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Freshening of Antarctic Intermediate Water in the South Atlantic Ocean in 2005–2014

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Abstract. Basin-scale freshening of Antarctic Intermediate Water (AAIW) is reported to have occurred in the South Atlantic Ocean during the period from 2005 to 2014, as shown by the gridded monthly means of the Array for Real-time Geostrophic Oceanography (Argo) data. This phenomenon was also revealed by two repeated transects along a section at 30° S, performed during the World Ocean Circulation Experiment Hydrographic Program. Freshening of the AAIW was compensated for by a salinity increase of thermocline water, indicating a hydrological cycle intensification. This was supported by the precipitation-minus-evaporation change in the Southern Hemisphere from 2000 to 2014. Freshwater input from atmosphere to ocean surface increased in the subpolar high-precipitation region and vice versa in the subtropical high-evaporation region. Against the background of hydrological cycle changes, a decrease in the transport of Agulhas Leakage (AL), which was revealed by the simulated velocity field, was proposed to be a contributor to the associated freshening of AAIW. Further calculation showed that such a decrease could account for approximately 53 % of the observed freshening (mean salinity reduction of about 0.012 over the AAIW layer). The estimated variability of AL was inferred from a weakening of wind stress over the South Indian Ocean since the beginning of the 2000s, which would facilitate freshwater input from the source region. The mechanical analysis of wind data here was qualitative, but it is contended that this study would be helpful to validate and test predictably coupled sea–air model simulations.

1 Introduction

Thermocline and intermediate waters play an important part in global overturning circulation by ventilating the subtropical gyres in different parts of the world oceans (Sloyan and Rintoul, 2001). They also constitute the northern limb of the Southern Hemisphere supergyre (Ridgway and Dunn, 2007; Speich et al., 2002).

Previous studies have addressed the variability of intermediate water. Wong et al. (2001) found that the intermediate water had freshened between the 1960s and the period 1985–1994 in the Pacific Ocean. Bindoff and McDougall (2000) reported that there had been freshening of water between 500 and 1500 db from 1962 to 1987 along 32° S in the Indian Ocean. Curry et al. (2003) showed a salinity reduction on the isopycnal surface of intermediate water for the period from the 1950s to the 1990s in the western Atlantic. The freshening variability can be traced back to the signature of water in the formation regions (Church et al., 1991). The freshening examples given above are in agreement with the enhancement of the hydrological cycle, in which the wet (precipitation (P) > evaporation (E), P dominance) subpolar regions have been getting wetter and vice versa for the dry (E dominance) subtropical regions over the last 50 years (Held and Soden, 2006; Skliris et al., 2014).

Antarctic Intermediate Water (AAIW) is characterized by a salinity minimum (core of AAIW) centered at the depths of 600 and 1000 m (Fig. 1), which lies within the potential density (with reference to sea surface) range of $\sigma_0 = 27.1$–$27.3\,\mathrm{kg\,m^{-3}}$ (Piola and Georgi, 1982). The AAIW is found from just north of the Subantarctic Front (SAF; Orsi et al., 1995) in the Southern Ocean and can be traced as far as 20° N

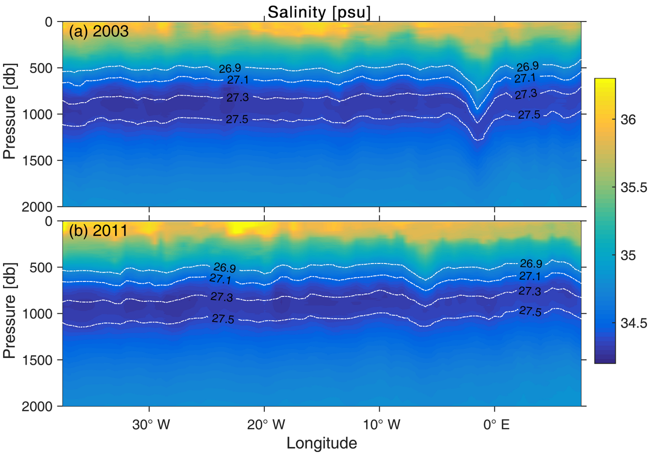

Figure 1. WOCE salinity sections along 30° S in the South Atlantic Ocean (positions shown in Fig. 2) observed in **(a)** 2003 and **(b)** 2011. Overlaid white solid–dotted lines are γ^n surfaces ranging from 26.9 to 27.5 kg m^{-3}, with a 0.2 kg m^{-3} interval.

(Talley, 1996). It is generally accepted that the variability of AAIW is largely controlled by air–sea–ice interaction (Close et al., 2013; Naveira Garabato et al., 2009; Santoso and England, 2004), but the argument about its origin and formation process continues. For example, there is the circumpolar formation theory of AAIW along the SAF, through mixing with Antarctic Surface Water (AASW) along isopycnals (Fetter et al., 2010; Sverdrup et al., 1942). Alternatively, it has been proposed that there is a local formation of AAIW in specific regions, as a by-product of Subantarctic Mode Water (SAMW) relating to deep convection (McCartney, 1982; Piola and Georgi, 1982). The first standpoint states that the AAIW is primarily derived from entirely subpolar sources; meanwhile the second one emphasizes the role that air–sea interaction plays in the oceans south of South America.

