Deep Atlantic Ocean Warming Facilitated by the Deep Western Boundary Current and Equatorial Kelvin Waves

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ABSTRACT

Increased heat storage in deep oceans has been proposed to account for the slowdown of global surface warming since the end of the twentieth century. How the imbalanced heat at the surface has been redistributed to deep oceans remains to be elucidated. Here, the evolution of deep Atlantic Ocean heat storage since 1950 on multidecadal or longer time scales is revealed. The anomalous heat in the deep Labrador Sea was transported southward by the shallower core of the deep western boundary current (DWBC). Upon reaching the equator around 1980, this heat transport route bifurcated into two, with one continuing southward along the DWBC and the other extending eastward along a narrow strip (about 4° width) centered at the equator. In the 1990s and 2000s, meridional diffusion helped to spread warming in the tropics, making the eastward equatorial warming extension have a narrow head and wider tail. The deep Atlantic Ocean warming since 1950 had overlapping variability of approximately 60 years. The results suggest that the current basinwide Atlantic Ocean warming at depths of 1000–2000 m can be traced back to the subsurface warming in the Labrador Sea in the 1950s. An inference from these results is that the increased heat storage in the twenty-first century in the deep Atlantic Ocean is unlikely to partly account for the atmospheric radiative imbalance during the last two decades and to serve as an explanation for the current warming hiatus.

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1. Introduction

The global warming in the twenty-first century has slowed down despite the continuing rapid accumulation of greenhouse gases in the atmosphere (Trenberth and Fasullo 2010; IPCC 2013). Different mechanisms have been proposed to explain this warming “hiatus,” especially in quantifying the contributions from internal variability and anthropogenic forcing (Fröhlich 2012; Solomon et al. 2010, 2011; Kaufmann et al. 2011; Meehl et al. 2011; Wu et al. 2011a). A common feature

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of these mechanisms is the uptake of excessive heat by global oceans, as inferred by the atmospheric radiative budget at Earth's surface (Katsman and van Oldenborgh 2011; Kosaka and Xie 2013; England et al. 2014; Chen and Tung 2014; Nieves et al. 2015; Lee et al. 2015; Liu et al. 2016; Gao et al. 2018), implying the great role of global oceans in modulating Earth's climate system.

Currently, there are two major theories of ocean heat uptake that have been proposed to explain the warming hiatus. One suggests that the La Niña-like decadal cooling in the eastern Pacific Ocean and intensified trade winds over the equatorial Pacific, corresponding to the negative phase of the interdecadal Pacific oscillation (IPO), had taken up the "missing heat" (Zhang et al. 1997; Kosaka and Xie 2013; England et al. 2014; Watanabe et al. 2014; Nieves et al. 2015). The intensified trade winds produce the equatorial ocean Kelvin waves and deepen the equatorial thermocline, leading to the upper oceans of larger depth to uptake surface heat (Liu et al. 2016). The other focuses on the vertical heat redistributing to the deeper oceans on decadal or multidecadal time scales (Meehl et al. 2011; Chen and Tung 2014). It was calculated that the surface warming hiatus has been accompanied by more than 30% of the total increment of ocean heat content in deep oceans below 750 m in the Atlantic and Southern Oceans and below 300 m in the Pacific and Indian Oceans (Meehl et al. 2011; Balmaseda et al. 2013). In addition, a large systematic increase of ocean heat content is not limited to tropical Pacific and Atlantic Oceans; rather, it is observed in multiple ocean basins (Drijfhout et al. 2014), with the deep Atlantic and Southern Oceans having been confirmed to be major reservoirs for deep ocean heat uptake (Chen and Tung 2014; Gao et al. 2018).

An important question that follows is through what physical processes the deeper ocean warming below the 700-m depth is materialized. It was proposed that ocean circulation, especially the meridional overturning circulation, has played an important role in redistributing the imbalanced heat from the surface to various depths, as well as spreading heat in deep oceans (Lee et al. 2015; Buckley and Marshall 2016; Liu et al. 2016, 2018). As shown in Zhang (2010), the advection process associated with the deep western boundary current (DWBC) is effective to carry the warming/cooling signal from the subpolar sinking region southward along the DWBC route. However, as is known, the deep meridional overturning circulations in the Atlantic Ocean are western boundary-trapped (Lozier 2012), which contrasts with the observed basinwide warming in the deep Atlantic Ocean. On the other hand, others suggested

that vertical ventilation has played an important role in quickly spreading warming vertically from surface to deeper layers (Chen and Tung 2014). While the assumed fast ventilation can explain the basinwide warming that appeared in static empirical orthogonal function modes (which vary or change simultaneously at the whole spatial domain or in the snapshot), which only provides an ocean state at a selected temporal location, there exist other possible mechanisms that can lead the same features revealed by EOF or snapshot. Therefore, it remains unclear whether the ventilation process is effective enough to deposit heat to the deep ocean, especially in the case of increased vertical stratification caused by surface warming in recent decades (Zhang et al. 2013).

