Southern Ocean forcing of the North Atlantic at multi-centennial time scales in the Kiel Climate Model

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A R T I C L E   I N F O

Available online 30 January 2014

Keywords:
Climate variability
Deep convection
Meridional ocean circulation
Bi-polar ocean seesaw

A B S T R A C T

Internal multi-centennial variability of open ocean deep convection in the Atlantic sector of the Southern Ocean impacts the strength of the Atlantic Meridional Overturning Circulation (AMOC) in the Kiel Climate Model. The northward extent of Antarctic Bottom Water (AABW) strongly depends on the state of Weddell Sea deep convection. The retreat of AABW results in an enhanced meridional density gradient that drives an increase in the strength and vertical extent of the North Atlantic Deep Water (NADW) cell. This shows, for instance, as a peak in AMOC strength at 30°N about a century after Weddell Sea deep convection has ceased. The stronger southward flow of NADW is compensated by an expansion of the North Atlantic subpolar gyre and an acceleration of the North Atlantic Current, indicating greater deep water formation. Contractions of the North Atlantic subpolar gyre enable warm water anomalies, which evolved in response to deep convection events in the Southern Ocean, to penetrate farther to the north, eventually weakening the AMOC and closing a quasi-centennial cycle.

Gyre contractions are accompanied by increases in sea level of up to 20 cm/century in some areas of the North Atlantic. In the Southern Ocean itself, the heat loss during the convective regime results in a sea surface height decrease on the order of 10 cm/century, with a maximum of 30 cm/century in the Weddell Sea. Hence, the impact of the Southern Ocean Centennial Variability (SOCV) on regional as well as North Atlantic sea level is of the same order of magnitude as the rise of global average sea level during the 20th century, which amounts to about 15–20 cm. This suggests that internal variability on a centennial time scale cannot be neglected a priori in assessments of 20th and 21st century AMOC and regional sea level change.

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1. Introduction

A teleconnection between the Southern Ocean and the North Atlantic referred to as the bi-polar ocean seesaw, involving asynchronous changes in the two regions, has been documented in many previous papers which dealt with climate fluctuations during the last glacial period and the transition to the current interglacial (e.g., Broecker, 1998; Blunier et al., 1998; Stocker, 1998; Seidov et al., 2001; Swingedouw et al., 2009). The bi-polar ocean seesaw is an oscillating meridional overturning regime driven by two deep water sources: the North Atlantic Deep Water (NADW) in the north, and Antarctic Bottom Water (AABW) in the south. A decrease (increase) in AABW production caused by surface salinity variations results in a retreating (advancing) bottom water mass in the deep Atlantic thereby increasing (decreasing) southward export of NADW, which consequently intensifies (reduces) NADW formation and northward surface currents in the Atlantic (Seidov et al., 2001). The AABW retreat results in an overall density decrease, first within a couple of decades in the South Atlantic, then also in the North Atlantic. The anomalous meridional density gradient caused by this delay is the driver of the enhanced NADW export (Swingedouw et al., 2009). Broecker (2001) hypothesized that such thermohaline circulation oscillations may be responsible for the 1500 year cycles in ice-rafted debris in the northern North Atlantic, which has been found by Bond et al. (1997).

The bi-polar ocean seesaw and related ocean circulation changes may be either internally driven through stochastic forcing by the atmosphere, or externally by e.g. low-frequency changes in solar radiation or volcanic activity (e.g., Otterå et al., 2010). In particular, surface freshening by meltwater input from the continental ice sheets lowering sea surface salinity is thought to have been one important driver of the bi-polar ocean seesaw. In contrast, the study by Toggweiler and Samuels (1980) and modeling results from Goosse and Fichefet (1999) stressed the

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0967-0645/$ – see front matter © 2014 Elsevier Ltd. All rights reserved.
http://dx.doi.org/10.1016/j.dsr2.2014.01.018
importance of increasing sea surface salinity, e.g. by brine rejection during sea ice formation, for Atlantic Meridional Overturning Circulation (AMOC) variability. The AMOC is a key element of the bi-polar ocean seesaw connecting the Northern and Southern Hemispheres. Seidov et al. (2001) by conducting a systematic set of ocean general circulation model experiments highlighted the role of Southern Ocean salinity anomalies in forcing the AMOC and found them much more influential than North Atlantic salinity anomalies. Here we also focus on the Southern Ocean influence on the AMOC, but concentrate on internal variability and investigate the link at centennial time scales.

The concept of the bi-polar ocean seesaw was also applied to multidecadal variability in the Atlantic during the instrumental period. The North Atlantic depicts pronounced basin-wide multidecadal sea surface temperature (SST) variability that is connected to the Southern Ocean SST and referred to as the Atlantic Multidecadal Oscillation or Variability (AMO/AMV). Folland et al. (1986), by investigating historical SST observations, were the first to identify the multidecadal dipolar SST anomaly pattern which is characterized by opposite changes in the North Atlantic and Southern Ocean. Climate models suggest that this pattern is potentially predictable at decadal time scales (e.g., Boer and Lambert, 2008). The existence of a twentieth century “bi-polar seesaw” of the Arctic and Antarctic surface air temperatures (SATs) has been postulated from observations by Chylek et al. (2010), but this was recently challenged by Schneider and Noone (2012) who investigated additional Antarctic stations. They did not find a robust correlation between Antarctic SAT and AMO/AMV. In many climate models, the decadal to centennial variations of the AMOC drive asynchronous SST changes in the North and South Atlantic (e.g., Latif et al., 2004; Knight et al., 2005; Park and Latif, 2008; Delworth and Zeng, 2012), while there is no robust pattern simulated in the high latitudes.

