Modulation of dense shelf water salinity variability in the western Ross Sea associated with the Amundsen Sea Low

Guijun Guo1,2, Libao Gao1,2 and Jiuxin Shi2,3

1 Center for Ocean and Climate Research, First Institute of Oceanography, Qingdao, People’s Republic of China
2 Qingdao National Laboratory for Marine Science and Technology, Qingdao, People’s Republic of China
3 Physical Oceanography Laboratory, Ocean University of China, Qingdao, People’s Republic of China

E-mail: guoguijun@fio.org.cn and gaolb@fio.org.cn

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Abstract

Dense shelf water (DSW) produced in the western Ross Sea (RS) is one of the major sources of Antarctic bottom water (AABW). Thus the understanding of long-term variability of DSW salinity and its controlling factors in the western RS is critical to assess the variability of globally distributed AABW. Here we analyze a long time record of hydrographic data (1984–2020) collected in the western RS, as well as sea ice drift vectors, surface wind speed, sea level pressure and Amundsen Sea low (ASL) indices. We confirm recent findings that there is a rapid increase of DSW salinity in the western RS after a minimum in 2013, although the DSW has experienced substantial freshening in the past few decades, indicating a significant multidecadal variability of DSW salinity in the western RS. Over the past four decades, multidecadal variability in the DSW salinity has been strongly coupled with westward zonal flow changes along the coastal current, and the post-2013 rapid enhancement of DSW salinity is accompanied by reduced freshwater input due to weakening of the westward zonal flow from the upstream Amundsen Sea (AS) into the RS. Large-scale circulation determining the strength of the zonal flow is closely linked to the ASL variability. The accelerated deepening of the ASL and the resulting southwestward extension of low pressure induce an eastward coastal current anomaly. This reduces the freshwater input from the AS to the RS and is responsible for the subsequent enhancement of DSW salinity in recent years in the western RS. These dynamical processes demonstrated here explain how the ASL changes modulate the DSW salinity in the western RS, and will help to understand the implication of climate changes in the Southern Ocean on AABW formation.

1. Introduction

The Southern Ocean is considered to be sensitive to climate changes and is warming more quickly than lower latitudes (Böning et al 2008, Gille 2008, Gao et al 2018). This rapid warming is ascribed to the remarkable strengthening of the circumpolar west-erlies, which is thought to increase the upwelling rate (Le Quéré et al 2007, Meredith et al 2012) of warm circumpolar deep water (CDW), as well as the transport of CDW onto the Antarctic continental shelves (Thoma et al 2008, Dinniman et al 2012, Spence et al 2014). This increase in oceanic heat over the shelf brought by CDW intrusion drives an increasing basal melt rate of the ice shelves and increases the freshwater input to shelf regions (Pritchard et al 2012, Rignot et al 2013, Paolo et al 2015, Rintoul et al 2016, Adusumilli et al 2020), which in turn reduces the density of dense shelf water (DSW), suppresses Antarctic bottom water (AABW) formation (Williams et al 2016) and has a profound implication for global thermohaline circulation (Lumpkin and Speer 2007, Marshall and Speer 2012).

The Ross Sea (RS, figure 1(a)), the largest embayment of the Southern Ocean around Antarctica, provides the densest DSW on the continental shelf resulting from the vigorous release of brine through sea ice production during autumn and winter (Jacobs et al 1970, Orsi and Wiederwohl 2009).
Figure 1. (a) Map of the Ross Sea and the observational stations in the Terra Nova Bay polynya (TNBP, dashed box). The inset zooms into the TNBP stations, with sampling years color coded the same as in (b) and (c). White dashed and gray lines in (a), respectively, represent the general currents along the shelf break and on the shelf, and thick blue arrows indicate the paths of CDW intrusion onto the shelf (referred to Budillon et al. (2003), Smith et al. (2014)). The thick red line represents the section at the AS/RS gate for zonal flow estimates. (b) and (c), Annual mean profiles of salinity and temperature, color coded by sampling years. Salinity profiles increased rapidly after 2013 are highlighted by triangles. Surface freezing temperature (~−1.9 °C) is shown as a vertical dashed line in (c). Topographical features in (a) are: TNB = Terra Nova Bay; DIT = Drygalski Ice Tongue; DT = Drygalski Trough.

