Multidecadal Warming and Density Loss in the Deep Weddell Sea, Antarctica

VOLKER H. STRASS, GERD ROHARDT, TORSTEN KANZOW, MARIO HOPPEMA, AND OLAF BOEBEL

Alfred-Wegener-Institut Helmholtz-Zentrum für Polar- und Meeresforschung, Bremerhaven, Germany

(Manuscript received 16 April 2020, in final form 10 August 2020)

ABSTRACT: The World Ocean is estimated to store more than 90% of the excess energy resulting from man-made greenhouse gas–driven radiative forcing as heat. Uncertainties of this estimate are related to undersampling of the subpolar and polar regions and of the depths below 2000 m. Here we present measurements from the Weddell Sea that cover the whole water column down to the sea floor, taken by the same accurate method at locations revisited every few years since 1989. Our results show widespread warming with similar long-term temperature trends below 700-m depth at all sampling sites. The mean heating rate below 2000 m exceeds that of the global ocean by a factor of about 5. Salinity tends to increase—in contrast to other Southern Ocean regions—at most sites and depths below 700 m, but nowhere strongly enough to fully compensate for the warming effect on seawater density, which hence shows a general decrease. In the top 700 m neither temperature nor salinity shows clear trends. A closer look at the vertical distribution of changes along an approximately zonal and a meridional section across the Weddell Gyre reveals that the strongest vertically coherent warming is observed at the flanks of the gyre over the deep continental slopes and at its northern edge where the gyre connects to the Antarctic Circumpolar Current (ACC). Most likely, the warming of the interior Weddell Sea is driven by changes of the Weddell Gyre strength and its interaction with the ACC.

KEYWORDS: Ocean; Southern Ocean; Ocean circulation; Ocean dynamics; In situ oceanic observations

1. Introduction

Assessing the heat budget of planet Earth is an essential prerequisite for verifying the concept of anthropogenic global warming and the models describing this (e.g., Hansen et al. 2011; Palmer 2012; Trenberth and Fasullo 2012; von Schuckmann et al. 2016). Incomplete measurements of the distribution and temporal changes of heat within the climate system compartments—atmosphere, land, ice, and ocean—bear the risk of leading to incorrect or biased conclusions. When Earth surface temperature measurements indicated a stalled rise of global mean temperature, while atmospheric CO₂ concentrations continued to increase during the 15-yr period following the 1998 El Niño, the paradigm of anthropogenic global warming was challenged. Doubt about its existence was cast among the general public, and used to raise opposition against reduction of man-made CO₂ emissions (e.g., Lewandowsky et al. 2016; Medhaug et al. 2017).

Since 2012 global mean surface temperatures, however, picked up rising and meanwhile have reached new record highs (e.g., <https://crudata.uea.ac.uk/cru/data/temperature/>; GISTEMP Team 2020; Lenssen et al. 2019; NOAA 2020; Zhang et al. 2019). The 1998–2012 slowdown of surface temperature increase, often termed the “global warming hiatus,” and its ending are now widely explained and mostly attributed to variable distributions of heat between the surface of the ocean and its deep interior (Kosaka and Xie 2013; Chen and Tung 2014; Lee et al. 2015; Fyfe et al. 2016).

The World Ocean takes a pivotal role for the heat budget of planet Earth. It features a heat capacity more than 1000 times that of the atmosphere (Levitus et al. 2005; Schmitt 2018) and the oceans are likely to have taken up more than 90% of the excess energy from anthropogenic greenhouse gas–driven radiative forcing (e.g., Levitus et al. 2012; Balmaseda et al. 2013; Rhein et al. 2013; von Schuckmann et al. 2016). Robust estimates of the ocean heat budget are thus urgently needed (e.g., Palmer et al. 2011; Gleckler et al. 2016) to better constrain the climate model sensitivity to perturbations arising from changes in greenhouse gas concentrations (Dessler and Forster 2018; Lewis and Curry 2018) and hence to improve future projections.

Existing global estimates of ocean heat content are based on measurements from different instruments such as expendable bathythermographs (XBTs), autonomous profiling floats (Argo), and conductivity–temperature–depth sondes (CTDs) with inherently different precisions and characteristic profiling depth ranges, namely, 700 m, 2000 m, and full ocean depth, respectively (Abraham et al. 2013; Rhein et al. 2013). Data taken by these instruments also cover different periods of time; for instance, XBTs dominated the number of vertical temperature profiles from the 1970s to 1990s until the establishment of the Argo program, which now delivers more profiles per year than were available before. However, vertical profiles for the deep ocean below the typical 2000-m depth of Argo floats are still sparse, accounting for just a few percent of the overall number of temperature profiles (Durack et al. 2018). Undersampled by any kind of temperature records are the

Denotes content that is immediately available upon publication as open access.

Corresponding author: Volker H. Strass, volker.strass@awi.de

This article is licensed under a [Creative Commons Attribution 4.0 license](http://creativecommons.org/licenses/by/4.0/) (<http://creativecommons.org/licenses/by/4.0/>).

DOI: 10.1175/JCLI-D-20-0271.1

FIG. 1. Study area and location of data records. Map of the Atlantic sector of the Southern Ocean with the generalized circulation of the Weddell Gyre; red curved arrows symbolize the advection and circulation of warm water masses into and within the Weddell Sea; light blue and dark blue curved arrows indicate the formation and outflow of bottom water. Repeated CTD measurements were collected along two transects: the prime meridian (red straight line N: northern Weddell outflow, and blue line E: eastern Weddell inflow) and at the tip of the Antarctic Peninsula across the western outflow (green line W). Further long-term records were obtained from CTD station profiles positioned (location n) on the prime meridian, and (locations e and ω) on the Weddell Sea cross section. The area in light gray indicates the deep Weddell Basin analyzed here; it is bounded by the 700-m isobath in the south and west, a straight line running from the South Orkney plateau to 55°S, 0° in the north, and the prime meridian section N–E in the east. Also shown are the coastline (thick black) and the shelf ice edge (thin black), and the isobaths of 1000 m (brown), 2500 m (yellow), 3500 m (cyan), 4000 m (light blue), and 5000 m (dark blue).

subpolar and polar ocean regions because of the logistic challenges they bear due to their remoteness and partial ice cover.

Undersampling of the subpolar and polar oceans not only adds uncertainty in global ocean assessments in terms of a possible bias. Of even more concern is that it is precisely the subpolar/polar regions where most of the exchange of properties between the sea surface and the deep ocean takes place. A central role in the meridional overturning circulation (MOC) of the global ocean takes the Southern Ocean (e.g., Marshall and Speer 2012), where water masses derived from the North Atlantic Deep Water (NADW), the Upper and Lower Circumpolar Deep Water (UCDW, LCDW), are upwelled into the surface layer owing to the almost circumpolar, wind-driven Antarctic Divergence. The lighter fraction of the upwelled deep water returns back northward with the upper branch of the MOC, and after modification by air–sea interaction subducts at the northern flank of the Antarctic Circumpolar Current (ACC) to form the Antarctic Intermediate Water (AAIW), which spreads northward and

across the equator centered around 1000-m depth (Talley 1996). The denser deep-water fraction is advected southward within the gyres of the Ross Sea and the Weddell Sea (Orsi et al. 1993; Fahrbach et al. 1994, 1995). Within the Weddell Gyre the inflowing heat-advecting Circumpolar Deep Water (CDW) (Reeve et al. 2016, 2019), locally termed Warm Deep Water (WDW), gains density by heat losses to the atmosphere and surrounding ice shelves, and by salinification through brine release during sea ice formation over the continental shelf (Gill 1973). This density gain results in the formation of Weddell Sea Deep and Bottom Waters (WSDW and WSBW). WSDW and WSBW make a major contribution to the Antarctic Bottom Water (AABW; e.g., Foster and Carmack 1976; Fahrbach et al. 1995; Orsi et al. 1999, 2002; Jullion et al. 2014), which spreads northward to fill most of the deep basins, and thus closes the deepest limb of the MOC.

To contribute to reducing the uncertainties in estimates of the global ocean heat budget, our study concentrates on the most severely undersampled parts, the subpolar and polar

TABLE 1. Summary of *Polarstern* cruises during which the database for this study was created. The transect along the prime meridian is marked N and E in Fig. 1, with the distance covered by data calculated from 50°S in the southward direction. The distance along the transect across the Weddell Sea from Joinville Island toward Kapp Norvegia (W and dashed green line in Fig. 1) runs from 63.058°S, 55.266°W to 71.231°S, 10.910°W with a break point at 67.743°S, 20.980°W. The transect off the Antarctic Peninsula (W in Fig. 1) coincides partly with the full Weddell Sea cross section but extends in observation time up to year 2019. The last line of each transect-related part in the table, termed the “used range,” indicates which subsample of the transect was analyzed for long-term trends.

Transect	Cruise	Month and year	Latitude/longitude	Distance (km)
Prime meridian	ANT-X/4	June 1992	55° to 69.5°S	555–2190
	ANT-XIII/4	May 1996	55° to 69.5°S	557–2184
	ANT-XV/4	May 1998	55° to 69.5°S	556–2182
	ANT-XXVIII/3	December 2000	55° to 69.5°S	557–2179
	ANT-XX/2	December 2002	55° to 69.5°S	556–2155
	ANT-XXII/3	February 2005	55° to 69.5°S	557–2177
	ANT-XXIV/3	February 2008	55° to 69.5°S	556–2180
	ANT-XXVII/2	December 2011	55° to 69.5°S	557–2180
	ANT-XXIX/2	December 2012	55° to 69.5°S	556–2180
	ANT-XXIX/6	June 2013	55° to 69.5°S	557–2180
	ANT-XXX/2	December 2014	51.5° to 69.5°S	159–2180
	Used range		55° to 69.5°S	556–2155
Weddell Sea cross section (Joinville Island to Kapp Norvegia)	ANT-VIII/2	September 1989	55° to 11.7°W	10–2192
	ANT-IX/2	December 1990	53.4° to 11.4°W	100–2195
	ANT-X/7	January 1993	53.7° to 11.4°W	80–2195
	ANT-XIII/4	May 1996	53.9° to 12°W	73–2182
	ANT-XV/4	April 1998	53.7° to 27.4°W	80–1376
	ANT-XXII/3	March 2005	54.7° to 13.7°W	32–2169
	ANT-XXIV/3	March 2008	53.3° to 17.3°W	105–1925
	ANT-XXVII/2	January 2011	54.2° to 11.7°W	56–2193
	Used range		53.3° to 17.3°W	105–1925
Western continental slope	ANT-VIII/2	September 1989	55° to 11.7°W	10–2192
	ANT-IX/2	December 1990	53.4° to 11.4°W	100–2195
	ANT-X/7	January 1993	53.7° to 11.4°W	80–2195
	ANT-XIII/4	May 1996	53.9° to 12°W	73–2182
	ANT-XV/4	April 1998	53.7° to 27.4°W	80–1376
	ANT-XXII/3	March 2005	54.7° to 13.7°W	32–2169
	ANT-XXIV/3	March 2008	53.3° to 17.3°W	105–1925
	ANT-XXVII/2	January 2011	54.2° to 11.7°W	56–2193
	ANT-XXIX/2	January 2013	54.9° to 46°W	15–482
	PS103	January 2017	54.4° to 46.5°W	48–457
	PS117	January 2019	53.7° to 46.5°W	80–457
	Used range		53.3° to 46.5°W	105–457

regions and the depths below those that are typically reached by autonomous instruments. To this end we analyze measurements from the Weddell Sea that cover the whole water column down to the sea floor, taken by the same accurate method during the last three decades.

