



Shear dispersion and delayed propagation of temperature anomalies along the Norwegian Atlantic Slope Current

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ABSTRACT

Using satellite altimetric sea surface height (ADT) data, we search for propagation of hydrographic anomalies along the Norwegian Atlantic Slope Current (NwASC) from the Svinøy section in the south to the Fram Strait in the north. Our analyses indicate that ADT anomalies, related to low-frequency temperature variations, propagate downstream with speeds of about 2 cm s^{-1} . Notably, this speed is nearly an order of magnitude slower than the speed of the NwASC, which in agreement with previously estimated propagation speeds of hydrographic anomalies along the flow. A conceptual tracer advection model, consisting of a thin current core interacting with an adjacent slow moving reservoir, is introduced to examine temperature anomaly propagation along the NwASC. It is shown that shear dispersion effects, resulting from cross-stream eddy mixing and velocity shear, can qualitatively explain the observed delayed propagation of hydrographic anomalies: low-frequency temperature anomalies move downstream with an effective velocity that corresponds to a mean velocity across the entire Atlantic Water layer, rather than the speed of Norwegian Atlantic Slope Current.

Keywords: Nordic Seas, propagation ocean heat anomalies towards the Arctic Ocean, satellite altimetry, shear dispersion and delayed propagation

1. Introduction

Heat transport associated with the poleward flow of Atlantic Water through the Nordic Seas plays a critical role for the climatic conditions in the Arctic Ocean, influencing the stratification and the sea ice cover (Rudels et al., 2005; Rudels, 2012; Carmack et al., 2015). Observations show that temperature anomalies of the Atlantic Water in the Nordic Seas propagate poleward with speeds of a few cm s^{-1} , and with a delay of a couple of years affect the Arctic Ocean sea ice extent and hydrography (Polyakov et al., 2005; Årthun et al., 2012; Carmack et al., 2015). Thus, observations of Atlantic Water temperature anomalies in the Nordic Seas can potentially be used for predicting the evolution of the Arctic sea ice cover a few years in advance (Årthun et al., 2012; Årthun et al., 2017).

Hydrographic anomalies carried by the Atlantic Water trace essentially the time-mean circulation in the Nordic Seas, reaching the Arctic Ocean via the Barents Sea and the Fram Strait as well as partly returning to the North Atlantic via the East Greenland Current (Mauritzen, 1996; Furevik, 2000; Chepurin

and Carton, 2012). In the eastern Nordic Seas, Atlantic Water is transported poleward in two main current branches (Mauritzen, 1996; Mork and Skagseth, 2010): The Norwegian Atlantic Slope Current (NwASC), which is a mainly barotropic flow tracing the continental slope, and the Norwegian Atlantic Frontal Current (NwAFC), which is a chiefly baroclinic frontal current; see Fig. 1. The bulk of the volume transport occurs in the NwASC, which is also the branch where the highest velocities are found. Here, the current speed is generally about 20 cm s^{-1} , but where the continental slope is steepest it can exceed 50 cm s^{-1} (Mork and Skagseth, 2010).

Propagation of heat/salinity anomalies from the subpolar North Atlantic, across the Greenland–Scotland Ridge, and further through the Nordic Seas toward the Arctic Ocean is a robust observational feature found in numerous investigations, including analyses of hydrographic data (Furevik, 2001; Sundby and Drinkwater, 2007; Skagseth et al., 2008; Chafik et al., 2015; Yashayaev and Seidov, 2017), satellite-based SST data (Furevik, 2000; Chepurin and Carton, 2012) and surface drifters (Koszalka et al., 2013). A common method to infer propagation speeds from hydrographic stations is to identify anomaly events and examine how they emerge with some delay at downstream locations

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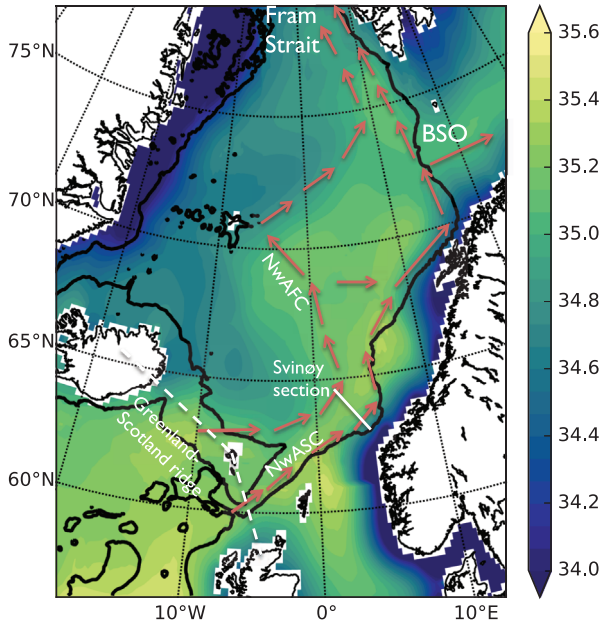


Fig. 1. Map of the Nordic Seas with surface salinity in shading (Zweng et al., 2013). The black contour is the 700 m depth contour. Indicated are the Fram Strait, Barents Sea opening (BSO), the Svinøy section, and parts of the Greenland–Scotland Ridge. In red are also indicated rough pathways of the NwAFC and the Norwegian Atlantic Slope Current (NwASC) and its fractionation at the BSO.

(Furevik, 2001; Sundby and Drinkwater, 2007; Yashayaev and Seidov, 2017). A related method for analysing propagation uses temperature or salinity correlations between two locations along the flow: The time lag for which the correlation reaches its maximum, is interpreted as a mean transit or advection time between the points (Jeffress and Haine, 2014). Using this approach, Skagseth et al. (2008) analyse mooring records of temperature in the Svinøy section and in the Barents Sea Opening (BSO) and find a maximum in correlation at a lag (or transit time) of two years, corresponding to a propagation speed of about 2 cm s^{-1} . Chepurin and Carton (2012) use lagged correlation analyses of satellite sea surface temperatures (SSTs) that trace propagation along a path from the North Atlantic, into and anti-clockwise around the Nordic Seas. Along this path, they estimate a transit time from the Svinøy section to the Fram Strait of 3 years, corresponding to a speed of about 3 cm s^{-1} . Using complex empirical orthogonal functions to analyse propagation of Atlantic Water anomalies, Årthun and Eldevik (2016) and Årthun et al. (2017) find propagation speed of about 2 cm s^{-1} both in an ocean model simulation and in observations of the sea surface temperature.

An intriguing and partly unresolved issue is the remarkable slow poleward propagation of Atlantic Water temperature/salinity anomalies¹ that has been inferred from observations and modelling studies. The studies mentioned above yield poleward propagation speeds of a few cm s^{-1} , which are roughly an order

of magnitude lower than the speeds in the cores of the Norwegian Atlantic Slope and Frontal Currents. Several mechanisms have been invoked to explain delayed propagation of hydrographic anomalies along current cores. Proposed mechanisms include the beta effect and effects related to air–sea heat exchange (e.g. Hansen and Bezdek, 1996; Krahmann et al., 2001), the interplay between velocity anomalies and along-stream hydrographic gradients (Sundby and Drinkwater, 2007), Lagrangian dispersion and cross-stream mixing (e.g. Koszalka et al., 2013; Jeffress and Haine, 2014). However, a firmly established explanation of the phenomena appears to be lacking.

In this paper, we focus on propagation of hydrographic anomalies along the NwASC. Specifically, we examine if anomaly propagation can be detected in satellite altimetric sea surface height data. Previous investigations have been based on mainly hydrographic and sea surface temperature data, but the relatively high temporal and spatial resolution of the satellite altimetry may provide additional information. We perform a lagged correlation analysis of satellite altimetric data along the Slope Current, exploiting some results on the relation between lagged correlations and propagation speeds presented by Jeffress and Haine (2014). Further, we apply a simple tracer advection model, known as the ‘leaky pipe’ model (Nilsson, 2000; Jeffress and Haine, 2014), to examine how purely oceanic processes can affect the propagation of hydrographic anomalies along the NwASC. The analysis shows that shear dispersion effects (Taylor, 1953; Young and Jones, 1991) can qualitatively explain the delayed propagation of hydrographic anomalies.

