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A Simple Shelf Circulation Model: Intrusion of Atlantic Water on the West Spitsbergen Shelf

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ABSTRACT

Barotropic flow along depth contours is found in accordance with standard geostrophic theory. A numerical model is developed that studies the deviation from such a flow. The model gives a good approximation of the dynamical processes on the West Spitsbergen Shelf (WSS) and shows that the West Spitsbergen Current (WSC), the main gateway of Atlantic water (AW) toward the Arctic, connects more easily to the Isfjorden Trough than anywhere else along the shelf. The circulation of AW in the troughs along the WSS is here named the Spitsbergen Trough Current (STC). From hydrographical and ocean current observations it is evident that the STC is primarily barotropic and driven by the sea surface height. A connection between the along-coast wind stress and the STC is established, and it is demonstrated how the increased occurrence of winter cyclones in Fram Strait during January–February accelerates and widens the WSC. Ultimately, this results in a strengthened STC and dominance of AW on the WSS. The STC represents a slower route of AW toward the Arctic Ocean and a large heat transport toward the West Spitsbergen fjords during winter (0.2–0.4 TW toward Isfjorden). Heat flux estimates show that half of the AW heat loss in the Isfjorden Trough is due to heat loss to the surrounding water masses, while the rest is lost to the atmosphere. Sea ice production along West Spitsbergen has been reduced, or even nonexistent, in some fjords since 2006. Here, the authors argue that this is a consequence of the strong southerly wind periods along the WSS during winter.

1. Introduction

During the past few decades, the Arctic has warmed approximately twice as rapidly as the entire Northern Hemisphere (Serreze and Barry 2011; Francis and Vavrus 2012; Stroeve et al. 2012). In Svalbard, the warming during winter has been particularly strong over recent decades, with an increase in winter temperature of 2°–3°C decade−1 at Svalbard Airport (Førland et al. 2011). The essential atmospheric processes controlling the climate in Svalbard are the large-scale transfer of heat and moisture from lower latitudes, which is strongly controlled by cyclones (Førland et al. 2011). Recent changes in these large-scale atmospheric circulation patterns have brought warm Atlantic water (AW) from the West Spitsbergen Current (WSC) onto the West Spitsbergen Shelf (WSS) and into the fjords even during winter (Cottier et al. 2007; Nilsen et al. 2008; Pavlov et al. 2013). This has halted sea ice from forming and opened up large areas of ice-free waters west and north of Svalbard (Cottier et al. 2007; Tverberg et al. 2014; Onarheim et al. 2014) with a potential impact on the Arctic ecosystem (Hegseth and Tverberg 2013; Lydersen et al. 2014). More frequent episodes of AW intrusion in Arctic regions support the establishment of a number of boreal species, for example, the blue mussel (Berge et al. 2005).

Convergence, mixing, and exchange of AW, Arctic water (ArW), and freshwater from land characterize the WSS. Within an annual cycle, the waters on the shelf and in the adjacent fjords switch from a state of Arctic dominance (cold and less saline in winter) to one of Atlantic dominance (warm and saline in summer) and back (Nilsen et al. 2008). Episodes of AW intrusion on the WSS represent an increased oceanic heat flux into...
the fjord systems and toward the glaciers. Currently there is enormous interest and activity in investigating the role of ocean heat on the stability of ocean-terminating glaciers (Christoffersen et al. 2011; Rignot et al. 2010, 2012; Luckman et al. 2015). Propagation of warm oceanic waters into fjords, with the potential to increase the melt rates of glaciers, has been identified as a likely mechanism leading to the acceleration, thinning, and retreat of glaciers (Holland et al. 2008; Straneo et al. 2010; Inall et al. 2014; Jackson et al. 2014; Luckman et al. 2015). The Greenland ice sheet is a critical source of freshwater to the North Atlantic, and it is via exchanges through the fjords that this freshwater is transported into the oceanic system. The western fjords of Svalbard provide an ideal opportunity to study interactions between glacial, oceanic, atmospheric, sedimentological, and biological processes. While AW influences all the fjords, they vary widely in character; from the heavily glaciated Hornsund with numerous calving glaciers to Isfjorden with only limited glacier cover. Improved understanding of the circulation within these fjords and on the adjacent WSS is important because their response to regional atmospheric, oceanic, and glacier variability is crucial to understand the past and forecast the future behavior of the glaciers.

As observed and reported over a hundred years ago by Helland-Hansen and Nansen (1909), even the surface flow follows the topographic features that form sub-basins, ridges, and plateaus hundreds or even thousands of meters below (Nilsen and Nilsen 2007). In the eastern Fram Strait, the WSC is topographically guided and flows along the Barents Sea continental slope with streamlines of $fH$ (Coriolis parameter/water column depth) and has been traditionally described as a barotropic flow along the Barents Sea Shelf and the WSS break (Aagaard et al. 1987; Manley 1995). The WSC is described in more detail in recent literature where the WSC is separated into a barotropic and baroclinic branch (see Fig. 1 for an illustration; Schauer et al. 2004; Teigen et al. 2010, 2011). Significant heat is lost from the WSC as it flows northward through communication with the WSS (Saloranta and Haugan 2004; Nilsen et al. 2006). Eddy fields generated by barotropic and baroclinic instability (Teigen et al. 2010, 2011) represent a
A vital mechanism for advecting water isopycnally from the warm core to the surface and across the shelf break (Nilsen et al. 2006). Saloranta and Svendsen (2001) found that there is no subsurface density front associated with the Arctic temperature–salinity front at the West Spitsbergen shelf break and concluded that barotropic rather than baroclinic instability is responsible for cross-frontal exchange between the WSC and the shelf waters. However, Tverberg and Nøst (2009) reported observations of variable horizontal density gradients on the continental shelf. Using a numerical model, they demonstrated that the observed density differences caused by the surface heat flux are likely to give rise to frontal instabilities that start a residual overturning circulation across the shelf edge front. V. Tverberg et al. (2015, unpublished manuscript) argue that eddy activity along the shelf edge front is essential for residual overturning to take place and that there must be some degree of topographic steering to bring the AW from the shelf edge to the inner shelf. In the steady state barotropic circulation model studied here without eddy diffusion, emphasis is put on how warm and saline AW in the WSC is directly guided onto the shelf by following isobaths. We employ the quasigeostrophic potential vorticity equation that also supports flows that can break the rigid constraint of strict topographic steering. Moreover, we study the effect of varying the position of the barotropic WSC over the shelf break and focus on the circulation response on the WSS by developing a single-layer model.

The continental shelf adjoining the west coast of Spitsbergen is complex, with alternating shallow banks (50–100-m depths) and deep troughs (200–400-m depths) cutting across the shelf. Here, model results are compared with observations in the Isfjorden Trough (Isfjordrenna) on the WSS (Fig. 1). The deep Isfjorden Trough should have a significant signature on the mean circulation (e.g., Sutherland and Cenedese 2009) owing to the tendency to conserve potential vorticity. Repeated hydrographic cross section data, direct current measurements, and results from a simple barotropic potential vorticity model presented here indicate that a steady, cyclonic, topographically trapped vortex resides over the Isfjorden Trough. The model is based on the theoretical considerations by Pedlosky (1987) and inspired by the application of Lagerloef (1983) on the shelf of Kodiak Island, Alaska. Here, we present a mechanism for transporting heat toward the glaciated fjords, and moreover, a mechanism to precondition the shelf area with warm AW for further exchange with the Arctic fjords during winter. The main hypothesis is that AW flooding events on the WSS have become more frequent during the recent decades because of changes in the atmosphere pressure field and the winter cyclone tracks around Svalbard (Rogers et al. 2005; Francis and Vavrus 2012; Barnes et al. 2014). We present a possible link between the wind-forced circulation on the WSS, the observed wintertime AW intrusion on the WSS and in the fjords, and the lack of sea ice around Svalbard.
Moreover, we reveal the shelf circulation pattern on the WSS by using in situ observations in combination with a simple barotropic shelf circulation model.

Section 2 introduces the study area and presents the data used. Section 3 goes through the quasigeostrophic potential vorticity equation, how this equation can be solved numerically, and how the equation applies to the WSS region. Results are presented and discussed in section 4 and the paper is concluded in section 5.

