On the Interplay between the Circulation in the Surface and the Intermediate Layers of the Arctic Ocean

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ABSTRACT

The circulation of the Arctic Ocean has traditionally been studied as a two-layer system, with a wind-driven anticyclonic gyre in the surface layer and a cyclonic boundary current in the Atlantic Water (AW) layer, primarily forced remotely through inflow and outflow to the basin. Here, an idealized numerical model is used to investigate the interplay between the dynamics of the two layers and to explore the response of the circulation in each of the layers to a change in the forcing in either layer. In the model, the intensity of the circulation in the surface and AW layers is primarily set by the ocean surface stress curl intensity and the inflow to the basin, respectively. Additionally, the surface layer circulation can strongly modulate the intensity of the intermediate layer by constraining the lateral extent of the AW current on the slope. In contrast, a change in the AW current strength has little effect on the surface layer circulation. The intensity of the circulation in the surface layer adjusts over a decade, on a time scale consistent with a balance between Ekman pumping and an eddy-induced volume flux toward the boundary, while the circulation in the AW layer adjusts quickly to any change of forcing (~1 month) through the propagation of boundary-trapped waves. As the two layers have different adjustment processes and time scales, and are subject to forcing that varies on all time scales, the interplay between the dynamics of the two layers is complex, and more simultaneous observations of the circulation within the two layers are required to fully understand it.

1. Introduction

In low temperature environments, such as the Arctic, ocean density changes are largely determined by salinity variations. As such, the surface layer in the Arctic Ocean comprises fresh and cold water coming from the Pacific Ocean through the Bering Strait, river runoff, an excess of precipitation over evaporation, and sea ice melt. In the Canadian basin, part of this freshwater mass in the surface layer is advected into the Beaufort Gyre and eventually exits into the North Atlantic Ocean along both sides of Greenland through the Davis Strait and Fram Strait (e.g., Coachman and Aagaard 1974; Lique et al. 2010). Below this fresh surface layer, warm and salty water is advected from the North Atlantic to the Arctic Ocean through the eastern part of the Fram Strait and Barents Sea. As it penetrates the Arctic basin, Atlantic Water (AW) descends beneath the fresher, colder mixed layer in the Nansen basin and circulates around the Eurasian and Canadian basins in a cyclonic, topographically steered, pan-Arctic boundary current (e.g., Coachman and Barnes 1963; Aksenov et al. 2011). The deepest part of the Arctic basin is filled with dense water masses formed by brine rejection during sea ice formation on the continental shelves or in the Barents Sea (Jones et al. 1995). According to observations, advection in this deeper layer is very slow (Aagaard 1981), resulting in long residence times of several decades (Ostlund et al. 1987).

The large-scale circulation in the Arctic Ocean has traditionally been considered a system of two independent layers. (Given its sluggish nature, the deepest layer can be seen as nondynamically active and is thus considered at rest in the remainder of this paper.) The forcing mechanisms of the circulation in the two active layers have been examined separately. In the surface layer, the large-scale circulation is primarily wind driven, with an atmospheric high pressure system over the Canadian basin and circulates around the Eurasian and Canadian basins in a cyclonic, topographically steered, pan-Arctic boundary current (e.g., Coachman and Barnes 1963; Aksenov et al. 2011). The deepest part of the Arctic basin is filled with dense water masses formed by brine rejection during sea ice formation on the continental shelves or in the Barents Sea (Jones et al. 1995). According to observations, advection in this deeper layer is very slow (Aagaard 1981), resulting in long residence times of several decades (Ostlund et al. 1987).

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and subsequent downwelling (Proshutinsky et al. 2002, 2009). Observations and model results suggest that the amount of freshwater accumulated in the Beaufort Gyre exhibits large variability on a seasonal-to-decadal time scale, and this variability is thought to be related to variations of the wind stress curl (Proshutinsky et al. 2009; Giles et al. 2012; Karcher et al. 2012).

The forcing mechanisms of the pan-Arctic boundary current within the AW layer remain poorly understood. Results from ocean–sea ice models (Karcher et al. 2007) and a barotropic idealized model (Yang 2005) have suggested that the flow in the AW layer is mainly driven by buoyancy loss over the Barents and Nordic Seas through the injection of positive potential vorticity into the Eurasian basin via the St. Anna Trough and Fram Strait. Such a circulation might also be influenced by the Eurasian basin via the Neptune effect; Holloway 1992; Nazarenko et al. 1998). Although the relative contributions of these different forcing mechanisms still need to be determined, most of these previous studies suggest that it is remote forcing outside of the Arctic basin that explains the variability of the AW layer circulation within the Arctic Ocean.

