Evolving Relative Importance of the Southern Ocean and North Atlantic in Anthropogenic Ocean Heat Uptake

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(Manuscript received 23 March 2018, in final form 19 June 2018)

ABSTRACT

Ocean uptake of anthropogenic heat over the past 15 years has mostly occurred in the Southern Ocean, based on Argo float observations. This agrees with historical simulations from phase 5 of the Coupled Model Intercomparison Project (CMIP5), where the Southern Ocean (south of 30°S) accounts for $72\% \pm 28\%$ of global heat uptake, while the contribution from the North Atlantic north of 30°N is only 6%. Aerosols preferentially cool the Northern Hemisphere, and the effect on surface heat flux over the subpolar North Atlantic opposes the greenhouse gas (GHG) effect in nearly equal magnitude. This heat uptake compensation is associated with weakening (strengthening) of the Atlantic meridional overturning circulation (AMOC) in response to GHG (aerosol) radiative forcing. Aerosols are projected to decline in the near future, reinforcing the greenhouse effect on the North Atlantic heat uptake. As a result, the Southern Ocean, which will continue to take up anthropogenic heat largely through the mean upwelling of water from depth, will be joined by increased relative contribution from the North Atlantic because of substantial AMOC slowdown in the twenty-first century. In the RCP8.5 scenario, the percentage contribution to global uptake is projected to decrease to $48\% \pm 8\%$ in the Southern Ocean and increase to $26\% \pm 6\%$ in the northern North Atlantic. Despite the large uncertainty in the magnitude of projected aerosol forcing, our results suggest that anthropogenic aerosols, given their geographic distributions and temporal trajectories, strongly influence the high-latitude ocean heat uptake and interhemispheric asymmetry through AMOC change.

1. Introduction

More than 90% of Earth's energy imbalance (EEI), mainly caused by greenhouse gas (GHG) emissions from industrial development, is stored in the ocean (Rhein et al. 2013; Trenberth et al. 2014). Increases in global ocean heat content (OHC) are robustly detected in both observations and model simulations (Levitus et al. 2012; Rhein et al. 2013; Roemmich et al. 2015; Riser et al. 2016; Liu et al. 2016; Cheng et al. 2016, 2017; Gleckler et al. 2016), illustrating the ocean's dominant role in slowing the rate of surface warming (von Schuckmann et al. 2016; Exarchou et al. 2015). The

excess heat, defined as the change in heat content since preindustrial time, enters the ocean through air–sea heat flux. The pattern of ocean heat uptake (heat taken up at the surface) is strongly associated with a multitude of complex climate processes, such as the concentration of atmospheric CO₂, aerosols, changing ocean circulation, cloud feedback, and eddies (Morrison et al. 2016).

The Southern Ocean has received much recent attention in the discussion of global heat uptake and heat storage. The dominance of the Southern Ocean in heat uptake is partly associated with wind-driven upwelling of cold deep water (Bryan et al. 1988; Manabe et al. 1990; Morrison and Hogg 2013; Frölicher et al. 2015). The upwelling of pristine deep water delays the Southern Ocean warming at the surface, then takes up a great amount of excess heat from the atmosphere, and then carries the heat content anomaly northward and downward into the thermocline via the overturning

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

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-18-0170.s1>.

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Publisher's Note: This article was revised on 18 October 2018 to add the open access designation that was missing when originally published.

DOI: 10.1175/JCLI-D-18-0170.1

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circulation (Morrison et al. 2016; Armour et al. 2016). The enhanced heat uptake in the Southern Ocean has a profound impact on tropical climate change, displacing the intertropical convergence zone (ITCZ) and monsoons (Hwang et al. 2017).

GHGs are the most important radiative forcing that warms the global climate. Much previous work focuses on ocean heat uptake in the Southern Ocean in response to increased CO₂ (the most important component in GHGs) (Bryan et al. 1988; Manabe et al. 1990, 1991; Kuhlbrodt and Gregory 2012; Marshall et al. 2015; Armour et al. 2016). For instance, in response to a quadrupling of CO₂ in phase 5 of the Coupled Model Intercomparison Project (CMIP5), enhanced heat uptake is found in both the Southern Ocean and the northern North Atlantic (Marshall et al. 2015). This suggests that the northern North Atlantic also plays an important role in the ocean uptake of anthropogenic heat. Climatologically, strong heat loss from the ocean to the atmosphere takes place in the subpolar North Atlantic (SPNA), leading to the formation of North Atlantic Deep Water (NADW) (Talley et al. 2011). The deep water formation is part of the Atlantic meridional overturning circulation (AMOC) (Böning et al. 2006) through which the SPNA is tied to the ocean circulation and global heat transport (Robson et al. 2016; Zhang and Yan 2017; Heuzé 2017). An AMOC slowdown in response to the increasing CO₂ results from the enhanced stratification of the upper water column due to the increased buoyancy flux into the SPNA (Thorpe et al. 2001; Gregory et al. 2005; Buckley and Marshall 2016). Associated with a weakening AMOC, less excess heat is transported to high northern latitudes, resulting in a sea surface temperature (SST) cooling tendency in the SPNA and increased heat flux into to ocean (Wood et al. 1999; Russell and Rind 1999; Weaver et al. 2007; Kim and An 2013; Rugenstein et al. 2013; Winton et al. 2013; Gregory et al. 2016).

Anthropogenic aerosols are a second major radiative forcing for global climate change. Aerosols and GHGs have opposite effects: while GHGs warm Earth, aerosols cool it. The global historical GHG effective radiative forcing (ERF) is estimated at $2.5 \pm 0.4 \text{ W m}^{-2}$, while the ERF of the historical aerosols is about $-1.0 \pm 0.4 \text{ W m}^{-2}$ in 2000 relative to 1850 based on the CMIP5 models (Shindell et al. 2015). Besides the difference in the magnitude of their ERF, GHG and anthropogenic aerosol forcings also show different spatial distributions. Because of their short residence time and localized emissions, anthropogenic aerosols are mainly distributed in the Northern Hemisphere (NH), while GHGs are well mixed in the atmosphere. The geographically distributed anthropogenic aerosols can lead to cross-equatorial energy transport in both the ocean and

atmosphere (Kang and Xie 2014; Acosta Navarro et al. 2017). The interhemispheric asymmetric climate responses, such as a southward shift of the ITCZ and cross-equatorial wind, are dominated by the inhomogeneous aerosol forcing (Wang et al. 2016a,b).

Anthropogenic ozone change is another important radiative forcing agent responsible for regional climate change. The anthropogenic stratospheric ozone hole over Antarctica has been the focus of extensive Southern Hemisphere climate studies. Antarctic stratospheric ozone depletion has enhanced Southern Hemisphere westerly winds, affecting the surface climate around Antarctica (Thompson et al. 2011; Waugh et al. 2013). Southern Ocean sea surface temperature shows fast and slow responses—rapid cooling followed by slow but persistent warming—due to ozone-hole forcing (Marshall et al. 2014; Ferreira et al. 2015). The ozone forcing fields, however, have large variations across the CMIP5 models with either interactive or prescribed ozone (Eyring et al. 2013; Frölicher et al. 2015).