2. The West Spitsbergen Shelf and the Isfjorden Trough

Figure 2 shows the WSS, including Van Mijenfjorden, Van Keulenfjorden, Kongsfjorden, and the Isfjorden system. In Fig. 2a, only the area that defines the model domain (Fig. 2b) is contoured with blue colors. The 500-m isobath defines the shelf break and the seaward boundary of the model domain. The troughs leading to the above-mentioned fjord systems are included in the model domain while the fjord proper is not included in the model grid setup. The Isfjorden Trough can guide AW all the way from the WSC to the mouth area of Isfjorden (Nilsen et al. 2008). The deepest area of the WSS is found in the mouth of Isfjorden (Svensksunddypet) and the effect of this deep depression is included in the model, as shown by the bottom contours in Fig. 2b. The entrance to Isfjorden is east of this area, behind the coast line, as described in Nilsen et al. (2008).

a. Hydrography and ocean current

Sections of temperature and salinity across and along the Isfjorden Trough are repeated typically in April and September and the stations included in the Forlandet section are indicated by red dots in Fig. 2a. Concurrent acoustic Doppler current profiler (ADCP) sections show generally a nearly barotropic eastward current on the southern side of the trough, and a more baroclinic westward current on the northern side (Fig. 3). The stratification on the WSS is generally weaker during winter and spring, as seen in Fig. 4 showing the Forlandet section for April 2011. The southern side of the section is occupied with AW (salinity above 34.9 psu) in the whole water column, and the water is guided eastward (toward the coast) along the southern trough slope. A vertical front between the colder and less saline ArW (temperature below 0.0°C; Loeng 1991) can be seen toward station 196 south of the trough. Modified AW and ArW close to freezing point temperature are seen to flow seaward on the northern side of the section. Despite the apparent baroclinic structure, the flow seems to be guided by topography on the northern side as well, which indicates a geostrophic balance where the surface elevation plays the major role for the shelf and trough circulation pattern (e.g., Nilsen and Nilsen 2007). The two red boxes on the southern side of the Forlandet section in Fig. 4 indicate two possible AW inflow routes, and the two purple boxes represent the concurrent outflow of the cyclonic circulation pattern that takes place in the Isfjorden Trough. The tallest red box is the modified AW taking the long path around Tampen (Fig. 2a), and the shorter red box is the modified AW taking the short path along Lexryggen (Fig. 2a). These pathways are further illustrated in Fig. 5 and explained in section 4 as a major result of this combined observation and model study.

Figure 5 gives an overview of the circulation pattern for the ArW (blue arrow in the surface) and the AW (red arrow in deeper layers) in the Isfjorden Trough and the connection to Isfjorden and Forlandsundet. The temperature and potential density in the section along the southern side of the Isfjorden Trough and into the mouth of Isfjorden (green circles in Fig. 5) from April 2011 is shown in Fig. 6 and hereinafter called the front section. The salinity data are not shown, but the 34.9 psu isohaline, which defines AW, follows the 1.5°C isothermal (dashed line) in Fig. 6. AW occupies the whole water column at station 44, before we enter a frontal zone in the surface to middepth when going eastward. The density front between Atlantic-type water and ArW on the WSS defines the West Spitsbergen Polar Front (WSPF) close to the mouth of Isfjorden. We adopt the ArW definition from Loeng

FIG. 3. ADCP data and density data calculated from CTD data collected in the Forland section for September 2010 from R/V Håkon Mosby. The color-contoured east (positive)–west velocity is given in cm s\(^{-1}\) and the black-contoured sigma-\(\tau\) lines are in kg m\(^{-3}\). The ADCP section was collected immediately after the CTD section, with the station numbers as indicated at the top of the figure. The ADCP data are 1-min averaged and not corrected for tidal velocity since, for example, the strongest tidal constituent in the region, the M\(_2\) tide, is found from moored instruments to have a major axis velocity between 3 and 4 cm s\(^{-1}\).
(1991) and Skogseth et al. (2005), where ArW originates from the Barents Sea and Storfjorden area, being transported by the East Spitsbergen Current and further by the South Cape Current toward the WSS. It is appropriate to reintroduce the Spitsbergen Polar Current (SPC), as first named by Helland-Hansen and Nansen (1909), which is the coastal current associated with the WSPF in Fig. 6. Hence, the SPC is a surface current (associated with the WSPF) along the west coast of Spitsbergen carrying ArW from Storfjorden and the Barents Sea.

In Fig. 6, AW is seen to penetrate the mouth area of Isfjorden. Below the WSPF depth (~150-m depth), AW circulates along the southern trough and mouth slope, topographically guided around the ArW that resides in the middle of the mouth where it occupies the whole water column (e.g., station 40 in Fig. 6). The front section is not able to exactly follow the AW, but it clearly indicates the topographically steered cyclonic circulation of AW in the mouth area of Isfjorden. To capture the AW penetrating into the mouth east of the WSPF, temperature, salinity, and current at several depths were

Fig. 4. (a) Temperature, (b) salinity, and (c) potential density in the Forlandet section for April 2011. The red boxes on the southern side of the section illustrate the two routes of AW that are topographically guided toward the mouth of Isfjorden. The long route around Tampen where AW is found to flow along isobaths deeper than 200 m (the left red box) and the shorter route over Lexryggen where AW flows coastward along isobaths shallower than 200-m depth. The modified AW flowing westward on the northern side is illustrated with purple boxes, and AW flowing deeper than 150-m depth are able to be topographically guided into the mouth area of Isfjorden and return westward in the deeper layers. The cold surface current on the northern side is the SPC, illustrated by blue circles. The AW shallower than 150 m on the southern side meets the SPC at the mouth of Isfjorden (Fig. 5), turns northward, and returns westward along the northern side of the Isfjorden Trough, flowing side by side with the SPC.
measured from moored instruments at the mouth of Isfjorden (the I1 mooring, turquoise circle in Fig. 5) during four 1-yr time periods (see Table 1). The mooring was redeployed each year at around 78.06°N, 13.52°E at 203–214-m depth from typically September to the following September. Spikes in the temperature and salinity data have been identified and replaced by linearly interpolated values using a window size of 3.5 days and a threshold value of three standard deviations from the detrended mean. Temperature and salinity time series from instruments in homogeneous layers are calibrated from conductivity, temperature, and depth (CTD) profiles obtained at the start and end of the yearlong time series. Temperature time series from the VEMCO loggers and salinity time series from the Aanderaa RCM9, recording Doppler current profiler (RDCP), and SeaGuard were adjusted from comparison with the temperature and salinity time series from the SeaBird SBE37s that were mounted above and below the respective instruments during time periods with homogeneous water columns (wintertime).

b. Wind

We use the latest global atmospheric reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), the ERA-Interim dataset (Dee et al. 2011). The four times per day 10-m wind vector is converted to surface wind stress and gridded on a 75 km × 75 km grid. This is the hindcast grid established by the Norwegian Meteorological Institute (MET Norway; Reistad and Iden 1998). A pronounced example of synoptic wind forcing in Fram Strait and the WSS is shown in Fig. 7 for the winter of 2006. Here, strong low-pressure systems had a trajectory through Fram Strait instead of the usual route into the Barents Sea. Such low-pressure trajectories set up southerly winds along the west coast of Spitsbergen that create a surface Ekman transport toward the shore and negative wind stress curl patterns on the WSS. The seasonal signals (1995–2012) of the along-coast surface wind stress over the southern WSS (Fig. 8a) have a southerly wind component during the winter months with a maximum during January–March. The signal has a large variability and the wind stress can quickly change from a southerly to a northerly wind direction. Nevertheless, the southerly wind signal in the seasonal along-coast wind stress is reflected in the wind stress curl time series (Fig. 8b) from the same area with negative values during the winter months. As shown in Fig. 7, this is due to an eastward/coastward reduction in the southerly wind component that creates a convergence pattern over the WSS. This curl pattern, together with a surface Ekman transport toward the coastline (eastward), is collecting surface water over the shelf where the surface water is stacked up against the west coast of Spitsbergen. The convergence of water on the shelf will change the sea surface tilt (sea level increasing toward the coast) and can both set up a northward flow on the shelf and influence the barotropic branch of the WSC at the shelf break.

3. A simple shelf flow model

Oceanographic data on the WSS and in the Isfjorden Trough demonstrate two important characteristics of
the flow: the flow pattern 1) shows a strong tendency for along-isobath flow in the vicinity of the Isfjorden Trough and 2) does not exhibit a significant seasonal variation from winter to autumn on the southern side of the trough. The data constitute an example of the Taylor–Proudman effect (Taylor 1917; Proudman 1916), which is likely to occur in the ocean under appropriate conditions. This theory holds that steady flow in a rotating, constant-density fluid does not vary along the axis of rotation (z axis) or, equivalently, that barotropic, geostrophic flow follows isobaths. The effect is modified by nonlinearity, stratification, and viscosity. The data from the Isfjorden Trough (Figs. 3, 4) indicate a cyclonic circulation pattern over the trough, a pattern consistent with the conservation of potential vorticity. For this reason this study uses an inviscid potential vorticity model, and a scaling argument following Pedlosky (1987) will be given in the following section. Table 2 lists the appropriate scales and scaling parameters. Most important are the Rossby number $\varepsilon$ and the Ekman numbers ($E_h$, $E_v$, the horizontal and vertical Ekman numbers, respectively), all of which must be very small for the lowest-order steady flow to be geostrophic. The following scaling has been implemented: $E_h^1/2; E_v^1/2 \ll F^{1/2}/2 \sim \varepsilon \ll 1$. The effect of bottom eddy viscosity is initially retained ($E_v^{1/2}/2 \sim \varepsilon$) through a bottom Ekman pumping term in order to discuss a more general vorticity balance and to strengthen our hypothesis of a steady, barotropic, inviscid quasigeostrophic potential vorticity equation in an f plane as our governing equation.

