1. Introduction

Driven by the polar easterlies over the Antarctic coastal regions (Large and Yeager 2009), the Antarctic Slope Current (ASC) is a roughly coherent westward current encircling Antarctica (Thompson et al. 2018; Thompson et al. 2020), with a typical velocity of 0.1–0.3 m s$^{-1}$. Coupled with the Antarctic Slope Front (ASF) (Jacobs 1991), the ASC directly regulates the cross-slope heat transport that increases the heat content of Antarctic shelf waters (Dinniman et al. 2011; Heywood et al. 2014; Schmittko et al. 2014) and enhances the basal melting of floating Antarctic ice shelves (Pritchard et al. 2012; Rignot et al. 2014). According to the comprehensive review of Thompson et al. (2018), three different types of the ASC/ASF are classified based on the dynamic mechanisms responsible for the cross-slope and along-slope structures: 1) Fresh Shelf, corresponding to relatively fresh and cold shelf water; 2) Dense Shelf, corresponding to relatively saline and cold shelf water; and 3) Warm Shelf, corresponding to relatively warm shelf water. Over Warm Shelf regions in West Antarctica, the rapid thinning of the ice shelves (Pritchard et al. 2012) has been confirmed to be correlated with the enhanced cross-slope exchanges and the increase in the heat content of the shelf waters (Rignot et al. 2013; Shepherd et al. 2018).

In contrast to West Antarctica, the ASC encircles the East Antarctic shelf and is mostly categorized as Fresh Shelf regions, and the coupled ASF acts to isolate the cold shelf waters from the warm modified Circumpolar Deep Water (mCDW) offshore (Whitworth et al. 1998). The warm mCDW in the deep ocean flows southward along the density surfaces approaching the continental shelf until it rises to a depth comparable to the continental shelf (Meredith et al. 2011). Off Fresh Shelf regions, the ASF is characterized by isopycnals that tilt downward and poleward toward the continental slope (Orsi and Whitworth 2005; Peña-Molino et al. 2016). Over the slope, density surfaces directly intersect the seafloor, and thus mCDW is blocked from traveling further south due to the constraint of the conservation of potential vorticity (PV; Pedlosky 1987). However, in certain locations around East Antarctica, warm mCDW is allowed to ventilate the continental shelf in the presence of submarine troughs (Fretwell et al. 2013; Greenbaum et al. 2015; Herráiz-Borreguero et al. 2015; Rintoul et al. 2016; Liu et al. 2017; Nitsche et al. 2017; Silvano et al. 2017, 2019; Nakayama et al. 2021), such as the Sabrina Coast (110°–125°E) and Prydz Bay (60°–90°E). As the westward
flowing ASC encounters the longitudinal variations in the seafloor depth, such as a trough intersecting the continental shelf break, the dynamic influences of topography on the ASC can modify the ASC structure and accommodate cross-slope density surfaces. The onshore extension of the isopycnals in the trough enables warm mCDW to be advected across the shelf break in a dense inflow layer. Therefore, submarine troughs provide a pathway for the cross-slope transfer of mass and tracers (Moffat et al. 2009; St-Laurent et al. 2013; Zhang et al. 2016). Physical mechanisms that govern the mCDW intrusion at the submarine troughs are the primary focus of our investigation.

A variety of mechanisms have been proposed to be responsible for cross-slope exchanges. Klinck and Dinniman (2010) described six dynamics for CDW intrusion: 1) bottom Ekman transport, 2) the inertial overshoots where the isobaths curve in front of the flow (Dinniman et al. 2003; Dinniman and Klinck 2004), 3) vertical excursion of isopycnals due to the strengthening of the Antarctic Circumpolar Current (Klinck and Dinniman 2010), 4) eddy-mediated transport (Nast et al. 2011; Zhang et al. 2011a; Stewart et al. 2018), 5) atmospherically forced exchanges (Thoma et al. 2008), and 6) a buoyancy-driven flow induced by ice shelf basal melting (St-Laurent et al. 2013). The second mechanism regarding inertial overshoots can be physically recognized as the dynamic influences of topography on the ASC. Based on an idealized model representative of the Bellingshausen Sea, St-Laurent et al. (2013) further summarized three dynamic mechanisms responsible for onshore flows within a coastal trough: 1) melt-driven flow inside the trough, 2) topography Rossby waves (TRWs) induced by mean flow, and 3) wave-topography interaction (Zhang et al. 2011b). The relative importance of the latter two mechanisms depends on the strength of the ASC and its distance to the shelf break (St-Laurent et al. 2013). Therefore, the critical implications of submarine troughs on the cross-slope transfer of masses and tracers have been highlighted. However, the dynamic mechanisms that govern the onshore transport within submarine troughs over Fresh Shelf regions in East Antarctica are still not clear.