The paper is structured as follows: Section 2 describes the satellite altimetric sea surface height (ADT) data and introduces some analysis methods. Section 3 presents the analysis of the ADT data and Section 4 examines theoretically how shear dispersion effects can slow down the propagation of hydrographic anomalies. Section 5 gives a summary and discussion of the main findings.

2. Data and method

The altimeter data are produced by Ssalto/Duacs and distributed by Aviso, with support from Cnes (<http://www.aviso.altimetry.fr/duacs/>). The product used here is the Absolute Dynamic Topography (ADT) and depicts the dynamic sea surface height; a geostrophic approach on the ADT gives the surface absolute geostrophic current. The ADT has thus been corrected for the inverted barometer effect and tides. We use the global gridded data on a $1/4^\circ$ spherical grid provided as daily fields from 1 January 1993 to 31 December 2014.

2.1. Method

We average the daily ADT into monthly means. This approach ensures that short-time variability, for example due to weather

events such as the passing of single low pressure systems (Richter et al., 2009), is filtered out. We then remove the seasonal cycle by constructing a monthly climatology from the whole 21-year period and subtracting it from every monthly value. In order to track the current core, we interpolate the ADT to a specific depth contour along the continental shelf. This method of tracking the current is possible since the NwASC closely follow the bathymetry (Helland-Hansen and Nansen, 1909; Jakobsen et al., 2003; Sjøland et al., 2008; Broomé and Nilsson, 2016). The depth contour of choice is here the 700 m contour, which lies in centre of the NwASC (Fig. 1). Chafik et al. (2015) further discuss the NwASC's relation to specific depth contours and find the 700 m depth contour to best track the centre of the current. The sensitivity of the analysis to the choice of depth contour will be discussed in Section 3. The ADT along 700 m depth contour is shown in Figs. 2a and b.

To investigate possible propagation of anomalies in the NwASC, we perform a lagged correlation analysis of the deseasonalised ADT along the 700 m depth contour. A base point is chosen in the Svinøy section (Fig. 1), near 63°N 4°E, since the section has long been a key location for monitoring the Atlantic inflow to the Nordic Seas (see Mork and Skagseth, 2010, and references therein). We then cross-correlate all the data along the 700 m depth contour north of the Svinøy section, up to about 80°N, against the base point for positive/negative lags. In the lagged correlation analyses shown here, the linear trend has been removed. However, the important correlation results pertaining to propagation are qualitative similar whether or not the linear trend is removed.

The cross-correlation $\rho_{xy}(\tau)$ between two time ADT series along the NwASC x_t and y_t , $t = 1, 2, \dots, n$, for a certain time lag τ is formulated as

$$\rho_{xy}(\tau) = \frac{1}{n - \tau} \frac{1}{\sigma_x \sigma_y} \sum_{k=1}^{n-\tau} x_k y_{k+\tau}, \quad (1)$$

where σ_x and σ_y are the standard deviations of x and y , respectively.

2.2. ADT and connection to hydrographic properties

The transport of buoyancy, or heat, along the NwASC is the focus of this study. However, it might not be immediately apparent that the ADT reflects hydrographic properties in this sense. Hence, the connection between the sea surface height observed by satellites and the steric height of heat and freshwater anomalies should first be established.

The satellite ADT, say η , can be divided into two components controlled by different effects (Broomé and Nilsson, 2016)

$$\eta = \eta_B + \eta_S, \quad (2)$$

Here, η_B reflects the dynamic pressure at the bottom, which is controlled by the mass anomaly in the water column, whereas η_S is steric height signal resulting from depth-integrated buoyancy anomalies. Using data from the Svinøy section, Richter and Maus (2011) examine anomalies in satellite-derived ADT and in steric height computed from hydrographic observations, and find that these anomalies on inter-annual timescales are well correlated. In the Atlantic Water, positive anomalies in thermosteric height normally coincide with negative anomalies in halosteric height. However, the thermosteric signal has a higher amplitude and gives the dominating contribution to the steric height, both in variability and in the time mean. Thus, along the NwASC a positive steric height anomaly corresponds usually to a positive anomalous heat content (Richter and Maus, 2011).

3. Northward propagation revealed in the satellite altimetry

3.1. Lagged cross-correlation analysis

Figure 2a shows the monthly ADT interpolated to the 700 m depth contour for the period 1993–2015. The ADT generally decreases northward, mainly reflecting a decrease in steric height as the buoyancy diminishes poleward (Walin et al., 2004; Chafik et al., 2015). The exception is the two local maxima encountered around 64°N and 70°N, where the topographic slope/curvature is large and the core of the NwASC is shifted slightly offshore (Broomé and Nilsson, 2016).

Both the raw monthly and the deseasonalised ADT fields (Figs. 2a and b) exhibit an overall positive trend over the 22-year period, which is coincident with a positive regional trend in heat content (Skagseth and Mork, 2012) as well as in AW volume inflow (Hansen et al., 2015; Årthun et al., 2012). Further, Fig. 2b shows a few positive and negative inter-annual anomalies in the ADT along the NwASC. Focusing on the southern segment near 64°N, one can identify a strong positive ADT anomaly centred around 2003, which coincides with in situ measurements of high temperatures and salinities in the Svinøy section (see e.g. Mork and Skagseth, 2010; Richter and Maus, 2011). Conversely, the negative ADT anomaly around 1994/1995 coincide with a period of lower Atlantic Water temperatures in the Svinøy section. It can be noted that the NwASC heat transport variations in the Svinøy Section tend to be dominated by velocity rather than temperature anomalies, but that the temperature anomalies have longer time scales than the velocity anomalies (Orvik and Skagseth, 2005).

In Fig. 2b, however, there is no obvious tilt in the ADT that would indicate a northward propagation of anomalies. To provide further insight, a lagged cross-correlation (described in Section 2) is hence calculated and shown in Fig. 3. Here, there are two distinct features of the correlation pattern that merit consideration. The first is the highest correlation, which is found at zero lag. This essentially instantaneous correlation varies slightly northward along the NwASC, but generally decreases with