In the South Atlantic, AAIW constitutes the return branch of the Meridional Overturning Circulation (MOC) (Donners and Drijfhout, 2004; Speich et al., 2007; Talley, 2013). As an open-ocean basin, the South Atlantic is fed by two different sources of AAIW (Sun and Watts, 2002). The first is younger, fresher and has a lower apparent oxygen utilization (AOU) and originates from the Southeast Pacific (McCartney, 1977; Talley, 1996) and the winter waters west of the Antarctic Peninsula (Naveira Garabato et al., 2009; Santoso and England, 2004). These source regions of AAIW are mostly dominated by the net surface freshwater flux from atmosphere to ocean ($P > E$), which facilitates the freshening of AAIW with time. The second is the older, saltier and higher AOU AAIW which comes from the Indian Ocean, transported by the Agulhas Leakage (AL) as Agulhas rings (Fig. 2). The mixture of the above two types of AAIW can lead to a transition of hydrographic properties across the subtropical South Atlantic (Boebel et al., 1997).

The influence of AL on the variability of AAIW in the South Atlantic has been demonstrated to be substantial (Hummels et al., 2015; Schmidtko and Johnson, 2012), as 50–60 % of the Atlantic AAIW originates from the Indian Ocean (Gordon et al., 1992; McCarthy et al., 2012), with increased (decreased) transport of AL relating to salinification (freshening) of AAIW. AL has apparently increased during the period from the 1950s to the early 2000s (Durgadoo et al., 2013; Lübbecke et al., 2015), but there have been no studies addressing the influence of AL on the AAIW in South Atlantic since 2000.

With the instigation of the Array for Real-time Geostrophic Oceanography (Argo) program, in situ hydrographic observation has tremendously expanded since 2003 (Roemmich et al., 2015), particularly in the Southern Ocean (SO) where historical data are sparse and intermittent. This decreases the uncertainty of estimates for the research on both seasonal and decadal variations of subsurface and intermediate waters.

The present work reports the freshening of AAIW in the South Atlantic over the preceding decade (2005–2014) using gridded monthly data based on Argo data. Against the background of an enhanced hydrological cycle, decreased transport of AL contributed to such freshening and may be driven by a weakening of wind stress in the South Indian Ocean during the same period.

2 Data and methods

Based on individual temperature (T) and salinity (S) profiles from Argo, International Pacific Research Centre (IPRC) gridded monthly-mean data for the period 2005–2014 have been produced using variational interpolation. The IPRC data have 27 levels from 0 to 2000 m depth vertically, on a nominal 1° × 1° grid globally and at monthly temporal resolution (http://apdrc.soest.hawaii.edu/projects/Argo/data/gridded/On_standard_levels/index-1.html). To reduce the error from low vertical resolution of data when computing the hydrographic values on isopycnal surfaces, T and S profiles were first interpolated onto 1 m vertical depth intervals using a spline method in the intermediate water depth, and a linear method in the thermocline depth. Because the IPRC data were interpolated from randomly distributed Argo profiles, it is necessary to demonstrate the robust nature of their signals by comparing them with the other Argo gridded products. As a result, the Japan Agency of Marine-Earth Science and Technology (JAMSTEC, Hosoda et al., 2008) T and S data from 2005 to 2014, with 1° longitude and 1° latitude resolution, were also collected for comparison and verification. The number of Argo profiles is rapidly increasing year by year, and part of their distribution has been outlined in previous studies, inter alia Hosoda et al. (2008) and Roemmich et al. (2015).

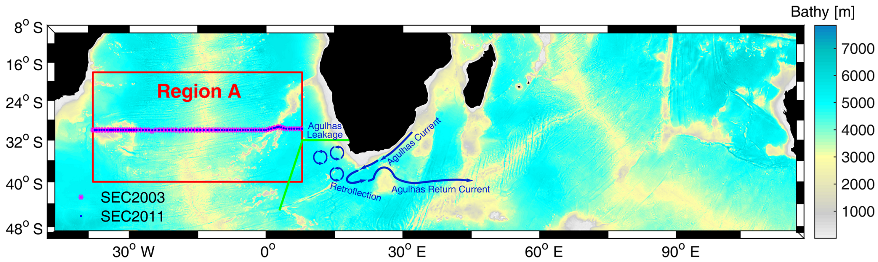

Figure 2. Bathymetry of the South Indian–Atlantic oceans. Color shading is ocean depth. Red box delineates the area for the basin-wide average of gridded data (hereafter referred to as Region A). The green line shows the Good Hope section, which is used to calculate the leakage transport to the South Atlantic. Magenta stars represent transatlantic CTD stations measured in 2003, with blue dots showing the 2011 measurements. The Agulhas Current, Retroflection, Agulhas Return Current and Agulhas Leakage (as eddies) are also shown.