While the changes of onshore heat transport at a trough in response to the strength and position of the ASC have been discussed extensively in previous literature (Dinniman et al. 2011; St-Laurent et al. 2013), the dynamic mechanisms that drive the onshore flow and govern the vertical structure and temporal evolution of the onshore flow still need to be further clarified. For a barotropic ocean without stratification, the currents intend to flow along the isobaths by following the Taylor–Proudman theorem. However, the dynamic effects of stratification on the currents overshooting to a submarine trough over Fresh Shelf regions are still not clear. In particular, the strength of the ASC has a pronounced seasonal cycle with maximum transport in the austral winter (Mathiot et al. 2011; Armitage et al. 2018), yet how the onshore transport at a submarine trough responds to the seasonality of the ASC remains unexplored. These motivate the need for an improved mechanistic understanding of dynamic processes that determine the transport across the continental slope and toward the ice shelves within a submarine trough. This study proceeds to a more in-depth analysis of the dynamic mechanisms that govern the onshore mCDW intrusion within a typical trough over a Fresh Shelf case.

The paper is organized as follows. Section 2 describes the high-resolution, eddy-resolving process model configuration representative of the ASC off Fresh Shelf regions around East Antarctica. Section 3 explores the mechanisms responsible for the dominant forcing, the vertical structure, and the seasonal evolution of the onshore flow within a trough. Section 4 summarizes and provides a discussion of the results.

2. Model description and experimental design

a. Model description

The numerical simulations are conducted with the Massachusetts Institute of Technology General Circulation Model (Marshall et al. 1997), including an ice shelf component (Losch et al. 2010). To represent an idealized Fresh Shelf segment in East Antarctica, the model domain is placed at 80°–90°E, 65°–68°S, including a deep ocean area, a steep continental slope, a shallow continental shelf, a submarine trough, and a sub-ice-shelf cavity (Fig. 1a). The geographic coordinate in this idealized model only represents a typical configuration of a submarine trough in East Antarctica but not a specific location. The primary model parameters are shown in Table 1.

The horizontal grids are placed on a spherical orthogonal projection, with 800 × 600 horizontal bins. Accordingly, the zonal grid spacing ranges from ~520 m at the southern boundary (68°S) to ~587 m at the northern boundary (65°S), and the meridional grid spacing is a constant of ~555 m. Such a high horizontal resolution allows the model to properly resolve mesoscale processes over the Antarctic continental shelf region where the Rossby deformation radius decreases to ~5 km (Chelton et al. 1998). The seafloor depth is zonally uniform and range from 3000 m in the northern deep ocean to 500 m over the southern continental shelf, separated by a steep slope at 66°–66.5°S ( ~5.5 km wide; Fig. 1b). A meridional trough (40 km wide and 700-m maximum depth) perpendicular to the slope extends from the shelf break to the sub-ice-shelf cavity, approximately following the submarine channels in Prydz Bay and the Sabrina Coast. The trough mouth is defined as the northern extension of the trough that extends to the north of shelf break (84.5°–85.5°E, 66°–65.8°S). The static and thermodynamically active ice shelf draft is 100 m thick at the ice calving front (67°S) and gradually increases to 500-m depth at the grounding line (68°S). Such a geometry configuration generally matches the typical topography in the Sabrina Coast and Prydz Bay (e.g., the Reynolds Trough, the Prydz Channel, and the Amery Depression).

The vertical discretization of the model has 100 levels. To satisfactorily resolve the vertical structure of cross-slope inflow in response to the dynamic influences of a submarine trough, the model has 70 levels with uniform intervals of 10 m in the upper 700-m depth. From 700- to 1700-m depth, 20 levels are discrete with 50-m intervals to capture the lower portion of the ASC over the slope. For the abyssal ocean,
10 levels are discrete with 130 m layer thickness from 1700- to 3000-m depth. Partial cells are employed to improve the accuracy of topography and ice draft, with a minimum wet-cell fraction of 0.3.

b. Open boundary forcing and initial conditions

This system is assumed to be primarily forced by the geostrophic zonal flow corresponding to the ASC observed in East Antarctica. Therefore, the large-scale ASF/ASC is prescribed on the zonal open boundaries by setting the potential temperature $\theta$ (Fig. 2a), the salinity $S$ (Fig. 2b), and the zonal velocity $u$ (Fig. 2c) on the basis of observational hydrographic datasets. The zonal boundary conditions for $\theta$ and $S$ are derived from the Baseline Research on Oceanography, Krill and the Environment West (Williams et al. 2010), the Japanese Whale Research Program, and the Marine Mammals Exploring the Oceans Pole to Pole (Treasure et al. 2017). The zonal boundary conditions for $u$ are derived from long-term mooring observations from Heywood et al. (1999) and Peña-Molino et al. (2016).

