1

Multidecadal Ocean Temperature and Salinity Variability in the Tropical North Atlantic:

Linking with the AMO, AMOC and Subtropical Cell

4

Chunzai Wang [1]

Liping Zhang [2 & 1]

7

8

[1] NOAA Atlantic Oceanographic and Meteorological Laboratory

Miami, Florida

11

[2] Cooperative Institute for Marine and Atmospheric Studies

University of Miami

Miami, Florida

15

16

19

20

Corresponding author address: Dr. Liping Zhang, Physical Oceanography Division,
NOAA/Atlantic Oceanographic and Meteorological Laboratory, 4301 Rickenbacker Causeway,
Miami, FL 33149.   E-mail: Liping.Zhang@noaa.gov.

24

Abstract

The Atlantic multidecadal oscillation (AMO) is characterized by the sea surface warming

(cooling) of the entire North Atlantic during its warm (cold) phase.   Both observations and most

of CMIP5 models also show that the warm (cold) phase of the AMO is associated with a surface

warming (cooling) and a subsurface cooling (warming) in the tropical North Atlantic (TNA)

Ocean.   It is further shown that the warm phase of the AMO corresponds to a strengthening of

the Atlantic meridional overturning circulation (AMOC) and a weakening of the Atlantic

subtropical cell (STC), both of which induce an anomalous northward current in the TNA

subsurface ocean.   Because the mean meridional temperature gradient of the subsurface ocean

is positive due to the temperature dome around 9°N, the advection by the anomalous northward

current cools the TNA subsurface ocean during the warm phase of the AMO.   The opposite is

true during the cold phase of the AMO.   It is concluded that the anticorrelated ocean

temperature variation in the TNA associated with the AMO is caused by the meridional current

variation induced by variability of the AMOC and STC, but the AMOC plays a more important

role than the STC.   Observations do not seem to show an obvious anticorrelated salinity relation

between the TNA surface and subsurface oceans, but most of CMIP5 models simulate an

out-of-phase salinity variation.   Similar to the temperature variation, the mechanism is the

salinity advection by the meridional current variation induced by the AMOC and STC associated

with the AMO.

1. Introduction

One of the most important climate variations in the Atlantic is the Atlantic multidecadal oscillation (AMO) which is a basin-wide mode in the entire North Atlantic and is defined by the North Atlantic sea surface temperature (SST) anomalies (e.g., Enfield et al. 2001; Knight et al. 2005). The AMO has significant regional and global climate associations, such as the Northeast Brazilian and African Sahel rainfall (Folland et al. 1986; Rowell et al. 1995; Folland et al. 2001; Rowell 2003; Wang et al. 2012), the Atlantic warm pool (Wang et al. 2008; Zhang et al. 2012a; Wang et al. 2013) and Atlantic hurricanes (Goldenberg et al. 2001; Wang and Lee 2009), and North American and European summer climate (Enfield et al. 2001; McCabe et al. 2004; Sutton and Hodson 2005). In spite of its importance, the mechanism of the AMO is still unclear. One of the most popular arguments is that the AMO is induced by the Atlantic meridional overturning circulation (AMOC) variations and associated heat transport fluctuations (Folland et al. 1986; Gray et al. 1997; Delworth and Mann 2000; Knight et al. 2005). Some of modeling studies indicate that the solar variability and/or volcanoes play a role in the AMO (Hansen et al. 2005; Otterå et al. 2010), or the external forcing aerosols can be as a primary driver for the AMO (Booth et al. 2012). A recent observational study shows that a positive feedback between the SST and dust aerosol in the North Atlantic via Sahel rainfall variability may be a mechanism for the AMO (Wang et al. 2012).

As a basin-wide SST mode in the North Atlantic, the AMO warming or cooling is also manifested in the tropical North Atlantic (TNA) SST anomalies. That is, the TNA SST variation on multidecadal timescales is in phase with the AMO. The variations of the TNA

2

1   SST and upper ocean temperature are very important because they directly affect hurricane

2   activity in the North Atlantic (e.g., Shay et al. 2000; Wang et al. 2008).   Here, we focus on the

3   TNA ocean temperature variations associated with the AMO.   Zhang (2007) and Wang et al.

4   (2010) showed that the TNA SST anomalies are inversely related to or anticorrelated with the

5   subsurface ocean temperature anomalies in association with the AMO.   In other words, the

6   TNA SST anomalies vary out-of-phase with the subsurface ocean temperature anomalies on

7   multidecadal timescales.   In a study for comparing several ocean reanalysis products by Corre

8   et al. (2012), the variation of the TNA subsurface temperature with the AMO can also be seen

9   (their Fig. 10).   Zhang (2007) suggested that the anticorrelated change between the TNA

10   surface and subsurface temperatures is a distinctive signature of the AMOC variation based on

11   the coupled GFDL_CM2.1 model and thus can be taken as an AMOC fingerprint.   The

12   mechanism of the anticorrelated change proposed by Zhang (2007) is associated with the

13   AMOC-induced TNA subsurface thermocline adjustment.

14   In the present paper, we further show that the anticorrelated multidecadal variation does

15   exist in observational data and climate models of the Coupled Model Intercomparison Project

16   phase 5 (CMIP5).   Using observational data and CMIP5 models, we demonstrate that the TNA

17   subsurface cooling (warming) during the warm (cold) phase of the AMO is largely due to the

18   meridional advection by the anomalous northward (southward) current, whereas the subsurface

19   thermocline adjustment plays a secondary role.   The present paper further shows that the

20   anomalous northward (southward) current is attributed to variability of the AMOC and the

21   shallow Atlantic subtropical cell (STC) (e.g., Malanotte-Rizzoli et al. 2000; Zhang et al. 2003).

1   In addition, the paper also discusses variability of the TNA surface and subsurface salinity on

2   multidecadal timescales.

3   The paper is organized as follows.   Section 2 briefly introduces the data sets and methods

4   used in this paper.   The assessment of the anticorrelated multidecadal variation between TNA

5   surface and subsurface temperatures in observational data and CMIP5 models is presented in

6   Section 3.   The mechanisms of the anticorrelated multidecadal variation are analyzed and

7   examined in Section 4.   Section 5 shows a linkage among the AMO, AMOC, STC, and the

8   TNA surface and subsurface ocean temperature variation.   Section 6 briefly investigates the

9   multidecadal variation of salinity in the TNA Ocean.   The paper is concluded with a summary

10   and discussion in Section 7.

11

12   **2. Data sets and methods**

13   Several data sets are used in this study.   The first one is the improved extended

14   reconstructed SST (ERSST), with a 2° latitude by 2° longitude resolution (Smith and

15   Reynolds 2004).   The second data set is the Simple Ocean Data Assimilation (SODA)

16   (Carton and Giese 2008).   The SODA uses an ocean general circulation model (GCM) to

17   assimilate available temperature and salinity observations.   The product is a gridded data set

18   of oceanic variables with monthly values at a 0.5°×0.5° latitude-longitude horizontal

19   resolution and 40 vertical levels (The SODA product is sometimes called observational data).

20   The ocean model surface boundary conditions are taken from a new atmospheric data set

21   designated as 20CRv2 (20th Century Reanalysis version 2), which contains the

1   synoptic-observation-based estimate of global tropospheric variability spanning 1871 to 2008

2   at 6-hourly temporal and 2° spatial resolutions (Compo et al. 2011).   The version 2.2.4 of the

3   SODA data is used, with the time covering from 1871 to 2008.   Additionally, the objectively

4   analyzed temperature and salinity version 6.7 (Ishii et al. 2006) at 24 levels in the upper ocean

5   of 1500 m from 1945-2010 is also used.   The analysis is based on the World Ocean Database,

6   the global temperature-salinity in the tropical Pacific from IRD/France, and the Centennial in

7   situ Observation Based Estimates (COBE) SST.   The Ishii et al. analysis also includes the

8   Argo profiling buoy data in the final several years and the XBT depth bias correction.

9   Eighteen coupled GCM output data sets of the "historical" simulations provided to the

10   Intergovernmental Panel on Climate Change Fifth Assessment Report (IPCC-AR5) are also used

11   in this study.   The modeling center and country, IPCC model name, and temporal coverage are

12   shown in Table 1.   The model data can be downloaded from the website of the Coupled Model

13   Intercomparison Project phase 5 (CMIP5; Taylor et al. 2012) (http://cmip-pcmdi.llnl.gov/cmip5/).

14   The historical runs are forced by observed atmospheric composition changes which reflect both

15   anthropogenic (greenhouse gases) and natural sources (volcanic influences, solar forcing,

16   aerosols and emissions of short-lived species and their precursors) and, for the first time,

17   including time-evolving land cover.   The historical runs cover much of the industrial period

18   from the mid-nineteenth century to the present and are sometimes referred to as "twentieth

19   century" simulations.   All of the coupled GCMs are coupled freely without flux correction.

20   Variables of surface heat flux, wind, temperature, salinity, and current are used in this study.

21   Several statistical methods are used in this study, such as the lead-lag correlation and linear

5

1 regression.   To demonstrate possible mechanisms of the surface and subsurface ocean

2 temperature variations in the TNA, we conduct the heat budget analysis of the zonal mean

3 temperature.   The ocean heat budget analysis is based on the temperature equation in which the

4 local temperature change depends on ocean temperature advection terms and surface heat flux.

5 Our major purpose is to qualitatively rather than quantitatively assess the relative importance of

6 the surface heat flux and ocean advections in generating the low-frequency temperature variation.

7 Since we focus on the zonal mean surface and subsurface temperature anomalies, the ocean

8 advection in the zonal direction can be neglected.   All the data are linearly detrended and

9 smoothed by a 7-year low-frequency filter (we use the fifth order Butterworth low-pass filter).

10

11 **3. Anticorrelated variation of TNA surface and subsurface ocean temperatures**

12 *a. Observed results*

13   We first use the SODA and Ishii data to calculate the AMO indices (Figs. 1a, e) on

14 low-frequency (or multidecadal) timescales.   We then regress the Atlantic zonal mean ocean

15 temperatures onto the multidecadal AMO indices (Figs. 1b, f).   Consistent with the previous

16 studies of Zhang (2007) and Wang et al. (2010), the TNA Ocean shows an anticorrelated

17 variation between the surface and subsurface temperatures on multidecadal timescales.   It can

18 be seen that the surface temperature is characterized by a basin-wide and uniform warming

19 (cooling) in the entire North Atlantic with a maximum magnitude occurring in the subpolar

20 region during the warm (cold) phase of the AMO, in agreement with the definition of the AMO

21 index.   The penetration depth of the surface warming is latitude-dependent.   The higher the

6

1   latitude is, the deeper the surface warming penetrates.   On the contrary, the TNA subsurface

2   ocean temperature tends to become cool (warm) during the warm (cold) phase of the AMO.   In

3   the TNA Ocean, the anomalous cooling appears below 100 m and penetrates down to 1500 m,

4   with a maximum cooling around 200 m between 8°N and 20°N.   North of 20°N, the location of

5   cooling anomaly becomes deeper.

6   The anticorrelated surface and subsurface temperature variation or vertical dipole

7   temperature structure in the TNA occurs only on multidecadal timescales.   The low panels of

8   Fig. 1 show the regressions of the TNA ocean temperature anomalies at each depth onto the

9   TNA SST anomalies (averaged in the region of 8°N-20°N and coast to coast) on interannual and

10 multidecadal timescales from the SODA and Ishii data.   It is shown that the dipole structure in

11 the TNA appears only on multidecadal timescales, whereas on interannual timescales the

12 temperature profile exhibits a uniform positive regression between the surface and subsurface

13 oceans.   This suggests that the TNA dipole temperature anomaly should originate from the

14 relatively slow ocean adjustment, particularly in the high latitude where the ocean has a long

15 time memory.

16   To further demonstrate the anticorrelated variation, we plot the multidecadal time series of

17 temperature anomalies at the sea surface and 200 m, and vertically averaged temperature

18 anomalies from 0-400 m using the SODA and Ishii data (Fig. 2).   Since the Ishii data are

19 available only from the middle of the 20th century, for the purpose of consistency we plot both

20 the time series starting from 1945.   The anticorrelated or out-of-phase relationship between the

21 SST anomalies and temperature anomalies at 200 m is obvious in both the data sets, with the

1   correlation coefficients of -0.95 and -0.93 in the SODA and Ishii data, respectively.   When the

2   TNA surface ocean is warm (cold), the subsurface temperature at 200 m is cold (warm).   Since

3   the subsurface variation is deeper and larger than the surface (Figs. 1b, f), the vertically averaged

4   temperature anomalies between 0-400 m mainly follow the subsurface temperature anomalies

5   such as those at 200 m.

6

7   *b. CMIP5 simulations*

8   The AMO indices from 18 CMIP5 model simulations and the ERSST observation are

9   shown in Fig. 3.   CMIP5 models show a significant decadal to multidecadal variation of the

10   AMO; however, they strongly disagree both in phase and strength with the observation.   To

11   some degree, CMIP5 simulations share a relatively better resemblance with the observation than

12   CMIP3 models (Medhaug and Furevik 2011), particularly in the last two decades.   The two

13   main discrepancies between the model and observation are during the early 20th century

14   (1900-1920) when the models underestimate the cooling, and during the subsequent mid-century

15   warming (1925-1965) when the models are generally too cool.   This could be due to errors in

16   the observations, inadequacy in the modeled response to external forcing or forcing that is not

17   included in model simulations, or due to different internal variations in different models.   The

18   contributions of external forcing and internal variation to the North Atlantic SST in climate

19   models are recently in debate (e.g., Ting et al. 2009; Booth et al. 2012; Terray 2012; Zhang et al.

20   2013).   It is possible that a combination of both effects is attributed to the 20th century

21   multidecadal SST variation in the North Atlantic; however, their relative importance remains an

1   issue to be determined.

2   The regression of the Atlantic zonal mean temperatures onto the AMO index for individual

3   models is shown in Fig. 4.   Of 18 CMIP5 models, 12 models are able to simulate the

4   anticorrelated variation between the TNA surface and subsurface ocean temperatures:

5   bcc-csm1-1, CanESM2, CNRM-CM5, CSIRO-MK3-6-0, GFDL-ESM2G, GFDL-ESM2M,

6   GISS-E2-R, HadGEM2-ES, MIROC5, MIROC-ESM-CHEM, MPI-ESM-P and MRI-CGCM3

7   (Figs. 4a-l).   These models simulate a surface warming (cooling) anomaly in the TNA Ocean

8   and a subsurface cooling (warming) anomaly with the maximum around 200 m between

9   8°N-20°N during the warm (cold) phase of the AMO.   The other 6 models do not well simulate

10   the TNA subsurface cooling during the warm phase of the AMO: FGOALS-g2, GISS-E2-H,

11   IPSL-CM5A-LR, IPSL-CM5A-MR, MIROC-ESM, and CCSM4 (Figs. 1m-r).   In the following

12   section, we will examine physical mechanisms that are responsible for the vertical temperature

13   dipole structure associated with the AMO.