In CMIP5 historical runs that include all of the anthropogenic (i.e., GHGs, aerosols, and ozone) and natural (i.e., solar irradiance and volcanoes) radiative forcings, the Southern Ocean (south of 30°S) takes up as much as 75% of global oceanic heat gain (Frölicher et al. 2015). Based on observations, the Southern Ocean has experienced continuous and monotonic long-term warming since around the 1960s (Gille 2002; Cheng et al. 2017). Furthermore, Roemmich et al. (2015) show that the extratropical Southern Hemisphere (20°–60°S) contributes 67%–98% of global upper ocean (above 2000 m) heat content change during the Argo era since 2006. This large fraction of heat uptake in the Southern Ocean contrasts with much smaller ocean heat uptake in the SPNA in both historical simulations and observations, even taking into account the different sizes of the two regions. Could the relative importance of the SPNA and Southern Ocean change in the future as a result of changes in the relative strengths of the anthropogenic forcings?

The present study explores the underlying mechanisms for the long-term change of regional ocean heat uptake. By using the single-forcing simulations of CMIP5, we find that GHG radiative forcing itself is not enough to account for regional pattern of historical ocean heat uptake, especially in the SPNA. Historically, the impact of anthropogenic aerosols, which are mostly confined to the NH, is to nearly completely offset heat uptake due to GHGs in the SPNA. We find that this results from compensation of AMOC change due to GHGs by that due to anthropogenic aerosol forcing. In contrast to this small historical heat uptake in the SPNA, heat uptake is large in the Southern Ocean, where the GHG forcing dominates, somewhat reinforced by ozone

TABLE 1. CMIP5 models used in this study. Numbers shown in columns of “model experiments” give the number of ensemble members used in various simulations: GHG single-forcing runs (GHG), anthropogenic aerosol single-forcing runs (AERO), historical runs (HIST), RCP4.5, and RCP8.5. (Expansions of acronyms are available online at <http://www.ametsoc.org/PubsAcronymList>.)

Institute	Model	Reference	Model experiments				
			GHG	AERO	HIST	RCP4.5	RCP8.5
CCCma	CanESM2	von Salzen et al. (2013)	5	5	5	5	5
NCAR	CCSM4	Meehl et al. (2012)	3	3	3	3	3
NSF-DOE-NCAR	CESM1-CAM5	Hurrell et al. (2013)	1	3	3	3	3
CSIRO-QCCCE	CSIRO-Mk3.6.0	Rotstayn et al. (2012)	10	10	10	10	10
NOAA GFDL	GFDL CM3	Donner et al. (2011)	3	3	3	3	1
NOAA GFDL	GFDL-ESM2M	Dunne et al. (2013)	1	1	1	1	1
NASA GISS	GISS-E2-R	Schmidt et al. (2014)	5	5	5	5	1
IPSL	IPSL-CM5A-LR	Dufresne et al. (2013)	1	1	1	1	1
NCC	NorESM1-M	Iversen et al. (2013)	1	1	1	1	1

forcing, and where aerosol forcing is weak. As anthropogenic aerosol emissions are expected to decrease while GHGs will continue to rise, the relative importance of the SPNA and Southern Ocean in future ocean heat uptake cannot be extrapolated from the historical change. Based on observations, historical period simulations, and future projections, our goal is to show the shifting relative importance of the two regions. We will show that the North Atlantic’s role in global ocean heat uptake rises dramatically in the twenty-first century.

The rest of the paper is organized as follows. [Section 2](#) describes the methods and data used. [Section 3](#) discusses the surface heat flux response patterns under different model simulations over the historical period from 1861 to 2005. [Section 4](#) investigates the patterns of ocean heat uptake in future projections from 2006 to 2100. [Section 5](#) discusses the response of the AMOC to anthropogenic forcing and its impact on regional heat uptake. [Section 6](#) further investigates the effect of anthropogenic aerosol reduction in future projections. [Section 7](#) compares OHC changes between observations and simulations. [Section 8](#) is a summary.

2. Data and methods

a. Simulations

This study uses outputs of climate model simulations from the World Climate Research Program’s CMIP5 ([Taylor et al. 2012](#)). We choose nine models that perform all of the following five experiments: GHG single-forcing (GHG), anthropogenic aerosol single-forcing (AERO), historical all-forcing simulations (HIST), and the representative concentration pathway (RCP) RCP4.5 and RCP8.5 scenarios. The numbers of ensemble members for each model and each experiment are listed in [Table 1](#). According to the conservation of energy, the heat budget for a particular body of water is ([Talley et al. 2011](#))

$$Q_T = Q_{\text{net}} + Q_V, \quad (1)$$

where Q_T is the rate of change in heat of the body of water, Q_{net} is the net surface heat flux, and Q_V is the advective heat flux by ocean circulation. The term Q_{net} is derived by summing turbulent flux (latent and sensible heat flux) and radiative flux (shortwave and longwave radiation) at the surface. In steady state (e.g., prior to the industrialization), $Q_T \approx 0$. With increasing GHGs, ocean takes up and stores the anthropogenic heat. Hereafter, we call changes in Q_{net} (represented by the linear trend) the anthropogenic ocean heat uptake (through the surface), and time integration of Q_{net} relative to the preindustrial steady state the cumulative heat uptake.

Here, we focus on the changes of Q_{net} and SST during the historical period (1861–2005) and RCP scenarios (2006–2100). The multimember average for each model is obtained first and then the multimodel ensemble mean is made to reduce internal variability. In GHG (AERO) runs, GHGs (anthropogenic aerosols) are the only time-varying forcing agent, with other forcings fixed at the preindustrial level. Besides GHG and AERO, we also use the ozone single-forcing runs (OZONE) from three CMIP5 models ([Cionni et al. 2011](#)) to investigate the effect of ozone on ocean heat uptake. Because there are many fewer ozone ensemble members and because the trade-off between AERO and GHG forcing dominates the ocean heat uptake, our analysis of ozone effects is not exhaustive, and is mostly included in the supplemental materials. The HIST runs are forced by historical estimates of changes in anthropogenic GHGs and aerosols, natural forcing (volcanoes and isolation), and changes in land use ([Sheffield et al. 2013](#)).

We also use 38 realizations from the CESM Large Ensemble Project (LENS), all of which use the same external forcing and model with small differences in the atmospheric initial condition ([Kay et al. 2015](#)). Therefore, the effects of external forcing and internal climate

TABLE 2. Observed data used in this study. Annual mean is applied to all the observed data.

