The Role of an Indian Ocean Heating Dipole in the ENSO Teleconnection to the North Atlantic European Region in Early Winter during the Twentieth Century in Reanalysis and CMIP5 Simulations

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(Manuscript received 23 April 2020, in final form 2 November 2020)

ABSTRACT: In this study the role of an Indian Ocean heating dipole anomaly in the transition of the North Atlantic–European (NAE) circulation response to El Niño–Southern Oscillation (ENSO) from early to late winter is analyzed using a twentieth-century reanalysis and simulations from phase 5 of the Coupled Model Intercomparison Project (CMIP5). It is shown that in early winter a warm (cold) ENSO event is connected through an atmospheric bridge with positive (negative) rainfall anomalies in the western Indian Ocean and negative (positive) anomalies in the eastern Indian Ocean. The early winter heating dipole, forced by a warm (cold) ENSO event, can set up a wave train emanating from the subtropical South Asian jet region that reaches the North Atlantic and leads to a response that spatially projects onto the positive (negative) phase of the North Atlantic Oscillation. The Indian Ocean heating dipole is partly forced as an atmospheric teleconnection by ENSO, but can also exist independently and is not strongly related to local Indian Ocean sea surface temperature (SST) forcing. The Indian Ocean heating dipole response to ENSO is much weaker in late winter (i.e., February and March) and not able to force significant signals in the North Atlantic region. CMIP5 multimodel ensemble reproduces the early winter Indian Ocean heating dipole response to ENSO and its transition in the North Atlantic region to some extent, but with weaker amplitude. Generally, models that have a strong early winter ENSO response in the subtropical South Asian jet region along with tropical Indian Ocean heating dipole also reproduce the North Atlantic response.

KEYWORDS: Atmosphere–ocean interaction; Atmospheric circulation; El Niño; Model evaluation/performance; North Atlantic Oscillation; Tropical variability

1. Introduction

The El Niño–Southern Oscillation (ENSO) teleconnection to North Atlantic–European (NAE) region is an active field of current research (e.g., van Loon and Madden 1981; Kiladis and Diaz 1989; Halpert and Ropelewski 1992; Rodó et al. 1997; Moron and Ward 1998; Van Oldenborgh et al. 2000; Brönnimann 2007; Lorenzo et al. 2011; Herceg-Bulić et al. 2012, 2017). This teleconnection is much weaker than the ENSO teleconnection to the Pacific–North American (PNA) region, but is nevertheless the major source for seasonal predictability in the Euro-Atlantic region (e.g., Domeisen et al. 2015; Scaife et al. 2017). Several pathways have been proposed for ENSO teleconnections to the North Atlantic region, including the canonical tropospheric pathway including the tropical Northern Hemisphere (TNH) pattern (e.g., Wallace

and Gutzler 1981; Jiménez-Esteve and Domeisen 2018; and many others) and a stratospheric one (Ineson and Scaife 2009; Butler et al. 2014; Domeisen et al. 2019). Recent studies have reported that in models and reanalysis there are significant nonlinearities and asymmetries between the ENSO response in the North Pacific (Frauen et al. 2014; Garfinkel et al. 2019; Jiménez-Esteve and Domeisen 2019), Arctic stratosphere (Rao and Ren 2016; Trascasa-Castro et al. 2019; Weinberger et al. 2019), and North Atlantic (Hardiman et al. 2019; Jiménez-Esteve and Domeisen 2020; Trascasa-Castro et al. 2019; Weinberger et al. 2019). ENSO teleconnections are initiated in the tropical central-eastern equatorial Pacific by a positive heating anomaly induced by a warm ENSO (El Niño) event. Due to an approximate equilibrium in the thermodynamic equation between the adiabatic cooling and the diabatic heating term, the heating is compensated mainly by rising motion, which leads to upper-level divergence (e.g., Trenberth et al. 1998; Hamouda and Kucharski 2019), while almost opposite conditions are observed in case of a cold ENSO (La Niña) event. The upper-level divergence may act as a Rossby wave source (Sardeshmukh and Hoskins 1988) and induce an extratropical wave train, which propagates northeastward from the source region toward the PNA region, eventually reaching the Atlantic region during boreal winter. Transient eddy feedbacks are also essential to zonally extend the anomalies to the North Atlantic (e.g., Li and Lau 2012a,b; Jiménez-Esteve and Domeisen 2018).