14

15   **4. Mechanisms of surface and subsurface ocean temperature variations**

16   *a. Long-term mean states*

17   To help the understanding of the ocean temperature variations in the TNA, we first present

18   the oceanic and atmospheric mean states in the TNA.   Long-term mean surface net heat flux in

19   the TNA is characterized by a heat loss from the ocean in the west and a heat gain in the east

20   (Fig. 5a).   The former is due to the strong trade wind-induced turbulent heat flux, whereas the

21   latter is largely associated with the radiative heat flux.   Surface freshwater flux (evaporation

1   minus precipitation) exhibits a freshwater gain in the Intertropical Convergence Zone (ITCZ)

2   region (Fig. 5b) where the surface wind convergence is accompanied with the ascending motion.

3   In the north of the ITCZ, the TNA is covered by a freshwater loss as a result of the surface wind

4   divergence and the descending motion.   Fig. 5c displays the vertical structure of the zonal mean

5   ocean temperature in the TNA region.   In the surface, the temperature gradually decreases with

6   the latitude, resulting from the solar radiation.   Thus, the mean meridional SST gradient is

7   negative near the surface, except near the equator because of the equatorial upwelling (Fig. 6a).

8   In the subsurface, it is interesting to see that there is a temperature dome at 9°N where the ITCZ

9   is located and the isotherms on both sides tend to tilt upward (Fig. 5c).   On the equatorward side

10   of the temperature dome, the mean meridional temperature gradient in the subsurface ocean is

11   negative, while the opposite is true on the poleward side (Fig. 6a).   Similar to the temperature

12   structure, the zonal mean salinity is also represented by a salinity dome at 9°N (Fig. 5d).

13   Therefore, the mean meridional salinity gradient also features a dipole pattern across 9°N (Fig.

14   6b).   In the surface, salinity is low in the ITCZ region and high in the subtropics (Fig. 5d) and

15   thus gives rise to a positive meridional salinity gradient north of the dome (Fig. 6b), which can

16   be expected from the surface net freshwater flux (Fig. 5b).   Further inspection finds that the

17   maximum positive meridional temperature and salinity gradients in the subsurface locate in the

18   depth of about 200 m between 8-16°N (Figs. 6a, b), which collocates with the maximum

19   multidecadal subsurface TNA temperature variation (Fig. 1).

20   The vertical temperature gradient is always positive in the TNA with the maximum gradient

21   in the upper layer from the equator to 15°N (Fig. 6c).   Contrary to the temperature, there is a

10

1   dipole structure in the vertical salinity gradient, with a negative gradient in the upper 100 m and

2   a positive salinity gradient below (Fig. 6d).   The surface negative salinity gradient can be

3   attributed to the surface freshwater gain and the intrusion of subtropical high salinity in the

4   subsurface (Fig. 5b).   As expected, the zonal mean meridional ocean velocity displays a

5   northward current above the Ekman layer, which results from the trade wind-induced northward

6   Ekman drift (Fig. 5e).   On the contrary, water tends to flow southward in the interior that can be

7   explained by the Sverdrup relation.   The zonal mean vertical ocean velocity in the TNA is

8   generally characterized by an upwelling south of 14°N and a downwelling in the north (Fig. 5f),

9   which is primarily due to the Ekman transport-induced vertical motion.

10

11   *b. Physical mechanisms in observational data*

12   We first examine the meridional and vertical advection terms contributing to the

13   multidecadal subsurface temperature variation.   Fig. 7 shows the regression of various zonal

14   mean terms onto the multidecadal AMO index in the SODA data (Note that the Ishii data do not

15   have ocean currents and cannot be used for the ocean heat budget).   The similar results can be

16   obtained by computing the zonal advection terms at each grid point and then performing the

17   zonal mean and regression onto the AMO index.   It is clearly seen that the TNA subsurface

18   cooling anomalies can be broadly explained by the total meridional advection (Fig. 7a), while the

19   total vertical advection plays a much smaller role except north of 14°N (Fig. 7e).   The advection

20   by the ocean currents is further decomposed into two components: Anomalous (denoted by

21   prime) and mean (denoted by overbar) currents.   It is shown that the total meridional advection

11

1　mainly arises from the advection by the anomalous meridional current ($-V'\partial\bar{T}/\partial y$, Fig. 7b),

2　whereas the advection by the mean meridional current is secondary ($-\bar{V}\partial T'/\partial y$, Fig. 7c).

3　Compared to these two terms, the nonlinear term is small ($-V'\partial T'/\partial y$, Fig. 7d).　A remarkable

4　feature is that the subsurface cooling between 8°N-20°N in Fig. 1b corresponds to the

5　contribution due to the advection term by the anomalous current in Fig. 7b ($-V'\partial\bar{T}/\partial y < 0$　in

6　the TNA subsurface ocean).　Because the mean meridional temperature gradient is positive at

7　the TNA subsurface ocean ($\partial\bar{T}/\partial y > 0$, Fig. 6a), an anomalous northward current must occur in

8　the subsurface ocean, i.e., V'　must be positive.　Since the zonally averaged field is considered

9　here, it is expected that the anomalous northward current is associated with the flow variations of

10　the AMOC and/or the Atlantic subtropical cell (STC).　We will come back this issue later.

11　　Similar to the total meridional advection, the total vertical advection primarily depends on

12　the advection by the anomalous vertical velocity (Fig. 7e vs Fig. 7f), while the advection by the

13　mean vertical velocity plays a much smaller role (Fig. 7e vs Fig. 7g).　The nonlinear term can

14　be ignored (Fig. 7h).　The advection by the anomalous vertical velocity ($-W'\partial\bar{T}/\partial z$) tends to

15　cool temperature in the upper 400 m north of 14°N and to warm temperature south of 14°N.

16　Because　$\partial\bar{T}/\partial z$　is always positive, this implies that there is an anomalous downwelling

17　(upwelling) south (north) of 14°N, consistent with the wind distribution (and its corresponded

18　Ekman transport and vertical motion) associated with the AMO (Fig. 8a).　The effect of the

19　advection by the mean vertical velocity tends to be opposite to that by the anomalous vertical

20　velocity (Fig. 7g vs Fig. 7f).

1   The contributions of both $-\bar{W}\,\partial T'/\partial z$ and $-\bar{V}\,\partial T'/\partial y$ to the TNA subsurface

2   temperature cooling are small.   These suggest that the anomalous temperature gradients induced

3   by the local process or remote forcing such as the thermocline adjustment (Zhang 2007) cannot

4   produce the subsurface temperature cooling on multidecadal timescales.   The result does not

5   support the hypothesis by Zhang (2007) who suggested that the ocean thermocline adjustment

6   associated with the AMOC is important for the multidecadal subsurface cooling in the TNA.   In

7   summary, the TNA subsurface cooling associated with the warm phase of the AMO is largely

8   attributed to the advection by the anomalous northward current.   North of 14°N, the advection

9   by an anomalous upwelling also contributes to the TNA subsurface cooling.

10   Next, we examine the TNA surface warming associated with the warm phase of the AMO.

11   The total meridional advection shows a surface cooling in the TNA region (Fig. 7a).   The

12   surface cooling is contributed by both the advections of the anomalous and mean meridional

13   currents ($-V'\partial\bar{T}/\partial y$ and $-\bar{V}\,\partial T'/\partial y$, Figs. 7b, c).   Given the surface mean states (Figs. 5, 6),

14   this indicates that there exists an anomalous surface southward current and positive SST

15   meridional gradient anomaly.   The anomalous surface southward current is largely due to the

16   anomalous westerly wind-induced Ekman drift (Fig. 8a).   It is also seen that there is an obvious

17   boundary to separate the surface and subsurface cold meridional advection in Fig. 7b.   This

18   boundary approximately collocates with the ocean mixed layer, and has a negligible meridional

19   advection because of the nearly zero anomalous meridional current (the transition from the

20   subsurface northward to surface southward flows) and temperature gradient.

13

1   The total vertical advection can explain the surface warming south of 14°N (Fig. 7e) as a

2   result of an anomalous downwelling induced by the weakened trade winds.   However, it tends

3   to cool the surface temperature north of 14°N due to the anomalous upwelling (Fig. 7e, f).

4   Therefore, both the meridional and vertical advection terms cannot explain the surface warming

5   in the TNA north of 14°N associated with the warm phase of the AMO.   In fact, the surface

6   warming in the TNA north of 14°N is primarily due to the surface net heat flux (Fig. 8a).   The

7   westerly wind anomalies associated with the warm phase of the AMO reduce the loss of

8   turbulent heat flux and then warm the surface ocean.

9

10   *c. Physical mechanisms in CMIP5 simulations*

11   As shown earlier, 12 of 18 CMIP5 models simulated the out-of-phase variation of the TNA

12   surface and subsurface ocean temperatures reasonably well.   Fig. 9 shows the 12 model

13   ensemble mean of advection term regressions onto the model AMO index.   Overall, the result is

14   broadly similar to that in the SODA data.   In particular, the advection by anomalous meridional

15   current is a major contribution to the TNA subsurface ocean cooling (Fig. 9b).   In comparison

16   with the SODA data, the advection by the mean meridional current is well organized, with the

17   TNA subsurface ocean warming and cooling south and north of 13°N, respectively (Fig. 9c).

18   The TNA subsurface cooling north of 13°N is an obvious feature in spite of a weak magnitude.

19   This indicates that the advection of the anomalous meridional temperature gradient by the mean

20   meridional current can partly contribute to the subsurface cooling in the TNA, although it is not a

21   dominant factor.

14

1   Like the result in the SODA data, the anomalous downwelling south of 14°N disfavors

2   (favors) the TNA subsurface cooling (surface warming), whereas the opposite is true north of

3   14°N.   In the meantime, the advection by the mean vertical velocity tends to offset but cannot

4   overwhelm the effect of advection by the anomalous vertical velocity (Figs. 9f, g).   As a result,

5   the total vertical advection (Fig. 9e) is mainly determined by the advection due to the anomalous

6   vertical velocity.   We have to keep in mind that the contribution of the meridional advection to

7   the TNA multidecadal temperature variation is much larger than that of the vertical advection

8   (Fig. 9a vs Fig. 9e).

9   Similar to the SODA data, the model-simulated surface warming in the TNA associated with

10   the warm phase of the AMO is also largely due to the surface net heat flux induced by the

11   weakened trade winds (Fig. 10).   Nearly all the models display a cyclonic wind stress and a

12   weakening of the trade winds in the North Atlantic, consistent with the 20CRv2 reanalysis.

13   Even in the models that do not well simulate the out-of-phase relationship between the TNA

14   surface and subsurface ocean temperatures, the net surface flux still tends to warm the TNA

15   surface ocean during the warm phase of the AMO (Figs. 10m-r).   This suggests that the heat

16   flux is important for the surface ocean temperature in the TNA.

17   In summary, the heat budget analyses from both observational data and CMIP5 models show

18   that the anomalous northward current plays an important role in the subsurface ocean cooling

19   (warming) in the TNA associated with the warm (cold) phase of the AMO.   Given that the

20   subsurface anomaly occurs from the mixed layer down to 800-1500 m, it is expected that the

21   anomalous northward current may relate to both the deep AMOC and shallow STC variability.

1   In the following sections, we will discuss the AMOC and STC relations to the TNA ocean

2   temperature variation associated with the AMO.

3

4   **5. Link of the AMO with the AMOC and STC**

5   *a. Relation to the AMOC*

6   Both the SODA data and most of CMIP5 models show that during the warm (cold) phase of

7   the AMO, the anomalous northward (southward) meridional current is responsible for the TNA

8   subsurface ocean cooling (warming).   To examine AMOC relation to this variation, we first

9   calculate the AMOC streamfunctions.   The long-term mean streamfunctions for the SODA data

10   and CMIP5 models are shown in Fig. 8b and Fig. 11, respectively.   The mean AMOC

11   streamfunctions show highly varying structures.   The positions of the maximum overturning are

12   typically found at 600-1500 m depth and between 20°N and 60°N.   There are large differences

13   in how the models reproduce the lower overturning cell.   Several models show either an absent

14   or a very weak lower overturning cell of the Antarctic Bottom Water (AABW), while other

15   models show the AABW all the way north to 60°N.   Hydrographic observations show that the

16   AABW almost disappears north of 35°N (Johnson 2008).   CMIP5 models show a long-term

17   mean overturning circulation range from 13 to 33 Sv.   Based on hydrographic data, the

18   estimates of the AMOC strength are 14-18 Sv at 24°N (Ganachaud and Wunsch 2000; Lumpkin

19   and Speer 2003), 18.7±5.6 Sv at 26.5°N (Cunningham et al. 2007), and 13-19 Sv at 48°N

20   (Ganachaud 2003).   Generally speaking, CMIP5 models basically capture the feature of the

21   AMOC although the magnitude and location are a little bit different from observations.

16

1   The AMOC index in the TNA Ocean is defined as the maximum streamfunction in the

2   latitude band of 5°N-20°N.   Similar results can be obtained if we choose a specific latitude in

3   the TNA (not shown).   To reduce and/or exclude the surface wind driven overturning, we have

4   used a further criterion that the maximum should be located deeper than 300 m.   The AMOC

5   indices in the TNA for the SODA data and CMIP5 models are shown in Fig. 8c and Fig. 12.   To

6   emphasize the multidecadal variability, we have detrended and smoothed the indices by a 7-year

7   low-frequency filter.   An initial impression is that these TNA AMOC indices vary differently.

8   However, a close examination finds that the TNA AMOC indices in the SODA data and most of

9   CMIP5 models vary with their individual AMO indices.   The feature can be clearly seen from

10   the correlation plots in Fig. 8e and Fig. 13.   At the zero time lag, the TNA AMOC and AMO

11   indices are positively correlated in the SODA data and the 12 CMIP5 models that are able to

12   simulate the TNA subsurface ocean cooling (Fig. 8e and Figs. 13a-l).   For the 6 models that are

13   unable to simulate the subsurface ocean cooling, the correlations are negative except the nearly

14   zero correlation of IPSL-CM5A-MR (Figs. 13m-r).   This implies that the TNA AMOC is

15   strengthened (weakened) during the warm (cold) phase of the AMO.   It also means that there is

16   an anomalous northward (southward) current in the TNA upper ocean layer during the warm

17   (cold) phase of the AMO, which in turn induces the TNA subsurface cooling (warming).   This

18   is consistent with the result of the heat budget analysis shown earlier for the SODA data and the

19   12 models showing realistic subsurface temperature anomalies.