A centennial mode of open ocean deep convection in the Weddell Sea has been documented in two control integrations of the Kiel Climate Model (Martin et al., 2013). The flip-flop behavior of deep convection in one of the two integrations, which is further analyzed here, is clearly seen in both the Weddell Sea mixed layer depth (Fig. 1A) and Weddell Sea SST (Fig. 1B). The Weddell Sea deep convection is the major source of AABW in the Kiel Climate Model (KCM). It is driven by heat accumulated at mid-depth in the Atlantic–Indian sector of the Southern Ocean (SO, 70°S–80°E, south of 40°S) below 1600 m. (D) southward NADW transports [Sv] at 30°N (black) and 30°S (red) and northward AABW transport at 30°S (blue). (E) density difference [kg/m³] in the Atlantic between 30°–40°S and 30°–40°N below 1200 m. (F) transport anomaly [Sv] through Drake Passage as a measure of ACC strength variations. In all panels gray shaded areas mark time periods of the non-convective regime, i.e. without deep convection in the Weddell Sea. (See the web version of this article for color figures.)

Fig. 1. Time series of annual mean (A) mixed layer depth [m] averaged over the Weddell Sea convection region (WSCR, 58°–68°S, 35°W–10°E), (B) SST [°C] averaged over the same area, (C) deep ocean heat content (HC) anomaly [10¹² J] for the Atlantic–Indian sector of the Southern Ocean (SO, 70°W–80°E, south of 40°S) below 1600 m. (D) southward NADW transports [Sv] at 30°N (black) and 30°S (red) and northward AABW transport at 30°S (blue), (E) density difference [kg/m³] in the Atlantic between 30°–40°S and 30°–40°N below 1200 m, and (F) transport anomaly [Sv] through Drake Passage as a measure of ACC strength variations. In all panels gray shaded areas mark time periods of the non-convective regime, i.e. without deep convection in the Weddell Sea. (See the web version of this article for color figures.)

2. Climate model

We analyze a 1300-yr long control integration of present-day climate conditions of the Kiel Climate Model (KCM). The KCM (Park et al., 2009) consists of the ECHAM5 atmosphere general circulation model (AGCM) on a T31 (3.75° × 3.75°) grid coupled to the NEMO ocean sea-ice GCM on a 2° Mercator mesh with 0.5° meridional resolution in the equatorial region. No form of flux correction or anomaly coupling is used. The model employs constant levels of greenhouse gas concentrations characteristic of the present climate with a CO₂ concentration of 348 ppmv. The KCM simulates a rich spectrum of internal variability. In particular, the model simulates multidecadal SST variability in the North Atlantic and North Pacific Oceans with realistic period and spatial structure (Park and Latif, 2010). Compared to the original model version used by Park and Latif (2008) and Park and Latif (2010) the version used here employs a slightly modified parameterization in the sea ice model yielding a thicker sea ice cover in the Southern Ocean (Martin et al., 2013, their Fig. 1), which agrees well with observations (Latif et al., 2013, their Fig. 5). Nevertheless, this
version, like the previous one, simulates pronounced Southern Ocean multi-centennial variability (Fig. 1) as shown in Martin et al. (2013).

3. Results

To demonstrate the presence and timing of the SOCV we present time series of key variables such as mixed layer depth (MLD), sea surface temperature (SST), and heat content (HC) averaged over the Weddell Sea convection region (WSCR) defined as the area 58°–68°S and 35°W–10°E (Fig. 1). The identification of two regimes, i.e. convective and non-convective, enables us to present maps of differences in selected variables such as sea surface height and heat content. We use 100-yr averages to discuss the mean conditions during a regime, but 10-yr means to describe the peak phase conditions of a regime. An MLD threshold of 60 m is used to identify the timing of deep convection regime shifts in the Weddell Sea, because the spatially averaged annual mean MLD in the WSCR is always less (greater) than 60 m during the non-convective (convective) regime. The MLD is a runtime diagnostic and refers to the shortest vertical distance between the surface and the depth with a density gradient greater than 0.01 kg/m³ (Madec, 2008).

We define two idealized indices to diagnose phases of non-sinusoidally occurring convective events: a square wave marking the regime periods and a triangle wave emphasizing the regime shifts. The square wave equals 1 during years with active deep convection in the Weddell Sea and 0 otherwise. It emulates a quantity that shifts between two states. The triangle wave equals 1 during the year of convection onset and −1 during the year of convection shutdown; years between regime shifts are interpolated linearly. Thus the triangle wave emulates a quantity that changes continuously over the course of the regime. We find, for instance, that mean MLD and SST in the WSCR are good indicators of the presence of a certain regime. The two are highly correlated with the square wave at lags 0 and 1 year, respectively (Table 1). However, these two quantities are not suitable indicators of regime shifts. Peaks of the MLD and SST occur with lags of 45 and 49 years with respect to the triangle wave.