A right-handed Cartesian system is used with the y axis oriented northward along the slope break of the WSS. The x axis is positive shoreward and the origin is located at the reference depth $D$, below the flat sea level (Fig. 9). In this study, $D = 500$ m represents the shelf break depth where the WSC attains its maximum northward velocity ($V_{\text{max}}$) on average. The local depth is $H = D + \eta - h_B$, where $h_B$ is the height of the seafloor above the reference level and $\eta$ is the sea level displacement. Further, $u, v, w$ are the velocity components in the $x, y, z$ directions, respectively; $f$ is the Coriolis parameter; $g$ is the gravitational acceleration; and $A_\eta$ is the vertical kinematic eddy viscosities. By scaling the independent variables ($x, y, z, t$) by ($L, L, L, T$) and the dependent variables ($u, v, w, \eta$) by [V, V, V(D/L), (fVL)/g], the governing equations can be expressed in terms of nondimensional variables. Assuming the fluid to be barotropic, the horizontal momentum and the vertically integrated continuity equation yields

$$
\varepsilon \left[ \frac{\partial \eta}{\partial t} + u \frac{\partial \eta}{\partial x} + v \frac{\partial \eta}{\partial y} \right] - \nu = \frac{\partial^2 \eta}{\partial x^2}
$$

$$
\varepsilon \left[ \frac{\partial \eta}{\partial t} + u \frac{\partial \eta}{\partial x} + v \frac{\partial \eta}{\partial y} \right] + u = \frac{\partial^2 \eta}{\partial y^2}
$$

$$
\varepsilon F \frac{\partial^2 \eta}{\partial t^2} + \left( u \frac{\partial^2 \eta}{\partial x^2} + v \frac{\partial^2 \eta}{\partial y^2} \right) \left( \varepsilon F \eta - \frac{h_B}{D} \right)
$$

$$
\varepsilon F \frac{\partial^2 \eta}{\partial t^2} + \left( 1 + \varepsilon F \eta - \frac{h_B}{D} \right) \left( u \frac{\partial^2 \eta}{\partial x^2} + v \frac{\partial^2 \eta}{\partial y^2} \right) = 1/2 E_v^{1/2},
$$

where $\varepsilon = V/(fL)$ is the Rossby number, $\zeta = \partial u/\partial x - \partial v/\partial y$ is the relative vorticity, $F = (f^2 L^2)/(gD)(L^2 R^2)$ is the influence of the surface elevation where $R$ is the barotropic Rossby radius, and $E_v = 2A_\eta/(fD^2)$ is the vertical Ekman number. Next, we expand all the dependent variables in power series in the small Rossby number. For example,

$$
v(x; \varepsilon) = \sum_{n=0}^{\infty} \varepsilon^n v_n(x)
$$
is the expansion of the velocity vector. To lowest order, $O(1)$, the horizontal momentum equation is

$$
u_0 = \frac{\partial \eta_0}{\partial x}, \quad u_0 = -\frac{\partial \eta_0}{\partial y},$$  

(3)

which are the geostrophic velocities and equivalent to the definition of $\eta_0$ as a streamfunction, hence, $\psi = \eta_0$ and $\nabla^2 \psi = \xi_0$, where $\nabla^2 = \partial^2/\partial x^2 + \partial^2/\partial y^2$. Substitution of (3) into the continuity [(1)] shows that

$$\left( \frac{\partial u_0}{\partial x} + \nu_0 \frac{\partial}{\partial y} \frac{h_B}{D} \right) = \nabla \left( \frac{h_B}{D} \right) = 0,$$

(4)

and states that the flow is, to lowest order, confined to flow along isobaths. A more dynamically relevant equation can be derived if the Rossby number is considered large enough to break the rigid constraint of strict topographic steering in (4), that is, that the motion is sufficiently distant from the exact geostrophic mode. This is possible when the order of magnitude of $h_B/D$ is reduced, or more precise, is restricted to be of the same order as the Rossby number, thus $h_B/D \sim \epsilon$. For our situation this implies that the relative vorticity

![Fig. 7. The surface wind stress (arrows) and wind stress curl (contoured colors) pattern in Fram Strait and the oceans surrounding Svalbard (Barents Sea, Greenland Sea, and the Arctic Ocean). Among events of intense southerly winds along the WSS, a case of time mean spatial pattern is shown for 18–25 Dec 2005. The black parallelogram marks the region were a spatially averaged time series of the wind stress is constructed, the red asterisk is the position of the corresponding wind stress curl time series, and the turquoise circle marks the position of the II current meter mooring in the mouth of Isfjorden.](image)

![Fig. 8. The seasonal signal of (top) the along-coast surface wind stress and (bottom) the wind stress curl in the positions marked in Fig. 7. The graphs give the monthly time mean from 1995 until 2012, and the bars give the standard deviation for each month.](image)
WSC and the topographic steering of the different troughs incindent on the shelf have equal importance, and the Rossby number scaling will determine which of these two will dictate the flow field on the shelf. To better monitor and scale the dominance of either relative vorticity or topographic steering, a new variable \( \eta_B \) for the seafloor topography is introduced:

\[
\eta_B = h_B/D,
\]

where \( O(\eta_B) \sim 1 \) in order to satisfy the constraint \( h_B/D \sim \varepsilon \). If \( h_B/D \) is much greater than \( O(\varepsilon) \), \( \eta_B \) is then far greater than \( O(1) \), and we return to the purely geostrophic motion flowing along isobaths, that is, (4). Next, scaling arguments and data will be used to decide if the effect of friction is important to retain or not while developing a vorticity balance for a steady state flow over varying topography.

The second governing equation is derived from an elimination of \( v_1 \) from the \( O(\varepsilon) \) horizontal equation [3] and use of the \( O(\varepsilon) \) continuity equation [1] to yield the quasigeostrophic potential vorticity equation

\[
\left( \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} + v_0 \frac{\partial}{\partial y} \right) (\xi - F \eta_0 + \eta_B) + \frac{r}{2} \xi_0 = 0, \tag{6}
\]

where \( r = E_v/\varepsilon \) and \( F \) is a relative measure of the vorticity induced by displacement of the free surface. In the strictly inviscid limit, where \( r \ll 1 \) (i.e., \( \varepsilon \gg E_v^{1/2} \)), it can be shown (Pedlosky 1987) that, to \( O(\varepsilon^2) \), the dimensionalized version of (6) is equivalent to the more familiar inertial potential vorticity conservation equation

\[
\frac{D}{Dt} \left( \frac{\xi + f}{H} \right) = 0. \tag{7}
\]

Henceforth, we will consider the steady-state version of (6) and use the geostrophic streamfunction \( \psi = \eta_0 \). For a continental shelf, \( F \ll 1 \), which is a rigid lid approximation, and terms of \( O(F) \) can be neglected. Equation (6) is now rewritten

\[
\left( \frac{\partial \psi}{\partial x} \frac{\partial}{\partial y} - \frac{\partial \psi}{\partial y} \frac{\partial}{\partial x} \right) \frac{h_B}{D} + \varepsilon \left( \frac{\partial \psi}{\partial x} \frac{\partial \psi}{\partial y} - \frac{\partial \psi}{\partial y} \frac{\partial \psi}{\partial x} \right) \frac{h_B}{D} + \varepsilon \nabla^2 \psi = 0. \tag{8}
\]

Equation (8) represent a vorticity balance between three terms: the vorticity induced by flow across isobaths, which is of order \( \nabla (h_B/D) \); a vorticity advection term of order \( \varepsilon \); and a bottom boundary layer pumping term of order \( 1/2E_v^{1/2} \). The topography term is generally the largest one on the shelf. The small deviation between streamlines and isobaths may be balanced by the advective term or pumping term, or both, depending on the relative scaling of \( \varepsilon \) and \( 1/2E_v^{1/2} \). The linear vorticity balance between the first and the last term, when \( 1/2E_v^{1/2} \gg \varepsilon \), states that positive (negative) vorticity must exist at a point if flow is toward deeper (shallower) water, or alternatively, it states that the maximum relative vorticity occurs over the maximum topographic gradient. On the other hand, when \( \varepsilon \gg 1/2E_v^{1/2} \) and there is a balance between the first and second term,

\[
\left( \frac{\partial \psi}{\partial x} \frac{\partial}{\partial y} - \frac{\partial \psi}{\partial y} \frac{\partial}{\partial x} \right) \left( \frac{h_B}{D} + \varepsilon \nabla^2 \psi \right) = 0. \tag{9}
\]