Within the framework of a coordinated experiment from the Arctic Ocean Model Intercomparison Project (AOMIP; Proshutinsky et al. 2001), Holloway et al. (2007) and Karcher et al. (2007) have analyzed the circulation in several coupled sea ice–ocean models. They have reported that the circulation of the AW layer differs widely across models, with a mean flow that can be cyclonic or anticyclonic depending on the model. Their results suggest that the circulation in the AW layer could be partly driven by the simulated surface circulation, as they found that, in the Canadian basin, models with a too strong Beaufort Gyre and a too deep halocline (compared to observations) have very weak cyclonic or anticyclonic AW flow. Zhang and Steele (2007) also point out the importance of the choice of vertical mixing in numerical simulations for the direction and intensity of the AW circulation. Recently, Karcher et al. (2012) have used observations of iodine-129 isotopes combined with model results to reveal the existence of some variability in the intensity and the pathway of the circulation in the AW layer. They suggest that the interplay between the circulation in the surface and intermediate layers could be a real dynamical feature in the Arctic Ocean rather than just a possible model deficiency; an intensification of the anticyclonic wind-driven circulation in the surface layer would tend to limit or even maybe reverse the AW layer flow in the Canadian basin. However, the details of the mechanisms connecting the two layers remain largely unknown. In these different studies, a complete understanding of the processes that set up the circulation, and in particular the interplay between the two layers, is limited because of the complexity of the general circulation models, which does not allow the authors to isolate and investigate one process at a time. Here, we develop a simple idealized model to investigate the interaction between the circulation in the two layers of the Arctic Ocean. Our study builds upon previous work by Yang (2005) and Spall (2013), who represent the Arctic Ocean as a circular basin to investigate the processes controlling its circulation.

The remainder of this paper is organized as follows: The model setup used for this study is described in section 2, followed by a description of the results from the control run in section 3. The interplay between the circulation in the two layers is investigated as we explore the response of the circulation in each of the layers to a change in the forcing in either layer, looking at both the mean state (section 4) and the transient response to a change of forcing (section 5). Conclusions and discussion are given in section 6.

## 2. The numerical experiment

The Massachusetts Institute of Technology primitive equation general circulation model (MITgcm; Marshall et al. 1997) is used to examine the circulation in the Arctic Ocean. Following the approach of Yang (2005), Spall (2013), and Davis et al. (2014), the domain consists of a circular, semiclosed basin with a 1500-km diameter, connected to a sponge region by a 375-km-wide channel that allows inflow and outflow to the basin. The depth in the interior of the domain is 1500 m, with a slope all around the boundary over a 150-km-wide region (Fig. 1). Although the Arctic basin is up to 4500 m deep at some locations, our basin is certainly deep enough to allow the existence of the two upper, active layers, and our study does not require a deeper basin.

The model configuration is set on an $f$-plane centered at the North Pole ($f_0 = 1.454 \times 10^{-4} \text{ s}^{-1}$). It uses a grid with a horizontal resolution of 15 km and 30 evenly spaced vertical levels of 50 m. Following Spall (2013) and Davis et al. (2014), the internal Rossby deformation radius based on an average salinity of 32.3 in the surface layer and 35.0 in the AW layer (resulting in a reduced gravity $'g'$ of 0.02 m s$^{-2}$) and a surface layer thickness of 400 m is $L_d = \sqrt{(g')/f_0} \approx 20$ km. This is also consistent with the results of Nurser and Bacon (2014), who used a climatology and estimated a ~15-km Rossby radius in the interior of the Arctic basin. Thus, our
model resolution is too coarse to fully resolve the mesoscale activity, and the role of eddies in the Arctic Ocean dynamics is beyond the scope of the present study. The model uses no slip boundary conditions on the sides and the bottom, and a quadratic bottom drag is applied with coefficient $5 \times 10^{-3}$. Horizontal viscosity and diffusivity are set with Laplacian coefficients of 500 and $50 \, \text{m}^2 \, \text{s}^{-1}$, respectively. The vertical viscosity and diffusion are set to $10^{-2} \, \text{m}^2 \, \text{s}^{-1}$. The equation of state from Jackett and McDougall (1995) is used to compute density from temperature, salinity, and depth.

All the simulations presented in this paper are initialized at rest, with uniform temperature and salinity profiles everywhere in our domain that represent average conditions in the interior of the Arctic Ocean (Fig. 1). The profiles are taken from the Polar Science Center Hydrographic Climatology (PHC; Steele et al. 2001) and smoothed to eliminate any detailed vertical structure. Note that our initial temperature profile does not show the subsurface maximum commonly observed in the interior of the Canadian basin (a signature of the Summer Pacific Water; e.g., Timmermans et al. 2014). Our model configuration also does not include a representation of the Bering Strait (and its inflow), the Arctic shelves, sea ice component, any river runoff input, or any atmospheric forcing (except for wind stress). Thus, it does not include the processes required to set and maintain the fresh surface layer and the halocline (Aagaard et al. 1981; Rudels et al. 1996; Anderson et al. 2013). Although of importance, understanding these processes is beyond the scope of our study. Our approach is instead to initialize with simple profiles of temperature and salinity, which exhibit a fresh and cold surface layer lying on top of a warmer and saltier layer and to restore temperature and salinity toward the initial profiles everywhere, with a 10-yr time scale. This time scale results in a restoring force that is weak enough to maintain the stratification without influencing the dynamics within our basin. Sensitivity experiments suggest that our results show very little sensitivity to the choice of the restoring time scale. To absorb any waves propagating out of the domain, the restoring time scale is ramped up to a stronger value (30 days) at the southern boundary, where we also restore velocity toward a zero profile (no restoring force is applied to the velocity outside of this forcing region).

All the experiments are run for 10 yr, which is sufficient to reach a quasi-steady state. For instance, the different integrated quantities examined for the control run to describe the intensity of the transport in each layer (see the following section) do not vary by more than 10% over the following 90 yr of integration.