Observation	Reference	Data	Depth	Period
Argo	Roemmich and Gilson (2009)	Temperature	0–2000 m	2006–16
IAP	Cheng et al. (2016)	Temperature	0–2000 m	1940–2015
EN4-G10	Good et al. (2013); Gouretski and Reseghetti (2010)	Potential temperature	0–5500 m	1900–2016
EN4-L09	Good et al. (2013); Levitus et al. (2009)	Potential temperature	0–5500 m	1900–2016

variability can be isolated by analyzing the ensemble mean and deviations, respectively. Historical and RCP8.5 external forcing from 1920 to 2100 (ensemble member 1 starts from 1850) is applied in LENS. We also use a modified RCP8.5 ensemble from CESM with aerosols fixed at their 2005 level (2005AERO) (Xu et al. 2018) to identify the effect of changing anthropogenic aerosols in future projections. The difference between LENS and 2005AERO (15 realizations) is their aerosol trajectories.

An AMOC index is defined as the maximum volume transport streamfunction at 30°N and is calculated for the first-member run (r1i1p1) from each CMIP5 model. The AMOC anomaly, relative to the 1861–80 average in each model, is used to calculate the multimodel mean of the AMOC change. We also obtain the AMOC anomaly from the ensemble member 1 of LENS and 2005AERO. The AMOC anomaly of LENS is relative to the 1861–80 mean, and the AMOC anomaly of 2005AERO is relative to the annual mean of 2006.

Annual anomalies and the 11-yr running average are calculated for the analysis of air–sea heat flux and the AMOC intensity. We interpolate all model outputs to a regular $1^\circ \times 1^\circ$ latitude–longitude grid to facilitate comparison.

b. Observations

Observed ocean temperature is compared with the simulations to assess their validity. Argo floats have been profiling ocean temperature in the upper 2000 m since 2004 (Gould et al. 2004; Roemmich and Gilson 2009). Considering the large coverage gaps in 2004 and 2005 (Roemmich et al. 2015), we use the monthly gridded Argo temperature data (Roemmich and Gilson 2009) from the beginning of 2006. For a longer time series, extending to the deep ocean, we use the optimal interpolated EN4 potential temperature product from the Met Office Hadley Centre, which is available on a $1^\circ \times 1^\circ$ grid with 42 vertical levels from the surface to about 5500 m during 1900–2016 (Good et al. 2013). The Levitus et al. (2009) and Gouretski and Reseghetti (2010) methods are used to correct the XBT bias in EN4-L09 and EN4-G10 versions, respectively. We also

use the monthly mean temperature product from the Institute of Atmospheric Physics (IAP) (Cheng et al. 2016, 2017) from 1940 to 2015 on a regular $1^\circ \times 1^\circ$ grid with 41 vertical levels in the upper 2000 m. The EN4 and IAP products include shipboard, mechanical bathythermograph (MBT), XBT, and Argo temperature profiles. Observational data are listed in Table 2.

In this study, OHC is calculated by integrating the temperature profile within a certain layer and then multiplying by reference values of density (1025 kg m^{-3}) and specific heat ($3985 \text{ J kg}^{-1} \text{ K}^{-1}$) of seawater. OHC in different layers (0–300 m, 0–700 m, 0–2000 m, or below 2000 m) is calculated for each observational dataset (Argo, EN4, and IAP). Subsurface potential temperature from LENS is also used to calculate the OHC and compare with observations. To suppress high-frequency noise, we apply a 5-yr running average to the time series of the observed and LENS OHC. The time evolution of OHC is relative to the climatology from 1975 to 2012. Empirical orthogonal function (EOF) analysis is applied to the 0–2000-m OHC in the northern North Atlantic (north of 30°N) of IAP data from 1950 to 2015. A similar EOF analysis is applied to EN4-G10 from 1950 to 2016. A 5-yr running average is applied before EOF analysis. The principal components are normalized by their respective standard deviations.

3. Historical responses to anthropogenic forcing

a. Ocean heat uptake

The CMIP5 multimodel ensemble mean net surface heat flux Q_{net} trends for the historical period (1861–2005) are shown in Fig. 1. Positive (negative) trends indicate that the ocean is being heated (cooled). In HIST runs (Fig. 1c), the Southern Ocean dominates ocean heat uptake. In GHG runs (Fig. 1a), heat uptake in the northern North Atlantic is greatly intensified, while the Southern Ocean remains similar to HIST. Moreover, the long-term trends of Q_{net} and SST (Fig. 2) in the Southern Ocean and SPNA are broadly of opposite sign. That is, where SST cools in the SPNA and is neutral in the high southern latitudes, the Q_{net} trend is positive, toward less cooling. Where SST has warmed fastest, in the midlatitude Gulf Stream, the Kuroshio, and the Brazil

FIG. 1. Trend of Q_{net} ($\text{W m}^{-2} \text{ decade}^{-1}$) of the ensemble mean of 9 CMIP5 models in the (a) GHG runs (1861–2005), (b) AERO runs (1861–2005), and (c) HIST runs (1861–2005). Positive indicates excess heat absorbed by the ocean. Stippling indicates regions exceeding 95% statistical significance computed from the two-tailed t test.

and Agulhas Currents, the trend in Q_{net} is to greater cooling. The change of Q_{net} is thus damping the SST change in these critical regions, as shown by [Armour et al. \(2016\)](#). This correlation suggests that the air–sea

heat flux Q_{net} is a strong function of the air–sea temperature difference, that that difference is largely due to changes in SST rather than in air temperature, and that the SST changes are due to anomalous advection.

FIG. 2. Trend of SST ($^{\circ}\text{C} \text{ decade}^{-1}$) of ensemble mean of 9 CMIP5 models in the (a) GHG runs (1861–2005), (b) AERO runs (1861–2005), and (c) HIST runs (1861–2005). Positive indicates warming of surface seawater.

FIG. 3. Time series of surface heat uptake area integrated over the (a) SO (south of 30°S) and (b) NA (north of 30°N) in the twentieth century from the ensemble mean of CMIP5 models. Different colors show the responses of HIST (black), GHG (red), and AERO (blue) runs. The shadings denote model uncertainties as one standard deviation across models. An 11-yr running average is applied to the time series.

In the GHG runs, the Southern Ocean and the northern North Atlantic are the most active regions that take up the anthropogenic heat (Fig. 1a). Elsewhere the trend in Q_{net} is small and not well organized in space. To compare the trends in Southern Ocean (SO) and northern North Atlantic (NA) heat uptake poleward of 30° latitude quantitatively, the surface heat flux is integrated over the SO (south of 30°S) and NA (30° – 70°N , 80° – 10°W). The time evolution of heat flux for each experiment referenced to the 1861–80 average is shown in Fig. 3. For the SO over the twentieth century, the long-term heat flux trend from HIST runs (black) is $0.128 \text{ PW century}^{-1}$, which is close to the result from GHG runs (red; $0.178 \text{ PW century}^{-1}$). HIST runs capture the cooling effects of volcanic aerosols following volcano eruptions (Agung in 1963, El Chichón in 1982, and Pinatubo in 1991) (Ding et al. 2014; Smith et al. 2016), which are not included in the GHG or AERO runs. In contrast with the SO, for the NA, heat uptake is

large in GHG ($0.091 \text{ PW century}^{-1}$) but is small ($0.030 \text{ PW century}^{-1}$) in HIST, which includes all radiative forcings.