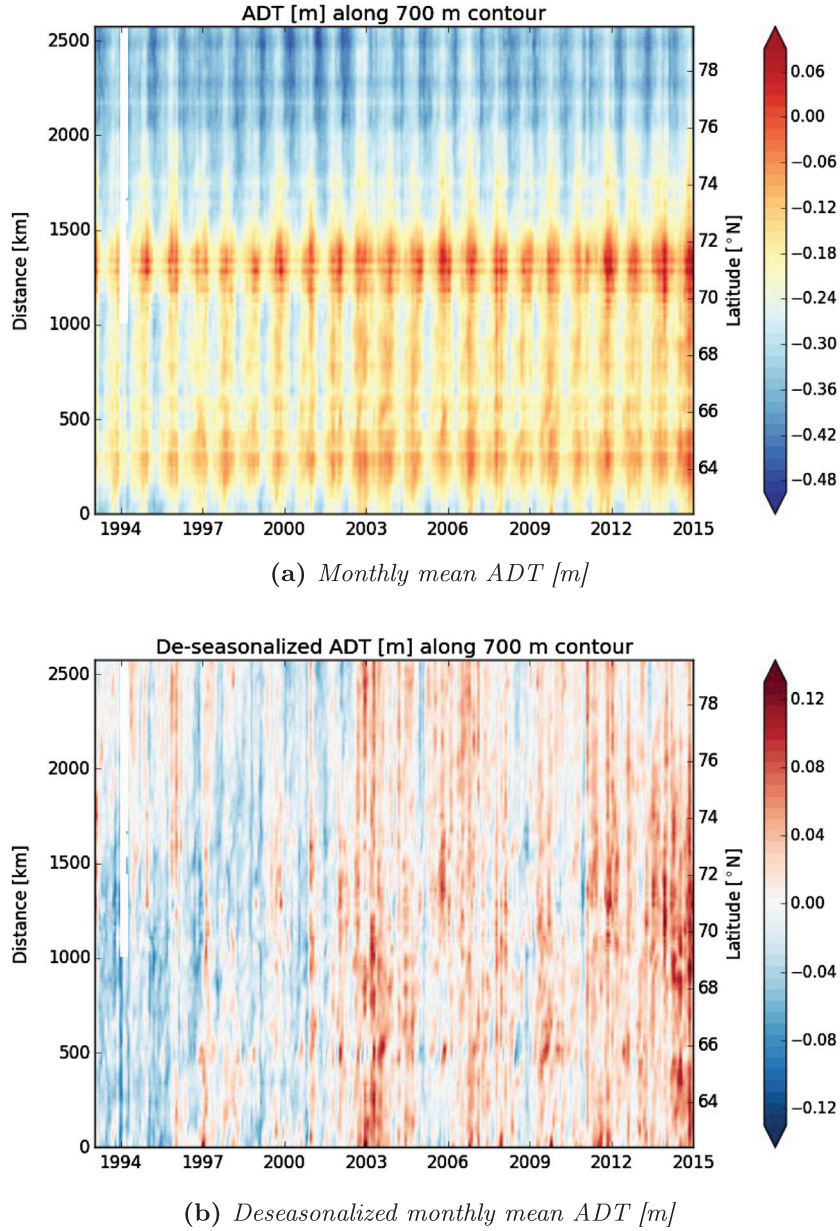


Fig. 2. (a) Monthly mean and (b) deseasonalised monthly mean ADT [m] interpolated to the 700 m depth contour along NwASC. Note that the y-axis, which is linear in latitude, is not necessarily linear in distance.

distance from the base point in the Svinøy section, from 1 – the correlation with the base point itself – to approximately 0.4 about 2500 km further north. The second pattern emerging is a band of elevated correlation at increasing positive lag with increasing distance from the Svinøy section.

To begin with, we focus on the nearly instantaneous high correlations in Fig. 3, which from an inspection of Fig. 2b can be connected to ADT variations that are fairly coherent along the NwASC. These variations, mainly forced by large-scale atmo-

spheric circulation patterns (Mork and Blindheim, 2000; Furevik, 2001; Richter et al., 2012), are associated with velocity increases and decreases of the topographically steered Slope Current (Skagseth et al., 2004) as well as with basin-scale sea level fluctuations. These fluctuations are predominantly barotropic, and their response time scales are set by the speed of barotropic Kelvin and shelf waves (Gill and Schumann, 1974). As these waves propagate from the Svinøy section to the Fram Strait within a couple of days, our monthly averaged ADT data

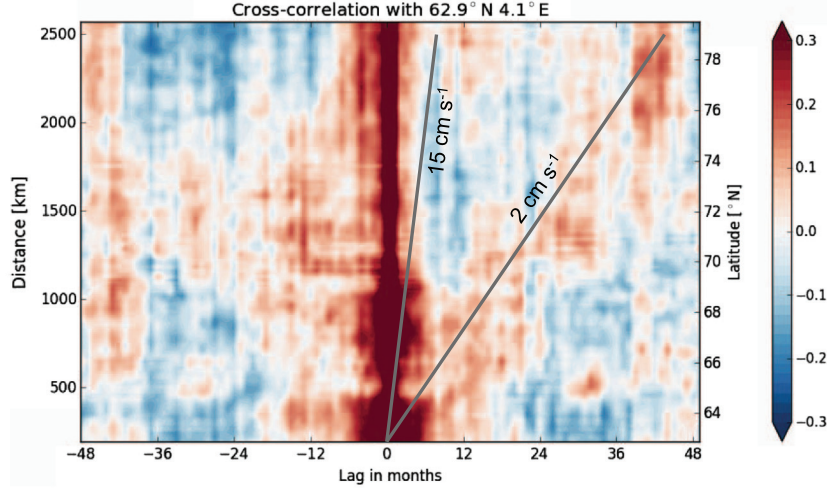


Fig. 3. Lagged cross-correlation of points along the 700 m depth contour (along NwASC) with the base point near $63^\circ\text{N } 4^\circ\text{E}$. The ADT has been deseasonalised prior to calculating the correlation. Following a straight line along the elevated correlation at increasing time lag gives an estimate of a propagation speed of about 2 cm s^{-1} . This can be compared to a mean current speed of 15 cm s^{-1} .

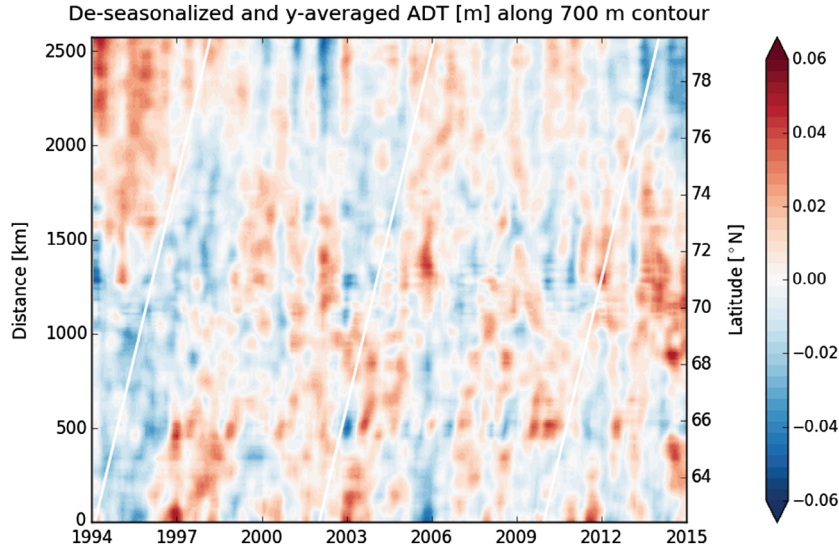


Fig. 4. The spatially varying ADT anomaly $\eta'(y, t)$ along the 700 m isobath in the NwASC; see Equations (3a,b). The anomaly $\eta'(y, t)$ is computed by subtracting the spatial mean from each individual month of the deseasonalised anomalies along the NwASC (Fig. 2b). This simple procedure highlights propagating anomalies and, as side effect, removes most of the linear trend present in Fig. 2b. The white lines represent a 2 cm s^{-1} propagation speed. Note that the anomaly $\eta'(y, t)$ shown here has been slightly smoothed with a 4-month running mean filter.

captures a quasi-equilibrium barotropic signal with a high coherency along the slope.

The second distinct pattern in Fig. 3, with elevated correlation along a tilted lag–distance segment, presumably represents poleward movements of ADT anomalies along the NwASC that are associated with steric height variations. Between Svinøy and Fram Strait, a distance of about 2500 km, the elevated corre-

lation has a maximum at a lag of about 40 months. Fitting a straight line through the elevated area of correlation gives an average propagation speed of about 2 cm s^{-1} , which is comparable to estimates of propagation speed of hydrographic anomalies of $2\text{--}3 \text{ cm s}^{-1}$ from other studies (see Furevik, 2000; Skagseth et al., 2008; Chepurin and Carton, 2012; Koszalka et al., 2013, and additional references in Section 1).

It is instructive to compare the two patterns with elevated downstream correlation emerging in Fig. 3 with lagged correlation analyses between moored current meters in the Svinøy section and the Barents Sea Opening presented by Skagseth et al. (2008); see their fig. 2.7. They use monthly filtered data and find that the velocity correlation between the two moorings is maximum at zero lag (around 0.4), whereas the temperature correlation reaches its maximum at a lag of about 2 years (around 0.8). This supports the notion that the correlation patterns in Fig. 3 reflect barotropic signals that are fairly coherent along the NwASC (Skagseth et al., 2004) as well as slowly moving steric-height anomalies.