3. Control run

The goal of our study is to examine the interplay between the circulation in the surface and the intermediate layers of the Arctic Ocean. Although our model configuration is a gross simplification of the real Arctic basin topography and water mass structure, we are able to set up a simulation (labeled as control run hereinafter) that reproduces some of the basic features important for the
Arctic Ocean dynamics, including a two-layer circulation system with characteristics close to those observed in the Canadian basin of the Arctic Ocean. In this section, we first describe the forcing applied to our model and the characteristics of the control run.

a. Forcing

The circulation in the Arctic surface layer is primarily wind driven by the Beaufort high over the Canadian basin, resulting in the anticyclonic Beaufort Gyre. In the ice-covered Arctic, the amount of stress received at the surface of the ocean results from the surface wind stress over the ice-free ocean and the ice–water stress under sea ice, with their relative magnitudes scaled by the ice concentration (Yang 2009). Here, our model setup does not include a sea ice component in order to keep our configuration as simple as possible. We thus follow the approach detailed in Davis et al. (2014), and we prescribe an idealized anticyclonic surface stress centered over the circular basin and directly applied at the ocean surface (Fig. 2).

Along any diameter within the domain, the ocean surface stress in the \( x \) and \( y \) directions is described by

\[
\tau^{(x)} = \sin(\theta) \left[ \frac{1}{r} \int r \cos^2(r) \, dr \right],
\]

and

\[
\tau^{(y)} = -\cos(\theta) \left[ \frac{1}{r} \int r \cos^2(r) \, dr \right],
\]

respectively, where \( \theta \) is the angle that lines connecting each grid point with the center of the domain make with the positive \( x \) axis, and \( r \) is the radial distance between each grid point and the center of the domain (i.e., \( r^2 = x^2 + y^2 \)). Within the channel and sponge region, the stress is described by

\[
\tau^{(x)} = C \left( \frac{y}{r^2} \right),
\]

and

\[
\tau^{(y)} = C \left( \frac{x}{r^2} \right),
\]

where \( x \) and \( y \) are the distances to each grid point from the center of the domain along the \( x \) and \( y \) axes, respectively. The quantity \( C \) is a scale factor that ensures that the curl of the stress field is continuous at the domain/channel boundary. This stress formulation ensures that the curl of the ocean stress field reaches its maximum magnitude in the center of the domain and decreases to zero at the boundaries and in the outflow region (Fig. 2). The stress field is first normalized between 0 and 1. We then set the magnitude of the surface forcing by multiplying the normalized fields by a constant \( A \), which is thus the scaling factor.

**Fig. 2.** Normalized curl of the ocean surface stress (color), with surface stress vectors overlaid in white. The gray box indicates the forcing region where we restore temperature and velocity toward the forcing profiles, which are based on the anomaly profile shown on the right.
factor for both ocean surface stress and the ocean surface stress curl. For the control run, \( A \) is set to 0.02 N m\(^{-2}\), similar to the choice made by Davis et al. (2014) based on surface ocean stress simulated by the state-of-the-art Drakkar model (Lique et al. 2009; Lique and Steele 2012, 2013). In the following, we will refer to \( A \) as the maximum value of the ocean surface stress, but one should keep in mind that it also scales linearly with the surface stress curl.

The circulation in the Arctic intermediate layer is thought to be remotely forced outside of the basin (Yang 2005; Karcher et al. 2007). We choose not to force the flow through open boundary conditions in order to avoid the generation of waves in the domain and possible issues with volume conservation (Marchesiello et al. 2001). Instead, our approach is to strongly restore temperature and zonal velocity forcing profiles, with a 1-day time scale, in a small forcing region (including the slope) indicated in Fig. 2 in order to simulate a flow that propagates into the circular basin as a cyclonic current in the intermediate layer along the slope. Our model configuration does not include the processes that set the depth of the AW in the interior of the Arctic basin and so our forcing profiles are configured to directly force a flow in the intermediate layer (between 200 and 1200 m). The normalized anomaly profile (Fig. 2) has a maximum at a depth corresponding to the temperature maximum of the initial temperature profile (Fig. 1). The temperature and zonal velocity forcing profiles applied in the forcing region are then obtained by scaling this normalized anomaly profile empirically to obtain an inflow through the channel to the main basin comparable with volume and heat transports observed through the Fram Strait. For the control run, the normalized anomaly profile (Fig. 2) is multiplied by \(-0.25\) m s\(^{-1}\) to obtain the velocity forcing profile (the negative sign ensures that the current flows with the slope on its right), while the temperature forcing profile is obtained by multiplying the normalized anomaly profile by 1.5°C and adding it to the initial temperature profile (Fig. 1). We acknowledge that the velocity and temperature forcing profiles applied in the forcing region exhibit much larger values than the temperature and velocity variations observed upstream of the entrance to the Arctic. However, such strong anomalies (compared to the background state in the domain) are required to obtain a flow through the channel comparable to observations at the entrance of the Arctic basin.