The time series of cumulative heat uptake since 1861 (not shown) is similar to the time series of heat uptake. In HIST runs during the twentieth century, the global ocean takes up 263 ± 102 zettajoules (ZJ; $1 \text{ ZJ} = 10^{21} \text{ J}$) and the SO takes up 190 ± 81 ZJ, accounting for $72\% \pm 28\%$ of global ocean heat uptake, consistent with Frölicher et al. (2015). During the same period, the NA takes up 16 ± 82 ZJ, contributing only 6% of global heat uptake. The ratio of cumulative heat uptake between SO and NA is 11.8 ± 13.9 . Thus, the Southern Ocean dominates ocean heat uptake in the historical period. In GHG runs, the global ocean takes up 570 ± 152 ZJ, which is double the global ocean heat uptake in the HIST runs. The SO maintains about the same level of heat uptake (255 ± 94 ZJ) while the NA takes up 137 ± 41 ZJ, accounting for $24\% \pm 11\%$ of global cumulative ocean heat uptake. The NA heat uptake in GHG runs increases by around 7.5 times relative to HIST runs. The discrepancy of regional cumulative heat uptake between HIST and GHG indicates the influence of factors other than GHG.

b. Compensation between GHG and anthropogenic aerosols

Anthropogenic aerosols are an important radiative forcing that cools the global climate (Forster et al. 2007). The global map of long-term trend of Q_{net} due to the anthropogenic aerosol radiative forcing (AERO) is shown in Fig. 1b. The geographically distributed anthropogenic aerosols cause enhanced SST cooling (Fig. 2b) with more heat loss in the NH (Fig. 1b) than the Southern Hemisphere (SH). The intensive heat loss takes place in the NA rather than the North Pacific even though aerosols are prevalent throughout the NH. This is primarily due to the greater sensitivity of the SPNA, where the mean mixed layers are about 10 times deeper than in the North Pacific (Holte et al. 2017). The Q_{net} trend pattern in AERO runs broadly opposes that in GHG runs (Figs. 1a,b). The compensation between GHGs and aerosols is remarkable in the SPNA. This compensation is also clear in the zonally integrated trend in Q_{net} from the CMIP5 ensemble mean (Fig. 4). The cross-correlation in the zonally integrated pattern is -0.60 between AERO and GHG runs, illustrating an overall similarity in the responses of ocean heat uptake to GHG and aerosol forcings (Xie et al. 2013, 2015). There is a linear relationship between regional ocean heat uptake and global radiative forcing shown in Fig. S1 in the online supplemental material. We sum the Q_{net} trends from GHG and AERO runs (defined as

FIG. 4. Zonally integrated Q_{net} trend ($\text{TW lat}^{-1} \text{ decade}^{-1}$) in the GHG (red), AERO (blue), and HIST (black) runs from 1861 to 2005. The brown curve denotes the linear combination of responses of GHG and AERO (GHG+AERO). The top-right corner indicates the correlation coefficient between heat flux responses of the GHG and AERO runs in the first row and HIST and GHG+AERO in the second row. Positive represents the heat gain of the ocean. Similar calculations using the smaller ensemble of ozone runs are shown in Fig. S2.

GHG+AERO), denoted by the brown curve (Fig. 4). The cross-correlation in meridional pattern between GHG+AERO and HIST is very high (0.95), indicating that oceanic heat uptake is forced to first order by GHGs and anthropogenic aerosols. It also suggests that the sum is a good approximation for the all-forcing simulations.

Tables 3 and 4 compare cumulative ocean heat uptake in the SO and NA in total amount and percentage to the global uptake from various runs. The SO to NA ratio of uptake is about 2:1 for GHG (warming), 1:2.6 for AERO (cooling), and 12:1 in the HIST (warming). The dominant SO uptake in HIST is a result of compensating NA uptake in response to GHG (137 ZJ) and AERO (−134 ZJ). The results from GHG+AERO and HIST are particularly close, implying that the historical response of heat uptake is mainly due to the

combination of GHG and aerosol radiative forcing. Therefore we hypothesize that the impact of anthropogenic aerosols on ocean heat uptake accounts for much of the discrepancy between the HIST and GHG runs.

Other forcing agents, such as ozone, may also affect the pattern and magnitude of ocean heat uptake. The zonally integrated Q_{net} trend of ozone runs is represented by the green curve (the ensemble mean of 3 CMIP5 models) in Fig. S2. Here the result represents the effect of ozone concentration from both stratosphere and troposphere. The global average response of Q_{net} trend is smaller in the single forcing OZONE runs than in GHG and AERO runs. The Southern Ocean, however, shows marked heat uptake: enhanced westerly wind stress in the Southern Ocean due to stratospheric

TABLE 3. Cumulative ocean heat uptake (ZJ) in the SO (south of 30°S), NA (30°–70°N, 80°–10°W) and global ocean simulated by the CMIP5 models. Values in parentheses indicate the periods used to calculate the change of ocean heat uptake for different runs. The model uncertainties are estimated as one standard deviation across the models. CSIRO-Mk3.6.0 is excluded in HIST runs due to the negative trend of global cumulative heat uptake.

	GHG (1900–2000)	AERO (1900–2000)	GHG+AERO (1900–2000)	HIST (1900–2000)	RCP4.5 (2006–2100)	RCP8.5 (2006–2100)
SO	255 ± 94	−52 ± 62	202 ± 116	190 ± 81	829 ± 145	1187 ± 208
NA	137 ± 41	−134 ± 71	3 ± 62	16 ± 82	443 ± 114	632 ± 115
Global	570 ± 152	−283 ± 153	287 ± 156	263 ± 102	1580 ± 114	2468 ± 179

TABLE 4. The percentage of the SO and NA (defined as in Table 3) relative to global ocean cumulative heat uptake simulated by the CMIP5 models. The table structure is as in Table 3.

Percentage	GHG (1900–2000)	AERO (1900–2000)	GHG+AERO (1900–2000)	HIST (1900–2000)	RCP4.5 (2006–2100)	RCP8.5 (2006–2100)
SO/Global	45% $\pm$ 10%	18% $\pm$ 25%	71% $\pm$ 51%	72% $\pm$ 28%	52% $\pm$ 9%	48% $\pm$ 8%
NA/Global	24% $\pm$ 11%	47% $\pm$ 26%	1% $\pm$ 42%	6% $\pm$ 39%	28% $\pm$ 7%	26% $\pm$ 6%

ozone depletion (Thompson et al. 2011; Ferreira et al. 2015) causes enhanced equatorward Ekman transport and hence this heat uptake. Furthermore, we find that the Q_{net} trend is smaller in ozone runs than that in GHG and AERO runs in the midlatitudes of the NH. We also sum the Q_{net} trends from GHG, AERO, and OZONE runs (defined as GHG+AERO+OZONE) (Fig. S2). We note an offset between HIST and GHG+AERO+OZONE, which may result from the nonlinear effect of forcing agents when prescribed together, but could instead be due to the small number of CMIP5 ozone single-forcing experiments.