3.2. Revealing anomaly propagation by suppressing barotropic variability

Although signs of propagation are visible in the lagged ATD correlations along the NwASC, there are as noted above no evident indications of propagation in the Hovmöller diagrams shown in Fig. 2. One reason could be that the slowly propagating signal is masked by stronger nearly coherent variations of the sea level along the NwASC. A straightforward way to suppress structures associated with long-range spatial coherence and/or high propagation speeds is to simply remove the mean ADT along the NwASC from each month in the time series. For this purpose, we decompose the ATD along the 700 m isobath as

$$\eta(y, t) = \bar{\eta}(t) + \eta'(y, t), \quad (3a)$$

where $\eta(y, t)$ is the de-seasonalised ADT along the NwASC, and we have introduced

$$\bar{\eta}(t) \equiv \frac{1}{y_2 - y_1} \int_{y_1}^{y_2} \eta(y, t) dy, \quad \eta'(y, t) \equiv \eta(y, t) - \bar{\eta}(t). \quad (3b)$$

Here, y is a coordinate along the isobath, and y_1/y_2 start/end point of the isobath segment. As defined here, $\eta'(y, t)$ is the spatially varying ADT anomaly along the NwASC. Due to strong topographic steering along the continental slope, we anticipate that the dynamic bottom pressure is nearly constant along the depth contours on monthly or longer time scales (Nøst and Isachsen, 2003; Broomé and Nilsson, 2016). Accordingly, we expect that $\bar{\eta}(t)$ is associated primarily with barotropic dynamics tied to bottom pressure variations, i.e. related to η_B ; see Equation (2).

Figure 4 shows the spatially varying ADT-anomaly component $\eta'(y, t)$, which is computed from the deseasonalised ADT anomalies (Fig. 2b); a 4 month running-mean filter has been used to smooth the data. Evidently, signs of propagation along the NwASC becomes more recognisable once the along-contour mean has been removed for each month in the ADT time series. The Hovmöller diagram in Fig. 4 indicates that some of the

inter-annual ADT anomalies, which are recorded in the south near Svinøy, migrate poleward with speeds of a few cm s^{-1} . By comparing with in situ observations in the Svinøy section (Mork and Skagseth, 2010; Richter and Maus, 2011), one can relate the low ADT anomalies emerging around 1994 and 2005 to recordings of anomalously low temperatures and steric height, whereas the high ADT anomalies emerging around 1997 and 2003 corresponds to anomalously high temperatures and steric height in the Atlantic Water. Figure 4 also indicates anomalies with shorter length scales and higher frequencies that appear to be generated along the continental slope, rather than being advected into the domain from the south. These features may be related disturbances of the Atlantic Water created locally by wind-forced upwelling and/or air-sea fluxes (Lien et al., 2014). Note also that the Hovmöller diagram indicates a range of propagation speeds, which is compatible with the broad band of elevated lagged correlations in Fig. 3.

3.3. Sensitivity analysis

Although the elevated lagged correlations along the NwASC are rather low, we have pointed to several features indicating that the signal is related to propagating steric height anomalies.

Additional motivation can be found by examining cross-correlations along alternative spatial paths in the region. If lagged correlations are computed on the 700 m isobath in the NwASC using a base point that is shifted westward from continental slope in the Svinøy section into the central Norwegian Sea, then the indications of propagation disappear. When the base point is kept in the Svinøy section but the ADT is interpolated to the 500 or 900 m depth contours instead (both following along the continental slope northward), the correlation pattern becomes similar to that in Fig. 3. However, the pattern disappears if deeper depth contours, located closer to the western limit of the Atlantic Water, are used (not shown). These sensitivity tests show that the elevated lagged correlation pattern is only found along paths in the Slope Current or within the Atlantic Water slightly offshore, thus corroborating that it reflects propagation of anomalies.

4. Mechanisms for slow propagation of hydrographic anomalies

Numerous studies show that air-sea fluxes are important for creating large-scale oceanic heat anomalies and that ocean-atmosphere interactions may partly influence the propagation of oceanic heat anomalies (see e.g. Hansen and Bezdek, 1996; Nilsson, 2000; Furevik, 2001; Krahmann et al., 2001). Here, we focus on purely oceanic mechanisms that can reduce the speed of hydrographic disturbances relative to the mean current in NwASC. If these disturbances would behave as tracers, they would be essentially advected by the mean flow in the current core move poleward with speeds of about $20\text{--}40 \text{ cm s}^{-1}$.

Furthermore, the Slope Current flows along topography with shallow water to the right, which implies that the phase speed of Kelvin waves and topographic waves adds to the mean current (Gill and Schumann, 1974; Yang and Pratt, 2013). Accordingly, barotropic wave disturbances will travel poleward much faster than the mean flow and so will also Kelvin wave like baroclinic disturbances. Thus, wave propagation effects are unlikely to explain the reduced velocity of the observed hydrographic anomalies along the NwASC.

However, even if the hydrographic anomalies behave essentially as passive tracers,² tending to move with the mean flow, effects of velocity shear and mixing in the cross-stream direction can reduce the downstream propagation. The underlying mechanism is essentially the classical shear dispersion effect highlighted by Taylor (1953). To appreciate the relevance of the shear dispersion effect in the present context, it is instructive to recapitulate a few features of the Atlantic Water (AW) circulation in the eastern Nordic Seas. Figure 1 shows that the Atlantic Water layer extends from the inner current core (NwASC), via a broad section with weak mean currents (Jakobsen et al., 2003; Mork and Skagseth, 2010), to the outer current core (NwAFC). Thus, there is significant horizontal velocity shear across the Atlantic Water layer. Furthermore, the time-mean temperature and salinity are essentially homogenous across the entire AW layer, indicating that heat and salt are vigorously mixed from the current cores into the region of weak mean flow (Isachsen et al., 2012). Thus, if anomalously warm or cold AW enters the NwASC from the south, the hydrographic anomaly will be advected poleward and simultaneously mixed in the cross-stream direction into the region of weak mean flow. This interplay between downstream advection and cross-stream mixing in the NwASC gives rise to a shear dispersion effect (Taylor, 1953; Young and Jones, 1991), which will be elaborated below.

Taking a Lagrangian perspective, Koszalka et al. (2013) provide an illuminating picture of ‘delayed propagation’ in terms of transit times along the NwASC. Examining how a cluster of surface drifters move downstream after being released in the Svinøy section, they note how some drifters remain in the current core, whereas a large fraction of them stray off into adjacent low-speed waters. Hence, the effective path of a Lagrangian parcel generally becomes longer and its average speed slower than that of the current core. This yields a probability distribution of Lagrangian transit times from the Svinøy section to a given downstream location. Based on the drifter data, Koszalka et al. (2013) find mean transit times comparable to the observationally inferred hydrographic anomaly speeds.

4.1. The ‘leaky pipe’ model

We now go onto examine an Eulerian tracer advection model that in a simple way illustrates Taylor’s shear dispersion effect and also connects to concepts of Lagrangian transit times. The model,

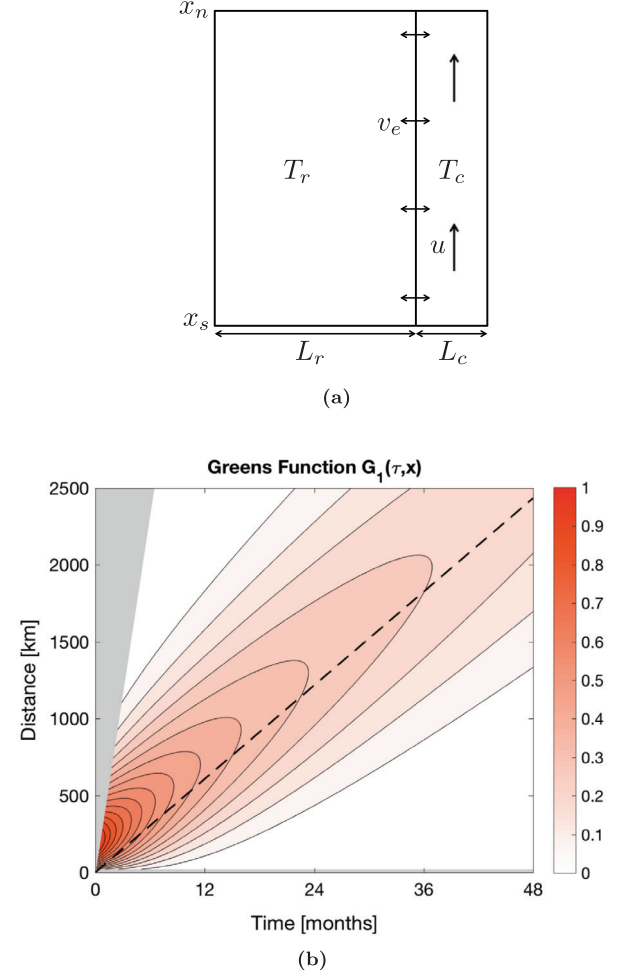


Fig. 5. (a) Illustration of the leaky pipe model. The narrow current (the ‘leaky pipe’) has width L_c , temperature T_c and uniform flow u . The reservoir has width L_r , temperature T_r and no flow. The two regions interact through a mixing described by an eddy velocity v_e . (b) The Green’s function $G_1(\tau, x)$ [Equation (13)], where τ is the time lag, with parameter values representative for the NwASC (Section 4.2). The Green’s function gives an estimate of the tracer–tracer correlation function [Equation (17)], which means that the Green’s function can be compared to the lagged correlation in Fig. 3.