makes the RS an optimal site for AABW formation and export to the global ocean (Orsi et al 1999, Gordon et al 2009, 2015), contributing ~25% of all AABW formation on Antarctic shelves (Orsi et al 2002). DSW outflows in the RS occur at the Drygalski Trough and the Glomar-Challenger Trough, where the DSW is dominated by high salinity shelf water and ice shelf water respectively. High salinity
shelf water in the northern Drygalski Trough has freshened by −0.06 between 1995 and 2006, while the ice shelf water in the northern Glomar-Challenger Trough has freshened by −0.04 during 1998–2006 (Budillon et al 2011). Changes of these thermohaline characteristics in the RS linked to AABW formation will likely have substantial impacts at the global scale by impacting thermohaline circulation (Johnson et al 2008, Orsi and Wiederwohl 2009, Jacobs and Giulivi 2010).

The ocean circulation on the RS continental shelf consists of two main inflows from the east, the Antarctic Coastal Current and Antarctic Slope Current (figure 1(a)) (Budillon et al 2003, Smith et al 2014), which are driven by easterly wind along the coast. The westward inflows are critical to carry freshwater from the upstream Amundsen Sea (AS) into the RS, contributing to the long-term variability of DSW salinity in the RS (Jacobs and Giulivi 2010, Nakayama et al 2014). In the past few decades, ocean measurements in the RS have revealed marked decreases (−0.03 decade⁻¹) in the salinity of DSW (Jacobs et al 2002, Jacobs and Giulivi 2010). However, beginning in 2014 the DSW salinity in the RS has rapidly increased (Castagno et al 2019). In addition to changes in the freshwater inflow from upstream, several other mechanisms are also potentially responsible for the variability of DSW salinity, such as sea ice production in coastal polynyas, Ross Ice Shelf melting, changes in CDW intrusion onto the shelf region and precipitation (Jacobs et al 2002, Castagno et al 2019).

Sea ice production in Terra Nova Bay polynya (TNBP) and RS polynya (RSP) shows little change with time (Tamura et al 2016), and is believed to govern the DSW salinity only within the seasonal cycle (Assmann and Timmermann 2005). Changes in CDW intrusion onto shelf (Castagno et al 2017, 2019), precipitation and Ross Ice Shelf melting (Jacobs et al 2002, Jacobs and Giulivi 2010) are also believed to have little impact on DSW salinity. Basal melt rates of ice shelves in West Antarctica have large interannual and decadal variability (Paolo et al 2018, Adusumilli et al 2020). During 1994–2018, meltwater fluxes from basal melt of ice shelves in the AS sector increase to a maximum in 2009 and then decline incrementally to near steady-state values after that (Adusumilli et al 2020). However, mass loss of ice sheet in West Antarctica has increased from 1979 to 2017 (Shepherd et al 2018, Rignot et al 2019), indicating an increasing freshwater flux into the ocean via basal melt of ice shelves as well as iceberg calving (Depoorter et al 2013). This would decrease rather than increase the salinity of DSW in the western AS after 2014 (Castagno et al 2019). Therefore, long-term freshening of DSW in the RS is ascribed to increased freshwater transport from the upstream AS in earlier studies (Jacobs and Giulivi 2010, Budillon et al 2011), and the rapid increase of DSW salinity since 2014 is also related to large-scale forcing that reduces the freshwater input to the RS (Castagno et al 2019). The westward coastal current is critical to freshwater inflow to the RS (Nakayama et al 2014). However, in the presence of sea ice in Antarctic regions, satellite altimetry cannot measure SSH well enough to get geostrophic currents, hence the exact mechanism linking the DSW salinity variability to large-scale circulation and climate phenomena remains largely uncertain.

Several studies have documented the possible link between the advection of low-salinity water from increased mass loss of ice sheets upstream and the long-term freshening of the RS (Nakayama et al 2014, Smith et al 2014, Dinniman et al 2018). In contrast, Assmann and Timmermann (2005) argue that the steady decrease in DSW salinity reported by Jacobs et al (2002) is an aliasing artifact due to irregular sampling and deduce a multidecadal oscillation in DSW salinity superimposed on the long-term freshening trend from model results. The long time series of DSW salinity in Jacobs et al (2002) and Assmann and Timmermann (2005) are derived from the RSP north of Ross Island, which is one of the most densely sampled regions at that time. However, the DSW formed in the RSP (S < 34.76) is less saline than that formed in the TNBP (S > 34.8) (Orsi and Wiederwohl 2009), and changes in the DSW of the TNBP are believed to have greater effects on AABW formation than that of the RSP (Jenderisie et al 2018). Thus, understanding the long-term changes in DSW in the TNBP and their governing factors is critical to assess AABW changes in Antarctic region.

