

Spatial Characteristics and Dynamic Mechanisms of the Antarctic Slope Current in the Ross Sea

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Key Points:

- Using a numerical model, we identified both surface-intensified and bottom-intensified slope currents in the Ross Sea
- Both barotropic and baroclinic pressure terms contribute to the spatial heterogeneity of the Antarctic Slope Current
- The conversion between Mean Kinetic Energy and Mean Available Potential Energy reverses in different slope regimes

Supporting Information:

Supporting Information may be found in the online version of this article.

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Abstract Coupled with the Antarctic Slope Front (ASF), the Antarctic Slope Current (ASC) mostly encircles Antarctica and has variable structures. Two regimes of the ASC/ASF have been identified in the Ross Sea, but the spatial characteristics of the ASC/ASF have not been depicted in detail, and the dynamical mechanisms of the two ASC/ASF regimes are still unclear. Using an eddy-permitting coupled regional ocean-sea ice-ice shelf model, we investigate the spatial characteristics and energetics of the ASC/ASF in the Ross Sea. Based on the simulated results, two distinct structures of the ASC/ASF have been identified in two regimes: (a) in Regime I, the westward ASC is characterized by a surface-intensified flow, predominantly driven by the barotropic pressure gradient; (b) in Regime II, the ASC is characterized by a bottom-intensified westward flow, dominated by the baroclinic pressure gradient, especially prominent where Dense Shelf Water (DSW) overflows. Furthermore, by conducting an energy budget analysis in the framework of the Lorenz Energy Cycle, we found that in Regime I the ASC is maintained by Ekman transport that piles up water against the slope and drives a conversion from Mean Kinetic Energy (MKE) to Mean Available Potential Energy (MAPE), whereas in Regime II the ASC is strengthened as a bottom-intensified current through DSW overflows, which release MAPE into MKE.

Plain Language Summary The Antarctic Slope Current (ASC), coupled with the Antarctic Slope Front (ASF), flows along the Antarctic continental shelf edge. Since cross-slope flow has to cross the ASF, the ASC can regulate the cross-slope exchanges of water masses, heat, salinity, and nutrients. Therefore, it is necessary to investigate the characteristics of the ASC and its responsible mechanisms. Numerical modeling of the Ross Sea reveals two distinct spatial patterns: a surface-intensified type with maximum flow near the surface and a bottom-intensified type with strongest flow residing on the seafloor. Momentum budget analysis shows the surface-intensified ASC is dominated by sea surface height gradients, while the bottom-intensified ASC is primarily driven by cross-slope density gradients. Energy analysis demonstrates that in the surface-intensified regime, Mean Kinetic Energy (MKE) is converted into Mean Available Potential Energy (MAPE), sustaining the shelf-break density front and inhibiting on-shelf warm water intrusion. Conversely, in the bottom-intensified regime, Dense Shelf Water outflow converts MAPE to MKE, accelerating the ASC, and then the ASC transports the kinetic energy downstream. Although the spatial resolution of this numerical model is not high enough to fully resolve mesoscale processes, this study still provides a benchmark result for future high-resolution simulations and in situ observations.

1. Introduction

Coupled with the Antarctic Slope Front (ASF; Jacobs, 1991), the Antarctic Slope Current (ASC) flows along the continental slope and almost encircles Antarctica (Thompson et al., 2018; Whitworth et al., 1998). Three mechanisms have been proposed to drive the ASC/ASF: (a) the mechanical forcing of polar easterlies (Gill, 1973; Jacobs, 1991; Mathiot et al., 2011); (b) the thermal wind shear driven by cross-slope density gradients (Gordon et al., 2004; Huneke et al., 2022, 2023; Thompson & Heywood, 2008, 2020); (c) the tidal forcing (Flexas et al., 2015; Si et al., 2022; Stewart et al., 2019). Acting as a potential vorticity barrier (Pedlosky et al., 1987) separating waters over the shallow continental shelf from those in the deep ocean (Thompson et al., 2018), the ASC/ASF has a great impact on cross-slope exchanges of heat, salt, and nutrients (Baines & Condie, 1998;

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Dinniman et al., 2011; Heywood et al., 2014; Liu et al., 2022; Prézelin et al., 2004; Schmidtke et al., 2014). Warm Circumpolar Deep Water (CDW) on the northern flank of the ASC/ASF can intrude onto the continental shelf, leading to ice shelf thinning in West Antarctica (Pritchard et al., 2012; Shepherd et al., 2018). Dense Shelf Water (DSW) on the southern flank of the ASC/ASF spills into the deep ocean and forms Antarctic Bottom Water (AABW), which is a mixture of DSW and ambient waters (Gordon et al., 2015; Orsi et al., 1999). AABW constitutes a critical lower limb of the global overturning circulation (Cessi, 2019; Evans et al., 2017; Marshall & Speer, 2012), while CDW intrusions threaten the stability of ice shelves, potentially contributing to global sea level rise (Rignot et al., 2013; Tinto et al., 2019; Wang et al., 2023). Therefore, it is of great interest to investigate the spatiotemporal evolution of the ASC/ASF that regulates cross-slope exchanges.

Around Antarctica, the Ross Sea is the second-largest marginal sea after the Weddell Sea, with the largest ice shelf, the Ross Ice Shelf, located to its south. Intensive cross-slope exchanges have been identified in the Ross Sea. Modified CDW (MCDW) intrudes onto the continental shelf within submarine troughs in the Ross Sea (Budillon et al., 2003; Dinniman et al., 2011; Morrison et al., 2020; Orsi & Wiederwohl, 2009; Wang et al., 2024), and the DSW overflow on the Ross Sea continental slope accounts for ~20% of the AABW production (Meredith, 2013; Padman et al., 2009). Cross-slope exchanges must cross the ASC/ASF, and thereby it is important to understand the mechanisms driving the ASC/ASF and associated characteristics in the Ross Sea.

Previous studies have documented some characteristics of the ASC/ASF in the Ross Sea. According to the classification based on the thermodynamic characteristics of the ASC/ASF (Thompson et al., 2018), the ASC/ASF exhibits the Fresh Shelf regime in the eastern Ross Sea and the Dense Shelf regime in the western Ross Sea. The Fresh Shelf regime features relatively fresh and cold shelf water, while the Dense Shelf regime features relatively saline and cold shelf water. According to the dynamic classification of the ASC (Huneke et al., 2022), the ASC/ASF consists of a surface-intensified regime in the eastern Ross Sea and a bottom-intensified regime in the western Ross Sea. The surface-intensified regime features large velocities in the upper layer, whereas the bottom-intensified regime features a bottom-intensified flow with reduced velocity toward the upper layer. Generally, the thermodynamic and dynamic classifications align well (Huneke et al., 2022; Thompson et al., 2018); that is, the surface-intensified ASC occurs over the eastern Fresh Shelf regime, and the bottom-intensified ASC occurs over the western Dense Shelf regime. However, there are also misfits between the thermodynamic and dynamic classifications. For example, the northern tip of the Iselin Bank (IB) is recognized as a transition zone between the Fresh Shelf and Dense Shelf regimes by Thompson et al. (2018), yet the slope around the IB is categorized as a bottom-intensified regime by Huneke et al. (2022). Such discrepancies may be relevant to the complex physical characteristics of the ASC/ASF. To improve our understanding of the mechanisms governing the ASC in the Ross Sea, this study aims to clarify the spatial structure and energetics of the ASC/ASF. Given the limited observations over the Ross Sea continental slope, numerical models are useful for exploring the characteristics of the ASC/ASF and the responsible mechanisms.

We utilize an eddy-permitting numerical model to examine the spatial patterns and energetics of the ASC in the Ross Sea. This study is organized as follows. Section 2 describes the model configuration and methodologies. Section 3 presents the spatial characteristics and a dynamical analysis of the ASC. Section 4 summarizes the results with a discussion.

2. Model Description and Methods

2.1. Model Description

Based on the Massachusetts Institute of Technology General Circulation Model (Marshall et al., 1997), an eddy-permitting coupled regional ocean-sea ice-ice shelf model for the Ross Sea is configured by Yan et al. (2023), which includes a dynamic and thermodynamic sea ice model (Zhang & Hibler, 1997; Zhang & Rothrock, 2000) and a thermodynamic ice shelf model (Losch et al., 2010), with tidal forcing included.