**Table 2. Scales and scaling parameters appropriate to the WSS and WSSl.**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>( V_{max} )</td>
<td>( \sim 20–36 \text{ cm} \text{s}^{-1} )</td>
<td>Maximum current over the 500 m isobath in the barotropic WSC</td>
</tr>
<tr>
<td>( V )</td>
<td>( \sim 20 \text{ cm} \text{s}^{-1} )</td>
<td>Velocity scale (average current in the barotropic WSC)</td>
</tr>
<tr>
<td>( L )</td>
<td>( \sim 20 \text{ km} )</td>
<td>Length scale (width of the barotropic WSC profile)</td>
</tr>
<tr>
<td>( f )</td>
<td>( 1.43 \times 10^{-4} \text{ s}^{-1} )</td>
<td>Coriolis parameter (( -78^\circ \text{N} ))</td>
</tr>
<tr>
<td>( A_v )</td>
<td>( -1 \times 10^{-4} \text{ m}^2 \text{s}^{-1} )</td>
<td>Vertical eddy viscosity</td>
</tr>
<tr>
<td>( A_h )</td>
<td>( -10 \text{ m}^2 \text{s}^{-1} )</td>
<td>Horizontal eddy viscosity</td>
</tr>
<tr>
<td>( D )</td>
<td>( 500 \text{ m} )</td>
<td>Bottom depth at max WSC velocity and reference depth</td>
</tr>
<tr>
<td>( h_B )</td>
<td>( \sim 300 \text{ m} )</td>
<td>Characteristic trough height above the reference depth</td>
</tr>
<tr>
<td>( e )</td>
<td>( V(FL) = 0.07–0.13 )</td>
<td>Rossby number</td>
</tr>
<tr>
<td>( \eta )</td>
<td></td>
<td>Trough aspect ratio</td>
</tr>
<tr>
<td>( \xi )</td>
<td></td>
<td>Vertical Ekman number</td>
</tr>
<tr>
<td>( \zeta )</td>
<td></td>
<td>Horizontal Ekman number</td>
</tr>
</tbody>
</table>

**Fig. 9.** Schematic of the one-layer model where \( \eta \) is the sea surface elevation, \( H \) is the total depth, \( h_B \) is the bottom variation, and \( D \) is a reference depth.
which represents the conservation along a streamline of the quantity \( h \partial \psi / \partial t + \nu \nabla^2 \psi \), the potential vorticity. Equation (9) states that the relative vorticity must increase (decrease) downstream if flow is toward deeper (shallower) water. Hence, (9) states that the maximum vorticity gradient occurs over the maximum topographic gradient. In other words, if friction and bottom boundary layer pumping was dominating \( (\nu E_{\nu} / \nu^2 \gg \epsilon) \), the flow would be downslope on the southern side of the Isfjorden Trough with a distinct return flow along the deep part of the trough (the trough axis). On the contrary, (9) is used here because the data strongly suggested potential vorticity conservation with cyclonic flow around a deep trough, that is, a Taylor–Proudman column of cyclonic relative vorticity associated with greatest depth. The scaling factors (Table 2) show that \( \epsilon > \nu E_{\nu}^{1/2} \) points toward the potential vorticity model. For the Isfjorden Trough, the Rossby number is small enough for inertial terms to be important to the ageostrophic components of the momentum and vorticity balance.

a. The numerical potential vorticity model

The inertial potential vorticity equation [(9)] is chosen because the data indicate potential vorticity conservation; (9) states that potential vorticity (after dividing by \( \epsilon \)), \( \nabla^2 \psi + \eta_B \), is conserved along a streamline. An equivalent expression is

\[
\nabla^2 \psi + \eta_B = K(\psi),
\]

(10)

where the potential vorticity is a function of \( \psi \). The function \( K(\psi) \) is a value that must be determined for each streamline. Equation (10) is a Poisson equation where \( G(x, y, \psi) = [\eta_B - K(\psi)] \) will be prescribed on the boundary of the (bottom topography) grid. The finite difference scheme used to solve (10) incorporates a 75 × 35 rectangular grid with the row \( j = 1:75 \) and column \( k = 1:35 \) subscripts representing the \( y \) and \( x \) axes (Fig. 2). The grid spacing, \( \Delta x = \Delta y \), is 4 km and the finite difference notation for (10) is

\[
-\frac{1}{4} \psi_{j,k-1} - \frac{1}{4} \psi_{j,k+1} - \frac{1}{4} \psi_{j+1,k} - \frac{1}{4} \psi_{j-1,k} = \frac{\Delta x^2}{4} [\eta_{B,j,k} - K(\psi_{j,k})]\]  

(11)

with \( \psi_{j,k}^{+1} \) denoting the nodal solution at a position \((j, k)\) in the grid, indicated by the superscript \((t + 1)\) forward in the iteration sequence, while \( i \) indicates known values on the grid around the node. Equation (11) comprises a five-point Gauss formula, and from the resulting set of equations, a square sparse coefficient matrix \( A \) can be generated for the internal nodes (those lying inside the area of interest). The problem is then reduced to solving a linear system

\[
A \psi = \frac{\Delta x^2}{4} G,
\]

(12)

where \( A \) is a square matrix bounded by the number of internal grid points or number of grid points within the boundary values (specified in the next section). The grid is constructed of bathymetry from the WSS and fjord mouth areas based on the International Bathymetric Chart of the Arctic Ocean (IBCAO; Jakobsson et al. 2008). For convenience, the \( y \) axis was located at the seaward boundary and was made a streamline to include the dynamic effect of the WSC. The 500-m isobath is chosen as the \( y \) axis where the isobath meanderings are straightened to follow the \( y \) axis (Figs. 2a,b). The origin is placed at the 500-m isobath slightly north of the South Cape latitude, and the \( x \) axis is positive toward east/the coast, where the coast line experiences distortions due to the straightening of the 500-m isobath. Because of a 6-m shallow region in northern Forlandsundet, the island Prins Karls Forland is considered a peninsula connected to Spitsbergen (Fig. 2b). This makes the boundary values easier to specify since the shoreline is assumed to be a streamline where \( \psi = 0 \).

b. Boundary conditions: The West Spitsbergen Current

The upper slope branch of the WSC can be described as a topographically guided barotropic jet flowing along isolines of planetary potential vorticity \( f/H \) and is referred to here as the barotropic WSC branch (Teigen et al. 2011). The WSC is the boundary condition at the seaward boundary of our model domain and is controlled and varied by the strength of the maximum current speed and the horizontal current profile width. Analyses of current meter data suggest that the current is asymmetric (Teigen et al. 2010), with a sharper current shear on the shore side than on the ocean side (Fig. 10). An asymmetric (but still analytically differentiable) idealized current profile was therefore proposed in Teigen et al. (2010) constructed by multiplying a Gaussian profile with a sigmoid curve, yielding a skew Gaussian profile (Fig. 10a). Sections across the West Spitsbergen Current taken with ship-mounted ADCPs (Walczowski et al. 2005) also support this result.

In our simplified version of the barotropic WSC branch, the \( y \) axis is a streamline and fixed at the seaward boundary on the 500-m isobath where the average maximum current speed position of the WSC is found (Teigen et al. 2010; Figs. 2, 10a). The \( x \) axis is the
upstream (southern) boundary where a cross slope distribution of \( \psi_0 \) \((y = 0 \text{ or } j = 1)\) is imposed \((j = 1, k = 1:35)\), representing the eastern half of the fitted skewed jet WSC profile (Fig. 10a), so that the function \( K(\psi) \) can be determined. The sea level \( \psi_0 \) increases from the maximum current position \((j = 1, k = 1)\) at the shelf break toward the shelf where \( \psi \) is set to the reference level zero in the initial setup of the numerical iterations. Furthermore, the streamfunction \( \psi \) is only allowed to be negative in order to be coupled to the boundary current, the WSC. A parabolic function is used assuming \( \partial^2 \psi_0 / \partial y^2 = 0 \) and \( \partial^2 \psi_0 / \partial x^2 = -(\xi_0 L) / V \) at \( y = 0 \ (j = 1, k = 1:35)\), where \( \xi_0 \) is now the absolute value of the relative vorticity \((|\xi_0|)\) along the \( x \) axis at \( y = 0 \). Making the current profile more general (now in continuous function form) by assuming that the maximum WSC can increase when translating the velocity scale \( V = 20 \text{ cm s}^{-1} \) an eastward distance \( x = a \), and utilizing that \( \psi_0 = 0 \) at the coast \( x = L_x / L \), we arrive at the following boundary condition along the \( x \) axis \((y = 0)\):

\[
\psi_0 = -\frac{1}{2} \frac{\xi_0 L}{V} \left[ x^2 - \left( \frac{L_x}{L} \right)^2 \right] + \frac{\xi_0}{V} \left( 1 + \frac{1}{V} \right) \left( x - \frac{L_x}{L} \right),
\]

(13)

and from this,

\[
u_{\text{WSC}} = \frac{\partial \psi_0}{\partial x} = -\frac{\xi_0 L}{V} \left( x - \frac{a}{L} \right) + 1.
\]

(14)

There exists a position \( x_0 = V / (\xi_0 L) + a / L \) on the shelf where the \( v_{\text{WSC}} \) profile given in (14) reaches a zero northward velocity. A more dynamically important zero