which has been referred to as a leaky pipe model, describes tracer advection in a current core that exchanges tracer concentration with a surrounding stagnant reservoir (Nilsson, 2000; Waugh and Hall, 2005; Jeffress and Haine, 2014). The geometrical setting of the ‘leaky pipe’ model is illustrated in Fig. 5.

We denote the tracer/temperature in the current and reservoir T_c and T_r , respectively. The lateral widths of the two regions are L_c and L_r , respectively, and for simplicity, the vertical extent of the model is taken as constant. The governing tracer conservation

equations are

$$L_c \left(\frac{\partial T_c}{\partial t} + u \frac{\partial T_c}{\partial x} \right) = -v_e(T_c - T_r) + f(x, t) \quad (4a)$$

$$L_r \frac{\partial T_r}{\partial t} = -v_e(T_r - T_c), \quad (4b)$$

where u is the time-mean current speed in the core region, v_e an eddy velocity that provides mixing between the current core and the reservoir, and $f(x, t)$ a source term.

Some illuminating features of the two coupled tracer equations can be obtained by considering the asymptotic behaviour at large times in the absence of the source term $f(x, t)$. In this limit, the leaky pipe model has dynamics that resembles the shear dispersion problem considered by Taylor (1953). Specifically, Taylor examined advection and diffusion of a tracer in a laminar Poiseuille pipe flow: a key result is that at large times, the sectionally averaged tracer concentration is governed by a simplified advection–diffusion equation, where the advection speed is given by the sectionally averaged velocity and shear dispersion gives rise to an enhanced effective along-stream diffusivity (Taylor, 1953; Young and Jones, 1991). The leaky pipe model has similar physics in the limit of large times, and it can be viewed as a two-compartment representation of tracer advection in a shear flow. To illustrate this point, we follow the original heuristic arguments of Taylor (1953; see also Young and Jones, 1991).

The first step is to derive an equation for the sectionally averaged tracer concentration. For this purpose, we introduce sectionally averaged tracer and velocity fields

$$T_E \equiv \frac{L_c T_c + L_r T_r}{L_c + L_r}, \quad u_E \equiv u \frac{L_c}{L_c + L_r}, \quad (5)$$

and a tracer difference given by $\Delta T \equiv T_c - T_r$. By adding Equations (4a) and (4b), taking $f(x, t) = 0$, dividing with cross-stream width $L_c + L_r$, and using the definitions above, we obtain an equation for the sectionally averaged tracer concentration

$$\frac{\partial T_E}{\partial t} + u_E \frac{\partial T_E}{\partial x} + \frac{u_E L_r}{L_c + L_r} \frac{\partial \Delta T}{\partial x} = 0. \quad (6)$$

Here, we note that in the limit of very strong cross-stream tracer mixing (i.e. large v_e), it follows from Equations (4a) and (4b) that $T_c \approx T_r$. In this case, $\Delta T \approx 0$ and hence Equation (6) implies that the sectionally averaged tracer concentration is advected by the effective velocity u_E , which is slower than the current speed u . In fact, this is the approximate behaviour of the leaky pipe model in the limit of large times. To verify this, one takes the difference of the Equations (4a) and (4b), which results in several terms. However, by considering the limit of large times, specified as $t \gg L_c/v_e$, one can neglect the time-derivative

terms. Furthermore, by taking the limit of long along-stream tracer perturbations, specified by $L_x \gg u L_c/v_e$, one finds that cross-stream tracer variations are small compared with the sectionally averaged tracer, i.e. $|\Delta T| \ll |T_E|$. Within this limit, the dominant balance in the tracer-difference equation is

$$L_c u \frac{\partial T_E}{\partial x} = -v_e \Delta T. \quad (7)$$

This equation has a simple physical interpretation that is analogous to the one given by Taylor (1953): the cross-stream velocity difference/shear between the current core and the reservoir creates tracer differences by advecting the sectionally averaged tracer concentration, which are dissipated by cross-stream mixing. Finally, using Equation (7) to eliminate ΔT in Equation (6), one obtains

$$\frac{\partial T_E}{\partial t} + u_E \frac{\partial T_E}{\partial x} = \kappa_E \frac{\partial^2 T_E}{\partial x^2}, \quad \kappa_E \equiv \frac{u_E^2 L_r}{v_e}. \quad (8)$$

Here, κ_E is the effective diffusivity caused by the shear dispersion. As in the classical shear dispersion problem, κ_E is proportional to the square of the sectionally averaged velocity u_E and inversely proportional to the mixing, represented by the eddy velocity v_e in this case.

When applying the leaky pipe model to the NwASC, it is obvious that the current core is generally narrow compared to the reservoir width, i.e. $L_c \ll L_r$. In this regime Equation (5) gives that $u_E \ll u$: the effective advection speed is hence reduced significantly compared to the current speed. Based on the simple two-compartment model, one anticipates that an effective tracer-advection velocity along the NwASC can be estimated as

$$u_E(x) = A^{-1} \iint u(x, y, z) \, dz \, dy, \quad (9)$$

where A is the cross-sectional area of the Atlantic Water, y the coordinate normal to the flow and z the vertical coordinate. As shown in Fig. 1, the westward extent of the Atlantic Water varies along the Slope Current. By defining Atlantic Water as having a salinity above 35, one finds that in the narrower Svinøy section the cross-section area A is about $1 \cdot 10^8 \, \text{m}^2$ and in the wider Lofoten Basin section it is about $5 \cdot 10^8 \, \text{m}^2$. Further, the volume transports in the NwAFC and NwASC are roughly $2 \cdot 10^6$ and $3 \cdot 10^6 \, \text{m}^3 \, \text{s}^{-1}$, respectively (Mork and Skagseth, 2010). Accordingly, the effective velocities, based on the total Atlantic Water volume transport, are 5 and $1 \, \text{cm} \, \text{s}^{-1}$ in the Svinøy and Lofoten sections, respectively; using the NwASC volume transport these velocities reduce to 3 and $0.6 \, \text{cm} \, \text{s}^{-1}$. This simple estimate shows that the Equation (9) yields effective tracer advection velocities that are comparable to the

observed propagation speeds of hydrographic anomalies along the NwASC.

4.2. Lagged correlation and Green's functions

Before discussing an explicit solution of the leaky pipe model in relation to ADT data along the NwASC, we will introduce some mathematical concepts that aid the physical interpretation. We follow [Jeffress and Haine \(2014\)](#), who show that information on the tracer–tracer correlations can be obtained from the Green's function of a linear advection model.

