocean.

In fact, Fig. 8 shows that the North Atlantic Ocean
freshens during this time. Note that historical data coverage within the East Greenland Current, the Canadian Archipelago, and Baffin Bay is very poor. This means that we cannot explicitly follow the freshwater outflow, only its effects on the deep basins immediately adjacent. There is also a widespread cooling in the subpolar gyre, possibly caused in part by melting sea ice and mixing of cold melt water.

The North Pacific Ocean shows two distinct regions in Fig. 8: a fairly uniformly rising DH area in the northeast and a noisier falling DH area to the south and west. As shown in Fig. 7, the northeast Pacific Ocean DH increase arises from both freshening and warming over the 50-yr record (Levitus et al. 2005) but is more strongly forced by warming over the 1970s–90s. The similar spatial pattern of thermosteric and halosteric change over the North Pacific Ocean seen in Fig. 8 contradicts the finding of Overland et al. (1999), who found nearly orthogonal temperature and salinity differences in response to the mid-1970s PDO shift. However, our analysis does not extend quite as far south as theirs did.

**b. Sector zonal means**

Figure 9 shows sector zonal mean changes in DH, $DH_s$ and $DH_f$ between the late 1960s and the early 1990s, using a format similar to Fig. 2. The dynamic height difference between the North Pole and Nordic Seas decreased by 32% (8 cm relative to the 25-cm mean), owing mostly to southward freshwater transfer. Similarly, the DH gradient between the Nordic Seas and the North Atlantic subpolar gyre declined by 18% (8 cm relative to the 45-cm mean), owing in part to cooling of the subpolar gyre. The net effect of these changes is a weakening, or “filling in” of the Nordic Seas DH minimum, indicated in Fig. 9 by shading. This leads to a decrease in the pressure gradients associated with flows from the south (the MOC) and from the north (Fram Strait and the AFT), possibly implying a slowing of these ocean transports during this period.

**Table 1. Steric height changes (cm) over the Northern Seas for 1950–2000 and for 1965–90 (North Atlantic subpolar gyre, Nordic Seas, and Arctic Ocean), and 1970–95 (northeast Pacific Ocean). Arctic Ocean DH changes are not available and thus are assumed to be identical to the observed DH changes (salinity dominates the density at cold temperature). Similarly, Arctic Ocean DH changes are assumed to be zero, with half the uncertainty of the (slightly warmer) Nordic Seas. Also, Arctic Ocean trends from 1950 to 1990 are assumed to continue through 2000. Changes in the total North Pacific Ocean (north of 40°N) are of similar sign to those in the northeast Pacific Ocean shown here, but generally smaller owing to partly compensating trends to the southwest (Fig. 8). Uncertainty in each region is 1.96 × standard deviation of the slope (bold indicates 95% significance). Uncertainty in the combined regions (final two rows) is computed from the sum of individual regions’ slope variances and the variance of each region’s slopes relative to the total mean slope.**

<table>
<thead>
<tr>
<th>Region</th>
<th>$DH_s + DH_f = DH$</th>
<th>$DH_s + DH_f = DH$</th>
</tr>
</thead>
<tbody>
<tr>
<td>North Atlantic gyre</td>
<td>$5.1 \pm 1.5$</td>
<td>$4.7 \pm 1.5$</td>
</tr>
<tr>
<td>Nordic Seas</td>
<td>$3.5 \pm 1.1$</td>
<td>$3.8 \pm 1.2$</td>
</tr>
<tr>
<td>Arctic Ocean</td>
<td>$-1.0 \pm 0.4$</td>
<td>$-0.7 \pm 0.3$</td>
</tr>
<tr>
<td>Northeast Pacific Ocean</td>
<td>$2.7 \pm 0.6$</td>
<td>$1.7 \pm 0.7$</td>
</tr>
<tr>
<td>Weighted total*</td>
<td>$2.3 \pm 1.1$</td>
<td>$2.0 \pm 1.2$</td>
</tr>
<tr>
<td>Weighted total (no Pacific)</td>
<td>$-0.2 \pm 0.9$</td>
<td>$-0.5 \pm 1.7$</td>
</tr>
</tbody>
</table>

* Weighted by area of maximum data coverage in each region: North Atlantic subpolar gyre = $4.20 \times 10^6$ km$^2$, Nordic Seas = $1.52 \times 10^6$ km$^2$, Arctic Ocean = $4.66 \times 10^6$ km$^2$, and northeast Pacific Ocean = $5.48 \times 10^6$ km$^2$. 
scale of about 25 yr, implying a total feedback period of 50 yr. Thus we propose that even as the MOC gained strength during 1965–90, cold and fresh arctic outflows were accumulating in the North Atlantic, eventually decreasing the steric height gradients and decelerating the MOC.

Of course, steric height is only part of the total sea surface height (SSH). Mass inputs (not included in this study) can also change SSH, although some studies have shown fairly good correspondence between large-scale patterns of SSH and DH trends (Cabanes et al. 2001; Proshutinsky et al. 2004), while others show a not unexpected decrease in the correlation of SSH and DH variability with increasing latitude (Fig. 3d; Guinehut et al. 2006).