According to the typical hydrography of Fresh Shelf regions, $\theta$ over the shelf is set to $-1.9^\circ$C (close to the surface freezing point) for the entire water column, and $S$ increases from 34.4 psu at the sea surface to 34.55 psu on the sea floor. A sharp pycnocline over the slope separates the cold and fresh shelf region from the warm and salty deep zone. In the deep ocean, the warm core of mCDW is set as $0.5^\circ$C at 400-m depth, with the maximal salinity of 34.7 psu. A thermocline above the mCDW linearly warms up from $1.8^\circ$C at 50-m depth to $0.5^\circ$C at the mCDW core depth, capped by a 40-m-thick surface layer at $-1.9^\circ$C. The halocline above the mCDW shares the same depth with the thermocline, with $S$ linearly increasing from 34.55 psu at 50-m depth to 34.7 psu at 400-m depth. A fresher layer overlies the halocline, with $S$ ranging from 34.4 psu at the sea surface to 34.5 psu at 40-m depth. Beneath the mCDW core, $\theta$ gradually cools down to
ingly, the sloping potential density $D$ and $S$ from the deep ocean toward the continental slope and $0.5$ km. Since we set $\Delta \theta$ as $0$. The stress is assumed to be included by the open boundaries. By using such a simplified method similar to St-Laurent et al. (2013), we can conduct a clean analysis of the dynamic response of the ASC to a submarine trough over the Fresh Shelf regions.

c. Experimental design

Two numerical experiments are conducted to investigate the dynamic effects of a submarine trough on the ASC and the consequent mCDW intrusion. In the first experiment, hereafter referred to as the CONTROL, $\theta$, $S$, and $u$ are held constant at the open boundaries (see appendix A for model evaluation). The CONTROL is integrated for 3 years, and 6-h mean outputs in the last year are used to investigate the dynamic mechanisms governing the mCDW intrusion. In the second experiment, hereafter referred to as the SEASON, the boundary conditions $\theta$ and $S$ are still held constant. Yet, a sinusoidal function is superposed on $u$ to represent the seasonality of the ASC. For each grid at the zonal boundaries, the amplitude of the superposed sinusoidal function is half of the constant $u$, and the phase is characterized by a maximal value at 30 June. In a seasonal cycle, the ASC transport speeds up from $\sim 2$ Sv at the beginning of January to $\sim 6$ Sv at the end of June and slows down to $\sim 2$ Sv at the end of December (Fig. 2e). Such an idealized seasonal cycle of the ASC is on the basis of mooring observations and modeling results (Heywood et al. 1999; Mathiot et al. 2011; Peña-Molino et al. 2016).

The SEASON uses the last output of the CONTROL as the initial conditions and approaches an equilibrium state after integration of 2 years, with annually repeated open boundary forcing. Thereafter, the SEASON is integrated for

\[\Delta \mu \text{ from } -1.8^\circ \text{ to } 0.5^\circ \text{C} \text{ at 3000-m depth, and } S \text{ freshens to 34.65 psu at the seafloor.}\]

To represent the ASF, the thermocline (from $-1.8^\circ$ to 0.5°C) and halocline (34.5 – 34.7 psu) tilt southward and downward from the deep ocean toward the continental slope and intersect the seafloor below 500-m depth (Figs. 2a, b). Accordingly, the sloping potential density $\sigma$ surfaces (27.8 – 27.84 kg m$^{-3}$) over the slope serve as a PV barrier that prevents mCDW from intruding the shallow shelf (Fig. 2d). As the seafloor deepens from 500 to 700 m in the submarine trough (black lines in Figs. 2a–d), $\sigma$ classes hosting the pycnocline can reach the shelf and offer an intrusion route for mCDW. The prescribed ASC lies between 65.87° and 65.57°S, characterized by the maximal velocity of $-0.3$ m s$^{-1}$ at 65.75°S. Since the ASC is in geostrophic balance (Fig. 2e), its vertical shear is consistent with the thermal-wind balance with $\sigma$ surfaces deepening to the south over the slope (Fig. 2d). At zonal open boundaries, an integral of $u$ over the ASC area estimates the westward transport at $\sim 4$ Sv (1 Sv = $10^8$ m$^3$ s$^{-1}$) (the red line in Fig. 2e). The $\theta$ and $S$ at the northern open boundary are the same as the north of the zonal boundary conditions, and the meridional velocity $v$ at all open boundaries is set as $0$.

The model is initialized with zonally uniform hydrography that is equal to the values at zonal open boundaries. Since we intend to examine the dynamic influences of topography, the atmospheric forcing and the sea ice model are not included to exclude surface influences. The effect of large-scale surface stress is assumed to be included by $\theta$, $S$, and $u$ forcing over open boundaries. By using such a simplified method similar to
one more year and saved with 6-h mean output. To exclude the influences of the model adjustment induced by the initial sudden change from the CONTROL to the SEASON, we only use the last (third) year model output of the SEASON to investigate the dynamic response of mCDW intrusion to the seasonality of the ASC.