We are interested in the lagged correlation between an upstream (x_1) and downstream (x_2) point, defined as

$$S(\tau, x_2, x_1) \equiv \frac{1}{2t_1} \int_{-t_1}^{t_1} T_c(x_1, t) T_c(x_2, t + \tau) dt, \quad (10)$$

where $2t_1$ represents the time interval over which observations are available and τ is here the time lag. Further, we specify the tracer variations at the upstream point x_1 as

$$T_c(x_1, t) = F(t), \quad (11)$$

where $F(t)$ is an arbitrary function. If the forcing term f in Equation (4a) is taken to be zero, the tracer distribution downstream of x_1 can be calculated as

$$T_c(x_2, t) = \int_{-\infty}^{\infty} G_1(t - t', x_2 - x_1) F(t') dt', \quad (12)$$

where G_1 is a Green's function solution of Equations (4a) and (4b) defined as ([Jeffress and Haine, 2014](#))

$$G_1(T, X) = \frac{v_e}{L_c} \sqrt{\frac{\epsilon X}{T}} I_1(2\sqrt{\epsilon XT}) \exp(-\epsilon T - X) H(T) + \exp(-X) \delta(T). \quad (13)$$

Here, I_1 is the modified Bessel function of the first order, H the unit step function and δ Dirac's delta function. For notational convenience, we have introduced

$$X \equiv \frac{x_2 - x_1}{u L_c / v_e}, \quad T \equiv \frac{t - t'}{L_c / v_e} - X, \quad \epsilon \equiv L_c / L_r. \quad (14)$$

By inserting Equations (11) and (12) into Equation (10), one finds after some algebra that the lagged correlation can be calculated as ([Jeffress and Haine, 2014](#))

$$S(\tau, x_2, x_1) = \int_{-\infty}^{\infty} G_1(\tau - t', x_2 - x_1) C_{FF}(t') dt', \quad (15)$$

where $C_{FF}(t')$ is the auto-correlation function of the tracer at x_1 , namely

$$C_{FF}(t') \equiv \frac{1}{2t_1} \int_{-t_1}^{t_1} F(t) F(t + t') dt. \quad (16)$$

In the limit when $F(t)$ has the character of rapidly de-correlating white noise, $C_{FF}(t')$ approaches a delta function, which implies that Equation (15) simplifies to

$$S(\tau, x_2, x_1) \approx G_1(\tau, x_2 - x_1). \quad (17)$$

Hence in this limit the Green's function is an estimator of the lagged tracer correlation along the flow. As discussed by [Jeffress and Haine \(2014\)](#), the addition of the stochastic forcing term f along the flow and a Gaussian – rather than a delta function – de-correlation of $C_{FF}(\tau)$ primarily act to broaden the tracer–tracer correlation function in space and time. Furthermore, [Jeffress and Haine \(2014\)](#) show that Equation (17) applies not only for the leaky pipe model, but also for a broader class of linear advection models. Hence, if the elevated lagged correlation pattern for large lags and separations in Fig. 3 arises from linear advective dynamics, then it should approximate the corresponding Green's function for large lags and separations.

4.3. Model interpretation and comparison with observations

Figure 5b shows the Green's function G_1 for a set of parameters representative for the NwASC: Based on conditions around the Svinøy section ([Mork and Skagseth, 2010](#)), we take the width of the current L_c to be about 40 km and the reservoir width L_r to be 260 km, yielding $\epsilon = L_r / L_c = 0.15$. The mean speed, averaged over the current core, is $u = 15 \text{ cm s}^{-1}$, which according to Equation (5) gives an effective tracer advection velocity $u_E = 2 \text{ cm s}^{-1}$. The eddy velocity v_e is the parameter that is most difficult to estimate. One method to estimate the eddy velocity is to write $v_e \sim \kappa_e / L_e$, where κ_e and L_e are the eddy diffusivity and length scale, respectively. [Isachsen and Nøst \(2012\)](#) presents estimates of eddy diffusivities and along the NwASC they find values in the range 1000–2000 $\text{m}^2 \text{ s}^{-1}$. Using $L_e \sim L_c$, we obtain $v_e \sim 2\text{--}4 \text{ cm s}^{-1}$.

When interpreting the Green's function physically, it is instructive to recall that $G_1(\tau, x_2 - x_1)$ gives an estimate of the lagged tracer–tracer correlations in a linear advective model; see [Jeffress and Haine \(2014\)](#) and Equation (17). Thus, Fig. 5b illustrates that the tracer correlation in the leaky pipe model has a near- and a far-field region (note that the contribution from the δ function in Equation (13) is suppressed for visual clarity). The near-field region extends downstream a distance on the order of $L_c(u/v_e) \sim 200 \text{ km}$. Here, the downstream lagged correlations

are high along the line in the t - x plane defined by mean-flow speed u . In the near-field region, the reservoir tracer T_r has not yet had time to respond to the variations of the current core tracer T_c . By taking $T_r = 0$ and $f = 0$ in Equation (4a) one sees that, in this small-scale and high-frequency limit, T_c anomalies move with speed u are damped exponentially over the time scale L_c/v_e .

For larger distances and longer time lags, the highest correlation is instead encountered near a line in the t - x plane that represents propagation with the slower effective advection speed u_E . Figure 5b also shows that in the far field, the tracer correlation broadens for increasing lags and distances. In this regime, the sectionally averaged tracer concentration will approximately obey Equation (8): One can show (Nilsson, 2000) that for large distances and times the asymptotic form of the Green's function [Equation (13)] is given by

$$G_1(\tau, x) \approx \frac{v_e}{L_c} \sqrt{\frac{\epsilon L_c}{4\pi\tau v_e}} \exp[-(\tau u_E - x)^2/(\kappa_E \tau)]. \quad (18)$$

Consistent with the advective-diffusive Equation (8), this describes a Gaussian envelope propagating with the speed u_E and broadening due to the effective diffusivity $\kappa_E = L_r u_E^2/v_e$. For the present parameters, the induced diffusivity is comparable to the estimated values of eddy diffusivities along the NwASC (Isachsen and Nøst, 2012). It is important to note, however, that the along-stream effective diffusivity κ_E only becomes manifest on tracer disturbances of long along-stream length scales. Hence, the damping rate due to this diffusivity will be relatively weak, despite that κ_E could be comparable to typical eddy diffusivities.

A cross-section of the Green's function at a fixed distance x in Fig. 5b can also be interpreted as a probability distribution of transit times τ of tracer anomalies between two locations along the flow (Koszalka et al., 2013; Jeffress and Haine, 2014). Similar to the Green's function, the broad band of elevated lagged correlations in the ADT field (Fig. 3) can hence be interpreted as a probability distribution of transit times. The lagged correlations visible in Fig. 3 suggest a broad range of transit times corresponding to an average propagation speed around 2 cm s^{-1} , qualitatively consistent with the Lagrangian transit times estimated by Koszalka et al. (2013).

5. Discussion and conclusions

In the present study, we have used satellite altimetric sea surface height (ADT) data along the NwASC to examine poleward propagation of temperature anomalies. A lagged correlation analysis of the ADT along the NwASC (Fig. 3) resulted in a slanted pattern of elevated correlation that corresponds to a propagation speed of about 2 cm s^{-1} . This is in agreement with propagation speeds of Atlantic Water anomalies in the eastern Nordic Seas

inferred from hydrographic measurements and sea surface temperature data ((Furevik, 2000); Sundby and Drinkwater, 2007; Skagseth et al., 2008; Chepurin and Carton, 2012; Årthun et al., 2017). Further, in the Svinøy section, the ADT data correlates, on inter-annual time scales, fairly well with in situ hydrographic measurements and derived steric height anomalies (Mork and Skagseth, 2010; Richter and Maus, 2011). This suggests that the altimetric data captures migrating low-frequency steric height signals, associated with Atlantic Water temperature anomalies. Compared to hydrographic data, the higher temporal and spatial resolution of the satellite data are potentially advantageous for detecting and examining large-scale hydrographic anomalies. A disadvantage of the altimetric ADT data, however, is that they also record barotropic dynamics associated with large length scales and high propagation speeds (Fig. 2). These mainly wind-forced barotropic variations are generally larger than the steric ones, which reduces the signal to noise ratio and presumably contributes to the relatively weak lagged correlations found in Fig. 3. To some degree, the barotropic variability can be suppressed by removing the spatial mean sea level variations along the NwASC (Fig. 4).