Two hydrographic cruises of repeated transects along 30° S were conducted during the World Ocean Circulation Experiment (WOCE) Hydrographic Program (http://www.nodc.noaa.gov/woce/wdiu/diu_summaries/whp/index.htm). Their locations are presented in Fig. 2. The first transect consisted of 72 stations in 2003 by the R/V *Mirai* (Japan, Kawano et al., 2004); the second was in 2011 with 81 stations sampled from the *Ronald H. Brown* (United States, Feely et al., 2011). These two transects not only occupied almost identical station positions in the subtropical South Atlantic, but were also conducted in the same season (November and October respectively). Furthermore, the time interval between the two sections from November 2003 to October 2011 is very similar to the period covered by the IPRC data (January 2005–December 2014) and can therefore be used to validate those results.

To smooth out some of the higher frequency variability (i.e. mesoscale eddies and internal waves), the investigation of halocline variation should be along neutral density surfaces (McCarthy et al., 2011; McDougall, 1987). The layer of AAIW is defined using neutral density (γ^n, unit: $\mathrm{kg\,m^{-3}}$; Jackett and McDougall, 1997) instead of potential density, with the upper and lower boundaries being 27.1 γ^n and 27.6 γ^n (Goes et al., 2014), respectively.

Monthly 10 m wind fields between years 1980 and 2014 from the ERA-Interim archive at the European Centre for Medium Range Weather Forecasts (ECMWF) (http://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/) were used to investigate the decadal variability of wind stress (WS) over the South Indian Ocean. Another reanalysis wind product of National Centers for Environmental Prediction Department of Energy Atmospheric Model Intercomparison Project reanalysis 2 (NCEP-2, http://www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis2.html) was also used for the period 1980–2014. Additionally, the satellite-derived wind products of the Quick Scatterometer (QuikSCAT) for 2000–2007 and the Advanced Scatterometer (ASCAT) for 2008–2014 (both in ftp://ftp.ifremer.fr/ifremer/cersat/products/gridded/MWF/L3/) were used to compare and verify the decadal variability of WS revealed by the ERA-Interim wind product. In this work, the WS over open ocean was calculated from 10 m wind field data using the equation adopted in Trenberth et al. (1989).

Reanalysis data including precipitation (P) and evaporation (E) from the ERA-Interim were used to reveal the freshwater input from the atmosphere to the ocean surface in the preceding decade.

The Simple Ocean Data Assimilation version 3.3.1 (SODA3.3.1, http://www.atmos.umd.edu/~ocean/), which is forced by the Modern-Era Retrospective Analysis for Research and Applications Version 2 (MERRA-2), spans the 36-year period 1980–2015 (Carton et al., 2017). The global simulated velocity field at specified depths provided by SODA makes it possible to evaluate the transport of AL.

3 Freshening of Antarctic Intermediate Water

3.1 Freshening observed from Argo gridded products

The Argo gridded products provide a globally distributed and continuous time series of T and S profiles down to 2000 m ocean depth. The present work focused on the AAIW in the South Atlantic Basin (Fig. 2, Region A), which encompasses most of the subtropical gyre and a part of the tropical regimes (Boebel et al., 1997; Talley, 1996). Computed from the Argo gridded data of IPRC, the biennial mean $\theta-S$ diagram (Fig. 3a) clearly shows that the AAIW has experienced a process of progressive basin-scale freshening during the period from January 2005 to December 2014. The linear trend of salinity (Fig. 3b) further reveals that the freshening takes up most of the AAIW layer but with a little salinification in the deeper part. Except around the 27.42 γ^n neutral density surface, the AAIW variation is significant at the 95 % confidence level, using the F-test criteria. In comparison with Fig. 3a, it was found that the cutoff point of transformation from salinity decrease to increase is near the salinity minimum. Above the salinity minimum, the shift of $\theta-S$ trends towards cooler and fresher values along density surfaces and seems to be a response to the warming and fresh-

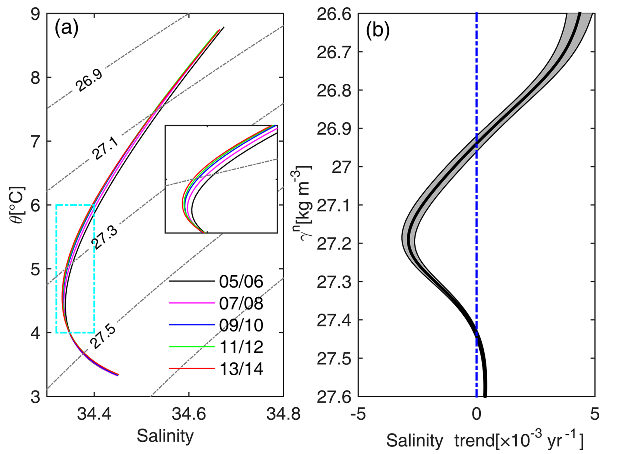

Figure 3. (a) Biennial mean θ–S diagram averaged over Region A for IPRC data with γ^n surfaces superimposed (grey solid–dotted lines). The inserted figure is the magnification of the area delineated by cyan solid–dotted box. The corresponding time for each θ–S curve is listed in their bottom-right corner (i.e. 05/06 for 2005–2006). **(b)** Salinity trend along γ^n surfaces for period January 2005–December 2014 is displayed by the thick black line, and the 95 % confidence intervals (F-test) are represented by the light grey shadings, calculated from IPRC data.