2. Data and methods

Focusing on the Weddell Sea and aiming at the assessment of thermohaline changes at the highest possible precision, we solely analyze ship-lowered CTD casts. All casts were taken during cruises of the research icebreaker FS *Polarstern*. For the analysis of long-term trends, we use only measurements that were repeatedly made at the same locations (Fig. 1), which were visited every few years beginning in 1989 (Table 1). This way we avoid as far as possible artifacts arising from gridding or spatial interpolation.

The CTD sensors were always calibrated before and after the cruise. In addition, salinity was measured with laboratory

salinometers during the cruises in water samples taken from the CTD rosette bottles and used for calibration of the sensor-derived salinities. The accuracies of the CTD measurements and the applied processing procedures (sensor calibration, data validation, etc.) have been described in earlier work using data from these transects (Fahrback et al. 2004, 2011) and in detail in Driemel et al. (2017). Provided therein are also quality codes for each cruise dataset.

The final calibrated salinities are presented on the Practical Salinity Scale (no units). For measurements of temperature, we use the unit degrees Celsius (°C) and for changes of temperature the unit kelvin (K). Regarding the vertical coordinate, we use depth (in meters) for the sake of readability of the results, applying the approximation $1 \text{ m} \approx 1 \text{ dbar}$ of pressure.

For ease of comparison with the literature we conduct our ocean warming analysis in various depth layers that correspond to those used in previous studies, such as the top 700 and 2000 m. The 700-m depth contour moreover coarsely separates

FIG. 2. Vertical distribution of water masses across the Weddell Gyre. A vertical section of potential temperature along the prime meridian is shown between 70° and 55°S, depicting the distribution of the different major water masses: the Warm Deep Water (WDW), the Weddell Sea Deep Water (WSDW), and the Weddell Sea Bottom Water (WSBW). The WDW is found between the 0°C isotherms; the lower −0.7°C isotherm defines the boundary between the WSDW and the WSBW. The shown temperature distribution represents the long-term mean obtained by averaging over the data collected during the 11 repeat transects indicated in Table 1. Horizontal dotted lines depict the separation of depth layers used in the analysis.

the water column in the Weddell Sea into two strata with different physical characteristics (Fig. 2): 1) an upper stratum comprising the surface mixed layer occupied by the cold Antarctic Surface Water (AASW), which is subject to strong seasonal variations, and the core of the advected WDW; and 2) a lower stratum, containing the less variable and only weakly stratified water masses WSDW and WSBW. In addition, the 700-m isobath roughly demarcates the interior Weddell Basin from the adjacent shelves (Fig. 1).

Since this paper focuses on multidecadal changes, and in order to smooth out seasonal to interannual variability, we use the following approach. Long-term trends of potential temperature T , salinity S , and neutral density γ^n in different layers at the locations shown in Fig. 1 are derived from linear regression of the time series data. For sections N, E, and W and location n, the time period considered is 1989–2019, whereas for locations e and ω this is 1989–2011. Data along sections are averaged in layers before linear regression.

After estimation of the temperature and salinity trends in each layer and determination of the layers' horizontal areas A and volumes V , defined by bottom topography as illustrated in Fig. 1, their mean densities ρ and specific heat coefficients (c_p), changes of heat content Q (J) and subsequently the heating rate q (W m^{-2}) are calculated according to

$$Q = V\rho c_p \Delta T \text{ (J)}, \quad (1)$$

and

$$q = \frac{V}{A} \rho c_p \frac{dT}{dt} \text{ (W m}^{-2}\text{)}, \quad (2)$$

where ΔT denotes the temperature change over the observation period and dT/dt is the temperature trend. The layer areas A are defined in two alternative ways. In case of layers that extend from the surface down to a certain depth, it is given by the area encircled by that lower depth level, following the bottom topography. For the deeper layers that extend from 700 m or from deeper toward the sea floor, the area is defined by bathymetry at their upper depth level. The northern and eastern area limits are given by a straight line running from the South Orkney plateau to 55°S, 0°E and the prime meridian section N–E, respectively (Fig. 1).

The errors in our estimates of heating rates are calculated from propagation of individual errors. The individual errors in the trends of temperature, salinity, and density and in specific heat coefficients are determined as errors of the means (standard deviation divided by $\sqrt{6}$) obtained from the three sections and three locations. For the individual errors in volumes and areas, we assume 10% uncertainty.

To reveal the patterns of long-term change, we evaluate how the multidecadal temperature variations are distributed in the vertical–horizontal plane along the sections. To eliminate as much as possible the seasonal and interannual variability, we proceed as follows. Analogous to the calculation of the temperature and salinity trends in certain layers along a section, trends are determined by linear regression on a grid of 5-km horizontal and 50-m vertical spacing along a section to detect local differences in warming over the period of observations. These trends are then superimposed to the mean temperature distributions to identify the changes during the covered time period of 22 years between the first year, 1992 or 1989, and the last year, 2014 or 2011, for the prime meridian or Weddell cross

section, respectively. Trends in density, calculated from temperature, salinity, and pressure, are further exploited to determine the differences in geostrophic current shears between the beginning and the end of the observation period. As reference for calculating the geostrophic shear we assume zero velocity close to the sea floor. This assumption is based on knowledge of the deep-reaching current structure of the Weddell Gyre (e.g., [Fahrbach et al. 1994](#); [Reeve et al. 2019](#)). Across the ACC through Drake Passage, [García et al. \(2002\)](#) obtained a geostrophic flow field by assuming a reference velocity of zero at the sea floor, which was very close to transports assessed by other methods.

In search of mechanistic explanations for the observed changes we also assess the vertical stability of the upper water column by calculating the squared Brunt–Väisälä frequency:

$$N^2(\rho) = -\frac{g}{\rho} \frac{\partial \rho}{\partial z}, \quad (3)$$

where g is gravity and z the vertical coordinate. To separate the thermal $N^2(T)$ and haline $N^2(S)$ contributions to stability (see, e.g., [Strass and Nöthig 1996](#)), we split into the terms using the thermal and haline expansion coefficients. As above, we calculate $N^2(\rho)$ as well as $N^2(T)$ and $N^2(S)$ at the beginning and the end of our observational period from reconstructed density, temperature, and salinity fields, respectively, obtained from superimposing their long-term trends (as determined by linear regression of the time series records) on their mean distributions.

3. Results

a. Long-term trends and rates

[Table 2](#) shows the obtained trends together with their upper and lower bounds as well as the correlation coefficients R^2 . Within all depth layers below 700 m at all three stations (n- ω) and sections (N-W) shown in [Fig. 1](#), long-term (over up to 30 years) trends of increasing temperatures were found. The trends were determined by linear regression, which explains between 20% and 92% of the variance. Temperature trends in the top 700 m, in contrast, are not significantly different from zero, most likely because the changes in this layer are dominated by intra- and interannual variability and not by multiannual trends. [Figure 3](#) displays the temperature time series and trends for the water column above and below 700 m.

For all layers defined below 700 m, almost all of the temperature trends (i.e., the slopes of linear regressions) determined for the individual stations and sections are not significantly different from each other ([Table 2](#)). The mean rates of temperature increase below 700 m vary between 2.1 and 2.4 mK a⁻¹ ([Table 3](#)). For layers extending from the surface to greater depths (i.e., to 2000 or 3000 m or the sea floor), the mean temperature increases are in the range 1.8–2.2 mK a⁻¹. Based on the hence justified assumption of spatially uniform warming of each layer, we calculated the heat content change and the heating rate of each volume according to Eqs. (1) and (2). The results are shown in [Fig. 4](#), as well as in [Table 3](#) together with their uncertainties.

For the entire deep layer (i.e., from 700-m depth to the sea-floor), we obtain a heat content change Q of $2.37 \pm 0.47 \times 10^{21}$ J.

The heating rate of the layer from 700 m to bottom we estimate as 0.92 ± 0.20 W m⁻². For the entire water column, from the surface to bottom, the Weddell Sea heating rate is estimated to 1.04 ± 0.28 W m⁻².

The significant warming of the Weddell Sea below 700 m gives rise to the question how density is affected. Possible density changes are critical because the WDSW and the WSBW represent precursors of the AABW, and thus drive the lower limb of the global ocean overturning circulation. To assess density changes we first look into shifts in salinity.

Changes in salinity observed at our repeat stations and sections are highly variable between locations and over time ([Fig. 5](#)). Long-term multidecadal salinity trends at the different locations determined in the same layers as before for temperature are in the range between -1.5×10^{-3} and 0.2×10^{-3} a⁻¹. In the top 700 m the trends are nowhere significantly different from zero, although the overall tendency points to freshening; averaged over all six sections and stations, the freshening trend in the upper 700 m amounts to $-0.43 \pm 0.22 \times 10^{-3}$ a⁻¹ ([Table 4](#)). In the water column below 700 m, in contrast, trends that are significantly different from zero (4 out of 6) indicate an increase in salinity; averaged over all six sections and stations, the salinification trend in the deep water layers below 700 m is determined to $0.07 \pm 0.03 \times 10^{-3}$ a⁻¹. Mean salinification trends are determined also individually for all the different layers below 700 m ([Table 4](#)). However, freshening trends also exist below 700 m in two out of the six analyzed time series but are not significantly different from zero ([Fig. 5d](#)). If the upper 700 m are included in the calculation of mean trends, freshening dominates for all layers. Looking again at salinity changes at the different repeat sections and stations, it is interesting to note that—while statistically not significantly different from zero—the strongest freshening trends above and below 700-m depth are observed in the western outflow regime of the Weddell Gyre (section W in [Figs. 1](#) and [5](#)). A dominance of salinification and the same distribution pattern of freshening/salinification between repeat stations and sections as below 700 m is also found in the deepest layers, from 3000 m to bottom or from 4000 m to bottom (not shown here).

As a consequence of the changes in temperature and salinity, density decreases for almost all layers and locations considered in the Weddell Sea ([Fig. 6](#)). Of the overall 78 multidecadal trends determined by linear regression in the various layers and at the different locations, only one in the top 700 m is positive (but statistically insignificant); all others are negative, indicating density decrease ([Table 2](#) and [Fig. 6](#)). The overall mean density trend in the upper 700 m is determined to be $-0.75 \pm 0.32 \times 10^{-3}$ kg m⁻³ a⁻¹ and in the deeper layer from 700 m to the bottom of the Weddell Sea to be $-0.43 \pm 0.09 \times 10^{-3}$ kg m⁻³ a⁻¹. Density decreases also in the deepest layers, from 3000 m to bottom or from 4000 m to bottom, which contain the densest water masses. The uniform decrease in density indicates that density rather follows the uniform warming trend in temperature than the more variable salinity changes. Locally, the strongest decrease in density is observed in the Weddell outflow regime, along the section down the continental slope of the Antarctic Peninsula (section W) and at

TABLE 2. Long-term trends of potential temperature T , salinity S , and neutral density γ^n in different layers at the locations (indicated in the first column) shown in Fig. 1. For sections N, E, W, and location n, the time period considered is 1989–2019, whereas for locations e and ω this is 1989–2011 (see Table 1). The trends are derived from linear regression of the time series data and are indicated together with their lower and upper bounds of the 95% confidence interval and the squared correlation coefficient R^2 .