Figure 9 shows rising Pacific Ocean DH trends, also sustained over the total 1950–2000 period (although recently reversing). This should accelerate, not deceler-
ate, the pressure gradient forcing for the AFT. This acceleration might have occurred not via Fram Strait, but rather through the Canadian Arctic Archipelago and Baffin Bay to the Labrador Sea, which shows some DH increase in Fig. 8. Some support for this idea is provided by Zweng and Muenchow (2006), who analyzed Canadian data not yet incorporated into WOD01. They found a freshening in the southward-flowing Baffin Island current during the latter half of the twentieth century, although the severe summer bias in their data introduced a significant level of uncertainty. The data thus suggest that during 1965–95, the Fram Strait branch of the AFT may have slowed even as the Canadian Arctic branch accelerated.

c. Volume fluxes between the Arctic and North Atlantic Oceans

Can the changes observed in the North Atlantic Ocean during the Great Salinity Anomaly of the 1970s be completely explained by sea ice and liquid (ocean) fluxes from the Arctic Ocean? Freshening in the subpolar gyre amounted to \( \sim 200 \) cm (Fig. 5), or \( \sim 8400 \) km\(^3\) when multiplied by an area of \( 4.2 \times 10^6 \) km\(^2\). For the Nordic Seas, Fig. 6 indicates a freshening of \( \sim 160 \) cm, or \( \sim 2400 \) km\(^3\) when multiplied by an area of \( 1.5 \times 10^6 \) km\(^2\). Thus the North Atlantic Ocean freshened by \( \sim 11000 \) km\(^3\) over 10 yr. This is far more than can be explained by the mid-1970s Arctic Ocean freshwater loss.
of about 2000 km$^3$ (Fig. 4; Arctic Ocean area = 4.7 × 10$^6$ km$^2$). Further, sea ice export is not likely to explain the residual of ~9000 km$^3$, since this would require a positive annual export anomaly of about 900 km$^3$ (i.e., 30%–50% of the mean) over 10 yr, a value far above that seen in model simulations (Karcher et al. 2005; Rothrock and Zhang 2005). Thus other factors must play a major role. A similar calculation for heat transports could also be performed, although it would be more complex, owing to the more complex relationship between thermosteric height and heat content [Eq. (4)]. A full accounting of freshwater and heat budgets is beyond the scope of the present study.

6. Conclusions

Our main conclusions are as follows:

1) Net steric height increase over our domain: Steric height over the Northern Seas increased during 1950–2000, mostly from freshening (Table 1). While local temperature changes were also important, the net effect over our domain was small.

2) Filling of the Nordic Seas DH minimum: Steric height plots centered on the Arctic Ocean (Figs. 2 and 3) clearly show two major ocean pathways: the AFT and the upper limb of the MOC. Our analysis indicates that the steric height minimum that in part drives both of these filled in over the North Atlantic and Arctic Oceans during 1965–90 (Fig. 9), possibly indicating a slowing of these circulations. A DH rise in the northeast Pacific Ocean during this time seemingly contradicts this result, possibly resolved by an accelerated flow through the Canadian sector of the Arctic and North Atlantic Oceans. More recent data indicate that some (but not all) areas show reversed DH trends relative to the previous 25 yr, highlighting the strong decadal-scale variability in these time series.

3) GSA-ice versus GSA-ocean: Anomalous Arctic Ocean sea ice export in the late 1960s (GSA-ice; Fig. 4) contributed to a freshening of the Nordic Seas (DH rise; Fig. 6), a freshening and cooling in the North Atlantic subpolar gyre (DH drop; Fig. 5), and a small salinification of the Arctic Ocean as sea ice growth rebounded in the following few years (DH drop). We also find a stronger Arctic Ocean salinification during the mid-1970s as liquid freshwater in the upper few hundred meters was exported southward (GSA-ocean; Fig. 4), which contributed to a sustained freshening of the North Atlantic Ocean. However, the freshening in our analysis levels off by the 1980s, about 10 yr earlier than in the study by Curry and Mauritzen (2005). The pattern of Arctic Ocean DH change over the period 1965–90 (Fig. 8) indicates a cyclonic shift in the front separating Pacific influence (high DH) from Atlantic influence (lower DH), similar to that observed during the 1990s. Our analysis of Arctic Ocean dynamic height fields supports the capacitor model of freshwater storage and release proposed by Proshutinsky et al. (2002), where we find a storage period of ~15 yr and a release period of ~5 yr.

4) Insufficient Arctic Ocean sources for observed GSA freshening: A freshwater volume analysis indicates that the 1970s North Atlantic Ocean freshening was much larger than can be explained by the sum of sea ice and liquid (ocean) freshwater export from the Arctic Ocean (section 5c). Thus other (as yet undetermined) sources must have played a major role.

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