As a result of the forcing in the forcing region, the inflow volume and heat transports through the channel (computed along the section shown in Fig. 1) are 7 Sverdrups (Sv; 1 Sv = 10\(^6\) m\(^3\) s\(^{-1}\)) and 35 TW, respectively, for the control experiment. This is roughly comparable to the volume and heat transports of the AW inflow observed through the Fram Strait; from mooring observations, Schauer et al. (2004, 2008) estimate the volume and heat transports to be 12–13 Sv and 16–40 TW, respectively, while Tsubouchi et al. (2012) estimate these transports to be 4.1 Sv and 25 TW, based on an inverse model applied to the Arctic basin. It should be noted that estimation of the transports largely depends on the location of the mooring array, since an important part of the northward transport through the mooring array at 79°N used by Schauer et al. (2004, 2008) recirculates just north of the Fram Strait and does not penetrate further into the Arctic basin. It thus does not contribute to the transport of the boundary current in the high Arctic. Eliminating the transport associated with the recirculation, Marnela et al. (2013) reduce the estimate of inflow transport through the Fram Strait to 4.6 Sv. The inflow through our model channel is also comparable to the transport of the circumpolar boundary current of 5 ± 1 Sv observed in the Laptev Sea by Woodgate et al. (2001).

b. Mean fields

After 10 years of constant forcing, the control experiment reaches a quasi-steady state. Figure 3 shows the sea surface height (SSH) over the circular basin averaged over year 10 of the simulation as well as the normal velocity, salinity, and temperature on the section across the basin indicated in Fig. 1. As expected from the choice of the forcing applied to our model domain, a two-layer system is obtained. In the surface layer, an anticyclonic circulation develops, with a maximum SSH at the center of the domain of 0.6 m and an SSH difference between the center of the gyre and the edge of 0.7 m. This is larger than the SSH difference observed by satellite averaged over the period 1995–2010, with the difference between the center and the edge of the gyre reaching 0.4 m (Giles et al. 2012). Consequently, the model maximum velocities in the surface gyre (\(-9.5\) cm s\(^{-1}\)) are also larger than the geostrophic velocities computed from SSH satellite observations, which range from 1.9 to 5.5 cm s\(^{-1}\), depending on the year considered (Giles et al. 2012). Water is accumulated in the center of the gyre because of the convergence in the horizontal Ekman transport, which in turn leads to downwelling and deformation of the salinity field. In the center of the domain, the isohalines deepen by 50–100 m in the control simulation, which is consistent with the deepening across the Beaufort Gyre found in the PHC (Steele et al. 2001) as discussed in Davis et al. (2014).

Below the surface layer, a cyclonic current is trapped along the slope and carries a strong signature in
temperature. The temperature and velocity maximum are around 400 and 300 m, respectively, with values reaching around 2°C and 5 cm s\(^{-1}\), respectively. On the eastern side of the basin, across the section indicated in Fig. 1, the transport intensity along the slope is \(\sim 2\) Sv, which is comparable to the transport of the AW boundary current observed in the Canadian basin (3 ± 1 Sv; Woodgate et al. 2001). The model temperatures overestimate the AW temperatures observed in the boundary current in the Canadian basin, which are around 1°C (McLaughlin et al. 2009). A signature of the boundary current is also visible in the SSH field, with values increasing again toward the edge of the basin (with a difference across the current of about 0.1 m), while the isopycnals tend to tilt downward. The horizontal velocity field at the depth of the AW current core (not shown) reveals that part of the inflow recirculates just north of the channel and exits the basin again quickly, which explains why the intensity of the boundary current is weaker than the inflow.

c. Sensitivity to vertical mixing and topography

Using the Pan-Arctic Ice Ocean Model and Assimilation System (PIOMAS) ocean–sea ice model, Zhang and Steele (2007) have performed sensitivity experiments in which they vary the value of the vertical mixing. Their results suggest a large sensitivity of the simulated stratification and the intensity of the circulation in the surface and AW layers to the choice of vertical mixing. In their model, they found an optimal value of background diapycnal diffusivity of \(10^{-6}\) m\(^2\) s\(^{-1}\) (using the KPP scheme; Large et al. 1994) to improve the realism of their model results. To test this hypothesis, we perform a set of experiments in which we set the vertical mixing to \(10^{-5}, 10^{-4}, \) and \(10^{-4}\) m\(^2\) s\(^{-1}\), and we compare the results to the control run \((K_z = 10^{-5}\) m\(^2\) s\(^{-1}\)). Results are shown in Fig. 4. When we decrease the level of vertical mixing to lower values than the control run, the results show only little sensitivity and the stratification and the circulation in the surface and intermediate layers are mostly not affected by the change of the level of vertical mixing. On the other hand, when we increase the vertical mixing to \(K_z = 10^{-4}\) m\(^2\) s\(^{-1}\), the stratification is strongly altered. As in the model of Zhang and Steele (2007), the simulation with a strong vertical mixing produces a deeper wind-driven gyre and a less sharp stratification. However, unlike the model of Zhang and Steele (2007), this does not alter significantly the AW circulation in our model, and the intensity of the AW current remains similar (1.9 vs 2 Sv in the control run). Our results are also largely consistent with the study of Spall (2013), who finds that a change of the vertical
mixing only alters significantly the depth of the halocline (and the associated freshwater content) and the intensity of the AW boundary current when the value is on the high end of the range ($K_z$).

Observations suggest that the level of vertical mixing in the Arctic Ocean is low compared to other regions of the global ocean and of the order of $10^{-6}$–$10^{-5}$ m$^2$ s$^{-1}$ (Guthrie et al. 2013; Lique et al. 2014). Within that range, our model shows almost no sensitivity to the choice of $K_z$, so we choose to set the value to $10^{-5}$ m$^2$ s$^{-1}$ for the remainder of this study.