4. Future projections

The global aerosol optical depth (AOD) and CO_2 concentrations used for the historical (HIST) and future projection (RCP8.5) runs of GFDL-CM3 over the twentieth and twenty-first century are shown in Fig. 5. Anthropogenic aerosol forcing, represented by AOD, in CMIP5 historical simulations increases rapidly over the twentieth century due to industrial development. It peaks at the beginning of the twenty-first century, and then is projected to decline during the RCP scenarios as countries begin to restrict aerosols out of the need to improve air quality and protect human health (Westervelt et al. 2015; Rotstayn et al. 2013). Unlike aerosols, GHGs including CO_2 are projected to increase steadily. Thus, aerosol forcing offsets the warming effect of GHGs to some extent over the historical period; its reduction amplifies the GHG warming relative to the historical period in future projections (Xie et al. 2015).

The Q_{net} trends projected until 2050 in the RCP4.5 and RCP8.5 scenarios are shown in Fig. 6. The zonally integrated patterns of Q_{net} trend (Fig. 6c) differ markedly from HIST (Fig. 4) in the NH extratropics. The biggest difference is that heat uptake is projected to greatly intensify in the northern North Atlantic (Figs. 6a,b vs Fig. 1c). Comparison of the time series of NA and SO ocean heat uptake over the period from 1900 to 2100 (Fig. 7a) shows a rapid increase in heat uptake takes place in the NA (solid curves) at the beginning of the twenty-first century, while the SO heat uptake continues its historical rise. Given the steady rise of CO_2 concentration, this rapid increase in NA uptake

is associated with the change of anthropogenic aerosol forcing. With anthropogenic aerosol emissions declining in future projections, the NA heat uptake increases to the same order of magnitude as the SO heat uptake (Fig. 7a), despite its much smaller area. In the future, more and more excess heat is projected to be stored in the ocean through the North Atlantic.

Heat flux trends in the SO and NA in various runs and their ensemble means are shown in Fig. 7b, separated according to single-forcing GHG and AERO runs, as well as the HIST and projected RCP4.5 and RCP8.5 runs. The relative trends in the SO versus the NA for the GHG and AERO are shown by the straight line fits. These straight lines envelop the future scenarios (RCP4.5 and RCP8.5) despite the large spread across models, which implies that GHG and AERO remain the dominant forcings in the future projections. According to Tables 3 and 4, from 2006 to 2100 the percentage of global heat uptake contributed by the SO decreases to 52% $\pm$ 9% (48% $\pm$ 8%) for the RCP4.5 (RCP8.5). The contribution of the NA increases to 28% $\pm$ 7% for the RCP4.5 and 26% $\pm$ 6% for the RCP8.5. The SO and NA ratio of cumulative heat uptake is about 2:1, which is much smaller than that during the historical period (12:1). That is, the relative importance of the SO and NA in anthropogenic heat uptake evolves due to the very different trajectories of GHGs (increasing) and anthropogenic aerosols (decreasing).

FIG. 5. Global CO_2 concentration (red curve; ppm) and global mean ambient AOD at 550 nm (blue curve) from 1900 to 2100. Data are from the HIST (1900–2005) and RCP8.5 (2006–2100) runs of GFDL-CM3.

FIG. 6. The Q_{net} trend ($\text{W m}^{-2} \text{ decade}^{-1}$) of the ensemble mean of 9 CMIP5 models in (a) RCP4.5 (2006–50) and (b) RCP8.5 (2006–50). Positive indicates excess heat absorbed by the ocean. Stippling indicates regions exceeding 95% statistical significance computed from the two-tailed t test. (c) Zonally integrated Q_{net} trend in RCP4.5 (orange) and RCP8.5 (purple) from 2006 to 2050.

5. AMOC responses to anthropogenic forcing

The long-term mean (1960–2005) meridional overturning streamfunction is illustrated in Fig. 8, for the Atlantic Ocean north of 30°S and for the Southern Ocean south of 30°S, based here on the first realization of the LENS. The Atlantic meridional overturning circulation index is defined as the maximum streamfunction at 30°N. The multimodel ensemble mean change in AMOC intensity, relative to the average over 1861–80 in each model, is shown in Fig. 9. The AMOC has almost no trend ($-0.03 \text{ Sv decade}^{-1}$; $1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$), only decadal variability, in the HIST runs covering the twentieth century. In contrast, the GHG and AERO runs weakened ($-0.21 \text{ Sv decade}^{-1}$) and strengthened ($0.19 \text{ Sv decade}^{-1}$) the AMOC, respectively, hence nearly compensating each other, as seen above for North Atlantic heat uptake (Delworth and

Dixon 2006). The robustness of this result is apparent in the eight CMIP5 models that are averaged in Fig. 9, six of which similarly show only a small AMOC trend, both positive and negative ($-0.1 < \text{AMOC trend} < 0.1 \text{ Sv decade}^{-1}$), in their HIST runs, despite large variation (from ~ 10 to $\sim 30 \text{ Sv}$) in the mean AMOC intensity among models (Fig. S3). Thus, even though the global GHG radiative forcing is larger than the aerosol forcing in magnitude (from about 2:1 to 3:1), their impacts on AMOC strength are comparable.

In future projections, with continuing increase in GHG but decrease in aerosol forcing, the AMOC slows down dramatically in RCP4.5 ($-0.42 \text{ Sv decade}^{-1}$ from 2006 to 2050) and RCP8.5 ($-0.67 \text{ Sv decade}^{-1}$ from 2006 to 2100) (Fig. 9). These trends are much larger than the trend in the historical period because the GHGs are not compensated by aerosols.

FIG. 7. Time series of (a) heat uptake in the NA (30° – 70° N, 80° – 10° W; solid curves) and SO (south of 30° S; dashed curves) relative to the average of 1861–80 for various ensembles of CMIP5. Shading denotes model uncertainties as one standard deviation across models. (b) SO vs NA in heat uptake trend. The multimodel means are represented by hollow circles, with each model run denoted by small dots. The dashed red (blue) lines for the historical single forcing runs GHG (AERO) are plotted through the multimodel means (hollow circles), with their slopes based on all runs in the ensemble. The ensemble mean responses from LENS (purple) and 2005AERO (green) are also shown in the scatterplot (filled circle), with individual members denoted by smaller dots.

