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Widespread freshening in the Seasonal Ice Zone near 140°E off the Adélie Land Coast, Antarctica, from 1994 to 2012

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[1] Long-term water mass changes during 1994–2012 are examined from nine repeat hydrographic sections in the Seasonal Ice Zone along 140°E, off Antarctica. Significant freshening trends are detected within most of the water masses from the bottom to surface. Bottom Water freshened by 0.008–0.009 decade^{−1} below isopycnal surfaces and its layer thickness decreased by 120–160 dbar decade^{−1} throughout the study period. In addition to general thinning, the layer thickness was anomalously thin in 2012, suggesting a possible link with the sudden calving of the Mertz Glacier Tongue and subsequent reduction in sea-ice production. Winter Water freshened by 0.03 decade^{−1} throughout the study period, with significant interannual variability. In the offshore region, a long-term increase in precipitation can explain a substantial portion of the freshening trend. The Lower Circumpolar Deep Water on the continental slope underwent freshening at the same rate as the Bottom Water during the last two decades. Modified Shelf Water also shows robust freshening at a rate of 0.03 decade^{−1}. Combined with the freshening of near-surface and Bottom Water masses in this region, these data indicate freshening of the entire water column over the continental slope. This widespread freshening is broadly consistent with the enhancement of the global hydrological cycle, together with a possible acceleration of land ice melting.

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1. Introduction

[2] The Southern Ocean plays pivotal roles in the meridional overturning circulations of the global oceans. The overlying wind system induces poleward upwelling of deep water, and the subsequent transformations through air-sea-

ice fluxes produce surface and Bottom Waters that are exported equatorward [e.g., Rintoul *et al.*, 2001]. Salinity and freshwater flux, which is defined as precipitation plus glacial and sea-ice meltwater minus evaporation, are key factors that affect the overturning circulations. The freshwater flux provides positive buoyancy forcing, which transforms upwelled deep water into lighter surface water [e.g., Warren *et al.*, 1996]. Precipitation dominates evaporation in the Southern Ocean and glacial melt supplies freshwater to the ocean. Sea ice also contributes to the freshwater redistribution: melting of sea ice adds freshwater, while sea-ice formation increases salinity of the underling sea water there. Strong salt gain due to intense sea-ice formation in coastal polynyas increases the density of shelf water until the water sinks down the continental slope and entrains surrounding fluid to form Bottom Waters [e.g., Baines and Condie, 1998].

[3] Together with changes in momentum and heat flux between the atmosphere and the ocean, changes in freshwater flux will affect the overturning circulation. Changes in surface and subsurface salinity have been detected globally from the 1950s to the 2000s; sea surface salinity increased in regions of high salinity and decreased in regions of low salinity [Boyer *et al.*, 2005; Hosoda *et al.*, 2009; Durack and Wiffels, 2010]. This observed pattern is broadly consistent with an enhancement of the global hydrological cycle [e.g., Held and Soden, 2005].

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Concurring with this global signature, near-surface water masses in the Southern Ocean have freshened [Boyer *et al.*, 2005; Böning *et al.*, 2008], possibly at a rate exceeding that inferred from the IPCC class model (CMIP-3) ensembles [Helm *et al.*, 2010]. Changes in local precipitation–evaporation (P–E) balance, contributions from the formation and melting of sea ice, and melting of land ice including icebergs and ice shelves can significantly affect the oceanic salinity. However, these factors vary spatially, especially the freshwater derived from the melting of land ice, and their temporal changes are not yet completely quantified. Given differing geographical and climatological environments, regional analysis of water mass changes is invaluable in precisely describing the variability and inferring the cause of those changes.

[4] Among vast region surrounds the Antarctic, the region off the Adélie Land Coast near 140°E is a particularly important region, characterized by the strong wind stress curl [e.g., Trenberth *et al.*, 1989], intense sea-ice production [e.g., Tamura *et al.*, 2008, 2012], and Bottom Water formation [Rintoul, 1998; Williams *et al.*, 2010]. Here, Antarctic Bottom Water is formed through the export of Dense Shelf Water, originating from intense sea-ice production in the George V Land polynya system, which joins the Bottom Water flowing westward from the Ross Sea.

[5] In the latter half of the 20th century, changes have been observed in the water masses constituting the two overturning cells in this region. For the water masses within the lower cell, a freshening trend was observed in the Bottom Water (BW) in the region off the Adélie Land Coast and Australian–Antarctic Basin [Jacobs, 2004; Aoki *et al.*, 2005a; Rintoul, 2007; Johnson *et al.*, 2008; Jacobs and Giulivi, 2010; Purkey and Johnson, 2013]. For the period 1995–2008, Shimada *et al.* [2012] estimated a salinity trend of -0.004 ± 0.0022 decade⁻¹ and a temperature trend of 0.0021 ± 0.0024 °C decade⁻¹ as a meridional average on isobars along 140°E using the WOCE Hydrographic Program (WHP) SR3 and Japanese repeat hydrographic sections. Freshening of the Ross Sea Shelf Water and Bottom Water (RSBW) has contributed significantly to the freshening of the Antarctic Bottom Water in this region [Rintoul, 2007; Jacobs and Giulivi, 2010; Shimada *et al.*, 2012]. However, in previous analyses of long-term changes in water properties, the contributions of interannual variability have not been properly estimated. Recently, calving of the Mertz Glacier Tongue occurred in February 2010 [Young *et al.*, 2010], and subsequent shrinkage of the polynya led to abrupt reduction in sea-ice production [Tamura *et al.*, 2012] and dense water formation [Shadwick *et al.*, 2013]. This change may result in a decrease in BW production [Kusahara *et al.*, 2011]. Quantifying the long-term and interannual changes of BW properties here is important, given their possible impact on deep-water temperature changes further downstream in the South and North Pacific [Kawano *et al.*, 2006; Masuda *et al.*, 2010].

[6] Significant changes in salinity have also been observed in the water masses of the upper overturning cell. In the Indian Ocean area, including the region off the Adélie Land Coast, however, detailed studies of the salinity change and its causes have been relatively limited. Using SURVOSTRAL repeated XBTs, thermosalinographs, and WHP SR3 data during 1992–2004 in the Antarctic Zone,

mainly north of 58°S, Morrow *et al.* [2008] revealed that surface water freshened and warmed at a rate of -0.077 decade⁻¹ and 0.67 °C decade⁻¹, respectively, and Winter Water (WW) cools at a rate of -0.25 °C decade⁻¹. However, the change in upper-ocean salinity has not been adequately revealed, and descriptions of property changes are limited, especially in the Seasonal Ice Zone (SIZ) south of 60°S.

[7] Property changes are also reported within the subsurface Circumpolar Deep Water (CDW), which is transported from the lower-latitude deep oceans, upwells, and compensates for the equatorward export of bottom and surface waters. Aoki *et al.* [2005b] described warming and salinification along isopycnals during the period from the 1950s to the 1990s south of the Polar Front of the Antarctic Circumpolar Current (ACC) between 30 and 160°E. Morrow *et al.* [2008] reported that the CDW layer warmed at a rate of 0.04 – 0.05 °C decade⁻¹ in the Antarctic Zone north of 58°S. However, the property changes in the waters near the southern end along 140°E are also not clearly traced.

[8] On the continental shelf off the Adélie Land Coast, even less is known about the long-term changes in water mass properties. Lacarra *et al.* [2011] observed interannual salinity variability: the Dense Shelf Water was fresher in 2009 than in 2008 in the Adélie Depression around 145°E. In the D’Urville Trough at 140°E, Aoki *et al.* [2005a] showed salinity to be fresher in 2002 than in 1994. Long-term freshening has been reported in the Ross Sea Shelf [Jacobs and Giulivi, 2010] and in the northwest Weddell Sea Shelf [Hellmer *et al.*, 2011], and hence the observed salinity difference off the Adélie Land Coast may be influenced by long-term, large-scale variability patterns. There are few studies describing the long-term salinity changes in this shelf region, however [Shadwick *et al.*, 2013].

[9] While several studies have documented water mass changes in this region, the nature and causes of changes in the SIZ remain poorly understood. Moreover, to quantify the long-term trend, the relative contributions of interannual and longer-term variations should be separated to reduce any possible aliasing effect from inadequate temporal sampling. From the 1990s, the 140°E section near the continent has been occupied relatively frequently by Australian WHP SR3, French SURVOSTRAL and ICOTA/ALBION/CEAMARC [Lacarra *et al.*, 2011], and Japanese research cruises. Hence, this section is a scientifically valuable transect along which to quantitatively evaluate and investigate the oceanic long-term changes.

[10] To investigate the causes of the salinity changes, an understanding of the temporal variation of freshwater fluxes is necessary. The high-latitude Southern Ocean is affected by freshwater fluxes from precipitation–evaporation (P–E), formation and melting of sea ice, and melting of land ice including icebergs and ice shelves. Recently, some progress has been made in estimating time series for these fluxes. Estimates of P–E became available from atmospheric reanalysis and other observational data sets [e.g., Kanamitsu *et al.*, 2002], and time series of sea-ice production rates have been derived from satellite data and heat flux calculations [Tamura *et al.*, 2008, 2011]. Although the accuracy of these products is difficult to assess and rates of land-ice melt are largely unknown, the estimates provide at least a qualitative measure of temporal

variability of freshwater input. Comparisons with changes in ocean salinity could in turn validate the freshwater flux estimates.

[11] The purpose of this paper is to analyze and quantify the long-term and interannual variability of the water masses that constitute the overturning cells in the SIZ along 140°E off the Adélie Land Coast. The causes of the near-surface salinity variability are then examined and compared with the available data sets of freshwater fluxes.

2. Analysis Domain and Data Sources

[12] Here, we introduce the geographical settings and climatological water mass distributions in the analysis domain. We then describe the hydrographic data used, including the data sets of freshwater fluxes.

2.1. Flow Field and Water Mass Distribution

[13] The 140°E transect crosses the eastern portion of the Antarctic-Australian Basin (Figure 1). While the continental landmass in this sector extends relatively far north, the ACC migrates to the south, so the fronts of the ACC are relatively close to the continent (Figure 2a). The southern branch of the polar front (PF-S) is situated around 60°S, the northern and southern branches of the southern ACC front at 62 and 63°S, and the southern boundary of ACC (SB) at 64.5°S on the continental slope [Sokolov and Rintoul, 2002, 2009; Chaigneau *et al.*, 2004; Aoki *et al.*, 2006]. Along the 63–64°S latitude band, which coincides with the latitude band of the Antarctic Divergence, a series of cyclonic eddies have been observed across the transect [Hirawake *et al.*, 2003; Aoki *et al.*, 2007]. A prominent bathymetric bump at 63°S coincides with the location of the eddy at 140°E. The maximum sea-ice edge (SIE) in winter (defined by 30% sea-ice concentration, after Yuan and Martinson [2000]) is located around 62.5°S. On the upper continental slope at 65°S, the Antarctic Slope Front (ASF), associated with westward flow, divides the shelf and offshore regimes [Aoki *et al.*, 2006].