Around the Arctic basin, the steepness of the bathymetry varies considerably. This results in large spatial variations in the velocity of the AW that flows around the basin within a current trapped along the slope (Lique and Steele 2012). The bathymetry chosen for our study has a slope of 100 (1500-m depth change over a distance of 150 km from the coast), which is a reasonable value for the average slope around the Eurasian and Canadian basins. To get some insight into the dependence of our results on the choice of the slope, we run two sensitivity experiments in which we double or divide by two the slope. Results are shown in Fig. 5. Although the forcing is similar to that in the control run, the change of the slope, including in the forcing region, results in different inflow through the channel in the two additional experiments. When the slope is twice as steep as in the control run, the inflow through the channel reduces to 4.4 Sv, and the intensity of the AW current reduces to ~1 Sv, while these quantities increase to 11.7 and 4.2 Sv, respectively, when the slope is half as steep. In the three simulations, the stratification remains the same. The main difference is in the lateral expansion of the AW current on the slope; a steeper slope results in a narrower current. Inversely, when the slope is more gradual, the core of the current expands laterally, limiting the spatial extension of the surface wind-driven circulation and thus reducing its intensity. The SSH in the center of the basin is 0.44 and 0.68 m in the runs with more gradual and steeper slopes, respectively, compared to 0.58 m in the control run.

In summary, although our model remains very idealized compared to the real Arctic and our results depend on some of the specific choices we have made in the design of our model domain and numerics, the control simulation is able to reproduce some of the key dynamical features of the Arctic basin, that is, a fresh wind-driven anticyclonic gyre in the upper layer, with a boundary-trapped cyclonic flow in the AW layer below. We now proceed to use this model to investigate the mechanisms driving the Arctic circulation, where our aim is to understand the processes at play rather than reproducing the exact details of the flow, which are somewhat model dependent (e.g., Karcher et al. 2007).
4. Time-averaged circulation in response to forcing

From the analysis of a subset of the AOMIP models, Karcher et al. (2007) show that the intensity of the AW layer circulation in the Canadian basin is mostly set by the intensity of the vertical flux of potential vorticity (PV), which is related to the local wind stress over that region. This suggests that the intensity of the circulation in one layer (surface or AW) could be modulated by the intensity of the forcing usually associated with the other layer. To investigate this link between the two layers, we perform a series of 12 sensitivity experiments in which we vary the intensity of the ocean surface stress curl (by changing $A$) and/or the inflow through the channel (by changing the temperature and zonal velocity forcing profiles in the forcing region). In this section, we examine the quasi-steady state of the 12 simulations after 10 years of constant forcing. All the results presented in the following are annual averages for the last year of the simulations.

Figure 6 shows the results of the experiments in terms of the intensity of the circulation in each layer versus the intensity of the forcing. The SSH in the center of our basin is considered a proxy for the intensity of the circulation in the surface layer, while the intensity of the circulation in the AW layer in the basin is approximated by the northward transport through the eastern part of the section indicated in Fig. 1. For a given inflow through the channel, the SSH at the center of the basin almost follows a linear relationship when plotted against the intensity of the ocean surface stress $A$. The same is true when we examine the freshwater content in our circular basin or any measure of the surface gyre intensity. This result is in agreement with the previous studies of Proshutinsky et al. (2002, 2009) and Davis et al. (2014), who found a linear relationship between the wind stress curl over the gyre and the freshwater content in the region. For the channel inflow imposed in the control run (7 Sv, circle symbols), the linear fit between the SSH and the maximum ocean surface stress $A$ has a coefficient of 25 m$^3$ N$^{-2}$. In our model, a change of the channel inflow intensity has only a small influence on the intensity of the surface gyre. When the maximum surface stress is kept constant at 0.02 N m$^{-2}$, the maximum SSH only varies between 0.55 and 0.7 m when the channel inflow varies from 0 to 15 Sv (blue symbols).

Similarly, the intensity of the AW current in the basin increases almost linearly with the intensity of the channel inflow for a given surface stress forcing. For the surface stress applied in the control run (0.02 N m$^{-2}$), the linear relationship between the two transports has a coefficient of 0.23 (blue symbols). However, unlike the circulation in the surface layer, the intensity of the circulation in the AW layer is very sensitive to the intensity of the circulation in the other layer. For a fixed inflow of ~7 Sv as in the control run, the intensity of the AW current varies from 1.5 to 7.5 Sv when the surface stress decreases from 0.04 N m$^{-2}$ to zero (circle symbols).

Figure 7 shows the normal velocities along a section crossing the basin, as well as the isohalines 33, 34, and...
34.8, for a subset of simulations in which we vary the surface forcing or the intensity of the channel inflow. For a fixed surface forcing, the AW current on the slope tends to expand spatially in response to an increase of the inflow through the channel, while the stratification remains the same. When the intensity of the AW current is enhanced, the wind-driven gyre is thus slightly reduced in its intensity and its spatial expansion in depth. On the other hand, for a given inflow through the channel, an increase of the surface stress leads to an intensification of the gyre and the Ekman pumping in the surface layer and an increase of the downwelling in the center of the basin. As the slope of the isohalines (or isopycnals) tilts more and more and becomes steeper, the AW current is further constrained on the slope, and hence the integrated transport of the current is reduced, although the maximum velocity in the core of the current remains roughly the same.