3. Results

a. The features of mCDW intrusion

In the CONTROL, the annual mean 27.82 kg m$^{-3}$ isopycnal is the upper limit of density classes hosting warm mCDW (Fig. 3). To the east of the trough mouth, the isohalines of 34.5–34.55 psu reach the shelf break, yet the −1.8°C isotherm intersects on the upper slope (Figs. 3a,b). Consequently, the 27.82 kg m$^{-3}$ isopycnal is confined to the south of 65.9°S due to the sharp front of $\theta$. Such steep isopycnals tiling acts as a PV barrier that effectively prevents mCDW from ventilating the cold and fresh shelf region (Fig. 3g). The isopycnal of 27.82 kg m$^{-3}$ at the eastern side of the trough mouth is roughly as deep as upstream (Figs. 3c,d). However, the seafloor is deeper than 500 m, and mCDW accesses the shelf break beneath the 27.82 kg m$^{-3}$ isopycnal. At the deepest cross section of the trough (85°E), the 27.82 kg m$^{-3}$ isopycnal directly extends onto the shelf, and the PV barrier vanishes below 500-m depth (Fig. 3h). In the absence of the PV
FIG. 3. (a),(b) The annual mean $\theta$ ($^\circ$C) and $\bar{S}$ (psu) cross section at upstream (86.25$^\circ$E); white lines are $\sigma$ (kg m$^{-3}$) in contour intervals of 0.02 kg m$^{-3}$. (c),(d) As in (a) and (b), but for the cross section at the upstream edge (85.37$^\circ$E). (e),(f) As in (a) and (b), but for the cross section at midtrough (85$^\circ$E). (g) The potential vorticity ($10^{-10}$ m$^{-1}$ s$^{-1}$) cross section at upstream (86.25$^\circ$E). (h) As in (g), but for cross section at midtrough (85$^\circ$E). The locations of the cross sections are marked in Fig. 1b.
barrier, warm mCDW spreads southward and floods the trough between the 27.82 kg m\(^{-3}\) isopycnal and the seafloor (Figs. 3e,f).

Warm mCDW is confined within the trough and extends to the ice shelf front (Figs. 4a–c). The warmest mCDW are identified at the trough mouth occupying the lower portion of the water column from the seafloor to \(<450\)-m depth (Fig. 4a). Under the ice shelf front, the mCDW volume dramatically decreases, with \(\theta\) cooling down to \(<-1.5^\circ\text{C}\) (Fig. 4c), implying the blocking effects of the ice shelf front on the inflow (Wåhlin et al. 2020). The onshore flow closely concentrates along the eastern side of the trough, and the outflow hugs the western flank (Figs. 4d–f). Steered by the eastern slope of the trough, the onshore inflow is characterized by bottom intensification and gradually enhanced southward (Figs. 4d–f). Such vertical shear of the onshore flow is coupled with the local isopycnals tilting eastward and upward (Figs. 4a–c). The southward enhancement of inflow is expected since the outflow induced by the ice shelf basal melting must be compensated by equivalent inflow (St-Laurent et al. 2013).

The onshore volume and heat transport across the shelf break of the trough (only the southward transport is included) shows the temporal evolution of mCDW intrusion (Fig. 5). The change of onshore heat transport is apparently determined by the onshore volume transport (i.e., \(v\)) rather than the changes of \(\theta\). Significant high-frequency oscillations are presented in the time series, characterized by periods shorter than 1 week. The high-frequency pulsing is linked to the TRWs discussed in section 3d. In contrast to the CONTROL, the time series from the SEASON exhibit a notable seasonal cycle, with a minimum in February and a maximum at the end of March. In the SEASON, the ASC transport prescribed at open boundaries peaks at the end of June. Therefore, it is interesting to find that the seasonality of mCDW intrusion is out of phase with the seasonal cycle of the ASC. The mechanism responsible for the mCDW intrusion peaking in March is discussed in section 3e.

b. First mechanism: Bottom pressure torque

To clarify the dominant mechanism driving the time-mean onshore transport, the full vorticity budget is assessed. Without atmospheric forcing, the full vorticity budget is written as
where $\zeta_t$ is the vorticity tendency, $f$ is the planetary vorticity, $\beta = \partial f/\partial y$, $w$ is the vertical velocity, $u$ is the horizontal velocity vector, and $V_{isco}$ is the parameterized viscous term. On the RHS of Eq. (1), the terms are referred to as term I, the meridional transport term; term II, the stretching term; term III, the advection term; term IV, the viscosity term. The long-term mean vorticity tendency is negligible as the model approaches a steady state, and then Eq. (1) is rearranged as

$$\beta \nu = f w_z - \nabla \times (u \cdot \nabla u) + \nabla \times (V_{isco}),$$

(2)

Since the ASC is relatively weak, the along-slope flow tends to turn onshore at the upstream corner of the trough and shift slightly downstream due to inertia. Accordingly, the critical region is located at the trough mouth where the westward flowing ASC initially diverts onshore. The annual mean vorticity budget at the trough mouth reveals the dominant mechanism driving the time-mean onshore flow (Fig. 6). Compared to the terms on the RHS of Eq. (2), the meridional transport term has minimal values (Fig. 6a). The large terms on the RHS of Eq. (2) are the stretching term and the advection term (Figs. 6b,c). The positive advection term is largely counterbalanced by the negative stretching term, indicating that deepening of topography dominates the dynamics. The negative stretching term is also partially counteracted by the noisy and positive viscosity term in the bottom layer. Yet, the viscosity term only plays a vital role near the seafloor in response to the bottom friction (Fig. 6d).