[14] Water mass properties are closely related to the flow field and air-sea-ice fluxes, which are constrained by bathymetric and geographic features. The near-surface water in the majority of the analysis domain is covered by fresh and cold Antarctic Surface Water (AASW; Figure 2b). From fall to winter, the mixed layer of this AASW deepens, integrating the overlying shallow mixed layer. From spring to summer, a warm and fresh shallow layer develops near the surface. Beneath this shallow summer mixed-layer is the WW, which largely preserves the water properties of the previous winter mixed layer. Water mass properties in the WW core therefore reflect the integrated variability of air-sea-ice interactions, and so it is one of the key layers in tracing the effects of climate variability. Under the WW lays the warm and saline CDW, whose salinity maximum has its origin in the North Atlantic Deep Water. It upwells and becomes the ultimate source of the water masses in the Southern Ocean. According to its different pathways, CDW is subdivided to Upper and Lower varieties. The depths of the WW and CDW cores are shallowest at the latitude of the Antarctic Divergence and then deepen equatorward.

[15] Density surfaces of the BW deepen to the north and extend to around 58°S south of the South Indian Ridge.

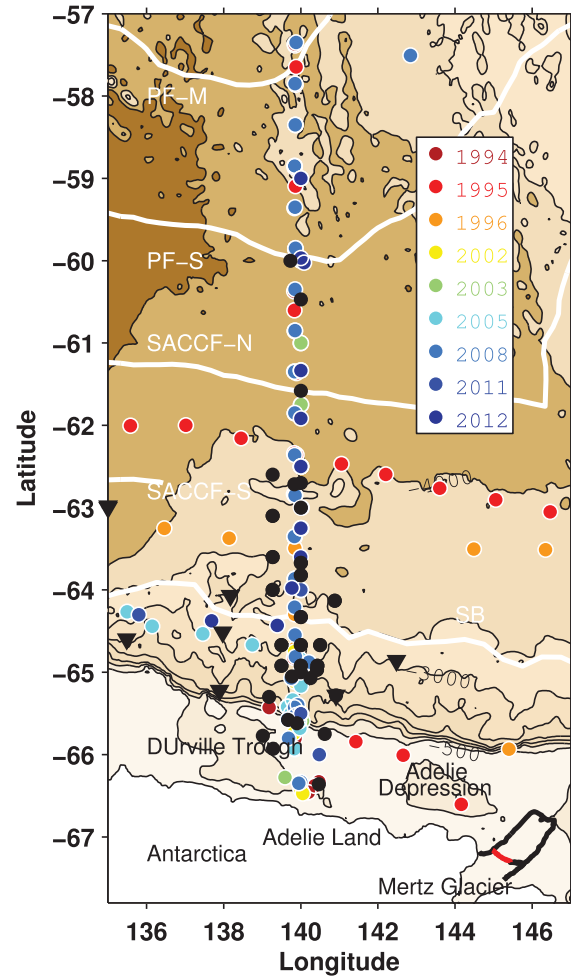


Figure 1. Location of summertime observations near 140°E. Colors of the dots indicate the observation years (see legend). Black dots (triangles) denote observations from historical data (those for Bottom Water) from the Southern Ocean Database. White solid lines denote the fronts of the Antarctic Circumpolar Currents (Middle and Southern branches of Polar Fronts, Northern and Southern branches of Southern ACC Fronts after Sokolov and Rintoul [2009], and Southern Boundary of ACC as the southernmost latitude of 1.5°C potential temperature maximum derived from climatological mean field). Thick solid lines denote the Mertz Glacier Tongue before (black) and after (red) the calving in February 2010.

BW found in this region is a mixture of RSBW and down-slope flow of dense water exported through the Adélie/Mertz Sill off the Adélie Land/George V Land coasts just to the east (Adélie Land Bottom Water (ALBW)) [Rintoul, 1998; Williams *et al.*, 2010]. The Adélie Depression around 145°E is the major source of Dense Shelf Water. Here a large polynya was located to the west side of the Mertz Glacier Tongue extended equatorward until the sudden calving of the edge in February 2010 [Tamura *et al.*, 2012].

[16] In this analysis, we define the core of each water mass as follows: Winter Water by a temperature minimum, and Upper and Lower CDW by temperature and salinity

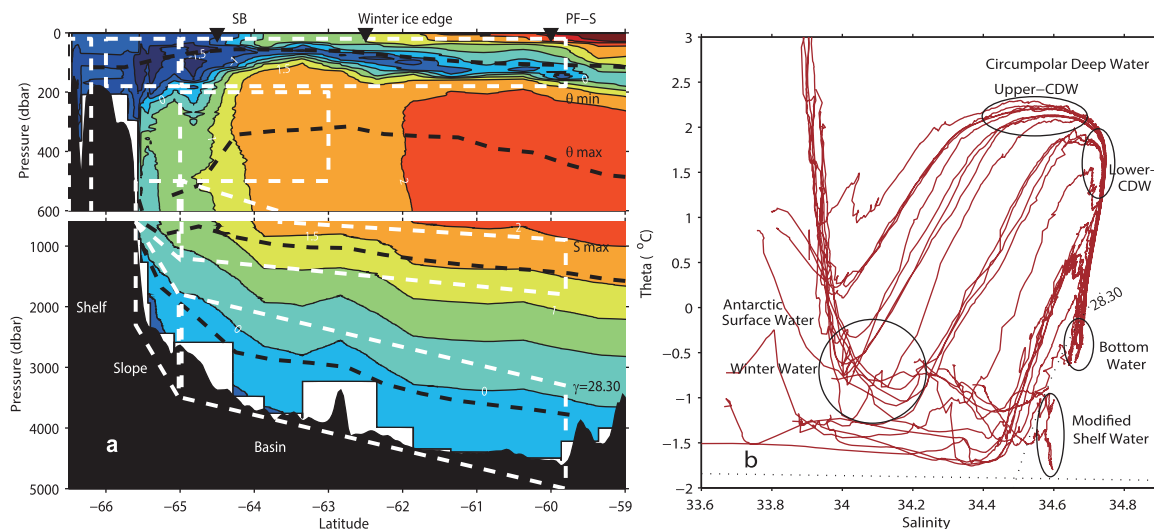


Figure 2. (a) Vertical section of potential temperature along 140°E of WHP SR3 cruise in January 1994. Contour interval is 0.5°C. Pressure levels of potential temperature minimum and maximum, salinity maximum, and neutral density of 28.30 kg m⁻³ are indicated by broken black lines. Latitudes of climatological mean ACC fronts (Southern branch of Polar Front and Southern Boundary of ACC) and winter sea-ice edge are indicated by triangles. The analysis subdomains are also shown with the boxes outlined with dashed white lines. (b) Potential temperature—salinity diagram of sea waters for the same 1994 cruise. Core properties of water masses are indicated by schematic ovals.

maxima, respectively. Analysis of properties is carried out at the core of each water mass. BW is defined as the water whose neutral density is greater than the density criteria of 28.30 and 28.27 kg m⁻³, depending on the region, and the analysis of salinity and potential temperature is conducted for averages within those parts of the water column that meet the density criteria both for each section and climatological section (i.e., according to the fixed density and pressure surfaces).

[17] At the southern end of the 140°E meridian lies the D'Urville Trough, whose maximum depth is around 1000 dbar. Modified CDW is advected into this trough, beneath which is the Modified Shelf Water (MSW) [Orsi and Wiederwohl, 2009; Lacarra et al., 2011].

2.2. Hydrography Data

[18] During the summers (January–February) in the period from 1994 to 2012, nine top-to-bottom conductivity–temperature–depth (CTD) transects, complete or partial, were conducted south of 60°S along 140°E (Figure 1). In 1994, 1995, 1996, and 2008, WOCE Hydrographic Program (WHP) SR3 transects were occupied by RSV Aurora Australis. In the 2000s, intensive repeat hydrography was begun by the Japanese Antarctic Research Expedition (JARE) and Kaiyodai Antarctic Research Expedition (KARE). In 2002 and 2003, JARE observations were conducted on R/V Tangaroa. In 2003, 2005, 2007, 2011, and 2012, KARE and KARE/CEAMARC transects were occupied by R/V Umitaka Maru. Not all the cruises covered the whole transect south of 60°S. The southern end of the 2012 KARE cruise was at 64.4°S due to the presence of sea ice. In contrast, the 2003 KARE and 2007 KARE/CEAMARC cruises were conducted only on the shelf. CTD observations are conducted to WOCE standards; the temperature and salinity data are estimated to be accurate to within

0.002°C and 0.002. Data are obtained within 20 m from the seafloor, with a few exceptions of 30 m due to rough sea conditions. Hence, the bottommost property was taken to be the deepest level of observation.

[19] To clarify regional characteristics, the analysis domain is meridionally divided into three regions: offshore in the deep basin and over the continental rise (Basin); on the continental slope (Slope); and within the D'Urville Trough on the continental shelf (Shelf; Figure 2). The northern limit is set to 60°S, around the PF-S, due to the limitations of data coverage. The boundary between the Slope and Basin domains is set to 65°S, which roughly coincides with the location of ASF and the 3000 dbar isobath. The southern boundary of the Slope is set to 66°S.

[20] To further examine longer-term water mass variability, hydrographic data before 1994 were obtained from the Southern Ocean database (Orsi, A. H., and T. Whitworth III, WOCE Southern Ocean Atlas, <http://woceSOatlas.tamu.edu>). Only the summer data (in December, January, and February) were used. For the subsurface analysis (for layers of WW, CDW, and MSW), data are selected from the region between 139 and 141°E south of 60°S. For the BW analysis, data availability is even more limited. However, the spatial variability is less than that near the surface, and hence data were analyzed at eight stations by Eltanin in 1969 and 1971, between 135 and 142.5°E (downstream of the Adélie Sill). For the data in the D'Urville Trough, one profile taken in December 2000 by R/V Nathaniel B. Palmer was added to the analysis. The data were obtained from the World Ocean Database 2009 [Boyer et al., 2009].