We performed sensitivity experiments in which either the surface stress or the velocity and temperature forcings upstream from the channel inflow in the model were switched off. Results are shown in Fig. 8. Although extreme and unrealistic, these two experiments help us shed some light on the processes at play for the Arctic Ocean dynamics. When the channel inflow is switched off, the wind-driven circulation is slightly enhanced compared to the control run, with a SSH in the center of the basin of 0.7 m and deeper penetration within the water column. The velocity section corresponding to that simulation also reveals the existence of a region with velocities close to zero along the slope, corresponding to the region where the AW current develops in the control run. The area of this zero-velocity region likely depends on the intensity of the wind-driven circulation as well as the slope of the bathymetry. When the surface stress is switched off, the intensity of the AW current increases and even exceeds the intensity of the channel inflow (pink points in Fig. 6), likely because of an inertial recirculation in the interior of the basin, which depends on the details of the bathymetry at the northern end of the channel. The velocity section shows that the core of the AW current has a similar shape to

![Figure 6](image-url)

**FIG. 6.** (top left) Time average over year 10 of the SSH in the center of the basin against the maximum surface stress \( A \) and the (top right) inflow through the channel; (bottom left) the intensity of the AW current in the basin against \( A \) and the (bottom right) inflow through the channel. Each point represents a different simulation. The simulations with the same surface stress coefficient are indicated with the same color, while the simulations with the same channel inflow are indicated with the same symbol. The control run is indicated with a filled blue circle.
that in the control run, although the current expands further off the slope and up to the surface (Fig. 8).

Our sensitivity experiments demonstrate that the mean state of the circulation in the surface layer is mostly set by the intensity of the surface stress, with very little influence from the circulation in the AW layer below. In contrast, the intensity of the circulation in the AW layer is influenced by both its remote forcing (the intensity of the channel inflow), which sets the stratification within the slope region, and the intensity of the circulation in the surface layer, which constrains the lateral extension of the core of the AW current. These results support the findings of Karcher et al. (2012), although we could not find any evidence of a regime with reversed AW circulation in our model, even for the case with no inflow through the channel (in this case, the intensity of the AW current reduces to zero). Although we acknowledge that the existence of such a regime might just not be possible in our model setup, evidence from observations in the study of Karcher et al. (2012) was suggestive but inconclusive, and it would be interesting to test this hypothesis using direct observations from current meters.

5. Adjustment process

In the previous section, we have examined the time-averaged circulation in the basin, to which we have applied a 10-yr constant forcing. This represents an Arctic Ocean subject to constant forcing for at least 10 years, allowing a full dynamical adjustment to the forcing. In reality, the Arctic Ocean is subject to forcing that varies on all time scales. As explained in greater detail by Davis et al. (2014), the intensity of the stress applied at the surface of our basin results from both the wind forcing and the sea ice conditions in the Canadian basin. Both the winds and the sea ice conditions exhibit variations on seasonal-to-decadal time scales (Yang 2006, 2009; Proshutinsky et al. 2009; Spreen et al. 2011), resulting in large variations of the ocean surface stress. Similarly, moorings deployed through the Fram Strait or at the entrance of the Canadian basin near the Lomonosov Ridge capture large variations of the transport intensity on all time scales (Schauer et al. 2004, 2008; Beszczynska-Möller et al. 2012; Woodgate et al. 2001). For our model setup, this corresponds to variations affecting the inflow through our channel.
As the forcing of both the surface and the intermediate layers exhibits large variability, it is interesting to examine the adjustment time scales and processes involved in the response in each layer to a change of forcing in one or the other layer (or both). In this section, we analyze some transient experiments. All the simulations start from the end of year 10 of the control run. We then modify instantaneously the intensity of one or the other forcing (or both), run the simulations for 10 additional years, and examine the transient adjustment. For clarity, we only present results from simulations in which the surface stress and/or the channel inflow are increased. Simulations with reduced intensity of the forcing have also been run and exhibit reverse behavior (i.e., the model behaves linearly).

At the initial stage of all our experiments, any change of forcing (surface stress and/or inflow) leads to a redistribution of volume. After 3 days (which is our output frequency), the net volume flux through the channel reaches its maximum deviation from zero. When the surface stress is kept constant and the inflow through the channel is modified, the net volume flux through the channel (which is zero when the basin is adjusted) converges toward zero in \( \frac{1}{10} \) days. When the surface stress is modified, the net transport through the channel takes \( \frac{1}{50} \) days to reach equilibrium. This adjustment corresponds to a redistribution of volume between the circular basin and the region below the channel, which occurs initially through the propagation of gravity waves (Luneva et al. 2012).

For comparison, we also run an additional experiment in which the surface stress and the channel inflow remain the same as in the control run for 10 additional years. In that case, although none of the forcing applied to our basin includes any time variability, the intensity of the AW circulation in the basin exhibits some intrinsic variability (Fig. 9). The circulation in the surface layer exhibits almost no intrinsic variability, regardless of the proxy chosen to define the intensity of the surface circulation. The amount of intrinsic variability affecting the intensity of the AW circulation appears to depend on the intensity of the surface layer circulation, with a tendency for reduced variability when the wind forcing increases (Fig. 9). More analysis is required to fully understand the mechanism that underlies this intrinsic variability, which is related to small-scale structure within the AW current core along the slope because of the existence of eddies, meanders, and shifts of the current core, which develop over time.