This model is discretized on an orthogonal curvilinear grid mesh, with a horizontal resolution of ~5 km × 5 km. 70 vertical layers range from 10 m at the top layer to 600 m at the bottom layer. To better depict the seafloor depth and the geometry of sub-ice-shelf cavities, the partial cell scheme is employed to improve the representation of topography that is derived from the Refined Topography Data Set 2 (Schaffer et al., 2016; Figure 1a). Open boundary conditions, including the velocity, temperature, and salinity, are set up based on the Estimating the Circulation and Climate of the Ocean, Phase II (ECCO2; Menemenlis et al., 2008), the Marine Mammals

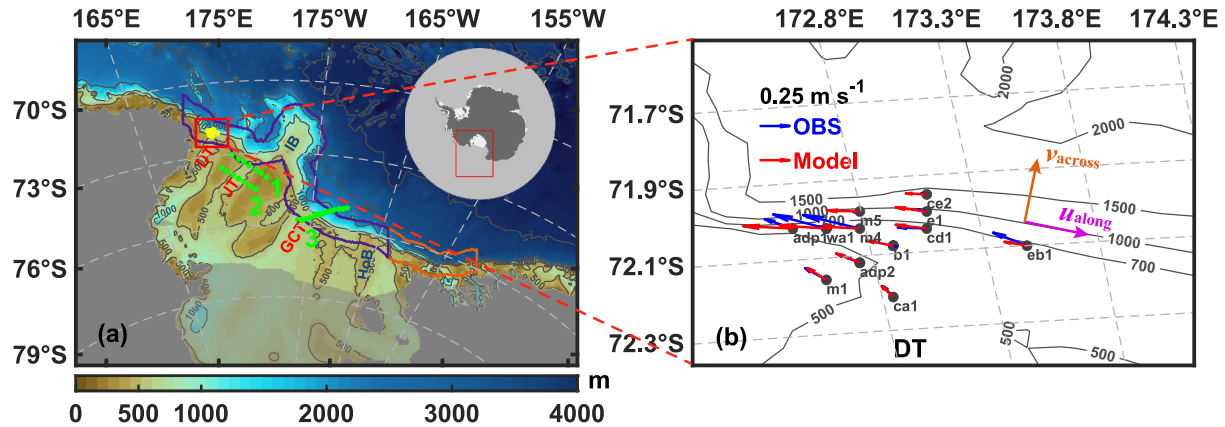


Figure 1. (a) The model bathymetry (color shading) with ice shelves (semitransparent regions) in the focused region. The red box in the inset shows the entire model domain. The purple and orange enclosed areas indicate the regimes shown in Figures 5, 8, 10–12. These encompass a 50-km-wide band surrounding the 1,000-m isobath. The yellow dots mark the mooring stations from the Antarctic Slope (AnSlope) program. The green dots on transects 1 and 2 show the conductivity-temperature-depth (CTD) stations from the Processes Regulating Iron Supply at the Mesoscale (PRISM) project, while green dots on transect 3 indicate the stations of Expendable Bathythermograph (Expendable Bathythermograph) from the PRISM project. DT: Drygalski Trough; JT: Joides Trough; GCT: Glomar Challenger Trough; IB: Iselin Bank; HoB: Houtz Bank. (b) Comparison of the velocity between the modeling results and the AnSlope mooring observations (OBS denotes the observations). The arrows represent the time-mean (OBS: 2003–2005; Model: 2014–2023) and depth-averaged (from the shallowest current meter ~200 m to the sea floor) velocity. Blue arrows indicate the observed velocity from the AnSlope mooring data, and red arrows indicate the simulated velocity from the model. The purple and orange arrows indicate the representative unit vectors (with a reference magnitude of 0.5 m s⁻¹) for the along-isobath velocity and the cross-isobath velocity (u_{along} and v_{across} defined in Equations 1a, 1b) at an arbitrarily selected point along the 1,000 m isobath.

Exploring the Oceans Pole to Pole observations (MEOP, Roquet et al., 2013, 2014; Treasure et al., 2017), and the World Ocean Atlas 2018 (WOA18, Locarnini et al., 2018; Zweng et al., 2019). The open boundary conditions differ from those used in Yan et al. (2023), which were constructed based on the Southern Ocean State Estimate (SOSE, Mazloff et al., 2010; Verdy & Mazloff, 2017), WOA18, and MEOP. Tidal forcing at the open boundaries, incorporating 10 major tidal constituents, is obtained from the Circum-Antarctic Tidal Simulation (Padman et al., 2002, 2003). Sea ice boundary conditions, including snow thickness, sea ice thickness, concentration, and velocity, are derived from the SOSE and the National Snow and Ice Data Center (Fetterer et al., 2017). The atmospheric forcing is derived from the European Center for Medium-Range Weather Forecasts Reanalysis v5 data set (ERA5, Hersbach et al., 2023), which has a spatial resolution of 0.25° × 0.25° and a temporal resolution of 1 hr. More parameters and configurations are referred to by Yan et al. (2023) in detail. The model integration spans from 2000 to 2023. To exclude the spin-up phase (Figures S1 and S2 in Supporting Information S1), daily outputs from the last 10 years (2014–2023) are used for analyzing the long-term mean characteristics and dynamics of the ASC.

2.2. Momentum Budget

Generally, the ASC tends to flow zonally, but the convoluted Antarctic continental slope does not always align with the zonal direction. To properly capture the characteristics of the ASC/ASF, we project the velocity from the geographic coordinate onto the directions along and across a fixed isobath. The along-isobath flow u_{along} and the cross-isobath flow v_{across} are obtained by projecting the horizontal velocity onto the parallel and normal directions of the topographic gradient:

$$u_{\text{along}} = (u, v) \cdot \left(\frac{1}{|\nabla h|} \frac{\partial h}{\partial y}, -\frac{1}{|\nabla h|} \frac{\partial h}{\partial x} \right) \quad (1a)$$

$$v_{\text{across}} = (u, v) \cdot \left(\frac{1}{|\nabla h|} \frac{\partial h}{\partial x}, \frac{1}{|\nabla h|} \frac{\partial h}{\partial y} \right) \quad (1b)$$

where h is the topography, defined as a positive value representing the distance from the undisturbed sea surface to the seafloor. ∇ is the horizontal nabla operator, and u and v are the zonal and meridional velocities, respectively.

By substituting the terms of Equations 1a and 1b into the momentum equations, the along-isobath and across-isobath horizontal momentum equations can be written as follows:

$$\underbrace{\frac{\partial u_{\text{along}}}{\partial t}}_{\text{tendency}} = \underbrace{-u_{\text{along}} \frac{\partial u_{\text{along}}}{\partial x_{\text{along}}} - v_{\text{across}} \frac{\partial u_{\text{along}}}{\partial y_{\text{across}}} - w \frac{\partial u_{\text{along}}}{\partial z}}_{\text{advection}} + \underbrace{f v_{\text{across}}}_{\text{Coriolis}} + \underbrace{\left[\underbrace{\left(-g \frac{\partial \eta}{\partial x_{\text{along}}} \right)}_{\text{I}} + \underbrace{\left(- \int_z^0 \frac{g}{\rho_0} \frac{\partial \rho}{\partial x_{\text{along}}} dz \right)}_{\text{II}} \right]}_{\text{pressure}} + \underbrace{V^x_{\text{along}}}_{\text{viscosity}} \quad (2a)$$

$$\underbrace{\frac{\partial v_{\text{across}}}{\partial t}}_{\text{tendency}} = \underbrace{-u_{\text{along}} \frac{\partial v_{\text{across}}}{\partial x_{\text{along}}} - v_{\text{across}} \frac{\partial v_{\text{across}}}{\partial y_{\text{across}}} - w \frac{\partial v_{\text{across}}}{\partial z}}_{\text{advection}} + \underbrace{(-f u_{\text{along}})}_{\text{Coriolis}} + \underbrace{\left[\underbrace{\left(-g \frac{\partial \eta}{\partial y_{\text{across}}} \right)}_{\text{I}} + \underbrace{\left(- \int_z^0 \frac{g}{\rho_0} \frac{\partial \rho}{\partial y_{\text{across}}} dz \right)}_{\text{II}} \right]}_{\text{pressure}} + \underbrace{V^y_{\text{across}}}_{\text{viscosity}} \quad (2b)$$

where f is the Coriolis frequency, η is the sea surface height, ρ is the sea water density, $\rho_0 = 1,038 \text{ kg m}^{-3}$ is a constant reference density used in the model, $g = 9.8 \text{ m s}^{-2}$ is the gravitational acceleration, and V^x and V^y are the combination of viscosity and dissipation. V^x and V^y include the effects of the stress from winds and sea ice at the surface layer and friction at the bottom. The term on the LHS of denotes the tendency of the velocity. The first term on the RHS of denotes the advection term, the second term denotes the Coriolis term, the third term denotes the pressure term, and the last term denotes the viscosity and dissipation term. The pressure term consists of two components: (a) the barotropic component and (b) the baroclinic component. All terms in the momentum equations are derived from the daily outputs of the model spanning 2014–2023.

2.3. Lorenz Energy Cycle

To elucidate the dynamical mechanisms sustaining the ASC in the Ross Sea, we employ the Lorenz Energy Cycle (Lorenz, 1955; Storch et al., 2012; Wu et al., 2017) to quantify the energy budget. There are four energy reservoirs in the LEC, including the Mean Kinetic Energy (MKE, K_m), the Eddy Kinetic Energy (EKE, K_e), the Mean Available Potential Energy (MAPE, P_m), and the Eddy Available Potential Energy (EAPE, P_e). The four energy reservoirs are formulated as follows:

$$K_m = \int_V \frac{1}{2} \rho_0 (\bar{u}^2 + \bar{v}^2) dV \quad (3)$$

$$K_e = \int_V \frac{1}{2} \rho_0 (\overline{u'^2} + \overline{v'^2}) dV \quad (4)$$

$$P_m = - \int_V \frac{g}{2n_0} \overline{\rho'^2} dV \quad (5)$$

$$P_e = - \int_V \frac{g}{2n_0} \overline{\rho'^2} dV \quad (6)$$

where the overbar denotes the time mean (2014–2023), and the prime denotes the deviation of model daily outputs from the time mean. $\int_V dV$ indicates the volume integral. n_0 represents the vertical gradient of the time and area mean of local potential density. $\rho^*(x, y, z, t) = \rho(x, y, z, t) - \langle \rho(x, y, z, t) \rangle$, and $\langle \cdot \rangle$ represents the area mean at a

given depth, with the domain of area mean shown in Figure 1a. $\overline{\rho(x, y, z, t)}$ denotes the reference density defined by the area average of the time-mean density.