The purpose of this study is to infer the important physical processes involved in basinwide deep Atlantic Ocean warming from examining the evolution of deep-ocean temperature in reanalysis data to fill the missing piece in the physical mechanisms proposed in previous studies. By tracking the transport pathways of deep Atlantic Ocean heat anomaly spreading over the whole Atlantic basin on multidecadal or even longer time scales, a consistent physical explanation for basinwide deep Atlantic warming is proposed.

The paper is arranged as follows: Section 2 introduces the reanalysis data analyzed in this study and explains the adaptive and spatiotemporally local method that we use to extract the evolution of deep ocean temperature. Section 3 presents the detailed evolution of deep Atlantic Ocean temperature. Section 4 provides statistical justification for the diagnosed sequential warming along two pathways and their spreading. Section 5 explains new physical processes responsible for basinwide warming spreading in the deep Atlantic Ocean. With all these results and previously proposed physical mechanisms, we seek to synthesize a consistent physical diagram for how deep Atlantic Ocean heat uptake in the last seven decades has been materialized. Section 6 presents a summary and some discussions.

2. Data and methodology

a. Data

In this study, the quality-controlled Met Office observational potential temperature and salinity EN4.1.1 dataset with Gouretski and Reseghetti (2010) bias corrections is analyzed to elucidate the evolution of deep Atlantic Ocean warming (Good et al. 2013). The EN4 dataset can be downloaded from <http://www.metoffice.gov.uk/hadobs>. Compared with previous versions of EN datasets, EN4 includes significant improvements

FIG. 1. EEMD decomposition of the averaged north subtropical Atlantic Ocean temperature (1000–2000 m). (left) The original data (black) and successive remainders R_j after an additional EEMD component C_j is extracted. The red line is the ultimate EEMD trend R_n . (right) Each line represents an EEMD component C_j from high to low frequency.

in data quality control and provides uncertainty estimates. EN4 has assimilated a collection of ocean temperature and salinity profiles obtained across the global oceans over the period from January 1900 to present. The resolution of EN4 is 1° by 1° in latitude and longitude, with 42 vertical levels. The main observational data sources for EN4 came from the World Ocean Database (WOD, latest version WOD09), the Coriolis dataset for Reanalysis (CORA), the data compiled during the Arctic Synoptic Basin-wide Oceanography (ASBO) project, the data from the Global Temperature and Salinity Profile Program (GTSP), and the Argo data from the Argo Global Data Assembly Centers (GDACs) since 2000. In the locations where no observation was available, EN4 data relax to the climatology of the *World Ocean Atlas 1998*, an average over 1971–2000. Because the observations are sparse and the quality of the data is unguaranteed before 1950 (Good et al. 2013), our present analysis only applies to the temperature data after January 1950. The monthly Ishii dataset (Ishii and Kimoto 2009) was also analyzed simultaneously for cross-validation and comparison (<http://rda.ucar.edu/datasets/ds285.3>). The Ishii dataset

contains subsurface temperature and salinity at 24 levels in the upper 1500 m of oceans.

b. The multidimensional ensemble empirical mode decomposition method

The multidimensional ensemble empirical mode decomposition (MEEMD) method (Wu et al. 2009) is used to extract the multidecadal and longer time-scale evolution of ocean temperature in the original EN4 dataset. MEEMD is developed based on the empirical mode decomposition (EMD; Huang et al. 1998; Fig. S1 in the online supplementary material) and ensemble empirical mode decomposition (EEMD; Wu and Huang 2009; see also our online supplementary materials).

In MEEMD, individual temperature time series $x_s(t)$ from the EN4 dataset at any spatial grid point s is decomposed naturally using EEMD into a small number of amplitude–frequency-modulated oscillatory components $C_{j,s}$ ($j = 1, 2, \dots, n$) and a residue trend $R_{n,s}$ (Fig. 1), with the trend containing at most one extremum or being monotonic (Wu et al. 2007, 2011a) after all the identifiable oscillatory components are extracted, that is,

$$x_s(t) = \sum_{j=1}^n C_{j,s}(t) + R_{n,s}(t). \quad (1)$$

This time-varying trend $R_{n,s}$ obtained from EEMD decomposition requires no a priori function form (basis) and has low sensitivity to the addition of new data (Wu et al. 2007, 2011a). It is noted that EEMD is a sparse decomposition method; often, the total number of components n is less than the logarithms $\log_2 N - 1$, where N is the length of the series subjected to EEMD decomposition. For example, if the time series is 1000 data points, the EEMD decomposition results in less than 9 oscillatory components and a secular trend.

The MEEMD (Wu et al. 2009) takes the advantage of the high locality and robustness of EEMD. There are two types of MEEMD, with one for the decomposition of spatial data such as images and the other for temporal-spatial data such as gridded climate data. It is the latter type of MEEMD that we use in this study. After all the time series of a climate variable are decomposed into components, $C_{j,s}$ ($j = 1, 2, \dots, n$) and a residue trend $R_{n,s}$, at all spatial grids, as illustrated in Fig. 1, $C_{j,s}$ from all individual spatial locations are projected onto their original spatiotemporal locations to form the spatiotemporal evolution pattern of time scales corresponding to all $C_{j,s}$. Since the characteristic time scale of $C_{j,s}$ resulted from EEMD is not sensitive to small data perturbations and the temperature data difference between neighboring grids is usually small, all j th components from all spatial locations contain almost identical time scales, and different ranked components [e.g., j th and $(j + 1)$ th] at any spatial location are dramatically different. Thus, the combination process from EEMD component to MEEMD spatiotemporal evolving pattern is a natural selection.