[21] To reduce spatial aliasing on temporal variability, we calculated 0.5° × 0.5° climatological fields of temperature, salinity, and neutral density at the standard depths (53 layers from 0 to 6000 m) using World Ocean Database 2009. Instead of usual climatological fields whose

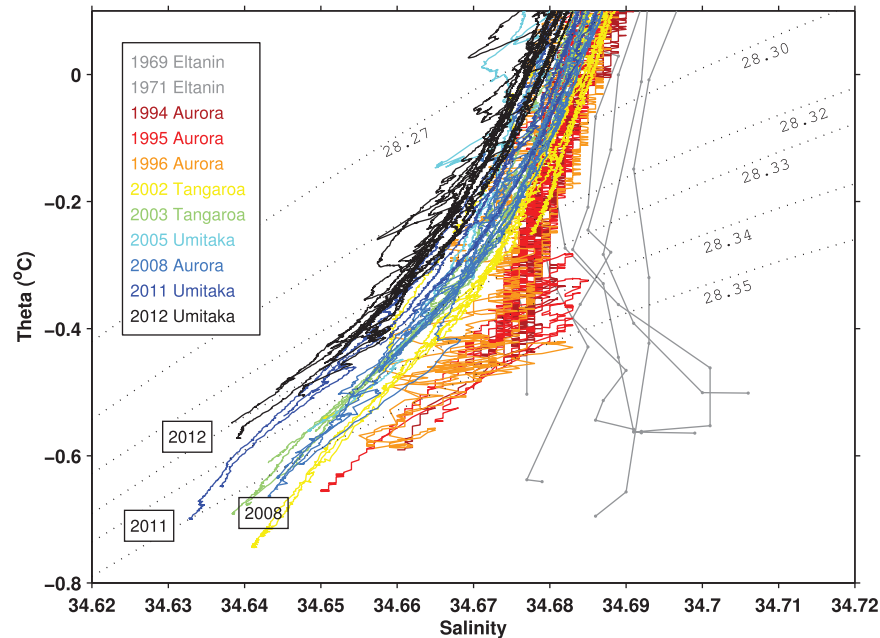


Figure 3. Potential temperature–salinity diagram of observations for 60–65°S band near 140°E. Legend denotes observation year and vessel. Color lines indicate the CTD observations in the legends. Gray dots denote bottle observations in 1969 and 1971 by R/V Eltanin. Background contours in broken lines are neutral density surfaces.

weighting functions are dependent only on latitudinal and longitudinal distances, we created a climatology that takes isobath direction into account to retain the structure following the local bathymetry.

2.3. Freshwater Fluxes

[22] To investigate the causes of observed salinity changes, data sets of P-E, sea-ice production, and SIE location were obtained for the main period of interest (1993–2012). Meteoric water transport from lower latitudes is the ultimate reason for the relatively fresh nature of the near-surface layers in the Southern Ocean. For precipitation, estimates from IsoGSM, which originates from NCEP/DOE R2 [Yoshimura *et al.*, 2008; Kanamitsu *et al.*, 2002] were used. For evaporation, the OaFlux data set (<http://oafux.whoi.edu>) was used. These data sets are provided from 1979.

[23] Near-shore salinity could be affected by sea-ice production in the polynya. Time series of sea-ice production in Adélie and Mertz Depression regions were obtained from satellite microwave observations [Tamura *et al.*, 2008, 2012]. Thin ice thickness in the polynyas is estimated from 85 and 37 GHz brightness temperature data obtained from the Special Sensor Microwave Imager (SSM/I) using the Tamura *et al.* [2007] algorithm. By using the ice thickness, sea-ice production is estimated from heat flux calculation during the freezing period by assuming that all of the heat loss at the surface is used for ice formation [Tamura *et al.*, 2008]. In addition, the potential effect of offshore sea-ice cover was also examined with time series of satellite-derived SIE. The latitude of maximum SIE, defined by a sea-ice concentration of 30% derived from SSMI [Yuan and Martinson, 2000], is investigated as a proxy for possible sea-ice melting/freezing at the SIE.

3. Results

[24] Temporal changes in water mass properties during the 18 year period from 1994 to 2012 are examined for the BW, WW, and CDW on the continental slope and in the deep basin with high-resolution CTD data. Property changes of the MSW on the shelf are also studied for the 14 year period from 1994 to 2008. Salinity changes of the WW layer are then compared with the time series of freshwater fluxes and their proxies.

3.1. Bottom Water

[25] It has been reported that water mass characteristics of BW off the Adélie Coast are subject to significant change. BW freshened from 1994 to 2008 [e.g., Aoki *et al.*, 2005a; Rintoul, 2007; Shimada *et al.*, 2012]. Therefore, near-bottom salinity and temperature properties are reexamined for the region 60–65°S, augmented with the updated observations.

[26] After 2008, a general freshening continued up to the summers of 2011–2012 (Figure 3). BW salinity was freshest in 2011. Potential temperature at the bottom was coolest in 2002, and was warmer in 2008–2012. In 2012, it was distinctively warmer than in 2011. Potential temperature exhibits considerable fluctuation against the long-term trend, compared to that of salinity. Through a combination of these salinity and temperature changes, the density of the Bottom Water is at a minimum in 2012.

[27] To quantitatively estimate the long-term and inter-annual variability of BW in the Basin and Slope regions, bottom properties are averaged on isopycnal and isobaric surfaces. The properties are averaged below the neutral density surface 28.30 kg m^{-3} at each point in the Basin

(28.27 kg m^{-3} on the Slope; because there is no water denser than 28.30 kg m^{-3} at some points on the upper slope). Vertical averages are also calculated for the fixed depth range whose climatological neutral density is larger than 28.30 kg m^{-3} in the Basin and 28.27 kg m^{-3} on the Slope. Their values are spatially averaged for the respective region each year. They are then separated into the trend and residual components, representing the long-term and interannual variability, respectively. SR3 observations at 62.8°S in 1994, 1995, and 2008 are omitted because sudden shoaling of isopycnals, due to the steep bathymetry in the area, causes significant departures from the other years' data that do not assist in resolving the bathymetric feature.

[28] In the Basin, salinity measurements reveal a dominant freshening trend within the isopycnal layer; salinity decreases significantly at a rate of $-0.0083 \pm 0.0023 \text{ decade}^{-1}$ (error is the 95% one-tail confidence interval, unless otherwise specified hereafter; Figure 4a). The trend component explains 87% of the total variance. The trend and variance are summarized in Table 1 along with other relevant variables. Potential temperature, however, does not reveal a monotonic pattern (Figure 4b). After the minimum in 2002, the averaged temperature was consistently warmer. The trend throughout the period ($-0.014 \pm 0.025^\circ\text{C decade}^{-1}$) explains only 14% of the total variance. Because the freshening trend dominates the statistically insignificant temperature trend, the spatially averaged density at the bottom decreased. It exceeded 28.35 kg m^{-3} during the former half of the period, 1994–2003 but became less than 28.35 kg m^{-3} during the latter half, 2005–2012 (Figure 4c). In 2012, it was even lower than 28.34 kg m^{-3} . The rate of bottom density change is $-0.0082 \pm 0.0062 \text{ kg m}^{-3} \text{ decade}^{-1}$. In addition to this decrease in bottom density, the BW thickness (denser than 28.30 kg m^{-3} surface) decreased from around 900 to 700 m during the period (Figure 4d). The rate of the layer thickness change is significant at $-122 \pm 60 \text{ dbar decade}^{-1}$, which explains 68% of the total variance of the layer thickness. The BW layer, hence, thinned by more than 20% during the study period.

[29] Analysis on isobaric coordinates reveals similar trends. Freshening trends are significant ($-0.0064 \pm 0.0028 \text{ decade}^{-1}$; Figure 5a) with its mean estimate slightly less than the trend on isopycnals. Temperature trend is positive ($0.017 \pm 0.038^\circ\text{C decade}^{-1}$), although it is statistically insignificant. The decreasing trend in density is almost the same as found in isopycnal coordinates.

[30] BW on the Slope generally reveals similar variability to that in the Basin both within isopycnal and isobaric layers, with a larger contribution of interannual variability (Figures 4e–4h and 5e–5h). The salinity trend was significant at $-0.0093 \pm 0.0054 \text{ decade}^{-1}$ and $-0.0070 \pm 0.0046 \text{ decade}^{-1}$, respectively, while the temperature change was insignificant (-0.004 ± 0.057 and $0.038 \pm 0.060^\circ\text{C decade}^{-1}$). The variance of the salinity anomaly explained by the trend is 65% and 59%. Density at the bottom decreased at a rate of $-0.016 \pm 0.010 \text{ kg m}^{-3} \text{ decade}^{-1}$. The rate of the layer thickness change is $-163 \pm 137 \text{ dbar decade}^{-1}$, which explains 47% of the total variance. Magnitudes of salinity, density, and layer thickness trends agree well with those in the Basin. Analysis on isopycnal and isobaric coordinates indicates consistent trends dominated by significant freshening. Hence, the freshening and layer shrinking

trends are significant over an area extending from the continental slope to the deep basin.

3.2. Winter Water

[31] In the WW core, salinity and temperature have stronger meridional gradients compared to those of the BW. The magnitudes of the spatial variability are larger than that of the temporal variability. Generally, the core salinity is higher to the south; it is around 34.0 at 60°S and increases to 34.4 at $65\text{--}66^\circ\text{S}$. This tendency is at least partially attributed to high sea-ice production [Rintoul and Bullister, 1999]. The core temperature is around 0°C at 60°S and decreases to near the surface freezing point on the slope. The core pressure is 100–150 dbar at 60°S , shoals to around 50 dbar at $64\text{--}65^\circ\text{S}$, and deepens again to the south. Since the spatial variability is large compared to the temporal variability and the observation locations differ from year to year, it is necessary to remove the effects of this spatial variation to extract the temporal variability. To remove the spatial dependence, climatological mean properties on the climatological temperature minimum surface for the three parameters are calculated and removed in both Slope and Basin, respectively. Anomalies of more than two standard deviations were excluded. The anomalies are then spatially averaged for each year in the two regions: Slope and Basin.

[32] On the Slope, the salinity anomaly was generally higher in the first half of the study period and reached a maximum in 2003 (Figure 6d). It was lower in the late 2000s and achieved a minimum in 2011. The salinity anomaly shows a significant decreasing trend of $-0.034 \pm 0.027 \text{ decade}^{-1}$. The trend component explains around half (51%) of the total variance (these values are summarized in Table 2).

[33] Variables other than salinity do not reveal any significant trend on the Slope. The temperature anomaly increased at a statistically insignificant rate of $0.057 \pm 0.12^\circ\text{C decade}^{-1}$ with the trend explaining 13% of the variance. Pressure anomalies in the WW core do not show significant trend.