A substantial component of interannual atmospheric variability in the Pacific Sector of Antarctica, as a potential driver of ocean variability, is associated with the AS low (ASL) (Turner et al 2017, Raphael et al 2019). The ASL is a highly dynamic and mobile climatological low pressure system located in the Pacific Sector of the Southern Ocean (Hosking et al 2013). The intensity and location of the ASL play a key role in modulating the zonal and meridional winds between the Antarctic Peninsula and the RS (Turner et al 2017, Paolo et al 2018, Raphael et al 2019), thus the wind-driven circulation and horizontal transport in this region are also expected to be related to the ASL changes. Here we used the hydrographic datasets collected in the TNBP over the past four decades, as well as the available sea ice drift vectors, surface wind speed, sea level pressure (SLP) and the ASL indices, to investigate how the ASL changes modulate DSW salinity in the western RS through affecting large-scale circulation. In particular, we investigated the response of westward zonal current on the shelf to the ASL changes, which determines the freshwater input from the upstream AS into the RS, to provide new insight into the impacts of climate changes on DSW production, which is critical to AABW formation.
2. Data and methods

2.1. Hydrographic data

The temperature and salinity profiles collected in the TNBP during 17 years within the period of 1984–2020 were used in this study (figure 1). Profiles of years before 2013 were obtained from the World Ocean Database (WOD13, table S1 (available online at stacks.iop.org/ERL/16/014004/mmedia)) (Boyer et al 2013), and only data records with maximum depths larger than 600 m were used. There were also available datasets in 2010 and 2011 in WOD13 collected by instrumented seals, but they were not used in this study due to their relatively low salinity accuracy (~0.05) (Roquet et al 2014). The hydrographic data profiles in and after 2013 were collected by Chinese National Antarctic Research Expedition cruises using a Sea-Bird SBE 911 plus conductivity–temperature–depth (CTD) profiler, except for the XCTD deployment in 2018. The XCTD data was validated by a comparison test during the summer cruise in 2020, which showed that the salinity profiles derived from two XCTDs agreed well with that obtained from the SBE 911 plus CTD profiler (figure S1(a)) and the salinity differences were generally less than 0.01 (figure S1(b)). To minimize seasonal aliasing, only data from the more heavily sampled summer months (table S1) were used and all the profiles were interpolated to 10 m bins linearly for an easier comparison. This work focused on deep layers below 600 m to eliminate the effects of short-term processes at the surface. The horizontal standard deviation (std) of the salinity among all stations in each year below 600 m were generally less than 0.01 (figure S2), while the std at each layer was set to 0 when only one profile was available in a particular year. The highest std occurred in 1988, in which the observational stations covered the largest area in TNBP (inset map in figure 1(a)), but the averaged std value over the layer below 600 m was still less than 0.01. This is small compared to the scale of the multidecadal variability of DSW salinity (figure 1(b)), indicating that the spatial differences for observational stations are unlikely to alias the results.

2.2. Ocean current calculation

Daily surface wind velocity was used to calculate mean ocean current in the RS by subtracting wind effect from ice motion vectors. The daily sea ice motion data (NSIDC-0116, version 4.1) for 1979–2018 was obtained from the National Snow and Ice Data Center with a resolution of 25 km (Tschudi et al 2019). The daily surface wind was derived from the mean values of 10 m wind velocity at 00, 06, 12 and 18 (UTC) in ERA5 reanalysis, which is produced by the European Centre for Medium-Range Weather Forecasts on a high resolution 0.25° × 0.25° (Copernicus Climate Change Service 2017). After mapping the daily ice motion vectors linearly onto the same grid as for the surface wind data, we used a linear relation rule described by Thorndike and Colony (1982) to relate ice motion ($u_{ice}$, $v_{ice}$), surface wind velocity ($U_{10}$, $V_{10}$) for the same day, speed reduction factor $F$, turning angle $\theta$, and mean ocean current ($\bar{u}_{oc}$, $\bar{v}_{oc}$).