It is noted that this decomposition inherits the temporal locality of EEMD (Wu et al. 2011a,b; Ji et al. 2014). In spatial domain, the whole decomposition process is independent; therefore, MEEMD is absolutely spatially local. The spatiotemporal locality of the method itself guarantees that any coherent spatiotemporal structures obtained using MEEMD are not a result of the method itself; rather, they are intrinsic to the data. This feature has been demonstrated by previous studies (Wu et al. 2009; Ji et al. 2014; Feng et al. 2014a,b).

Since MEEMD trends are time varying, the instantaneous warming rate at any given temporal location does not reflect well the warming amount from a reference time to that particular temporal location. In contrast, for linear trend, a known trend slope can uniquely determine the warming amount for any given temporal episode. To better reveal the spatiotemporal evolution of deep Atlantic Ocean warming using

MEEMD trends and to facilitate the comparison of the MEEMD trends with the corresponding linear ones, new quantities need to be defined from MEEMD trends. Here, we define a quantity called “accumulated warming” as the warming at a particular temporal location t given by the MEEMD trend plot from its reference value at 1950, that is, $\text{Trend}(t) - \text{Trend}(1950)$ (Ji et al. 2014). Such a definition allows us to clearly track the spatial spreads of any localized signals in subsequent times after 1950 and eliminate the effect of trend state at 1950 since the starting trend signal is zero everywhere. The temporal and spatial locality of MEEMD trends provides an improved measure to diagnose the underlying variation and change of physical information of data (Huang and Wu 2008; Wu et al. 2011a; Ji et al. 2014; Franzke 2014). More detailed information about EMD/EEMD/MEEMD can be found in the online supplemental materials. The corresponding MATLAB EEMD code, which we used in this study, can be downloaded from <http://rcada.ncu.edu.tw/research1.htm>.

c. The gridwise fitted linear trend

A widely used trend analysis method is the least squares straight line fitting to a time series or its subsections (e.g., Fig. 2.22 of Hartmann et al. 2013). To obtain the spatial coherence of such trends, the slopes or the starting and end point differences of linear trends at individual grids are displayed/pieced together. The linear fitting is carried out for the whole selected data temporal domain and its subsections when necessary for comparison with MEEMD trends.

3. The evolution of deep Atlantic Ocean warming: Two pathways

The spatiotemporal evolution of the accumulated warming averaged over depths ranging from 1000 to 2000 m extracted using MEEMD is displayed in Fig. 2 (and supplemental movie S1). This particular depth range was chosen for two reasons: 1) the shallower core of the DWBC (Lozier 2012) exists within this depth range, and 2) warming/cooling signals in layers within this depth range show consistent spatiotemporal evolution structure in EN4. The spatiotemporal evolutions of warming/cooling trends at different depths above 1000 m are not consistent with each other (Figs. S2 and S3), and coherent signatures of upper-hemispheric overturning cells (Zhang 2010; Lozier 2012) do not emerge from the temperature record on multidecadal or longer time scales.

One notable feature in Fig. 2 (movie S1) is that the deep Atlantic warming spreads along two distinct routes. In the early 1960s, accumulated warming greater

FIG. 2. Spatiotemporal evolution of the MEEMD trend of deep Atlantic Ocean temperature. (a)–(f) MEEMD trends (K) ending in 1960, 1970, 1980, 1990, 2000, and 2010, respectively. The gray box in (f) indicates the sinking region in the Labrador Sea (45°–60°N, 35°–50°W). The two pathways along the DWBC (black) and the equator (blue) will be further analyzed. (g) The spatial structure of temperature increase based on a time-unvarying linear trend over the entire data domain from 1950 to 2014.

than 0.05 K had already emerged in the Labrador Sea and the region of the North Atlantic immediately south of Greenland, while elsewhere in the Atlantic showed little sign of warming (less than 0.01 K). It is noted here that the selection of 0.05 K as a criterion of noticeable warming is only for convenience, and the spatiotemporal structure of the warming evolution is not sensitive to the criterion as long as it is within the range between 0.03 and 0.07 K. In the next two decades, this warming increased and expanded westward and then southward along a route matching the DWBC. By 1980, warming greater than 0.05 K had already reached the equatorial Atlantic. In the 1980s and 1990s, the warming continued to enhance, but its route split into two: one extending southward along the DWBC and the other eastward

along a narrow strip between 2°S and 2°N centered at the equator (Fig. S4). By the end of the twentieth century, the equatorial warming had reached the eastern boundary of the tropical Atlantic and showed signs of southward and northward expansion along the eastern Atlantic coast. By 2010, most regions of the deep Atlantic Ocean had warmed, while the subpolar deep North Atlantic had shown cooling greater than 0.05 K. It is noted that the final basinwide accumulated warming identified using MEEMD shares the similar structure of the corresponding one obtained using linear trend for the period 1950–2010 (Fig. 2g). However, the linear trends cannot provide useful information of evolution.