The energy budget equations are:

$$\frac{dK_m}{dt} = G(K_m) + C(K_e, K_m) + C(P_m, K_m) + D(K_m) + B(K_m) \quad (7)$$

$$\frac{dK_e}{dt} = G(K_e) + C(K_m, K_e) + C(P_e, K_e) + D(K_e) + B(K_e) \quad (8)$$

$$\frac{dP_m}{dt} = G(P_m) + C(P_e, P_m) + C(K_m, P_m) + D(K_m) + B(K_m) \quad (9)$$

$$\frac{dP_e}{dt} = G(P_e) + C(P_m, P_e) + C(K_e, P_e) + D(K_e) + B(K_e) \quad (10)$$

where the energy transfer pathways consist of the generation term (G), the conversion term (C), the dissipation term (D), and the boundary flux term (B). D denotes the energy dissipation rate induced by the viscosity, and it is calculated as the residual of Equations 7–10, respectively.

$G(K_m)$ and $G(K_e)$ denote the surface power input provided by the time-mean and time-varying stresses from both winds and sea ice. $G(P_m)$ and $G(P_e)$ denote the generation of the MAPE and the EAPE by surface buoyancy fluxes that create density variances through heat and freshwater fluxes. The rates of generation terms are calculated as follows:

$$G(K_m) = \int_s \overline{\tau_{x,s} u_s} + \overline{\tau_{y,s} v_s} dS \quad (11)$$

$$G(K_e) = \int_s \overline{\tau'_{x,s} u'_s} + \overline{\tau'_{y,s} v'_s} dS \quad (12)$$

$$G(P_m) = - \int_s \frac{\overline{\rho^*}}{n_0} B_0 dS \quad (13)$$

$$G(P_e) = - \int_s \frac{\overline{\rho'}}{n_0} B'_0 dS \quad (14)$$

where $\tau_{x,s}$ and $\tau_{y,s}$ are the surface stresses from both winds and sea ice, u_s and v_s are the surface velocities, and $\int_s dS$ denotes the area integral at the surface. $B_0 = \frac{\alpha Q^*}{\rho_0 C_w} + \rho_0 \beta S E^*$ represents the surface buoyancy flux, and α and β are the temperature and salinity expansion coefficients, respectively. C_w denotes the heat capacity of seawater, and S denotes the salinity. Q^* is the net air-sea heat flux, and E^* is the net air-sea freshwater flux.

The conversion terms are calculated as:

$$C(P_m, K_m) = - \int_V g \overline{\rho^* w} dV \quad (15)$$

$$C(K_e, K_m) = \int_V (\rho_0 \overline{u' u_h'} \cdot \nabla \bar{u} + \rho_0 \overline{v' u_h'} \cdot \nabla \bar{v}) dV \quad (16)$$

$$C(P_e, K_e) = - \int_V g \overline{\rho' w'} dV \quad (17)$$

$$C(P_e, P_m) = - \int_V \frac{g}{n_0} \overline{\rho' u_h'} \cdot \nabla_h \bar{\rho^*} dV \quad (18)$$

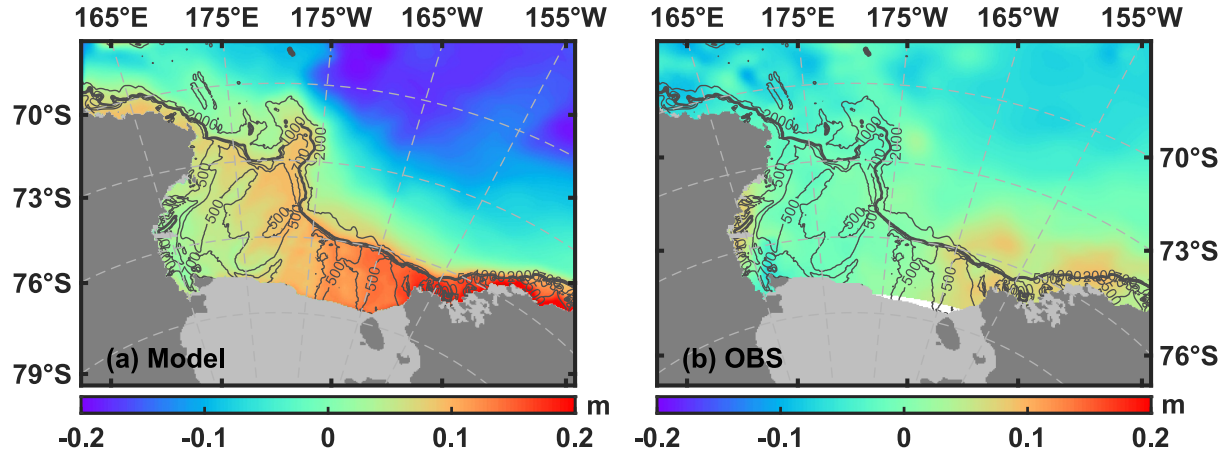


Figure 2. (a) Sea surface height from the model. The gray lines denote the isobaths (at 500, 700, 1,000, and 2,000 m). (b) As in (a), but from the CNES-CLS22.

where w is the vertical velocity, u_h is the horizontal velocity vector. A positive value of $C(X, Y)$ denotes that the energy is converted from X to Y , satisfying $C(X, Y) = -C(Y, X)$.

The boundary flux terms are calculated as:

$$B(K_m) = \oint_{\Omega^s} \left(\overline{u_h} \frac{1}{2} \rho_0 (\overline{u^2} + \overline{v^2}) + \overline{u_h P^*} \right) \cdot \mathbf{n} d\Omega^s \quad (19)$$

$$B(K_e) = -\oint_{\Omega^s} \left(\overline{u_h} \frac{1}{2} \rho_0 (u'^2 + v'^2) + \overline{u_h P'} \right) \cdot \mathbf{n} d\Omega^s \quad (20)$$

$$B(P_m) = \oint_{\Omega^s} \left(\overline{u_h} \frac{g}{2n_0} \overline{\rho^{*2}} - \overline{\rho^* u_h' \rho'} \frac{g}{n_0} \right) \cdot \mathbf{n} d\Omega^s U \quad (21)$$

$$B(P_e) = \oint_{\Omega^s} \left(\overline{u_h} \frac{g}{2n_0} \overline{\rho'^2} \right) \cdot \mathbf{n} d\Omega^s \quad (22)$$

where $P^*(x, y, z, t) = P(x, y, z, t) - \langle \overline{P(z)} \rangle$, $\langle \overline{P(z)} \rangle$ is the area average of the time-mean hydrostatic reference pressure, $\mathbf{n}$ denotes the unit normal vector of the boundary surface Ω , and Ω^s represents the area of the boundary surface.

3. Results and Discussion

3.1. Model Validation

In this study, we evaluated model performance through comprehensive hydrographic observations, with particular focus on the circulation and water mass properties over the continental shelf and slope. Figure 1b shows the comparison between the simulated velocities and observations from the AnSlope moorings near the mouth of the Drygalski Trough (DT) trough on the continental slope. The AnSlope moorings, while limited to depths >248 m (shallowest at M1), confirm the ability of this model to represent the ASC at depths. The model reproduces the velocity direction and magnitude in the lower layer (from ~ 200 m depth to the sea floor) over the western slope. Figure 2 compares the simulated SSH with observations from the CNES-CLS22, which is constructed based on satellite and in situ observations (Jousset, 2022). The model captures the main large-scale SSH features, including an elevated SSH over the eastern continental shelf and a depressed SSH northeast of the shelf, consistent with the cyclonic circulation of the Ross Gyre. While the strong cross-slope gradient in the model is not so evident in the CNES-CLS22, the simulated SSH over the continental slope shows good agreement with observational estimations based on the CryoSat-2 radar altimetry (Figure 4a in Armitage et al., 2018). This agreement gives us confidence in the representation of the ASC/ASF by the model. Figure 3 shows the comparison between the simulated potential temperature θ and observations from the Processes Regulating Iron