$$\begin{bmatrix} u_{ice} \\ v_{ice} \end{bmatrix} = F \begin{bmatrix} \cos \theta & -\sin \theta \\ \sin \theta & \cos \theta \end{bmatrix} \begin{bmatrix} U_{10} \\ V_{10} \end{bmatrix} + \begin{bmatrix} \bar{u}_{oc} \\ \bar{v}_{oc} \end{bmatrix}.$$

Following the forms shown by Kimura and Wakatsuchi (2000) and Kimura (2004), the speed reduction factor $F$, and turning angle $\theta$, were calculated by using a least squares technique as:

$$\theta = \arctan \left( \frac{\sum U_{10}v_{ice} - \sum V_{10}u_{ice}}{\sum U_{10}u_{ice} + \sum V_{10}v_{ice}} \right),$$

$$F = c_1 + c_2 - c_3 + c_4 \frac{\sum U_{10} + \sum V_{10}}{210},$$

where

$$c_1 = \cos \theta \sum U_{10}u_{ice},$$
$$c_2 = \sin \theta \sum V_{10}u_{ice},$$
$$c_3 = \sin \theta \sum U_{10}v_{ice},$$
$$c_4 = \cos \theta \sum V_{10}v_{ice},$$

and $U_{10}$, $V_{10}$, $u_{ice}$ and $v_{ice}$ are yearly mean values of $U_{10}$, $V_{10}$, $u_{ice}$ and $v_{ice}$. $U_{10}$, $V_{10}$, $u_{ice}$ and $v_{ice}$ are ($U_{10} - \bar{U}_{10}$), ($V_{10} - \bar{V}_{10}$), ($u_{ice} - \bar{u}_{ice}$) and ($v_{ice} - \bar{v}_{ice}$), respectively. Generally, thick ice and high ice concentration in the Southern Ocean cause a large ice stress gradient and lower the value of $F$, hence suppress the ice motion (Steele et al 1997, Kimura 2004). With the yearly mean ice motion, surface wind, calculated $F$ and $\theta$, the yearly mean ocean current ($\bar{u}_{oc}$, $\bar{v}_{oc}$) can be calculated as:

$$\begin{bmatrix} \bar{u}_{oc} \\ \bar{v}_{oc} \end{bmatrix} = \begin{bmatrix} \bar{u}_{ice} \\ \bar{v}_{ice} \end{bmatrix} - F \begin{bmatrix} \cos \theta & -\sin \theta \\ \sin \theta & \cos \theta \end{bmatrix} \begin{bmatrix} U_{10} \\ V_{10} \end{bmatrix}.$$

In this study, only sea ice drift and surface wind data in April–October were used due to rare sea ice coverage on the shelf in other months. The yearly mean ocean current, ($\bar{u}_{oc}$, $\bar{v}_{oc}$), in each year were derived by combining daily ice motion vectors and surface wind velocity in April–October in the corresponding year via the formula above, and time variation of $F$, $\theta$ and ($\bar{u}_{oc}$, $\bar{v}_{oc}$) in each year were assumed to be negligible in this formula (Kimura and Wakatsuchi 2000, Kimura 2004).

The existence of landfast sea ice may lead to large bias in mean ocean current calculation. Nishashi and Ohshima (2015) report that there is little landfast ice along the Ross Ice Shelf calving front due to the occurrence of RSP in winter seasons. Most landfast sea ice occurs west of the RS, but only occupies a narrow region fairly close to the coast. Meanwhile, sea ice motion data close to the Antarctica coast were excluded (figure S3), eliminating the effect of landfast ice on the mean ocean current calculation. Kimura (2004) report that the ocean current derived
using this method in the Southern Ocean is fairly reasonable compared with previous model results, convincing us the reliability of this method used to calculate ocean current. The derived mean ocean current averaged over 1979–2018 in the RS also agrees well with the results presented in Kimura (2004) (figure S3). Based on the yearly mean ocean current in the RS region, the westward zonal flow along the coastal current from the upstream AS to the RS was also examined. This current is critical to the inflow of freshwater from the AS into the RS, which is considered among the main contributors to the long-term freshening in the RS (Jacobs and Giulivi 2010). The calculated surface ocean current was used to estimate barotropic component of zonal flow changes assuming no vertical shear. This barotropic assumption likely overestimates the net transport and variability of the zonal flow because the baroclinic structure is omitted due to the lack of subsurface measurements beneath the ice. However, the barotropic mode is reported to dominate the transport variability of the coastal current in the RS (Dotto et al 2018), thus variability of the calculated ocean current here can characterize the actual changes of the current on the shelf to a large extent. Hence the zonal flow was estimated as the mean zonal velocity of the derived ocean current across a section at the AS/RS gate extending from a point (77° S) on the shelf northward to 76° S along 160° W (figure 1(a)). The derived ocean current close to the coast was excluded when doing the zonal flow analysis due to the uncertainty caused by landfast ice.