[34] In the Basin, trends of almost similar magnitudes are seen (Figure 6a and Table 2). The salinity anomaly trend is $-0.028 \pm 0.024 \text{ decade}^{-1}$, which explains 42% of the variance. The magnitude of the trend component is similar to that on the Slope, although interannual variability is much larger. Temperature and pressure anomaly trends were not statistically significant.

[35] Overall, a freshening trend of about $-0.03 \text{ decade}^{-1}$ is detected in both the Slope and Basin. However, temperature and core pressure do not show statistically significant trends.

3.3. Circumpolar Deep Water

[36] Outcropping of CDW and subsequent transformation by air-sea fluxes result in water masses supplying the upper and lower overturning cells in the Southern Ocean [Rintoul et al., 2001]. Hence, changes in the properties of CDW would also induce changes in other water masses. Like the WW, there are strong meridional gradients of temperature and salinity along the CDW core. Salinity in the core of the Upper CDW increases poleward from 34.55 at 60°S to 34.7 at 65°S , while temperature cools from around

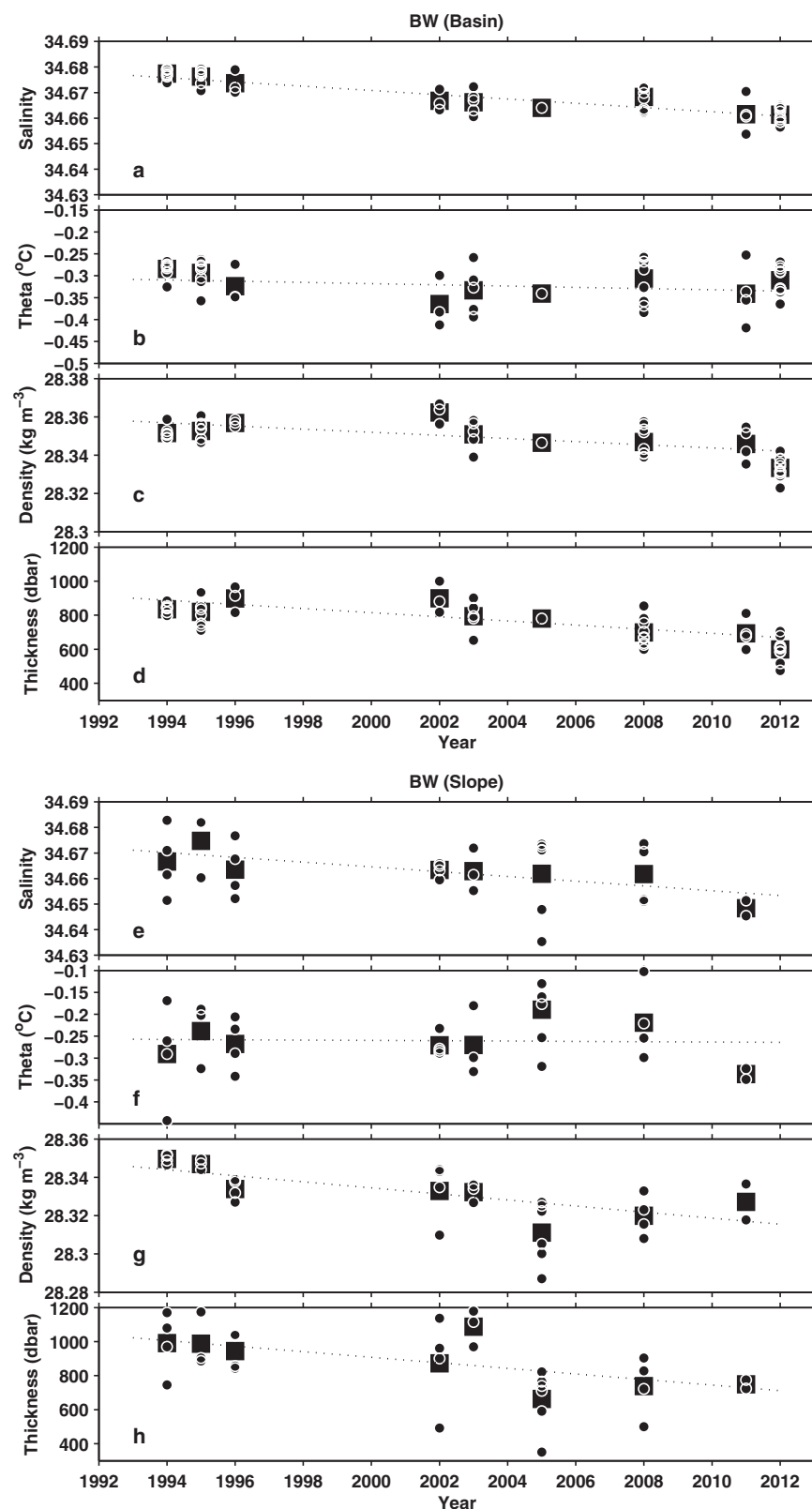


Figure 4. Time series of (a and e) salinity, (b and f) potential temperature, (c and g) bottom density, and (d and h) layer thickness (in pressure unit) for the Antarctic Bottom Water in the Basin (a–d) and on the Slope (e–h) averaged below the neutral density criteria of 28.30 kg m^{-3} (28.27 kg m^{-3}) in the Basin (on the Slope), respectively. Each dot corresponds to the CTD data from a station and black squares denote the cruise averages. The broken lines denote the linear regressions to the cruise averages.

Table 1. Long-Term Trends and Explained Variances of Water Mass Properties for Bottom Water Averaged on Isopycnal and Isobaric Coordinates^a

	Salinity Trend (decade ⁻¹) Variance	Theta Trend (°C decade ⁻¹) Variance	Density Trend (kg m ⁻³ decade ⁻¹) Variance	Thickness Trend (dbar decade ⁻¹) Variance
<i>Isopycnal</i>				
Bottom Water	-0.0083 ± 0.0023	-0.0140 ± 0.0251	-0.0082 ± 0.0062	-121.7 ± 59.6
(Basin)	0.87	0.14	0.47	0.68
Bottom Water	-0.0093 ± 0.0054	-0.0037 ± 0.0569	-0.0158 ± 0.0102	-163.4 ± 136.9
(Slope)	0.65	<0.01	0.60	0.47
<i>Isobaric</i>				
Bottom Water	-0.0066 ± 0.0028	0.0170 ± 0.0380	-0.0082 ± 0.0042	–
(Basin)	0.74	0.09	0.66	–
Bottom Water	-0.0070 ± 0.0046	0.0380 ± 0.0604	-0.0150 ± 0.0101	–
(Slope)	0.59	0.20	0.58	–

^aError bar is 95% one-tail confidence interval.

2.2°C at 60°S to 1.0°C at 65°S. The properties of the Upper CDW at the temperature maximum were examined at 63–65°S, which coincides with the latitude band of the Antarctic Divergence. Latitudes north of 63°S were excluded because the presence of cyclonic eddies significantly affects the properties on this scale and the data coverage here does not resolve the eddy completely. Anomalies were calculated relative to the climatological fields, as done for WW.

[37] The salinity anomaly of the Upper CDW increased at a rate of 0.0078 ± 0.0066 decade⁻¹ (at 90% one-tail confidence level; Figure 7a). The variance explained by this trend was 29%, which is not a dominant factor. The temperature anomaly increased at a rate of 0.060 ± 0.047 °C decade⁻¹ (at 90% one-tail confidence level; Figure 7b). The pressure anomaly showed insignificant change. Therefore, long-term trends suggest a shift toward saltier and warmer conditions, although for all variables the variances of the interannual variability predominate over those of long-term trends.

[38] For the Lower CDW, the water properties at the salinity maximum were examined in the same way. The climatological mean salinity at the salinity maximum is relatively constant (~ 34.74) for 60–63°S, and freshens toward the south, while the mean temperature is 1.5–1.6°C for 60–63°S and gets cooler toward the south. To measure this particular water mass, the analysis was carried out in both the Slope and the Basin.

[39] On the Slope, the salinity anomaly decreased at a rate of -0.0056 ± 0.0044 decade⁻¹ (Figure 8d). Temperature and pressure trends are insignificant when compared with their interannual variabilities. In the Basin, the temperature anomaly increased (Figure 8b) although the rate is statistically insignificant. Salinity and core pressure did not change significantly on a long-term scale. Hence, the changes found for Lower CDW differ in the two domains. The salinity trend is strong and comparable to that of the BW on the slope, but becomes insignificant further offshore.

3.4. Modified Shelf Water in the D’Urville Trough

[40] On the shelf, bathymetry and external forcing conditions are complex, and oceanic properties tend to mostly reflect the local conditions, and hence spatial proximity is

most critical in extracting temporal variability. In the D’Urville Trough, we can compare, relatively closely, 7 (8) stations within 10 n miles (20 n miles) in 4 (5) years of 1994, 2000 (December), 2002 (2003), and 2008, spanning 14 years (Figure 9). The average bottom depth of the CTD casts is around 800 dbar. The salinity below the maximum temperature in MSW was relatively homogeneous; therefore, we averaged the profiles from the bottom to the nearest local temperature-maximum layer (indicated by squares in Figure 9). The averaged thickness of this layer is roughly 250 dbar.

[41] From 1994 to 2008, salinity of the MSW decreased at a rate of -0.030 ± 0.009 decade⁻¹. The magnitudes of the decreases from 1994 to 2002 and from 2002 to 2008 are nearly equal. This freshening rate roughly concurs with the salinity anomaly trend of WW on the Slope and in the Basin, indicating that this freshening, of a similar magnitudes regionally, is a broad-scale feature, although the layer thickness is substantially thicker on the shelf.

3.5. Relationship Between Salinity Change of Winter Water and Freshwater Fluxes

[42] The salinity change in the WW layer is relatively well sampled in time, and hence, it is feasible to compare it with the effects of freshwater fluxes.

[43] On the Slope, close to the shelf, local freshwater flux from sea-ice production or melting and P-E can both be significant contributors to salinity change. Since the cause of the higher salinity to the south is partly attributed to high sea-ice production in the George V Land polynya system, the variation in ice production could alter the WW salinity on the Slope. The WW salinity anomalies were therefore compared with sea-ice production in the winters preceding the observed summers.