Location	Depth range (m)	T (mK a ⁻¹)	Lower (mK a ⁻¹)	Upper (mK a ⁻¹)	R^2	$S \times 10^3$ (a ⁻¹)	Lower (a ⁻¹)	Upper (a ⁻¹)	R^2	γ^n (g m ⁻³ a ⁻¹)	Lower (g m ⁻³ a ⁻¹)	Upper (g m ⁻³ a ⁻¹)	R^2
n	0–700	-3.00	-9.74	3.75	0.06	-0.20	-1.26	0.85	0.01	-0.90	-2.63	0.83	0.08
n	0–2000	0.29	-2.17	2.74	0.00	-0.02	-0.38	0.35	0.00	-0.56	-1.28	0.16	0.17
n	0–3000	0.72	-0.96	2.41	0.06	0.00	-0.24	0.24	0.00	-0.50	-1.01	0.02	0.24
n	0–bottom	1.20	0.20	2.21	0.32	0.01	-0.13	0.15	0.00	-0.46	-0.78	-0.15	0.41
n	700–2000	2.04	1.06	3.02	0.59	0.09	0.05	0.13	0.66	-0.29	-0.51	-0.07	0.37
n	700–3000	1.84	1.04	2.64	0.63	0.07	0.04	0.10	0.60	-0.30	-0.48	-0.12	0.48
n	700–4000	1.76	1.03	2.48	0.66	0.05	0.02	0.08	0.49	-0.31	-0.48	-0.15	0.54
n	700–bottom	1.81	1.24	2.39	0.76	0.04	0.01	0.08	0.34	-0.35	-0.48	-0.22	0.70
n	1000–4000	1.72	1.02	2.41	0.67	0.05	0.02	0.08	0.44	-0.32	-0.48	-0.16	0.57
n	2000–4000	1.57	0.97	2.17	0.69	0.03	-0.01	0.06	0.19	-0.32	-0.47	-0.18	0.63
n	3000–4000	1.56	0.95	2.17	0.68	0.02	-0.01	0.06	0.11	-0.33	-0.48	-0.19	0.64
n	3000–bottom	1.81	1.38	2.25	0.85	0.02	-0.03	0.07	0.06	-0.39	-0.50	-0.29	0.83
n	4000–bottom	2.00	1.67	2.34	0.92	0.02	-0.04	0.07	0.04	-0.44	-0.52	-0.35	0.90
N	0–700	-1.08	-11.51	9.36	0.01	-0.26	-0.93	0.40	0.08	-0.41	-1.82	0.99	0.05
N	0–2000	1.75	-2.24	5.73	0.10	0.03	-0.22	0.29	0.01	-0.39	-1.19	0.42	0.11
N	0–3000	1.89	-0.98	4.76	0.20	0.07	-0.11	0.26	0.09	-0.35	-0.98	0.29	0.15
N	0–bottom	2.01	0.04	3.98	0.37	0.09	-0.05	0.23	0.19	-0.36	-0.82	0.10	0.26
N	700–2000	3.27	0.57	5.97	0.45	0.19	0.10	0.29	0.71	-0.37	-0.93	0.19	0.20
N	700–3000	2.79	0.63	4.96	0.49	0.18	0.09	0.26	0.72	-0.33	-0.78	0.12	0.24
N	700–4000	2.60	0.74	4.46	0.53	0.16	0.08	0.24	0.70	-0.33	-0.71	0.05	0.30
N	700–bottom	2.56	1.00	4.13	0.60	0.15	0.07	0.23	0.69	-0.35	-0.67	-0.02	0.40
N	1000–4000	2.52	0.77	4.27	0.54	0.16	0.08	0.24	0.70	-0.33	-0.68	0.03	0.33
N	2000–4000	2.10	0.82	3.37	0.61	0.14	0.07	0.21	0.68	-0.29	-0.55	-0.03	0.41
N	3000–4000	1.95	0.93	2.96	0.68	0.12	0.05	0.20	0.61	-0.30	-0.51	-0.08	0.52
N	3000–bottom	2.16	1.39	2.93	0.82	0.12	0.03	0.20	0.52	-0.36	-0.53	-0.20	0.73
N	4000–bottom	2.55	1.75	3.36	0.85	0.11	0.01	0.21	0.39	-0.47	-0.65	-0.30	0.80
e	0–700	1.34	-7.42	10.09	0.01	0.06	-1.15	1.27	0.00	-0.11	-1.14	0.92	0.01
e	0–2000	1.59	-1.75	4.93	0.09	0.05	-0.35	0.44	0.01	-0.27	-0.58	0.05	0.24
e	0–3000	1.53	-0.86	3.91	0.15	0.03	-0.22	0.28	0.01	-0.30	-0.49	-0.11	0.53
e	0–bottom	1.61	-0.10	3.31	0.28	0.01	-0.14	0.16	0.00	-0.35	-0.50	-0.20	0.71
e	700–2000	1.76	0.74	2.78	0.57	0.04	-0.05	0.13	0.09	-0.34	-0.57	-0.11	0.48
e	700–3000	1.61	0.78	2.45	0.62	0.02	-0.07	0.11	0.02	-0.35	-0.57	-0.13	0.52
e	700–4000	1.60	0.85	2.35	0.67	0.01	-0.08	0.10	0.00	-0.36	-0.56	-0.15	0.57
e	700–bottom	1.67	0.78	2.55	0.61	0.00	-0.10	0.10	0.00	-0.38	-0.59	-0.18	0.60
e	1000–4000	1.57	0.86	2.29	0.68	0.00	-0.09	0.10	0.00	-0.37	-0.58	-0.16	0.58
e	2000–4000	1.49	0.88	2.10	0.72	-0.01	-0.11	0.08	0.01	-0.37	-0.57	-0.17	0.61
e	3000–4000	1.56	0.97	2.15	0.76	-0.02	-0.12	0.09	0.01	-0.38	-0.56	-0.20	0.66
e	3000–bottom	1.71	0.74	2.68	0.58	-0.02	-0.15	0.10	0.02	-0.42	-0.62	-0.22	0.67

TABLE 2. (Continued)

Location	Depth range (m)	T (mK a ⁻¹)	Lower (mK a ⁻¹)	Upper (mK a ⁻¹)	R^2	$S \times$ 10^3 (a ⁻¹)	Lower (a ⁻¹)	Upper (a ⁻¹)	R^2	γ^n (g m ⁻³ a ⁻¹)	Lower (g m ⁻³ a ⁻¹)	Upper (g m ⁻³ a ⁻¹)	R^2
e	4000–bottom	1.86	0.43	3.29	0.43	-0.03	-0.20	0.13	0.02	-0.46	-0.69	-0.23	0.63
E	0–700	-2.28	-10.27	5.70	0.04	-0.10	-0.58	0.37	0.03	0.18	-0.42	0.79	0.05
E	0–2000	0.62	-2.86	4.10	0.02	0.10	-0.08	0.27	0.14	-0.01	-0.41	0.40	0.00
E	0–3000	1.13	-1.44	3.70	0.10	0.11	-0.01	0.23	0.31	-0.10	-0.45	0.25	0.04
E	0–bottom	1.26	-0.72	3.24	0.19	0.11	0.02	0.20	0.46	-0.14	-0.44	0.17	0.11
E	700–2000	2.18	0.57	3.79	0.51	0.20	0.13	0.27	0.82	-0.11	-0.47	0.25	0.05
E	700–3000	2.17	0.79	3.55	0.59	0.17	0.11	0.24	0.79	-0.18	-0.50	0.14	0.15
E	700–4000	2.00	0.80	3.20	0.61	0.15	0.08	0.22	0.74	-0.20	-0.49	0.10	0.20
E	700–bottom	1.95	0.82	3.08	0.63	0.14	0.07	0.21	0.71	-0.21	-0.48	0.07	0.25
E	1000–4000	2.09	0.96	3.23	0.66	0.15	0.08	0.22	0.73	-0.23	-0.52	0.06	0.27
E	2000–4000	1.96	1.00	2.92	0.70	0.12	0.05	0.19	0.61	-0.28	-0.53	-0.02	0.39
E	3000–4000	1.90	1.06	2.74	0.74	0.10	0.02	0.18	0.49	-0.30	-0.54	-0.07	0.48
E	3000–bottom	1.88	1.04	2.73	0.74	0.11	0.02	0.19	0.46	-0.31	-0.53	-0.09	0.52
E	4000–bottom	1.77	0.82	2.72	0.67	0.11	0.00	0.21	0.38	-0.29	-0.51	-0.07	0.51
ω	0–700	4.05	-2.11	10.20	0.16	-0.67	-1.56	0.23	0.20	-1.35	-2.42	-0.29	0.42
ω	0–2000	3.51	0.82	6.20	0.43	-0.15	-0.41	0.12	0.12	-0.85	-1.26	-0.43	0.65
ω	0–3000	2.97	1.06	4.89	0.52	-0.08	-0.24	0.08	0.11	-0.71	-1.02	-0.40	0.70
ω	0–bottom	4.40	2.48	6.33	0.70	-0.02	-0.11	0.07	0.03	-1.07	-1.63	-0.50	0.61
ω	700–2000	3.23	2.05	4.40	0.77	0.14	0.02	0.26	0.37	-0.53	-0.78	-0.29	0.68
ω	700–3000	2.64	1.79	3.50	0.81	0.09	-0.01	0.20	0.28	-0.48	-0.71	-0.24	0.65
ω	700–4000	3.01	2.15	3.87	0.85	0.09	0.00	0.18	0.30	-0.58	-0.86	-0.30	0.66
ω	700–bottom	4.34	2.26	6.42	0.66	0.14	0.04	0.24	0.47	-0.84	-1.34	-0.35	0.56
ω	1000–4000	2.85	2.04	3.65	0.85	0.08	-0.01	0.17	0.25	-0.57	-0.84	-0.29	0.66
ω	2000–4000	2.34	1.69	2.98	0.85	0.04	-0.05	0.14	0.09	-0.50	-0.76	-0.25	0.63
ω	3000–4000	2.24	1.66	2.82	0.87	0.03	-0.07	0.14	0.04	-0.48	-0.72	-0.25	0.65
ω	3000–bottom	3.89	2.12	5.66	0.68	0.11	0.01	0.21	0.37	-0.78	-1.19	-0.36	0.60
ω	4000–bottom	3.18	2.33	4.04	0.87	0.15	0.03	0.26	0.44	-0.57	-0.81	-0.33	0.73
W	0–700	3.06	-4.19	10.30	0.09	-1.42	-2.85	0.01	0.36	-1.92	-3.45	-0.39	0.47
W	0–2000	2.68	-0.49	5.85	0.29	-0.55	-1.11	0.02	0.35	-1.05	-1.74	-0.36	0.57
W	0–3000	2.34	-0.08	4.75	0.35	-0.42	-0.84	0.00	0.36	-0.90	-1.42	-0.39	0.64
W	0–bottom	2.05	-0.05	4.16	0.35	-0.37	-0.73	-0.01	0.37	-0.80	-1.24	-0.36	0.65
W	700–2000	2.42	1.46	3.38	0.78	0.02	-0.06	0.09	0.03	-0.49	-0.71	-0.26	0.72
W	700–3000	2.04	1.38	2.69	0.85	-0.02	-0.09	0.04	0.05	-0.49	-0.68	-0.31	0.80
W	700–4000	1.76	1.13	2.38	0.82	-0.04	-0.10	0.03	0.14	-0.45	-0.63	-0.28	0.79
W	700–bottom	1.72	1.08	2.37	0.80	-0.04	-0.10	0.03	0.14	-0.45	-0.63	-0.27	0.78
W	1000–4000	1.59	1.05	2.14	0.83	-0.05	-0.11	0.02	0.24	-0.46	-0.62	-0.29	0.81
W	2000–4000	1.12	0.39	1.85	0.57	-0.09	-0.16	-0.02	0.46	-0.43	-0.61	-0.25	0.77
W	3000–4000	0.65	-0.32	1.62	0.20	-0.10	-0.18	-0.01	0.43	-0.30	-0.49	-0.11	0.59
W	3000–bottom	0.74	-0.20	1.67	0.26	-0.09	-0.18	-0.01	0.41	-0.31	-0.49	-0.13	0.63
W	4000–bottom	1.72	0.00	3.44	0.36	-0.04	-0.16	0.09	0.05	-0.43	-0.73	-0.14	0.55