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

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-20-0269.s1>.

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DOI: 10.1175/JCLI-D-20-0269.1

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Also, the Walker circulation may be modified by an El Niño event and can thus induce sinking motion and therefore reduced rainfall in remote tropical regions (Cai et al. 2019). This compensating sinking motion to a warm ENSO event in the tropical Atlantic region can induce a Gill-type response there, which could initiate further processes (García-Serrano et al. 2017). ENSO also induces some changes in the tropical Indian Ocean (Zhong et al. 2005; Roxy et al. 2011), and that may affect its connections to the NAE circulation, which is discussed in the current study by using the Twentieth Century Reanalysis and the simulations from phase 5 of the Coupled Model Intercomparison Project (CMIP5; Taylor et al. 2012).

It has been noted that the ENSO teleconnection to the NAE region shows substantial intraseasonal variations from November to March (King et al. 2018a; Ayarzagüena et al. 2018). In early winter the response to a warm ENSO event is a negative geopotential height anomaly in the northern North Atlantic that spatially projects onto the positive phase of the North Atlantic Oscillation (NAO), whereas in late winter (from January onward) a dipole response spatially projecting onto the negative phase of the NAO is found (Bladé et al. 2008; King et al. 2018b). Therefore, considering the winter season as December to February leads to a weakening of already weak signals (e.g., Mezzina et al. 2020). It has been noted that uninitialized model simulations do not reproduce well the observed transition of the ENSO response in the NAE region (Bladé et al. 2008; Ayarzagüena et al. 2018), whereas initialized seasonal forecast do reproduce the transition (Ayarzagüena et al. 2018; Abid et al. 2021).

Several mechanisms have been proposed for the early to late winter ENSO teleconnection transition to the NAE region. Ayarzagüena et al. (2018) investigated the role of Gulf of Mexico and Caribbean Sea rainfall responses to ENSO in the transition, whereas Abid et al. (2021, hereinafter AKM2021), examined the role of the Indian Ocean in the transition. AKM2021 demonstrated that a positive ENSO event induces a rainfall (heating) dipole in the Indian Ocean in early winter (December). This heating dipole generates an upper-level negative geopotential height response located close to the subtropical South Asian jet (SAJET), which initiates a wave train that crosses the North Pacific Ocean and eventually reaches the NAE region. The weakening of the Indian Ocean heating dipole response to ENSO in late winter is related to a southward shift of the Indian Ocean rainfall climatology (AKM2021). Thus, the seasonal variation of the Indian Ocean heating dipole response to ENSO is a good candidate for the ENSO response transition in the NAE region.

Several studies have demonstrated that the tropical Indian Ocean has a strong impact on extratropical and in particular on NAE climate anomalies from intraseasonal to multidecadal time scales (e.g., Bader and Latif 2005; Cassou 2008; Molteni et al. 2015; Yu and Lin 2016; Lee et al. 2019). Some studies also showed that the Indian Ocean warming (Zhong et al. 2005) may counteract the direct boreal winter ENSO-induced northern annular mode (NAM) and PNA response (e.g., Annamalai et al. 2007; Fletcher and Kushner 2011; Fletcher and Cassou 2015). ENSO teleconnections in CMIP5 simulations (Taylor et al. 2012) have been investigated in many

previous studies (e.g., Weare 2013; Langenbrunner and Neelin 2013; Ayarzagüena et al. 2018). However, a detailed analysis of the CMIP5 simulations regarding the role of the tropical Indian Ocean in the early to late winter ENSO response transition in the NAE circulation anomalies is still missing.