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**Fig. 10.** (a) The time-averaged (October 2007 to June 2008) skewed jet current profile of the fitted skewed jet WSC profile constructed from observations of the WSC across Fram Strait at 78.83°N. The black arrow shows the averaged maximum velocity in the WSC profile, and the line marked by \( v_{\text{WSC}} \) is the velocity profile used to represent the WSC in the model. (b) The calculated relative vorticity of the WSC. (c) A fitted skew jet profile of the WSC (every hour), from October 2007 to June 2008, at the 200–250-m level at mooring F0–F5, from Teigen et al. (2010). Mooring positions are marked in all figures and the mooring depth is given in parenthesis in (a). The vertical black line through all subfigures marks the value [in (b)] chosen for the relative vorticity \( \xi_0 \) in the model.
crossing is where the sea surface elevation reaches the reference level \( \psi_0 = 0 \):

\[
x_{\psi_0} = 2\left(\frac{a}{L} + \frac{V}{\xi_0 L}\right) - \frac{L_s}{L}.
\]

(15)

For the sea surface elevation to hug the shelf break this position has to be far enough east in order to generate any current on the shelf and in troughs, and henceforth, for properties in the WSC to be transported across the shelf. Conclusively, the primary role of the WSC in the model is to set the height of the sea surface material isopleths along the shelf slope.

Two assumptions influence the physical interpretation of the model results: 1) the streamfunction \( \psi \) is only allowed to be zero or negative and 2) we choose a constant relative vorticity \( \xi_0 \) for the upstream boundary condition at \( y = 0 \), that is, a linearly decreasing WSC profile from the maximum current \( V_{\text{max}} \) and toward the zero crossing on the shelf. The first assumption controls the sea surface elevation and the geostrophic velocities. Moreover, the formula for zero crossing \([\text{(15)}]\) determines where the WSC set up a pressure gradient force on shelf areas. By choosing a constant relative vorticity, we strongly simplify the true nature of the WSC (Fig. 10). As seen in Figs. 10a and 10b, the yearly averaged current profile is a skewed Gaussian jet profile with a highly varying relative vorticity across the West Spitsbergen Slope (WSSl).

The constant relative vorticity chosen for our model is given by the dashed line in Fig. 10b, \( \xi_0 = 0.4762 \times 10^{-5} \text{s}^{-1} \), and represents the absolute value of the eastern part of the WSC profile (Fig. 10a). This value of \( \xi_0 \) is chosen such that the linear current profile with a maximum current of \( V_{\text{max}} = V = 20 \text{ cm s}^{-1} \) (the yearly average in Fig. 10a) at \( x = 0 \) is not able to be topographically steered into troughs indenting the WSS, as shown in Fig. 2b where the contour lines just follow the WSS. No flow into the trough is achieved by letting the streamfunction \( \psi \) have a zero crossing in a grid point west of an isobath that leads toward the troughs in the slope region. A small increase in \( V_{\text{max}} \) or a shift in the position of \( V \) toward the shelf (i.e., increasing \( a \)) will immediately initiate currents in the troughs. This paper will study the development and dependency of the shelf flow on \( a \) (or \( V_{\text{max}} \)).

Figure 10c shows that the barotropic WSC can extend farther onto the shelf when \( V_{\text{max}} \) increases or \( V \) shifts toward the shelf, and this is captured in our simple linear velocity profile in (14).

Solutions to (12) are found through Gaussian elimination. Initially, all values of \( \psi \) are set to zero except at the boundaries (\( j = 1 \) and \( k = 1 \)). Along \( j = 1, \psi_{1,k} = \psi_0 \) was computed according to (14), and then \( \psi_{j,1} \) was set to \( \psi_{1,j} = \psi_0(k = 1) \) for all \( j \) to initialize the seaward boundary. For each successive iteration by Gaussian elimination, a new \( \psi_j^{k+1} \) is calculated from (11) based on older values. Since \( K(\psi) \) is conserved along streamlines [see Eq. (10)], the function \( K(\psi_{j,k}) \) is determined by searching along the upstream boundary for the particular value to which \( \psi_{j,k} = \psi_0(k) \) and calculating \( K(\psi) = -C_{1j} \psi_0(k = 1, k) \) at that point \( k \). Moreover, this \( K(\psi_{j,k}) \) is then used to calculate \( G_{j,k} \) in (12). After each iteration a fractional change is calculated based on the old values, and a solution is reached when the fractional change between iterations converges toward a small value.

4. Results and discussions

a. Shelf circulation

Here, we present model results where the simple linear current profile representation of the WSC [(14) and Fig. 10a] is moved eastward onto the shelf. Thus, the position of the characteristic WSC velocity \( V = 0.20 \text{ m s}^{-1} \) will be moved up the slope (eastward) and onto the shelf by increasing \( a \) in (14). If the maximum velocity of WSC increases or the characteristic velocity \( V \) is shifted even farther onto the shelf, shallower isobaths will be exposed to the WSC and thereby able to guide AW farther onto the shelf. This is clearly illustrated in Figs. 11a and 11c, corresponding to the boundary conditions \( a = 2 \text{ km} \) and \( a = 8 \text{ km} \), respectively, where water masses from the WSC circulate a larger number of troughs and shallower areas on the WSS when the characteristic velocity of the WSC is moved farther eastward.

Table 3 lists some of the important model parameters and volume transport results in the Forlandet section (Figs. 2, 3) as a function of \( a \), the position of the characteristic current in the WSC. The second column gives \( x_{\psi_0} \), that is, how far east the sea surface elevation of the WSC reaches onto the shelf. This dominates the final circulation pattern, and Figs. 11a–d show this pattern for \( a = 2, 3, 8, \) and \( 14 \text{ km} \), respectively. When \( a = 2 \text{ km} \) the characteristic velocity of the WSC has climbed up 50 m from the 500-m isobath and the WSC sea surface tilt is able to generate current in the Isfjorden Trough only. Deep bottom slopes are able to guide AW in along the southern side of Isfjorden Trough and out along the northern side. No currents are generated above isobath shallower than 200 m because of surface elevation \( \psi_0 = 0 \) when \( H < 200 \text{ m} \), and this is the reason for lack of AW transport into the Kongsfjorden Trough (Fig. 2) at this stage. Forcing the WSC farther onto the shelf (Fig. 11b with \( a = 3 \text{ km} \)) initiates flow above the 200-m isobath, and a cyclonic flow pattern is also generated in the Kongsfjorden Trough.

The WSC connects easier to the Isfjorden Trough than anywhere else along the shelf because the Isfjorden
Trough is deeper than the other troughs (Fig. 11). When increasing $a$ in the model runs, Fig. 11 shows that the AW finds new paths into the trough toward Isfjorden. At $a = 3$ km (Fig. 11b), all the AW circulates around Tampen (Fig. 2) and continues eastward on the southern side of the Isfjorden Trough. In Fig. 11c, when $a = 8$ km, circulation is stronger in Eggbukta, and some of the isobaths in Eggbukta lead into the Isfjorden Trough along Lexryggen. This pathway is captured by the Forlandet section that now experiences two major branches transporting AW into the Isfjorden Trough. When $a = 14$ km (Fig. 11d) the path from Eggbukta over Lexryggen seems to dominate the inflow of AW toward Isfjorden. At this stage the Kongsfjorden Trough is receiving modified AW directly from the Isfjorden Trough, and Bellsund in front of Van Mijenfjorden is also in connection with the WSC. Hence, the entire WSS can be influenced by the warm and saline AW if the WSC is forced far enough onto the shelf.

Deviation from this general geostrophic shelf circulation pattern and strict topographic steering can be seen...
in both observations and model results in the Forlandet section (Fig. 2). An across-slope and across-trough flow is more pronounced in the northern part of the Forlandet section compared to the southern part for the model runs with $a < 8$ km (Figs. 11a,b). Furthermore, for $a = 3$ km in Fig. 11b, the two streamlines following the 200-m bathymetry line along the southern side of the through connects in the area close to the mouth of Isfjorden since the maximum vorticity gradients occur over the maximum topography gradients [when $O(\eta_B) \sim 1$] and the flow will therefore have a tendency toward deeper water. The slightly across-slope flow and flow toward deeper water is reflected in the summer observations where a westward return flow (Fig. 3) on the northern side of the Isfjorden Trough is flowing along deeper bathymetry than compared to the initial eastward current on the southern side after following the trough–slope cyclonic circulation pattern. Moreover, the winter CTD observations (Fig. 4) also show that AW taking the long path around Tampen (Fig. 2a) on the southern side of the Isfjorden Trough is found to flow along deeper isobath in the westward return flow in the northern part of the Forlandet section. As explained in the next section, this happens when the relative vorticity term in (10) becomes significant and can relieve the flow from strict topographic guiding.

