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# A Lagrangian view of Pacific water inflow pathways in the Arctic Ocean during model spin-up

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A B S T R A C T

In this study, we identify the routes of Pacific water within the Arctic Ocean using velocity fields, derived from the spin-up of a numerical model, and representing different circulation states within the basin. Lagrangian analysis shows there are two major routes of Pacific inflow water circulating in the Arctic Ocean, a Transpolar route and an Alaskan route. Those two routes transport more than 70% of the Pacific water, ~50% of which flows through the central Canadian Arctic Archipelago (CAA), to the Atlantic. The outer edge (close to the coast) of Pacific inflow water routes, especially within the interior basin, is close to the 7 m isopleth of the upper (above 227 m) freshwater content. The proportion of Pacific water flowing along the two routes significantly changes with the spatial distribution of freshwater within the Canadian Basin. When more freshwater occupies the Beaufort Gyre (during the 5th year of spin-up), almost all the Pacific water entering the central CAA is from the Transpolar route. However, with a much weaker (flattened) Beaufort Gyre due to the loss of a significant amount of freshwater, ~65% of the Pacific water entering the central CAA is from the Alaskan route, resulting in younger Pacific water reaching the central CAA. Thus, we propose that not only the amount of freshwater but also its spatial distribution within the Canadian Basin play an important role in the Arctic Ocean circulation system, although the total volume transport (~0.35 Sv, 1 Sv = $10^6$ m$^3$ s$^{-1}$) of Pacific water through the central CAA shows little variation with time.



## 1. Introduction

Pacific water inflow through the Bering Strait provides the Arctic Ocean with large amounts of heat, freshwater and nutrients, playing an important role in Arctic oceanographic and ecological processes. Seasonal and inter-annual variability of this inflow may act as a possible trigger to the start of seasonal sea ice melting and explains much of the western Arctic sea ice reduction in the past decade (Woodgate et al., 2006, 2010; Shimada et al., 2006). It feeds about one third of the total freshwater input (8450 km$^3$ year$^{-1}$) into the Arctic Ocean, directly affecting the structure of the halocline, especially within the Canadian Basin (Bauch et al., 1995; Woodgate and Aagaard, 2005). In the highly productive western Arctic shelf, it is the primary source of nutrients during ice melting and the open-water period (Coachman et al., 1975; Grebmeier et al., 2006).

Pacific water enters the Arctic Ocean through Bering Strait forced by a meridional sea level gradient (Coachman and Aagaard, 1966; Aagaard et al., 2006). The mean annual northward volume transport is about 0.8 Sv (1 Sv = $10^6$ m$^3$ s$^{-1}$), with high seasonal and inter-annual variations (Roach et al., 1995; Woodgate and Aagaard, 2005; Woodgate et al., 2005b, 2006). The circulation of Pacific water within the Chukchi Sea is dominated by geostrophic balance with three major pathways (Herald Valley in the west, Barrow Canyon in the east and the Central Channel in the middle) (Weingartner et al., 2005; Woodgate et al., 2005a; Pickart and Stossmeister, 2008; Panteleev et al., 2010; Winsor and Chapman, 2004; Spall, 2007). Pacific water will travel for five to nine months undergoing physical and biochemical modifications before reaching the Canadian Basin (Panteleev et al., 2010). Using a regional eddy-resolving coupled sea ice-ocean model as well as satellite data, Watanabe (2011) investigated the Beaufort shelf break meso-scale eddies and the shelf-basin exchange of summer Pacific water, and concluded that both meso-scale eddies and wind-driven Ekman transport play roles in conveying Pacific water off the shelf.

The fraction and distribution of Pacific water within the Arctic Ocean have been extensively studied using chemical tracers and nutrients. Silicate (Jones and Anderson, 1986; Bauch et al., 1995) and $PO_4^*$ (Broecker et al., 1998; Ekwurzel et al., 2001) concentrations have been used to distinguish Pacific inflow water from other water masses in the Arctic Ocean, however, their concentrations could be significantly affected by biological processes, particular during the ice-free season. Thus, Jones et al. (1998) proposed that

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the ratio of nitrate versus phosphate (N:P) could be used to identify between waters of Pacific and Atlantic origin. Utilizing this N:P relationship, Pacific water was found to be the major source of freshwater in the Canadian Basin, particularly at depths ranging 50–200 m (Jones et al., 2008; Yamamoto-Kawai et al., 2008; Carmack et al., 2008). Also the inter-annual variation of Pacific water influence in the central Arctic was revealed by observations (Alkire et al., 2007; Alkire et al., 2010). For the circulation, Jones et al. (1998) found the surface Pacific water might split into two branches near the Chukchi Plateau. One branch flows eastward along the northern Alaska coast, the CAA and Greenland, then exits through Fram Strait. The other branch, after mixing with Atlantic source water, enters the deep basin along the Mendeleyev Ridge, recirculating in the Canada and Makarov Basins, then exits these basins across the Lomonosov Ridge north of Greenland, and flows out of the Arctic Ocean through Fram Strait as well. Deeper Pacific water along the second route was also shown in later observations (Jones et al., 2008). Steele et al. (2004) studied historic temperature and salinity profile data within the Arctic Ocean, and found similar pathways of summer Pacific water within the Arctic Ocean when the Arctic Oscillation (AO) index is high. But when the AO index is low, the Pacific inflow water might be all entrained into the Beaufort Gyre (Steele et al., 2004). However, observations are still sparse in space and time, and systematically direct measurements of currents under the permanent sea ice are still not available yet. Thus it limits our understanding of the pathways of Pacific inflow.