By taking an integral of Eq. (2) over depth and scaling with the seawater density $\rho$, the full vorticity equation gives the barotropic vorticity equation written as

$$\beta \int^H_0 \rho w dz = -\rho w_b - \int^H_0 \rho \nabla \times (u \cdot \nabla u) dz + \int^H_0 \rho \nabla \times (V_{isco})dz,$$

(3)

where $\eta$ is the sea surface elevation, $H$ is the undisturbed water depth, and $w_b$ is the vertical velocity at the seafloor. The terms in Eq. (3) are referred to as term I, the advection of geostrophic vorticity; term II, the bottom pressure torque; term III, the advection of relative vorticity; term IV, the curl of viscosity. Term II in Eq. (3) can also be written as $\nabla p_b \times \nabla H$ (Hughes and de Cuevas 2001; Spence et al. 2012), where $p_b$ is
the bottom pressure. Therefore, term II reflects the dynamic influences of the topography on the flow when the bottom pressure contours are not parallel with the isobaths. The bottom pressure torque is also extensively used to explain the joint effect of baroclinicity and relief (Mertz and Wright 1992) on the western boundary currents, such as the Gulf Stream and the Kuroshio (Spence et al. 2012; Liu et al. 2020).

In the eastern flank of the trough mouth, the bottom pressure torque plays a leading role in the barotropic vorticity balance (Fig. 6f), and it is mainly counterbalanced by the...
advection of relative vorticity (Fig. 6g). The viscosity term plays a secondary role in counteracting the bottom pressure torque (Fig. 6h). Finally, the residual of the RHS of Eq. (3) is balanced by the onshore advection of geostrophic vorticity on the LHS (Fig. 6e). Therefore, it is clear that the bottom pressure torque drives the time-mean flow. The ASC flows westward along the isobaths over the slope from the eastern open boundary until it encounters the steep trough. As the isobaths (the depth ranging from 500 to 700 m) suddenly bend southward at the eastern trough mouth, the ASC tends to jump across the isobaths to a deeper depth due to its inertial overshot and breaks the parallelism between the bottom pressure contours and the isobaths ($\nabla p_b \times \nabla H \neq 0$). As a result, the bottom pressure torque leads the southern fraction of the ASC onshore into the trough.

c. Second mechanism: Topography beta spiral

It is reasonable to expect a bottom-intensified inflow due to the bottom pressure forcing; however, the mechanism governing the vertical structure needs to be further clarified. Two model profiles (M1 and M2 in Fig. 6e) at the eastern trough mouth show the vertical structure of the velocity and the corresponding $\theta$ (Fig. 7). A striking spatial pattern is characterized by the anticlockwise rotation of the horizontal velocity vector with increasing depth, yet the velocity magnitude decreases slightly with depth (Fig. 7a). Therefore, the bottom-intensified inflow manifests the anticlockwise rotation of the horizontal velocity with increasing depth. In addition, the rotation of the flow is most significant at the thermocline depth, implying a strong linkage between the rotation and the vertical stratification. In the bottom frictional layer (~40 m above the seafloor), the clockwise rotation of the velocity vector may be correlated with the bottom Ekman spiral that is to the right of the overlying flow in the Southern Hemisphere. This study focuses on the water column above the bottom frictional layer.

This vertical structure of the current rotation is referred to as the topographic beta spiral (TBS; Pedlosky 1987; Huang 2010; Yang et al. 2018). With the geostrophic and hydrostatic approximations, the vertical shear of the horizontal velocity angle is determined as

$$\frac{\partial \varphi}{\partial z} = \gamma \left( \nu \frac{\partial \rho}{\partial z} \right)$$

(4)

where $\varphi = \tan^{-1}(\nu/\omega)$ is the horizontal velocity angle, $\gamma = g/f(\rho_0)$, and $\rho_0 = 1028$ kg m$^{-3}$ is the reference density (appendix B). Based on Eq. (4), the horizontal flow rotation with depth is regulated by the stratification $\rho_0$ and the vertical velocity $\omega$.

The meridional cross section of $\varphi$ shows that the northernmost discernible TBS is identified at the northern tip of the trough mouth at 65.8°S (Fig. 8a). As the ASC is oriented toward the shelf break, the current in the lower layer rotates faster than in the upper layer until it reaches ~65.9°S. Thereafter, the anticlockwise rotation of the flow in the lower layer slows down, and $\varphi$ in the upper layer increases rapidly, consistent with the TBS at M2 shown in Fig. 7b. Parameter $\varphi_z$ calculated from Eq. (4) agrees well with the velocity from model output (Fig. 8b), featuring an anticlockwise rotation with increasing depth and peaking at ~400-m depth where the pycnocline sits. Thus, the TBS is expected to be the dominant mechanism governing the vertical structure of the onshore flow. Significant differences of $\varphi_z$ between the simulated velocity and Eq. (4) are found in the bottom layer (550-m depth below), implying that the bottom Ekman dynamics overtakes the TBS near the seafloor.