**b. Volume and heat fluxes**

Figure 12 shows the volume transport in the Forlandet section (Fig. 2) in the Isfjorden Trough when $a = 8$ km. The Forlandet section (Fig. 2) is chosen since the section seems to capture all the flow branches in the trough and gives a representative picture of the changes in the volume transport when the boundary values are changed.

Table 3 gives an overview of the volume transport in the Forlandet section for the different $a$ cases. It is seen that the eastward transport, on the order of 0.02 Sv ($1$ Sv = $10^6$ m$^3$ s$^{-1}$), to the Isfjorden Trough increases with increasing $a$, that is, an increasing eastward transport of AW from the WSC to the WSS. The eastward transport is nearly compensated by the westward transport on the northern side of the trough, but the total volume transport shows a net positive inflow on the order of 0.001 Sv. The total volume transport reaches a maximum (0.0025 Sv) in both observations and model results in the Forlandet section (Fig. 2). An across-slope and across-trough flow is more pronounced in the northern part of the Forlandet section compared to the southern part for the model runs with $a < 8$ km (Figs. 11a,b). Furthermore, for $a = 3$ km in Fig. 11b, the two streamlines following the 200-m bathymetry line along the southern side of the through connects in the area close to the mouth of Isfjorden since the maximum vorticity gradients occur over the maximum topography gradients [when $O(\eta_B) \sim 1$] and the flow will therefore have a tendency toward deeper water. The slightly across-slope flow and flow toward deeper water is reflected in the summer observations where a westward return flow (Fig. 3) on the northern side of the Isfjorden Trough is flowing along deeper bathymetry than compared to the initial eastward current on the southern side after following the trough–slope cyclonic circulation pattern. Moreover, the winter CTD observations (Fig. 4) also show that AW taking the long path around Tampen (Fig. 2a) on the southern side of the Isfjorden Trough is found to flow along deeper isobath in the westward return flow in the northern part of the Forlandet section. As explained in the next section, this happens when the relative vorticity term in (10) becomes significant and can relieve the flow from strict topographic guiding.

**Table 3. Model parameters and the Isfjorden Trough volume transport across the Forlandet section.** Variable $a$ is the eastward position of the WSC characteristic velocity $V$, $x_0$ is the eastward position of the zero crossing of $\phi_0$, $V_{\text{max}}$ is the maximum velocity of the WSC at the 500-m isobath, $e$ is the Rossby number, $V_{TE}$ is the eastward transport on the southern side of the Isfjorden Trough, $V_{TW}$ is the westward transport on the northern side, and $VT$ is the total transport in the section.

<table>
<thead>
<tr>
<th>$a$ (km)</th>
<th>$x_0$ (km)</th>
<th>$V_{\text{max}}$ (m s$^{-1}$)</th>
<th>$e$</th>
<th>$V_{TE}$ (m$^3$ s$^{-1}$)</th>
<th>$V_{TW}$ (m$^3$ s$^{-1}$)</th>
<th>$VT$ (m$^3$ s$^{-1}$)</th>
</tr>
</thead>
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<tr>
<td>0</td>
<td>7.9</td>
<td>0.200</td>
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<td>$1.059 \times 10^4$</td>
<td>$-1.049 \times 10^4$</td>
<td>0.000</td>
</tr>
<tr>
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<td>8.9</td>
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<td>$-1.049 \times 10^4$</td>
<td>0.010</td>
</tr>
<tr>
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<td>0.205</td>
<td>0.07</td>
<td>$1.059 \times 10^4$</td>
<td>$-1.049 \times 10^4$</td>
<td>0.010</td>
</tr>
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<td>0.010</td>
</tr>
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<td>$-1.529 \times 10^4$</td>
<td>1.229</td>
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<tr>
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</tr>
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<td>$-3.338 \times 10^4$</td>
<td>0.334</td>
</tr>
</tbody>
</table>
around $a = 8\text{ km}$ (Fig. 12) and then decreases again. An explanation is given through the conservation of potential vorticity. When $a$ is low, the circulation on the shelf follows the deep bottom contours in the troughs. The deviation from the reference depth $D$, $h_B$, is then small, and thus, $h_B/D \sim \varepsilon$, and the relative vorticity term in (10) becomes more important. This will relieve the flow from strict topographic guiding and initiate cross-trough flow and cyclonic circulation around deep depressions, as seen around the Svensksunddypet in the mouth of Isfjorden.

When $a$ increases beyond $8\text{ km}$, the WSC water masses follow shallower isobaths on the shelf (Fig. 11d). Because of conservation of potential vorticity the AW is more easily guided over these bottom contours because the trough slopes are steeper on the upper part of the slopes (Fig. 2). Moreover, the relative vorticity term in (10) becomes less important since $h_B/D$ approaches the trough aspect ratio (Table 2) and becomes much larger than $\varepsilon$. It is therefore less easy for the water to deviate from a strict topographic steering, and this is reflected in the decreasing total volume transport in Table 3. Hence, it is possible for the cyclonic flow in the Isfjorden Trough to deviate from standard geostrophic theory until the shorter path following Lexryggen becomes dominant (Figs. 5, 11). Time series of speed and direction (not shown) from the I1 mooring (Table 1) show that the velocity vector rotates between 10° and 15° clockwise when the speed exceeds 30–40 cm s$^{-1}$. As will be shown later, this occurs during winter months when the trough flow in the Isfjorden Trough becomes an order of magnitude stronger, and it is most pronounced during the years when the shorter path following Lexryggen dominates. Hence, the velocity vectors are less aligned with the bathymetry and show a deviation from the standard geostrophic theory when the flow follows deeper bathymetry in the longer route around Tampen.

Heat transport is determined by the difference in temperature between the temperature of the water masses being advected into or out of a volume as compared to the temperature representative for the water masses being replaced. Thus, if Fig. 11 represents the averaged circulation pattern for four different winter seasons, it is obvious that the heat transport onto the WSS will be very different between these cases because of the circulation of AW from the WSC and the contrast between AW temperatures and the ArW temperatures. During winter when the contrast between the $2^\circ$–$3^\circ\text{C}$ AW and the ArW close to freezing point temperature is largest, the heat input from the WSC to the WSS is noticeable in the sea ice cover and air temperatures. Using the eastward barotropic volume transports given in Table 3, the eastward heat transports on the southern side of the trough can reach 0.2–0.4 TW for $a$ between 2 and 8 km. It is interesting to follow this AW as it circulates the Isfjorden Trough and estimate how much heat is released to the atmosphere on the way. By assuming no heat is lost to the surrounding water masses, but instead all the heat is lost to the atmosphere, the heat loss over a test area marked by the black rectangle (approximately 200 km$^2$) in Fig. 12 can be estimated. Teigen (2011) estimated that the heat loss from the ocean to the atmosphere on the WSS reaches $\sim 500\text{ W m}^{-2}$ in winter. Using this heat loss over our test area, we estimate that the water masses cool approximately $1^\circ\text{C}$ from the left-hand side (Forlandet section) to the right hand side of the rectangle (the Isfjorden mouth). This is close to the measurements during April 2011 showing a temperature reduction by $2^\circ\text{C}$ over the same distance. It also indicates that the heat loss to the surrounding water masses cannot be neglected but instead accounts for around half the heat loss.

### c. Wind forcing

Large-scale wind systems can force warm AW from the WSC onto the shelf by changing the sea surface elevation on the shelf and the tilting sea surface in conjunction with the WSC. The strength of the along-coast wind is the key parameter that influences the dynamics of the shelf circulation, and the wind direction will vary from month to month. Figure 13 shows a pronounced interannual variability in the along-coast wind stress and the monthly wind stress curl on the southern WSS. We see a southerly wind stress dominance in the time series between 1995 and 2012, and it is especially the winter months November–April that experience the strongest wind stress forcing on the shelf. This is because of the winter cyclones, emanating from the Icelandic low and traversing the Nordic Seas, which are often forced to enter the Fram Strait between Svalbard and Greenland instead of the normal route into the Barents Sea (Zhang et al. 2004; Rogers et al. 2005). Atmospheric blocking caused by high-pressure systems over Scandinavia and Europe (Rogers et al. 2005; Häkkinen et al. 2011) often modify the cyclone pathway, typically causing cyclones to enter the Greenland Sea and thereby inducing a strong southerly flow over Svalbard and bringing warm and humid air masses from the south. Occurrences of winter cyclones in Fram Strait have increased during the last decade because of stronger atmospheric blocking (Häkkinen et al. 2011) and especially during January–February. This occurs prior to a buildup of a higher pressure over Svalbard and the Eurasian basin during the coldest winter months March–April, or before the cyclones go into the Barents Sea where the atmospheric blocking ceases. A shift in wind direction from southerly to northerly is evident in Fig. 13a for some of the winter
months. Winter 2005/06 is a clear example when the wind turned from southerly in December–January to extreme northerly winds in February, as reported in Cottier et al. (2007) (highlighted with black arrows in Fig. 13). A two-layered WSS then changed from a surface water convergence/ pycnocline downwelling to a surface water divergence/ pycnocline upwelling area as shown in Fig. 13b for the calculated wind stress curl field time series over the southern WSS. Cottier et al. (2007) used a mooring at the mouth of Kongsfjorden and Isfjorden (I1 mooring as positioned in Figs. 2, 7) to monitor the water column response to the changing wind stress forcing. Here, we will revisit the forcing mechanisms and use the I1 mooring with a longer time series and clarify the water column response to an along-shelf wind in the Isfjorden Trough and the southern WSS.