To quantify the spreading speed of the warming signal along the DWBC route and the equatorial route, we

FIG. 3. MEEMD trends of the deep Atlantic Ocean warming along two paths, (a) the DWBC and (b) the equator (lines in Fig. 2f). The orange lines represent the temperature anomaly (a) spatially averaged over 5° zonally and a vertical depth of 1000–2000 m at the labeled latitude along the DWBC, and (b) spatially averaged over 5° meridionally and a vertical depth of 1000–2000 m centered at the labeled longitude along the equator, with the thick red line being their corresponding MEEMD trends. Zero reference lines are shown as thin solid black lines for each plot. Dashed black lines represent the linear fits for the zero-crossing points of the MEEMD trends.

further examined the spatially averaged ($5^\circ \times 5^\circ$) temperature anomalies along these two routes and their corresponding MEEMD trends (Fig. 3), respectively. At northern latitudes (north of 35°N), multidecadal variability of temperature anomaly is large (Van Aken et al. 2011; Fig. 3a), leading to relative inaccuracy in trend estimation. This inaccuracy is rooted from neighboring EEMD components being not absolutely orthogonal between them and often having a correlation of 0.2, meaning that about a few percent of variance of the multidecadal variability may be contained in the trend. When the multidecadal component is large but the true trend is small, the EEMD-extracted trend has relatively large error. A similar problem also exists in linear fitting. With this in mind, the speed of the southward-extending warming signal along the DWBC is estimated using only the zero-crossing temporal locations of MEEMD trends for latitudes south of 35°N . This estimation gives a speed of 1.4 cm s^{-1} of the southward warming signal spreading from 30°N to 30°S . It is noted that this estimated speed is not sensitive to the selection of either the zero-crossing temporal locations or the temporal locations of any given threshold of warming smaller than 0.1 K . Our estimated warming spread speed falls within the range from 1 to 2 cm s^{-1} inferred from transient tracer ages at around

1800-m depth level (Richardson and Fratantoni 1999) to modeled temporally mean meridional speed (1.5 cm s^{-1}) of the DWBC averaged over the 1800–4000-m depth range (Zhang 2010), but smaller than the modeled tracer-based estimation of the DWBC speed of 5 cm s^{-1} in the whole subpolar mixing regime dominated by decadal or shorter time-scale variability (Waugh and Hall 2005).

Based on the temporal locations of MEEMD trend zero crossing (Fig. 3b, the dashed black line), the calculated eastward warming spread speed at the equator is 1.3 cm s^{-1} . This speed is approximately two orders of magnitude smaller than the Kelvin wave speed associated with the thermocline depth change (Sarachik and Cane 2010) and also smaller than the observed speed range from 3 to 20 cm s^{-1} of the equatorial deep jets (Richardson and Fratantoni 1999; Gouriou et al. 1999). This discrepancy needs to be explained. There exists previous modeling evidence of deep Atlantic Ocean signal propagation associated with the equatorial Kelvin wave (Kawase and Sarmiento 1986; Kawase 1987; Huang et al. 2000; Goodman 2001; Cessi et al. 2004; Marshall et al. 2015). With the small vertical stratification in the deep Atlantic Ocean, the speed of 1.3 cm s^{-1} is not out of the question. However, the equatorial Kelvin waveguide consistent with such a small Kelvin

FIG. 4. The instantaneous warming rates (K decade^{-1}) of EEMD temperature trends in (a) the Labrador Sea and (b) along the DWBC.

wave speed is extremely narrow (with a meridional characteristic scale less than a degree in latitude), much smaller than the meridional scale we diagnosed. In section 5, we will propose a skeleton Kelvin wave–horizontal diffusion theory to explain our diagnosed warming evolution structure.

We also observed an enlarging cooling area in the deep Atlantic Ocean north of the Gulf Stream extension region first in the 1960s and reduced warming along the DWBC north of 30°N in the 1990s and 2000s (Fig. 2). This feature implies a potential hidden multidecadal variability in the overall basinwide warming. To illustrate this feature more clearly, we analyzed the evolution of the warming rate (the temporal derivative of temperature anomaly emphasizing multidecadal variability in trend; Fig. 4) so as to uncover the multidecadal contribution to the MEEMD trends, both in the Labrador Sea from surface to 2000-m depth (Fig. 4a) and along the DWBC from 50°N to 30°S (Fig. 4b). In the Labrador Sea, the subsurface warming propagated downward to a depth of approximately 1500 m in about 15 years spanning the 1950s and 1960s (Fig. 4a), then extended southward along the DWBC, and reached 30°S in the 2000s (Fig. 4b). The interplay of warming and cooling episodes has a period of about 60 years. The surface cooling at the Labrador Sea from the 1950s and 1970s propagated downward and later southward, with similar propagating characteristics to those seen in the subsurface warming of the Labrador Sea that appeared in the 1950s and 1960s.