20   Previous studies suggested that the AMOC variations have a latitudinal dependence (Mignot

21   and Frankignoul 2005; Msadek and Frankignoul 2009; Zhang 2010).   These studies showed that

17

the subpolar AMOC variation leads the subtropical and tropical AMOC variations by several

years (about 5 years in GFDL_CM2.1 model) and the length of time lag is mainly determined by

the advection speed in the North Atlantic deep water formation region (Zhang 2010).   Here, we

also analyze the high-low latitudinal dependence of the relationship between the AMOC and

AMO in CMIP5 simulations.   In the higher latitudes, most of the models show that the AMOC

leads the AMO (not shown).   As the latitude decreases, the leading time becomes smaller and

the relationship between the AMOC and AMO becomes more and more synchronized.   Thus,

the warm (cold) phase of the AMO corresponds to a strengthened (weakened) TNA AMOC in

the nearly zero time lag.

The positive correlations between the AMO and TNA AMOC in the SODA data and CMIP5

models maximize in different ways.   Some models show zero time lag (e.g., GFDL-ESM2G).

Some models show that the maximum correlation occurs when the TNA AMOC leads by several

years (e.g., GFDL-ESM2M), while the correlation in the SODA data maximizes when the AMO

leads the TNA AMOC by 5 years (Fig. 8e).   This inconsistency may arise from the different

adjustment time of the TNA AMOC (determined by the advection speed) to northern Atlantic

deep convection.   The other possibility may be from external factors' interference such as

aerosols.

b. Relation to the STC

The Atlantic subtropical cell (STC) is a shallow overturning cell connecting the subtropics

with the equatorial region of the Atlantic Ocean (e.g., Malanotte-Rizzoli et al. 2000; Zhang et al.

18

1   2003).   Water mass that is subducted into the thermocline in the eastern subtropics of both

2   hemispheres is swept westward by the north and south equatorial currents, respectively.   The

3   water mass eventually reaches to the equatorial undercurrent, drifts eastward and upwells in the

4   eastern equator.   In conjunction with the surface northward Ekman current, it forms a shallow

5   (mainly in the upper ocean of 250 m) overturning cell, i.e., the Atlantic STC.   To obtain the

6   Atlantic STC variability in the Northern Hemisphere, we choose two methods to define the STC

7   index.   One is similar to the Pacific STC (McPhaden and Zhang 2002, 2004; Schott et al. 2007,

8   2008), which chooses the coast-to-coast transport across 9°N as an STC index.   The other

9   definition is similar to the AMOC index, which is defined as the maximum streamfunction in the

10   latitude band of 5°N-20°N in the upper 200 m.   Although there are some magnitude differences

11   in the two definitions, the phase variability is very similar.   So we only show results from the

12   second definition in this paper.   Similar results can be obtained by using the first definition, in

13   particular for the normalized STC index and correlations with the AMO index.

14   The Atlantic STC indices in the SODA data and CMIP5 models are displayed in Fig. 8d and

15   Fig. 14.   The multidecadal STC variability is different from the TNA AMOC variability (Fig.

16   8c vs Fig. 8d and Fig. 12 vs Fig. 14).   For the SODA data and 12 CMIP5 models which are able

17   to simulate the out-of-phase relationship between the TNA surface and subsurface temperatures,

18   the STC indices are negatively correlated with the TNA AMOC indices at the zero time lag,

19   whereas for the other 6 models the correlations are either positive or nearly zero (not shown).

20   Since we are interested in examining the contribution of the meridional current to the TNA

21   subsurface temperature variation associated with the AMO, we focus on the lead-lag correlations

19

1 between the AMO and STC indices. All the models and SODA data show a negative

2 correlation between the AMO and STC indices with a zero time lag (Figs. 8f, 15). This

3 indicates that the northern Atlantic STC is significantly weakened (strengthened) during the

4 warm (cold) phase of the AMO. The weakening (strengthening) of the STC is accompanied by

5 an anomalous northward (southward) current in the subsurface (about 100-200 m) and an

6 anomalous southward (northward) Ekman drift in the surface. The anomalous subsurface

7 northward current associated with the AMO produces the multidecadal subsurface cooling as

8 exhibited in Figs. 1b, f and Fig. 4, at least in the upper 200 m. Of course, the weakening of the

9 STC cannot induce the cooling anomaly in the ocean deep layer (because of its shallow feature)

10 which is due to the AMOC variation as discussed in last subsection.

11 The STC variation associated with the AMO can be explained by the surface wind variation,

12 particularly the trade winds and the associated subtropical wind stress curl (Fig. 8a and Fig. 10).

13 The weakening (strengthening) of the surface trade winds and a positive (negative) subtropical

14 wind stress curl anomaly correspond to a decrease (an increase) of the STC. This is because the

15 interior southward Sverdrup transport, which is a key element of the subsurface branch of the

16 STC, is primarily determined by the subtropical wind stress curl. Meanwhile, the surface

17 northward Ekman transport (the upper branch of the STC) is mainly determined by the strength

18 of the trade winds. Because all models and observational data show a weakening

19 (strengthening) of the trade winds and a positive (negative) subtropical wind stress curl during

20 the warm (cold) phase of the AMO, the Atlantic STC is all weakened (strengthened) as shown in

21 Figs. 15, 8f.

20

1

c. Interpreting subsurface cooling by the enhanced AMOC and weakened STC

As shown in previous sections, the warm (cold) phase of the AMO is accompanied by the

subsurface ocean cooling (warming) in the TNA which is mainly attributed to the meridional

advection by an anomalous northward (southward) in the subsurface ocean.   Our analyses have

further shown that the warm (cold) phase of the AMO is associated with a strengthening

(weakening) of the TNA AMOC in the SODA data and 12 CMIP5 models, but the opposite is

true for the other 6 CMIP5 models.   The SODA data and these 12 models do show the TNA

subsurface ocean cooling (warming) during the warm (cold) phase of the AMO, whereas these 6

models do not well simulate the subsurface cooling (warming).   However, all 18 models and the

SODA data show that the warm (cold) phase of the AMO corresponds to a weakening

(strengthening) of the Atlantic STC.   The strengthening (weakening) of the AMOC features an

anomalous northward (southward) current in the TNA subsurface ocean, and so does the

weakening (strengthening) of the STC.   These suggest that both the variations of the AMOC

and STC can contribute to the subsurface ocean cooling (warming), but the AMOC contribution

seems to be more important.

We first consider the case for the SODA data and the 12 CMIP5 models that are able to

simulate the out-of-phase variation between the TNA surface and subsurface ocean temperatures.

If the ocean mixing in the deep water formation region increases, the associated high latitude

AMOC strengthens (e.g., Stocker et al. 2007; Zhang et al. 2012b).   Then, the AMOC signal

propagates southward accompanied with the advection, coastal and equatorial Kelvin waves, and

21

Rossby waves (e.g., Zhang 2010).   After several years later, the AMOC in the TNA reaches to

the maximum, induces an anomalous northward current in the upper 1500 m, and thus generates

a TNA subsurface ocean cooling between 8°N-20°N where the mean meridional temperature

gradient is positive due to the 9°N temperature dome.   Meanwhile, at the surface the

strengthened AMOC leads to a substantial North Atlantic and TNA surface warming (the warm

phase of the AMO) as a result of the increased net heat flux and increased northward heat

transport.   The TNA surface warming decreases the surface pressure, induces an anomalous

surface convergence, and thus generates a weakening of the trade winds and a positive wind

stress curl anomaly in the subtropical North Atlantic.   On one hand, the weakened trade winds

tend to warm SST by decreasing the turbulent heat flux loss and increasing the Ekman drift

induced downwelling.   On the other hand, the weakened trade winds and positive wind stress

curl anomaly reduce the shallow STC strength and the associated subsurface southward interior

current, generating an anomalous northward current.   Thus, like the AMOC, the STC also acts

to cool the TNA subsurface ocean.

The situation is different for the 6 CMIP5 models that are not well simulate the out-of-phase

variation between the TNA surface and subsurface ocean temperatures.   These models show

that the warm (cold) phase of the AMO is associated with a weakening (strengthening) of the

AMOC in the TNA.   So the models cannot simulate the anomalous northward (southward)

current induced by the AMOC in the TNA subsurface ocean that is required for the subsurface

cooling (warming) during the warm (cold) phase of the AMO.   However, these 6 models are

able to simulate the inverse relationship between the shallow STC and AMO because the STC is

1   mostly wind-driven (the models do simulate the feature of the weakened trade winds).   This

2   means that the simulated warm (cold) phase of the AMO is associated with a weakening

3   (strengthening) of the STC.   The weakening (strengthening) of the STC induces an anomalous

4   northward (southward) current which can make a contribution to the subsurface ocean cooling

5   (warming) in the TNA.   However, this contribution is not large enough for the models to fully

6   simulate the out-of-phase variation between the subsurface and surface ocean temperatures in the

7   TNA.

8

9   **6. Multidecadal salinity variation in the TNA Ocean**

10   Given the feature of the TNA surface and subsurface temperature variation, a natural

11   question is: Does salinity show a similar variation in the TNA Ocean?   Using the pentadal

12   salinity data of the World Ocean Database 2005 (Boyer et al. 2006), Wang et al. (2010) showed

13   that the salinity anomalies do not seem to display an out-of-phase relationship between the

14   surface and subsurface oceans in the TNA associated with the AMO (see their Fig. 10).   Here

15   we also calculate the regressions of the Atlantic zonal mean ocean salinity onto the AMO index

16   for the SODA and Ishii data (Figs. 16s, t).   The SODA regression is similar to the result of

17   Wang et al. (2010), which shows a uniform freshening in the TNA surface and subsurface oceans.

18   The Ishii salinity regression does not seem to show an obvious out-of-phase relation either

19   although there is a positive salinity regression near the surface north of 15°N (Fig. 16t).   We

20   have to keep in mind that the salinity data may not be reliable, especially during earlier time.

21   The regressions of the Atlantic zonal mean ocean salinity onto their individual AMO indices

1   for the 18 CMIP5 models are shown in Figs. 16a-r.   For most of the 12 models that are able to

2   simulate the multidecadal TNA temperature variation, they also produce the out-of-phase

3   relation in the TNA surface and subsurface ocean salinity (Figs. 16a-l).   During the warm (cold)

4   phase of the AMO, these models show the negative salinity anomalies in the TNA subsurface

5   ocean and the positive salinity anomalies in the surface ocean.   For the 6 models that are unable

6   to simulate the out-of-phase temperature variation, they cannot well produce the negative salinity

7   anomalies in the TNA subsurface ocean either (Figs. 16m-r).

8   We also calculate the salinity variation contribution by the ocean advection terms.   The

9   salinity variation is largely determined by the meridional salinity advection.   The 12 model

10   ensemble mean of the regressions of the meridional salinity advection terms onto the AMO index

11   is shown in Fig. 17.   It is shown that the total meridional advection can explain the decreased

12   salinity in the TNA subsurface ocean and the increased salinity in the TNA surface ocean (Fig.

13   17a).   A further decomposition shows that the salinity advection by the anomalous meridional

14   current plays a dominant role and the other terms are secondary (Figs. 17b-d).   Given the

15   positive mean meridional salinity gradient ($\partial \overline{S} / \partial y > 0$), Fig. 17b indicates that the TNA

16   subsurface ocean has an anomalous northward current, whereas the surface ocean flow is

17   anomalously southward.   These are consistent with the results in the last section that these 12

18   models are associated with a strengthening of the TNA AMOC and a weakening of the Atlantic

19   STC during the warm phase of the AMO, both of which result in an anomalous northward

20   current in the subsurface ocean.   The surface anomalous southward flow is due to the Ekman

21   transport induced by the weakened trade winds associated with the warm phase of the AMO.

1   Thus, the model simulated AMOC and STC can also contribute to the salinity variation of the

2   TNA subsurface and surface oceans on multidecadal timescales.

3

4   **7. Discussion and conclusion**

5   The paper uses observational data and CMIP5 model simulations to show and examine the

6   variations of the Atlantic multidecadal oscillation (AMO), Atlantic meridional overturning

7   circulation (AMOC), subtropical cell (STC) and surface wind in the North Atlantic.   It is shown

8   that the tropical North Atlantic (TNA) features an anticorrelated variation in the surface and

9   subsurface ocean temperatures associated with the AMO.   That is, the warm (cold) phase of the

10 AMO is accompanied by a warm (cold) surface ocean and a cold (warm) subsurface ocean in the

11 TNA.   Given that there is no significant rapid TNA subsurface response to changes in the

12 radiative forcing, this multidecadal TNA subsurface temperature anomaly can be taken as a

13 proxy or fingerprint for the AMOC variation (Zhang 2007).   Here we show that the

14 anticorrelated surface and subsurface temperature variation in the TNA does indeed link with the

15 variations of both the AMOC and STC.   We found that the TNA subsurface temperature

16 variation is largely attributed to the advection by the anomalous meridional current ($-V'\partial \bar{T} / \partial y$)

17 which is induced by the AMOC and STC variations.   Because the mean meridional temperature

18 gradient in the TNA subsurface is positive ($\partial \bar{T} / \partial y > 0$) due to the temperature dome at 9°N,

19 the anomalous subsurface current induced by the AMOC and STC basically controls the

20 variation of the TNA subsurface temperature.   The advections of the anomalous temperature

25

1 gradients by the mean currents ($-\bar{W}\,\partial T'/\partial z$ and $-\bar{V}\,\partial T'/\partial y$) are of secondary importance in

2 controlling the TNA subsurface temperature variation.   This suggests that the temperature

3 gradient changes induced by the local process or remote forcing such as the thermocline

4 adjustment (Zhang 2007) cannot produce the subsurface temperature cooling associated with the

5 warm phase of the AMO.

6   The variations of the AMOC, STC, ocean temperature and surface wind with the AMO can

7 be summarized in Fig. 18.   A strengthening of the AMOC is associated with the AMOC

8 anomaly signal in the high latitude that propagates southward, which is accompanied with the

9 advection, coastal and equator Kelvin waves, and Rossby waves (e.g., Zhang 2007).   Several

10 years later, the strength of the TNA AMOC reaches to its maximum, which induces an

11 anomalous northward current in the upper ocean of 2000 m and thus generates a TNA subsurface

12 cooling between 8°N-20°N where the mean meridional temperature gradient is positive due to

13 the 9°N temperature dome.   Meanwhile, in the surface, the strengthened AMOC leads to a

14 substantial North Atlantic and TNA surface warming (i.e., the warming during the warm phase

15 of the AMO).   The TNA surface warming decreases the surface pressure, induces an anomalous

16 surface convergence and thus generates a weakening of the trade winds and a positive wind

17 stress curl anomaly over the subtropical North Atlantic.   On one hand, the weakened trade

18 winds tend to warm the surface ocean by decreasing the turbulent heat flux loss and increasing

19 the Ekman drift induced downwelling.   On the other hand, the weakened trade winds and

20 positive wind stress curl anomaly weaken the STC and the associated subsurface southward

21 interior current, generating an anomalous northward current and thus a subsurface cooling in the

1   TNA.   The opposite is true for a weakening of the AMOC.

2   The mechanism is operated in the SODA data and 12 CMIP5 model simulations.   In the

3   SODA data and 12 CMIP5 models, the warm (cold) phase of the AMO is associated with a

4   strengthening (weakening) of the TNA AMOC and a weakening (strengthening) of the STC,

5   both of which induce an anomalous northward (southward) subsurface current and thus cool

6   (warm) the TNA subsurface ocean.   However, 6 CMIP5 models cannot simulate the anomalous

7   northward (southward) current induced by the AMOC in the TNA subsurface ocean during the

8   warm (cold) phase of the AMO.   But these 6 models are able to produce a weakening

9   (strengthening) of the STC associated with the warm (cold) of the AMO because the STC is

10   mostly wind-driven (these models do simulate the feature of the weakened trade winds).   The

11   weakening (strengthening) of the STC does induce an anomalous northward (southward) current

12   which can make a contribution to the subsurface ocean cooling (warming) in the TNA.