The along-slope velocity from the ~50- and ~190-m depths at the I1 mooring are presented in Fig. 14a for the same period as the temperature observations in Fig. 14b. These two depths represent the upper and lower layer in a two-layered representation of the water column, respectively. Both the upper and lower layers are cooled during autumn and winter, but the cooling is often interrupted with intrusion/advection of warmer water. This is visible each year, and during winter 2012, the temperature hardly went below 1°C. Our hypothesis is
that these warm water signals during winter are due to inflow of AW as described by our shelf circulation model and triggered by air–ocean interaction processes. Figure 14 shows that the along-slope velocities at the I1 mooring (Fig. 2a) increase during the winter months (December–March). These are also the months when the along-coast wind stress and the wind stress curl on the southern WSS are found to be strongly positive and negative, respectively (Fig. 13).

A surface water convergence due to a negative wind stress curl increases the sea surface elevation on the shelf and creates downwelling of the water column pycnoclines. Hence, a strong wind stress field over the WSS will change the sea level elevation and change or strengthen the barotropic pressure field on the shelf. This is evident in the current meter time series at mooring I1 (Fig. 14), where the velocity is nearly equal in both 50- and 190-m depths during the winter months. The water column is 210-m deep in this position, and the current vector response indicates a dominating barotropic velocity driven by the sea surface tilt across the shelf. An Ekman transport toward the coastline (eastward) and a negative wind stress curl on the WSS will also influence the surface elevation at the shelf break. The barotropic WSC branch (Teigen et al. 2010) at the shelf break is driven by the positive gradient (positive eastward) in the surface tilt over the shelf break, and a convergence of surface water on the shelf can strengthen this gradient and also move the surface tilt eastward over shallower bathymetry. Figure 10 confirms that the WSC is getting stronger during the winter months and that the current is moving onto shallower areas of the shelf break. Hence, an increase in the sea surface elevation on the WSS can change the characteristic of the
barotropic WSC branch and place warm AW over isobaths that directly guide the warm water into troughs and toward the fjords along West Spitsbergen.

To test the shelf water response to wind forcing, a correlation analysis (Table 4) between the concurrent time series of the shelf wind stress (Fig. 7, black parallelogram) and ocean temperature (Fig. 7, turquoise circle) at I1 is performed for all the years with I1 data. The correlation is performed over an autumn–winter–spring period (between September and May) when the low-pressure systems are strong and can influence the WSS. A method to evaluate the statistical significance of the cross-correlation is presented in appendix A.

For all listed years in Table 4, and in both upper and lower layers, there is a significant negative correlation with a zero time lag between the along-coast wind stress and the ocean temperature. Such a result appears when there is a decrease in temperature in response to southerly wind stress (positive) or a temperature increase as a response to northerly wind stress (negative). Southerly wind will set up a surface onshore Ekman transport. The onshore Ekman transport together with water convergence due to the wind stress curl field create an increased surface tilt toward the coast that will speed up the SPC (blue arrow in Fig. 5) and advect more ArW from the south and into the mouth area of Isfjorden. Furthermore, a negative wind stress curl will force the WSPF density lines (Fig. 6) downward and expose the lower temperature sensor to colder temperatures. When the wind stress turns to northerly, the contrary will occur with a downward-tilting sea surface toward the coast and an upwelling of the WSPF. The resulting (barotropic) surface current is then opposing the (baroclinic) SPC, and as described in Cottier et al. (2007), this can break down the geostrophic control connected to the WSPF and allow a larger transport of warm AW toward the fjord mouth in the lower layer. Hence, a negative along-coast wind stress will also result in a negative cross correlation with the ocean temperature in both layers (Table 4). It should be noted that this two-layer coastal upwelling/downwelling response is not captured by our barotropic model.

Positive time lag for the cross-correlation peaks and the corresponding correlation level are given in the last two columns of Table 4, respectively. Not all the correlations are statistically significant, but the peak in the time-lagged cross-correlation calculations are considered to have a physical significance. All correlations between the along-coast wind and the ocean temperature at I1 are positive and signify a time-lagged advection of warm AW toward the mouth of Isfjorden and I1 after a period of southerly wind forcing. The physical explanation is given by our barotropic model (Fig. 11). After a southerly wind period and corresponding negative wind stress curl field on the WSS (Fig. 7), the barotropic branch of the WSC is forced up the slope over shallower isobaths. This will set up a circulation in the Isfjorden Trough that guides AW along the southern trough slope toward the mouth of Isfjorden. The time lags in Table 4 give an indication of how long it takes for AW to get from the shelf break to the mouth of Isfjorden, and they seem to cluster in two different categories, around 90 and 50 days time lag. When the characteristic WSC velocity $V$ is displaced an eastward distance less than 8 km (Figs. 11a,b), the AW has to circulate around Tampen (Fig. 2), and the circulation distance from Tampen to I1 is approximately 75 km. The model (Fig. 11) gives an average velocity of 0.01 m s$^{-1}$ along the southern slope of the Isfjorden Trough, and it will take approximately 87 days to travel the distance between Tampen and I1. When $V$ is displaced an eastward distance larger than 8 km (Figs. 11c,d), the AW will also circulate along Lexryggen (Fig. 2), which is a much shorter distance to travel, that is, approximately 45 km. With the same average velocity (0.01 m s$^{-1}$) along the slope of Lexryggen, it will only take around 52 days to get AW from the shelf break to I1. These estimates correspond well with the correlation time-lag results in Table 4. Hence, the positive correlation with two distinct time lags between the synoptic wind forcing on the WSS and ocean temperature at mooring I1 reflects two different circulation patterns as demonstrated in the barotropic model. The AW has predominantly taken the shorter Lexryggen route because of a more easterly displaced $V$ in years with a time lag around 50 days. In years with approximately 90 day time lag, the AW has taken the longer route around Tampen, which most likely also represents the normal trough circulation pattern throughout the year with $V_{\text{max}} \sim V$ and a less displaced $V$ over shallower isobaths.

Table 4 shows that the longer route around Tampen was dominating during winter 2010/11, and as explained

<table>
<thead>
<tr>
<th>Year</th>
<th>Water depth (m)</th>
<th>$r$ at $\tau = 0$</th>
<th>$\tau_1$ (days)</th>
<th>$r$ at $\tau = \tau_1$</th>
</tr>
</thead>
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<tr>
<td>2005/06</td>
<td>50</td>
<td>-0.63</td>
<td>54</td>
<td>0.43</td>
</tr>
<tr>
<td>2005/06</td>
<td>190</td>
<td>-0.36</td>
<td>54</td>
<td>0.25</td>
</tr>
<tr>
<td>2006/07</td>
<td>50</td>
<td>-0.40</td>
<td>52</td>
<td>0.18</td>
</tr>
<tr>
<td>2006/07</td>
<td>190</td>
<td>-0.42</td>
<td>49</td>
<td>0.15</td>
</tr>
<tr>
<td>2010/11</td>
<td>50</td>
<td>-0.32</td>
<td>80</td>
<td>0.20</td>
</tr>
<tr>
<td>2010/11</td>
<td>190</td>
<td>-0.40</td>
<td>80</td>
<td>0.30</td>
</tr>
<tr>
<td>2011/12</td>
<td>50</td>
<td>-0.34</td>
<td>97</td>
<td>0.35</td>
</tr>
<tr>
<td>2011/12</td>
<td>190</td>
<td>-0.49</td>
<td>41 and 97</td>
<td>0.25 and 0.16</td>
</tr>
</tbody>
</table>
in the previous sections, this route following deeper bathymetry can deviate from a strict topographic guided current and have an across-slope velocity component. More circulation around deep depressions and shallower ridges are revealed in the numerical model simulations for the longer route circulation pattern (Figs. 11a,b), and the front section collected in April 2011 also indicates a cyclonic circulation around the deep Svensksunddypet. The baroclinic geostrophic velocity referring to the surface pressure is calculated in Fig. 6b and shows a bottom intensified cyclonic circulation pattern in the same area as the model. The surface velocity and the barotropic velocity component is given by the absolute dynamical topography (ADT; see, e.g., Rio and Hernandez 2004), and data from Archiving, Validation, and Interpretation of Satellite Oceanographic Data (AVISO; not shown because of uncertain error estimates close to the coastline) show a depression in the mouth of Isfjorden, indicating a cyclonic circulation above Svensksunddypet. Hence, observations and results from the numerical model show some reconciliation in both standard geostrophic theory and the deviation from such a flow.