We note that the use of cross-correlations is limited, as the time series are non-stationary. Further, the control integration is rather short given the multi-centennial quasi-periodicity, leading to large uncertainties in the correlation coefficients. However, correlation analysis will help to introduce the basic relationships between the chosen indices, and these relationships will be subsequently supported by other analyses.

3.1. Southern Ocean deep convection and Antarctic Bottom Water

Open ocean convection is the process by which Antarctic Bottom Water (AABW) is produced in the KCM, and the Weddell Sea is most important in this respect. The Weddell Sea convection index, as defined by the annual mean mixed layer depth (MLD) in the WSCR, exhibits strong multi-centennial variability (Fig. 1A). The deep convection is also reflected in Weddell Sea SST averaged over the same region (Fig. 1B). A phase of deep convection goes along with anomalously warm SST in this region because the deeper waters brought up are warmer (and saltier) than the surface waters. In contrast, anomalously cold SST characterizes a phase without deep convection because the shallower mixed layer quickly loses heat to the cold atmosphere in winter. Periods of the non-convective regime are marked by gray shading in Fig. 1.

Cross-correlations were computed as a function of the time lag using the triangle wave as an index for regime changes in the Weddell Sea. The peak SST in the WSCR is reached almost 50 years after the convection onset (Fig. 2A). During periods of active deep convection, the mid (600–1600 m) and the deep ocean (below 1600 m) lose heat to the atmosphere. The heat depletion is not restricted to the convection region itself but extends far into the Atlantic and Indian sectors of the Southern Ocean (see Martin et al., 2013). Fig. 1C depicts the anomaly of the heat content of the deep ocean in the region 70°W–80°E and south of 40°S, termed SO HC. When the deep convection in the Weddell Sea sets in this heat reservoir starts to shrink, eventually losing heat in the order of 10²³ J over several decades, and is virtually depleted after about a century (e-folding time scale is 75 years). This is a necessary condition for convection shutdown (Martin et al., 2013) and the main reason for the centennial time scale of the SOCV. This is also expressed by the correlation of the SO HC with the triangle wave reaching a minimum about a century after the shutdown (Table 1, dashed red line in Fig. 2A). During the non-convective regime, heat is accumulated in the WSCR, mostly at mid-depth but also in the deep ocean, with a maximum correlation occurring 34 and 23 years prior to convection onset, respectively (Table 1). It is mostly the effect of the solid sea ice cover of at least 1 m thickness that acts as a lid, preventing deep convection by strongly reducing ocean-atmosphere heat exchange (Martin et al., 2013).

The formation rate of AABW varies strongly with the convection flip-flop in the Weddell Sea reaching a maximum toward the end of the convective regime. Accordingly, the index of the northward AABW transport in the Atlantic, defined as the magnitude of the minimum in the streamfunction at 30°S, varies between 2 Sv and 7 Sv and peaks at the end of the convective regime (Fig. 1D, blue line). The AABW transport index is anti-correlated (r = −0.84, lag 1 yr) with the SO HC (Figs. 1 and 2B, blue line). The latter is highly positively correlated with the deep convection flip-flop represented by the triangle wave (r = 0.81, lag = 3 yr; see also Table 1). The oscillatory behavior is well developed during the second half of the integration (Fig. 1): AABW extremes occur 100–150 years before and after convection onset (or heat content maximum) near years 4100 (Event 1) and 4400 (Event 2). This elucidates the tight link between deep heat content variations, deep convection, and variations in bottom water formation (Table 1).

3.2. AMOC lower branch

The net southward transport of North Atlantic Deep Water (NADW), defined as the maximum of the overturning streamfunction at 30°S and approximately 800 m depth, varies between 10 Sv and 15 Sv (Fig. 1D, red line). The NADW index at 30°N (Fig. 1D, black line), again defined as the local maximum of the streamfunction, varies approximately in the same range as the NADW outflow index at 30°S. The shutdown of the deep convection in the Weddell Sea (identified by a transition from white to gray background shading in Fig. 1) is followed by a strengthening of the NADW cell, which lasts several decades. What causes this strengthening? And how is it related to the SOCV?

The AABW and NADW indices at 30°S are anti-correlated, with a correlation magnitude exceeding 0.65 at lag 0 (dashed red line in Fig. 2B). Thus, the NADW outflow is closely linked to the strength of the northward AABW transport. The correlation is significant at a level of 0.99 (Table 1). The cross-correlation function of the AABW index at 30°S with the NADW index at 30°N (solid black line in Fig. 2B) is very similar. The lack of a phase lag suggests that the strength of the NADW export varies approximately in anti-phase with that of the AABW cell. However, the oscillatory behavior is not as prominent as in the cross-correlation function between the AABW and Southern Ocean heat content (SO HC, dash-dotted blue line in Fig. 2B), since the NADW transport also varies strongly on shorter time scales. In particular, the NADW cell exhibits strong
The square wave equals 1 during years of the convective regime (white background in Fig. 1) and 0 otherwise (gray shading). The triangle wave equals 1 in years of deep convection onset and 0 at shutdown and is interpolated linearly in between. A t-test was used to determine the level of significance for each correlation. The number of independent samples for each of the two time series being correlated was estimated based on the e-folding time scale of their autocorrelation functions and the mean was applied to the test.