13   However, this contribution is not large enough for these 6 models to fully simulate the

14   out-of-phase variation between the subsurface and surface ocean temperatures in the TNA.

15   This suggests that the AMOC plays a more important role than the STC in the anticorrelated

16   variation of the TNA surface and subsurface ocean temperatures.   In other words, a positive

17   correlation between the TNA AMOC and the AMO is required for models to simulate the

18   anomalous meridional current in the TNA subsurface ocean and then to change the subsurface

19   ocean temperature.

20   For most of the 12 CMIP5 models that are able to simulate the multidecadal TNA

21   temperature variation, they also produce an out-of-phase relation in the TNA surface and

27

1   subsurface ocean salinity.   During the warm (cold) phase of the AMO, these models show the

2   negative salinity anomalies in the TNA subsurface ocean and the positive salinity anomalies in

3   the surface ocean.   The mechanism is similar to that of the temperature variation.   In these

4   models, the ocean salinity variations in the TNA are largely determined by the meridional

5   advection associated with the AMOC and STC variations.

6   The present paper shows that an anomalous meridional current in the TNA subsurface ocean

7   is required for simulating the subsurface ocean temperature and salinity variations associated

8   with the AMO.   Both the AMOC and STC can contribute to the subsurface meridional current.

9   However, a quantitative separation of the meridional flows induced by the AMOC and STC is

10   difficult by analyzing the data and model outputs only.   Numerical model experiments are

11   needed for addressing this issue.   For example, two parallel model runs can be conducted.   The

12   first run is a fully coupled model experiment.   The second run is the same as the first one but

13   with a wind prescribing to the climatological wind in the TNA region.   In this case, the STC

14   cannot be changed by the wind variation.   By comparing these two model experiments, we can

15   obtain the relative contribution of the AMOC and STC to the anomalous meridional current in

16   the subsurface ocean.   Assessing the quantitative effect of the AMOC and STC remains

17   challenge and is beyond the scope of this paper.

18   The variations of the ocean temperature in the TNA region are very important because they

19   directly affect hurricane activity in the North Atlantic (e.g., Shay et al. 2000).   However, almost

20   all of previous studies on long-term hurricane variations use the SST variations only (e.g.,

21   Goldenberg et al. 2001; Wang et al. 2008).   The present paper and others (Zhang 2007; Wang et

al. 2010) show that on multidecadal timescales the TNA surface and subsurface ocean

temperatures vary differently.   In other words, the multidecadal temperature variations in the

TNA are different if we use different upper ocean layers (see Figs. 1 and 2).   Thus, a caution

should be exercised for interpreting hurricane activity in terms of the multidecadal ocean

temperature variation in the TNA.

1 **References**

2 Booth, B. B. B., and Coauthors, 2012: Aerosols implicated as a prime driver of twentieth-century

3   North Atlantic climate variability. *Nature*, **484**, 228-232.

4 Boyer, T. P., and Coauthors, 2006: World Ocean Database 2005. NOAA Atlas NESDIS 60, U.S.

5   Government Printing Office, 190 pp.

6 Carton, J. A., and B. S. Giese, 2008: A reanalysis of ocean climate using Simple Ocean Data

7   Assimilation (SODA). *Mon. Wea.Rev.*, **136**, 2999–3017.

8 Compo, G. P., and Coauthors, 2011: The Twentieth Century Reanalysis Project. *Q. J. R.*

9   *Meteorol Soc.*, **137**, 1–28.

10 Corre, L., L. Terray, M. Balmaseda, A. Ribes, and A. Weaver, 2012: Can oceanic reanalyses be

11   used to assess recent anthropogenic changes and low-frequency internal variability of upper

12   ocean temperature? *Clim. Dyn.*, **38**, 877-896.

13 Cunningham, S. A., and Coauthors, 2007: Temporal variability for the Atlantic Meridional

14   Overturning Circulation at 26.5N. *Science*, **317**, 935–937, doi:10.1126/science.1141304.

15 Delworth, T. L., and M. E. Mann, 2000: Observed and simulated multidecadal variability in the

16   Northern hemisphere. *Climate Dynamics*, **16**, 661–676.

17 Enfield, D. B., A. M. Mestas-Nuñez, and P. J. Trimble, 2001: The Atlantic multidecadal

18   oscillation and its relation to rainfall and river flows in the continental U.S. *Geophys. Res.*

19   *Lett.*, **28**, 2077– 2080.

20 Folland, C. K., D. E. Parker, and T. N. Palmer, 1986: Sahel rainfall and worldwide sea

21   temperatures, 1901−85. *Nature*, **320**, 602– 607.

22 Folland, C. K., A. W. Colman, D. P. Rowell, and M. K. Davey, 2001: Predictability of northeast

23   Brazil rainfall and real-time forecast skill, 1987–98. *J Clim.*, **14**, 1937– 1958.

24 Ganachaud, A. and C. Wunsch, 2000: Improved estimates of global ocean circulation, heat

25   transport and mixing from hydrographic data. *Nature*, **408**, 453–457.

1   Ganachaud, A., 2003: Large-scale mass transports, water mass formation, and diffusivities

2   estimated from World Ocean Circulation Experiment (WOCE) hydrographic data. *J.*

3   *Geophys. Res*., **3213**, doi:10.1029/2002JC001565.

4   Goldenberg, S. B., C. W. Landsea, A. M. Mestas-Nuñez, and W. M. Gray, 2001: The recent

5   increase in Atlantic hurricane activity: Causes and implications. *Science*, **293**, 474– 479.

6   Gray, W. M., J. D. Sheaffer, and C. W. Landsea, 1997: Climate trends associated with

7   multidecadal variability of Atlantic hurricane activity, in Hurricanes: Climate and

8   Socioeconomic Impacts, edited by H. F. Diaz and R. S. Pulwarty. pp. 15– 53, Springer,

9   New York.

10   Hansen, J., and Coauthors, 2005: Earth's energy imbalance: confirmation and implications.

11   *Science*, **308**, 1431-1435.

12   Ishii, M., M. Kimoto, K. Sakamoto, and S. I. Iwasaki, 2006: Steric sea level changes estimated

13   from historical ocean subsurface temperature and salinity analyses. *J. Oceanography*., **62**

14   **(2)**, 155-170.

15   Johnson, G. C., 2008: Quantifying Antarctic Bottom Water and North Atlantic Deep Water

16   volumes. *J. Geophys. Res*., **113**, C05027, doi:10.1029/2007JC004477.

17   Knight, J. R., and Coauthors, 2005: A signature of persistent natural thermohaline circulation

18   cycles in observed climate. *Geophys. Res. Lett*., **32**, L20708, doi:10.1029/2005GL024233.

19   Lumpkin, R. and K. Speer, 2003: Large-scale vertical and horizontal circulation in the North

20   Atlantic Ocean. *J. Phys. Oceanogr*., **33**, 1902–1920.

21   Malanotte-Rizzoli, P., K. Hedstrom, H. Arango, and D. B. Haidvogel, 2000: Water mass

22   pathways between the subtropical and tropical ocean in a climatological simulation of the

23   North Atlantic Ocean circulation. *Dyn. Atmos. Oceans*, **32**, 331– 371.

24   McCabe, G. J., M. A. Palecki, and J. L. Betancourt, 2004: Pacific and Atlantic Ocean influences

25   on multidecadal drought frequency in the United States. *Proc Natl Acad Sci*., **101**,

26   4136-4141.

27   McPhaden, M. J., and D. Zhang, 2002: Slowdown of the meridional overturning circulation in

the upper Pacific Ocean. *Nature*, **415**, 603– 608.

McPhaden, M. J., and D. Zhang, 2004: Pacific Ocean circulation rebounds. *Geophys. Res. Lett.*, **31**, L18301, doi:10.1029/2004GL020727.

Medhaug, I., and T. Furevik, 2011: North Atlantic 20th century multidecadal variability in coupled climate models: sea surface temperature and ocean overturning circulation. *Ocean Sci.*, **7**, 389-404.

Mignot, J., and C. Frankignoul, 2005: On the variability of the Atlantic meridional overturning circulation, the NAO and the ENSO in the Bergen Climate Model. *J. Clim.*, **18**, 2361-2375.

Msadek, R., and C. Frankignoul, 2009: Atlantic multidecadal oceanic variability and its influence on the atmosphere in a climate model. *Climate Dyn.*, **33**, 45-62.

Otterå, O. H., and Coauthors, 2003: The sensitivity of the present-day Atlantic meridional overturning circulation to freshwater forcing. *Geophys. Res. Lett.*, **30**, 1898, doi:10.1029/2003GL017578.

Rowell, D. P., C. K. Folland, K. Maskell, and M. N. Ward, 1995: Variability of summer rainfall over tropical North-Africa (1906–92) observations and modelling. *Q J R Meteorol. Soc.*, **121**, 669–704.

Rowell, D. P., 2003: The impact of Mediterranean SSTs on the Sahelian rainfall season. *J Clim.*, **16**, 849– 862.

Schott, F. A., and Coauthors, 2008: Pacific Subtropical Cell variability in the SODA 2.0.2/3 assimilation. *Geophys. Res. Lett.*, **35**, L10607, doi:10.1029/2008GL033757.

Schott, F. A., W. Wang, D. Stramma, 2007: Variability of Pacific subtropical cells in the 50-year ECCO assimilation. *Geophys. Res. Lett.*, **34**, L05604, doi:10.1029/2006GL028478.

Shay, L. K., G. J. Goni, and P. G. Black, 2000: Effects of a Warm Oceanic Feature on Hurricane Opal. *Mon. Wea. Rev.*, **128**, 1366–1383

Smith, T. M., R. W. Reynolds, 2004: Improved extended reconstruction of SST (1854–1997). *J Clim.*, **17**, 2466–2477.

1   Stocker, T. F., A. Timmermann, M. Renold, and O. Timm, 2007: Effects of salt compensation on

2   the climate model response in simulations of large changes of the Atlantic meridional

3   overturning circulation. J. Climate, **20**, 5912–5928.

4   Sutton, R. T., and D. L. R. Hodson, 2005: Atlantic Ocean forcing of North American and

5   European summer climate. *Science*, **309**, 115–118.

6   Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012: An overview of CMIP5 and the experiment

7   design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498.

8   Terray, L., 2012: Evidence for multiple drivers of North Atlantic multidecadal climate variability.

9   *Geophys. Res. Lett.*, **39**, L19712, doi:10.1029/2012GL053046.

10   Ting, M., Y, Kushnir, R. Seager, and C. Li, 2009: Forced and internal twentieth-century SST

11   trends in the North Atlantic. *J. Climate.*, **22**, 1469–1481.

12   Wang, C., S. dong, E. Munoz, 2010: Seawater density variations in the North Atlantic and the

13   Atlantic meridional overturning circulation. *Clim Dyn.*, **34**, 953-968.

14   Wang, C., and S.-K. Lee, 2009: Co-variability of Tropical Cyclones in the North Atlantic and the

15   Eastern North Pacific. *Geophys. Res. Lett.*, **36**, L24702, doi:10.1029/2009GL041469.

16   Wang, C., S. K. Lee, and D. B. Enfield, 2008: Atlantic Warm Pool acting as a link between

17   Atlantic Multidecadal Oscillation and Atlantic tropical cyclone activity. *Geochem. Geophys.*

18   *Geosyst.*, **9**, Q05V03 doi:10.1029/2007GC001809.

19   Wang, C., S. Dong, A. T. Evan, G.R. Foltz, and S.-K. Lee, 2012: Multidecadal covariability of

20   North Atlantic sea surface temperature, African dust, Sahel rainfall and Atlantic hurricanes.

21   *J. Climate*, **25**, 5404-5415.

22   Wang, C., L. Zhang, and S.-K. Lee, 2013: Response of freshwater flux and sea surface salinity to

23   variability of the Atlantic warm pool. *J. Climate*, in press.

24   Zhang, D., M. J. McPhaden, and W. E. Johns, 2003: Observational evidence for flow between

25   the subtropical and tropical Atlantic: The Atlantic subtropical cells. *J. Phys. Oceanogr.*, **33**,

26   1783–1797.

27   Zhang, L., C. Wang, and L. Wu, 2012a: Low-frequency modulation of the Atlantic warm pool by

the Atlantic multidecadal oscillation. *Climate Dynamics*, **39**, 1661–1671.

Zhang, L., L. Wu, and J. Zhang, 2012b: Simulated response to recent freshwater flux change over the Gulf Stream and its extension: Coupled ocean–atmosphere adjustment and Atlantic–Pacific teleconnection. *J. Climate*, **24**, 15, 3971-3988.

Zhang, R., 2007: Anticorrelated multidecadal variations between surface and subsurface tropical North Atlantic. *Geophys. Res. Lett*., **34**, L12713, doi:10.1029/2007GL030225.

Zhang, R., 2010: Latitudinal dependence of Atlantic Meridional Overturning Circulation (AMOC) variations. *Geophys. Res. Lett*., **37**, L16703, doi:10.1029/2010GL044474.

Zhang, R., and Coauthors, 2013: Have aerosols caused the observed Atlantic multidecadal variability? *J. Atmos. Sci*., in press.