A related question is whether the warming/cooling of the deep Atlantic Ocean at the depth of 1000–2000 m

since 1950 is mostly associated with natural multidecadal variability or anthropogenically forced warming. The answer to this question is particularly important to understand whether the deep Atlantic Ocean warming has contributed to explain the missing heat at the surface that was related to the global warming hiatus in the last two decades. Since the analyzed global warming from 1850 to 1950 in both observations (Wu et al. 2011a) and anthropogenically forced warming in Earth system model simulations (Zhang et al. 2013) was only about 0.2 K at the surface and still small, it is arguable that such a structure reflects natural temperature variability on multidecadal time scales associated with the DWBC and its variability. The interplaying phases of warming and cooling with an approximately 60-yr period are consistent with model results (Frankcombe et al. 2010). This warming/cooling contrast near the two ends of the DWBC in the deep Atlantic Ocean coincided with the bipolar “seesaw” pattern at the surface (Chylek et al. 2010; Wu et al. 2011a).

While the DWBC pathway of warming/cooling can be justified by heat advection by the DWBC and was previously confirmed by many studies (e.g., Munoz et al. 2011; Williams et al. 2014; Häkkinen et al. 2015), the narrowness of the DWBC cannot explain the basinwide warming in the deep Atlantic Ocean. The equatorial pathway is a new finding, and further physical and statistical justifications need to be discussed. In the following two sections, we present our detailed arguments on why the above results are dependable both statistically and physically, assuming that EN4 data contain realistic information of deep Atlantic Ocean warming.

4. Statistical robustness of EEMD revealed spreading of warming

a. Linear fitting and its appropriateness

The revealed warming signal evolution pattern above cannot be obtained cleanly using traditional analysis methods, such as spatially pieced-together linear fitting of gridwise temperature time series (section 2c). The diagram of linearly fitted trends along the DWBC route and along the equatorial route is presented in Fig. 5 and the pieced-together trend evolution is in Fig. 6.

A visually identifiable difference shown in Figs. 3 and 5 is the mismatching degrees of sequentially delayed locations of the zero crossing in EEMD trends and linear trends, illustrated by the dashed lines that fit the zero-crossing points of the linear trends and of the EEMD trends along two routes. It is mathematically provable that the zero-crossing point of a zero-mean

FIG. 5. Linear trends of the deep Atlantic Ocean warming along two paths: along (a) the DWBC region and (b) the equator (Fig. 2f). The orange lines are the averaged temperature anomaly over (a) a zonally 5° vertically 1000–2000-m depth at the labeled latitude along the DWBC route, or over (b) a meridionally 5° vertically 1000–2000-m depth centered at the labeled longitude along the equator, and the thick red line is its corresponding linear trend. Thin black lines are reference zero lines and the thick dashed black line is the linear fit of the zero-crossing points of the linear trends.

linear trend of a stationary time series tends to stay within the immediate vicinity of the midpoint in the temporal domain. For a zero-mean nonstationary time series, the location constraint of the zero-crossing point associated with a linear trend (which does not necessarily have a zero mean) is not as stringent, but still remains largely true. This is the reason why the linear trends of sequentially delayed time series cannot characterize the sequential delay well. When multidecadal or longer time-scale sequential warming along continuous spatial locations is the focus, alternative methods that are able to capture sequential delay (i.e., having high locality) must be used. The EEMD trends indeed serve this purpose well. Along the DWBC, the zero-crossing fitting of the visually appealing (also mathematically rigorous) EEMD trend shows that it takes about 15 years for the warming/cooling signal of multidecadal or longer time scales to travel from 30°N to 30°S (corresponding to a speed of 1.4 cm s^{-1} if the curved coastal effect is not considered, and a larger speed if the curved coastal effect is taken into account). Along the equatorial path, it takes about 11 years for the similar signal to travel from 35°W to 5°E , corresponding to a speed of about 1.3 cm s^{-1} .

It is noted here that the nature of varying trends can be reflected, to some degree, by piecewise linear trends

(Fig. 6). The overall linear trends along a series of spatial grids cannot capture sequentially delayed warming defined as $\text{Trend}(t) - \text{Trend}(1950)$ along these grids as illustrated in Fig. 5b. To make the linear trends directly comparable to those from MEEMD trends (Fig. 2), we examine the linear trends of data from 1950 to a given temporal location t to obtain $\text{Trend}(t) - \text{Trend}(1950)$ but based on linear fitting (Fig. 6). Clearly, the sequential warming obtained from linear fitting along the DWBC and the equatorial region is not as clean as those from MEEMD trends although qualitatively similar over the long terms, with warming patches in isolated regions especially when the linear fitting ends at earlier decades. Because of the discontinuity of piecewise trends at neighboring temporal spans selected, a systematically continuing trend without containing variability of temporal scale longer than the selected temporal window for the calculation of piecewise trends is not obtainable. Any variability of time scales longer than the selected temporal window is contained in the piecewise trends.