FIG. 3. Temperature records and trends. Long-term records of potential temperature averaged over (a),(b) the top 700 m and (c),(d) the deep-water column from 700 m to the bottom and their temporal trends. Solid lines in (a) and (c) show mean temperature records composed of CTD sections, dashed lines temperature records obtained from repeat CTD stations; colors and lettering refer to Fig. 1, which shows the section and station positions in a map; the black dashed lines in (a) and (c) represent the trend lines obtained from linear regressions. Red dots and associated vertical black bars in (b) and (d) show the temperature increases per year and corresponding 95% confidence intervals obtained from the linear regressions applied to the six temperature records, which are indicated on the x axes of (b) and (d) by the same colored letters as in (a) and (c). The solid and dashed blue lines in (b) and (d) represent the overall mean temperature trend and standard error of the mean trend after averaging over all six records.

station ω (Figs. 1 and 6). While the above-average density decrease along section W results mainly from freshening, at station ω it is dominantly driven by strong warming (cf. Figs. 3, 5, and 6).

b. Patterns of change

Following the quantification of thermohaline changes as averages over major layers, we subsequently look into the distribution of the multidecadal trends on smaller horizontal

TABLE 3. Changes of temperature and of ocean heat content Q as well as energy flux q in different layers with their mean densities and specific values used in the calculation. The areas and volumes of layers that extend from the surface to a given depth are defined by the bathymetric contour that beigns the Weddell Basin in the south and west at that lower depth. Areas and volumes of layers that extend down from a given depth to the sea floor in contrast are defined by the topography at their upper surface. The northern and eastern boundaries of all areas are indicated by the gray shading in Fig. 1.

Layer (m)	Area ($\text{m}^2 \times 10^{12}$)	Volume ($\text{m}^3 \times 10^{15}$)	Density (kg m^{-3})	Specific heat ($\text{J k}^{-1} \text{K}^{-1}$)	Temperature trend (mK a^{-1})	Q ($\text{J} \times 10^{21}$)	Q error ($\text{J} \times 10^{21}$)	q (W m^{-2})	q error (W m^{-2})	Error (%)
0–700	4.06	2.84	1029.4	3977	0.51	0.12	0.30	0.05	0.12	255
700–bottom	4.06	12.09	1037.7	3917	2.41	2.37	0.47	0.92	0.20	22
0–2000	3.75	7.50	1032.4	3956	1.81	1.11	0.35	0.47	0.15	33
0–3000	3.37	10.11	1034.5	3943	1.84	1.15	0.34	0.71	0.18	25
0–bottom	4.06	14.93	1036.8	3928	2.18	2.65	0.65	1.04	0.28	27
2000–bottom	3.75	6.41	1040.5	3912	2.04	1.06	0.21	0.45	0.10	23
3000–bottom	3.37	2.53	1044.7	3881	2.15	0.44	0.09	0.21	0.05	22
4000–bottom	2.45	0.39	1047.4	3866	2.06	0.06	0.01	0.04	0.01	21

FIG. 4. Graphical summary of heating rates. The heating rates (W m^{-2}) determined for the various layers are displayed by the black boldface numbers within the bluish cells that represent the various layers; the numbers in red indicate the heating rates scaled by a factor of 0.71, the ratio of the global ocean to whole Earth surface area. The bottom topography is indicated in dark gray, referring for example to the western continental slope of the Weddell Basin along the green-colored section shown in Fig. 1. (a) Layers above and below 700 m; as the reference depth of the layer below 700 m that determines this layer's horizontal area the mean water depth of the Weddell Basin, 3680 m, is taken. (b) Layers extending downward from the surface to 2000 m, 3000 m, and to the bottom. (c) Layers extending upward from the sea floor to 4000, 3000, and 2000 m. The numerical values of the layer areas, volumes, mean densities and specific heats, temperature trends, heat content changes, and heating rates together with their errors are given in Table 3.

and vertical scales along sections (Fig. 1). The results regarding the distribution of long-term warming and cooling along the prime meridian and the Weddell Sea cross section are shown in Figs. 7 and 8.

Common to both sections (Figs. 7 and 8) is the large variability in the top few hundreds of meters, prominent through alternating patches of apparent cooling and warming with typical horizontal scales in the order of hundreds of kilometers. This high variability is in line with the result of the time series regression analysis that no trend significantly different from zero is found in the upper 700 m (Table 2 and Fig. 3). Figures 7a and 8a reveal that this variability is highest in the depth range 50–250 m, directly above the depth of the temperature maximum, which marks the core of the Warm Deep Water. As regards the Weddell Sea cross section (Fig. 8a), the temperature maximum depth moved down over the observation period by approximately 50 m. Hence, cooling dominates in approximately hundred meters above, particularly in the western half of the section. A comparable systematic vertical displacement of the temperature maximum depth is not identifiable along the prime meridian section (Fig. 7a). Here, vertical displacements of the temperature maximum are rather local, centered around 57° and 66°S and associated with exceptionally strong warming around 100-m depth.

Strikingly, the bands of enhanced warming centered around 57° and 66°S extend over full ocean depth, down to the sea floor

at those latitudes (Fig. 7b). Two other vertically coherent bands of warming between the temperature maximum and the bottom are observed, one south of 68°S above the continental slope and another at 62°S in the center of the prime meridian section. Within these bands of warming, the highest temperature increase is found at the top of the WDW layer, which is embedded between isopycnals $\gamma'' = 28.00 \text{ kg m}^{-3}$ and $\gamma'' = 28.27 \text{ kg m}^{-3}$. Also evident is that the WDW layer warming is stronger in the northern band than in the southern ones.

Similar bands of enhanced warming, vertically coherent from below the temperature maximum depth to the sea floor, are also observed along the Weddell Sea cross section (Fig. 8). Here, too, these bands are located at the outer edges of the Weddell Gyre above the continental slopes on either side of the Weddell Basin. Notable in both sections is that the vertically coherent bands of enhanced warming occurred where the isopycnals migrated downward over the time of observations (Figs. 7b and 8b). The bands of warming, however, are accompanied by vertically coherent bands of reduced warming or even cooling in the upper 1500–2000 m on their deep basin side in both cases, giving the distribution of warming along the Weddell Sea cross section a symmetric appearance relative to the middle of the section.

Along both sections, prime meridian and Weddell, the vertically coherent bands of warming appear connected with each other near the sea floor by a bottom layer of enhanced warming.

FIG. 5. As in Fig. 3, but for salinity.

The warming thus affects also the WSDW below isopycnal $\gamma^n = 28.27 \text{ kg m}^{-3}$ and deeper the WSBW with densities higher than $\gamma^n = 28.40 \text{ kg m}^{-3}$, a water mass that does not have a direct connection to areas outside of Weddell Sea.

c. Possible causes of long-term variations

For examining the causes of the multidecadal thermohaline changes in the Weddell Sea, we investigate possible long-term

changes of both stratification and the gyre circulation as an indicator of advection. We start by analyzing the near-surface stratification, leaning on the assumption that changes in the regional atmosphere–ocean exchange of momentum, heat, and water would be reflected in it.

Figure 9, showing the trend-derived stratification for 1992 and for 2014 in the top 500 m along the prime meridian, does not reveal marked differences both in terms of the magnitudes

TABLE 4. Mean multidecadal trends of potential temperature T , salinity S , and neutral density γ^n in different layers. The means and their standard deviations (Std.) are determined from the six trends that have been derived from linear regression of the multidecadal time series. They are graphically displayed in the right-hand panels of Figs. 3, 5, and 6.

Layer	Temperature (K a^{-1})		Salinity (a^{-1})		γ^n ($\text{kg m}^{-3} \text{ a}^{-1}$)	
	Mean $\times 10^{-3}$	Std. $\times 10^{-3}$	Mean $\times 10^{-3}$	Std. $\times 10^{-3}$	Mean $\times 10^{-3}$	Std. $\times 10^{-3}$
0–700	0.51	1.30	−0.43	0.22	−0.75	0.32
0–2000	1.81	0.54	−0.09	0.10	−0.52	0.16
0–3000	1.84	0.37	−0.05	0.08	−0.48	0.12
0–bottom	2.18	0.49	−0.03	0.07	−0.53	0.14
700–2000	2.49	0.26	0.11	0.03	−0.36	0.06
700–3000	2.22	0.19	0.09	0.03	−0.36	0.05
700–4000	2.20	0.22	0.07	0.03	−0.37	0.05
700–bottom	2.41	0.41	0.07	0.03	−0.43	0.09
1000–4000	2.14	0.20	0.07	0.03	−0.38	0.05
2000–4000	1.88	0.13	0.04	0.04	−0.37	0.04
3000–4000	1.80	0.11	0.03	0.03	−0.35	0.03
3000–bottom	2.15	0.36	0.04	0.04	−0.43	0.07
4000–bottom	2.06	0.32	0.05	0.03	−0.44	0.04

FIG. 6. As in Figs. 3 and 5, but for density.

and spatial distribution over the observation period. The vertical $N^2(\rho)$ maximum, usually coincident with the base of the mixed layer, is located within the top 110 m of the water column and is dominated by the haline stratification, $N^2(S)$. Both at the beginning and at the end of our 22-yr-long observation period the $N^2(\rho)$ maximum is located nearer to the sea surface in the southern half of the section, approximately half as deep, than in the northern part, with a local minimum at 64°S in the center. The latter position coincides with the northern flank of Maud Rise, which is known for its influence on the horizontal distribution of sea ice, and hence also on the release of fresh-water during seasonal ice melt and thus on $N^2(S)$ (e.g., Cisewski et al. 2011). A closer look reveals a latitudinal shift of the zone of southward decreasing $N^2(\rho)$ and $N^2(S)$ depths north of Maud Rise between 1992 and 2014. In addition, at the northern end of the section we find a downward displacement of the $N^2(T)$ maximum, which however is several times smaller in magnitude than $N^2(S)$. The $N^2(T)$ maximum is located below

the $N^2(\rho)$ maximum, and thus is unlikely to have resulted from surface forcing. The depth distributions of $N^2(\rho)$, $N^2(T)$, and $N^2(S)$ along the Weddell Sea cross section (not shown) are quite similar, with even less change over the observation period than along the prime meridian.

For examining possible changes of the gyre circulation, the geostrophic shear relative to the sea floor across the prime meridian section at the beginning and end of the observation period is estimated. The distribution pattern of geostrophic flow (Fig. 10) reveals mostly eastward flow in the northern half and dominantly westward flow in the south. This is the pattern expected from knowledge of the Weddell Gyre mean circulation (schematically represented in Fig. 1). The gyre axis (i.e., the change of direction between the eastward flow in the northern limb of the gyre lean to the ACC and the westward flow in the southern limb) is expected to cross the prime meridian at approximately 61°S (Reeve et al. 2019). This is in rough agreement with the geostrophic flow component displayed

FIG. 7. Vertical distribution of long-term temperature changes along the prime meridian (sections N and E in Fig. 1). The temperature changes are derived by applying the trends calculated from the temperature time series to the time interval covered by the data. The changes shown are thus representative for the developments over the 22 years from 1992 to 2014. Bluish colors indicate cooling, reddish colors warming; the dashed line separates cooling from warming. Shown are (a) the upper 700 m and (b) the full depth range. Also indicated as continuous lines (1992 in black and 2014 in gray) are the depths of the temperature maximum in (a) and selected isopycnals (from surface to bottom: $\gamma^n = 28.00, 28.27, 28.32, 28.40 \text{ kg m}^{-3}$) in (b); $\gamma^n = 28.00$ distinguishes AASW from WDW, 28.27 WDW from the WSDW, and 28.40 WSDW from the WSBW; $\gamma^n = 28.32$ marks the upper bound of the very dense water masses that are confined to the Weddell regime. The rise of bottom topography between 64° and 67°S signifies the western flank of Maud Rise.

in Fig. 10, which indicates a few more flow reversals, though. The reversals above and south of Maud Rise are explained by topographic steering of the flow around this seamount (Cisewski et al. 2011), which however has its peak rather near 3°E than at the prime meridian. The narrow swift westward current at the southernmost end of the section near the continent signifies the inflow into the Weddell Sea brought about by the merging of the southern periphery of the Weddell Gyre and the Antarctic Coastal Current associated with the Antarctic Slope Front.