Table 1
Timing of events and processes: cross-correlations of mixed layer depth (MLD, Fig. 1A), sea surface temperature (SST, Fig. 1B), sea surface height (SSH), heat content (HC, Fig. 1C), Antarctic bottom water (AABW) and North Atlantic deep water (NADW) transports (Fig. 1D), the strength of the Antarctic circumpolar current (ACC, Fig. 1F), and the North Atlantic current speed and subpolar gyre extent (Fig. 9B). Regions are defined as follows: Weddell Sea convection region (WSCR, 58°–68°S, 35°W–10°E; light gray contours in Fig. 4), Atlantic–Indian sector of the Southern Ocean (50°, 70°W–80°E, south of 40°S), and southern South Atlantic (35°–45°S, 35°W–0°). Heat content is computed separately for the mid (600–1600 m) and deep ocean (below 1600 m). Time series of the AABW and NADW transports as well as the North Atlantic current speed and subpolar gyre extent were low-pass filtered by an 11-yr running mean. A square wave emulates the contrast of the regime mean states, a triangle wave the timing of regime shifts. The square wave equals 1 during years of the convective regime (white background in Fig. 1) and 0 otherwise (gray shading). The triangle wave equals 1 in years of deep convection onset and 0 at shutdown and is interpolated linearly in between. A t-test was used to determine the level of significance for each correlation. The number of independent samples for each of the two time series being correlated was estimated based on the e-folding time scale of their autocorrelation functions and the mean was applied to the test.

| Leading variable | Lagging variable | Correlation r | Significance p | Lead in years |
|------------------|-----------------|---------------|----------------|---------------|
| Square wave      | MLD             | 0.87          | 0.99           | 0             |
|                  | SST             | 0.91          | 0.99           | 1             |
|                  | SSH             | -0.87         | 0.99           | 13            |
| Triangle wave    | MLD             | 0.65          | 0.99           | 45            |
|                  | SST             | 0.61          | 0.99           | 49            |
|                  | SSH             | -0.82/0.66    | 0.99/0.99      | -33/72        |
|                  | HC (600 m to 1600 m) | 0.81/-0.65    | 0.99/0.99      | -34/73        |
|                  | HC (below 1600 m) | 0.85/-0.58    | 0.99/0.98      | -23/100       |
|                  | SO HC (below 1600 m) | 0.81/-0.43    | 0.99/0.92      | -3/101        |
| SO HC (below 1600 m) | AABW 30°S     | -0.84/0.74    | 0.99/0.99      | 1/137         |
| AABW 30°S        | NADW 30°S       | -0.68/0.43    | 0.95/0.96      | -6/66         |
|                  | NADW 30°N       | -0.70/0.22    | 0.99/0.82      | 0/67          |
| WSCR SSH         | WSCR HC (600m to 1600 m) | -0.97         | 0.99           | 0             |
|                  | ACC strength    | -0.74         | 0.99           | 5             |
|                  | S. South Atl. SSH | -0.85         | 0.99           | 20            |
| NADW 30°N        | North Atl. current zonal speed | 0.73         | 0.99           | 1             |
|                  | North Atl. subpolar gyre extent | 0.45         | 0.99           | -11           |

Fig. 2. (A) Cross-correlations of an idealized triangle wave with annual mean sea surface temperature (SST, Fig. 1B) in the Weddell Sea convection region (WSCR) (solid black) as well as heat content (HC) in the Southern Ocean’s (SO) Atlantic and Indian sectors (40°–90°S, 70°W–80°E) below 1600 m (Fig. 1C) (dashed red). The triangle wave equals 1 in the first year of the convective regime and 0 in its last and is interpolated linearly in between. (B) Cross-correlation of the annual mean SO HC with the magnitude of the northward AABW transport at 30°S (dash-dotted blue); cross-correlations of the AABW transport magnitude with the southward transport magnitudes of North Atlantic Deep Water (NADW) at 30°N (solid black) and 30°S (dashed red); all time series in (B) were low pass filtered by an 11-yr running mean. Peak correlations of both panels referred to in the main text are listed. (See the web version of this article for color figures.)

multidecadal variability originating in the North Atlantic (Park and Latif, 2008). We conclude that the NADW cell responds to the Weddell Sea deep convection without feeding significantly back on the deep convection in the Southern Ocean. This can be inferred from the rather weak correlations at negative lags greater than 50 years (Fig. 2B).

We now describe in more detail the evolution of the AABW and NADW cells after the two major deep convection events labeled “Event 1” and “Event 2” in Fig. 1. Deep convection events before the model year 4100 are of more irregular character and may still be under the influence of model spin up, because the simulation presented here was branched off of another multi-millennial control simulation with a mean state featuring more active deep convection in the Weddell Sea (Martin et al., 2013).