**FIG. 7.** (a) The annual mean horizontal velocity vector profile at M1 station (85.375°E, 65.9°S), with corresponding $\theta$ (°C) in color. (b) As in (a), but for the M2 station (85.375°E, 66°S). M1 and M2 are located on the isobath of 620-m depth.
Based on the vertical shear of the horizontal velocity angle $\varphi_z$ derived from Eq. (4), we calculate the estimated horizontal velocity angle $\varphi_b$ by integrating Eq. (4) upward from 550 m depth, with the simulated $\varphi = \tan^{-1}(v/u)$ at 550-m depth as the lower reference level. As expected, $\varphi_b$ shows quite good congruence with $\varphi$ from the simulated velocity (Fig. 8c), and the differences between $\varphi$ and $\varphi_b$ are attributed to the overestimated $\varphi_z$ of Eq. (4) at the pycnocline layer depth. Since $\varphi_z$ in Eq. (4) is a production of $\gamma$, the inverse of $|u|^2$, $w$, and $\rho_c$, it is unlikely to find the dominant factor directly. By assessing these factors separately (Figs. 8d,e), the vertical structure of $\rho_c$ is most similar to the pattern of $\varphi_z$, implying the notable influences of the stratification.

For the M2 station, the profiles $\varphi_z$ and $\varphi$ also confirm that the TBS determines the vertical shear of the flow angle from the sea surface to the thermocline (Figs. 9a,b). The horizontal momentum equations are

$$\frac{\partial u}{\partial t} = -u \frac{\partial u}{\partial x} - v \frac{\partial u}{\partial y} - w \frac{\partial u}{\partial z} + \frac{f v}{\rho_c} \text{ Coriolis}$$

$$+ \left[ -g \frac{\partial \eta}{\partial x} - \int_0^z g \frac{\partial p}{\partial x} \, dz \right] + \left[ \frac{\partial}{\partial z} \left( \frac{1}{\rho_c} \frac{\partial u}{\partial z} \right) + D \right],$$

where $\rho_c$ is the density of the pycnocline.
The page contains mathematical equations and diagrams related to the dynamics of the ocean. The text is discussing the momentum balance and the role of various forces and terms in the equations. The diagrams illustrate the depth profiles and budget terms for momentum. The text mentions the TBS mechanism and the influence of topography on the flow. The figures show the distribution of certain parameters over time and depth.
ASC crosses the contours of $f/H$ and descends to the deeper trough mouth by inertial overshoot, the water column tends to stretch and generate a cyclonic vortex in response to the PV conservation. The vortex propagation is constrained by the isobaths that act as waveguides. While the cyclonic vortices are advected with the mean flow and continuously pass through the zonal cross section shown in Fig. 4a, the high-frequency oscillations associated with TRWs are detected on the time series of onshore volume and heat transport (Fig. 5).

An analytical solution of the periodicity of TRWs is given by Fennel and Schmidt (1991) as

$$ T = \frac{4\pi H_{\text{shelf}}}{H_{\text{trough}} f} \left[1 - \frac{2R_1}{d} \ln \left(2 - 2 \cos \left(\frac{\pi H_{\text{trough}}}{H_{\text{shelf}}} \right) \right) \right]^{-1}, $$

where $T$ is the TRWs period, and $R_1$ is the first baroclinic Rossby radius of deformation estimated using the model output. Therefore, the TRWs period is determined by the topography ($H_{\text{shelf}}$ and $H_{\text{trough}}$) and $R_1$ correlated with the stratification. Following Eq. (7) employed in St-Laurent et al. (2013), it gives a periodicity of ~4.0 days by substituting the parameters used in the CONTROL and $R_1 = \sim 8.5$ km estimated over the eastern trough mouth. Note that the periodicity of TRWs from the CONTROL may not be exactly equal to this analytical solution, as the analytical periodicity is derived from a highly simplified quasigeostrophic ocean model.

Since the cyclonic vortices are coupled with relatively lower sea surface height $\eta$, the $\eta$ evolution is straightforward to reveal the TRWs characteristics. The Hovmöller diagram of $\eta$ at midtrough shows a periodicity of ~4 days, with onshore phase propagation (Fig. 10). The power spectrum of $\eta$ within the trough shows peaks near the period band of the analytical TRWs (Fig. 11a), confirming that the high-frequency oscillations shown in Fig. 5 are induced by the TRWs generated at

![Fig. 11.](image1.png)

![Fig. 12.](image2.png)
the trough mouth and propagating onshore along the isobaths. In addition, the integral of the $\eta$ power spectrum over the simulated TRWs frequency band shows that the energetic region within the trough is located in the northeastern flank (Fig. 11b).

e. Dynamic responses of mCDW intrusion to the ASC seasonality

Compared to the steady monthly-mean time series in the CONTROL, the onshore volume and heat transport in the SEASON show a remarkable seasonal cycle (Fig. 5). The strength of the ASC can directly determine the mCDW onshore transport by the bottom pressure torque discussed in section 3b. However, mCDW intrusion peaks at the end of March, leading the maximum of the ASC transport by three months (Fig. 2e).