Differences of the geostrophic flow patterns between 1992 and 2014 (Figs. 10a,b) are visible in the width and partly also the strength of particular current bands. Notable is a southward widening by about 1° of latitude and strengthening of the band

FIG. 8. Vertical distribution of long-term temperature changes along the Weddell Sea cross section (solid and dashed green line in Fig. 1). Otherwise as in Fig. 7, but for the 22 years 1989–2011.

of eastward flow associated with the northern gyre limb. While three flow reversals between 60° and 62°S made assessment of the gyre axis in 1992 difficult, the axis is identified more clearly at 61°S in 2014. The strengthening of the eastward northern gyre limb at its southern flank is accompanied by a wider and stronger westward flow between 61°S and Maud Rise in 2014 compared to 1992. Also increased in width and strength is the westward flow south of 67°S above the Antarctic continental slope. Taken together, the changes of the geostrophic current perpendicular to the prime meridian section suggest an enhancement of the gyre strength from 1992 to 2014 that is related to a southward displacement of the northern eastward flowing limb and an increase of the westward flow in the southern limb.

Comparison of the geostrophic flow fields (Fig. 10) with the long-term temperature changes (Fig. 7) reveals the following. The widest band of strongest warming between 57° and 59°S coincides with the strongest eastward geostrophic current associated with the northern limb of the Weddell Gyre. A closer look reveals that the maximum warming occurs right at the northern edge of the strongest current, at the transition of the Weddell Gyre into the ACC. This coincidence of warming and velocity increase is also suggested by the slopes of isopycnals in Fig. 7b, with the warming maximum found where the steepest isopycnal slopes moved southward between 1992 and 2004. The second most pronounced band of surface to bottom warming, identified near Maud Rise, seems to be related to the

FIG. 9. Changes in stratification between 1992 and 2014 in the upper water column along the prime meridian (sections N and E in Fig. 1). As a measure of stratification, the squared Brunt–Väisälä frequency N^2 is displayed. Shown are (a),(d),(g) N^2 based on the vertical density profile, (b),(e),(h) the contribution of the vertical temperature profile to N^2 , and (c),(f),(i) the contribution of the vertical salinity profile to N^2 . The temperature, salinity, and density fields used to calculate N^2 , as well as its thermal and haline contributions, at the (left) beginning (1992) and (center) end (2014) of the time interval covered by the data are derived from superposition of the least squares trends to the long-term means. The continuous black lines in (a)–(f) and blue and red lines in (g)–(i) indicate the depths of the N^2 maxima. Note that different depth ranges are used: 500 m in (a)–(f) and, to make the small differences visible, 300 m in (g)–(i).

circulation around this topographic structure. A comparable band of strong warming, even stronger by magnitude but not reaching the surface, is found near the continental slope where the flow associated with the southern limb of the gyre and the Antarctic Slope Front/Coastal Current sets eastward. Those two warming bands are also associated with both a downward displacement and stronger tilt of the isopycnals (Fig. 7). The fourth band of warming (Fig. 7), vertically coherent from just below the temperature maximum to the sea floor, is located close to 62°S directly south of the gyre center, where the flow changed from varying around zero in 1992 to clearly westward, hence into the Weddell Sea, in 2014.

In contrast to the prime meridian section, the Weddell Sea cross section does not cut through the Weddell Gyre axis. Located south of the gyre axis, the section runs along the gyre's southern limb (Reeve et al. 2019), often parallel rather than perpendicular to streamlines. Also the flow in the inner Weddell Sea is rather barotropic, thus much less vertically sheared than at the prime meridian (Reeve et al. 2019). Moreover, the currents in the western outflow region (Fahrback et al. 1994; Thompson and Heywood 2008; Naveira Garabato et al. 2019) are known to be bottom-intensified. Because of all these reasons and also a lack of sufficient current measurements for an assessment of reference level velocities during the observation period, we refrain from presenting changes in geostrophic shear velocities across the Weddell Sea section.

4. Discussion

a. Thermohaline changes and heating rates

We have analyzed long-term hydrographic time series composed of high-quality shipborne measurements, which were conducted between 1989 and 2019 at six different repeat stations and sections distributed over the Weddell Sea. Part of this dataset was previously included in analyses of global ocean heat content (e.g., Purkey and Johnson 2010; Desbruyères et al. 2016, 2017). However, the data were not exploited in depth to add to the specific understanding of regional heat, stratification, and circulation changes in Weddell Sea. The present study extends until most recently the work on decadal-scale variations of water mass properties in the deep Weddell Sea started by Fahrback et al. in 2004. In addition to previous studies, we looked into the pattern of changes in the Weddell Sea, employing a new approach of superimposing the determined long-term trends on mean distributions. This approach was introduced to minimize influences of seasonal and interannual variability. It allowed to reconstruct distribution patterns at different instances of time, here the start and the end of the observation period, whereby the differences solely result from the multiannual trends.

In the top 700-m layer the long-term station/section-mean warming trend was not found to be statistically significantly different from zero. With a temperature increase of 0.51 mK a^{−1} it is also much smaller than the warming trend of 14 mK a^{−1} of the

FIG. 10. Vertical distributions of the eastward geostrophic flow across the prime meridian section relative to the sea floor. The density fields used to calculate the geostrophic shear are derived by extrapolation of density trends calculated from the temperature and salinity time series to the begin and end of the time interval covered by the data. The flow fields shown are calculated for (a) 1992 and (b) 2014. The dashed line indicates the zero-velocity contour. The rise of bottom topography between 64° and 67° S represents the western flank of Maud Rise.

global surface ocean during the observation period 1989–2018 (NOAA 2020). Such lack of warming of the Southern Ocean surface layer south of the ACC, compared to the warming farther north, has been noted previously (e.g., Armour et al. 2016).

Below 700-m depth, in contrast, the temperature records show a significant warming trend that has been approximated by linear regression at all locations and in all investigated layers. When viewed together, the temperature time series below 700 m do not provide evidence of a reduced warming during the so-called hiatus period 1998–2012. Neither do they reveal an accelerated warming during the 2000s as suggested by Cheng et al. (2019) using Argo float data in the upper 2000 m of the global ocean, nor do our temperature records support the suggestion (Desbruyères et al. 2016) of different trends before and after year 2000 in the whole deep Southern Ocean. Interesting to note, though, is that the warming rate of $2.41 \pm 0.41 \text{ mK a}^{-1}$ that we determined in the interior of the Weddell Sea below 700 m agrees with the $2.4 \pm 0.6 \text{ mK a}^{-1}$ found by Zenk (2019) from hydrographic stations repeated since 1991 at the sill of the Vema Channel in the South Atlantic, where AABW spills over from the Argentine Basin into the Brazil Basin farther north.

Using the observed warming rates (temperature increases), we have calculated the heat content changes in several depth layers and subsequently the corresponding heating rates of the layers. When comparing our heating rates with estimates from the literature, in particular those obtained from global ocean heat content changes, we have to take into account that the latter are frequently normalized to the surface area of the entire Earth. To be compatible with this convention, we need to multiply our heating rates by a factor of 0.71, which is the ratio of the global ocean surface area to that of the entire Earth.

Applying this scaling factor, the Weddell Sea surface-to-bottom heating rate of 1.04 W m^{-2} reduces to 0.74 W m^{-2} (Fig. 4), which is almost identical to (or, taking the uncertainties into account, not different from) the full-depth global ocean average of 0.71 W m^{-2} determined by Desbruyères et al. (2017). Also for the Argo depth range, surface to 2000 m, our Weddell Sea heating rate of 0.33 W m^{-2} falls in the range $0.020\text{--}0.65 \text{ W m}^{-2}$ of existing global ocean estimates (Levitus et al. 2012; Rhein et al. 2013; Johnson et al. 2016; Cheng et al. 2019; Zanna et al. 2019). However, when comparing the heating rate estimates for the surface and the deeper layers, striking differences between our values and global ocean averages become apparent. For the upper 700 m our heating rate of 0.04 W m^{-2} is nearly one order of magnitude smaller than the global average of 0.31 W m^{-2} (Desbruyères et al. 2017), and not significantly different from zero. The reverse holds for the deeper layer below 2000 m; here our Weddell Sea heating rate of 0.32 W m^{-2} is roughly 5 times larger than the global ocean average of 0.065 W m^{-2} (Desbruyères et al. 2016).

These results support the view that the Weddell Sea (and the Southern Ocean at large) makes a major contribution to the warming of the deep and abyssal global ocean (Purkey and Johnson 2010; Frölicher et al. 2015; Desbruyères et al. 2016, 2017; Durack et al. 2018; Sallée 2018). The importance for the global ocean results from the formation of AABW, for which the Weddell Sea is a major source region. [For recent reviews, see Purkey et al. (2018) and Vernet et al. (2019)]. The Southern Ocean, which makes up roughly 15% of the world's ocean surface area, is estimated to overproportionally contribute 67% of the heat content increase in the global ocean below 2000 m (Desbruyères et al. 2016).

Our observations of haline changes in the Weddell Sea partly differ from those reported from other Southern Ocean regions. Averaged over all our three repeat stations and three repeat sections, we diagnosed a freshening in the top 700 m and a salinification in the deeper layers. While the former agrees with most existing reports (e.g., Jacobs and Giulivi 2010; Swart et al. 2018), the latter is the opposite to the freshening of bottom water reported from the southern Indian and Pacific Oceans (Rintoul 2007; Menezes et al. 2017) and the Southern Ocean at large (Purkey and Johnson 2013). However, the salinification of the deep and bottom waters that we observed is not uniformly distributed over the Weddell Sea; at the western continental slope off the Antarctic Peninsula and associated with the northward flowing limb of the Weddell Gyre they are freshening (Table 2 and Fig. 5). This particular finding is in line with observations around the tip of the peninsula (Jullion et al. 2013), which document a decadal freshening of the Antarctic

Bottom Water that is exported from the Weddell Sea. Also in the top 700 m the freshening is not uniform (Hoppema et al. 2015).

b. Potential drivers of the observed multidecadal changes

1) AIR–SEA FLUXES

Could an altered air–sea flux related to a possible increasing air temperature in the Weddell Sea region be responsible for the observed multidecadal hydrographic changes? Direct heating of the ocean by the atmosphere is, however, hardly possible because of the climatological mean temperature difference between air and sea. On the eastern side of the Weddell Sea, at roughly the central Weddell Sea latitude, the long-term mean air temperature recorded at the Antarctic research base Neumayer (70°40'S, 8°16'W) is near -15.9°C , with a slight cooling trend of $-0.14^{\circ}\text{C decade}^{-1}$ (for the period 1981–2011; Klöwer et al. 2014). At the western side, at Marambio Base (64°14'S, 56°38'W), an annual mean of -8.1°C and a warming trend of $+0.21^{\circ}\text{C decade}^{-1}$ has been determined (for 1979–2018; Turner et al. 2020). These air temperatures are well below the coldest sea surface temperature set by the freezing point of approximately -1.85°C .