The enhanced AABW transport during the convective regime (Fig. 1D, blue line) is associated with an increase in average density in the South Atlantic below 1200 m, so that the density contrast between South and North Atlantic is balanced at the peak of AABW transport (Fig. 1D, blue line, and Fig. 1E). This forces the NADW transport at both 30°S and 30°N to a minimum at the end of the convective period (Fig. 1D, red and black lines).

The zonally averaged overturning streamfunction at convection shutdown, defined as the conditions averaged over the 10 years prior to the shutdown, shows very similar patterns for both Event 1 and Event 2 (Fig. 3A and C). At convection shutdown, the strength of the AABW cell is close to its maximum with relatively large transports reaching far north (green contours in Fig. 3). The NADW cell is relatively shallow extending barely to 2000 m depth and the overall cell strength is rather weak (red contours).

The shutdown of the deep convection in the Weddell Sea is associated with a decrease in the northward transport of AABW at 30°S (Fig. 1D, blue line). The AABW cell weakens with an e-folding
time of about 50 years. AABW transport and the SO HC are positively correlated with a time lag of 137 years (Table 1 and blue line in Fig. 2B), indicating that the AABW transport has reached equilibrium by then. During this time, the average density in the South Atlantic below 1200 m decreases as retreating AABW is replaced by NADW and the density gradient toward the North Atlantic increases (Fig. 1E). Consequently, the NADW cell strengthens, which is visualized by another 10-yr average of the overturning streamfunction computed 100 years after the shutdown of Weddell Sea deep convection (Fig. 3B and D). By this time, which is in the middle of the non-convective regime, the NADW cell has reached a maximum (Fig. 1D) and deepened, now extending well below 2000 m (Fig. 3B and D). The outflow of NADW at 30°S, for instance, has increased by about 3 Sv (Fig. 1D, red line). This effect, i.e. the reduction in northward AABW transport causing an increase in southward NADW export, is one example of the bipolar ocean seesaw. And since oscillations in SO HC drive the variations in AABW production and transport through the deep convection flip-flop, the AMOC variability is clearly linked with the SOCV.

3.3. Southern Ocean sea level and the ACC

The depletion of SO HC results in a significant drop in sea surface height (SSH) in the entire Southern Ocean, but most prominently in the Weddell Sea (Fig. 4). SSH is a mean dynamic topography calculated with filtered free surface formulation (Roulet and Madec, 2000) and represents the local divergence of currents and the freshwater influx (Madec, 2008). The differences are derived from two 100-yr means of SSH, which represent the convective and non-convective regimes in the Weddell Sea. The SSH difference amounts to almost 30 cm in the WSCR. In the convection region (indicated by the light gray contours of mixed layer depth in Fig. 4), SSH variability is highly correlated at 0.97 and in phase with the heat content changes at mid-depth (600–1600 m). The correlation of WSCR SSH peaks 33 years prior to convection onset (given by the triangle wave) and has a minimum 72 years after onset (Table 1). This century-long decline in sea level in the Atlantic Sector of the Southern Ocean causes a steepening of the meridional SSH gradient across the ACC and of the zonal SSH gradient through Drake Passage. Both gradients increase by about 10% during the convective regime. The steeper SSH gradients drive an increase in ACC strength by up to 20 Sv (Fig. 1F), which has a delay of 5 years relative to the SSH change in the WSCR (Table 1). While the geostrophic surface flow derived from the SSH gradient explains 80% of the ACC variability in the Atlantic sector, the associated temperature changes in the upper to mid ocean add some variability through the thermal wind balance, which by itself explains only 50% of the ACC variability in our simulation.

A section along 30°W extending from 30°S to 55°S shows the meridional structure of the zonal flow (Fig. 5A). Despite the coarse resolution of the ocean model grid, two separate jets can clearly be identified during the non-convective regime (red contours). The northern jet at ~36°S depicts the South Atlantic Current (SAC), and the southern jet at ~47°S marks the ACC. The steepening of the meridional SSH gradient in the Atlantic sector during the convective regime causes the SAC to shift southward (Fig. 5A, dashed blue contours). Also seen is the pronounced increase in zonal velocity in the southern jet, the ACC. The southward shift of the SAC reaches a maximum of 4° latitude toward the end of the convective regime, when it almost merges with the ACC. The magnitude – not the tendency – of this response to the deep convection variability, however, may be affected by the coarse resolution of our model.

The SAC is associated with the Subtropical Front, a sharp gradient in temperature and salinity between the South Atlantic and the Southern Ocean. In line with the southward shift of the SAC, the Subtropical Front expands southward. Fig. 5B displays the differences in potential temperature for the same section (color-filled contours) with the solid and dashed bold black lines depicting the 10 °C isotherms in the non-convective and convective regime, respectively. The warming associated with the migration of the front extends down to about 1000 m and has a
maximum of 5 °C at 250 m. This warming just north of the SAC causes a positive SSH anomaly, which lags the WCSR SSH by 20 years (Table 1) and further steepens the cross-ACC SSH gradient (Fig. 4).