2.3. Investigation of drivers of circulation changes
The maximum covariance analysis (MCA) (Bretherton et al 1992, Wallace et al 1992) was used to capture the dominant coupled modes between the yearly mean SLP and the circulation field derived from ice motion vectors and surface wind velocity in the RS during 1979–2018. As the derived ocean current represents mean state of the circulation in a year and the seasonal dependence is not considered (Kimura 2004), the yearly mean SLP was obtained by averaging the SLP data over all months in a year. Then the MCA was achieved by using singular value decomposition of the temporal covariance matrix. The resulting pairs of singular vectors represent the spatial pattern of each field, and the associated temporal coefficients depict the temporal variation in the spatial patterns. Yearly mean SLP was calculated from the ERA5 (Copernicus Climate Change Service 2017), the same reanalysis data as for the daily wind velocity. Then the yearly mean SLP field was analyzed via MCA with the derived yearly mean ocean current. The means and linear trends of both the SLP and ocean current were subtracted before the MCA was performed. The yearly ASL indices (central pressure and location of the ASL) (Hosking et al 2016) were also used to evaluate the potential drivers modulating the large-scale circulation and coastal currents on the shelf, which are found to be responsible for salinity changes in the TNBP (Jacobs and Giulivi 2010, Budillon et al 2011).

3. Results and discussions

3.1. Multidecadal variability of DSW salinity
Figure 1(b) shows the annual mean profiles of salinity and temperature beneath 200 m collected in the TNBP over the past four decades. The profiles suggest an overall decrease in DSW salinity since the 1980s, with a relatively stable temperature that remains at the surface freezing temperature of −1.9 °C, except for the cold Ice Shelf Water (colder than surface freezing temperature) signal observed in some individual years at 200–500 m. Figure 2(a) shows a time series of the vertical mean salinity of deep layers below 600 m depth in TNBP, represented as a change in the densest DSW in the western RS. The substantial decreasing trend of DSW salinity of deep layers before 2013 is estimated to be 0.052 ± 0.026 decade$^{-1}$ (significant at 99% confidence level), freshening at a higher rate than that estimated by Jacobs et al (2002) and Jacobs and Giulivi (2010). However, notably, the freshening trend appears to reverse beginning in about 2013. Although the absolute salinities at all depths after 2013 show lower values than those in the 1980s, there is a rapid increase in the salinity after a minimum in 2013 (figure 1(b)), indicating a recent rebound of DSW salinity in the western RS. To emphasize the multidecadal variability of DSW salinity, we focused on the detrended time series of salinity and its potential drivers, which emphasize the shorter term variability (figure 2).

3.2. Westward zonal flow changes
The rapid enhancement of DSW salinity in recent years is also reported by Castagno et al (2019). After evaluating the potential contributors to the long-term variability of DSW salinity (section 1), Castagno et al (2019) ascribe the rapid increase of DSW salinity in recent years to large-scale forcing which determines the freshwater input from upstream regions to the RS. However, the exact linkage between large-scale circulation modulated by atmospheric forcing and DSW salinity changes are not resolved because the ocean current is difficult to estimate in Antarctic regions. Here, based on yearly mean ocean current derived from ice motion vectors and surface wind velocity, we extended the studies in Castagno et al (2019) and illustrated how the ASL modulates the DSW salinity by affecting large-scale circulation that influences the freshwater input to the RS.