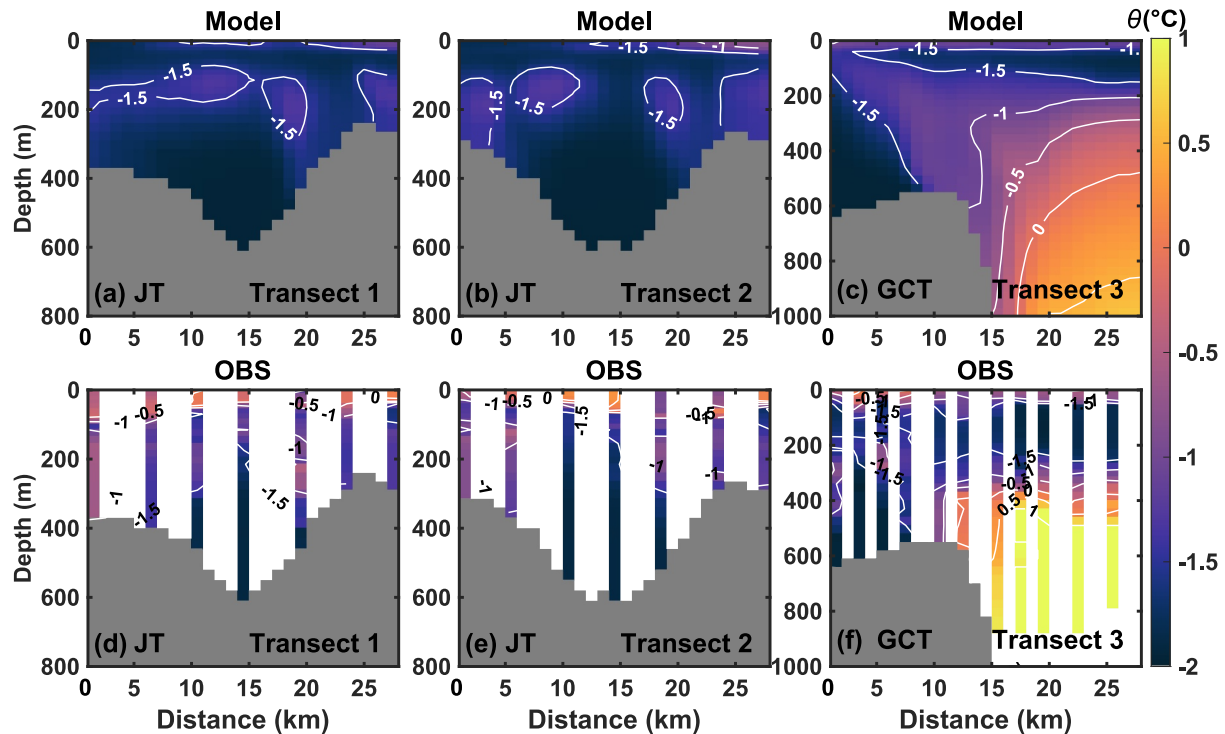


Figure 3. Comparisons of the potential temperature θ between model results (the top panels) and CTD and Expendable Bathythermograph (XBT) data (the bottom panels) from the Processes Regulating Iron Supply at the Mesoscale project (transects marked by green points in Figure 1a). The white lines indicate the θ , with a contour interval of 0.5°C . The left and middle panels show the zonal transects across the Joides Trough (JT), and the right panels show the meridional transect along the Glomar Challenger Trough (GCT). The CTD data at JT were collected in January 2012 and compared to the January climatology (2014–2023) of the model results. XBT data at GCT span December 2011–February 2012 and are compared to the December–February climatology (2014–2023) of the model results. The gray shading indicates the bedrock derived from the RTopo-2.

Supply at the Mesoscale (PRISM) project. Over the continental shelf, the model reproduces warm MCDW intrusions in the middle layers (~ 100 – 400 m depth) in the Joides Trough (JT) and Glomar Challenger Trough (GCT), and the model also reproduces cold DSW occupying the bottom layers (Figures 3a, 3b, 3d, and 3e). For the Antarctic Surface Water (AASW) in the surface layers in January, the simulated θ is approximately 1°C colder than observations. Over the continental slope, the model captures the transition from CDW to MCDW as it extends from the deep ocean to the GCT (Figures 3c and 3f). The simulated salinity S reproduces salty DSW in the deep troughs, although the simulated AASW is slightly fresher than observations (Figure S3 in Supporting Information S1). For a detailed evaluation of the simulated sea ice, see Figures S4–S5 in Supporting Information S1. Despite some biases between the model results and hydrographic observations, the model still reproduces the spatial distributions of water masses and velocities over the continental slope in the Ross Sea. Therefore, we propose that this model is useful for studying the spatial characteristics and the maintenance mechanism of the ASC/ASF in the Ross Sea.

3.2. The Spatial Pattern of the ASC/ASF

The simulated depth-averaged ASC, guided by topography, is generally westward (Figure 4a). Velocity measurements from Acoustic Doppler Current Profilers depict a westward flow throughout the water column over the slope in the north of the DT (Budillon et al., 2011; Visbeck & Thurnherr, 2009), and observed icebergs also drifted westward along the slope (Keys et al., 1990). Therefore, the simulated westward ASC agrees well with various observational results. Despite the depth-averaged ASC being westward, a distinct baroclinic structure with notable spatial asymmetry is also present, with an intensification in the upper 200 m over the eastern slope (Figure 4b) and a bottom intensification in the deeper layer over the western slope (Figure 4c). For clarity, velocity vectors at various vertical layers are shown in Movie S1.

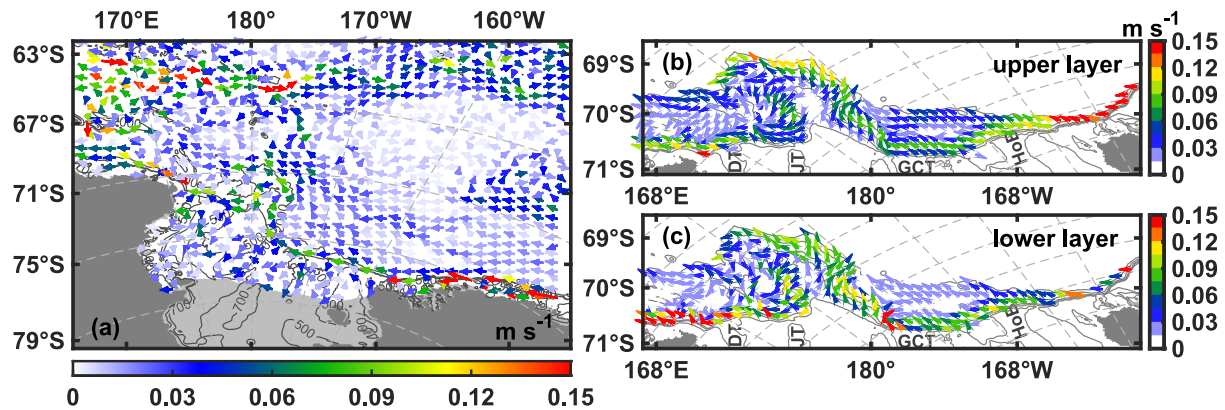


Figure 4. (a) The magnitude of time-mean (2014–2023) and depth-averaged velocity (U_h , $m s^{-1}$), with color shading indicating the velocity magnitude. Arrows show the directions of U_h . (b) The depth-averaged velocity vectors in the upper 200 m layer over the slope. (c) As in (b), but for the layer below 200 m depth.

The simulated barotropic streamfunction shows the horizontal distribution of large-scale circulations in the Ross Sea (Figure 5a). As the southern branch of the cyclonic Ross Gyre (broad negative values in the deep ocean in Figure 5a), the transport of the ASC is notably strong, evidenced by the strong cross-slope gradient in the barotropic streamfunction. Along the 1,000 m isobath, u_{along} shows the transformation of the ASC from a surface-intensified flow to a bottom-intensified flow (Figure 5b). According to the spatial patterns, the ASC can be classified into two regimes. In Regime I (the orange band in Figure 5b), the ASC flows into the Ross Sea from the