The atmosphere also responds to the regime changes in Weddell Sea deep convection. An increase in surface wind stress is simulated after convection shutdown. This is related to the increase of the meridional sea level pressure (SLP) gradient, which is caused by the surface cooling associated with the shutdown as discussed in Latif et al. (2013). The ACC, however, weakens during the non-convective regime (Fig. 1D), which rules out wind stress as the driver of multi-centennial ACC variability in our simulation.

3.4. AMOC upper branch

Fig. 6 shows the heat content difference between the convective and non-convective regimes in the upper (150–600 m) and mid to deep (1200–3000 m) Atlantic Ocean for Events 1 and 2 (see Fig. 1 for the timing of the events). The differences are calculated by subtracting a 10-yr average of the non-convective regime (NADW maximum) from a 10-yr average toward the end of the convective regime (AABW maximum, cf. Fig. 3). First, we note the great heat loss from the upper and deep layers caused by the deep convection in the Weddell Sea (dark blue colors in the Southern Hemisphere between 50° and 70 S). Second, as a consequence of this deep convection, the upper ocean gains heat north of the ACC (yellow and red colors in Fig. 6A and C). And third, while the upper ocean is generally warming, the mid to deep ocean cools (Fig. 6B and D) depicting the shoaling of the NADW cell during the two events when the AABW cell is well developed (Fig. 3A and C).

In the South Atlantic, the aforementioned shift of the Subtropical Front manifests itself as a strong heat content increase in the upper ocean at ~40° S. Hovmoeller diagrams of the zonal mean heat content in the upper (150–600 m, Fig. 7A) and mid to deep ocean (1200–3000 m, Fig. 7B) show that this strong warming (positive anomalies at ~40° S in Fig. 7A) is in phase with the prominent cooling of the mid to deep ocean south of the ACC (~50° S) displayed in Fig. 7B. The warming spreads further north within 10–20 years and is associated with an increased inflow from the Indian Ocean – enabled by the SAC shift – and a warmer, more extensive subtropical gyre, and found along the western boundary of the Atlantic Ocean and the path of the North Atlantic Current (Fig. 6A and C). The latter is also reflected in SSH (Fig. 4).

The heat content changes are in accordance with the advective mechanism described by Swingedouw et al. (2009). Interestingly, the slow response of the SAC, which is delayed by about 50 years...
to deep convection onset, causes the positive heat content anomalies to barely reach the equatorial region before the deep convection in the Southern Ocean ceases (shutdown years are indicated by solid black lines in Fig. 7A).

Heat content variability in the northern North Atlantic is dominated by strong multidecadal variability (Fig. 7A). However, the centennial variability south of 40°N occasionally “leaks” into northern latitudes and interferes with the more prominent multidecadal variability. This interaction produces a strong warming in the region 40°N–60°N about 130 (200) years after deep convection shutdown (onset) in the Southern Ocean. The increased heat content in the upper ocean may not only slow down deep water formation but affects the water properties of the deep water itself. 

NADW formation clearly transfers this signal to the deep ocean, where it then propagates southward (Fig. 7B). Here, positive (negative) heat content anomalies are associated with higher (lower) temperature and salinity. This affects the density of the NADW, which feeds back into the density difference between North and South Atlantic. That is, North Atlantic deep ocean density (averaged for depths below 1200 m) decreases during the convective regime because it warms (Fig. 7B) while South Atlantic density increases due to advancing AABW, with the letter having greater impact on the meridional density difference (Fig. 1E). In light of the multi-centennial cycle presented here it is important to note that due to the long delay the heat anomaly, which causes the NADW density to decrease, originates from the

Fig. 6. Heat content differences [MJ/m³] in the Atlantic for the upper branch of the AMOC between 150 m and 600 m (A and C) and the lower AMOC branch averaging depths 1200–3000 m (B and D) for Event 1 (A and B) and Event 2 (C and D). Differences are computed for the same periods as in Fig. 3: the decade just before convection shutdown in the Weddell Sea minus the decade 100 years after shutdown. (See the web version of this article for color figures.)

Fig. 7. Hovmoeller diagrams of zonal mean Atlantic heat content anomalies [MJ/m³] averaged over 150–600 m (A) and 1200–3000 m depth (B). The time series are low-pass filtered by applying an 11-yr running mean. Dashed black lines mark the onset of deep convection in the Weddell Sea, solid lines its shutdown.
previous deep convection event in the Weddell Sea (Fig. 7). Moreover, the larger amount of heat carried south by the lower branch of the AMOC arrives in the southern South Atlantic during the (next) convective regime. This additional heat supply to the Weddell Sea could possibly support the deep convection there by reducing the effective rate of heat depletion.

These results suggest, that the North Atlantic upper ocean heat content is likely more influenced by the adjustment in the deep ocean than by upper ocean anomalies, and that the enhanced NADW export causes the deep water formation in the subpolar North Atlantic to intensify, leading to an enhanced northward transport of warm subtropical waters. Additional analyses of the subpolar gyre presented in the next section will lend further support to this hypothesis.