The zonal mean velocity across the AS/RS gate section derived from yearly mean ocean current is presented in figure 2(b). This flow is critical to the spreading of freshwater from the AS westward to the continental shelf of the RS (Nakayama et al 2014). The mean zonal flow across the section (figure 1(a))
is $1.47 \pm 1.01 \text{ cm s}^{-1}$, and exhibits a significant multidecadal variability (figure 2(b)). Time series of the zonal flow anomaly evolve with time similar to the DSW salinity anomaly shown in figure 2(a). The salinity anomaly and their regression onto the zonal flow anomaly 4 years prior are closely correlated at 0.72 ($p < 0.01$). As all the salinity profiles were sampled in summer months (table S1) and deep DSW near the sea bed in TNBP is renewed every year in the austral winter due to strong polynya activity (Rusciano et al 2013), DSW salinity from these profiles is expected to represent that of previous year to a large extent. By contrast, the ocean currents are derived as yearly mean values from sea ice motion and surface wind data in April–October. This means that freshwater from the AS flowing across the gate section travels about 3 years before reaching the sampling locations in TNBP, and contributes to local salinity changes. The mean absolute velocity along the coast from the AS/RS gate to the TNBP is estimated to be $1.21 \pm 0.24 \text{ cm s}^{-1}$, enabling the water parcel to travel on average $1145 \pm 227$ km in 3 years after leaving the AS/RS gate, which roughly agrees with the distance from the gate to the TNBP along the coast ($\sim 1200$ km). For instance, salinity anomaly regressed onto the zonal flow anomaly 4 years prior increased by 0.039 in 2018 compared to that in 2013, comprised 46% of the salinity anomaly differences between 2018 and 2013. Hence, the multidecadal variability of DSW salinity is strongly modulated by the strength of the westward coastal current, which determines the volume of the freshwater input from the AS to shelf regions of the RS. By this mechanism, the weakened westward coastal current after 2010 (figure 2(b)) reduces freshwater input to the RS, and is responsible for the rapid enhancement of DSW salinity in recent years (figure 2(a)).

### 3.3. Evolution of large-scale circulation

Multidecadal variability of DSW salinity in the western RS is closely linked to the westward zonal flow along the coastal current (figure 2), which determines the freshwater input from the upstream AS to the RS. The westward coastal current is associated with large-scale circulation in the Ross Gyre (Dotto et al 2018) and here the yearly mean ocean current derived from ice motion vectors and surface wind velocity was used to assess long-term variability of large-scale circulation. Variability of large-scale circulation is associated with SLP (Dotto et al 2018), which is

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**Figure 2.** Time series of salinity in TNBP and zonal flow across the section at the AS/RS gate. (a) Annual mean salinity blow 600 m depth. The red line represents detrended salinity (left scale), the blue line represents the regression of detrended salinity onto the detrended zonal flow at the AS/RS gate section 4 years prior (left scale), and the orange line represents the actual salinity (right scale). Correlation coefficient and $p$ value between the detrended salinity and its regressed values are also depicted in (a). (b) Three-year running mean of detrended (blue, left scale) and actual (orange, right scale) zonal flow at the AS/RS gate. Closed circles in (b) represent corresponding data 4 years prior to the salinity records. Vertical standard deviation of the annual mean salinity profiles below 600 m depth in TNBP is overlaid as error bar on detrended salinity in (a). The error bar in (b) is the standard deviation among all grid cells along the gate section, estimated on the derived yearly mean ocean current. Standard deviation of the regressed detrended salinity induced by the ocean current is also overlaid (blue error bar in (a)).
Figure 3. Spatial pattern (a) and associated temporal coefficients (normalized by one standard deviation) (b) of the first MCA mode between the detrended mean ocean current and SLP data over the RS for 1979–2018. The 1000 m isobath is presented by the gray line in (a), and the amplitudes of spatial patterns are scaled by one standard deviation of the corresponding temporal coefficients. The correlation coefficient and p value between the temporal coefficients of the first MCA mode are shown in (b), along with the ASL index, by a blue line normalized by one standard deviation. The detrended ocean current was analyzed only at grids with data available for all studied years. The light shading in (b) indicates different phases dominated by negative/positive values for the normalized temporal coefficients of ocean current mode (orange line).

reported to be modulated to a large extent by the ASL in West Antarctica (Turner et al. 2013). Here, we applied MCA between the derived mean ocean current and the SLP data over the most recent four decades, 1979–2018, to examine the atmospheric forcing of the multidecadal variability of the circulation and its effects on westward transport of freshwater along coastal current. For the two fields used here, the leading modes of the MCA patterns represent the patterns of SLP anomalies in the RS that most strongly influence the ocean current anomalies and the resulting freshwater transport of the coastal current.