[44] Sea-ice production has significant interannual variability (Figure 10a). Among those years in which there were hydrographic observations (open circles), the production is highest in 2002 and lowest in 2010. In the mid-1990s, it was also relatively higher than the average from the 2000s. The correlation coefficient between the salinity and sea-ice production for the available 8 years was 0.45, which is not statistically significant. Hence, the variability of polynya sea-ice production may partly influence the

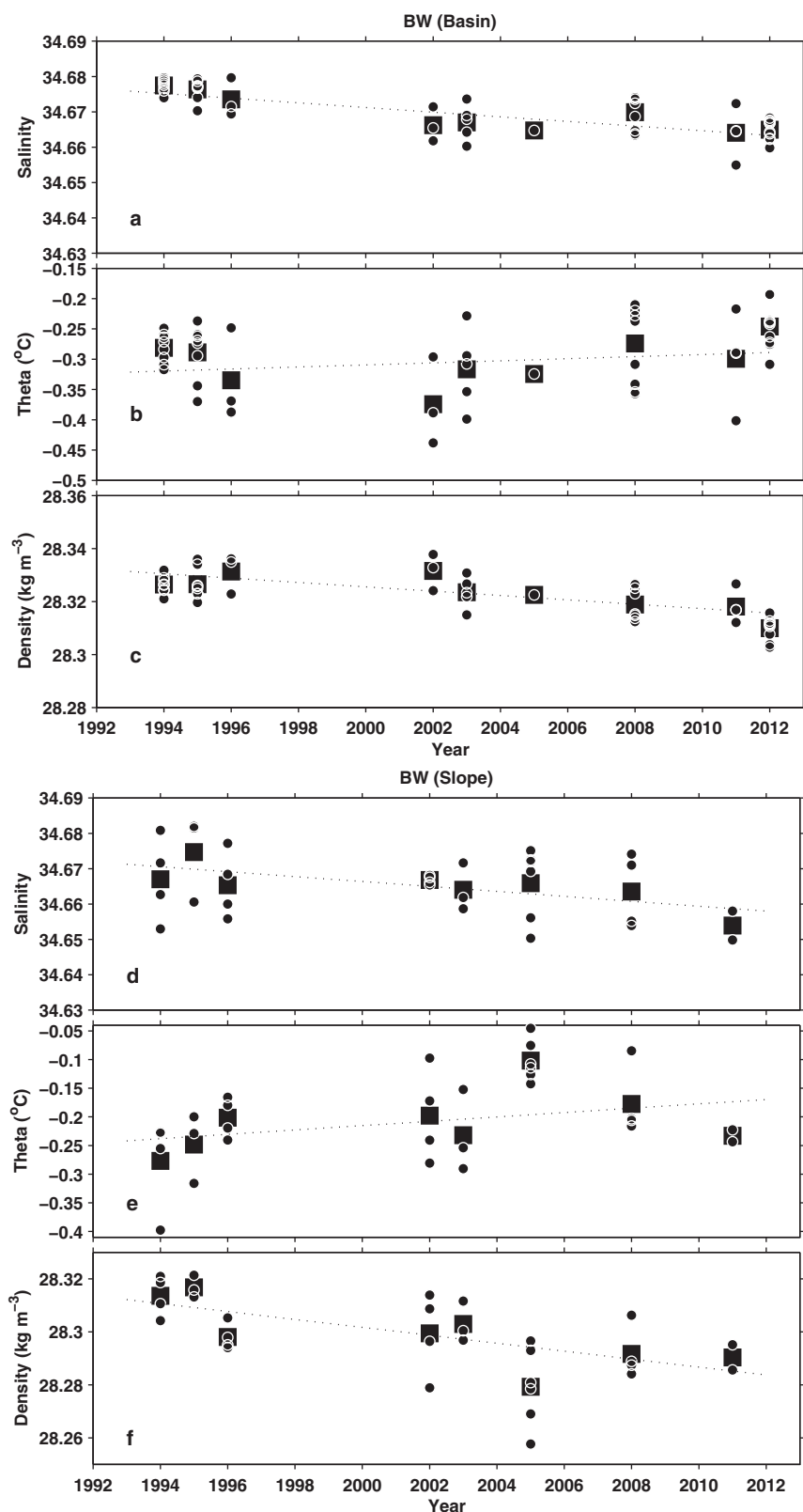


Figure 5. Time series of (a and d) salinity, (b and e) potential temperature, and (c and f) density for the Antarctic Bottom Water in the Basin (a–c) and on the Slope (d–f) averaged below the climatological pressure criteria derived from neutral density criteria of 28.30 kg m^{-3} (28.27 kg m^{-3}) in the Basin (on the Slope), respectively. Each dot corresponds to the CTD data from a station and black squares denote the cruise averages. The broken lines denote the linear regressions to the cruise averages.

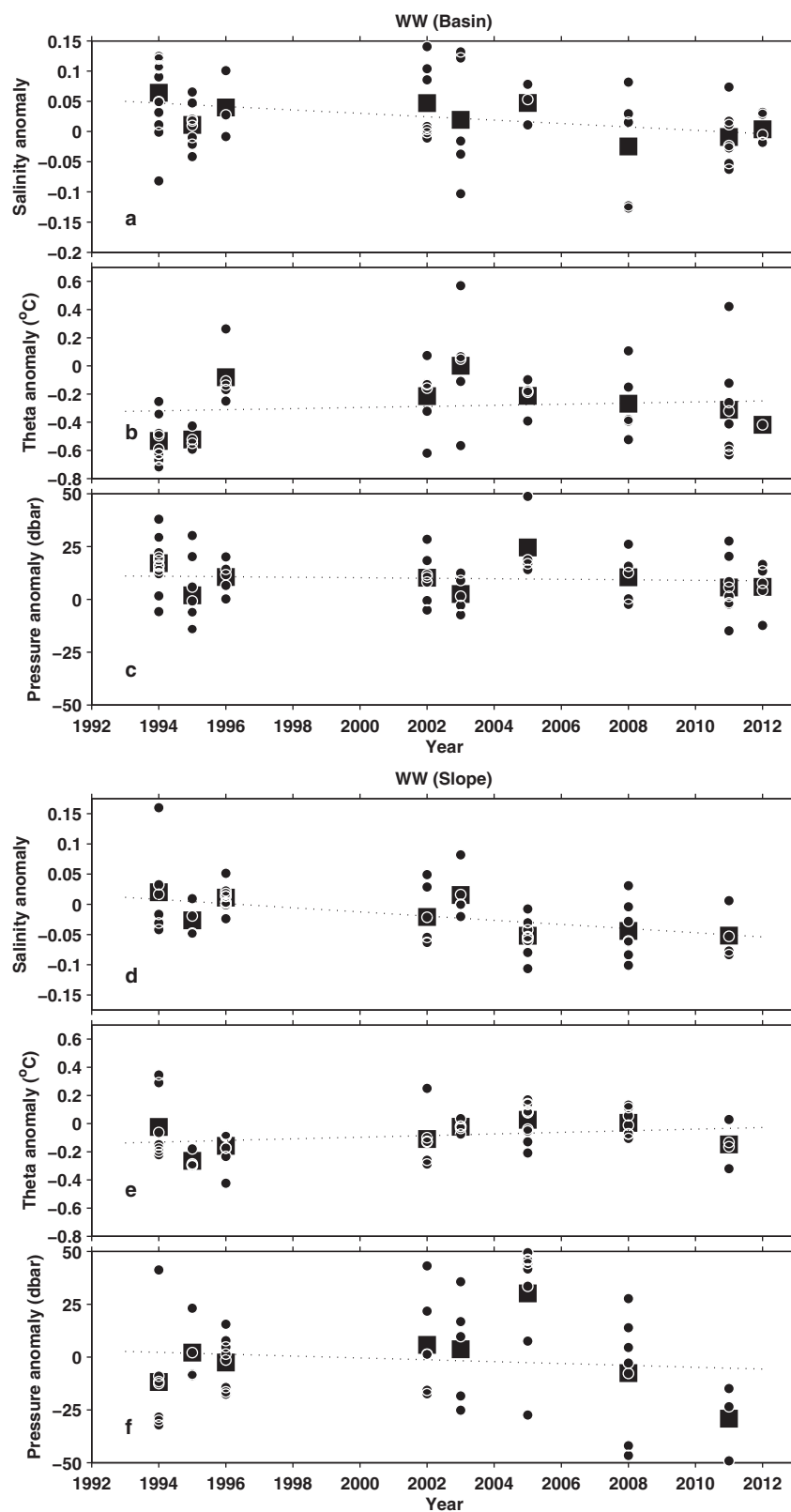


Figure 6. Time series of (a and d) salinity anomaly, (b and e) potential temperature anomaly, and (c and f) pressure anomaly of the Winter Water core in the Basin (a–c) and on the Slope (d–f), respectively. Each dot corresponds to the CTD data from a station and black squares denote the cruise averages. The broken lines denote the linear regressions to the cruise averages.

Table 2. Long-Term Trends and Explained Variances for Water Mass Properties^a

	Salinity Trend (decade ⁻¹) Variance	Theta Trend (°C decade ⁻¹) Variance	Pressure Trend (dbar decade ⁻¹) Variance
Winter Water (Basin)	-0.0281 ± 0.0237 0.42	0.0391 ± 0.1919 0.02	-0.121 ± 0.761 0.01
Winter Water (Slope)	-0.0345 ± 0.0270 0.51	0.0575 ± 0.1182 0.13	-0.441 ± 2.110 0.03
U-CDW (63–65°S)	$0.0078 \pm 0.0066^*$ 0.29	$0.0604 \pm 0.047^*$ 0.32	12.20 ± 39.96 0.05
L-CDW (Basin)	$4.8e-4 \pm 0.0029$ 0.01	0.0377 ± 0.0548 0.20	13.18 ± 33.02 0.08
L-CDW (Slope)	-0.0056 ± 0.0044 0.50	0.0099 ± 0.1940 <0.01	25.52 ± 91.90 0.05
Modified SW (D'Urville Trough)	-0.0305 ± 0.0092 0.87	— —	— —

^aError bar is 95% one-tail confidence interval. Note that * indicates 90% one-tail confidence interval.

interannual variability of WW salinity downstream, but it cannot be the dominant factor.

[45] With regard to the longer-term salinity trend, the sea-ice production again cannot be the primary cause. The trend in sea-ice production from the observed years (broken line in Figure 10a) indicates an overall decrease, which could be qualitatively consistent with the observed WW freshening. However, using the regression coefficient between sea-ice production and salinity anomaly, the freshening rate that can be statistically supported was estimated as an order of magnitude smaller than the observed salinity tendency, and hence the trend of sea-ice production cannot explain the long-term salinity trend.

[46] In addition to sea-ice production, the WW salinity anomaly was compared with the annually integrated P-E for the preceding winter (the year of integration spans from November to October, assuming the winter mixing ends in October; i.e., WW salinity in January 1994 is compared with the P-E flux integrated from November 1992 to October 1993). The local P-E does not have a statistically significant correlation. Therefore, the variability in the salinity anomaly in this region cannot be explained quantitatively by changes in sea-ice production or P-E.