1

2   Table 1.   The 18 CMIP5 models used in this study.

Sponsor, Country	Model Name	Temporal Coverage
Beijing Climate Center, China	bcc-csm1-1	1850/01-2012/12
Canadian Center for Climate Modeling and Analysis, Canada	CanESM2	1850/01-2005/12
National Center for Atmospheric Research (NCAR), USA	CCSM4	1850/01-2005/12
Météo-France/Centre National de Recherches Météorologiques, France	CNRM-CM5	1850/01-2005/12
Commonwealth Scientific and Industrial Research Organisation (CSIRO), Australia	CSIRO-Mk3-6-0	1850/01-2005/12
Institute of Atmospheric Physics, Chinese Academy of Sciences, China	FGOALS-g2	1900/01-2005/12
U.S. Department of Commerce/National Oceanic and Atmospheric Administration (NOAA)/Geophysical Fluid Dynamics Laboratory (GFDL), USA	GFDL-ESM2G	1861/01-2005/12
	GFDL-ESM2M	1861/01-2005/12
National Aeronautics and Space Administration (NASA)/Goddard Institute for Space Studies (GISS), USA	GISS-E2-H	1850/01-2005/12
	GISS-E2-R	1850/01-2005/12
Met office Hadley Centre, UK	HadGEM2-ES	1859/12-2005/11
Institute Pierre Simon Laplace, France	IPSL-CM5A-LR	1850/01-2005/12
	IPSL-CM5A-MR	1850/01-2005/12
Center for Climate System Research (University of Tokyo), National Institute for Environmental Studies, and Frontier Research Center for Global Change (JAMSTEC), Japan	MIROC5	1850/01-2005/12
	MIROC-ESM	1850/01-2005/12
	MIROC-ESM-CHEM	1850/01-2005/12
Max Planck Institute for Meteorology, Germany	MPI-ESM-P	1850/01-2005/12
Meteorological Research Institute, Japan	MRI-CGCM3	1850/01-2005/12

3

4

Figure Captions

3   **Figure 1.**   Variations of ocean temperature in the North Atlantic.   Shown are the normalized

4   (by standard deviation) AMO index from the (a) SODA and (e) Ishii data, the regression (°C) of

5   the Atlantic zonal mean temperature onto the normalized AMO index from the (b) SODA and (f)

6   Ishii data, and regression (°C per °C) of the Atlantic temperature averaged in the region of

7   8°N-20°N and coast to coast at each depth onto the SST on multidecadal (period of T higher than

8   7 years) and interannual (period of T lower than 7 years) timescales from the (c, d) SODA data,

9   and (g, h) Ishii data.   The AMO is defined by the detrended SST anomalies averaged in the

10   North Atlantic Ocean (0° to 60°N, coast to coast) and smoothed by a 7-year low-frequency filter.

11

12   **Figure 2.**   Multidecadal time series of TNA (8°N to 20°N, coast to coast) ocean temperature

13   anomalies at the sea surface and 200 m and vertically averaged temperature anomalies from

14   0-400 m using the (a, b, c) SODA and (d, e, f) Ishii data.   All time series are normalized by their

15   standard deviations.

16

17   **Figure 3.**   The AMO indices in observation (ERSST) and CMIP5 models.   The AMO index

18   (°C) is calculated by the detrended SST anomalies averaged in the North Atlantic Ocean (0° to

19   60°N, coast to coast) and smoothed by a 7-year low-frequency filter.

20

21   **Figure 4.**   Regression of the Atlantic zonal mean temperature (°C) onto the normalized AMO

22   index from CMIP5 models.

23

24   **Figure 5.**   Long-term mean background states in the TNA.   The mean (a) surface wind (m/s)

25   and net heat flux (positive downward, W/m^2) and (b) freshwater flux (evaporation minus

26   precipitation, mm/day) are from the 20CRv2 data.   The zonal (coast to coast) mean (c)

27   temperature (°C), (d) salinity (psu), (e) meridional velocity (10^{-2} m/s) and (f) vertical velocity

1   (10^{-6} m/s) are from the SODA data.

2

3   **Figure 6.**   Long-term mean (a) meridional temperature gradient (10^{-6} °C/m), (b) meridional

4   salinity gradient (10^{-6} psu/m), (c) vertical temperature gradient (10^{-2} °C/m), and (d) vertical

5   salinity gradient (10^{-2} psu/m) from the SODA data.

6

7   **Figure 7.**   Regression of advection terms (10^{-8} °C/s) onto the normalized AMO index from the

8   SODA data.   Show are the (a) total meridional temperature advection, (b) advection by the

9   anomalous meridional current, (c) advection by the mean meridional current, and (d) nonlinear

10 term.   (e-h) are the same as (a-d) but for the vertical temperature advection.   The gray line in

11 (b) denotes the mean mixed layer depth, defined by the depth where the difference with the

12 potential density at the first layer equals to 0.125 kg m^{-3}.

13

14 **Figure 8.**   (a) Regression of surface net heat flux (shading, positively downward, W/m^2), wind

15 stress (vector, N/m^2) and wind stress curl (contour, contour interval is 1.0×10^{-9} N/m^3, positive

16 and negative values are denoted by the green and gray lines, respectively) onto the normalized

17 AMO index, (b) long-term mean AMOC streamfunction (Sv), (c) the TNA AMOC index, (d) the

18 STC index, lead-lag correlations of the AMO with (e) the TNA AMOC index and (f) the STC

19 index.   In (e) and (f), the unit in x-axis is year, and positive (negative) year indicates the

20 AMOC/STC leads (lags) the AMO.   The horizontal dash lines represent the 90% confidence

21 level.   The AMO and streamfunction are from the SODA data, and the heat flux and wind are

22 from 20CRv2 reanalysis.   The TNA AMOC index is defined as the maximum streamfunction

23 below 300 m between 5°N to 20°N, and the STC index is defined as the maximum

24 streamfunction in the upper 250 m between 5°N to 20°N.

25

26 **Figure 9.**   Regression of advection terms (10^{-8} °C/s) onto the normalized AMO index.   The

27 regression is the ensemble mean of 12 CMIP5 models that simulate the TNA surface and

1 subsurface temperature variation reasonably well.   Show are the (a) total meridional

2 temperature advection, (b) advection by the anomalous meridional current, (c) advection by the

3 mean meridional current, and (d) nonlinear term.   (e-h) are the same as (a-d) but for the vertical

4 temperature advection.

5

6 **Figure 10.**   Regression of surface net heat flux (shading, positively downward, W/m^2), wind

7 stress (vector, N/m^2) and wind stress curl (contour, contour interval is 1.0×10^{-9} N/m^3, positive

8 and negative values are denoted by the green and gray lines, respectively) onto the corresponding

9 normalized AMO index in CMIP5 models.

10

11 **Figure 11.**   Long-term mean AMOC streamfunction (Sv) from CMIP5 models.

12

13 **Figure 12.**   The TNA AMOC index defined as the maximum streamfunction below 300 m from

14 5°N to 20°N from CMIP5 models.   All time series are normalized by their standard deviation.

15

16 **Figure 13.**   Lead-lag correlations between the AMO and TNA AMOC index from CMIP5

17 models.   The unit in x-axis is year.   Positive (negative) year in x-axis means the AMOC leads

18 (lags) the AMO.   The horizontal dash line indicates the 90% confidence level.

19

20 **Figure 14.**   The Atlantic STC index defined as the maximum streamfunction in the upper 250

21 m between 5°N to 20°N from CMIP5 models.   All time series are normalized by their standard

22 deviation.

23

24 **Figure 15.**   Lead-lag correlations between the AMO and STC index from CMIP5 models.

25 The unit in x-axis is year.   Positive (negative) year in x-axis means the STC leads (lags) the

26 AMO.   The horizontal dash line indicates the 90% confidence level.

27

Figure 16. Regression of the Atlantic zonal mean salinity (psu) onto the normalized AMO index from (a-r) CMIP5 models, (s) SODA data and (t) Ishii data.

Figure 17. Regression of salinity advection terms (10^{-8} psu/s) onto the normalized AMO index. The regression is the ensemble mean of 12 CMIP5 models that simulate the TNA surface and subsurface temperature variation reasonably well. Show are the (a) total meridional salinity advection, (b) advection by the anomalous meridional current, (c) advection by the mean meridional current, and (d) nonlinear term.

Figure 18. Schematic diagram showing the variations of the AMOC, STC, ocean temperature and surface wind with the AMO on multidecadal timescales.

1
2
3
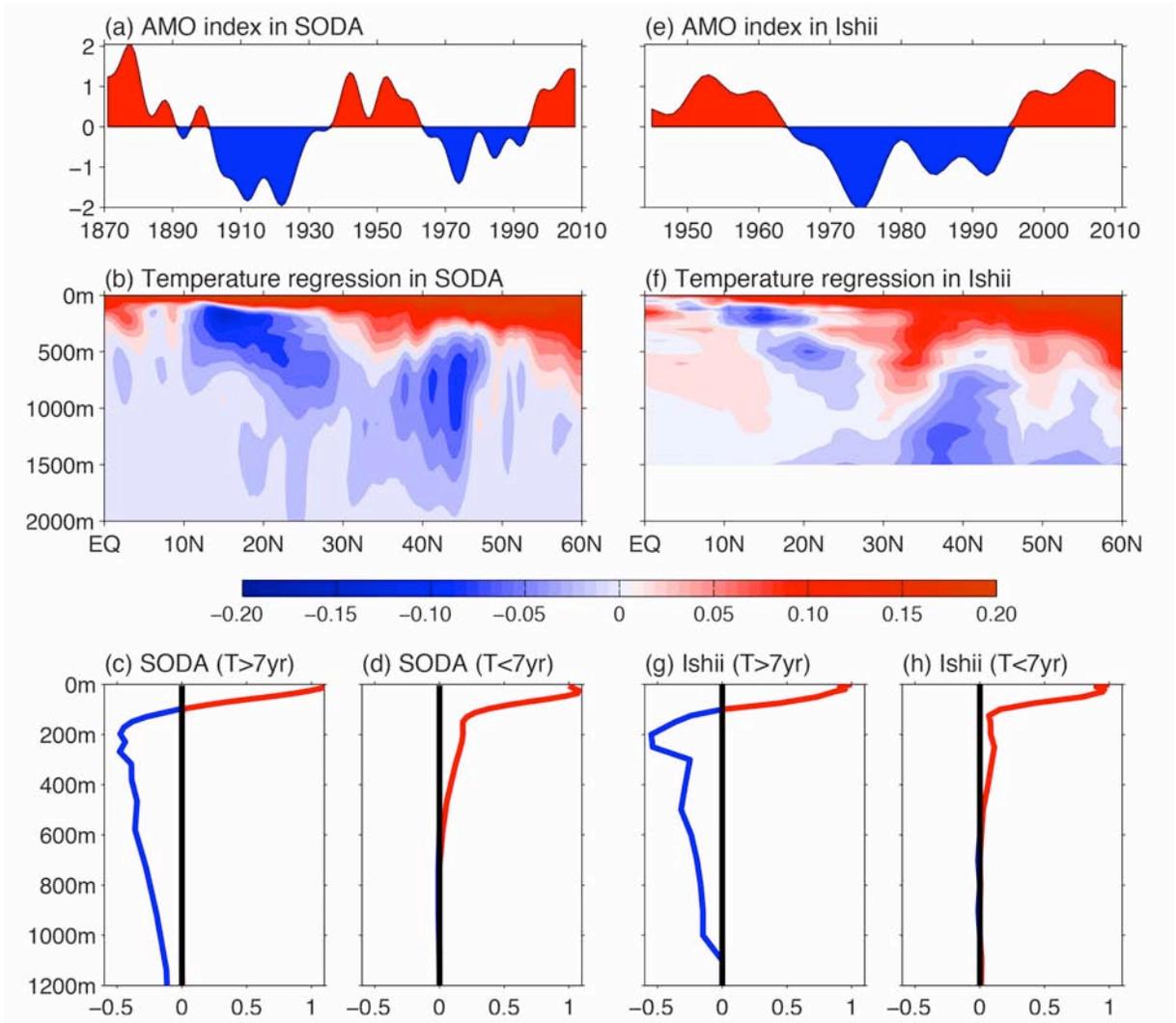
4
5

6 **Figure 1.** Variations of ocean temperature in the North Atlantic.   Shown are the normalized (by standard
7 deviation) AMO index from the (a) SODA and (e) Ishii data, the regression (°C) of the Atlantic zonal mean
8 temperature onto the normalized AMO index from the (b) SODA and (f) Ishii data, and regression (°C per °C) of the
9 Atlantic temperature averaged in the region of 8°N-20°N and coast to coast at each depth onto the SST on
10 multidecadal (period of T higher than 7 years) and interannual (period of T lower than 7 years) timescales from the
11 (c, d) SODA data, and (g, h) Ishii data.   The AMO is defined by the detrended SST anomalies averaged in the
12 North Atlantic Ocean (0° to 60°N, coast to coast) and smoothed by a 7-year low-frequency filter.
13
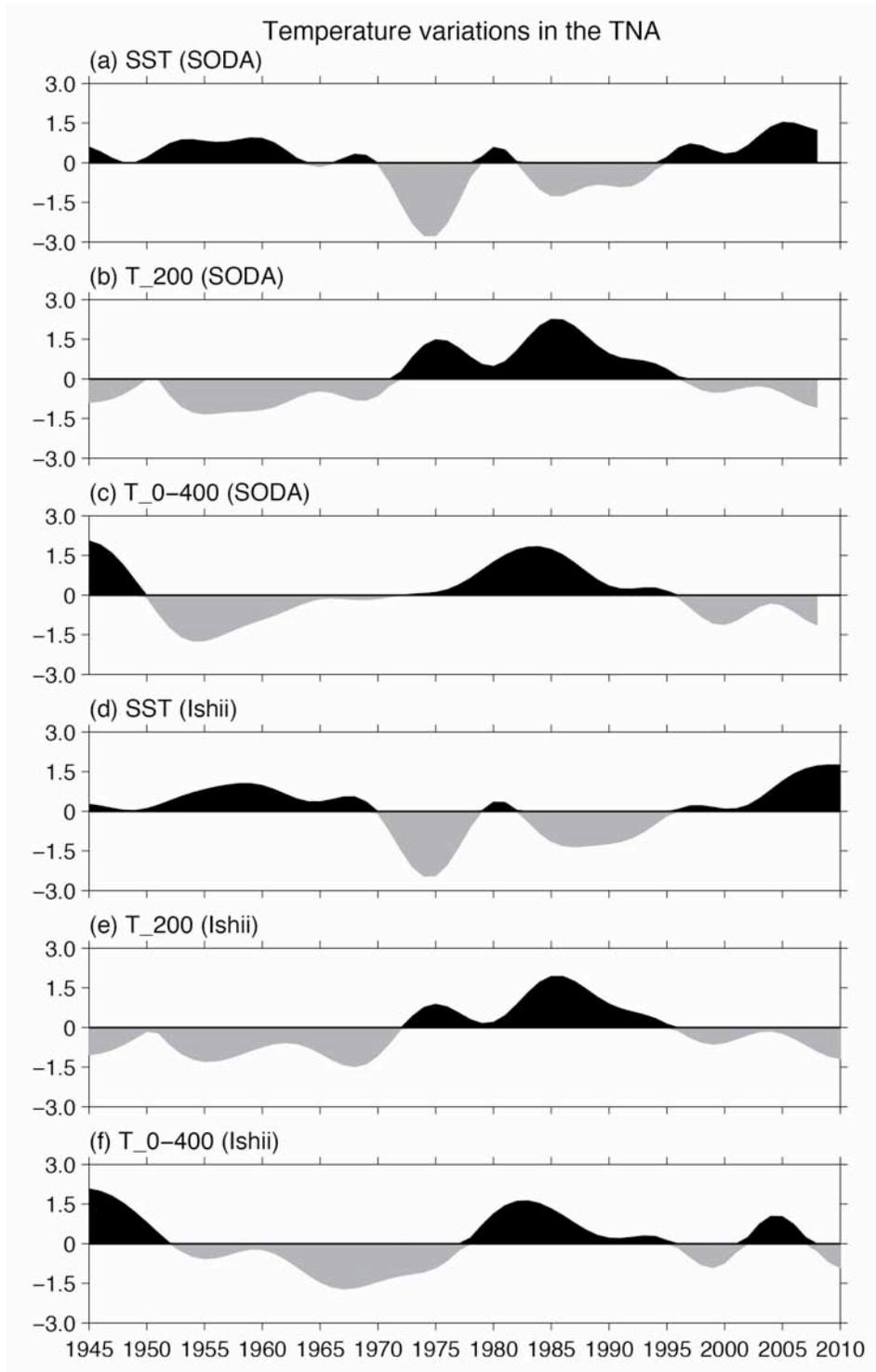
1
2
3   **Figure 2.**   Multidecadal time series of TNA (8°N to 20°N, coast to coast) ocean temperature anomalies at the sea
4   surface and 200 m and vertically averaged temperature anomalies from 0-400 m using the (a, b, c) SODA and (d, e, f)
5   Ishii data.   All time series are normalized by their standard deviations.
6