It is also noted that a linear trend cannot depict the time-varying nature of the trend along temporal domain, if there is any. There is neither a physical reason that the trend should definitely be linear, especially under the situations that warming could be possibly

FIG. 6. Piecewise linear trends of deep Atlantic Ocean temperature starting at 1950. (a)–(f) Linear trends (K) ending in 1960, 1970, 1980, 1990, 2000, and 2010, respectively.

accelerating. As argued and demonstrated in previous studies (Huang and Wu 2008; Wu et al. 2011a,b; Ji et al. 2014; Franzke 2014), the EEMD trend can capture time-varying characteristics well and is therefore a more natural choice for tracking sequential changes.

b. Robustness of zero-crossing point determined by an EEMD trend

The temperature anomaly data displayed in Figs. 3 and 5 show a common characteristic of red noise riding on a time-varying trend, with riding red noise having increasing amplitude. Since the noise is spatiotemporally different both in amplitude and timing, it is helpful to understand the sensitivity of the determined trend zero-crossing point to different levels of noise and to the different realizations of noise. To achieve this goal, we compose synthetic time series containing a trend of multidecadal and longer time-scale variability and change and a red noise series that has temporally increasing amplitude. The reason for us to use synthetic time series is that the exact temporal location of its zero-crossing point is known, so that the loyalty of the EEMD trend to the prescribed trend can be quantified.

The synthetic trend is prescribed as

$$\text{Trend}_{\text{syn}}(t) = \frac{1}{10} \exp \left[\frac{1}{15} (t - 1995) - 1 \right] + 0.02 \cos[2\pi(t - 1930)/65], \quad (2)$$

and the noise has a form of

$$\text{RN}_{\text{syn}}(t) = a[0.1 + 0.0138(t - 1950)]N(t), \quad (3)$$

where $N(t)$ is the monthly red noise of a unit standard deviation. For simplicity, this monthly red noise time series is obtained by taking a 13-month running mean of a white noise series. The term a is selected to make RN_{syn} have a desired standard deviation.

The red noise characteristic specified in the synthetic data is quite similar to those displayed in Figs. 3 and 5. Various examples of the combinations of the trend (with a standard deviation of 0.1°C) and noise of different amplitudes (with a standard deviation of 0.02° , 0.04° , and 0.08°C) are plotted in Fig. 7. As shown in Figs. 3 and 7, when RN_{syn} has a standard deviation value about 0.04°C , the synthetic time series appear to be similar to the deep ocean temperature anomalies in the eastern equatorial Atlantic Ocean.

FIG. 7. EEMD trend and its sensitivity to the level of noise. (a) The thin red line is the shape of noise and the thick red line is the specified noise. The brown lines are the synthetic time series summing up the specified trend (red thick lines) and the noise of different levels of amplitude (with standard deviations of 0.02° , 0.04° , and 0.08°C). The thick black lines are the corresponding EEMD trends. The thin black lines are the EEMD trends for noise standard deviations from 0.01° to 0.1°C , with an increment of 0.01°C in noise level. (b) The standard deviation of the error (the difference between the EEMD trend and the specified trend) as a function of the noise level. (c) The difference of the zero-crossing points of the EEMD trend and of the specified trend as a function of noise level, with a negative value meaning the zero-crossing point of the EEMD trend coming earlier.

Figure 7 also presents various calculations applied to synthetic time series. The results show that the trend error (defined as the standard deviation of an EEMD trend and the specified trend) is almost linearly proportional to the noise level, and so is the error of the determined zero-crossing point of the EEMD trend. In general, when the noise level of the synthetic data is comparable to that of the data, the zero-crossing point of the EEMD trend falls within the 10-month error range and its difference from the specified trend is quite small, which confirms that the EEMD trend loyally captures the true trend and the increasing fluctuation in data bears little effect on the zero-crossing points used for the later determination of the spreading of the warming signal.

Another issue hidden in Fig. 7 is the small end effect. From a pure data analysis perspective, we do not have the luxury of knowing the exact trend when a time series is given, just like what is specified in Eq. (2). In a more

philosophical perspective, we even do not know what is noise, since one researcher's noise may be another's signal. In this sense, what can be really quantified is the trend locality. In Fig. 7, the degree of locality can be visually seen. To a significant degree, EEMD is an adaptive natural filter and the EEMD trend unavoidably contains part of multidecadal or longer time-scale variability of noise but is quite local. It was previously demonstrated numerous times that EEMD has higher locality than other methods, such as the Fourier transform and wavelets (Wu and Huang 2009; Wu et al. 2011a,b), given that the stoppage criterion for sifting is local but not global as described in Wu and Huang (2009).

c. Uncertainties of zero-crossing point determined by an EEMD trend

The EEMD trend can vary if a different realization of noise [Eq. (3) with changed white noise] is

FIG. 8. EEMD trend and its sensitivity to different realizations of noise. (a) The thin black lines are the EEMD trends corresponding to different realizations of the same type of noise with a standard deviation of 0.04. The red line is the specified trend, the thick white line is the mean of all the black lines, and the two thin yellow lines define the one standard deviation spread of all the black lines. (b) The standard deviation of the error (the difference between the EEMD trend and the specified trend) for different realizations of noise. (c) The difference of the zero-crossing points of the EEMD trend and of the specified trend for different realizations of noise, with a negative value meaning the zero-crossing point of the EEMD trend coming earlier. The standard deviation of the histogram is 10 months.

considered a part of data. To quantify this effect, we further test the sensitivity to different realizations of noise series. The result is plotted in Fig. 8. It is shown that the ensemble mean of trends from cases of different red noise realizations almost overlap the specified trend everywhere except the two data ends. In addition, one standard deviation (0.012°C) is very small compared to the trend itself (about 0.4°C increasing from 1950 to 2015). A particular important quantity is the error of the trend zero-crossing point, which has a standard deviation of about 10 months (Fig. 8c).

d. Statistical significance of the sequential spreading of warming trend

With the expositions in sections 4b and 4c, we are now ready to discuss the statistical significance of the warming signal sequential spreading. The null hypothesis in this case is that “the warming signal spreading revealed using MEEMD, as illustrated in Fig. 3, is caused by red noise contained in the data.”