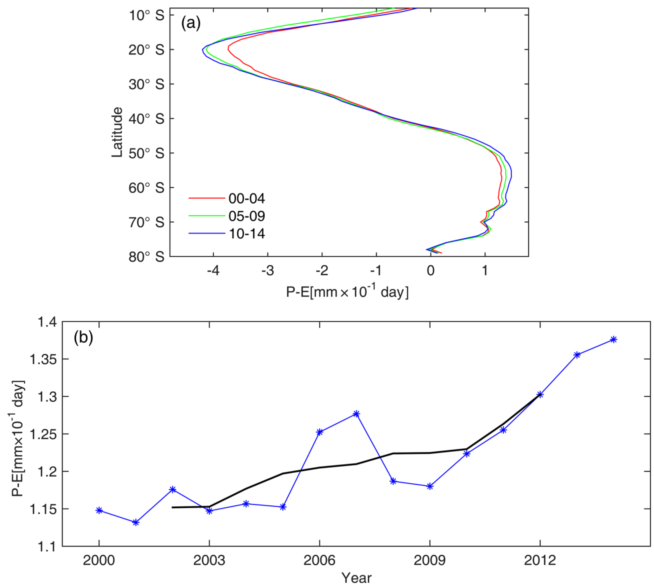

Figure 4. Calculated from ERA-Interim precipitation and evaporation data: **(a)** zonal mean (ocean areas only) of annual $P-E$ (freshwater input, mm day^{-1}); each line represents a 5-year averaged result. The corresponding time period (i.e. 00–04 for 2000–2004) is listed in the bottom-left corner. **(b)** Time series of annually $P-E$, averaged over the oceans in 45–65° S, 0–360° E band from 2000 to 2014 (blue star), and its 5-year running mean (black).

ening of surface waters where AAIW ventilates. Such thermohaline change has also been found in the Pacific and Indian oceans over a different time period (Wong et al., 1999). Church et al. (1991) and Bindoff and McDougall (1994) have researched the counterintuitive cooling of AAIW temperature induced by warming of surface water. They showed that a warming parcel in the mixed layer would subduct further equatorward, which would lead the $\theta-S$ curve to become cooler and fresher at a given density. The salinity decrease of the AAIW core indicates that such a change can only be induced by freshwater input from the source region, as mixing with more saline surrounding waters cannot give rise to a salt loss in the salinity minimum layer.

To demonstrate the robustness of AAIW variations revealed by the IPRC data, re-plots of Fig. 3a–b using another Argo gridded product from JAMSTEC are also shown for comparison (see Fig. S1 in the Supplement; only the AAIW layer is shown). Not only was the same variation along density surfaces in the AAIW layer found, but so too was a freshening of the salinity minimum. The isoneutral salinity increases in both IPRC and JAMSTEC data below the salinity minimum are quite small. The main discrepancy between them is that the salinity reduction in the JAMSTEC data is somewhat less than IPRC and at a higher 95 % confidence level (a mean of 0.006 between 27.1 γ^n and 27.6 γ^n).

The freshwater gain for the basin-scale salinity decrease of AAIW (mean salinity difference of 0.012 between 27.1 γ^n and 27.6 γ^n over a mean water mass thickness of 500 m) is estimated at 17 mm yr^{-1} in its source region. This assumes that the South Atlantic only experienced freshwater input and nothing changed, thus the relationship between the salinity in 2005 and 2014 per unit area was roughly $S_{2005} \cdot 500 = S_{2014} \cdot (500 + \Delta d)$. Here $S_{2005} = S_{2014} + 0.012$ and Δd is the freshwater gain during the covered period. However, the depth-integrated salinity change over the water column (between 26.6 γ^n and 27.6 γ^n) was 0.0014, since a salinity increase of thermocline water balances the observed freshening of AAIW. This salinity budget implies contemporary hydrological cycle intensification in the Southern Hemisphere, which is illustrated by the P minus E change from 2000 to 2014, with $P-E$ increasing in the subpolar region and vice versa in the subtropical region (Fig. 4a). In these cases, the thermocline (intermediate) water that ventilates in the high-evaporation (precipitation) subtropical (subpolar) regions gets more saline (freshened), as shown by the hydrographic observations (Fig. 3b).