Figure 10 shows snapshots of velocity anomalies after 15, 90, and 360 days in the transient experiment in which the channel inflow is kept constant, while the surface stress is enhanced at day 360 from 0.02 to 0.04 N m\(^{-2}\). Initially, for the first 40–50 days, the velocity anomaly is nearly barotropic and extends over the whole basin. The same signal can be seen in Fig. 9 (blue line); the SSH in the center of the basin shows an initial fast adjustment, corresponding to the volume adjustment in the circular basin. At that stage, the velocity field shows negative
anomalies over the whole eastern side of the basin, resulting in a reduction of the AW transport on the slope. This can be seen in Fig. 9, with a reduction of the AW transport to 1.3 Sv over ~50 days in response to the increased wind-driven circulation. After that initial stage, the velocity anomalies develop some baroclinic structure, and the SSH at the center of the basin adjusts over a longer time scale. The longer adjustment time scale is consistent with a balance between the wind-induced Ekman pumping and the eddy-induced volume flux toward the edge of the gyre, as discussed in the study of Marshall et al. (2002) of the dynamical equilibrium for a fresh or warm lens. Using a reduced-gravity model with a similar setup (same basin and same surface forcing with a maximum of 0.02 N m$^{-2}$), Davis et al. (2014) find that the balance between the Ekman pumping and the eddy flux occurs on a ~14-yr time scale. In our model, the input of vorticity by the surface stress is redistributed by the partially resolved mesoscale eddy field and ultimately dissipated through lateral and bottom friction. This results in the majority of the adjustment occurring over ~10 yr, although a small drift remains in the system even after 100 yr of simulation (not shown).

We now examine the adjustment of the basin in response to a change of inflow through the channel. Figure 11 shows snapshots of velocity anomalies after 6, 21, 90, and 360 days in a transient experiment in which the surface stress is kept constant, while the channel inflow increases from 7 to 11.7 Sv. Given our experimental setup, the channel inflow is set through a modification of the restoring velocity profile in our forcing region south of the channel, so that the channel inflow takes ~15 days to reach its new steady-state value (Fig. 9a, green line), although an anomaly is first seen after only 3 days. This is likely the time required by the fastest waves to propagate along the slope from our forcing region to the section across the channel where the inflow transport is...
computed. Velocity anomalies appear on the eastern side of the cross-basin section after 6 days and on the western side of the section after 21 days. The anomalies are nearly barotropic and mostly confined on the slope, within the core of the AW current. Given the stratification and the geography of our basin set on a $f$ plane, we estimate the Burger number $Bu = (L_d/L_c)^2$ (with $L_c$ being a characteristic length scale of the topography) to be $\sim 0.02$. For $Bu \ll 1$, we expect an adjustment through barotropic, topographic Rossby waves that propagate along the slope (with shallower depths on the right-hand side) at a speed of a few meters per second (Rhines 1970; Wang and Mooers 1976). Here, the anomaly propagates from the eastern side to the western side of the basin within 15 days, corresponding to a propagation speed of $\sim 3.5$ m s$^{-1}$. Within 30 days, the integrated AW transport has reached its maximum value, although a large amount of intrinsic variability affects the AW transport time series afterward (Fig. 9b) because of small-scale structures and shifts in the location of the current core (Fig. 11). The increase of the channel inflow also slightly affects the SSH in the center of the basin (compared to the control run; Fig. 9c). During an initial short period, the increase of the channel inflow leads to a SSH increase corresponding to the volume adjustment between the two sides of the channel. Afterward, the SSH in the center of the basin gradually decreases. Most likely, an intensified AW current provides an additional source of dissipation on the edge of the gyre at depth, resulting in a reduction of the gyre intensity and thus a smaller SSH in the center of the basin. After an initial fast adjustment, the intensity of the flow in the AW layer exhibits a slow declining trend during the entire simulation, which is likely the imprint of the decadal adjustment time scale of the gyre in the surface layer.

Last, we investigate a case in which both the surface stress and the channel inflow are increased (Fig. 12). Velocity anomalies along the cross-basin section exhibit a mixed pattern from the two preceding cases, with an initial barotropic response to the change in surface stress, on which the propagation of velocity anomalies with opposite signs on the slope is superimposed. Interestingly, the response to a change of the intensity of both the channel inflow and the surface stress is similar to the sum of the responses from changing each forcing independently, suggesting a strong linearity in the adjustment process. Compared to the case in which only the channel inflow is increased, the velocity anomalies have similar amplitude but are this time more constrained on the slope. This results in a smaller AW transport (Fig. 9b, red line). For the two transient experiments in which the channel inflow is

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**Fig. 10.** Results from the transient experiment in which the surface stress is increased from 0.02 to 0.04 N m$^{-2}$, while the channel inflow is kept constant. (a) Snapshot of the velocity (m s$^{-1}$) and (b)–(d) velocity anomalies [m s$^{-1}$; the anomalies are relative to panel (a)] along the section indicated in Fig. 1 at different times. The black lines correspond to the zero contour. Isohalines 33, 34, and 34.8 are indicated in white.
modified, Fig. 9b shows that the initial adjustment leads to a maximum value of AW transport after \( \sim 50 \) days, followed again by a decrease.