The aim of this study is to further investigate the ENSO response transition in the NAE region and in particular the role of the ENSO forced Indian Ocean heating anomalies using a long twentieth-century reanalysis dataset and also to analyze to what extent uninitialized coupled general circulation model (CGCM) simulations reproduce the observed transition. For this purpose, the models that participated in CMIP5 database are used.

The paper is organized as follows: The datasets used in the present study are discussed in section 2. Section 3 elucidates the method of analysis. Results investigating the role of an Indian Ocean heating dipole in the early-to-late winter ENSO response transition in the NAE region in reanalysis and CMIP5 models are discussed in section 4. Section 5 presents the summary and conclusions.

2. Datasets

To investigate the observed ENSO teleconnections to the NAE region, the monthly means of geopotential height (200 hPa), horizontal winds (200 hPa), pressure vertical velocity (500 hPa), and sea level pressure (SLP) at $2^\circ \times 2^\circ$ resolution are obtained from the National Oceanic and Atmospheric Administration (NOAA) Cooperative Institute for Research in Environmental Sciences (CIRES) Twentieth Century Reanalysis (20CR) V2c (Compo et al. 2011) along with the monthly sea surface temperature (SST; resolution $2^\circ \times 2^\circ$) from the NOAA Extended Reconstructed SST, version 5 (ERSST.v5; Huang et al. 2017).

To examine the fidelity of state-of-the-art climate models in reproducing the observed ENSO teleconnections, the historical simulations of 28 models that participated in CMIP5 database are used (for details regarding modeling groups and resolution see Table 1). These historical twentieth-century simulations are forced with the observed atmospheric composition changes that includes both natural (e.g., aerosols, volcanic impacts, solar forcing, and emanations of short-lived species and their precursors) and anthropogenic components (i.e., greenhouse gases and anthropogenic aerosols) as well as the time-varying estimates of land cover (Taylor et al. 2012).

Following previous studies (Joshi and Kucharski 2017; Kucharski and Joshi 2017; Joshi and Ha 2019), for fair comparison as well as for an equal weightage, the first ensemble member (i.e., r1i1p1) of each CMIP5 model has been used in this analysis; however, some models provide several realizations. The model outputs are freely available at the website <http://esgf-index1.ceda.ac.uk>, which is maintained by Earth System Grid Federation (ESGF). In this analysis, observational and model data are used for the common period 1901–2004. It is to be noted that the results presented in this study using the NOAA-CIRES 20CR dataset are quite similar to the ones obtained using the ECMWF Twentieth Century Reanalysis (ERA-20C; Poli et al. 2016) for the analysis period 1901–2004 and National Centers for Environmental Prediction

TABLE 1. List of CMIP5 models along with their modeling groups and resolution.