Against the background of hydrological cycle augmentation, the annual freshwater input in the AAIW ventilation region during the freshening period increased by 0.02 mm day^{-1}, about 17 % of the $P-E$ in 2005 (Fig. 4b). It is considered that the significant $P-E$ increase began around 2003 (Fig. 4b, 5-year running mean line), which means the observed freshened AAIW could be traced back to 2003. Though it was not possible to compute the direct freshwater input to the South Atlantic Basin in this study, the

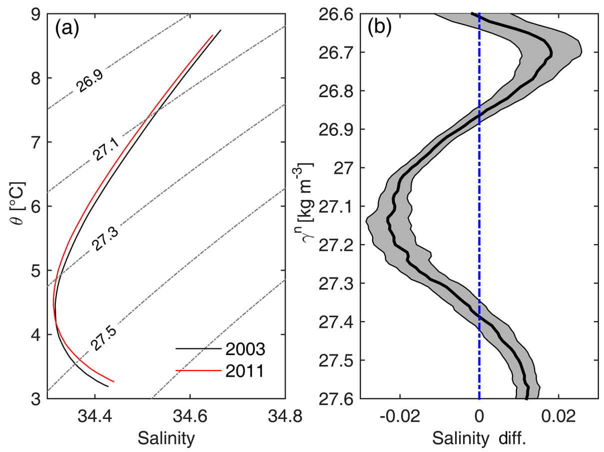

Figure 5. (a) The same as Fig. 3a but for sectional mean of WOCE hydrographic casts. The corresponding year for each θ–S curve is listed in the bottom-right corner. **(b)** Sectional mean differences (thick black line) of WOCE hydrographic data along γ^n and their 95 % confidence intervals (grey shadings, t test).

Argo-era freshening of AAIW is qualitatively consistent with the freshwater gain in its source region.

3.2 Freshening in the quasi-synchronous WOCE CTD observations

Here, two synoptic transatlantic sections from WOCE hydrographic program were used to explore the decadal freshening signal identified in the above subsection. Similar to Fig. 3a, the sectional mean $\theta - S$ diagram (Fig. 5a) displays the same shift of thermohaline values, including freshening of the salinity minimum, salinity reduction in the upper AAIW layer and vice versa in the lower layer. Compared to the $\theta - S$ curves of IPRC data (Fig. 3a), the curves of WOCE (Fig. 5a) seem to be, in general, cooler θ and fresher S. It is suggested that this is because the IPRC mean is weighted towards the warmer and saltier waters in the north.

Unlike the Argo gridded product which has a continuous time series of T and S data, there are only two sections in the WOCE observations. Instead of calculating the linear trend of salinity (as was done with the IPRC data), the difference in salinity observed in 2003 and 2011 was estimated (Fig. 5b). The light grey shading denotes the 95 % confidence interval using simple t test criteria and considering the number of degrees of freedom. Above the salinity minimum, the freshening of AAIW revealed by the IPRC and the WOCE data are quite similar, with the maximum appearing near 27.2 γ^n. Because the last WOCE observation terminated in 2011 and the salinity reduction would continue at least up to 2014, as displayed in Fig. 3a, the magnitude of the freshening in WOCE (Fig. 5b) is smaller than IPRC (Fig. 3b). In the water layer below the salinity minimum (around 27.41 γ^n), the salinity increase shown in the WOCE data is relatively large (Fig. 5b). This is thought to be because the salinity rise reached its maximum around 2011, which is shown in the time series of basin-wide averaged salinity on 27.45 γ^n- and 27.55 γ^n-density surfaces (see Fig. S2).

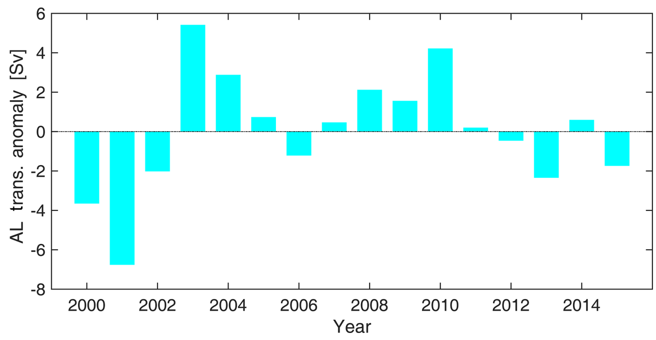

Figure 6. Computation of Agulhas Leakage transport anomaly from the SODA velocity field along the Good Hope line. Note that the depth integration is only for the AAIW layer.

For the salinification of thermocline water, there is a large discrepancy between IPRC and WOCE data on neutral density surfaces 26.6–26.7 γ^n (Fig. 5b). It is considered that this would not affect the salinity budget over the water column (Fig. 5b), given that the salt gain of thermocline water would balance the observed freshened AAIW. In conclusion, the general trend and consistency of the detail therein of the salinity change over the last 10-year time period, revealed by the IPRC and the WOCE data, leads us to state that the freshening of AAIW is a robust and valid finding.