Using a Lagrangian analysis, Lique et al. (2010) calculated the horizontal mass stream function based on the monthly velocity fields from a coupled ocean sea ice global model averaged over 1980–2001. They obtained similar patterns to Steele et al. (2004), but had almost all the Pacific water exiting the Arctic Ocean through the CAA channels, while observations (Jones et al., 2003) indicate that the Pacific water exported through Fram Strait could reach as far as 66°N. Thus, more work is needed to better understand the pathways of Pacific inflow water in the Arctic Ocean. Identifying the pathway of Pacific water within the Arctic Ocean will be helpful in knowing its fate and the associated impacts on the Arctic and subarctic system. In addition, the build-up and release of freshwater storage within the Arctic Ocean is dominated by large-scale atmospheric circulation variations (e.g., AO) (Proshutinsky et al., 2002; Häkkinen and Proshutinsky, 2004; Condron et al., 2009; Jahn et al., 2010b). Whether and how the Pacific water pathways will be impacted are still an open question. Also, do the two main routes transport similar amounts of Pacific water? If not, which one contributes more and what are the mechanisms? How sensitive are the answers obtained from numerical models to technical issues of spin-up and freshwater content drift?

To investigate the pathways of Pacific water, as well as its variability associated with different freshwater distributions within the Arctic Ocean, we use a Lagrangian method to track Pacific water based on velocity fields produced during the spin-up of a numerical simulation. We will first provide the model configuration and forcing data in Section 2, and then present the Pacific water inflow pathways based on velocity fields from year 5 in Section 3. The variation of Pacific water inflow pathways and distribution of freshwater content within the Canadian Basin is investigated as well in this section. Summary and discussion are given in Section 4.

## 2. Methods

### 2.1. Model description and configuration

In this study, we used a pan-Arctic regional configuration of the Nucleus for European Modeling of the Ocean (NEMO) numerical framework version 3.1 (Madec and the NEMO team, 2008). This coupled numerical model includes a three-dimensional (3D), linear free surface, hydrostatic, primitive-equation ocean generation circulation model and a dynamic-thermodynamic sea ice model, the Louvain-la-Neuve sea-ice model (LIM2) (Fichefet and Maqueda, 1997), with an elastic–viscous–plastic (EVP) ice rheology (Hunke and Dukowicz, 1997).

In the ocean module, the momentum equations are expressed in their invariant formulation (vorticity term + gradient of kinetic energy + vertical advection). An energy–entropy conserving scheme (Arakawa and Hsu, 1990), which conserves total energy and entropy only for horizontally non-divergent flow, is used to discretize the total vorticity term. The advection of tracer is formulated by the Total Variation Dissipation (TVD) scheme (Lévy et al., 2001). Parameterizations for the subgrid-scale processes include (i) a rotated Laplacian isopycnal scheme for lateral tracer diffusion with a maximum eddy diffusivity of 300.0 $m^2\ s^{-1}$ (proportional to the grid size); (ii) a horizontal bilaplacian momentum diffusion with an maximum eddy viscosity of $-1.5 \times 10^{11}$ $m^4\ s^{-1}$ (proportional to the cube of the grid size); (iii) a one and half order turbulent kinetic energy (TKE) closure scheme for vertical mixing combined with an enhanced vertical diffusion. The vertical eddy viscosity and diffusivity are chosen to be $1.0 \times 10^{-4}$ $m^2\ s^{-1}$ and $1.0 \times 10^{-5}$ $m^2\ s^{-1}$, respectively, and are enhanced to 10 $m^2\ s^{-1}$ if the Brunt–Vaisala frequency ($N^2$) is less than $-1.0 \times 10^{-12}$ $s^{-1}$.

To avoid the coordinate singularity (convergence of the meridians) near the North Pole in the standard latitude–longitude spherical grid, an orthogonal transformation method proposed by Murray (1996) is adopted to generate the horizontal model grid. The model domain covers the northern Bering Sea, the Arctic Ocean, the CAA, Nordic Seas and part of the North Atlantic Ocean (see Fig. 1, left) with a variable horizontal resolution of $\sim$ 11 km in the central CAA and $\sim$ 15 km in the Arctic Ocean. There are 46 levels in the vertical with layer thickness increasing smoothly from about 6 m at the surface to around 240 m at the bottom. The bathymetry data is derived from the global 1 min resolution relief dataset (ETOPO1,Amante and Eakins, 2009) provided by US National Geophysical Data Center (NGDC). Hanning smoothing with a minimum depth of 6.25 m is applied four times after the linear interpolation. Small islands, e.g., Diomede Islands at Bering Strait, are not resolved. Also narrow passages, such as Coronation Gulf and the adjacent waterways, are widened to have at least two tracer points, and Nares Strait is set to have at least three. Two open boundaries (Pacific and Atlantic) are used in our configuration (see Fig. 1, left).

### 2.2. Experiment setup

The atmospheric forcing data (10-m surface wind, 10-m air temperature and humidity, downward longwave and shortwave radiation, total precipitation and snowfall) come from the normal year data of the version 2 Coordinated Ocean-ice Reference Experiments dataset (CORE, Large and Yeager, 2004). The ocean starts from rest with initial 3D temperature and salinity fields from the Polar Science Center Hydrographic Climatology (PHC3.0, Steele et al., 2001). Initial ice (concentration: 0.95, thickness: 3.0 m) and snow (thickness: 0.5 m) are prescribed where sea surface temperature is close to the freezing point. The lateral ocean components (normal and along boundary velocities, temperature and salinity) are interpolated onto our model grids from the monthly averages over the period 1979–2004 of a global ocean simulation, ORCA025-KAB001 (Barnier et al., 2006). The inflows (outflows) at the Pacific and Atlantic open boundaries are adjusted linearly (proportional to the open boundary area) to keep the volume conserved. Temperature and salinity is restored within the buffer zone (shaded rectangles in Fig. 1, left) with a time scale of 40 days