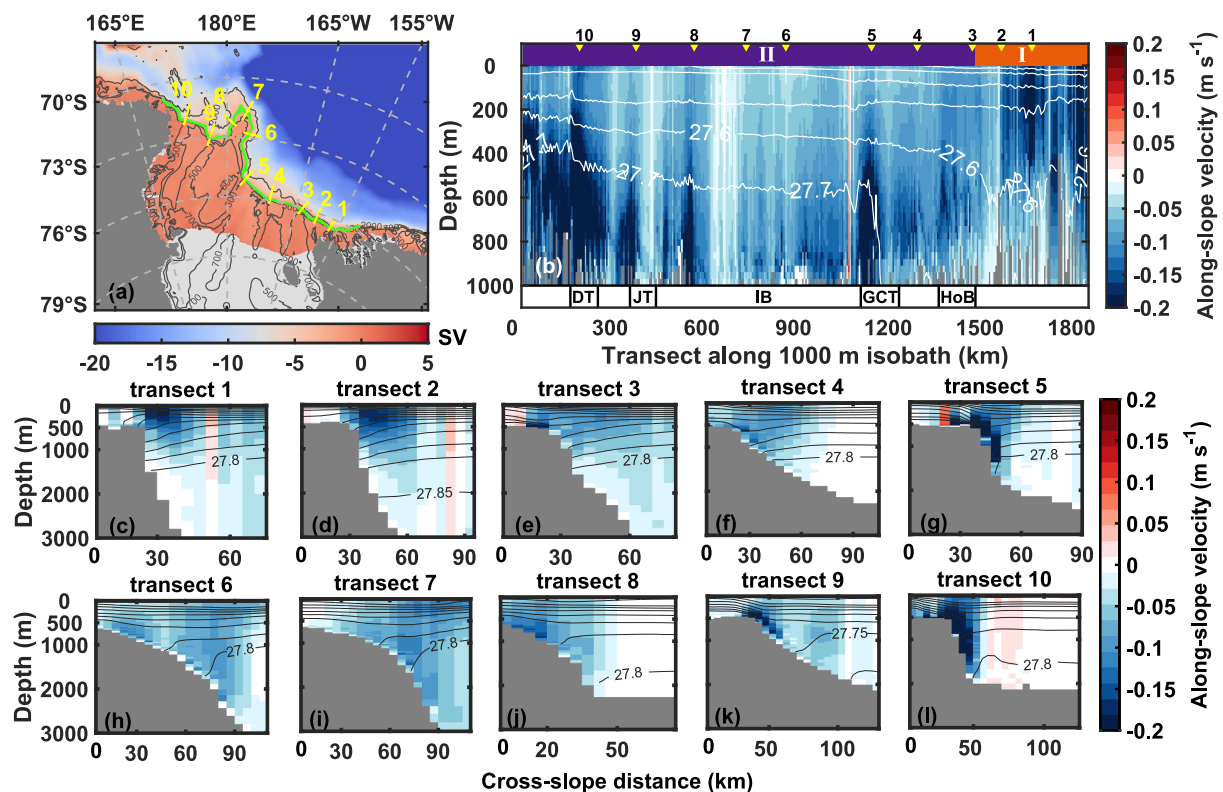


Figure 5. (a) The barotropic streamfunction ($1 Sv = 10^6 m^3 s^{-1}$), with positive (negative) values indicating anticyclonic (cyclonic) circulations. The thick green line denotes the 1,000 m isobath, where the along-slope transect is shown in (b). The algorithm for the detection of the 1,000 m isobath is described in the (Text S1 and Figure S6 in Supporting Information S1). The yellow lines show the cross-slope transects in (c–l). (b) u_{along} along the 1,000 m isobath, with negative values indicating that the current is to the left of the down-slope direction. Color bands on the top distinguish Regimes I and II. The spatial extensions of key topographic features to the south of the slope are labeled at the bottom. The white lines indicate the potential density σ_0 in contour intervals of $0.1 kg m^{-3}$. Yellow triangles denote the transect locations, which are also shown as yellow lines in (a). (c–l) u_{along} at the cross-slope transects, with negative values pointing to the left of the down-slope direction, and gray lines are σ_0 contours, with an interval of $0.05 kg m^{-3}$.

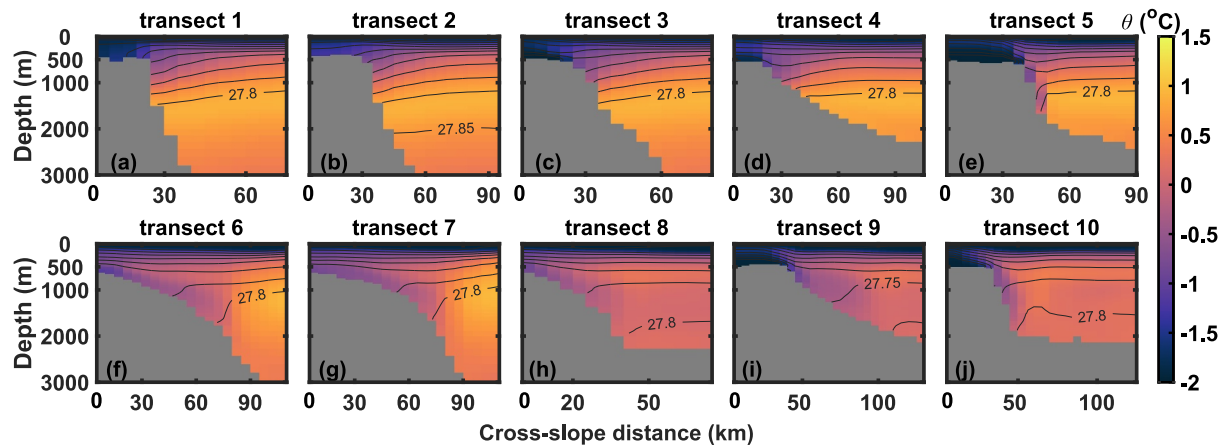


Figure 6. The distribution of the θ (color shading) along the cross-slope transects that are shown by yellow lines in Figure 5a. The black lines indicate the σ_0 isopycnals, with a contour interval of 0.05 kg m^{-3} . (a) θ and σ_0 along the transect 1. (b–j) As in (a), but the transects 2–10.

Amundsen Sea, featuring a surface-intensified westward flow. In Regime II (the purple band in Figure 5b), the deep flow gradually enhances, becoming a bottom-intensified flow to the north of the GCT. Similar bottom intensifications are also present further downstream, with a maximum present to the north of the DT.

To reveal the cross-slope extension of the ASC, 10 transects perpendicular to the slope are selected to show the u_{along} (Figures 5c–5l). In Regime I, the surface-intensified ASC at transects 1–2 is concentrated on the upper part of the continental slope (Figures 5c and 5d). As the ASC flows into Regime II, it becomes more bottom-intensified and more concentrated at the continental shelf break (Figure 5e). Thereafter, the meridional structure of the ASC seems to be associated with the strength of the bottom intensification. An enhanced undercurrent tends to sit on the shelf break and the upper slope (Figures 5g, 5k, and 5l), while a slowdown of the bottom flow tends to extend farther north (Figures 5f, 5h–5j). The bottom intensifications are associated with DSW outflows, which have also been found by Huneke et al. (2022), with a “V” shape of the isopycnals present over the slope (Thompson et al., 2018).

The cross-slope structure of the ASF is shown by θ and S in Figures 6 and 7, together with the isopycnals. In Regime I, cold and fresh shelf water occupies the upper layer on the outer continental shelf, with isopycnals of the ASF intersecting the slope (Figures 6a, 6b, 7a, and 7b). This is characterized as a Fresh Shelf region (Thompson et al., 2018). In Regime II, the cold and fresh shelf water is gradually replaced by cold and saltier shelf water (Figures 6c–6e, 6i, 6j, 7c–7e, 7i, and 7j). At transects 5, 9, and 10 that are across the western flanks of the GCT, JT, and DT, DSW approaches the shelf break and descends the continental slope (Figures 6e, 6i, and 6j; Figures 7e, 7i, and 7j). Overall, the ASF in Regime II is characterized as a Dense Shelf region (Thompson et al., 2018).

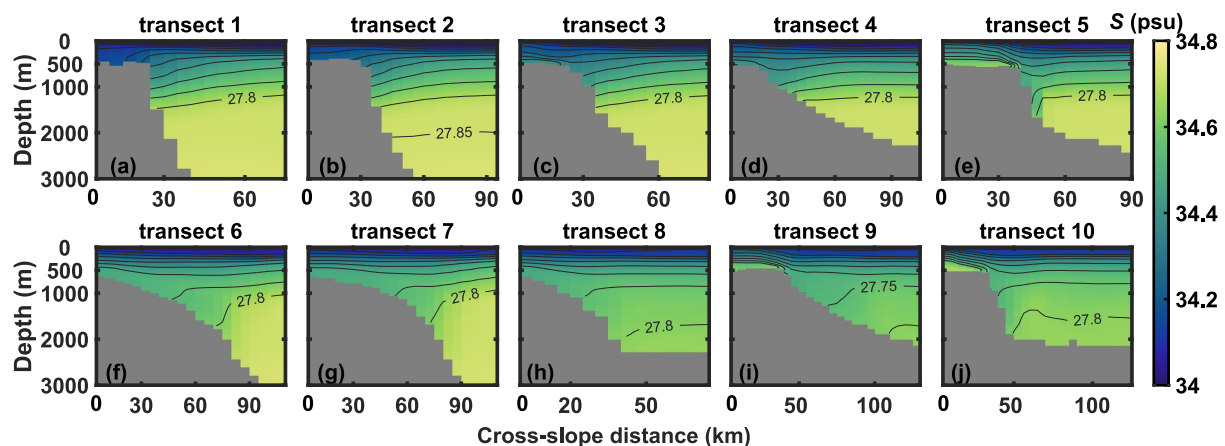


Figure 7. As in Figure 6, but for S .

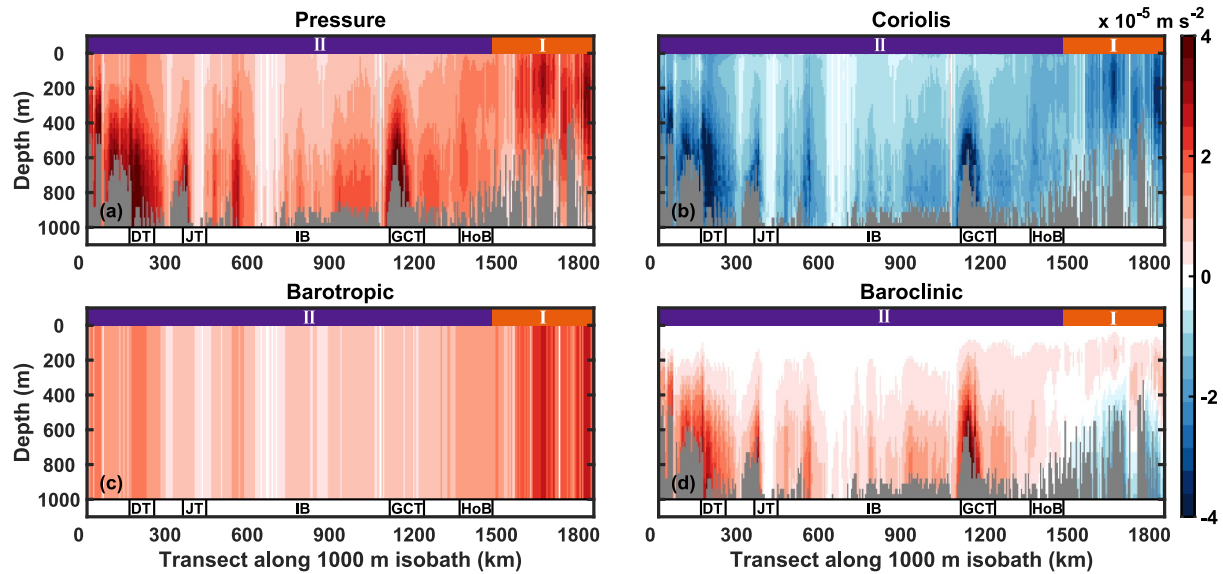


Figure 8. (a) The pressure term of the momentum budget of u_{along} along the 1,000 m isobath. (b) As in (a), but for the Coriolis term (c)–(d) The barotropic and baroclinic components of the pressure term, respectively.