41
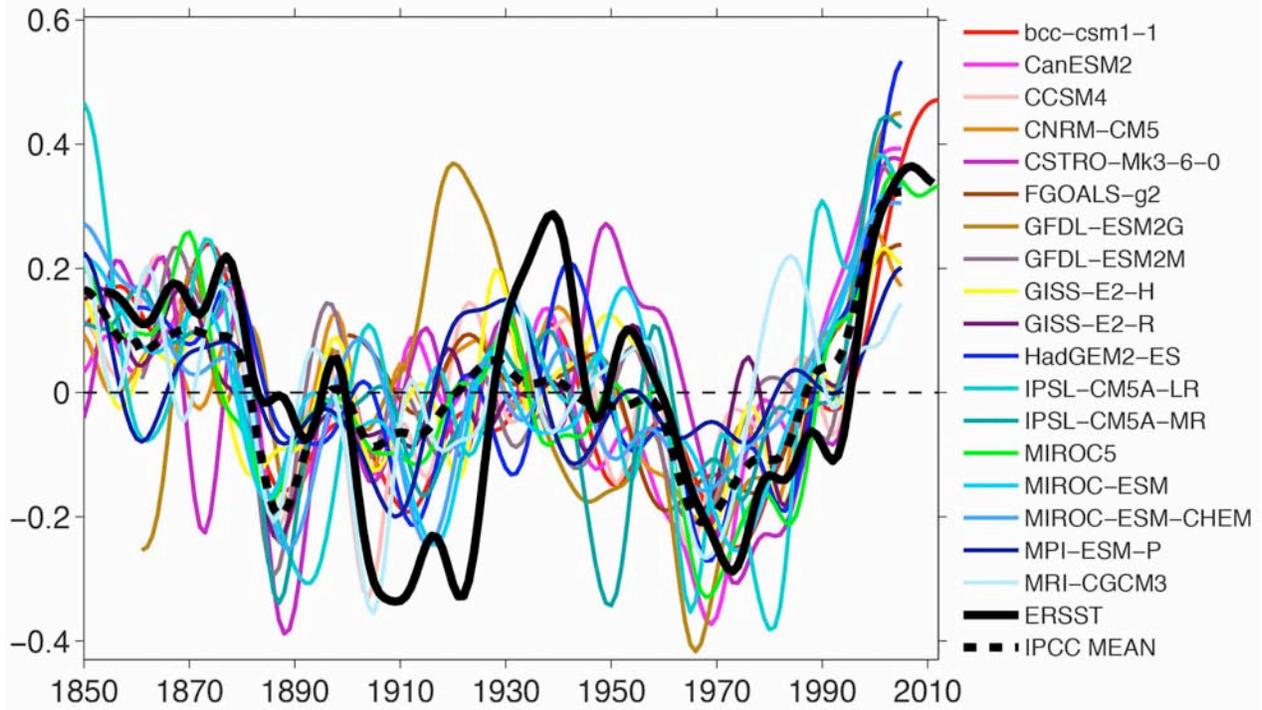
Figure 3. The AMO indices in observation (ERSST) and CMIP5 models. The AMO index (°C) is calculated by the detrended SST anomalies averaged in the North Atlantic Ocean (0° to 60°N, coast to coast) and smoothed by a 7-year low-frequency filter.
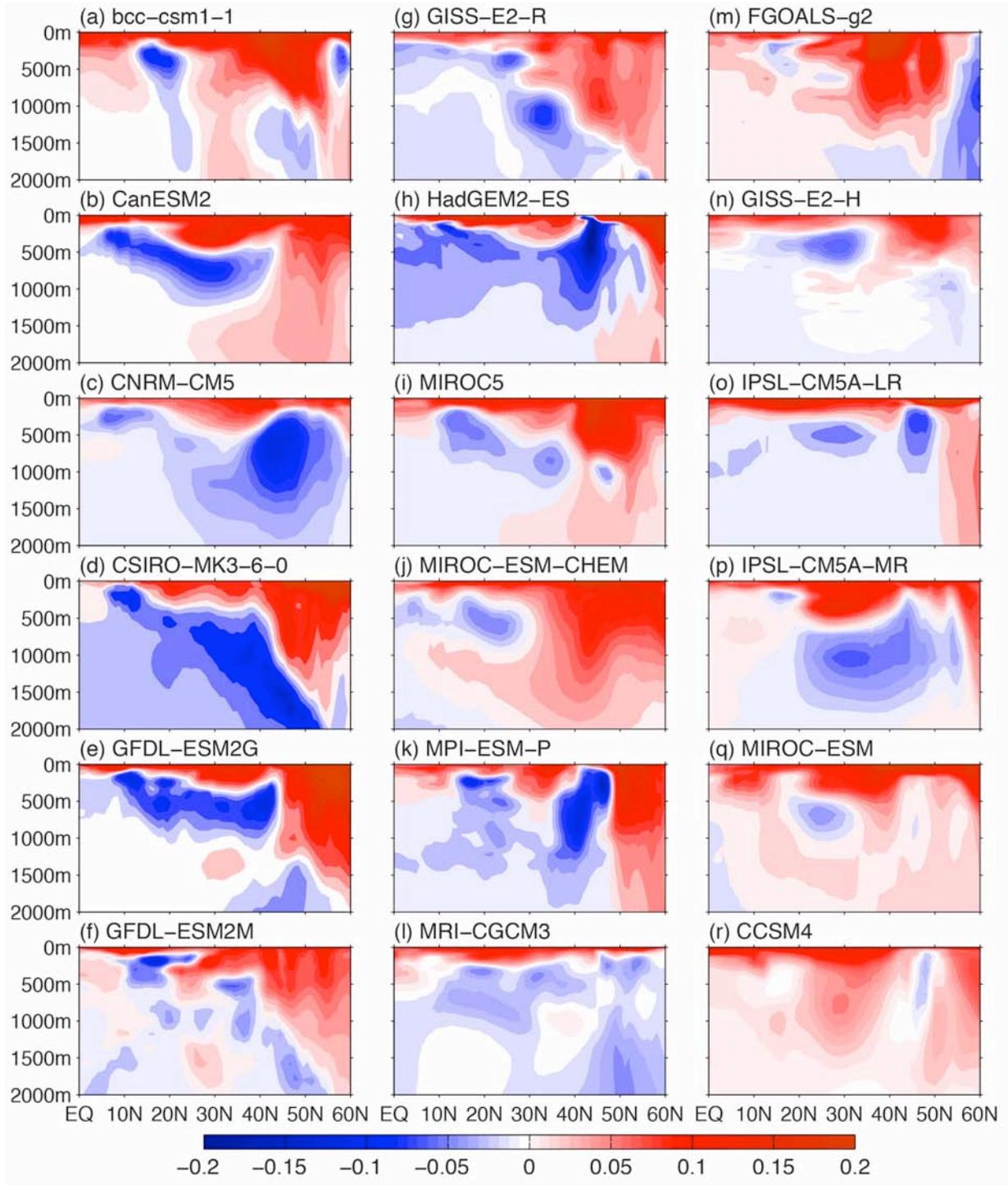
Figure 4. Regression of the Atlantic zonal mean temperature (°C) onto the normalized AMO index from CMIP5 models.

43

1

2
3
4   **Figure 5.**   Long-term mean background states in the TNA.   The mean (a) surface wind (m/s) and net heat flux
5   (positive downward, W/m^2) and (b) freshwater flux (evaporation minus precipitation, mm/day) are from the 20CRv2
6   data.   The zonal (coast to coast) mean (c) temperature (°C), (d) salinity (psu), (e) meridional velocity (10^{-2} m/s) and
7   (f) vertical velocity (10^{-6} m/s) are from the SODA data.
8

1
2
3
4

(a) Meridional temperature gradient   (b) Meridional salinity gradient

(c) Vertical temperature gradient   (d) Vertical salinity gradient
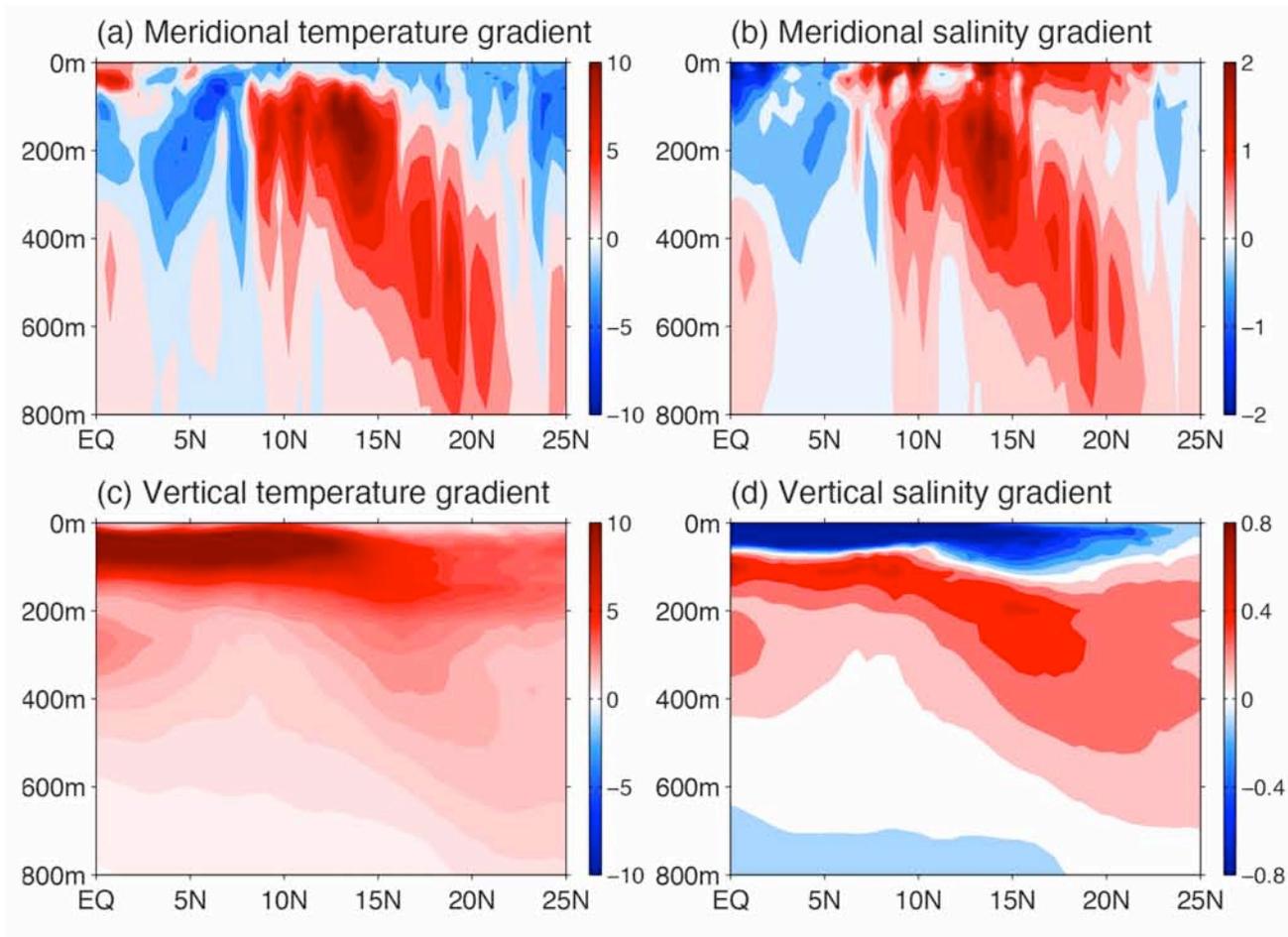
5
6
7 **Figure 6.**   Long-term mean (a) meridional temperature gradient (10^{-6} °C/m), (b) meridional salinity gradient (10^{-6}
8 psu/m), (c) vertical temperature gradient (10^{-2} °C/m), and (d) vertical salinity gradient (10^{-2} psu/m) from the SODA
9 data.
10

45

Figure 7. Regression of advection terms (10^{-8} °C/s) onto the normalized AMO index from the SODA data. Show are the (a) total meridional temperature advection, (b) advection by the anomalous meridional current, (c) advection by the mean meridional current, and (d) nonlinear term. (e-h) are the same as (a-d) but for the vertical temperature advection. The gray line in (b) denotes the mean mixed layer depth, defined by the depth where the difference with the potential density at the first layer equals to 0.125 kg m^{-3}.