The first leading MCA mode of the coupled ocean current anomaly and SLP anomaly field (figure 3(a)) accounts for 78% of the total squared covariance. The MCA1 patterns display an increased SLP and a westward ocean current anomaly in the ASL domain. The spatial pattern of the ocean current anomaly closely resembles that of the SLP, changing along the isolines of SLP anomalies. The associated temporal coefficients of the first MCA mode between the detrended ocean current and SLP field are significantly correlated ($r = 0.53$ with $p < 0.01$, figure 3(b)), indicating the dominant modulation of the SLP changes in horizontal transport. Figure 2 shows that the multidecadal variability of DSW salinity is modulated by zonal flow variability from the AS with 4 years lag via controlling freshwater input to the RS. Similar to the temporal evolution of the zonal flow at the AS/RS gate, the associated time series of ocean current patterns in figure 3(b) exhibit a dominant negative phase before 1987 (light red shading), followed by a dominant positive phase until 2011 (light blue shading) and a return to the negative phase beginning in 2012 (light green shading). The significantly coupled temporal coefficients obtained from MCA are closely related to the ASL index ($r = 0.93$, $p < 0.01$ and $r = 0.51$, $p < 0.01$ for SLP and ocean current anomaly, respectively) (figure 3(b)). This suggests that the ASL dominates the long-term variability of the freshwater transport along the coastal current.
input from the AS to the RS by affecting the zonal wind at the southern edge of the ASL domain, as reported by Raphael et al. (2019), which is linked to the barotropic transport of the coastal current.

To study the effect of the variability in the ASL on the large-scale circulation changes and the resulting freshwater input along the coastal current from the AS into the RS, we examined the relationship between the SLP and mean ocean current (figure 4) during each period shown in figure 3(b). These years (1979–2018) are crudely split into three phases according to whether the temporal coefficient of the ocean current mode are dominated by positive or negative values (figure 3(b)); notably, the observed DSW salinity in TNBP develops in a similar temporal structure with 4 years lag (figure 2(a)) and can also be classified in the same phases according to the positive/negative salinity anomalies. The mean ocean circulation shows a cyclonic circulation at the surface in all phases (figures 4(a), (c) and (e)), representing the main current system in the RS, the Ross Gyre, which controls the cross shelf exchanges and westward transport from the upstream AS in the gyre’s southern flank (Dotto et al. 2018). Because zonal wind is found to be closely associated with changes of strength and location in the ASL (Paolo et al. 2018, Raphael et al. 2019), the wind-driven circulation in the RS is also expected to be affected by the ASL. Distinct spatial patterns of surface circulation are detected as a response to the SLP distribution driven by the ASL changes for different phases. In Phase I (1979–1986) and Phase III (2012–2018), although the center of the ASL displaces toward opposite directions, the region of low pressure extends consistently toward southwest relative to that in Phase II (red and black contour in figure 4, respectively). This increases the zonal gradient of the SLP between the Antarctica continental high and the ASL in Phase I and Phase III, driving the meridional flow at surface to become more dominant on the eastern RS shelf (figures 4(b) and (f)). The decreased SLP is much more significant in Phase III (figure 4(e)), indicating an accelerated deepening of

![Figure 4](https://example.com/figure4.png)

**Figure 4.** Mean (left panels) and detrended (right panels) patterns of SLP (shading, hPa) and ocean current (vectors) in three phases. (a) and (b), Phase I (averaged over 1979–1986); (c) and (d), Phase II (averaged over 1987–2011); (e) and (f), Phase III (averaged over 2012–2018). Detrended ocean current is calculated only at grids with data available for all studied years. The 1000 m isobath is presented as a gray line. Red, yellow and black contour lines and triangles designate 981 hPa and mean locations of the ASL center in phase I, II, III, respectively.
the ASL after 2011 (figure 3(b)). This induces a strong eastward coastal current anomaly around the Antarctica (figure 4(f)), reduces the freshwater input to the RS from upstream regions, and results in a rapid increase of DSW salinity in recent years (figure 2(a)). In contrast, the slightly northeastward shift of the ASL (yellow contour in figure (4)) and positive SLP anomalies during Phase II (1987–2011, figures 4(c) and (d)) drive a westward anomaly of the coastal current, enhance the westward transport of freshwater from the AS and lead to the fresening trend before 2013. This suggests that the large-scale circulation changes are strongly modulated by the ASL, and the corresponding changes of westward transport of freshwater along the coastal current are responsible for the multi-decadal variability in DSW salinity.