A second possible driver of the trends we report could be altered radiative fluxes related to, for instance, changes in sea ice coverage that could modify the heat loss from ocean to atmosphere. Gradual increases of Antarctic sea ice extent by approximately 1% per decade are documented, with a marked decrease only from 2015 to 2018 (Parkinson 2019). Any change in air–sea heat or buoyancy fluxes—the latter also taking into account the impact of sea ice formation and melting on salinity—should have manifested themselves in changes of the near-surface stratification. However, we did not find evidence for this. It has to be noted that the majority of our data have been collected during austral summer, and might therefore be biased toward shallow N^2 maximum depths. However, the two autumn expeditions (May 1996 and 1998) and the only winter expedition in June 1992 also reveal (not shown here) depths of the N^2 maximum varying around 100 m. The consequential conclusion that the multidecadal temperature changes are not driven by surface fluxes in the Weddell Sea region is further confirmed by our finding that the warming is limited to the deeper layers, while the temperature trend in upper 700 m is insignificantly different from zero. However, the accumulation of heat in the deeper layers could also be the result of reduced open ocean deep convection that may have occurred in episodes that are not necessarily captured by the times of our measurements. Therefore, we looked into temperature–salinity diagrams (not shown here) of our data for indications of long-term changes in deep vertical mixing. Such indications but were not found.

Left as a third possibility of air–sea flux driven deep-ocean warming is that increased amounts of heat enter the deep Weddell Sea by way of dense water sinking along the continental slope of the Antarctic Peninsula. Our analysis does not provide evidence to support this either. However, our rather broad vertical and in case of sections also horizontal averaging might mask changes that could occur in the rather thin

slope-following layer of newly formed bottom water. Another argument against heat transfer from the shelves into the interior by slope convection is that shelf waters rather cooled than warmed during the past 50 years (Azaneu et al. 2013).

The hypothesis of regional atmospheric forcing being the cause of the observed multidecadal thermohaline changes in the interior Weddell Sea hence has to be rejected. In this respect the Weddell Sea differs from more northerly latitude bands ($<60^{\circ}\text{S}$) of the Southern Ocean, for which detection and attribution studies (Armour et al. 2016; Swart et al. 2018) as well as climate models (Frölicher et al. 2015; Liu et al. 2018) show an uptake of anthropogenic heat by upwelled UCDW, which subsequently moves northward as surface water and then subducts around 45°S .

2) ADVECTION

Left as the most likely explanation for the multidecadal warming of the Weddell Sea interior is a change in the advection of heat. This is suggested by two pieces of evidence contained in the pattern of warming.

First, temperature increased most in the layer that contains WDW, which is a water mass advected into Weddell Sea, thereby representing the main heat source of the Weddell Sea (e.g., Reeve et al. 2019). Water masses in the Weddell Sea are given names that usually differ from those of their source water masses in the ACC. A way to trace changes in Weddell Sea water masses back to changes in their source water masses is to follow isopycnals. The WDW is derived from the CDW and enters the Weddell Sea from the ACC with the eastern limb of the Weddell Gyre. The CDW is composed of the Upper Circumpolar Deep Water (UCDW) and the Lower Circumpolar Deep Water (LCDW) (Gordon 1967; Whitworth and Nowlin 1987). Along the southern flank of the ACC north of the Weddell Gyre, the UCDW and LCDW are found in the neutral density range $27.7\text{--}28.2\text{ kg m}^{-3}$ (Strass et al. 2017). The transition between the LCDW and the AABW, which overrides the ACC to enter the adjacent deep ocean basins further north, occurs at a neutral density of 28.27 kg m^{-3} (Orsi et al. 1999). The LCDW core roughly occupies the neutral density range from 28.04 to 28.08 kg m^{-3} (Donnelly et al. 2017). The strongest warming we observed in the Weddell Sea thus occurred in the UCDW to LCDW density range (see Figs. 7 and 8). Argo float data analyses, limited however to the upper 2000 m, indicated a deep-reaching warming between 0.02 and $0.04^{\circ}\text{C decade}^{-1}$ around the most inclined isopycnals at the southern flank of the ACC in the neutral density range $27.4\text{--}28.0$ (Böning et al. 2008; Sallée 2018). These warming rates are comparable with those we observe in the interior Weddell Sea. The Argo-based studies, however, do not disclose any warming occurring at the deeper isopycnals below the 2000-m Argo float depth limit. When moving southward in the ACC and entering the Weddell Sea, the LCDW rises from below 1500 m to about 500-m depth (Donnelly et al. 2017) as the combined result of wind-driven Ekman suction and of overriding the locally formed denser WSDW and WSBW. Increased advection of LCDW, which accounts for the deep salinity maximum in the ACC, can also explain the observed salinification in major parts of the deep Weddell Sea interior (Figs. 5c,d). Intensified advection of CDW moreover means

more “older” water is advected, which is consistent with the multidecadal increasing trends in TCO_2 and nitrate but decreasing trends in dissolved oxygen observed in the Weddell Sea by van Heuven et al. (2014).

Second, the vertically coherent bands of strongest warming coincide with locally deepened isopycnals (Figs. 7 and 8) and corresponding changes in geostrophic current shear (Fig. 10), which indicate a strengthening of the gyre circulation. A trend of increasing Weddell Gyre strength in parallel with an intensified wind stress curl was also diagnosed from satellite-based sea surface height (SSH) data by Armitage et al. (2018) for the years 2011–16 (i.e., for at least part of our observation period). Such an increase in gyre strength may be associated with the documented increase of the southern annular mode (SAM) index during the last decades (Latif et al. 2017), which goes along with both an increase of the westerly wind stress in the latitude band of the ACC and a southward shift of the wind stress maximum by about three degrees of latitude on zonal average (Lin et al. 2018). According to Liau and Chao (2017), the positive SAM correlated with an acceleration of the zonal eastward flow by a few millimeters per second at the southern flank of the ACC, which merges with the northern boundary of the Weddell Gyre, during the period 2003–15. The southward shift of the most tilted isopycnals at the northern Weddell Gyre/ACC boundary by $1\text{--}2^\circ$ of latitude diagnosed here (Fig. 10) is in line with Böning et al.’s (2008) finding obtained from a circumpolar Argo float analysis. Such a southward shift that is not accompanied by an increase in isopycnal tilt suggests that energy put into the ocean by stronger westerly winds is converted to mesoscale eddy kinetic energy (Meredith and Hogg 2006; Böning et al. 2008). Enhanced eddy activity may have contributed to an increased flux of heat from the ACC into the Weddell Sea (see also Fyfe et al. 2007).

The observed warming of the deep Weddell interior, of the densest AABW/WSDW class with $\gamma'' > 28.32 \text{ kg m}^{-3}$ that is not present in the ACC upstream of its confluence with the Weddell Gyre (Jullion et al. 2010), can only be explained by advection if it acts in concert with vertical mixing. This holds in particular for the WSBW with $\gamma'' > 28.40 \text{ kg m}^{-3}$ (Figs. 7, 8), which does not occur outside the Weddell Sea. Vertical mixing in the deep Weddell Sea is favored by a weak background stratification and through entrainment of surrounding water by newly formed dense water plumes during slope convection (Gordon et al. 1993; Fährbach et al. 1995). The flow of new bottom water along the steep topographic boundary generates submesoscale dynamical instabilities, which can provide an efficient mechanism for both vertical mixing and boundary–interior exchange (Naveira Garabato et al. 2019). If temperature increases in the deep interior by way of advection in the WDW/WSDW range as shown here, entrainment during slope convection can account for a transfer of heat into the bottom layer and explain the observed warming of the WSBW.

5. Conclusions

The persistent warming over up to 30 years seen in the interior Weddell Sea, which does not provide evidence of a stalled increase of temperatures during the 15-yr hiatus period

seen in surface data, confirms the importance of ocean observations that cover the full depth water column and also include the remote subpolar and polar regions. Sustained observations as those analyzed here contribute to tracking the energy budget of the Earth, and hence can prevent drawing the wrong conclusions regarding the response of the climate system to natural or anthropogenic forcing.

Noting that the Weddell Sea is one of the most severely undersampled oceanic regions and is therefore conjectured to be hiding potential surprises, it is remarkable that the surface-to-bottom heating rate of 0.74 W m^{-2} we determined is not significantly different from the full-depth global ocean average of 0.71 W m^{-2} (Desbruyères et al. 2017). This finding suggests that the range $0.5\text{--}0.9 \text{ W m}^{-2}$ (Wild 2017) of the Earth energy imbalance, obtained from combining direct observations taken at Earth’s surface and from space with climate model simulations and reanalyses, is valid and can probably be further narrowed down.

However, it is by far not certain that the Weddell Sea will transfer heat to the deep ocean at the same rate as during the investigated past three decades. The findings of the present study hint at two potential uncertainties of the future deep ocean heat uptake.

The first is related to the decrease of density of the WSDW and of the WSBW as predecessors of the AABW. Whether or not such density decrease will reduce the formation of AABW and hence the strength of the lower limb of the global ocean overturning circulation depends mainly on the contrast to the water masses overlying the AABW (e.g., Patara and Böning 2014). Observations of AABW in the Scotia Sea, the most direct pathway from the Weddell Sea to the Atlantic Ocean, indicate a recent recovery of the AABW supply to the Atlantic overturning circulation following a strong decline from the early 1990s to 2014 (Abrahamsen et al. 2019). The further development is hard to foresee and therefore needs continued measurement-based monitoring.

A second uncertainty results from the finding that the multidecadal warming of the deep Weddell Sea is possibly mediated by advection, which is driven by a southward shift and intensification of the Southern Ocean westerlies. The changes in the wind field over the Southern Ocean are coupled to the SAM. The future development of the SAM, though, is uncertain and a matter of scientific debate involving stratospheric ozone, greenhouse gases, and natural variability (e.g., Arblaster and Meehl 2006; Latif et al. 2017).

Continued increase of heat advection to the southern Weddell Sea would likely result in increased melting of the adjacent ice shelves, possibly at a rate higher than already predicted (Hellmer et al. 2017). Our observation that salinity has decreased in the outflow branch of the Weddell Gyre and in the newly formed bottom water, despite an overall salinity increase in the inflow regime, indicates an enhanced freshwater supply, for which increased ice shelf melting at the southern Weddell Sea periphery is a likely source. To assess the future contribution of Antarctic ice sheet melting to global sea level rise, sustained ocean observations in the Weddell Sea are indispensable.

Acknowledgments. We thank all those who contributed to the collection of our data: crews and captains of the research icebreaker *Polarstern*, technicians, engineers, and scientists. Basic funding was provided by the German Federal Minister of Education and Research (BMBF). In particular, we are grateful to the late Eberhard Fahrbach, who dedicated much of his professional lifetime to establishing and maintaining the Weddell Sea observation system, hence to creating the data base of this study. The constructive comments of three anonymous reviewers helped to improve the manuscript. All authors declare no conflicts of interest.

Data availability statement. The data are deposited and publicly available at <http://pangaea.de>.