We proposed that the apparent slow propagation of temperature and salinity anomalies along the NwASC primarily results from shear dispersion effects (Taylor, 1953; Young and Jones, 1991): A tracer injected in the current core will mix laterally into regions with weaker mean flow. The effective advection speed of the tracer thus becomes determined by a mean velocity taken over a lateral section extending outside the rapid current core, a mechanism that also can be understood from a Lagrangian perspective (Koszalka et al., 2013). Specifically, we used a two-compartment advection model (Nilsson, 2000; Waugh and Hall, 2005; Jeffress and Haine, 2014) to examine how shear dispersion affects propagation speeds and correlation features of a passive tracer. A key model result is that in the limit of large times and length scales, a balance between shear dispersion and lateral mixing gives rise to weakly damped disturbances that move with a sectionally averaged velocity [see Equations (8) and (9)]. Small-scale, high-frequency disturbances, on the other hand, propagate with the speed of the current core but are quickly attenuated by the lateral mixing. As a result, the lagged tracer correlations between well-separated points along the flow becomes dominated by large-scale, low-frequency disturbances that migrate with the effective advection velocity (Fig. 5b).

When invoking shear dispersion effects to explain delayed tracer transport along current cores, however, one faces some challenges. One difficulty is to define the lateral extent over which the tracer is mixed: unlike pipe and river flows, ocean currents are not confined between solid boundaries. Instead, fronts, topography and/or variations in lateral eddy diffusivity may provide 'dynamical confinement' on the flanks of the current cores, which sets an effective lateral width of a 'leaky pipe'. In the Nordic Seas, we anticipate that the extent of the Atlantic Water layer, between the front of the NwAFC in the

west and the topography along the NwASC in the east, defines an effective lateral width. Using the definition of the effective advection velocity u_E [Equation (9)], a rough estimate based on the Atlantic Water features gave propagation speeds of a few cm s^{-1} . Further, from the definition of u_E , one may expect that the propagation speed of hydrographic anomalies is slower near Lofoten, where the Atlantic cross-section area is larger, than near Svinøy, where the cross-section area is smaller (Fig. 1). To some extent, mean propagation velocities inferred from surface drifters in the Atlantic Water support this notion (Koszalka et al., 2013). However, it is not clear from the lagged correlations shown in Fig. 3 that there are any systematic variations of the propagation speed along the NwASC. Possibly, a longer ADT time series that captures a greater number of propagating low-frequency anomalies could reveal differential speeds along the flow.

Shear dispersion effects are likely to slow the propagation of hydrographic anomalies also along other current systems in the ocean. Similar to the Atlantic Water Current system in the eastern Nordic Seas, the East/West Greenland Current and its extension into the Labrador Sea has a structure with an off-shore front and steep topography towards the coast. Observations document salinity anomalies that propagate along this current system at a few centimeter per second (Dickson et al., 1988; Sundby and Drinkwater, 2007), which is broadly consistent with ‘leaky pipe’ dynamics. Similar slow propagation speeds of hydrographic anomalies are observed along the North Atlantic Current (Chepurin and Carton, 2012; Årthun et al., 2017), which is delimited by sharp front on its northwestern side but essentially lacks nearby topographic slopes in the southeastern direction. A leaky pipe model with a reservoir extending several thousand kilometer to the European continental slope in the east would yield very low propagation speeds that do not match the observations. However, the observed relatively weak eddy diffusivities away from the North Atlantic Current (Abernathey and Marshall, 2013) can possibly limit the effective width of a ‘leaky pipe’ to a couple of hundred kilometers, which yields anomaly propagation speeds that are broadly consistent with the observations.

Along the NwASC and in the entrance to the Barents Sea, the heat transport variations tend to be dominated by velocity variations (Orvik and Skagseth, 2005; Årthun et al., 2012). However, at lower frequencies, the heat transport variations become more dominated by temperature variations, indicating that the potential for long-range predictions of upstream Arctic climatic conditions are tied primarily to the propagation of Atlantic Water temperature anomalies (Årthun et al., 2017). In this context, it can be noted that the transit time of Atlantic Water heat anomalies through the Nordic Seas influences the Barents Sea and Arctic sea ice cover and hydrography in two related ways. First, the transit time determines how much heat the Atlantic Water will lose to the atmosphere before the sea ice edge is reached (Mauritzen, 1996; Koszalka et al., 2013). Second, the transit time controls the delay of the temperature response downstream in the Arctic

Ocean (see e.g. Årthun et al., 2012; Årthun et al., 2017). Thus, the mechanism controlling the propagation speed of hydrographic anomalies and, in particular, how the propagation speed depends on the large-scale atmospheric and oceanic conditions are dynamically interesting issues of relevance for Arctic Ocean variability. Concepts related to shear dispersion and the leaky pipe model discussed in the present study may, in combination with observations and modelling, aid the development of more refined mechanistic models for the propagation speed of Atlantic Water anomalies.

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No potential conflict of interest was reported by the authors.

Notes

1. A few studies have suggested asymmetries in propagation speeds of temperature and salinity anomalies (Yashayaev and Seidov, 2017) and of warm and cold temperature anomalies (Sundby and Drinkwater, 2007). The inferred speeds are still ‘slow’ but the robustness of these results is not fully clear.
2. Such ‘passive’ anomalies can have a steric-height signature at the sea surface and no bottom pressure anomaly; see LaCasce (2018).

References

- Abernathey, R. P. and Marshall, J. 2013. Global surface eddy diffusivities derived from satellite altimetry. *J. Geophys. Res. Oceans* **118**(2), 901–916. DOI:10.1002/jgrc.20066.
- Årthun, M. and Eldevik, T. 2016. On anomalous ocean heat transport toward the Arctic and associated climate predictability. *J. Climate* **29**(2), 689–704.
- Årthun, M., Eldevik, T., Smedsrud, L. H., Skagseth, Ø. and Ingvaldsen, R. 2012. Quantifying the influence of atlantic heat on Barents sea ice variability and retreat*. *J. Climate* **25**(13), 4736–4743.
- Årthun, M., Eldevik, T., Viste, E., Drange, H., Furevik, T. and co-authors. 2017. Skillful prediction of northern climate provided by the ocean. *Nat. Commun.* **8**. DOI:10.1038/ncomms15875.
- Broomé, S. and Nilsson, J. 2016. Stationary sea surface height anomalies in cyclonic boundary currents: Conservation of potential vorticity and deviations from strict topographic steering. *J. Phys. Oceanogr.* **46**(8), 2437–2456.
- Carmack, E., Polyakov, I., Padman, L., Fer, I., Hunke, E. and co-authors. 2015. Toward quantifying the increasing role of oceanic heat in sea ice loss in the new Arctic. *Bull. Am. Meteorol. Soc.* **96**(12), 2079–2105.
- Chafik, L., Nilsson, J., Skagseth, Ø. and Lundberg, P. 2015. On the flow of Atlantic water and temperature anomalies in the Nordic Seas