4. Conclusions

The production of DSW on the continental shelf of the RS contributes ~25% of the AABW to the global ocean (Orsi et al 2002). Hence, changes in DSW properties are expected to have global-scale impacts on heat exchanges through thermohaline circulation (Orsi and Wiederwohl 2009). Marked freshening of DSW has been detected and ascribed to increasing freshwater input resulting from the rapid melting of ice shelves in the upstream AS (Jacobs and Giulivi 2010, Budillon et al 2011). Using long-term records of observational temperature and salinity in the TNBP, we confirm recent studies showing a significant multi-decadal variability of DSW salinity in the western RS. The freshening trend of DSW salinity before 2013 is estimated to be 0.052 ± 0.026 decade$^{-1}$ (significant at 99% confidence level), and then is followed by an abrupt salinity enhancement thereafter. Earlier studies suggest that large-scale circulation changes and the resulting variability of freshwater input along coastal current from the AS to the RS are responsible for long-term changes of DSW salinity (Jacobs and Giulivi 2010, Castagno et al 2019). Based on the yearly mean ocean current derived from daily ice motion vectors and surface wind velocity, we find DSW salinity anomaly in TNBP is significantly correlated ($r = 0.72, p < 0.01$) with the zonal flow anomaly at the AS/RS gate with 4 years lag. With the mean absolute velocity along the coast from the AS/RS gate to the TNBP estimated to be 1.21 ± 0.24 cm s$^{-1}$, the low-salinity water parcels from upstream regions travel on average 1145 ± 227 km in 3 years after leaving the AS/RS gate along the coastal current, influencing DSW salinity of TNBP in summer months of the forth year.

The MCA results show that surface circulation controlling the westward transport on the southern flank of the Ross Gyre, i.e. the coastal current, is significantly coupled with SLP changes ($r = 0.53 \text{ with } p < 0.01$), and both principal components are closely linked to the temporal evolution of the ASL. Spatial pattern variability of the circulation is largely modulated by the strength and shift of the ASL. In 1979–1986 and 2012–2018, the deepening of the ASL extends the region of low pressure toward southwest. This increases the zonal gradient of the SLP in the RS and makes the meridional flow become more dominant, resulting in an eastward coastal current anomaly on the shelf and a reduction in the freshwater input to the RS along the coastal current. By contrast, in 1987–2011, the northeastward shift of the ASL induces a westward coastal current anomaly, increases freshwater inflow to the RS and explains the long-term freshening before 2013. Given the connection constructed between multi-decadal variability of the zonal flow and DSW salinity, it is reasonable to conclude that the deepening and southwestward extension of the ASL and the subsequent response of large-scale circulation determine the coastal current changes in upstream regions and the freshwater input to the RS, and are responsible for the multi-decadal variability of DSW salinity in the western RS.

Data availability

Temperature and salinity records from the WOD13 datasets are available at https://www.nodc.noaa.gov/OC5/WOD13/, and records in and after 2013 are held by the Chinese National Arctic and Antarctic Data Center (http://www.chinare.org.cn). The ERA5 hourly data on single levels of SLP and wind velocity at 10 m are from https://doi.org/10.24381/cds.adbb2d47. The daily ice motion vectors are available at https://doi.org/10.5067/TNAWUWO7QH7B. The ASL index is obtained from https://legacy.bas.ac.uk/data/a/absl/.

The data that support the findings of this study are available upon reasonable request from the authors.

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ORCID iDs

Guijun Guo  https://orcid.org/0000-0002-1437-1561
Libao Gao  https://orcid.org/0000-0002-0402-9340
Jiu Xin Shi  https://orcid.org/0000-0002-5825-8894

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