At the M1 station, the Hovmöller diagram of monthly-mean velocity fields from the SEASON shows a remarkable seasonal cycle (Fig. 12). The magnitude of the velocity and the prescribed ASC covary closely, with speed increasing from January to June and decreasing from July to December (Fig. 12a). In contrast to the steadily increasing $u$ from February to March (Fig. 12b), the $v$ profile above 400-m depth remarkably accelerates (Fig. 12c). Since there is no significant strengthening in the velocity magnitude above 400-m depth in March (Fig. 12a), such suddenly enhanced $v$ in the upper layer must be induced by the anticlockwise rotation of the current (Fig. 12d), i.e., the dynamic effects of TBS are expected to be responsible for the maximal onshore transport in March.

The change of the TBS from February to March is mainly regulated by the stratification. In the SEASON, the minimal peaks of $\varphi_z$ from both velocity fields and Eq. (4) rise from ~400-m depth in February up to ~350-m depth in March (Fig. 13a). More importantly, the minimum of $\varphi_z$ is also larger in March than in February, suggesting that the anticlockwise rotation of the upper flow is much stronger in March (Fig. 13b). Therefore, the enhanced onshore flow from February to March is primarily induced by the changes in the TBS structure. The changes of the pycnocline depth from February to March are similar to that of $\varphi_z$ (Fig. 13c), implying the critical role of the stratification. To qualitatively assess the contribution of each term ($w$, $|\bar{u}|^2$, and $\rho_z$) to the changes of $\varphi_z$, we calculate the intermediary $\varphi_z$ by switching the value of one term from February to March and holding the rest of the three terms to the value in February. (Fig. 13d). As expected, when $\rho_z$ is switched from the value in February to that in March, the profile of such an intermediary $\varphi_z$ is most close to that in March, confirming the key role of the stratification.

In the SEASON, the changes of the stratification from February to March are unique throughout the year (Fig. 14).
FIG. 14. (a) $\sigma$ (kg m$^{-3}$) cross section at 85.37°S in February from the SEASON. (b) As in (a), but for the difference between February and March. (c)-(n) The successive changes of monthly mean $\eta$ (m) from the SEASON. The yellow box in (d) highlights the lower $\eta$ center.
As the isopycnals hosting mCDW extends into the trough mouth (Fig. 14a), these isopycnals rising up to the south of 65.84°S and deepening down to the north of 65.84°S from February to March (Fig. 14b). With the geostrophic and hydrostatic approximations, the pycnocline depth is tightly related to the $h$ field. The successive changes of $h$ show an evident lower center at the trough mouth from February and March (Fig. 14d), and such a lower $h$ center must be associated with a shallower pycnocline (Fig. 14b). The monthly-mean evolution of $h$ may be correlated with the changes of the TRWs characteristics in response to the seasonal cycle of the ASC. In contrast to the unique pattern shown in Fig. 14d, there is no such a lower center over the trough mouth in Figs. 14c,e–n.

4. Concluding remarks

This study aims to clarify the mechanisms responsible for mCDW intrusion within a trough on a Fresh Shelf case in
East Antarctica. Three dynamic mechanisms are identified as follows: 1) we clarify the dominant role of the bottom pressure torque in driving mCDW intrusion over Fresh Shelf regions for the first time; 2) another novelty is the role of the TBS in regulating the vertical structure of the onshore flow; 3) the high-frequency oscillations of onshore transport are induced by the TRWs. The ASC flows westward along the zonal slope until it encounters the steep trough (Fig. 15). As the isobaths suddenly bend southward at the eastern trough mouth, the ASC overshoots to a deeper depth due to its inertia. For a barotropic case, the flow should follow the isobaths and turn onshore, yet the unique stratification over Fresh Shelf regions plays a key role in regulating the vertical structure of the onshore flow. To conserve the PV, the bottom pressure torque leads the southern fraction of the ASC into the eastern side of the trough, featured by a bottom-intensified flow. Meanwhile, the water columns stretch and generate cyclonic vortices (i.e., the TRWs) propagating along the isobaths within the trough. The bottom-intensified inflow manifests the anticlockwise rotation of the horizontal velocity with increasing depth, suggesting that the TBS predominantly governs the vertical structure of the horizontal velocity. Overall, the bottom pressure torque plays a dominant role in driving the time-mean inflow at the eastern side of the trough mouth, while the TRWs contribute to the high-frequency oscillations in the onshore volume and heat transport. These mechanisms occur simultaneously and govern the mCDW intrusion together. Since the strength of the ASC can regulate the bottom pressure, mCDW intrusion within the trough is expected to covary with the ASC. However, the maximal onshore volume and heat transport leads the strongest ASC by three months due to the changes of stratification. The simulated seasonality of mCDW intrusion coincides with the observations over the shelf near the Totten Glacier and the Amery Ice Shelf. Based on mooring observations in Prydz Bay, Herráez-Borreguero et al. (2015) documented the remarkable increase in mCDW intrusion from the beginning of February to the end of March. Observations from CTD-instrumented elephant seals revealed