REFERENCES

- Abraham, J. P., and Coauthors, 2013: A review of global ocean temperature observations: Implications for ocean heat content estimates and climate change. *Rev. Geophys.*, **51**, 450–483, <https://doi.org/10.1002/rog.20022>.
- Abrahamsen, E. P., and Coauthors, 2019: Stabilization of dense Antarctic water supply to the Atlantic Ocean overturning circulation. *Nat. Climate Change*, **9**, 742–746, <https://doi.org/10.1038/s41558-019-0561-2>.
- Arblaster, J. M., and G. A. Meehl, 2006: Contributions of external forcings to southern annular mode trends. *J. Climate*, **19**, 2896–2905, <https://doi.org/10.1175/JCLI3774.1>.
- Armitage, T. W. K., R. Kwok, A. F. Thompson, and G. Cunningham, 2018: Dynamic topography and sea level anomalies of the Southern Ocean: Variability and teleconnections. *J. Geophys. Res. Oceans*, **123**, 613–630, <https://doi.org/10.1002/2017JC013534>.
- Armour, K. C., J. Marshall, J. R. Scott, A. Donohoe, and E. R. Newsom, 2016: Southern Ocean warming delayed by circumpolar upwelling and equatorward transport. *Nat. Geosci.*, **9**, 549–554, <https://doi.org/10.1038/ngeo2731>.
- Azaneu, M., R. Kerr, M. M. Mata, and C. A. E. Garcia, 2013: Trends in the deep Southern Ocean (1958–2010): Implications for Antarctic Bottom Water properties and volume export. *J. Geophys. Res. Oceans*, **118**, 4213–4227, <https://doi.org/10.1002/jgrc.20303>.
- Balmaseda, M. A., K. E. Trenberth, and E. Källén, 2013: Distinctive climate signals in reanalysis of global ocean heat content. *Geophys. Res. Lett.*, **40**, 1754–1759, <https://doi.org/10.1002/grl.50382>.
- Böning, C. W., A. Dispert, M. Visbeck, S. R. Rintoul, and F. U. Schwarzkopf, 2008: The response of the Antarctic Circumpolar Current to recent climate change. *Nat. Geosci.*, **1**, 864–869, <https://doi.org/10.1038/ngeo362>.
- Chen, X., and K.-K. Tung, 2014: Varying planetary heat sink led to global-warming slowdown and acceleration. *Science*, **345**, 897–903, <https://doi.org/10.1126/science.1254937>.
- Cheng, L., J. Abraham, Z. Hausfather, and K. E. Trenberth, 2019: How fast are the oceans warming? *Science*, **363**, 128–129, <https://doi.org/10.1126/science.aav7619>.
- Cisewski, B., V. H. Strass, and H. Leach, 2011: Circulation and transport of water masses in the Lazarev Sea, Antarctica, during summer and winter 2006. *Deep-Sea Res. I*, **58**, 186–199, <https://doi.org/10.1016/j.dsr.2010.12.001>.
- Desbruyères, D. G., S. G. Purkey, E. L. McDonagh, G. C. Johnson, and B. A. King, 2016: Deep and abyssal ocean warming from 35 years of repeat hydrography. *Geophys. Res. Lett.*, **43**, 10 356–10 365, <https://doi.org/10.1002/2016GL070413>.
- , E. L. McDonagh, B. A. King, and V. Thierry, 2017: Global and full depth ocean temperature trends during the early twenty-first century from Argo and repeat hydrography. *J. Climate*, **30**, 1985–1997, <https://doi.org/10.1175/JCLI-D-16-0396.1>.
- Dessler, A. E., and P. M. Forster, 2018: An estimate of equilibrium climate sensitivity from interannual variability. *J. Geophys. Res. Atmos.*, **123**, 8634–8645, <https://doi.org/10.1029/2018JD028481>.
- Donnelly, M., H. Leach, and V. Strass, 2017: Modification of the deep salinity-maximum in the Southern Ocean by circulation in the Antarctic Circumpolar Current and the Weddell Gyre. *Ocean Dyn.*, **67**, 813–838, <https://doi.org/10.1007/s10236-017-1054-3>.
- Driemel, A., and Coauthors, 2017: From pole to pole: 33 years of physical oceanography onboard R/V *Polarstern*. *Earth Syst. Sci. Data*, **9**, 211–220, <https://doi.org/10.5194/essd-9-211-2017>.
- Durack, P. J., P. J. Gleckler, S. G. Purkey, G. C. Johnson, J. M. Lyman, and T. P. Boyer, 2018: Ocean warming: From the surface to the deep in observations and models. *Oceanography*, **31**, 41–51, <https://doi.org/10.5670/oceanog.2018.227>.
- Fahrbach, E., G. Rohardt, M. Schröder, and V. Strass, 1994: Transport and structure of the Weddell Gyre. *Ann. Geophys.*, **12**, 840–855, <https://doi.org/10.1007/s00585-994-0840-7>.
- , —, N. Scheele, M. Schröder, V. Strass, and A. Wisotzki, 1995: Formation and discharge of deep and bottom water in the northwestern Weddell Sea. *J. Mar. Res.*, **53**, 515–538, <https://doi.org/10.1357/0022240953213089>.
- , M. Hoppema, G. Rohardt, M. Schröder, and A. Wisotzki, 2004: Decadal-scale variations of water mass properties in the deep Weddell Sea. *Ocean Dyn.*, **54**, 77–91, <https://doi.org/10.1007/s10236-003-0082-3>.
- , —, —, O. Boebel, O. Klatt, and A. Wisotzki, 2011: Warming of deep and abyssal water masses along the Greenwich meridian on decadal time scales: The Weddell Gyre as a heat buffer. *Deep-Sea Res. II*, **58**, 2509–2523, <https://doi.org/10.1016/j.dsr2.2011.06.007>.
- Foster, T., and E. Carmack, 1976: Frontal zone mixing and Antarctic Bottom Water formation in the southern Weddell Sea. *Deep-Sea Res. Oceanogr. Abstr.*, **23**, 301–317, [https://doi.org/10.1016/0011-7471\(76\)90872-X](https://doi.org/10.1016/0011-7471(76)90872-X).
- Frölicher, T. L., J. L. Sarmiento, D. J. Paynter, J. P. Dunne, J. P. Krasting, and M. Winton, 2015: Dominance of the Southern Ocean in anthropogenic carbon and heat uptake in CMIP5 models. *J. Climate*, **28**, 862–886, <https://doi.org/10.1175/JCLI-D-14-00117.1>.
- Fyfe, J. C., O. A. Saenko, K. Zickfeld, M. Eby, and A. J. Weaver, 2007: The role of poleward-intensifying winds on Southern Ocean warming. *J. Climate*, **20**, 5391–5400, <https://doi.org/10.1175/2007JCLI1764.1>.
- , and Coauthors, 2016: Making sense of the early-2000s warming slowdown. *Nat. Climate Change*, **6**, 224–228, <https://doi.org/10.1038/nclimate2938>.
- García, M. A., I. Bladé, A. Cruzado, Z. Velásquez, H. García, J. Puigdefàbregas, and J. Sospedra, 2002: Observed variability of water properties and transports on the World Ocean Circulation Experiment SR1b section across the Antarctic Circumpolar Current. *J. Geophys. Res.*, **107**, 3162, <https://doi.org/10.1029/2000JC000277>.
- Gill, A. E., 1973: Circulation and bottom water production in the Weddell Sea. *Deep-Sea Res. Oceanogr. Abstr.*, **20**, 111–140, [https://doi.org/10.1016/0011-7471\(73\)90048-X](https://doi.org/10.1016/0011-7471(73)90048-X).