3.3. The Momentum Budget

The analysis of the momentum budget reveals the momentum balance of the ASC (Figure 8). Along the 1,000 m isobath, u_{along} is primarily determined by the balance between the pressure gradient and Coriolis terms (Figures 8a and 8b), while the influences of the advection and viscosity terms are much weaker (Figure S7 in Supporting Information S1), implying that the ASC is largely in the geostrophic balance. Since the ASC is mainly driven by polar easterlies (Gill, 1973), the ASC is expected to be directed westward around Antarctica, with higher pressure on the onshore side of the ASC (Figure 8a), except in the West Antarctica Peninsula, where the ACC hugs the slope (Thompson et al., 2018).

The pressure gradient force consists of barotropic and baroclinic terms (Figures 8c and 8d). The barotropic term is related to the cross-slope gradient of η , while the baroclinic term is associated with the cross-slope gradient of ρ . The positive barotropic term over the slope is in accordance with the prevailing polar easterlies (Figure 9), with a maximum in Regime I (Figure 8c). Driven by easterlies, the poleward Ekman drift piles up water on the shoreward side of the slope and results in a negative cross-slope gradient of η . Then, according to the geostrophic balance (Equation 2b), the positive barotropic pressure is balanced by a negative (westward) barotropic geostrophic flow. The strength of this positive barotropic term exhibits spatial variation, regulated by the distance between the slope and the coastline. In Regime I, the close coastline confines the easterly-driven poleward Ekman transport, establishing a strong gradient of η and consequently enhancing the positive barotropic contribution to the ASC. As the flow shifts westward into Regime II, the increasing distance from the coastline weakens the impact of coastal Ekman convergence, thereby attenuating the η gradient and reducing the magnitude of the positive barotropic term (Figure 8c).

In Regime I, the isopycnals above the 1,000 m isobaths tilt downward and southward (Figures 6a–6c and 7a–7c), resulting in a positive cross-slope gradient of ρ . Then, the baroclinic term gradually decreases with depth, with a negative baroclinic term in the deep layer, contributing to an intensified flow in the upper layer and a weakened flow in the lower layer. In Regime II, the southern flank of the “V” shape isopycnals tilts upward and southward (Figures 6d–6j and 7d–7j), resulting in a negative cross-slope gradient of ρ . Then, the baroclinic term gradually increases with depth, with a positive contribution of the total pressure term in the deep layer, contributing to an intensified bottom flow. The formation of the “V” shape isopycnals should be attributed to cross-slope DSW overflows that raise the isopycnals over the slope. Therefore, large values of the baroclinic term are present in the DT, JT, and GCT that are recognized as DSW pathways to the slope (Morrison et al., 2020; Smith et al., 2012).

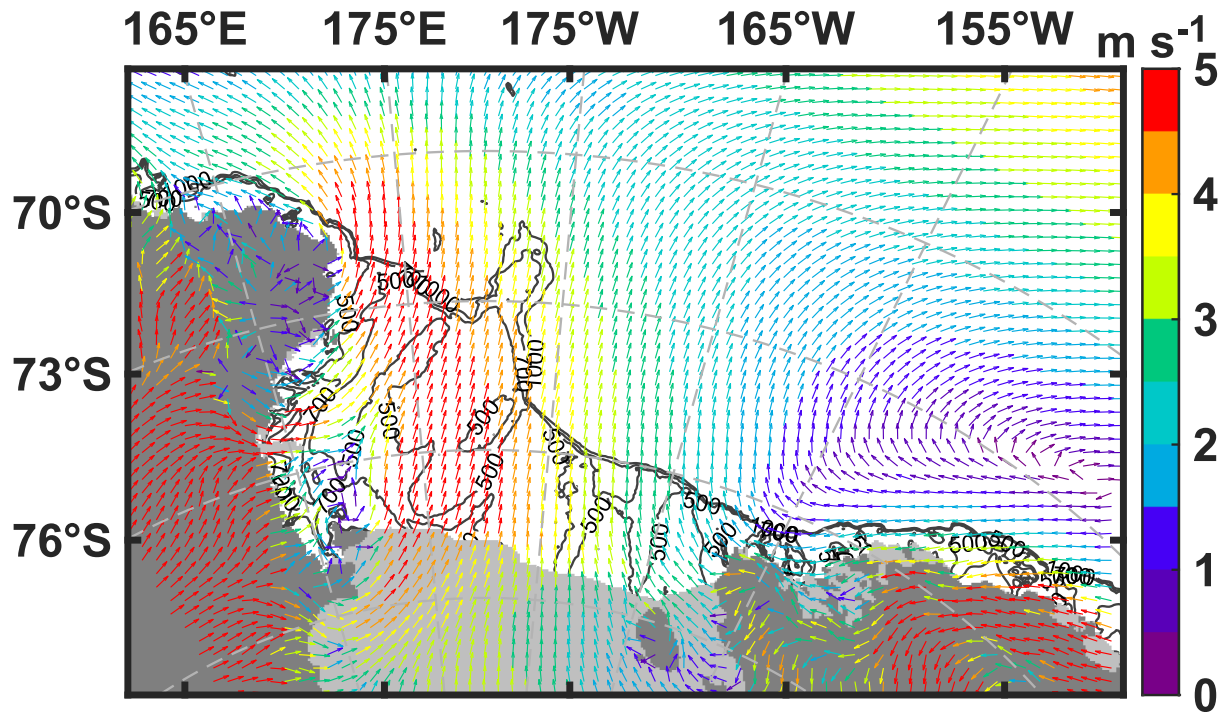


Figure 9. The time mean (2014–2023) winds from ERA5. Arrows show the directions of the wind, and color shading indicates the wind magnitude.

3.4. The Lorenz Energy Cycle Over the Slope

Using Equations 3–6, we analyze the four energy reservoirs, K_m , K_e , P_m , and P_e (Figure 10). As the time-mean kinetic energy reservoir, K_m quantifies the strength of the ASC. Over the slope, large K_m values are confined within narrow filaments along the 1,000 m isobath (Figure 10a). In Regime I, the maximal band is present in the

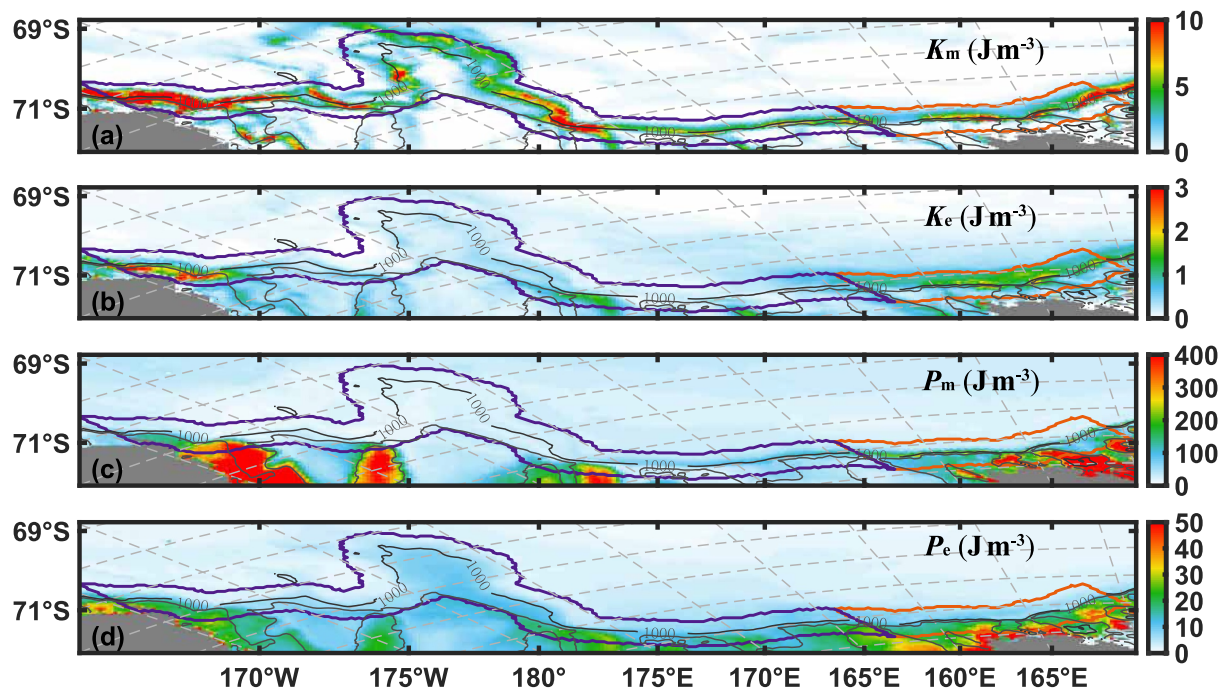


Figure 10. (a) Distributions of the depth-averaged K_m density. The areas enclosed by orange and purple lines represent the domains where the volume-integrated energy budget is calculated for Regimes I and II in Figure 13 (b)–(d) As in (a), but for K_e , P_m , and P_e , respectively.