[47] Considering the salinity anomaly in the Basin, the freshwater flux from P-E may be larger than that in the south and sea-ice melting can also be a controlling factor. The WW salinity was therefore compared with the annually integrated P-E for the preceding winter, as done for the Slope. In addition to P-E, the location of maximum SIE in the previous year was compared with WW salinity (i.e., winter water salinity in January 1994 is compared with maximum SIE latitude in 1993).

[48] The correlation with P-E estimates, derived for the 8 years of available hydrography data, is -0.59 , which is statistically significant at the 90% one-tail level (Figure 10b compare to Figure 6a). The negative correlation indicates the logical relationship that higher precipitation leads to a lower salinity. The standard deviation of P-E (0.09 m) is roughly comparable to that of the salinity anomaly (0.03) with the simple assumption that the freshwater input is distributed over the layer thickness of 75 m, which is the average depth of the WW core.

[49] Trends of P-E derived from both the entire 19 year period and the years in which ocean observations are avail-

able (solid and broken lines in Figure 10b) increased. As for the sea-ice production contribution on the Slope, the potential contribution of P-E was statistically estimated. Using the linear regression, P-E changes could explain -0.015 decade⁻¹, which is equivalent to 56% of the observed freshening in the Basin.

[50] Correlation with SIE latitude is -0.34 , which is not statistically significant (Figure 10c). If the SIE is the location of net melting of drifting sea ice, then an extended SIE will lead to greater melting and hence a lower salinity. However, the observed SIE retreat (both for observed and total years; solid and broken lines in Figure 10c) is not consistent with the observed freshening, given the negative correlation above.

[51] Overall, none of the factors considered (changes in P-E, polynya sea-ice production, and SIE location) can alone explain the observed salinity trend of around -0.03 decade⁻¹. Insufficient accuracy of the flux estimates and infrequent sampling might have obscured the relationship. In the Basin, increasing P-E could contribute substantially to the observed freshening trend, although other contributors are needed to fully explain the variability.

4. Discussion

[52] A significant freshening tendency was observed for all water masses, except CDW, in the study region. Here we first examine consistency of the observed freshening signal with previous studies. Additional analysis is attempted to investigate the continuity of the trend during the 18 years for the longer-time scale extending back to the 1970s. We then further investigate the causes of these signals and compare with results from other Antarctic margins.

4.1. Bottom Water

[53] Significant freshening of BW continued throughout the period 1994–2012. The present estimate of -0.0066 ± 0.0028 decade⁻¹ on isobars is slightly larger but agrees within uncertainties with the previous estimate of -0.0042 ± 0.0022 decade⁻¹ between 1995 and 2008 along the same section [Shimada *et al.*, 2012], which is reasonable given the differences between study regions and periods, and the presence of interannual variability.

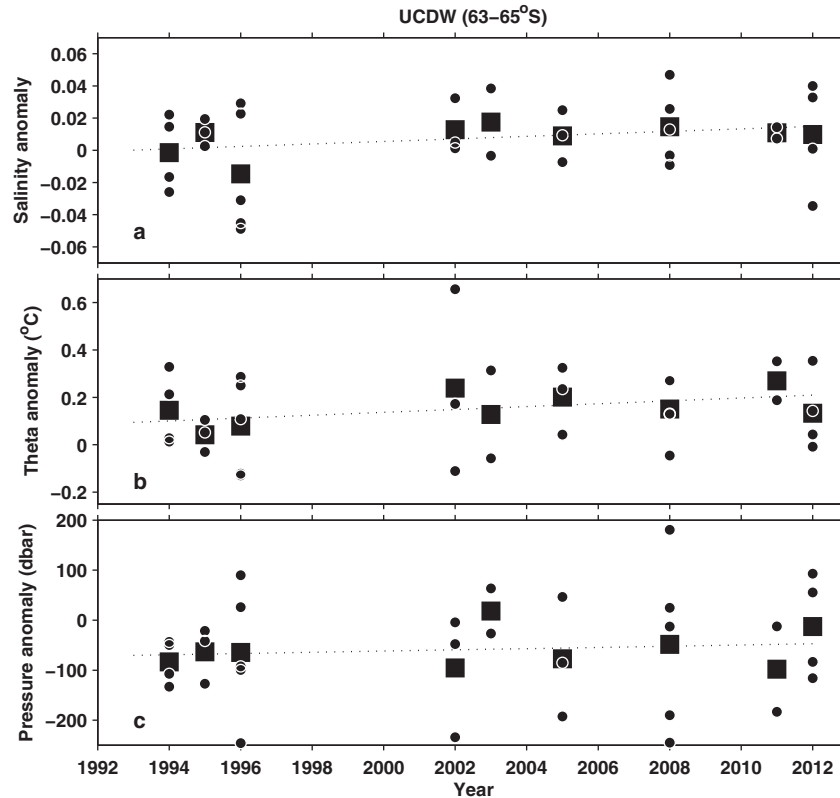


Figure 7. As for Figure 6 but for Upper Circumpolar Deep Water for the 63–65°S region.

[54] There is a possibility that the environmental conditions responsible for formation of this BW and its variation altered as a result of the sudden calving of Mertz Glacier Tongue and subsequent reduction of sea-ice formation [Tamura *et al.*, 2012; Shadwick *et al.*, 2013]. Whether the lowest salinity anomaly in 2011–2012 reflects this effect is statistically unclear. Calculating the trend without the data from 2011/2012 will result in a change of less than 5%. A change in temperature is yet more difficult to discern, since its interannual variability is large. However, the change in layer thickness is prominent. The estimated mean trend for the precalving period of 1994–2008 (-83 ± 65 dbar decade $^{-1}$; 90% one-tail interval) is 30% less than that (-122 ± 45 dbar decade $^{-1}$; 90% one-tail interval) for 1994–2012, although these estimates overlap within their error bars. Hence, we can say that a discernible thickness decrease, and hence reduction of BW volume, occurred in 2012. This change may be an effect from the calving event, but it is necessary to wait for the accumulation of future observations before the effect can be quantified.

[55] The observations around 1970 also indicate continuous freshening for more than three decades [Aoki *et al.*, 2005a; Rintoul, 2007], although the available data are limited to those obtained by R/V Eltanin. Comparisons were made with the historical observations to examine the continuity of the trends in the Basin (Figure 11). The salinity trend between 1969 and 2012 was -0.0055 ± 0.0015 decade $^{-1}$ and the mean estimate was 66% of that during 1994–2012. In the same way, the observed layer thickness trend was -86 ± 29 dbar decade $^{-1}$ and 70% of that obtained from the present rate. Thus, the overall freshening and thin-

ning tendencies are continuous from 1970, but it is likely that the trends have accelerated.

[56] Substantial portions of these BW changes have been attributed to property changes in the incoming Ross Sea Bottom Water [Jacobs and Giulivi, 2010; Shimada *et al.*, 2012]. From the observations of the deepest 600 m layer for 1500–2500 m depths in an area northwest of the Ross Sea, Jacobs and Giulivi [2010] derived salinity trends to be -0.008 decade $^{-1}$ from the 1950s to the 2000s. Shimada *et al.* [2012] observed a salinity change of -0.005 decade $^{-1}$ for the bottom 100 m, from 1969 to 1997 data around 150°E, which is consistent with the observed change of -0.0045 decade $^{-1}$ over the four decades in the northwest Ross Sea [Ozaki *et al.*, 2009]. The magnitude of the salinity change in the Ross Sea and for the downstream near 150°E is roughly comparable to the observed estimate along 140°E, and therefore more than half of the salinity change in the Australian-Antarctic Basin could be attributed to the changes in properties and transport of water masses in the Ross Sea [Shimada *et al.*, 2012].

[57] The decrease in layer thickness found in this study agrees well with the rate of global contraction of the Antarctic Bottom Water (AABW) [Purkey and Johnson, 2012]. The mean rate of isotherm descent in their study was estimated to be 14 m yr $^{-1}$ between the 1980s and 2000s.

4.2. Winter Water

[58] The freshening trend is significant for WW on the Slope and Basin. Morrow *et al.* [2008] detected surface water freshening of -0.077 decade $^{-1}$ for the region