Here we use two methods to reject this null hypothesis based on the characteristics of EEMD-identified trend zero-crossing points. From Fig. 3, the EEMD trend zero-crossing points along both the DWBC and the equator are steadily delaying. For the DWBC signal path, from 30°N to 30°S , seven zero-crossing points (corresponding to different latitudes) should fluctuate with respect to their mean zero-crossing temporal locations if the null hypothesis is true. If the temporal locations of these zero-crossing points are ordered temporally from small to large, there are 5040 (the factorial of 7) different orders (e.g., order 1: 0° , 30°N , 20°S , 10°S , 10°N , 20°N , 30°S ; order 2: 10°N , 30°N , 10°S , 20°S , 20°N , 0° , 30°S), with only one order going from 30°N to 30°S , meaning that the chance of the steady delay can only have a probability of 1 in 5040. Similarly, with 10 selected locations along the equator, the steady delay in warming would occur only with a chance of 1 in 3 628 800 (factorial of 10). Clearly, the chance of a steady delay of EEMD-identified trend zero-crossing points along the DWBC and along the equator is orders smaller than 1%. The null hypothesis

above is firmly rejected and the steady delaying diagnosed earlier can almost surely not be a case of being caused by randomness. It is also noted that this test is independent of the property of noise.

In the above test, we even did not consider the narrowness of zero-crossing spread caused by noise amplitude and noise realization. As we showed in Figs. 7 and 8, the standard deviation of an EEMD zero-crossing point caused by the noise amplitude is about 10 months. Considering that the zero-crossing points of the first and the last EEMD trends along the DWBC are separated by 180 months (15 years), meaning at least 18 standard deviations away from each other, the chance for that to be caused by randomness should be infinitesimally small. Along the equator, the zero-crossing points of the first and the last EEMD trends are separated by 132 months (13 standard deviations); again, the chance of that being caused by randomness is infinitesimally small.

In either way of testing, the steady sequential delays of the warming trend along the DWBC and the equator are at least significant at the 99.9% level.

5. The role of meridional diffusion in spreading the warming signal

Our results suggest that the DWBC has played a crucial role in spreading warming along the western coast region of the Atlantic Ocean through advection, consistent with the previous findings (Zhang 2010). However, the specific route of eastward expansion of the equatorially trapped warming has not been previously revealed, although short-period observations of ocean currents have shown the existence of an equatorial deep jet with similar meridional width (Gouriou et al. 2001).

The meridional structure of the warming along this route quite closely follows Gaussian curves centered at the equator (Fig. S4), and the warming signals propagate eastward, implying that the deep-ocean Kelvin waves are a likely candidate to spread warming to regions away from the western coastal region of the Atlantic Ocean. The modeling evidence of such a Kelvin wave signal in the deep Atlantic Ocean was seen in previous studies (Kawase and Sarmiento, 1986; Kawase 1987; Huang et al. 2000; Goodman 2001; Cessi et al. 2004; Marshall et al. 2015). However, the multidecadal time scales and low estimated propagation speed (1.3 cm s^{-1}) associated with this route of warming expansion appear to contradict the present belief that the oceanic variability involving equatorial Kelvin waves is often on interannual time scales or shorter (Brandt et al. 2011).

This contradiction is reconcilable when the role of meridional diffusion is taken into account. If this

propagation along the equator is associated with an equatorial Kelvin wave, the shape of the propagating warming signal should have a Gaussian shape centered at the equator, that is,

$$T \propto \exp\left(-\frac{y^2}{L^2}\right), \quad (4)$$

where y is the meridional distance from the equator and $L = \sqrt{c/\beta}$, with β being the meridional changing rate of the Coriolis parameter. With the estimated speed of 1.3 cm s^{-1} , the corresponding equatorial waveguide of latitudinal Gaussian width L should be about 0.2° , a meridional scale implying that the propagation of a multidecadal or longer time-scale warming signal can only show up along the grids at the equator in EN4 with a horizontal resolution of 1° by 1° in latitude and longitude. At first glance, the calculated width of 0.2° is inconsistent with the much wider meridional scale of warming in the equatorial zone (Fig. 2). However, because the smoothing scheme in EN4 has a spatial covariance scale of 400 km (Good et al. 2013), this equatorial waveguide is widened. A careful examination of the enlarged panels of Fig. 2 for the tropical regions (Fig. S4) leads to two findings: 1) the front portion of the eastward-propagating warming trend in the equatorial zone diagnosed from the EN4 is only about 1° – 2° wider on either side of the equator. With the error smoothing technique that has a decorrelation scale of 400 km adopted in the quality control of EN4 (Good et al. 2013), the propagating signal is likely to be artificially spread horizontally a few degrees by the quality control technique. This inferred feature is certainly consistent with our diagnoses; and 2) the back side portion of the eastward-propagating signal is significantly wider than the front portion, and the signal has been widening on both sides of the equator at longitudes east of 35°W .