The rapid slowdown of the AMOC in RCP4.5 and RCP8.5 is consistent with the rapid increase of the NA heat uptake (Fig. 7a), suggesting the influence of ocean circulation on the regional ocean heat uptake. When the climate warms due to GHG increase, surface temperature increases and salinity decreases (not shown, but due to increased precipitation and sea ice melt), which together decrease the upper ocean density. This strengthens the stratification and hence weakens deep convection and the AMOC intensity. When the AMOC slows, poleward heat transport diminishes over the North Atlantic, resulting in a cooling tendency in SPNA SST (Winton et al. 2013; Marshall et al. 2015; Drijfhout et al. 2012; Buckley and Marshall 2016). The decreasing SPNA SST is associated with intensified surface heat flux into the ocean (Rugenstein et al. 2013; Winton et al. 2013; Gregory et al. 2016). When more excess heat is absorbed by the SPNA, with increasing precipitation and sea ice melt, surface water in the SPNA becomes even less dense, increasing the stability of the water column and further weakening the AMOC (Delworth and Dixon 2006; Cheng et al. 2013; Menary et al. 2013). Thus there is a positive feedback between the SPNA surface heat flux and the AMOC intensity. Therefore it appears that the enhanced SPNA heat uptake is strongly tied to the change of ocean circulation, in particular the AMOC. The processes that govern heat uptake changes in the SO under anthropogenic forcings are entirely different, primarily affected by the climatological

currents (Marshall et al. 2015; Morrison et al. 2016; Liu et al. 2018).

6. Effect of anthropogenic aerosol reduction in future projections

To further investigate the effect of projected declining aerosols on regional heat uptake, we use two experiments from CESM: LENS and AERO2005 (section 2a).

FIG. 8. Long-term mean (from 1960 to 2005) meridional overturning streamfunction (Sv) in the Atlantic Ocean (north of 30° S) and in the Southern Ocean (south of 30° S) from the first realization of LENS. The streamfunction is calculated based on the seawater meridional velocity and is zonally integrated in the Atlantic and Southern Ocean, respectively. Red shading denotes clockwise circulation; blue shading denotes anticlockwise circulation. The AMOC index is defined as the maximum streamfunction at 30° N (dashed line).

FIG. 9. Time series of the AMOC intensity (Sv) in the CMIP5 multimodel ensemble runs listed in the figure. AMOC intensity is defined as the maximum volume transport streamfunction at 30°N and is calculated based on the multimodel mean of the first realization of each model relative to the average of 1861–80. The number in the parentheses denotes the long-term trend (dashed line) in each experiment.

The differences between the responses from 2005AERO, which fixes aerosols at the 2005 level, and RCP8.5 of LENS, which projects a decline in aerosols, reveal the effect of the reduced aerosols and their precursors (such as SO_2 and NO_2) in future projections. The long-term trends (2006–2100) of Q_{net} and SST from LENS and 2005AERO, and the difference between them, are shown in Fig. 10. For both LENS and 2005AERO, Q_{net} trend features an intense, positive heat uptake change in the NA (Figs. 10a,b). Ocean heat uptake is locally enhanced where SST warming (contours) is reduced, illustrating that surface heat flux is mainly driven by the ocean on multidecadal time scales (Gulev et al. 2013). For 2005AERO, the increased heat uptake is primarily attributed to continued GHG increase. Subtracting the Q_{net} response in 2005AERO runs from that in LENS (Fig. 10c), we find a remarkable positive trend in the SPNA, while changes are negligible elsewhere. Thus, the intensified heat uptake in the SPNA in LENS is partly attributed to the aerosol reduction.

Area-integrated heat uptake and cumulative heat uptake are calculated to quantify the contribution of the declining aerosols (Fig. 11). From 2006 to 2100, the NA heat flux trend is $0.428 \text{ PW century}^{-1}$ in RCP8.5 of LENS while only $0.287 \text{ PW century}^{-1}$ in 2005AERO (Fig. 11a). Despite the dominance of GHGs in radiative forcing in RCP8.5, one-third of that heat flux trend in the NA is due to the declining anthropogenic aerosols because the aerosol forcing is mostly restricted to the NH.

At the end of the twenty-first century, the heat uptake rate by the NA catches up with that of the SO in RCP8.5.

In the twentieth century, the time evolution of cumulative heat uptake of LENS shows small changes in the NA ($3 \pm 21 \text{ ZJ}$) but a remarkable increase in the SO ($190 \pm 11 \text{ ZJ}$) (Fig. 11b), consistent with the CMIP5 results (Table 3). The large standard deviation in the NA shows that internal variability is an important factor affecting the regional heat uptake. In RCP8.5 of LENS, at the end of the twenty-first century, the cumulative heat uptake change in the NA ($712 \pm 22 \text{ ZJ}$) is comparable to that in the SO ($964 \pm 13 \text{ ZJ}$). In 2005AERO, the cumulative heat uptake in the NA decreases to $503 \pm 17 \text{ ZJ}$. The reduced aerosols account for 29% of cumulative heat uptake in the NA in RCP8.5. In addition, the global ocean heat uptake decreases by 15% if aerosol emissions are fixed at 2005 levels. We can further deduce that with a weaker GHG forcing than RCP8.5 (such as RCP2.6), the relative contribution of the declining anthropogenic aerosols to heat uptake will become larger.

Moreover, in 2005AERO with aerosol fixed instead of declining from 2006 to 2100, the AMOC slows down by $-0.53 \text{ Sv decade}^{-1}$, less than half of the result from the LENS, in which aerosols decline ($-1.23 \text{ Sv decade}^{-1}$) (Fig. 11c). This shows that the declining anthropogenic aerosols permit more rapid slowdown of the AMOC (Menary et al. 2013) and result in enhanced heat uptake in the NA in the future.

FIG. 10. Trend of Q_{net} (shading; $\text{W m}^{-2} \text{ decade}^{-1}$) and SST (contours at $0.1 \text{ K decade}^{-1}$; dashed contours indicate negative trends) of (a) the ensemble mean of LENS, (b) the ensemble mean of 2005AERO, and (c) LENS minus 2005AERO from 2006 to 2100.

7. Comparison with observations of ocean heat content

For comparison with observations, we primarily use OHC rather than Q_{net} because OHC is based on subsurface temperature profiles, which are measured with high accuracy, while surface heat fluxes are indirectly calculated and suffer from large uncertainties in algorithm (Valdivieso et al. 2017). OHC change is determined by Q_{net} globally while the differences between OHC and Q_{net} trend patterns are due to ocean circulation. Note that our purpose is not to analyze the ocean heat budget, which is closed in the models, but rather to assess the model representation of ocean heat.

The spatially integrated, regional OHC changes in the upper 2000 m in the NA north of 30°N and SO south of 30°S are shown in Fig. 12, based on observations (Argo, IAP, EN4-L09, and EN4-G10) and simulations (LENS). In the NA, the observed OHC trend is relatively flat from 1960 to 1996, followed by a sharp rise afterward in all datasets, and then a decrease since 2006. In the SO, OHC features a robust rise after 1960. The contrast in OHC change between the NA and SO is consistent with the change of cumulative heat uptake obtained from the aforementioned simulations.