3.5. North Atlantic sea level and subpolar gyre

In the following, we show that the multi-centennial variability in the Southern Ocean also affects North Atlantic sea level by its strong impact on the AMOC. As depicted in Fig. 4 pronounced sea level differences are evident not only in the Southern Hemisphere but also in the northern North Atlantic. We now investigate the centennial sea level changes in the North Atlantic in more detail (Fig. 8). Sea level is depressed by $-2.5$ m in the center of the North Atlantic subpolar gyre in the long-term (1000-yr) average (Fig. 8A, gray contours). The subpolar gyre itself is clearly identified by the mean SSH contours. We apply an empirical orthogonal function (EOF) analysis to North Atlantic SSH between $30^\circ$N and $70^\circ$N to compute the leading variability pattern (EOF1). The EOF1 explains 25.7% of the SSH variability in this region and has its maximum along the eastern flank of the subpolar gyre where the North Atlantic Current (NAC) turns north (Fig. 8A, color-filled contours). We note that our simulation like those from many other climate models is biased, as the deep convection, usually residing in the Labrador Sea, is displaced to the southeast (Fig. 4, gray outlines). This is associated with a predominantly zonally flowing NAC. However, this will not affect the main mechanism described here. The time series of the principle component (PC1) associated with the EOF1 (black line in Fig. 8B) features highly significant power on centennial time scales (Fig. 8C). None of the other EOFs is associated with strong centennial variability, which was inferred from the spectra of the PCs lacking statistically significant centennial variability above the 99% level. The subsequent analysis is therefore restricted to the EOF1.

In the center of action of the EOF1, in the East Atlantic at $50^\circ$N and $20^\circ$W, the peak-to-peak SSH change amounts to about 20 cm/century. The centennial variability in the PC1 is superimposed by considerable decadal to multidecadal variability and becomes much clearer after applying a 101-yr running mean (red line in Fig. 8B). We now see that the centennial variability of the PC1 is coherent with the Southern Ocean Centennial Variability (SOCV). The low-pass filtered PC1 has a peak toward the end of the non-convective regime (gray shading) and then slowly declines during the convective regime. The decline continues into the next non-convective regime and minimum sea level is reached about 100–150 years after the maximum. This compares well with the oscillation timescale of the AABW variability (Section 3.2).

We compute composites of the local SSH anomalies using the PC1 as an index. The average SSH anomaly of years in which the PC1 exceeds 1 standard deviation (horizontal dotted blue lines in Fig. 8B) is depicted in red in Fig. 9A, the mean SSH of years with

Fig. 8. Dominant sea level changes in the North Atlantic. (A) Pattern associated with the 1st empirical orthogonal function (EOF1) of the sea surface height (SSH) between $35^\circ$N and $70^\circ$N, which explains 25.7% of the variability; gray contours depict the local SSH anomaly [m] averaged over 1000 years; for values smaller than $-2.0$ m the contour increment is 0.2 m (indicating the subpolar gyre center), for values greater than $-2.0$ m the increment is 0.1 m (depicting the North Atlantic Current). (B) Time series of the associated 1st principle component (PC1) in black; horizontal dotted blue lines mark ±1 standard deviation from the long-term mean; the bold red line depicts the PC1 low-pass filtered by a 101-yr running mean; gray shaded periods indicate the non-convective regime in the Southern Ocean. (C) Power spectrum of PC1, which has pronounced variability on centennial time scales; colored dashed lines indicate significance levels based in an AR(1) process. (See the web version of this article for color figures.)
the PCI being less than $-1$ standard deviation is illustrated by dashed blue contours. The SSH variations along the eastern flank of the subpolar gyre are related to the extent of the subpolar gyre. During years with anomalously positive SSH anomalies (solid red contours in Fig. 9A), the gyre covers a smaller area (compare the red and blue $-2.0$ m contours) and the depression in its center, which includes the deep convection sites, is less deep. A low-pass filtered time series of the gyre center area, defined as the area enclosed by the $-2.4$ m contour, is presented in Fig. 9B in red. In general, the gyre area is largest when the SSH anomaly is at its minimum and vice versa. More importantly, the gyre area increases when the NADW transport increases (Fig. 9B, blue line; same as in Fig. 1D, black line). Since the mean mixed layer depth within the subpolar gyre center does not exhibit enhanced power on centennial time scales (not shown), we suggest that the increase in NADW transport initiated by the SOCV is compensated by an expansion of the deep convection area in the North Atlantic – rather than intensification of the convection, i.e. deepening of the mixed layer – and an acceleration of the NAC. Although the link is rather weak with a correlation of 0.45 for the gyre extent and 0.73 for the NAC, it is significant at the 0.99 level (Table 1).

The gyre contraction, which in our model occurs primarily in a west-east direction, opens the passage for warmer waters from south of the NAC to penetrate farther north. In fact, it is after the gyre contraction that the positive heat content anomaly in the upper North Atlantic is found north of 40°N (Fig. 7A). Further discussion of the effect of a contracting subpolar gyre on the AMOC can be found in Hátún et al. (2005). We speculate that the increased heat content of the upper ocean, which spreads into the northern sinking region after the gyre contraction, is likely a key factor in reducing the NADW transport a decade later. The additional heat gets entrained into the NADW and eventually moves south within the lower AMOC branch (Fig. 7B).