1
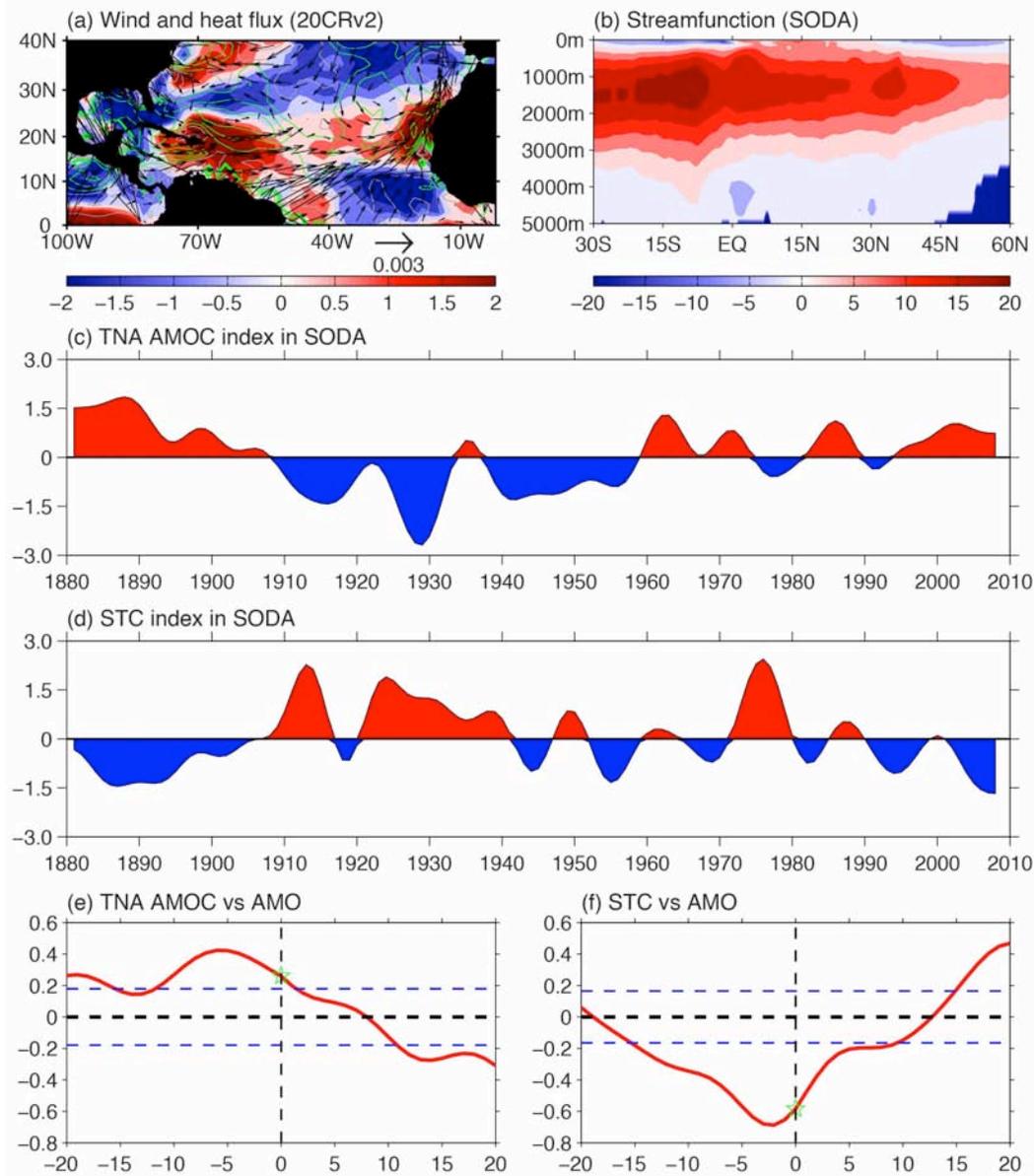
2

3

4 **Figure 8.** (a) Regression of surface net heat flux (shading, positively downward, W/m^2), wind stress (vector, N/m^2)

5 and wind stress curl (contour, contour interval is 1.0×10^{-9} N/m^3, positive and negative values are denoted by the

6 green and gray lines, respectively) onto the normalized AMO index, (b) long-term mean AMOC streamfunction (Sv),

7 (c) the TNA AMOC index, (d) the STC index, lead-lag correlations of the AMO with (e) the TNA AMOC index and

8 (f) the STC index. In (e) and (f), the unit in x-axis is year, and positive (negative) year indicates the AMOC/STC

9 leads (lags) the AMO. The horizontal dash lines represent the 90% confidence level. The AMO and

10 streamfunction are from the SODA data, and the heat flux and wind are from 20CRv2 reanalysis. The TNA

11 AMOC index is defined as the maximum streamfunction below 300 m between 5°N to 20°N, and the STC index is

12 defined as the maximum streamfunction in the upper 250 m between 5°N to 20°N.

13

Figure 9. Regression of advection terms (10^{-8} °C/s) onto the normalized AMO index. The regression is the ensemble mean of 12 CMIP5 models that simulate the TNA surface and subsurface temperature variation reasonably well. Show are the (a) total meridional temperature advection, (b) advection by the anomalous meridional current, (c) advection by the mean meridional current, and (d) nonlinear term. (e-h) are the same as (a-d) but for the vertical temperature advection.

Figure 10.   Regression of surface net heat flux (shading, positively downward, W/m^2), wind stress (vector, N/m^2) and wind stress curl (contour, contour interval is 1.0×10^{-9} N/m^3, positive and negative values are denoted by the green and gray lines, respectively) onto the corresponding normalized AMO index in CMIP5 models.

1
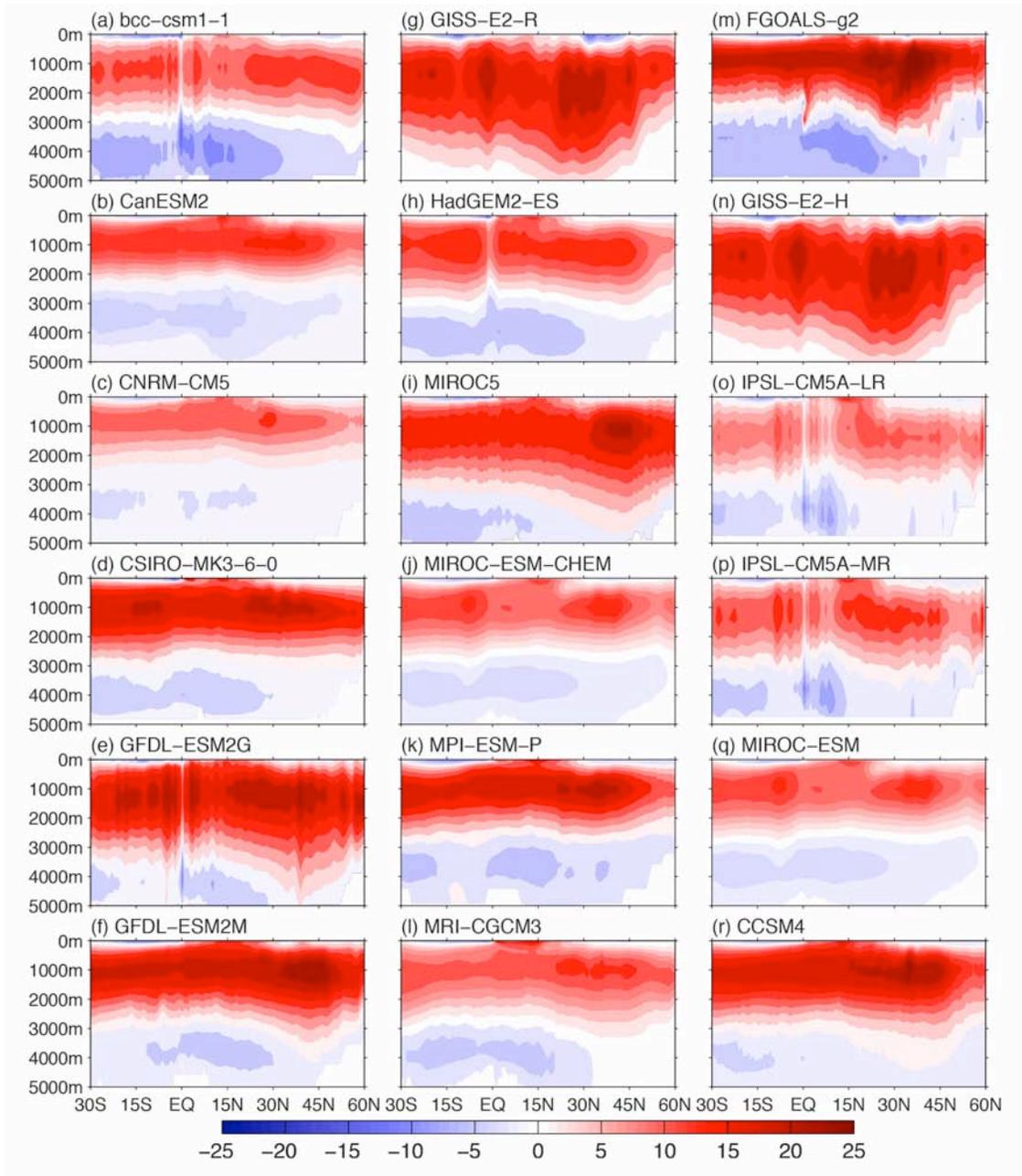
2

3

4 **Figure 11.** Long-term mean AMOC streamfunction (Sv) from CMIP5 models.

5

1
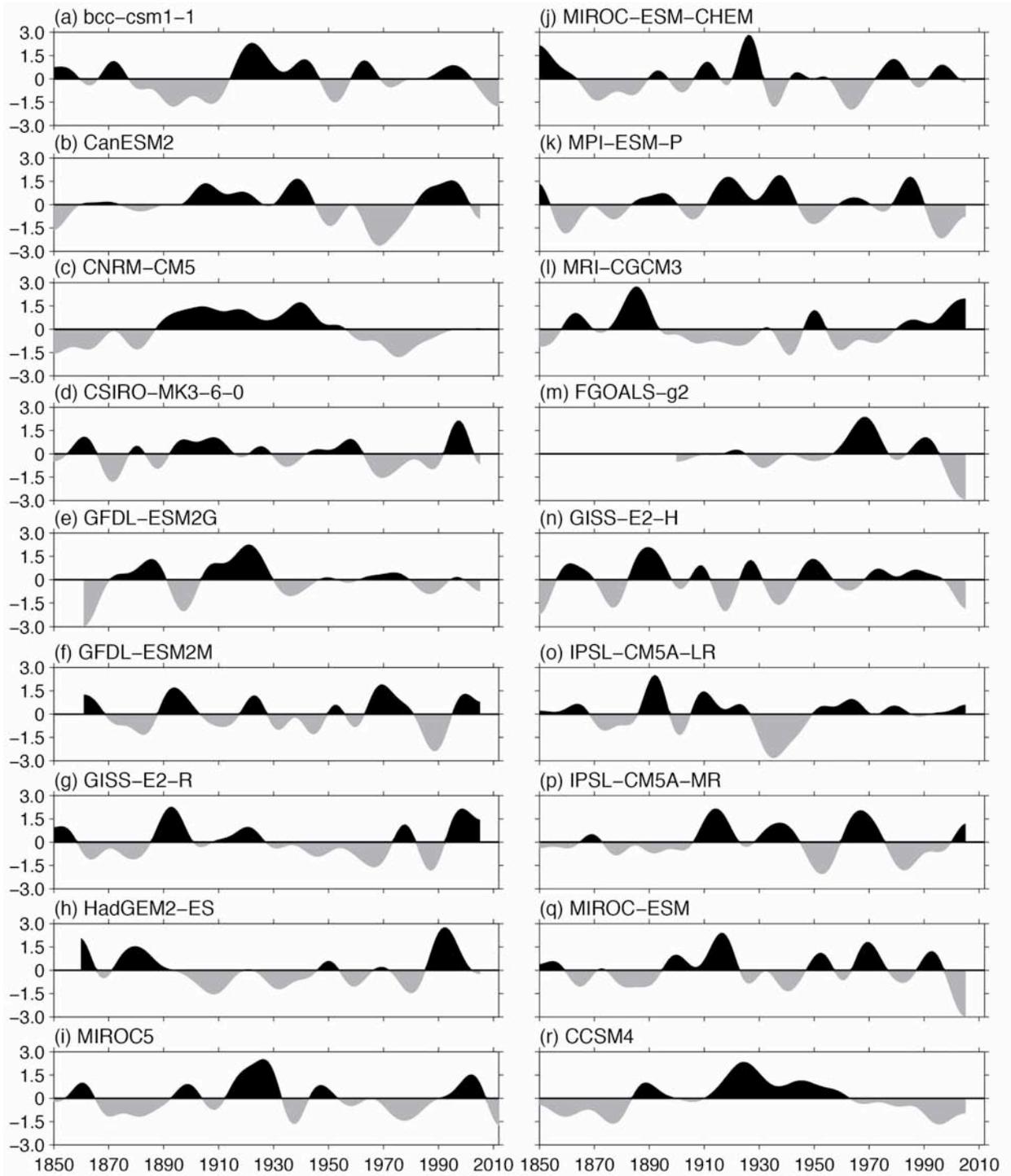
2
3
4   **Figure 12.**   The TNA AMOC index defined as the maximum streamfunction below 300 m from 5°N to 20°N from
5   CMIP5 models.   All time series are normalized by their standard deviation.
6