d. The AW circulation on the WSS: The Spitsbergen Trough Current

The WSC is the main provider of oceanic heat to the Arctic Ocean, and the heat transport is routed in two branches, the Yermak Branch (YB) and the Svalbard Branch (SB; Fig. 1). Here, we have discussed the modulation of the barotropic WSC branch (Teigen et al. 2010), which becomes the SB on the Yermak Plateau, through atmospheric forcing and corresponding changes in the sea surface height (SSH) on the WSS. Passages of strong cyclones in the Fram Strait pile up water along the West Spitsbergen coastline, which set up topographically guided currents into the troughs indenting the WSS. Hence, some of the AW in the barotropic WSC branch is forced to circulate in these troughs, as demonstrated by our simple shelf circulation model. This represents a longer and slower route of AW toward the Arctic Ocean, as illustrated with a dark red arrow in Fig. 1, and we name this flow the Spitsbergen Trough Current (STC). Our model demonstrates that an acceleration of the WSC (Fig. 10), driven by a change in the SSH on the WSS, starts a massive inflow of AW in the Isfjorden Trough and toward the mouth of Isfjorden. Different forcing situations, and thus, different maximum velocities $V_{\text{max}}$ and horizontal velocity profiles for the WSC, will determine the extent of the STC as shown in Fig. 11. Monthly wind stress forcing on the WSS is large during the winter months (Fig. 13) and the different flow patterns of the STC shown in Fig. 11 can prevail on a monthly time scale.

Since 2006, sea ice production along West Spitsbergen has been reduced, or even nonexistent, in some fjords (Nilsen et al. 2008; Teigen 2011; Onarheim et al. 2014). We argue that this started with low-pressure systems entering Fram Strait with corresponding strong southeasterly wind stress along the WSS during winter 2005/06 (Fig. 13). This brought AW toward the mouth of Isfjorden in accordance with our model. As described in Cottier et al. (2007), this event was followed by a breakdown of the geostrophic control of the WSPF due to a uniform strong northerly wind in Fram Strait during February 2006. A massive inflow of AW and warming of the fjord systems were captured by monitoring programs in Isfjorden and Kongsfjorden. Low-pressure systems returned in Fram Strait in March–April 2006 and resulted in the strongest southerly wind month (April) within our study period (Fig. 13). New pulses of AW on the WSS occurred because of the barotropic ocean response and initiation of the STC. A circulation pattern like in Figs. 11c and 11d may have been initiated since the cross-correlation estimate (Table 4) supports a shorter route along Lexryggen (Fig. 2). A similar wind forcing event took place during winter 2011/12 (not shown). This started the STC, as shown by the temperature time series at I1 (Fig. 14), with the entire water column occupied with warmer water during the winter months.

The STC is always initiated in the Isfjorden Trough after a forcing event, and this inflow of AW toward Isfjorden represents a large heat input to the fjord-shelf system during winter (0.2–0.4 TW). Moreover, the STC represents a slower and longer route of AW toward the Arctic Ocean and serves as a heat loss mechanism for the WSC since the water mass circulates longer in contact with the cold atmosphere and colder Arctic Water. When the STC water masses arrive at the Yermak Plateau during winter (not shown) the temperatures have decreased to below 0°C and represent some of the densest water masses on the shelf because of the high salinity from the AW origin.

The simple shelf circulation model presented in this paper is a barotropic model with a homogenous water column. Stratification is not included, but as described in appendix B, the effect of the surface layer close to the fjord mouth, associated with the WSPF and the SPC, can be included as a boundary condition if we specify an increased relative vorticity along the path of the SPC. Because of conservation of potential vorticity, the cyclonically circulating STC in the Isfjorden Trough will be translated toward deeper water when encountering the boundary with higher relative vorticity (stations 193–194 in Fig. 4).
5. Conclusions

Passages of strong cyclones in the Fram Strait pile up water along the West Spitsbergen coastline that modulates the SSH-driven barotropic WSC branch at the shelf break. This sets up topographically guided currents into the troughs indenting the WSS. Here, this flow is named the Spitsbergen Trough Current (STC). The position of the WSC over the shelf break is controlling the STC. The maximum current of the WSC is, on average, found over the 500-m isobath. If the maximum current of the WSC increases and the WSC is forced eastward over the shelf on the order of 2 km, AW starts to flow onto the shelf in the STC and the warm and saline water is guided into troughs and toward the fjords along the west coast of Spitsbergen, as demonstrated by our simple shelf circulation model. Since atmospheric forcing of the WSS is strongest during the winter months, the intrusion of AW is often largest during these months and plays a decisive role for the sea ice cover in fjords and shelf areas. Moreover, the STC also represents a novel heat loss mechanism for the WSC, as the AW experiences large changes on its longer and slower route toward the Arctic Ocean due to direct heat exchange with the atmosphere and to mixing with colder and fresher water masses on the shelf.

The combination of in situ observations and a shelf circulation model study have revealed the detailed current pattern of the STC in the Isfjorden Trough and the mouth of Isfjorden. The combination of observations and a numerical model have also revealed when the trough circulation follows standard geostrophic theory with topographic guiding and when the circulation deviates from this basic state where flow across isobaths is possible. A nearly homogeneous AW column can be topographically guided toward the mouth of Isfjorden along two routes on the southern slope of the Isfjorden Trough (Fig. 5): 1) the long STC route around Tampen where AW is found to flow along isobaths deeper than 200-m depth and able to be relieved from strict topographic guiding since the relative vorticity term in (10) becomes significant and 2) the shorter STC route over Lexryggen where AW flows coastward along isobaths shallower than 200-m depth. The AW flowing deeper than 150-m depth is able to be topographically guided into the mouth area of Isfjorden, whereas AW shallower than 150 m meets the SPC associated with the WSPF at the mouth of Isfjorden. Here, the STC turns northward and returns westward along the northern side of the Isfjorden Trough, flowing side by side with the SPC. The monthly strength and the curl pattern of the southerly along-coast wind stress on the WSS determine which route will dominate. Both STC routes toward Isfjorden represent a large heat input to the fjord-shelf system during winter (0.2–0.4 TW). Heat flux calculations show that half of the heat loss in the AW in the Isfjorden Trough has to be due to heat loss to the surrounding water masses, while the rest is lost to the atmosphere.

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APPENDIX A

Statistical Significance

The value beyond which a cross correlation in Table 4 becomes statistically significant remains to be determined, and we follow the method proposed by Davis (1976) and Chen (1982). Since the effective degrees of freedom (EDOF) may not be the same as the number of data points that enter into the calculation of the results in Table 4, a special effort is needed to obtain the EDOF for the winters that show large cross-correlation values. The relationship between the autoregressive nature and the EDOF is discussed in some detail by Davis (1976). He presented a method to calculate the EDOF by dividing the number of data points by the integral time scale. The integral time series is found by integrating the product of the discrete autocorrelation coefficients of wind stress and ocean temperature over the time series length. The integrated time scale determines the time period required to gain a new degree of freedom in the estimation of the cross correlation. Applying this method to our data, the integral time scale and EDOF were evaluated for all the winters in Table 4, and critical cross-correlation values for various significance levels can be found from the Student’s t distribution now that the EDOF has been estimated. A significant level of 5% is used for the null hypothesis. Given the correlation coefficient for two time series with a known EDOF, the Student’s t distribution computes the significance level at which the null hypothesis of zero correlation is disproved. If this value is less than 5% the null hypothesis can be rejected and the correlation coefficient is significant. Significant correlation coefficients are made bold in Table 4.
APPENDIX B

Current–Front Interaction

Based on results from the shelf circulation model, combined with the bathymetry data of the WSS, it is assumed that the first dynamical interaction on the WSS between the SPC and the AW circulation in the troughs, the STC, is taking place in the Isfjorden Trough, as illustrated in Fig. 5. The AW that circulates in the deepest part of the southern slope of the Isfjorden Trough, originating from Tampen (Fig. 2), is topographically guided eastward along this slope and continues into the mouth area of Isfjorden below the surface layer (Figs. 5, 6). But a large volume of the AW that circulates in the trough, and most likely the AW that comes from Lexryggen, turns northward in front of the mouth of Isfjorden when it encounters the WSPF and the SPC, that is, after station 43 in Fig. 5. A vertically homogeneous water column is seen on the southern slope of the Isfjorden Trough (Fig. 4). Because of conservation of potential vorticity in (7), the WSPF will be felt as a barrier when the barotropic AW column (from the sea floor to the surface) meets the front. When this AW column is flowing under the SPC, the water column is compressed and shrinks, increasing the potential vorticity in (7). Hence, in order to conserve potential vorticity, the AW column has to go northward over deeper water, follow the WSPF, and cross the mouth area of Isfjorden as a Taylor column. Figure 6 shows the AW column (station 44) up against the WSPF and the northward-flowing SPC in the surface layer. Figures 3 and 4 show that the AW column and the SPC continue to flow side by side in a westward direction along the northern slope of the Isfjorden Trough (stations 191–194).

REFERENCES