External forcing and internal variability as well as natural forcing affect the decadal-to-multidecadal climate change (Hansen et al. 2011). For instance, Terray (2012) concludes that the anthropogenic forcings account for the warming trend in the North Atlantic SST over the last three decades, while the internal variability is very strong in the SPNA, contributing to the multidecadal SST swings. Here we focus on how the external forcing and internal variability affect the OHC change in the NA. To isolate the contributions of forced and internal variability, we apply an EOF analysis to the OHC integrated over the upper 2000 m in the NA. The first two modes based on the IAP data are shown in Fig. 13. The leading mode explains about 43% of the total variance, with a dipole pattern between the subpolar and subtropical North Atlantic. The leading principal component (PC1) shows two major phase shifts during 1950–2015 (Fig. 13c), which is synchronized with the North Atlantic Oscillation (NAO) index from NOAA (<https://www.ncdc.noaa.gov/teleconnections/nao/>). The Atlantic multidecadal oscillation (AMO) index from NOAA (<https://www.esrl.noaa.gov/psd/data/timeseries/AMO/>) is also shown in Fig. 13b. The cross-correlation between PC1 and NAO is 0.77, suggesting that the first mode of atmospheric circulation significantly affects the OHC change in the North Atlantic north of 30°N (McCarthy et al. 2015; Delworth et al. 2017). The cooling of the SPNA and the warming of the subtropical gyre is

FIG. 11. Time series of (a) heat uptake and (b) cumulative heat uptake in the NA (orange and green) and SO (blue) relative to the average of 1861–80 in LENS and 2005AERO. The cumulative heat uptake in the SO is relative to the annual mean of 2006. (c) Time series of the annual mean AMOC intensity in the first realization of LENS (black) and 2005AERO (orange). For LENS, AMOC intensity is relative to the average of 1861–80. For 2005AERO, the annually AMOC index anomaly is relative to the annual mean of 2006.

associated with the positive phase of NAO (Lozier et al. 2008; Zhang and Yan 2017). Specifically, the enhanced westerlies in the SPNA during positive NAO drive increased heat loss (Duchez et al. 2016; Robson et al. 2016). The second principal component (PC2), which explains about 28% of the variance, is nearly

constant from 1960 to 1996, followed by a rapid rise. EOF2 features enhanced warming along the Gulf Stream and a cooling in the SPNA, resembling the trend of the ensemble mean of LENS (Fig. 15b). Therefore, EOF2 seems to reflect the response to external radiative forcing. PC2 suggests that the northern North Atlantic has become more active in anthropogenic heat uptake during the most recent two decades. We have also applied the EOF analysis to the OHC of EN4-G10 (Fig. S4). The spatial patterns and PCs are similar to those of the IAP product.

Based on the second mode of the EOF analysis (Fig. 13d), we select two 37-yr periods for the OHC change comparison: 1) 1960–96, when PC2 is relatively flat, and 2) 1979–2015, which includes a rapid increase of OHC and the NAO effect is nearly averaged out based on Fig. 13c. The comparison of the OHC trend between IAP data and the LENS ensemble mean for the first period, 1960–96, is shown in Fig. 14. The pattern of LENS ensemble mean behaves as a reference here, representing the external forcing effects on OHC change. There is broad agreement in the Southern Ocean where the zonally integrated heat gain from both IAP data and LENS shows a pronounced trend (Figs. 14c,d). The upper 2000 m of the SO accounts for about 74% (103%) of global heat storage change in IAP data (the ensemble mean of LENS). During this period, major differences between the observations and simulation are found in the North Atlantic. This is due to the internal multidecadal variability, captured in the observed EOF1 (Fig. 13c), which overwhelms the external forcing signal. For the second period of 1979–2015 (Fig. 15), in the NA the external forcing signal increases rapidly (Fig. 13d), leading to a different spatial pattern, which is similar to the pattern of LENS ensemble mean (Fig. 15b). The relative importance of the SO in global heat content change decreases to 39% and 51% in IAP data and LENS, respectively (Figs. 15c,d). Thus we show that the external forcing signal is stronger and detectable in recent decades. In addition, IAP data show much more tropical warming relative to the LENS ensemble, which seems to be related to internal variability. The largest discrepancies of OHC change between IAP and EN4 data are found in the Southern Ocean from 1960 to 1996 (cf. Fig. 14c and Fig. S5c), which is probably due to sparse observations over the period (Wang et al. 2018).

In future projections (RCP8.5 from 2015 to 2100), the North Atlantic and Arctic show remarkable increases in heat storage (Fig. 16a). During this period, the Southern Ocean only explains about 32% of global OHC change. This is comparable to the percentage of global anthropogenic heat absorbed by the whole Atlantic north of

North Atlantic

Southern Ocean

FIG. 12. Time series of OHC (0–2000 m) from observations (Argo, IAP, and EN4) and simulation (LENS) in the (a) NA and (b) SO. All the time series are relative to the 1975–2012 base period. The gray curves denote the OHC from each individual member of LENS. A 5-yr running mean is applied to the time series.

30°S (24%). Although the Southern Ocean OHC trend increases in recent decades (Fig. 17a), the percentage of the Southern Ocean to global OHC change decreases (Fig. 17b).

Heat transport due to ocean circulation (as represented by Fig. 7) can strongly affect the heat storage pattern (Winton et al. 2013; Frölicher et al. 2015; Tamsitt

et al. 2016) and is responsible for the discrepancies between the ocean heat storage and uptake patterns in the same regions (Frölicher et al. 2015; Armour et al. 2016). First of all, the SO peak is at 55°S in the heat uptake pattern (Fig. 4) but around 42°–45°S in heat content change pattern (Figs. 14c,d). The displacement is primarily due to advection by the equatorward Ekman

FIG. 13. The first two EOF modes of OHC above 2000 m from 1950 to 2015 from IAP data. Shown are (a),(b) EOF patterns and (c),(d) normalized PCs. A 5-yr running mean is applied to the PCs, NAO (blue), and AMO (green). The number at the upper right corner of (c) shows the cross-correlation between PC1 and NAO. The NAO index is from NOAA (<https://www.ncdc.noaa.gov/teleconnections/nao/>). The AMO index is from NOAA (<https://www.esrl.noaa.gov/psd/data/timeseries/AMO/>). The NAO and AMO indices are normalized. Compare with the EN4-G10 data analysis in Fig. S4.

transport (Tamsitt et al. 2016; Armour et al. 2016). The smaller percentage of the SO to global OHC change (32%) than the values in cumulative heat uptake shown in Table 4 may also be attributed to ocean adjustment. Furthermore, the much smoother OHC trend (Fig. 15d) with a nearly uniform increase in the top 300-m layer is most likely due to ocean ventilation. More work is needed to investigate the response of ocean heat uptake and storage to ocean circulation in a two-dimensional (latitude–longitude space) or three-dimensional (latitude–longitude–depth space) sense.