4 Decrease of Agulhas Leakage transport

AAIW in the South Atlantic is largely influenced by the AL through the intermittent pinching off of Agulhas rings (Fig. 2; Beal et al., 2011), transferring salty thermocline and intermediate water from the Indian Ocean to the South Atlantic (De Ruijter et al., 1999). The above discussion suggests that the freshening of AAIW was induced by the input of freshwater from the source regions, which consist of the southeast Pacific Ocean and the circumpolar subpolar oceans (see Sect. 1). As a result, if the transport of more saline water from the Indian Ocean decreased, it would promote the effect of this freshwater increase. In this section, the decrease of AL transport was evaluated by depth integration of the velocity field and further demonstrated by using an indirect indicator.

4.1 Evaluation from SODA velocity

In modeling studies, it is widely accepted to use a Lagrangian approach to quantify the leakage (Biastoch et al., 2009; van Sebille et al., 2009). Here, a simplified strategy was employed to compute the leakage by integrating the velocity within AAIW layer (approximately between 610 and 1150 m, according to Fig. 1), which was shown to result in a similar quantification to the Lagrangian one (Le Bars et al.,

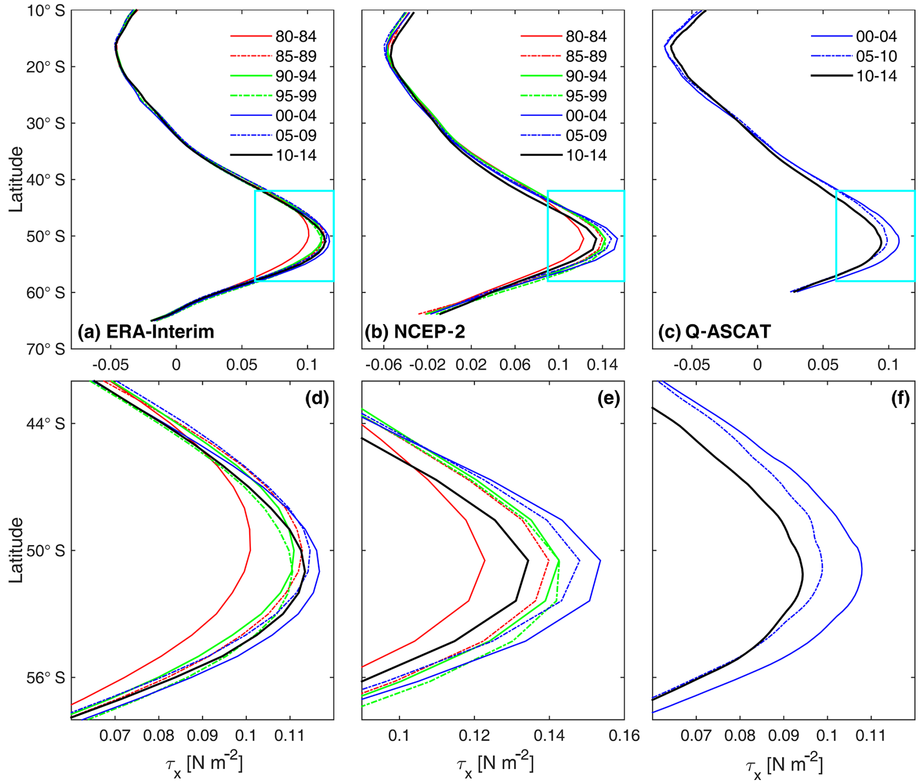

Figure 7. Zonally averaged wind stress calculated from the wind product of **(a)** ERA-Interim, **(b)** NCEP-2 and **(c)** QuikSCAT–ASCAT over the Indian Ocean (20–110° E) for different periods (i.e. 80–84 for January 1980–December 1984; 00–04 for January 2000–December 2004) listed in the top-right corners. Panels **(d, e, f)** are the magnification of cyan boxes in **(a, b, c)**, respectively.

2014). The depth integration is along the Good Hope section (green line in Fig. 2), using the cross-component velocity. Note that the leakage calculation is from the continent to the zero line of the barotropic streamfunction, which is the separation of the Agulhas regime and the Antarctic Circumpolar Current (Biastoch et al., 2015).

Before showing the transport computed from the SODA velocity data, it is necessary to verify that the SODA hydrographic data show the same freshening of AAIW as other datasets. AAIW in the South Atlantic was also found to have freshened during period 2005–2014, though with relatively small magnitude (Fig. S3). Yearly leakage computation within the AAIW layer was carried out for the period 2000–2015 (Fig. 6). It shows that the leakage in the early 2010s is smaller than that in the middle and late 2000s, forming a decreasing trend in a nearly 10-year period. This estimation of leakage seems to be consistent with the indirect estimate of AL transport given below.