For a fully (or very nearly) adjusted flow, as in the simulations analyzed in section 4, the downward-propagating influence of the wind-driven circulation on the AW circulation is much larger than the upward-propagating influence of the AW current on the surface layer circulation. However, our analysis of the adjustment to a change in forcing reveals that the adjustment time scale in the surface layer (\( \sim 14 \) yr) is much longer than the adjustment time scale of the AW circulation (\( \sim 1 \) month). It is thus not trivial to determine the imprint of the time variability of one layer onto the other layer. Further investigation of the interaction between variability in the two layers will be the subject of a forthcoming study, using a similar framework.

6. Summary and discussion

Compared to other regions of the globe, the dynamics that govern the circulation in the Arctic Ocean have been examined in a very limited number of studies and remain poorly understood. In the present study, a simple numerical model has been developed to examine in detail what sets the circulation of the surface and the intermediate layers, with a focus on the interplay between the two. Despite the very idealized nature of our setup, the model is able to reproduce reasonably well some key features of the Arctic Ocean circulation and in particular the two-layer circulation system found in the Canadian basin with an anticyclonic gyre in the surface layer forced by an ocean surface stress and a cyclonic topographically steered boundary current in the intermediate layer remotely forced by an inflow to the basin.

Results from sensitivity experiments in which we vary the intensity of one or the other (or both) forcings (and apply this forcing constantly for 10 years) reveal that, in the model, the strength of the circulation in the surface and AW layers is linearly linked to the ocean surface stress curl and the inflow to the basin, respectively. Additionally, the circulation in the surface layer can strongly modulate the strength of the circulation in the intermediate layer by constraining the lateral extent of the AW current on the slope. In contrast, a change of the AW current has little effect on the circulation within the surface layer. The downward influence of the wind-driven surface gyre on the AW circulation is consistent with the previous studies of Karcher et al. (2007, 2012), although our model does not exhibit a reversed AW circulation, even in the case of zero AW inflow to the basin. Based on numerical simulations performed with a similar idealized basin with different forcing, the study of Spall (2013) suggests that the AW boundary current is primarily forced by the salinity contrast between the salty AW and the upper freshwater layer. In his model setup (which does not include an anticyclonic wind stress

Fig. 11. As in Fig. 10, but for the transient experiment in which the channel inflow is increased from 7 to 11.7 Sv, while the surface stress is kept constant.
curl), the deepening of the halocline in the center of the basin is maintained by a lateral eddy salt flux from the boundary current, balanced by vertical diffusion in the interior of the domain. In turn, this balance controls the intensity of the AW boundary current. Our results also suggest a strong link between the depth of the surface layer (and the tilt of the isopycnals) and the intensity of the boundary current, although, in our model configuration, the thickness of the halocline is mostly set by the intensity of the surface stress curl. Further investigations are required to reconcile our model framework with that of Spall (2013) in order to better understand the full dynamics at play in the Arctic basin.

The full adjustment of the intensity of the circulation in the surface layer of our model to a change of forcing occurs over a decade on a time scale consistent with a balance between Ekman pumping and an eddy-induced volume flux toward the boundary discussed by Davis et al. (2014). On the other hand, the circulation in the AW layer adjusts quickly to any change of forcing (~1 month) through the propagation of boundary-trapped waves. In reality, both the ocean surface stress and the inflow of AW through the Fram Strait exhibit large variations on monthly to decadal time scales (Proshutinsky et al. 2009; Schauer et al. 2008). As the dynamics in the surface and intermediate layers adjust to a change of forcing on different time scales, determination of the possible imprint of the time variability of the circulation in one layer onto the circulation in the other layer is not straightforward. Our study calls for a simultaneous analysis of the time variations affecting the circulation in the surface and intermediate layers, based on realistic simulations and observations, since it is clear from our study that the strength of the circulation in one layer is potentially important in setting the strength of the circulation in the other, especially for the case of the AW layer in which the circulation appears to be strongly influenced by the upper layer.

The recent studies of Tsamados et al. (2014) and Martin et al. (2014) have suggested that the ongoing decline of the Arctic sea ice cover (Stroeve et al. 2012) has implications for the transfer of momentum from the wind through the sea ice pack, leading to an increase of the ocean surface stress. This increase of the ocean surface stress could result in an intensification of the Beaufort Gyre and an increase in the amount of freshwater accumulated in the Canadian basin of the Arctic Ocean (Giles et al. 2012; Davis et al. 2014). Our results suggest that the changes affecting the Arctic sea ice conditions might also have strong consequences for the circulation in the AW layer, as an intensification of the surface layer would tend to reduce the intensity of the AW boundary current in the Canadian basin. This reduction of the AW current might also have consequences for the heat budget of the Arctic Ocean, limiting the penetration of heat into the Canadian basin (Lique and Steele 2013; Itkin et al. 2014). This effect could potentially counterbalance the temperature increase of the
AW inflow through the Fram Strait (Polyakov et al. 2005) and represent a negative feedback for the ongoing sea ice melting (Polyakov et al. 2010).

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