- GISTEMP Team, 2020: GISS Surface Temperature Analysis (GISTEMP), version 4. NASA Goddard Institute for Space Studies, accessed 20 March 2020, <https://data.giss.nasa.gov/gistemp/>.
- Gleckler, P. J., P. J. Durack, R. J. Stouffer, G. C. Johnson, and C. E. Forest, 2016: Industrial-era global ocean heat uptake doubles in recent decades. *Nat. Climate Change*, **6**, 394–398, <https://doi.org/10.1038/nclimate2915>.
- Gordon, A. L., 1967: Structure of Antarctic waters between 20°W and 170°W. *Antarctic Map Folio Series*, V. C. Bushnell, Ed., American Geographical Society, Folio 6, 14 plates, 10 pp.
- , B. A. Huber, H. Hellmer, and A. Ffield, 1993: Deep and bottom water of the Weddell Sea's western rim. *Science*, **262**, 95–97, <https://doi.org/10.1126/science.262.5130.95>.
- Hansen, J., M. Sato, P. Kharecha, and K. von Schuckmann, 2011: Earth's energy imbalance and implications. *Atmos. Chem. Phys.*, **11**, 13 421–13 449, <https://doi.org/10.5194/acp-11-13421-2011>.
- Hellmer, H. H., F. Kauker, R. Timmermann, and T. Hattermann, 2017: The fate of the southern Weddell Sea continental shelf in a warming climate. *J. Climate*, **30**, 4337–4350, <https://doi.org/10.1175/JCLI-D-16-0420.1>.
- Hoppema, M., K. Bakker, S. M. A. C. van Heuven, J. C. van Ooijen, and H. J. W. de Baar, 2015: Distributions, trends and inter-annual variability of nutrients along a repeat section through the Weddell Sea (1996–2011). *Mar. Chem.*, **177**, 545–553, <https://doi.org/10.1016/j.marchem.2015.08.007>.
- Jacobs, S. S., and C. F. Giulivi, 2010: Large multidecadal salinity trends near the Pacific–Antarctic continental margin. *J. Climate*, **23**, 4508–4524, <https://doi.org/10.1175/2010JCLI3284.1>.
- Johnson, G. C., J. M. Lyman, and N. G. Loeb, 2016: Improving estimates of Earth's energy imbalance. *Nat. Climate Change*, **6**, 639–640, <https://doi.org/10.1038/nclimate3043>.
- Jullion, L., S. C. Jones, A. C. Naveira Garabato, and M. P. Meredith, 2010: Wind-controlled export of Antarctic Bottom Water from the Weddell Sea. *Geophys. Res. Lett.*, **37**, L09609, <https://doi.org/10.1029/2010GL042822>.
- , A. C. Naveira Garabato, M. P. Meredith, P. R. Holland, P. Courtois, and B. A. King, 2013: Decadal freshening of the Antarctic Bottom Water exported from the Weddell Sea. *J. Climate*, **26**, 8111–8125, <https://doi.org/10.1175/JCLI-D-12-00765.1>.
- , and Coauthors, 2014: The contribution of the Weddell Gyre to the lower limb of the global overturning circulation. *J. Geophys. Res. Oceans*, **119**, 3357–3377, <https://doi.org/10.1002/2013JC009725>.
- Klöwer, M., T. Jung, G. König-Langlo, and T. Semmler, 2014: Aspects of weather parameters at Neumayer station, Antarctica, and their representation in reanalysis and climate model data. *Meteor. Z.*, **22**, 699–709, <https://doi.org/10.1127/0941-2948/2013/0505>.
- Kosaka, Y., and S.-P. Xie, 2013: Recent global-warming hiatus tied to equatorial Pacific surface cooling. *Nature*, **501**, 403–407, <https://doi.org/10.1038/nature12534>.
- Latif, M., T. Martin, A. Reintges, and W. Park, 2017: Southern Ocean decadal variability and predictability. *Curr. Climate Change Rep.*, **3**, 163–173, <https://doi.org/10.1007/s40641-017-0068-8>.
- Lee, S., W. Park, M. Baringer, A. L. Gordon, B. Huber, and Y. Liu, 2015: Pacific origin of the abrupt increase in Indian Ocean heat content during the warming hiatus. *Nat. Geosci.*, **8**, 445–449, <https://doi.org/10.1038/ngeo2438>.
- Lenssen, N., G. Schmidt, J. Hansen, M. Menne, A. Persin, R. Ruedy, and D. Zyss, 2019: Improvements in the GISTEMP uncertainty model. *J. Geophys. Res. Atmos.*, **124**, 6307–6326, <https://doi.org/10.1029/2018JD029522>.
- Levitus, S., J. Antonov, and T. Boyer, 2005: Warming of the world ocean, 1955–2003. *Geophys. Res. Lett.*, **32**, L02604, <https://doi.org/10.1029/2004GL021592>.
- , and Coauthors, 2012: World ocean heat content and thermohaline sea level change (0–2000 m), 1955–2010. *Geophys. Res. Lett.*, **39**, L10603, <https://doi.org/10.1029/2012GL051106>.
- Lewandowsky, S., J. S. Risbey, and N. Oreskes, 2016: The “pause” in global warming: Turning a routine fluctuation into a problem for science. *Bull. Amer. Meteor. Soc.*, **97**, 723–733, <https://doi.org/10.1175/BAMS-D-14-00106.1>.
- Lewis, N., and J. Curry, 2018: The impact of recent forcing and ocean heat uptake data on estimates of climate sensitivity. *J. Climate*, **31**, 6051–6071, <https://doi.org/10.1175/JCLI-D-17-0667.1>.
- Liau, J.-R., and B. F. Chao, 2017: Variation of Antarctic Circumpolar Current and its intensification in relation to the southern annular mode detected in the time-variable gravity signals by GRACE satellite. *Earth Planets Space*, **69**, 93, <https://doi.org/10.1186/s40623-017-0678-3>.
- Lin, X., X. Zhai, Z. Wang, and D. R. Munday, 2018: Mean, variability, and trend of Southern Ocean wind stress: Role of wind fluctuations. *J. Climate*, **31**, 3557–3573, <https://doi.org/10.1175/JCLI-D-17-0481.1>.
- Liu, W., J. Lu, J.-P. Xie, and A. V. Fedorov, 2018: Southern Ocean heat uptake, redistribution, and storage in a warming climate: The role of meridional overturning circulation. *J. Climate*, **31**, 4727–4743, <https://doi.org/10.1175/JCLI-D-17-0761.1>.
- Marshall, J., and K. Speer, 2012: Closure of the meridional overturning circulation through Southern Ocean upwelling. *Nat. Geosci.*, **5**, 171–180, <https://doi.org/10.1038/ngeo1391>.
- Medhaug, I., M. B. Stolpe, E. M. Fischer, and R. Knutti, 2017: Reconciling controversies about the ‘global warming hiatus.’ *Nature*, **545**, 41–47, <https://doi.org/10.1038/nature22315>.
- Menezes, V. V., A. M. Macdonald, and C. Schatzman, 2017: Accelerated freshening of Antarctic Bottom Water over the last decade in the southern Indian Ocean. *Sci. Adv.*, **3**, e1601426, <https://doi.org/10.1126/sciadv.1601426>.
- Meredith, M. P., and A. M. Hogg, 2006: Circumpolar response of Southern Ocean eddy activity to a change in the southern annular mode. *Geophys. Res. Lett.*, **33**, L16608, <https://doi.org/10.1029/2006GL026499>.
- Naveira Garabato, A. C., and Coauthors, 2019: Rapid mixing and exchange of deep-ocean waters in an abyssal boundary current. *Proc. Natl. Acad. Sci. USA*, **116**, 13 233–13 238, <https://doi.org/10.1073/pnas.1904087116>.
- NOAA, 2020: NOAA National Centers for Environmental Information. Climate at a glance: Global time series, accessed 20 March 2020, <https://www.ncdc.noaa.gov/cag/>.
- Orsi, A. H., W. D. Nowlin Jr., and T. Whitworth III, 1993: On the circulation and stratification of the Weddell Gyre. *Deep-Sea Res. I*, **40**, 169–203, [https://doi.org/10.1016/0967-0637\(93\)90060-G](https://doi.org/10.1016/0967-0637(93)90060-G).
- , G. C. Johnson, and J. L. Bullister, 1999: Circulation, mixing, and production of Antarctic Bottom Water. *Prog. Oceanogr.*, **43**, 55–109, [https://doi.org/10.1016/S0079-6611\(99\)00004-X](https://doi.org/10.1016/S0079-6611(99)00004-X).
- , W. M. Smethie, and J. L. Bullister, 2002: On the total input of Antarctic waters to the deep ocean: A preliminary estimate from chlorofluorocarbon measurements. *J. Geophys. Res.*, **107**, 3122, <https://doi.org/10.1029/2001JC000976>.
- Palmer, M. D., 2012: Climate and Earth's energy flows. *Surv. Geophys.*, **33**, 351–357, <https://doi.org/10.1007/s10712-011-9165-8>.

- , D. J. McNeill, and N. J. Dunstone, 2011: Importance of the deep ocean for estimating decadal changes in Earth's radiation balance. *Geophys. Res. Lett.*, **38**, L13707, <https://doi.org/10.1029/2011GL047835>.
- Parkinson, C. L., 2019: A 40-y record reveals gradual Antarctic sea ice increases followed by decreases at rates far exceeding the rates seen in the Arctic. *Proc. Natl. Acad. Sci. USA*, **116**, 14 414–14 423, <https://doi.org/10.1073/pnas.1906556116>.
- Patara, L., and C. W. Böning, 2014: Abyssal ocean warming around Antarctica strengthens the Atlantic overturning circulation. *Geophys. Res. Lett.*, **41**, 3972–3978, <https://doi.org/10.1002/2014GL059923>.
- Purkey, S. G., and G. C. Johnson, 2010: Warming of global abyssal and deep Southern Ocean waters between the 1990s and 2000s: Contributions to global heat and sea level rise budgets. *J. Climate*, **23**, 6336–6351, <https://doi.org/10.1175/2010JCLI3682.1>.
- , and —, 2013: Antarctic Bottom Water warming and freshening: Contributions to sea level rise, ocean freshwater budgets, and global heat gain. *J. Climate*, **26**, 6105–6122, <https://doi.org/10.1175/JCLI-D-12-00834.1>.
- , W. M. Smethie, G. Gebbie, A. L. Gordon, R. E. Sonnerup, M. J. Warner, and J. L. Bullister, 2018: A synoptic view of the ventilation and circulation of Antarctic Bottom Water from chloro-fluorocarbons and natural tracers. *Annu. Rev. Mar. Sci.*, **10**, 503–527, <https://doi.org/10.1146/annurev-marine-121916-063414>.
- Reeve, K., O. Boebel, T. Kanzow, V. Strass, G. Rohardt, and E. Fahrbach, 2016: A gridded data set of upper-ocean hydrographic properties in the Weddell Gyre obtained by objective mapping of Argo float measurements. *Earth Syst. Sci. Data*, **8**, 15–40, <https://doi.org/10.5194/essd-8-15-2016>.
- , —, V. Strass, T. Kanzow, and R. Gerdes, 2019: Horizontal circulation and volume transports in the Weddell Gyre derived from Argo float data. *Prog. Oceanogr.*, **175**, 263–283, <https://doi.org/10.1016/j.pocean.2019.04.006>.
- Rhein, M., and Coauthors, 2013: Observations: Ocean. *Climate Change 2013: The Physical Science Basis*. T. F. Stocker et al., Eds., Cambridge University Press, 255–315.
- Rintoul, S. R., 2007: Rapid freshening of Antarctic Bottom Water formed in the Indian and Pacific Oceans. *Geophys. Res. Lett.*, **34**, L06606, <https://doi.org/10.1029/2006GL028550>.
- Sallée, J.-B., 2018: Southern Ocean warming. *Oceanography*, **31**, 52–62, <https://doi.org/10.5670/oceanog.2018.215>.
- Schmitt, R. W., 2018: The ocean's role in climate. *Oceanography*, **31**, 32–40, <https://doi.org/10.5670/oceanog.2018.225>.
- Strass, V. H., and E. Nöthig, 1996: Seasonal shifts in ice edge phytoplankton blooms in the Barents Sea related to the water column stability. *Polar Biol.*, **16**, 409–422, <https://doi.org/10.1007/BF02390423>.
- , H. Leach, H. Prandke, M. Donnelly, A. U. Bracher, and D. A. Wolf-Gladrow, 2017: The physical environmental conditions for biogeochemical differences along the Antarctic Circumpolar Current in the Atlantic sector during late austral summer 2012. *Deep-Sea Res. II*, **138**, 6–25, <https://doi.org/10.1016/j.dsr2.2016.05.018>.
- Swart, N. C., S. T. Gille, J. C. Fyfe, and N. P. Gillett, 2018: Recent Southern Ocean warming and freshening driven by greenhouse gas emissions and ozone depletion. *Nat. Geosci.*, **11**, 836–841, <https://doi.org/10.1038/s41561-018-0226-1>.
- Talley, L. D., 1996: Antarctic intermediate water in the South Atlantic. *The South Atlantic: Present and Past Circulation*, G. Wefer et al., Eds., Springer-Verlag, 219–238.
- Thompson, A. F., and K. J. Heywood, 2008: Frontal structure and transport in the northwestern Weddell Sea. *Deep-Sea Res. I*, **55**, 1229–1251, <https://doi.org/10.1016/j.dsr.2008.06.001>.
- Trenberth, K. E., and J. T. Fasullo, 2012: Tracking Earth's energy: From El Niño to global warming. *Surv. Geophys.*, **33**, 413–426, <https://doi.org/10.1007/s10712-011-9150-2>.
- Turner, J., G. J. Marshall, K. Clem, S. Colwell, T. Phillips, and H. Lu, 2020: Antarctic temperature variability and change from station data. *Int. J. Climatol.*, **40**, 2986–3007, <https://doi.org/10.1002/joc.6378>.
- van Heuven, S. M. A. C., M. Hoppema, E. M. Jones, and H. J. W. de Baar, 2014: Rapid invasion of anthropogenic CO₂ into the deep circulation of the Weddell Gyre. *Philos. Trans. Roy. Soc.*, **A372**, 20130056, <https://doi.org/10.1098/rsta.2013.0056>.
- Vernet, M., and Coauthors, 2019: The Weddell Gyre, Southern Ocean: Present knowledge and future challenges. *Rev. Geophys.*, **57**, 623–708, <https://doi.org/10.1029/2018RG000604>.
- von Schuckmann, K., and Coauthors, 2016: An imperative to monitor Earth's energy imbalance. *Nat. Climate Change*, **6**, 138–144, <https://doi.org/10.1038/nclimate2876>.
- Whitworth, T., III, and W. D. Nowlin Jr., 1987: Water masses and currents of the Southern Ocean at the Greenwich meridian. *J. Geophys. Res.*, **92**, 6462–6476, <https://doi.org/10.1029/JC092iC06p06462>.
- Wild, M., 2017: Progress and challenges in the estimation of the global energy balance. *AIP Conf. Proc.*, **1810**, 020004, <https://doi.org/10.1063/1.4975500>.
- Zanna, L., S. Khatiwala, J. M. Gregory, J. Ison, and P. Heimbach, 2019: Global reconstruction of historical ocean heat storage and transport. *Proc. Natl. Acad. Sci. USA*, **116**, 1126–1131, <https://doi.org/10.1073/pnas.1808838115>.
- Zenk, W., 2019: Abyssal currents. *Encyclopedia of Ocean Sciences*, 3rd ed. J. K. Cochran, H. J. Bokuniewicz, and P. L. Yager, Eds., Elsevier, 189–203, <https://doi.org/10.1016/b978-0-12-409548-9.11343-0>.
- Zhang, H.-M., and Coauthors, 2019: Updated temperature data give a sharper view of climate trends. *Eos, Trans. Amer. Geophys. Union*, **100**, <https://doi.org/10.1029/2019EO128229>.