The following calculation is to simply estimate the contribution of the AL transport change to our observed freshening. As shown by Fig. 6, the decreased rate of AL transport could be taken to be 2 Sv in a 10-year time period, assuming that this rate increased year by year in the study period (i.e., 0.2 Sv in the first year, 0.4 Sv in the second year, and so on). Following Sun and Watts (2002), here we take the salinity difference of $\Delta S = 0.1$ between the South Indian and the South Atlantic in the AAIW layers. The other parameters, including total seconds in a year, water thickness of the AAIW layer and the area of Region A, are taken to be $\Delta t = 365 \times 24 \times 3600$ s, $\Delta d = 500$ m and $\Delta s_{\mathrm{A}} = 1.09 \times 10^{13}$ m^2, respectively. Therefore, the salinity decrease from 2005 to 2014, induced by the change of AL transport, should be $(0.2 + 0.4 + \ldots + 2) \times 10^6 \times \Delta t \times \Delta S/(\Delta s_{\mathrm{A}} \times \Delta d)$. This results in a salinity reduction of 0.0064, which could account for approximately 53.0 % of the observed freshening revealed by the IPRC data. Though our estimate here is quite rough, we can state that, during 2005–2014, the AL significantly influenced the salinity change in the South Atlantic Ocean within the AAIW layers.

4.2 Weakening of the westerlies in the South Indian Ocean

Continuous measurements of the AL transport have never been realized before. An earlier study suggested that an increased AL transport correlates well with a poleward shift of the westerly winds (Beal et al., 2011). However, after using reanalysis and climate models, Swart and Fyfe (2012) argued that the strengthening of Southern Hemisphere surface westerlies has occurred without major transgressions in its

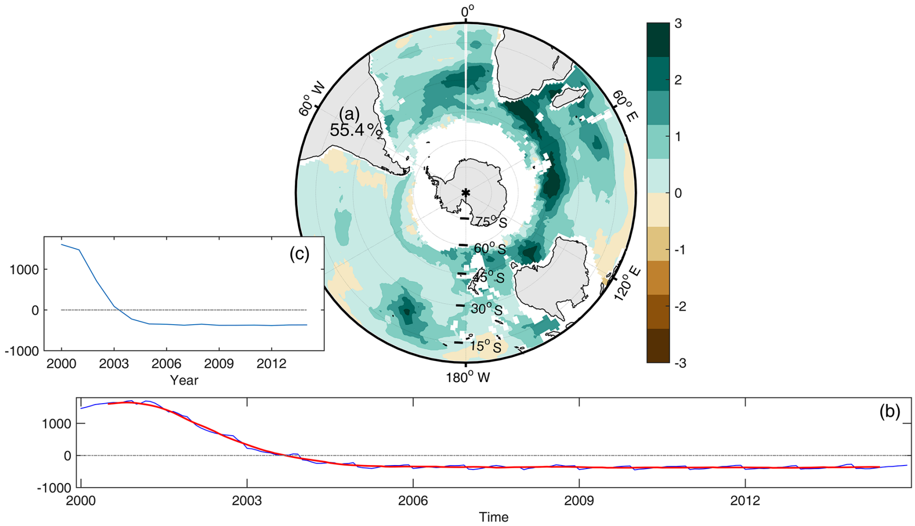

Figure 8. **(a)** Pattern and **(b)** time series (blue: monthly; red: 13-month smoothed) of EOF1 of salinity on 27.36 γ^n surface. **(c)** Yearly mean time series of EOF1. Calculated from SODA data.

latitudinal position over the period 1979–2010, during which period the AL has largely increased (Biastoch et al., 2009). A more recent study from Durgadoo et al. (2013) showed that the increase of AL is concomitant with an equatorward rather than a poleward shift of westerlies in their simulation cases. They also concluded that the intensity of westerlies is predominantly responsible for controlling this Indian–Atlantic transport. Many relevant studies agreed on this relationship, that the enhancement of westerly wind intensity is related to the increase of AL (Goes et al., 2014; Lee et al., 2011; Loveday et al., 2015).

The AL corresponds most significantly to westerly wind strength averaged over the Indian Ocean in contrast to that averaged circumpolarly or locally (Durgadoo et al., 2013). According to the work of Durgadoo et al. (2013), zonally averaged WS was calculated from the ERA-Interim wind product over the Indian Ocean (20–110° E) for every 5-year period since 1980 (Fig. 7a and d). Previous studies (Lee et al., 2011; Loveday et al., 2015) have found that the WS has increased considerably from the 1980s to the beginning of the 2000s (Fig. 7d), consistent with the contemporary increase of AL transport. Though there are oscillations during 1990s, the WS reached its peak around the years 2000–2004 (Fig. 7d), then began to decline. It can be concluded that the WS has weakened for period 2000–2014 (Fig. 7d), which implies a concurrent decrease of AL transport.

In addition to the ERA-Interim wind data, we have further checked the zonally averaged WS over the Indian Ocean (20–110° E), using another reanalysis product of NCEP-2 (Fig. 7b and e) and the combined QuikSCAT–ASCAT (Fig. 7c and f) satellite-derived wind products. The three zonally averaged WS agree that during the period 2000–2014, the westerlies reached a peak in the years 2000–2004, and then progressively subsided through 2005–2009 to 2010–2014. The process of gradual decline of WS is most pronounced in the NCEP-2 data. It is noteworthy that none of the three products show a significant meridional shift of the latitude of maximum WS from 2000 to 2014, in corroboration with the conclusion of Swart and Fyfe (2012).