8. Summary

The Southern Ocean accounts for $72\% \pm 28\%$ of global heat uptake in historical runs during the twentieth century in the CMIP5 ensemble. We have shown that the Southern Ocean's historical dominance relative to the same latitude range of the North Atlantic is due to

the compensating effects of anthropogenic aerosol cooling and GHG warming on the surface heat flux for the first order over the North Atlantic and AMOC intensity. In GHG single-forcing simulations for the historical period, the Southern Ocean's contribution to global uptake is much smaller at $45\% \pm 10\%$, with a large contribution from the northern North Atlantic ($24\% \pm 11\%$). Anthropogenic aerosols preferentially cool the NH, causing a large heat loss in the northern North Atlantic that opposes the GHG heating. The compensating aerosol effect is small over the Southern Ocean, and therefore this region dominates the historical global anthropogenic heat uptake, which is driven mainly by GHGs with a much smaller contribution from ozone (Fig. S2).

This pattern of historical heat uptake cannot be extrapolated into the future, however, as anthropogenic aerosols are projected to decrease due to environmental regulation. In future projections, the Southern Ocean

FIG. 14. The 1960–96 trends in OHC (denoted by $\text{OHC}_{\text{trend}}$) above 2000 m from (a) IAP data and (b) the ensemble mean of LENS. Also shown is the zonally integrated OHC trend from (c) IAP data and (d) the ensemble mean of LENS, in different ocean layers. The number at the upper right corner denotes the fraction of SO to global OHC change. Compare with the EN4-G10 data analysis in Fig. S5.

continues to absorb heat at an increasing rate, but the North Atlantic greatly increases its heat uptake due to the projected reduction in aerosol forcing. The consequence is that the Southern Ocean's percent contribution to global heat uptake decreases to about 50% while the northern North Atlantic uptake increases to about 27%. Using simulations from CESM, we show that the anticipated reduction in anthropogenic aerosols causes surface heat uptake to increase by about 30% in the northern North Atlantic and about 15% globally. The varying relative importance between the Southern Ocean and North Atlantic heat uptake between historical and future periods is determined by the different spatial distributions and trajectories of GHG and anthropogenic aerosol radiative forcing.

Heat losses (gains) in the SPNA are accompanied by strengthening (weakening) of the AMOC in the AERO (GHG) runs. In the GHG runs, as an example, less heat is advected to the high latitude North Atlantic due to the weakening AMOC intensity (Winton et al.

2013), leading to lower SSTs in the SPNA. This cooler surface water results in less heat loss from the ocean to the atmosphere, increasing the stratification of the upper ocean, which then further weakens the AMOC intensity. This positive feedback between AMOC and North Atlantic heat uptake sustains the weakening of AMOC intensity and intensification of regional heat uptake (Gregory et al. 2016; Buckley and Marshall 2016). This illustrates that ocean circulation change induced by anthropogenic forcing can strongly affect the pattern of heat uptake in the North Atlantic, while the Southern Ocean heat uptake is primarily driven by background ocean circulation (Marshall et al. 2015; Morrison et al. 2016; Liu et al. 2018). For instance, Liu et al. (2018) demonstrate that the background ocean circulation accounts for about 80% of Southern Ocean heat storage change, while the remaining 20% is due to wind-induced ocean circulation change.

It is likely that, in the late twentieth century, the lack of a significant observed decreasing trend in the AMOC, despite an increase in global ocean heat uptake, is due

(a) Observation: $\text{OHC}_{\text{trend}}$ of IAP (1979–2015) (b) Model: $\text{OHC}_{\text{trend}}$ of LENS (1979–2015)

(c) IAP: 1979–2015

(d) LENS: 1979–2015

FIG. 15. As in Fig. 14, but the OHC trend is calculated for the period from 1979 to 2015. Also compare with the EN4-G10 data analysis in Fig. S5.

to the strongly moderating influence of anthropogenic aerosols, which preferentially affect the North Atlantic. The recent observations of the AMOC intensity from the Rapid Climate Change (RAPID) array (Cunningham et al. 2007) at 26.5°N show that the AMOC has weakened substantially over the past decade

since 2004, consistent with from model simulations. However, the observed magnitude of the decreasing trend (about -0.5 Sv yr^{-1}) from 2004 to 2012 (Smeed et al. 2014; Roberts et al. 2014) is much larger than that projected to occur in the twenty-first century (about -0.1 Sv yr^{-1} ; Cheng et al. 2013). The internal AMOC variability

(a) Model: $\text{OHC}_{\text{trend}}$ of LENS (2015–2100)

(b) LENS: 2015–2100

FIG. 16. (a) OHC trend (denoted by $\text{OHC}_{\text{trend}}$) above 2000 m from the ensemble mean of LENS over 2015–2100. (b) Zonally integrated OHC trend in different layers, as in Fig. 14. The number at the top-right corner denotes the fraction of SO to global OHC change.

FIG. 17. (a) OHC trend above 2000 m in the SO and (b) fraction of the SO to global OHC change from observations (IAP and EN4-G10) and model (the ensemble mean of LENS). The error bars denote the 95% confidence interval.

seems to be responsible for the observed slowdown of the AMOC over the past decade, but the anthropogenic forcing effect cannot be ruled out (Smeed et al. 2014; Jackson et al. 2016). At least one decade of continuous observations is required to diagnose the influence of the external climate forcing on the AMOC (Roberts et al. 2014). Such observations will also test the projected increase in anthropogenic heat uptake in the subpolar North Atlantic.

GHG radiative forcing by itself fails to explain the pattern of historical ocean heat uptake change. Our study reveals the important role of anthropogenic aerosols relative to greenhouse gases in regional ocean heat uptake, but the large uncertainty in the magnitude and spatial pattern of anthropogenic aerosol forcing (Myhre et al. 2013; Rotstayn et al. 2015; Hansen et al. 2011) is a barrier to fully quantifying its impact on global and regional climate change. Further studies are necessary to improve the observation, understanding, and simulation of the complex aerosol effects on climate change. Another useful extension of our work should include quantifying the role of ocean circulation in high-latitude heat budgets.

Acknowledgments. J.-R. Shi is supported by the U.S. National Science Foundation (AGS-1637450) and the Southern Ocean Carbon and Climate Observations and Modeling project (SOCCOM) under the National Science Foundation Award (PLR-1425989). We acknowledge the World Climate Research Programme's Working Group on Coupled Modeling, which is responsible for CMIP5, and we thank the climate modeling groups for producing and making available the model output. The CESM LENS and 2005AERO simulations are available on the Earth System Grid (www.earthsystemgrid.org). The Argo data used here were collected and made freely available by the International Argo Program and by the national programs that contribute to it (<http://argo.ucsd.edu>).

The IAP product is available on the Climate Data Guide website (<https://climatedataguide.ucar.edu/climate-data/ocean-temperature-analysis-and-heat-content-estimate-institute-atmospheric-physics>). EN4 products are available on the Met Office Hadley Centre website (<https://www.metoffice.gov.uk/hadobs/en4/download-en4-2-0.html>).

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