- towards the Arctic Ocean. *J. Geophys. Res. Oceans* **120**(12), 7897–7918. DOI:10.1002/2015JC011012.
- Chepurin, G. A. and Carton, J. A. 2012. Subarctic and Arctic sea surface temperature and its relation to ocean heat content 1982–2010. *J. Geophys. Res. Oceans* **117**(C6), C06019.
- Dickson, R. R., Meincke, J., Malmberg, S.-A. and Lee, A. J. 1988. The ‘great salinity anomaly’ in the northern North Atlantic 1968–1982. *Progress Oceanogr.* **20**(2), 103–151.
- Furevik, T. 2000. On anomalous sea surface temperatures in the Nordic Seas. *J. Climate* **13**(5), 1044–1053.
- Furevik, T. 2001. Annual and interannual variability of Atlantic water temperatures in the Norwegian and Barents Seas: 1980–1996. *Deep Sea Research Part I* **48**(2), 383–404. DOI:10.1016/S0967-0637(00)00050-9.
- Gill, A. and Schumann, E. H. 1974. The generation of long shelf waves by the wind. *J. Phys. Oceanogr.* **4**(1), 83–90.
- Hansen, D. V. and Bezdek, H. F. 1996. On the nature of decadal anomalies in the North Atlantic sea surface temperature. *J. Geophys. Res.* **101**(C4), 8749–8758.
- Hansen, B., Larsen, K., Hátún, H., Kristiansen, R., Mortensen, E. and co-authors. 2015. Transport of volume, heat, and salt towards the Arctic in the Faroe Current 1993–2013. *Ocean Sci.* **11**(5), 743–757.
- Helland-Hansen, B. and Nansen, F. 1909. The Norwegian Sea – its physical oceanography based upon the Norwegian researches 1900–1904. *Rep. Norwegian Fishery Marine Investigations* **11**, 1–359.
- Isachsen, P. E., Koszalka, I. and LaCasce, J. H. 2012. Observed and modelled surface eddy heat fluxes in the eastern Nordic Seas. *J. Geophys. Res.* **117**(C08020), DOI:10.1029/2012JC007935.
- Isachsen, P. E. and Nøst, O. A. 2012. The air-sea transformation and residual overturning circulation within the Nordic Seas. *J. Marine Res.* **70**(1), 31–68.
- Jakobsen, P. K., Ribergaard, M. H., Quadfasel, D., Schmith, T. and Hughes, C. W. 2003. Near-surface circulation in the northern North Atlantic as inferred from Lagrangian drifters: variability from the mesoscale to interannual. *J. Geophys. Res. Oceans (1978–2012)* **108**(C8), 3251.
- Jeffress, S. and Haine, T. 2014. Correlated signals and causal transport in ocean circulation. *Q. J. R. Meteorol. Soc.* **140**(684), 2375–2382.
- Koszalka, I., LaCasce, J. and Mauritzen, C. 2013. In pursuit of anomalies—Analyzing the poleward transport of Atlantic Water with surface drifters. *Deep Sea Research Part II* **85**, 96–108.
- Krahmann, G., Visbeck, M. and Reverdin, G. 2001. Formation and Propagation of Temperature Anomalies along the North Atlantic Current. *J. Phys. Oceanogr.* **31**(5), 1287–1303.
- LaCasce, J. H. 2017. The prevalence of oceanic surface modes. *Geophys. Res. Lett.* **44** (21), 11097–11105. DOI:10.1002/2017GL075430.
- Lien, V. S., Gusdal, Y. and Vikebø, F. B. 2014. Along-shelf hydrographic anomalies in the Nordic Seas (1960–2011): locally generated or advective signals? *Ocean Dyn.* **64**, 1047–1059.
- Mauritzen, C. 1996. Production of dense overflow waters feeding the North Atlantic across the Greenland-Scotland Ridge. Part 1: evidence for a revised circulation scheme. *Deep Sea Research Part I* **43**(6), 769–806.
- Mork, K. A. and Blindheim, J. 2000. Variations in the Atlantic inflow to the Nordic Seas, 1955–1996. *Deep Sea Research Part I* **47**(6), 1035–1057.
- Mork, K. A. and Skagseth, Ø. 2010. A quantitative description of the Norwegian Atlantic Current by combining altimetry and hydrography. *Ocean Sci.* **6**, 901–911. DOI:10.5194/os-6-901-2010.
- Nilsson, J. 2000. Propagation, diffusion, and decay of SST anomalies beneath an advective atmosphere. *J. Phys. Oceanogr.* **30**(7), 1505–1513.
- Nøst, O. A. and Isachsen, P. E. 2003. The large-scale time-mean ocean circulation in the Nordic Seas and Arctic Ocean estimated from simplified dynamics. *J. Marine Res.* **61**(2), 175–210.
- Orvik, K. A. and Skagseth, Ø. 2005. Heat flux variations in the eastern Norwegian Atlantic Current toward the Arctic from moored instruments, 1995–2005. *Geophys. Res. Lett.* **32**(14), L14610. DOI:10.1029/2005GL023487.
- Polyakov, I. V., Beszczynska, A., Carmack, E. C., Dmitrenko, I. A., Fahrbach, E. and co-authors. 2005. One more step toward a warmer Arctic. *Geophys. Res. Lett.* **32**(17), n/a–n/a. DOI:10.1029/2005GL023740. 17605.
- Richter, K., Furevik, T. and Orvik, K. 2009. Effect of wintertime low-pressure systems on the Atlantic inflow to the Nordic Seas. *J. Geophys. Res. Oceans* **114**(C9), C09006.
- Richter, K. and Maus, S. 2011. Interannual variability in the hydrography of the Norwegian Atlantic Current: frontal vs. advective response to atmospheric forcing. *J. Geophys. Res. Oceans* **116**(C12), C12031.
- Richter, K., Segtnan, O. and Furevik, T. 2012. Variability of the Atlantic inflow to the Nordic Seas and its causes inferred from observations of sea surface height. *J. Geophys. Res. Oceans* **117**(C4), C04004.
- Rudels, B. 2012. Arctic Ocean circulation and variability – advection and external forcing encounter constraints and local processes. *Ocean Sci.* **8**, 261–286.
- Rudels, B., Björk, G., Nilsson, J., Winsor, P., Lake, I. and co-authors. 2005. The interaction between waters from the Arctic Ocean and the Nordic Seas north of the Fram Strait and along the East Greenland Current: results from the AO-02 Oden expedition. *J. Mar. Syst.* **55**, 1–30.
- Skagseth, Ø., Furevik, T., Ingvaldsen, R., Loeng, H., Mork, K. A. and co-authors. 2008. Volume and heat transports to the Arctic Ocean via the Norwegian and Barents Seas. In: *Arctic-Subarctic Ocean Fluxes* (eds. R. Dickson, J. Meincke and P. Rhines). Springer, Dordrecht, Netherlands, pp. 45–64.
- Skagseth, Ø. and Mork, K. A. 2012. Heat content in the Norwegian Sea, 1995–2010. ICES J. Marine Sci. *Journal du Conseil* **69**(5), 826–832.
- Skagseth, Ø., Orvik, K. A. and Furevik, T. 2004. Coherent variability of the Norwegian Atlantic Slope Current derived from TOPEX/ERS altimeter data. *Geophys. Res. Lett.* **31**(14), L14304.
- Søiland, H., Prater, M. and Rossby, T. 2008. Rigid topographic control of currents in the Nordic Seas. *Geophys. Res. Lett.* **35**(18), L18607.
- Sundby, S. and Drinkwater, K. 2007. On the mechanisms behind salinity anomaly signals of the northern North Atlantic. *Progress Oceanogr.* **73**(2), 190–202.
- Taylor, G. 1953. Dispersion of soluble matter in solvent flowing slowly through a tube. *Proc. R. Soc. London A: Math. Phys. Eng. Sci.* **219**, 186–203. The Royal Society.
- Walín, G., Broström, G., Nilsson, J. and Dahl, O. 2004. Baroclinic boundary currents with downstream decreasing buoyancy: a study of an idealized Nordic Seas system. *J. Marine Res.* **62**(4), 517–543.
- Waugh, D. W. and Hall, T. M. 2005. Propagation of tracer signals in boundary currents. *J. Phys. Oceanogr.* **35**(9), 1538–1552.

- Yang, J. and Pratt, L. J. 2013. On the effective capacity of the dense-water reservoir for the Nordic Seas overflow: some effects of topography and wind stress. *J. Phys. Oceanogr.* **43**(2), 418–431.
- Yashayaev, I. and Seidov, D. 2017. The role of the Atlantic Water in multidecadal ocean variability in the Nordic and Barents Seas. *Progress Oceanogr.* **132**, 68–127.
- Young, W. R. and Jones, S. 1991. Shear dispersion. *Phys. Fluids A Fluid Dyn.* **3**(5), 1087–1101.
- Zweng, M. M., Reagan, J. R., Antonov, J. I., Locarnini, R. A., Mishonov, A. V. and co-authors. (2013). World Ocean Atlas 2013, Volume 2: Salinity. *NOAA Atlas NESDIS* (eds. Levitus, S., and Mishonov, A.) Vol. 74, 39pp.