51

1
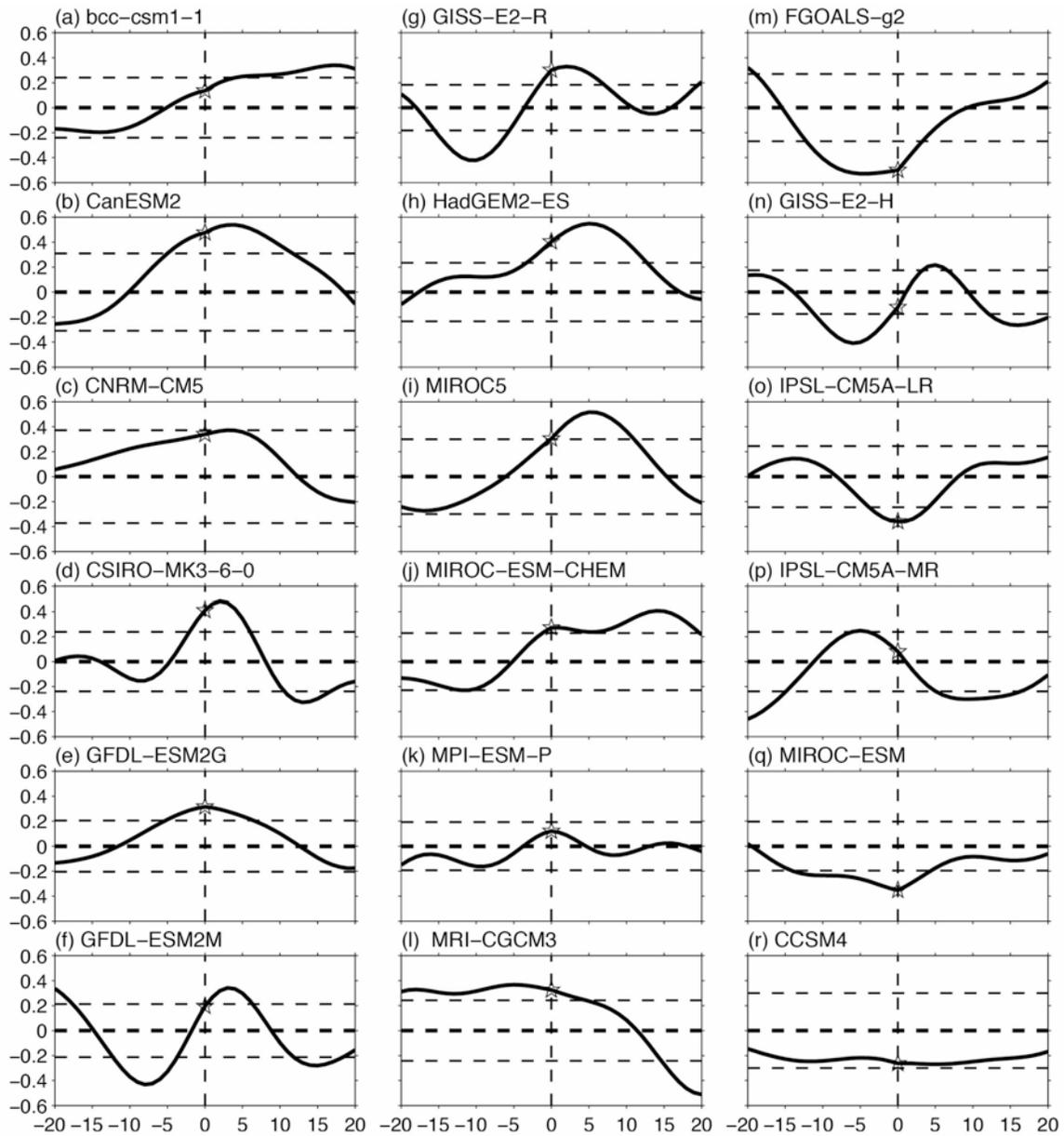
2
3
4 **Figure 13.**   Lead-lag correlations between the AMO and TNA AMOC index from CMIP5 models.   The unit in
5 x-axis is year.   Positive (negative) year in x-axis means the AMOC leads (lags) the AMO.   The horizontal dash
6 line indicates the 90% confidence level.
7
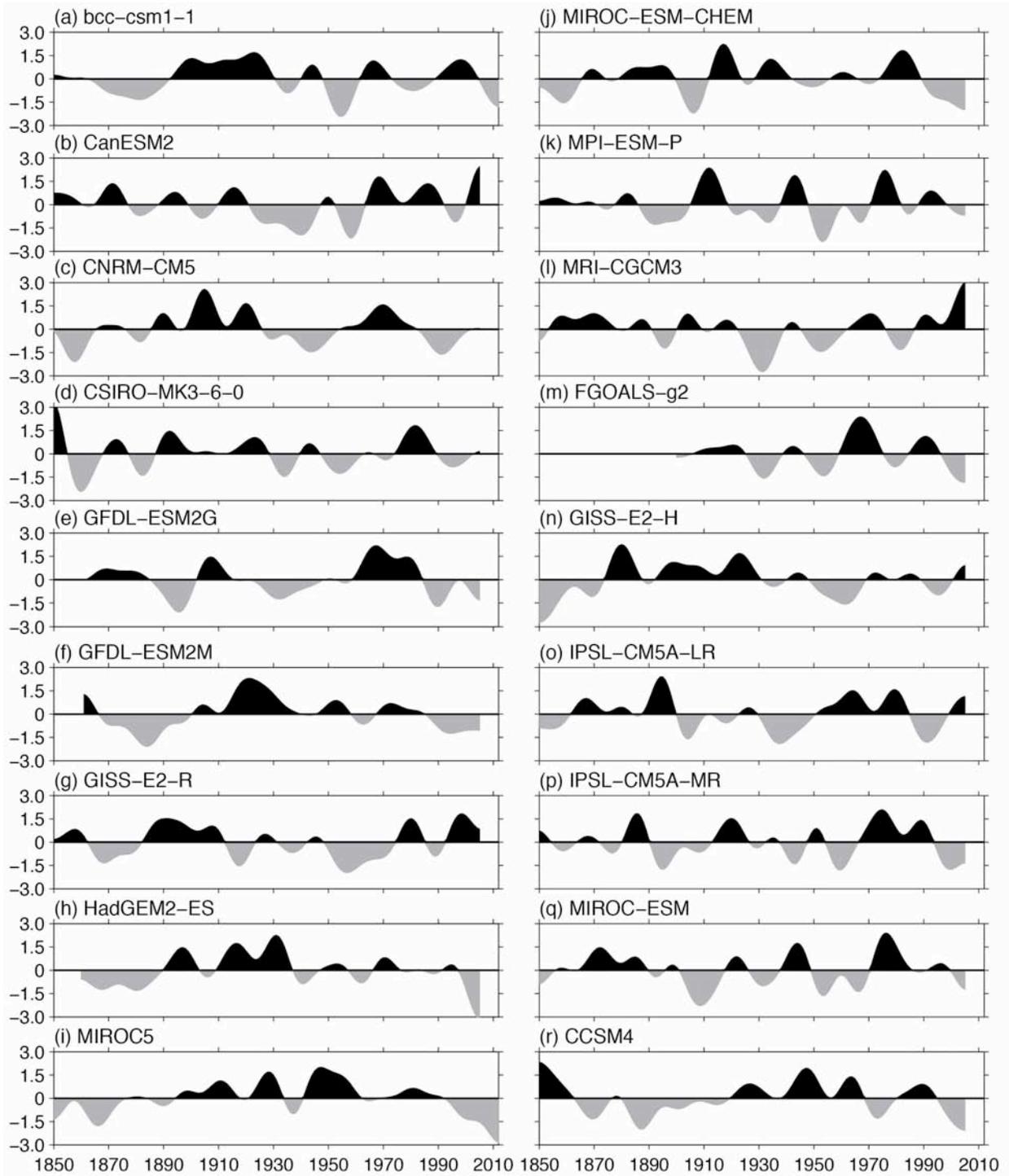
52

1

2
3
4 **Figure 14.** The Atlantic STC index defined as the maximum streamfunction in the upper 250 m between 5°N to
5 20°N from CMIP5 models. All time series are normalized by their standard deviation.
6

1
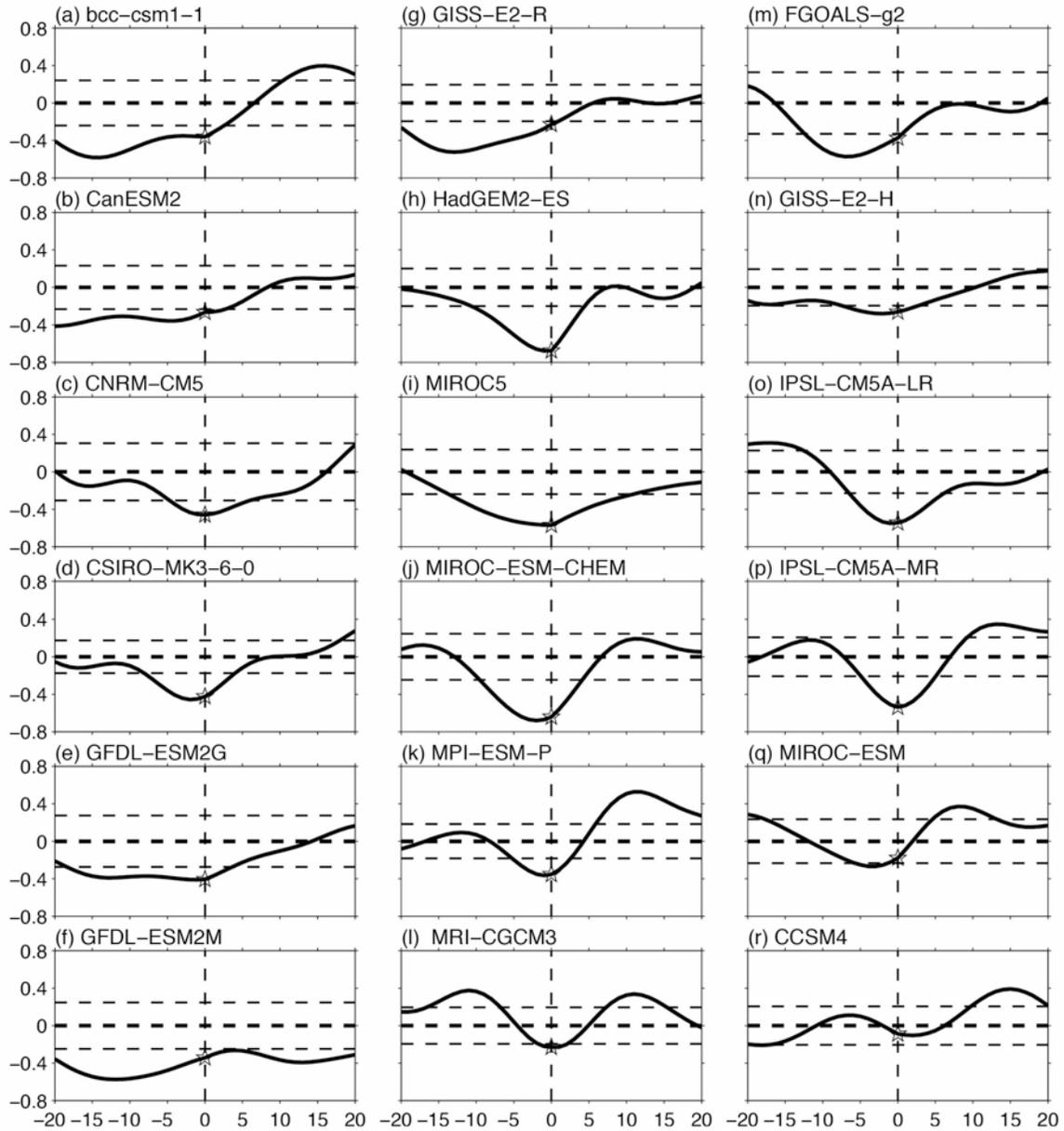
2

3

4 **Figure 15.**   Lead-lag correlations between the AMO and STC index from CMIP5 models.   The unit in x-axis is

5 year.   Positive (negative) year in x-axis means the STC leads (lags) the AMO.   The horizontal dash line indicates

6 the 90% confidence level.
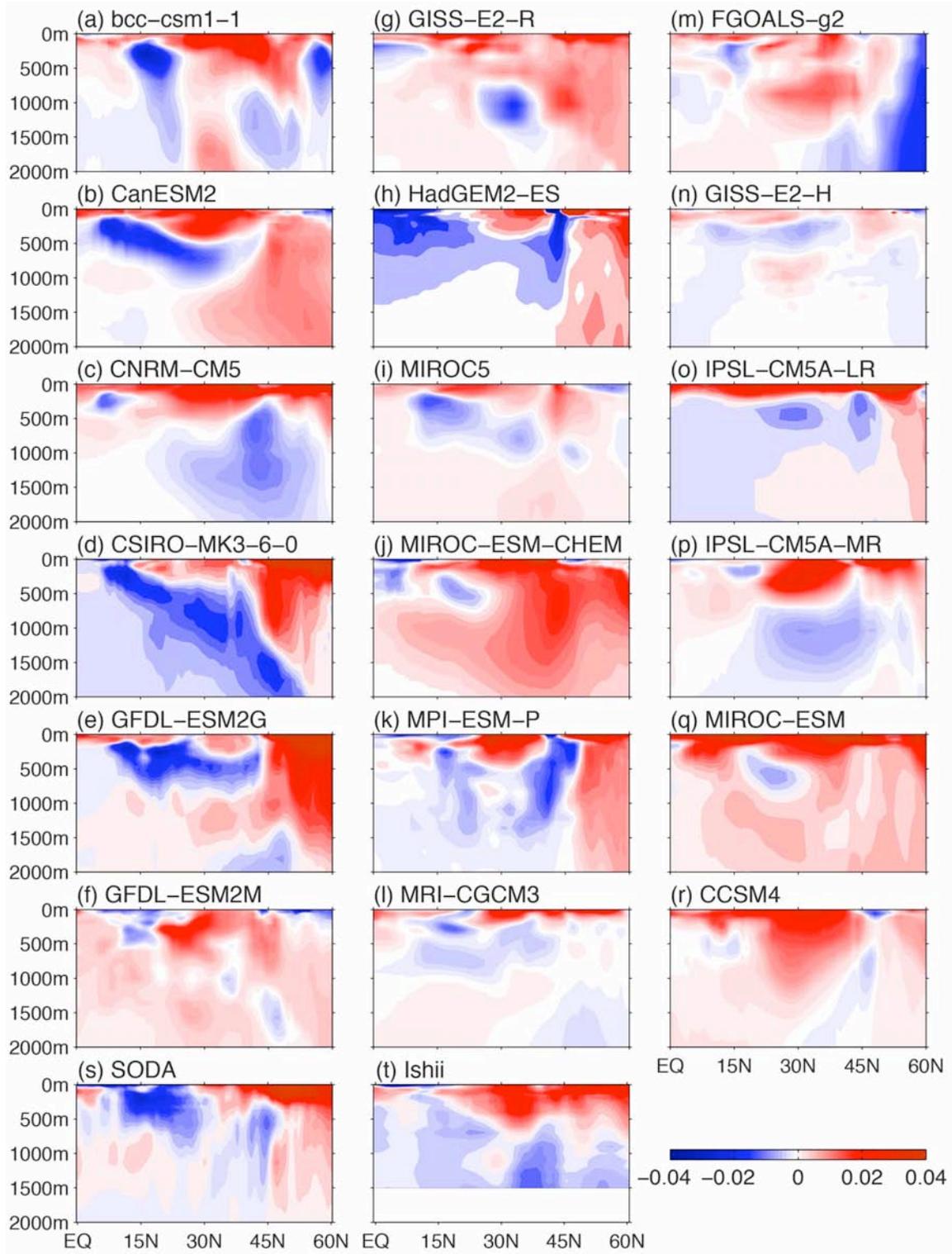
7

Figure 16. Regression of the Atlantic zonal mean salinity (psu) onto the normalized AMO index from (a-r) CMIP5 models, (s) SODA data and (t) Ishii data.

1
2
3

4
5

6   **Figure 17.**   Regression of salinity advection terms (10^{-8} psu/s) onto the normalized AMO index.   The regression
7   is the ensemble mean of 12 CMIP5 models that simulate the TNA surface and subsurface temperature variation
8   reasonably well.   Show are the (a) total meridional salinity advection, (b) advection by the anomalous meridional
9   current, (c) advection by the mean meridional current, and (d) nonlinear term.
10

1
2
3

Variations of the AMOC, STC, Ocean Temperature and Wind with the AMO

S

EQ

N

Warm

Cool

Anomalous STC

$$-V'\frac{\partial \overline{T}}{\partial y} < 0$$

Anomalous AMOC

4
5
6 **Figure 18.**   Schematic diagram showing the variations of the AMOC, STC, ocean temperature and surface wind
7 with the AMO on multidecadal timescales.
8