4. Discussion and conclusions

We have shown from a millennial control integration of the Kiel Climate Model (KCM) that the Southern Ocean Centennial Variability (SOCV) originating from a flip-flop of Weddell Sea deep convection can have a significant impact on the Atlantic Meridional Overturning Circulation (AMOC), the North Atlantic subpolar gyre and sea level. The SOCV is a result of competing buoyancy effects of surface freshening and mid to deep ocean warming that generate an oscillatory occurrence of open ocean deep convection in the Weddell Sea. As this deep convection is the major source of Antarctic Bottom Water (AABW) in the KCM, the northward AABW transport declines after the convection halts. The AABW decline enables an increase in southward North Atlantic Deep Water (NADW) export by enhancing the density difference between South and North Atlantic. This is seen as a slow strengthening of the AMOC after deep convection shutdown in the Weddell Sea. As a consequence of the greater NADW export, deep convection in the North Atlantic intensifies, which is indicated by an expansion of the North Atlantic subpolar gyre and a decrease in sea surface height in its center, and the North Atlantic Current accelerates. At the same time, the deep convection in the Weddell Sea initiates a northward propagation of a positive heat content anomaly in the upper Atlantic Ocean by redistributing heat from deeper layers and enhancing the South Atlantic subtropical gyre. The heat anomaly eventually reaches the northern North Atlantic leading to a decrease in deep water formation there and travels back south as a signal in the lower branch of the AMOC. The time scale of this link between Southern Ocean and North Atlantic deep convection regions is of the order of a century. The AMOC strength changes by up to 4 Sv and North Atlantic sea level variations are on the order of about 20 cm/century. Such changes are of the same order of magnitude as the global average sea level rise during the 20th century amounting to about 15–20 cm. This suggests that internal centennial variability cannot be ignored when assessing 20th century and projecting 21st century North Atlantic sea level trends.

In the KCM, there is a clear teleconnection between the Southern Ocean and the North Atlantic such that strong changes of the convective activity in the Weddell Sea are followed by a slow adjustment of the NADW cell. The changes in the KCM are consistent with the conceptual picture presented by Seidov et al. (2001), see their Fig. 1, in which the AABW cell and the NADW cell compete with each other. In Seidov et al. (2001), the cell that was directly perturbed by artificial surface freshening becomes weaker while the other strengthens, i.e. adding freshwater in the north weakens the NADW cell and strengthens the AABW cell. In our simulation, the situation is reversed. A prolonged phase of Weddell Sea deep convection drives an anomalously strong AABW cell which weakens the NADW cell with regard to both strength and vertical extent. After the halt of deep convection in the Weddell Sea, the AABW cell becomes less prominent and the NADW cell begins to slowly strengthen. The increase in NADW formation and export is mostly stopped by regional processes in the North Atlantic but also supported by the additional heat brought to the
upper North Atlantic from the Southern Ocean through northward advection by the western boundary currents. The density of the NADW itself is altered by the heat changes on a centennial timescale. Thus, the meridional density gradient associated with the bipolar ocean seesaw is not only influenced by advancing and retreating AABW as described in e.g. Swingedouw et al. (2009), but also by variations of NADW density linked to surface processes triggered by the SOCV.

In multi-millennial control simulations with different climate models, Delworth and Zeng (2012) as well as Cimatoribus et al. (2012) find salinity anomalies in the southern South Atlantic, not the Southern Ocean, to be instrumental in the generation of centennial-scale AMOC variability. We thus note that the generation of SOCV may be model dependent and Latif et al. (2013) show for the KCM that the time scale of the SOCV is very sensitive to model formulation. This relates to both model parameters and horizontal resolution.

Caution, however, should be taken when discussing the implications of the KCM results for past and future climate change. The model suffers from large biases, many of them being common to most climate models. In particular, the KCM produces Antarctic Bottom Water (AABW) mostly through open ocean deep convection in the Weddell Sea, which is unrealistic. Another bias concerns the too zonal path of the North Atlantic Current, which gives rise to a large cold SST bias in the central North Atlantic. Again, this bias is found in most climate models. Furthermore, the control simulation used in this study by definition does not take into account the effects of spatially and temporally varying external forcing, which likely played an important role in the climate evolution during the 20th century and will probably become increasingly important during the 21st century. Yet this study lends support to the hypothesis that internal variability on centennial time scales cannot be ignored and should be considered as a possible mechanism when discussing the origin of past and future climate changes on regional scales. Phenomena like the Medieval Warm Period or the Little Ice Age during the last millennium or the current hiatus in global warming may have contributions from internal centennial variability, and the Southern Ocean could have played a role. If the model results carry over to the real world, projections of the future climate cannot be treated as a “pure” boundary value problem, even when focusing on the changes toward the end of the 21st century.

Acknowledgments

This work was supported by the BMBF RACE Project and the EU NACLIM project. The model integrations were performed at the computing center of Kiel University. We thank two anonymous reviewers for their pointed comments, which helped to greatly improve this study.

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