The latter feature is indeed consistent with the solution to an inhomogeneous heat transport model that includes a very narrow equatorial strip of propagating warming signal and horizontal diffusion, that is,

$$\frac{\partial T}{\partial t} - \alpha \nabla^2 T = \exp\left(-\frac{y^2}{L^2}\right) f(x - ct), \quad (5)$$

where $f(x - ct)$ represents a zonally propagating Kelvin wave signal of a reduced gravity shallow-water model (Sarachik and Cane 2010), and α is the diffusion coefficient. When the propagation speed is significantly larger than the diffusion speed, the meridional shape of the temperature anomaly at a given longitude can be approximately described by a one-dimensional diffusion equation simplified from Eq. (5):

$$\frac{\partial T}{\partial t} - \alpha \frac{\partial^2 T}{\partial y^2} = \exp\left(-\frac{y^2}{L^2}\right). \quad (6)$$

For Eq. (6), the solution is

$$T(y, t') = \int_0^{t'} \int_{-\infty}^{\infty} \frac{1}{\sqrt{4\pi\alpha(t' - \tau)}} \exp\left[-\frac{(y - \eta)^2}{4\alpha(t' - \tau)}\right] \times \exp\left(-\frac{\eta^2}{L^2}\right) d\eta d\tau, \quad (7)$$

where t' is the local time starting at the arrival of the very first front of the signal $f(x - ct)$. The Green's function form of the solution (7) basically depicts that the meridional shape of the temperature perturbation should be close to a Gaussian shape centered at the equator with a characteristic meridional scale $\sqrt{4\alpha(t' - \tau)}$ when t' is sufficiently larger and L is small. It is also clear that the characteristic of meridional scale increases with time. For a near steady eastward-propagating signal at the equator, this solution essentially indicates that the horizontal signal should have a feature of wider meridional spread in the west and narrower meridional spread in the east, exactly as the evolution feature of the temperature anomaly shown in Fig. 2 and movie S1 illustrate.

6. Summary and discussion

In this study, we elucidated the deep Atlantic Ocean warming since 1950. The role of the DWBC in carrying surface unbalanced heating to the deep ocean was previously proposed. However, because of the narrowness of the DWBC, the deep Atlantic Ocean basinwide warming in the last six decades can hardly be explained. By using the MEEMD method to extract spatiotemporal evolution of multidecadal or longer time scales in the ocean reanalysis dataset EN4, we observed that the equatorial pathway of warming spreading has played a significant role. The likely candidate to materialize the equatorial pathway is the slow Kelvin wave. With the assistance of horizontal diffusion, the thermohaline circulation, and the equatorial Kelvin wave, the surface unbalanced warming is spread to whole deep Atlantic Ocean at depths of 1000–2000 m.

It is revealed that the current widespread warming/cooling in the deep Atlantic Ocean can largely be traced back to the subsurface warming/cooling in the Labrador Sea about 60 years earlier. The identical analysis has also been applied to Ishii data, and the results are qualitatively similar to those from EN4 in the layers between 1000 and 2000 m, although Ishii data only cover the top 1500 m of oceans (Figs. S5–S7).

We also confirmed that such an evolution pattern of the warming/cooling signal along the DWBC and the equator is extremely unlikely to be a result of randomness, with a chance significantly less than 0.1%. Further examination of Fig. 4b shows multidecadal variability of temperature anomaly overlapping a warming trend for all the latitudes from 50°N to 30°S along the DWBC. This result indicates that the present deep Atlantic Ocean warming is a result of both natural multidecadal temperature variability and the continuing anthropogenically forced warming, which is consistent with results diagnosed from surface data (Wu et al. 2011a; Tung and Zhou 2013).

The results that the current basinwide Atlantic Ocean warming at depths of 1000–2000 m can be traced back to the subsurface warming in the Labrador Sea in the 1950s implies that the deep Atlantic Ocean is unlikely to partly account for the atmospheric radiative imbalance during the last two decades and to serve as an explanation for the current warming hiatus. However, the quasi-60-yr cycle displayed in deep Atlantic Ocean warming and its warming/cooling contrast between the northern subpolar subsidence region of the DWBC and southern circumpolar current region coincides with the key bipolar “seesaw” warming/cooling signature at the surface (Chylek et al. 2010; Wu et al. 2011a). The latter results suggested that deep Atlantic Ocean can serve as a heat source/sink with its multidecadal memory modulating the global climate variability and change on multidecadal or longer time scales, although not as a simultaneous heat reservoir.

It is noted that our results are from analyzing ocean reanalysis datasets. Because of the lack of observations of interior ocean temperature assimilated to reanalysis datasets, especially in deep oceans where observations have been extremely sparse before the 1970s, some caution needs to be taken with our results. However, the physical processes revealed in this study should be common in climate system models that have similar thermohaline circulation and stratification in deep oceans to the observed.

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