Model	Institution	Resolution (lat × lon)
BCC-CSM1.1	Beijing Climate Center, China Meteorological Administration, China	64 × 128
BCC-CSM1.1-m	Beijing Climate Center, China Meteorological Administration, China	160 × 320
BNU-ESM	College of Global Change and Earth System Science, Beijing Normal University, China	64 × 128
CMCC-CESM	Centro Euro-Mediterraneo per I Cambiamenti Climatici, Italy	48 × 96
CMCC-CMS	Centro Euro-Mediterraneo per I Cambiamenti Climatici, Italy	96 × 192
CanESM2	Canadian Centre for Climate Modeling and Analysis, Canada	64 × 128
GFDL CM3	Geophysical Fluid Dynamics Laboratory, United States	90 × 144
GFDL-ESM2G	Geophysical Fluid Dynamics Laboratory, United States	90 × 144
GFDL-ESM2M	Geophysical Fluid Dynamics Laboratory, United States	90 × 144
GISS-E2-H	NASA Goddard Institute for Space Studies, United States	90 × 144
GISS-E2-R	NASA Goddard Institute for Space Studies, United States	90 × 144
HadCM3	Met Office Hadley Centre, United Kingdom	73 × 96
HadGEM2-AO	National Institute of Meteorological Research/Korea Meteorological Administration, South Korea	145 × 192
HadGEM2-ES	Met Office Hadley Centre, United Kingdom	145 × 192
INM-CM4	Institute for Numerical Mathematics, Russia	120 × 180
IPSL-CM5A-LR	Institut Pierre-Simon Laplace, France	96 × 96
IPSL-CM5A-MR	Institut Pierre-Simon Laplace, France	143 × 144
IPSL-CM5B-LR	Institut Pierre-Simon Laplace, France	96 × 96
MIROC5	Atmosphere and Ocean Research Institute (The University of Tokyo), National Institute for Environmental Studies, and Japan Agency for Marine-Earth Science and Technology, Japan	128 × 256
MIROC-ESM	Japan Agency for Marine-Earth Science and Technology, Atmosphere and Ocean Research Institute (The University of Tokyo), and National Institute for Environmental Studies, Japan	64 × 128
MIROC-ESM-CHEM	Japan Agency for Marine-Earth Science and Technology, Atmosphere and Ocean Research Institute (The University of Tokyo), and National Institute for Environmental Studies, Japan	64 × 128
MPI-ESM-LR	Max Planck Institute for Meteorology (MPI-M), Germany	96 × 192
MPI-ESM-MR	Max Planck Institute for Meteorology (MPI-M), Germany	96 × 192
MPI-ESM-P	Max Planck Institute for Meteorology (MPI-M), Germany	96 × 192
MRI-CGCM3	Meteorological Research Institute, Japan	160 × 320
MRI-ESM1	Meteorological Research Institute, Japan	160 × 320
NorESM1-M	Norwegian Climate Centre, Norway	96 × 144
NorESM1-ME	Norwegian Climate Centre, Norway	96 × 144

(NCEP)–National Center for Atmospheric Research (NCAR) reanalysis V1 (Kalnay et al. 1996) dataset for the shorter period 1948–2004 (figures not shown).

Since the resolution of the reanalysis and each model dataset differs, for ease of comparison the datasets are interpolated into a common latitude–longitude grid ($2.5^\circ \times 2.5^\circ$) using bilinear interpolation.

3. Methodology

The analysis is mainly based on regression maps, obtained by linearly regressing the field of interest onto the normalized index (e.g., Molteni et al. 2015). Thus, the units of all regression maps are the same as that of the field used for regression. The results presented in this study using

regression analysis are quite similar to those obtained using composite analysis (figures not shown), revealing that the results are robust, irrespective of the methodology adopted. Before doing regression analysis, all fields from reanalysis and models are linearly detrended so that the trends in the data do not influence the results; however, not detrending the data also leads to qualitatively comparable results.

a. Statistical significance test

In the case of reanalysis and CMIP5 models, the statistical significance of regression coefficients is assessed via a two-tailed Student's *t* test, whereas to test the statistical significance of the multimodel ensemble (MME) regressions a Student's one-sample *t* test is applied (Joshi et al. 2020).

b. Indices used

In this analysis, various indices have been used: 1) the Niño-3.4 index, defined as area-averaged SST anomalies (SSTAs) over 5°S–5°N, 170°–120°W; 2) the Niño-3 index, defined as area-averaged SSTAs over 5°S–5°N, 150°–90°W; and 3) the ENSO Modoki index (EMI), defined by following equation:

$$\text{EMI} = [\text{SSTA}]_A - 0.5 \times [\text{SSTA}]_B - 0.5 \times [\text{SSTA}]_C, \quad (1)$$

where the brackets in Eq. (1) represent the area-averaged SSTAs over each of regions A (10°S–10°N, 165°E–140°W), B (15°S–5°N, 110°–70°W), and C (10°S–20°N, 125°–145°E), respectively.