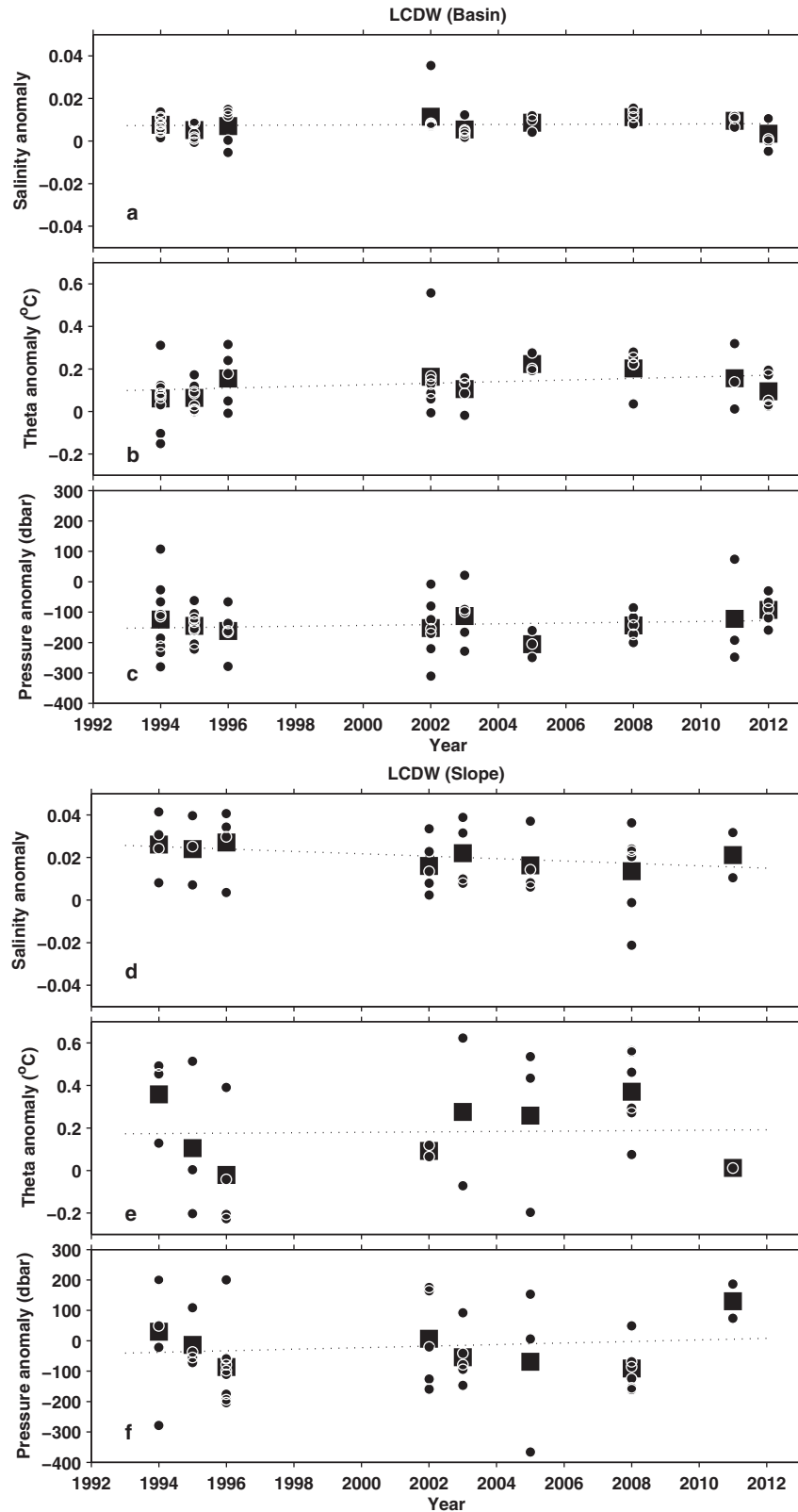


Figure 8. As for Figure 6 but for Lower Circumpolar Deep Water in the Basin (a–c) and on the Slope (d–f).

55–58°S over recent decades. They also attributed the freshening to increased precipitation. Their results cover a region slightly equatorward of the Basin domain and differ

in magnitude and depth but are basically consistent with those of the present study. On the Slope, however, there is little evidence of a trend and the cause of the observed

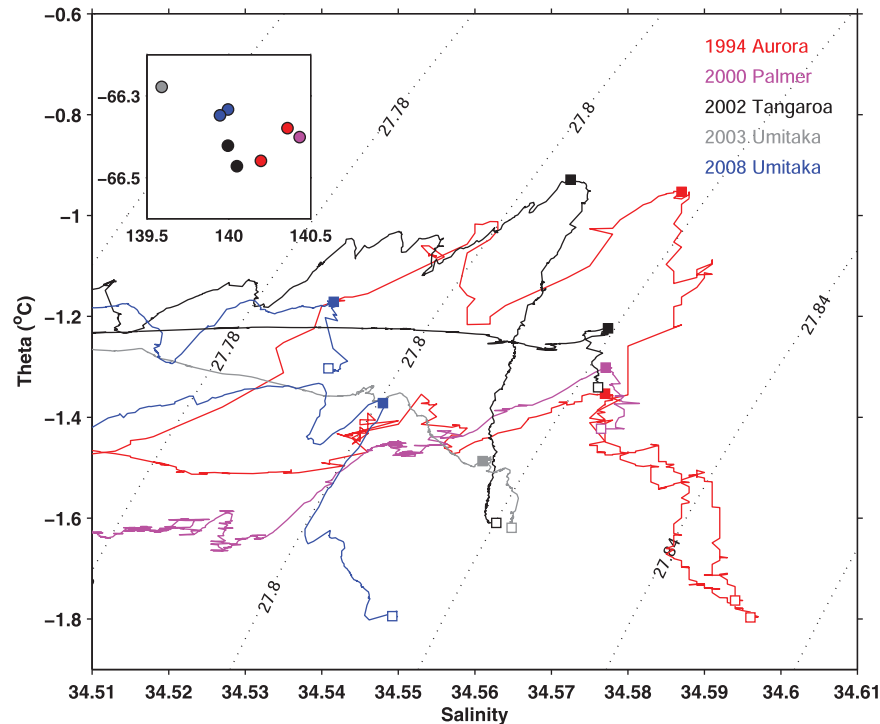


Figure 9. Potential temperature–salinity diagram of observations in the D’Urville Trough. Open (filled) squares denote measurements at the bottom (the local temperature maximum near the bottom). Legend denotes observation year and vessel. Color lines indicate the CTD observations in the legends. Background contours in broken lines are potential density referenced to the sea surface. Dots in the inset figure denote the observation points.

changes is not clear. In addition to changes in freshwater flux discussed above, changes in wind stress might contribute, for example, through changes in Ekman transport of freshwater. However, local zonal wind stress estimated from ERA-interim data set did not show significant correlation or trend in the easterly winds. It is possible that there is not a single dominant factor in this transition zone, and hence further records are needed to quantify the contribution of different drivers of salinity change.

[59] Historically, WW has been better sampled than BW. With bottle data from the 1960s to 1994s, salinity anomalies of the WW core were examined, although the sampling was mostly limited to standard depths. A salinity anomaly was calculated by subtracting the climatological fields as in section 3.2. The salinity anomaly generally increased from the 1960s to the 1980s, and decreased from the mid-1990s in both the Slope and Basin (Figure 12). This difference suggests the presence of multidecadal variability in WW salinity.

[60] The causes of this possible multidecadal variability are not clear. An increase in P-E from the 1980s to the 2000s, which was seen in IsoGSM/OAFlux, can potentially lead to freshening in this period. However, the possible salinification before 1980 is difficult to explain. Although there is no direct evidence, the polynya off the Mertz Glacier could have persisted from at least the 1960s to 2010s, due to the elongation of the Mertz Glacier Tongue during that period [Frezzotti *et al.*, 1998]; with the assumption that the polynya forms on the lee-side of the Tongue. If so, sea-ice production should have generally increased

throughout the period. The effect of the temporal variability of land ice melting is not clear. Multidecadal changes in those freshwater fluxes may result in changes of their relative contributions.

4.3. Modified Shelf Water in the D’Urville Trough

[61] During 1994–2008, significant freshening was observed in the D’Urville Trough. The cause of the freshening on the shelf, however, is not clear. Local P-E in the IsoGSM/OAFlux does not show any significant trend. Sea-ice production during 1993–2010 in the upstream polynya system increased, rather than decreased (Figure 10a). The behavior of land ice melt is not well observed and there is no evidence for increasing local ice shelf melt in the Mertz Glacier region. Advection from the east may be one reason, as in the case of the Ross Sea. Melt of Cook Ice Shelf could also be a cause [Tamura *et al.*, 2012], but further investigation is necessary to fully understand this component.

[62] The freshening tendency on the shelf over the recent two decades is not confined to this area. The magnitude of the freshening rate, $-0.03 \text{ decade}^{-1}$ is roughly equal to that of the Modified-CDW ($-0.04 \text{ decade}^{-1}$) and the High Salinity Shelf Water ($-0.03 \text{ decade}^{-1}$) on the Ross Sea continental shelf [Jacobs and Giulivi, 2010]. Freshening of the Ross Sea is attributed to upstream ice shelf melting of the West Antarctic Ice Sheet [Rignot *et al.*, 2008; Jacobs *et al.*, 2011]. Hellmer *et al.* [2011] found a change of $-0.05 \text{ decade}^{-1}$ (0.09 for the 17 year period consisting of 1989, 1997, and 2006) on the western Weddell Sea continental shelf. Possible causes of the freshening in this region

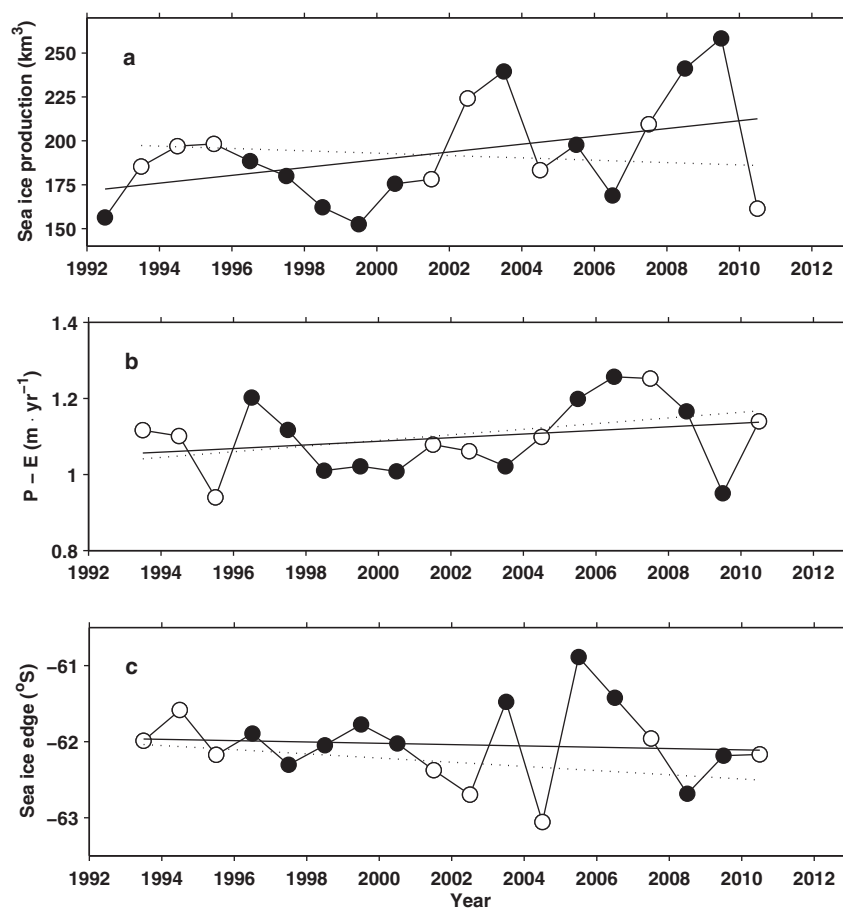


Figure 10. Time series of (a) annual sea-ice production in the polynya system, (b) annual precipitation—evaporation (P-E; average in the 60–65°S region), and (c) latitude of maximum winter sea-ice edge. Dots in Figures 10a and 10b denote the annual total values, derived as summation from the preceding November to October. Open circles denote values for the years that have oceanographic observations in the subsequent summer. Solid lines are linear regression for all years from 1993 to 2011, while broken lines are derived from the years of oceanographic observations only.

are the southward retreat of the summer SIE, and increased precipitation. Therefore, although the magnitudes of the freshening are similar, the factors contributing to the long-term variability are probably different from region to region.