4.3 Evidence from other works

Many efforts have been made to estimate AL transport, especially using model simulations (Lübbecke et al., 2015; Loveday et al., 2015). In recent years, Le Bars et al. (2014) provided the time series of AL transport over the satellite altimeter era, computed from absolute dynamic topography data, which can show the decadal variation of AL present. In their result (Fig. 8 in Le Bars et al., 2014), the anomalies of AL from satellite altimetry reached a peak around 2003 (annual average), and then began to subside, apart from a mid-2011 increase. In addition, their negative trend of AL (Fig. 9 in Le Bars et al., 2014) over the period from October 1992 to December 2012 indicates that the transport was reduced during the 2000s in contrast to the 1990s. Another study by Biastoch et al. (2015) should be of help in the present discussion. Though the time series of AL obtained from models did not show a distinct decline of AL transport in the last decade, which seems partly due to the data filter applied and the end of the time series (Fig. 4 in Biastoch et al., 2015), it displays a maximum of salt transport around 2000 (Fig. 5 in Biastoch et al., 2015). This peak and the subsequent decline of salt transport are consistent with the freshening of AAIW over the similar time period considered here.

Thus, in addition to the freshwater input that gave rise to the salt loss of the AAIW in the South Atlantic Ocean, reduced transport of AL or salt would further enhance this signal. Unfortunately, the analyses of the contributions from both the source region and the AL were only quantitative. Future work should be focused on the quantification of each factor based on model simulations.

5 Conclusions and discussions

The analysis of IPRC gridded data shows that the AAIW in the South Atlantic has experienced basin-scale freshening for the period from January 2005 to December 2014 (Fig. 3a and b), with freshwater input estimated at 17 mm yr^{-1} in its source region. Two transects of the WOCE hydrographic program observed in 2003 and 2011 also reveal the above variation of AAIW in the last decade (Fig. 5a and b).

This freshening in the intermediate water layer is thought to be compensated for by increased salinity in shallower thermocline water, indicating a contemporary intensification of hydrological cycle (Figs. 3b and 5b). In this case the freshwater input from atmosphere to ocean surface increased in the subpolar high-precipitation region and vice versa in the subtropical high-evaporation region (Fig. 4a). Over the last 10-year time period, significant freshwater gain began around 2003 (Fig. 4b), suggesting that the observed freshened AAIW could be traced back to this time.

Against the background of hydrological cycle intensification, the decrease of AL transport is proposed to contribute to the freshening of AAIW in the South Atlantic, associated with a weakening of westerlies over the South Indian Ocean. This decrease was revealed by the leakage evaluation along the Good Hope section. The mechanical analysis shows that the WS over the South Indian Ocean reached its peak around 2000–2004 and began to subside through 2005–2009 to 2010–2014 (Fig. 7), reversing its increasing phase from the 1950s to the beginning of the 2000s, during which period the AL had increased (Durgadoo et al., 2013; Lübbecke et al., 2015). This indirectly estimated variability of AL is consistent with other studies covering a similar period (Biastoch et al., 2015; Le Bars et al., 2014). As the AAIW carried by the AL is more saline relative to its counterpart in the South Atlantic Ocean, its decrease would promote the effect of freshwater input from the source region. Our estimate further suggests that such an induced freshwater input by AL could account for approximately 53 % of the observed freshening.

One might ask if there are any other sources that could significantly affect the AAIW in the South Atlantic Ocean, for example the Southeast Pacific (see Sect. 1). To clarify this question, we displayed the first pattern of empirical orthogonal function (EOF1) and its time series (called the principal components) of salinity on the 27.36 γ^n (around 27.2 σ_0) surface (Fig. 8) in the Southern Hemisphere, which explains 55.4 % of the variance. It shows that in 2000–2014, the most significant salinity reduction appeared in the South Indian Ocean, especially in the region of the Agulhas Current system. It also shows that compared to the west Atlantic, the east Atlantic (whose intermediate water is largely fed by its counterpart in the South Indian Ocean) experienced a major salinity reduction. In addition to these salinity changes, we also note that the salinity decrease in the southeast Pacific was considerably less than that in the South Indian and the South Atlantic. Therefore, it implies that the Southeast Pacific did not play an important role in our observed AAIW freshening.

The purpose of this work is to reveal the decadal freshening of AAIW in the South Atlantic Ocean over the last 10-year time period, and suggest the related contributing mechanism. Future work should be focused on the quantification of these two contributors, and the influence they have on the world ocean circulation, through modeling studies.

Data availability. The Argo data were collected and made freely available by the International Argo Program and the national programs that contribute to it (http://www.argo.ucsd.edu, http://argo.jcommops.org). The Argo Program is part of the Global Ocean Observing System. NCEP Reanalysis 2 data were provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at http://www.esrl.noaa.gov/psd/.

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Competing interests. The authors declare that they have no conflict of interest.

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