The tropical western–eastern Indian Ocean (TWEIO) index, defined as the difference of area-averaged vertical pressure velocity (omega) at 500 hPa over the eastern (10°S–10°N, 100°–140°E; area enclosed within the blue box in Fig. 2a) and western (10°S–10°N, 40°–80°E; area enclosed within the red box in Fig. 2a) Indian Ocean is also used (Abid et al. 2020). It is to be noted that the results of this study are not sensitive to slight variations in the eastern and western Indian Ocean domains used for creating the TWEIO index.

The omega field is used as a proxy for rainfall, since precipitation in reanalysis is considered less reliable. The omega-based TWEIO dipole index is highly correlated with rainfall index based on Global Precipitation Climatology Project (GPCP) observations (Adler et al. 2003) from 1980 to 2014 (correlation coefficient = 0.94). Because of a strong covariability of the early winter TWEIO index with ENSO (correlation coefficient = 0.54), the Niño-3.4 SSTs based index is linearly removed from the TWEIO index to assess the “pure” TWEIO impacts (e.g., Kucharski et al. 2008, 2009). All regressions are calculated without taking time lags into account, assuming that the responses are nearly instantaneous on monthly time scales. As a sensitivity test, we have confirmed that using a one-month lag between the index and the field gives very similar results.

c. Wave activity flux analysis

To analyze the source and direction of energy propagation, wave activity flux formulated by Takaya and Nakamura (2001) is applied. The wave activity flux is parallel to the local group velocity corresponding to the stationary Rossby waves and is independent of the wave phase (Plumb 1985; Takaya and Nakamura 1997, 2001). It is defined as follows:

$$W = \frac{p \cos \phi}{2|U|} \begin{pmatrix} \frac{U}{a^2 \cos^2 \phi} \left[\left(\frac{\partial \psi'}{\partial \lambda} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \lambda^2} \right] + \frac{V}{a^2 \cos \phi} \left(\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial \phi} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial \phi} \right) \\ \frac{U}{a^2 \cos \phi} \left(\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial \phi} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial \phi} \right) + \frac{V}{a^2} \left[\left(\frac{\partial \psi'}{\partial \phi} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \phi^2} \right] \\ \frac{f_0^2}{N^2} \left[\frac{U}{a \cos \phi} \left(\frac{\partial \psi'}{\partial \lambda} \frac{\partial \psi'}{\partial z} - \psi' \frac{\partial^2 \psi'}{\partial \lambda \partial z} \right) + \frac{V}{a} \left(\frac{\partial \psi'}{\partial \phi} \frac{\partial \psi'}{\partial z} - \psi' \frac{\partial^2 \psi'}{\partial \phi \partial z} \right) \right] \end{pmatrix}, \quad (2)$$

where p is the pressure normalized by 1000 hPa, f_0 is the Coriolis parameter at 45°N, a is Earth's radius, N is the Brunt–Väisälä frequency, ϕ is the latitude, and λ is the longitude. The geostrophic streamfunction (ψ) is defined as φ/f , where φ is the geopotential and $f (=2\Omega \sin \phi)$ is the Coriolis parameter with Earth's rotation rate (Ω). Also, U and V represents the basic states (i.e., the monthly climatological mean states), while ψ' denotes the perturbed streamfunction. In the present study, this perturbed streamfunction is the geopotential anomalies regressed onto the standardized “pure” TWEIO index (obtained by regressing TWEIO index onto the Niño-3.4 index).