4.4. Circumpolar Deep Water

[63] Saltier and warmer trends are detected in the Upper CDW. CDW warming of $0.04\text{--}0.05^{\circ}\text{C decade}^{-1}$ was observed at 300–500 m depths in the Antarctic Zone [Morrow *et al.*, 2008]. This estimate is consistent with the rate found in the present study.

[64] As with WW, salinity and temperature anomalies were derived from the historical record. Both salinity and temperature increased from the mid-1960s to the 1980s, which is consistent with the trends in the recent CTD record (Figure 13). Aoki *et al.* [2005b] described an increase in salinity (0.022 decade^{-1}) and temperature ($0.30^{\circ}\text{C decade}^{-1}$) on the isopycnals south of 60°S for $30\text{--}160^{\circ}\text{E}$ from the 1950s to 1990s. The overall picture is consistent.

[65] The structure of these changes can be explained by increased upwelling, which is also inferred from a decrease

in dissolved oxygen in this region [Aoki *et al.*, 2005b; Helm *et al.*, 2011]. This change is consistent with the general shift to the positive phase of the Southern Annular Mode from at least the 1970s [Thompson and Solomon, 2002; Visbeck, 2009].

[66] Changes in the Lower CDW are quantitatively similar to the changes in other nearby water masses that are spatially proximal. Offshore in the Basin, the Lower CDW warmed with a similar magnitude to the Upper CDW, although the rates are statistically insignificant. Salinity on the Slope freshened at the same rate as the BW. Together with the WW and MSW tendencies, freshening is suggested throughout the whole water column from the coast to the slope.

5. Summary and Conclusion

[67] Along the 140°E transect, significant changes in water mass properties, especially salinity, have been detected from the top to bottom in the period 1994–2012. These changes are compared with the longer-term historical observations from the 1960s. Overall, the Bottom Water freshened and its thickness decreased. These tendencies are

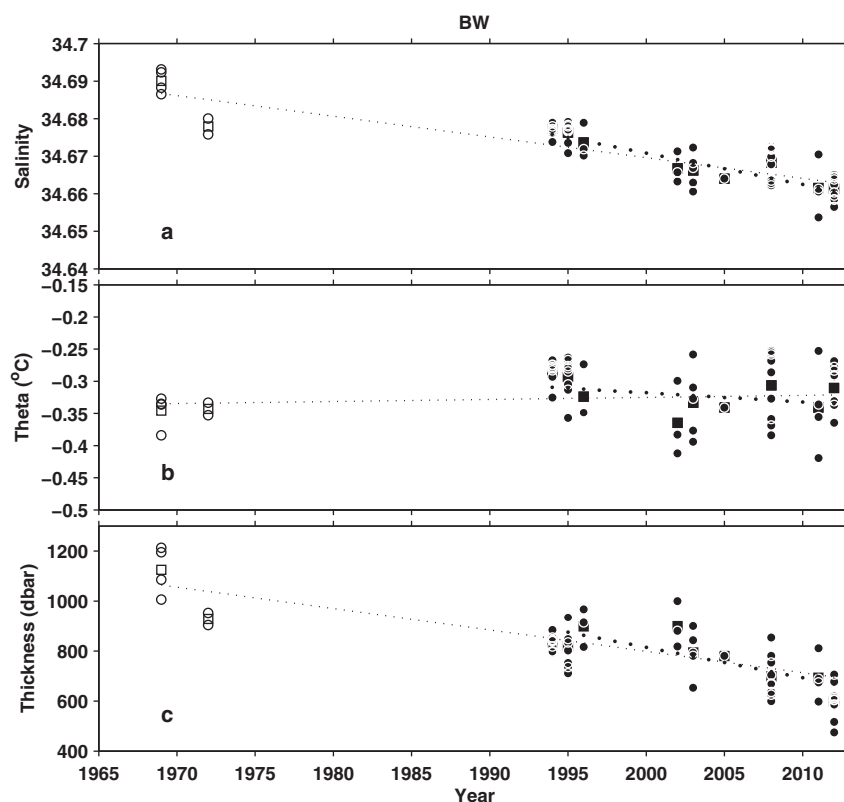


Figure 11. Time series of (a) salinity, (b) potential temperature, and (c) layer thickness of Bottom Water from the late 1970s to 2012 in the Basin. Dots and filled squares (annual average) are derived from the recent CTD observations as in Figure 4, and circles and open squares (annual average) are derived from bottle observations by R/V Eltanin. The broken lines are the linear regression from the recent CTD observations (the same with Figures 4a, 4b, and 4d).

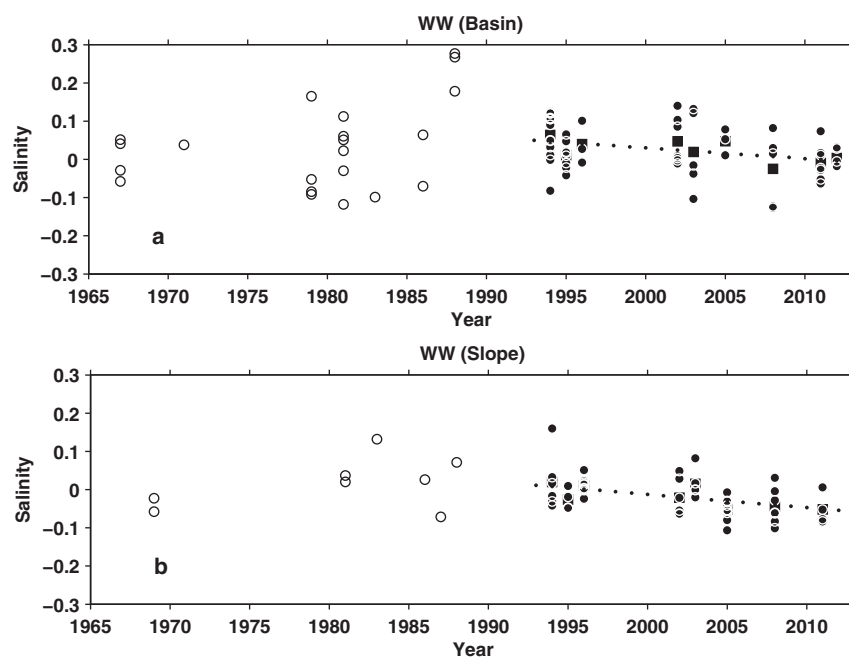


Figure 12. Time series of salinity anomaly of the Winter Water (a) in the Basin and (b) on the Slope. Dots and filled squares (annual average) are derived from the recent CTD observations as in Figures 6a and 6d, and circles are derived from bottle observations obtained from the Southern Ocean Database. The broken lines are the linear regression from the recent CTD observations as in Figures 6a and 6d.

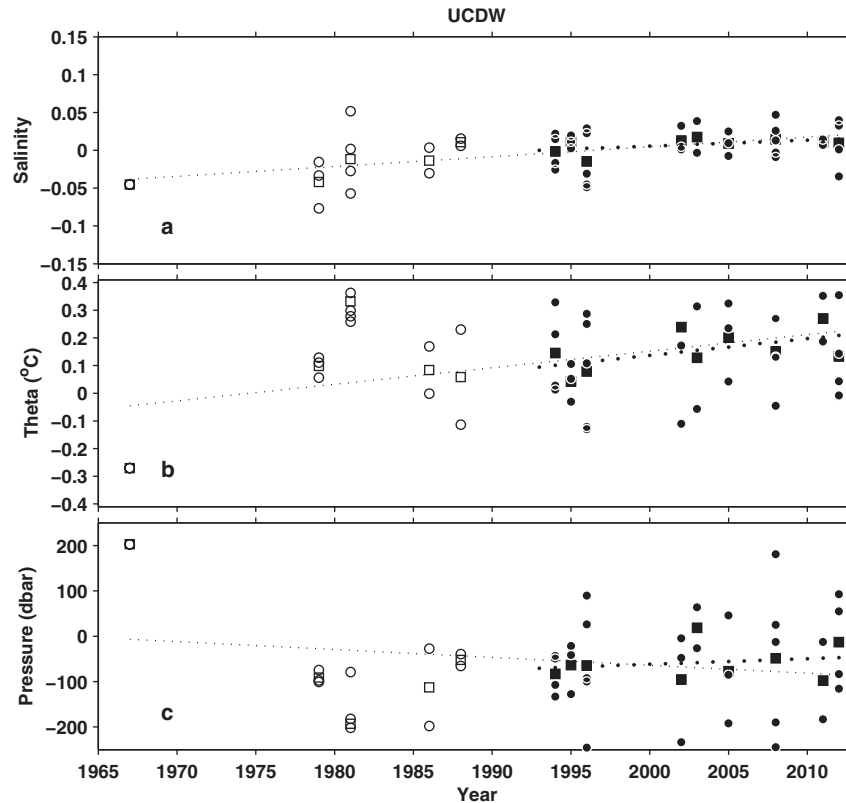


Figure 13. As for Figure 12, but for (a) salinity anomaly, (b) potential temperature anomaly, and (c) pressure anomaly for Upper Circumpolar Deep Water in 63–65°S.

consistent from the 1970s but their magnitudes may have accelerated prior to the 1990s. A prominent decrease in layer thickness was detected in 2012. Since there is a possibility that the sudden calving of the Mertz Glacier Tongue and subsequent reduction in sea-ice production has enhanced the reduction, further continuous observations in the coming years would be valuable in quantifying the effect.

[68] The WW and MSW have freshened at a rate of $-0.03 \text{ decade}^{-1}$, which is broadly consistent with the enhancement of the global hydrological cycle. The dominant causes of these changes differ from region to region. A substantial portion of the freshening at 60–65°S can be explained by an increase in precipitation, but this is not the case for the freshening on the shelf. Historical data show a salinity increase from the 1960s to around 1990s, suggesting the presence of multidecadal variability.

[69] A certain level of correlation between freshwater fluxes and salinity anomalies suggests a physical link between the two and supports the validity of the flux data sets, although further validations and comparisons with diverse data sets are necessary. The contribution of land ice has not been estimated here. Quantifying the estimates of the land ice water flux, including iceberg melting, is critical in linking to the changes in freshwater budget.

[70] This study describes the ongoing freshening throughout the water column in the SIZ at 140°E, except in the CDW, which is formed outside of the Southern Ocean. Given the possible impacts on global overturning circulation, further continuous efforts to observe salinity changes and the freshwater budget, and to survey the spatial extent

of these changes in the region are crucial. In addition to the continuation of hydrographic repeats, a buoy monitoring system capable of acquiring surface flux and water mass characteristics throughout the water column is desirable. To fully describe the status of the change, the development of a sustained observation system under the ice-covered region is necessary.

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