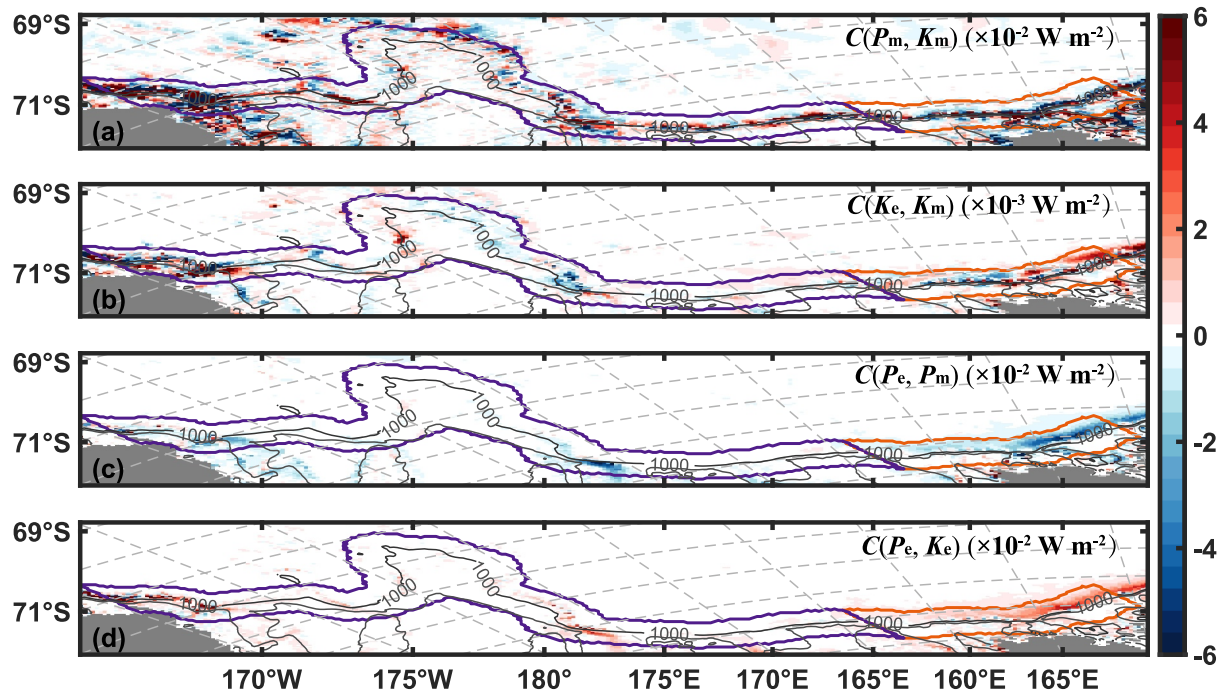


Figure 11. As in Figure 10, but for the vertically integrated energy conversion terms, (a) $C(P_m, K_m)$, (b) $C(K_e, K_m)$, (c) $C(P_e, P_m)$, and (d) $C(P_e, K_e)$.

eastern sector where the surface-intensified ASC is most pronounced (Figure 5b). In Regime II, the maximal bands of K_m are present to the northwest of the GCT, JT, and DT, coinciding with the pathways of DSW overflows (Figure 10a). Compared to K_m , K_e is much weaker (Figure 10b). In Regime I, larger K_e spreads to the north of the 1,000 m isobath. In Regime II, relatively large values of K_e are only present northwest of the DT, where the densest DSW spills over the slope in the Ross Sea. In contrast to K_m , large values of P_m are concentrated within submarine troughs over the continental shelf rather than over the slope (Figure 10c). In Regime I, strong P_m is located to the south of the 1,000 m isobath. In Regime II, relatively large values of P_m are identified in the GCT, JT, DT, and the upper slope downstream, which is filled by DSW. Although the values of P_e are weaker than P_m , the spatial distribution of P_e is similar to P_m (Figure 10d).

Vertical integrals of the energy conversion terms are shown in Figure 11. Both $C(P_m, K_m)$ and $C(K_e, K_m)$ exhibit their elevated values over the slope, yet $C(K_e, K_m)$ is roughly an order of magnitude weaker than $C(P_m, K_m)$ (Figures 11a and 11b). Since the sign of $C(P_m, K_m)$ and $C(K_e, K_m)$ often changes within a short distance, it is not easy to quantify the contributions of $C(P_m, K_m)$ and $C(K_e, K_m)$ without a domain integration. However, the similarity of the spatial distributions between $C(P_m, K_m)$ and K_m still implies that the release of P_m may play an important role in driving the ASC in the Ross Sea (Figures 10a and 11a). Over the slope, the baroclinic production is dominated by negative $C(P_e, P_m)$ and positive $C(P_e, K_e)$ (Figures 11c and 11d), implying that the energy cascade in the baroclinic pathway is $P_m \rightarrow P_e \rightarrow K_e$ (Chen et al., 2016; Stewart & Thompson, 2016).

The generation terms represent the energy production from the surface forcing (Figure 12). $G(K_m)$ and $G(K_e)$ represent the power supplied by the mechanical stress (from the winds or sea ice), and $G(P_m)$ and $G(P_e)$ represent the power supplied by the buoyancy flux. Over the slope, both the time-mean winds and sea ice velocity are generally northwestward (Figure 9 and Figure S8 in Supporting Information S1), implying that both wind and sea ice stresses contribute to the positive values of $G(K_m)$ over most of the slope, except in the western flank of the IB, where the ASC flows southward along the convoluted slope. Downstream of the DT, strong easterlies (Figure 9) align with the westward surface flow (Figure 4b), maximizing energy transfer into K_m (Figure 12a). $G(K_e)$ is broadly positive over the slope, and is maximum downstream of the DT (Figure 12b), highlighting the role of time-varying winds in forcing the time-varying currents downstream of the DT. $G(P_m)$ is weak over the slope (Figure 12c), implying that time-mean buoyancy fluxes over the continental slope are relatively weak, while $G(P_e)$ is comparable with $G(K_e)$ except downstream of the DT (Figure 12d). The high time-mean sea ice

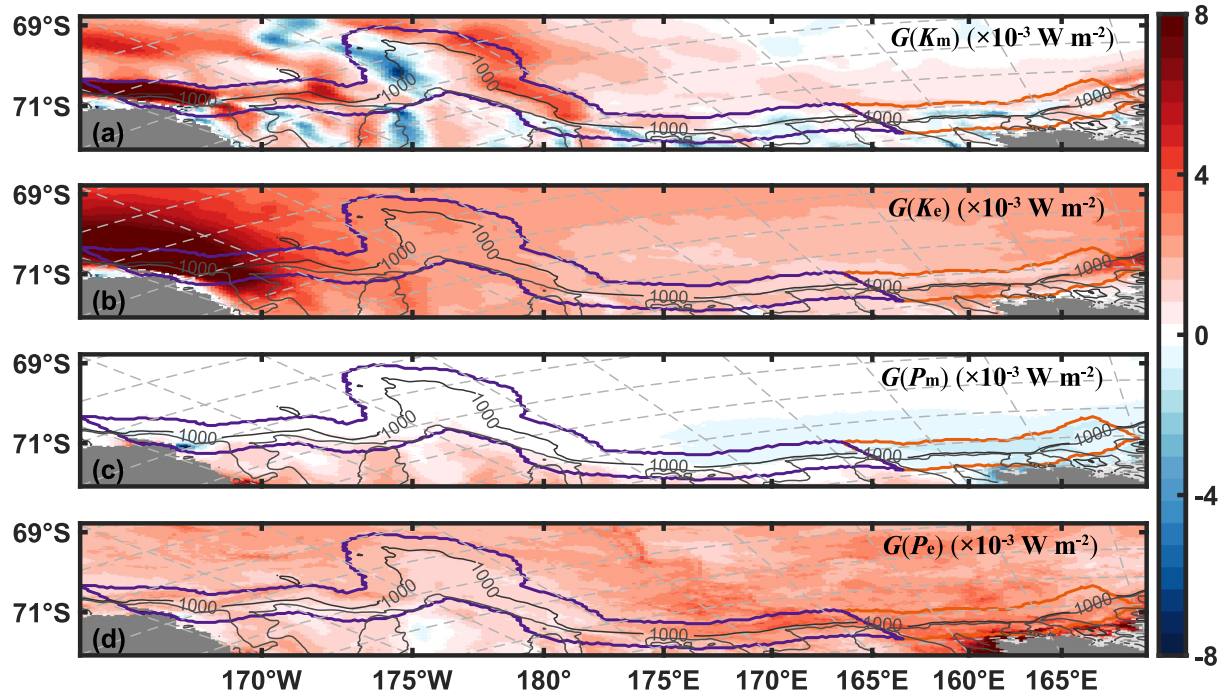


Figure 12. Horizontal distributions of the (a) $G(K_m)$, (b) $G(K_e)$, (c) $G(P_m)$, and (d) $G(P_e)$, respectively.

concentration over the slope likely contributes to the relatively weak $G(P_m)$ by insulating the ocean from the atmosphere.