**Fig. A1.** The snapshot of $\theta$ (°C) at 500-m depth in the first 180 days in the CONTROL. Gray curves are the ice shelf front and the isobaths (at 50-m intervals from 500 to 700 m, and 500-m intervals thereafter).
that mCDW is the dominant water mass in the eastern region of Prydz Bay from April through May (Williams et al. 2016). Based on observations from APEX profiling floats, Silvano et al. (2019) found that mCDW intrusion peaks in the early austral autumn in the Sabrina Coast.

There are some limitations in the model design and our analysis. The meridional position of the ASC/ASF is fixed, and thus the influences of the meridional migration of the ASC are not assessed. In the absence of surface atmospheric conditions and the sea ice component, the model cannot simulate the deep vertical convection that can cool down the water column and bring a salinification from brine rejection. Furthermore, the influences of high-frequency pulsing of the ASC, e.g., the tidal and synoptic oscillations, are still unclear.

The geometry and open boundary conditions are idealized and highly simplified in the simulations to match the typical hydrography. However, the mechanisms identified in the study should be qualitatively robust to represent the dynamic mechanisms responsible for mCDW intrusion within submarine troughs over Fresh Shelf regions, East Antarctica.

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Data availability statement. No datasets were generated or analyzed during the current study.

APPENDIX A

Model Evaluation

The CONTROL is integrated for 2 years to approach a steady state, and the monthly mean output is saved. During the first 90 days, mCDW flows southward along the eastern side of the trough and reaches the ice shelf front (Figs. A1a–c). Over the following 90 days, mCDW is blocked from traveling south by the ice shelf front and is directed northward along the western side of the trough (Figs. A1d–f). Mesoscale structures emerge between the inflow and the outflow, and such cyclonic eddies coincide with the vortexes propagation described by St-Laurent et al. (2013).

The model domain integrated heat content, kinetic energy and the bulk measures for the sub-ice-shelf cavity show that both the CONTROL and the SEASON are in the equilibrium state after integration of ~2 years (Figs. A2a,b). We further calculate the annual mean $\theta$, $S$, and barotropic streamfunction $\psi$ for the CONTROL in the second year. The $\theta$ and $\bar{S}$ at 500-m depth (the shelf bottom) show that mCDW intrusion is constrained within the trough, recognized as warmer and saltier water (Figs. A2c,d). The $\psi$ exhibits a cyclonic gyre centered over the trough, featuring a steady inflow along the eastern flank (Fig. A2e) and coinciding with the warmer and saltier tongue (Figs. A2c,d). The spatial patterns of $\theta$, $\bar{S}$, and $\psi$ are in qualitative agreement with the observed hydrography properties over Fresh Shelf regions.

Fig. A2. (a) Time series of the total kinetic energy ($m^2 s^{-2}$) and heat content (J) over the entire domain (yellow line and red line) and within the sub-ice-shelf cavity (blue line and green line) from the CONTROL. (b) As in (a), but for the SEASON. (c) The annual mean $\theta$ (°C) at 500-m depth from the CONTROL. (d),(e) As in (c), but for $\bar{S}$ (psu) and $\psi$ (Sv).
APPENDIX B

Derivation of the Topography Beta Spiral Equation

With the geostrophic and hydrostatic approximations, if the mixing and thermal forcing are negligible, the density balance in a steady state is

\[ \frac{\partial \rho}{\partial t} + u \frac{\partial \rho}{\partial x} + v \frac{\partial \rho}{\partial y} + w \frac{\partial \rho}{\partial z} = 0. \]  

(B1)

The continuity equation is

\[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0. \]  

(B2)

The thermal wind relation is

\[ f \frac{\partial u}{\partial z} = \frac{g}{\rho_0} \frac{\partial \rho}{\partial y}, \]  

(B3a)

\[ f \frac{\partial v}{\partial z} = -\frac{g}{\rho_0} \frac{\partial \rho}{\partial x}. \]  

(B3b)

Defining the angle of horizontal flow as \( \varphi = \tan^{-1}(u/v) \) and substituting Eqs. (B2), (B3a), and (B3b) to Eq. (B1), it gives the rotation rate of horizontal flow with depth:

\[ \frac{\partial \varphi}{\partial z} = -\frac{\gamma}{|u|} \left( w \frac{\partial \rho}{\partial x} \right), \]  

(B4)

REFERENCES