We summarize the energy budgets by quantifying the spatially integrated Lorenz Energy Cycle terms in Regimes I and II (Figure 13). Since this study focuses on the time-mean ASC, we mainly aim to clarify the budget of K_m . In Regime I, the primary energy source of K_m is $B(K_m)$, which amounts to ~ 163.2 MW (Figure 13a), which is associated with the advective flux and the pressure work (Jüling et al., 2018; Wu et al., 2017). An inverse energy cascade pathway from K_e to K_m has been identified, although $C(K_e, K_m)$ is relatively weak and amounts to ~ 15.1 MW. The power input by surface stress and the energy dissipation are also weak and comparable, with G

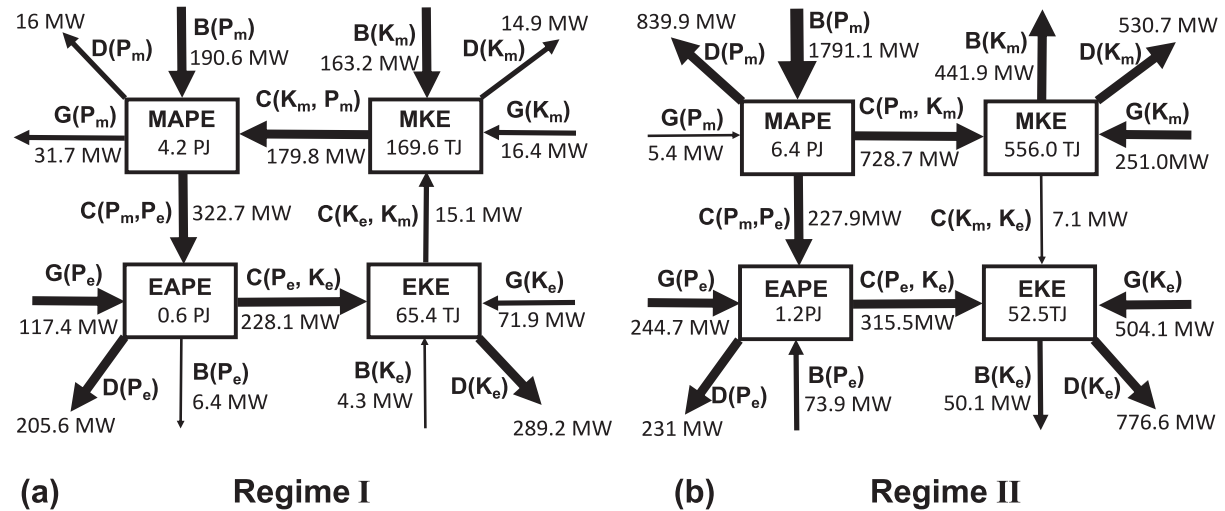


Figure 13. Domain-integrated energy budgets for Regimes I and II, respectively. Energy reservoirs are in units of PJ or TJ, $1 \text{ PJ} = 10^{15} \text{ J}$; $1 \text{ TJ} = 10^{12} \text{ J}$. Rates of generation, conversion, dissipation, and boundary fluxes are in units of MW, $1 \text{ MW} = 10^6 \text{ W}$. Arrows indicate the directions of energy transfer, with four thicknesses proportional to four orders of the flux magnitude, ranging from 10^0 MW to 10^3 MW .

(K_m) of ~ 16.4 MW and $D(K_m)$ of ~ 14.9 MW. Overall, in Regime I, a pronounced energy flow from K_m to P_m has been identified, implying that the ASC plays an important role in maintaining the existence of the ASF. There is an energy flow from K_e to K_m , yet the magnitude is negligible.

In Regime II, the primary energy source of K_m is P_m , with $C(P_m, K_m)$ reaching ~ 728.7 MW (Figure 13b). Such a strong energy conversion is roughly three times larger than $G(K_m)$, which amounts to ~ 251.0 MW. Yet, the energy cascade from K_m to K_e is two orders of magnitude smaller than $C(P_m, K_m)$, with $C(K_m, K_e)$ only reaching ~ 7.1 MW. Consequently, the kinetic energy dissipation and boundary flux are substantially enhanced, with $B(K_m)$ and $D(K_m)$ up to ~ 441.9 and ~ 530.7 MW, respectively. So, in Regime II, the ASC is mainly energized by the potential energy released from the ASF in the energy pathway $P_m \rightarrow K_m$.

4. Summary and Conclusions

In this study, we employ an eddy-permitting coupled ocean-sea ice-ice shelf regional model to investigate the spatial characteristics of the time-mean ASC in the Ross Sea. Based on the analysis of the momentum budget and the Lorenz Energy Cycle, we also explore the dynamical mechanisms governing the ASC.

Steered by the topography, the simulated ASC generally flows westward in the Ross Sea, yet two distinct structures of the ASC have been identified. In Regime I, the westward ASC features a surface intensification, and the core of the current is concentrated on the upper slope, with isopycnals intersecting on the upper slope. In Regime II, the ASC features a bottom intensification which tends to reside on the continental shelf break, with a “V” shape of isopycnals over the slope. Therefore, Regime I can be classified as the typical Fresh Shelf region, and Regime II as the Dense Shelf region (Thompson et al., 2018).

Along the 1,000 m isobath, the momentum budget of the ASC is dominated by the pressure gradient and Coriolis force. The barotropic pressure term dominates the flow direction, while the baroclinic pressure term contributes to the vertical shear. Driven by easterlies, the positive barotropic pressure term corresponds to a westward geostrophic flow. In Regime I, the baroclinic pressure term decreases with depth, resulting in a weakened bottom flow and an intensified surface flow. In Regime II, the baroclinic pressure term increases with depth on the onshore flank of the “V” shape ASF, resulting in an intensified bottom flow. Such bottom intensifications occur where DSW overflows down the slope.

There are also notable differences in the energy sources, transfers, and sinks of the ASC between Regimes I and II. In Regime I, the time-mean ASC is mainly sustained by the boundary advection and pressure work, with an energy pathway $P_m \leftarrow K_m$. The contributions of the generation term, dissipation term, and conversion from K_e are all negligible, while K_m is largely converted to P_m to enhance the gradient of the ASF. In contrast, in Regime II, K_m of the time-mean ASC sources energy from P_m in the energy cascade pathway $P_m \rightarrow K_m \rightarrow K_e$. The conversion between K_m and K_e is still weak, yet the dissipation term is considerably enhanced. Meanwhile, inferring from the boundary term, the advection of K_m downstream also plays an important role in the energy balance.

Our investigation has some limitations. The horizontal resolution of this model is insufficient to fully resolve mesoscale eddies in the Ross Sea (Mack et al., 2019), limiting the model representation of transient processes. Specifically, the time-varying terms in the Lorenz Energy Cycle may be underestimated by the model, and therefore, we have not discussed much of the time-varying terms in this study. Yet, Mack et al. (2019) have also shown that the relative energy partition between K_m and K_e did not change much over the continental shelf when the horizontal resolution increased from 5 to 1.5 km. Moreover, the influence of high-frequency processes (e.g., topographic Rossby waves, submesoscale eddies, and internal waves) on the location and strength of the ASC should be revisited in future studies. Therefore, future modeling studies focused on the ASC in the Ross Sea would benefit from an increase in both vertical and horizontal resolutions.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

The ERA5 product is available from the Copernicus Climate Data Store at <https://doi.org/10.24381/cds.adbb2d47> (Hersbach et al., 2023). MEOP Seal-CTD data is collected by the Integrated Marine Observing System (<https://meop.net/database/meop-databases/meop-ctd-database.html>). The RTopo-2 data is available through the PANGAEA Data Publisher for Earth and Environmental Science at <https://doi.pangaea.de/10.1594/PANGAEA.856844> (Schaffer & Timmermann, 2016). Southern Ocean State Estimate data is obtained from the Scripps Institution of Oceanography (http://sose.ucsd.edu/sose_stateestimation_data_05to10.html). WOA18 data is available through the National Centers for Environmental Information (<https://www.ncei.noaa.gov/access/world-ocean-atlas-2018/>). The AnSlope data is available from here <https://ocp.ldeo.columbia.edu/res/div/ocp/projects/anslope/Data.html>. The PRISM project is available from https://science.who.edu/users/olga/PRISM_RS/PRISM_RS.html. The CNES-CLS22 MDT data set is available from Jousset (2022) at <https://doi.org/10.24400/527896/a01-2023.003>. The AMSR-E ASI sea ice concentration data are accessible via the PANGAEA Data Publisher for Earth and Environmental Science at <https://doi.org/10.1594/PANGAEA.919778> (Melsheimer & Spreen, 2020). Sea ice data from the National Snow and Ice Data Center is available at <https://doi.org/10.7265/N5K072F8> (Fetterer et al., 2017). The model configuration and the simulated data used in this study are available at <https://doi.org/10.5281/zenodo.16028788> (Liu, 2025).

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