4. Results

a. Evolution of ENSO responses

Figures 1a–e show the regressions of the observed Niño-3.4 index onto 200-hPa geopotential height anomalies from NOAA-CIRES 20CR for the months November to March. In early winter (November and December) there is a clear indication of projection onto a positive NAO-like response in the NAE region, with a stronger impact on its northern lobe (i.e., the Icelandic low). In contrast, in late winter (January–March)

the canonical ENSO response prevails that involves a strengthening of the Aleutian low, and a positive geopotential height response over North America, which extends to Iceland. This is consistent with earlier findings (King et al. 2018a; Ayarzagüena et al. 2018; AKM2021). AKM2021 pointed out that strong changes in the ENSO-forced Indian Ocean heating response are accompanying the transition from early to late winter responses. The ensemble means of Niño-3.4 geopotential height (200 hPa) regressions (Figs. 1f–j), obtained by averaging the regression maps of geopotential height anomalies at 200 hPa onto the standardized Niño-3.4 index across all models, reveal that on average the models are able to reproduce the early (Figs. 1f,g) to late (Figs. 1h–j) winter ENSO response transition in the NAE region. However, the early winter signal in the North Atlantic region is weaker than in reanalysis.

Since rainfall is not assimilated, in this study we use 500-hPa pressure vertical velocity (omega) from NOAA-CIRES 20CR, which in tropical regions is a good proxy for rainfall and heating anomalies (e.g., Hamouda and Kucharski 2019). Figures 2a–e show the regression of the observed Niño-3.4 index onto omega from November to March. The omega field shows a dipole in the Indian Ocean with positive omega anomalies in the eastern and negative in the western basin,

FIG. 1. Regression maps of geopotential height anomalies at 200 hPa onto the standardized Niño-3.4 index for (a)–(e) NOAA-CIRES 20CR and (f)–(j) ensemble mean across all models (MME) for (top to bottom) November, December, January, February, and March. The black stippling indicates the grid points where the regression coefficients are statistically significant at the 90% confidence level, whereas the gray contours indicate the significance at the 95% confidence level. The unit of regression coefficients is m per standard deviation.

which is consistent with the TWEIO heating dipole identified by AKM2021 (positive heating corresponds to negative omega anomalies). However, the dipole is much stronger in early winter (November and December) compared to late winter (January–March). Furthermore, the MME means of Niño-3.4 forced 500-hPa vertical velocity (omega) for November to March is shown in Figs. 2f–j. The models also show a transition in the omega dipole response to ENSO, which is stronger in early winter (Figs. 2f,g) and becomes weaker in late winter (Figs. 2h–j).

The stronger TWEIO dipole response in early (particularly November) compared to late winter is also confirmed by the regression of the observed Niño-3.4 index onto the surface pressure from NOAA-CIRES 20CR, as shown in Figs. 3a–e. Also, the MME mean of SLP response to Niño-3.4 (Figs. 3f–j) shows the transition from early to late winter in the North Atlantic region, even if the signal in early winter is weak and slightly shifted to the south compared to reanalysis. As discussed in AKM2021, the TWEIO heating dipole in early winter excites a Gill-type response with an upper-level negative geopotential height anomaly, which induces an extratropical Rossby wave train that can propagate through the North Pacific and reach the North Atlantic projecting onto a positive NAO-type response there, whereas in late winter the canonical ENSO response prevails in the North Atlantic

projecting on a negative (positive) NAO phase for a warm (cold) ENSO event.

To investigate the ability of CMIP5 models in reproducing the early (December) to late (February) winter ENSO response transition, a Taylor diagram (Taylor 2001) is calculated, based on the spatial correlations between reanalysis and individual model 200-hPa height responses and their standard deviation in the NAE region (20°–80°N, 60°W–40°E) for early (December; Fig. 4a) and late winter (February; Fig. 4b). Out of 28 models under consideration, approximately 71% of models (i.e., 20 models) show a positive correlation in early winter, whereas in late winter this percentage increases to 93% (i.e., 26 models). This is quite predictable from the analysis of ensemble means; the models in late winter are in better agreement with the reanalysis compared to early winter, where fewer models show positive correlations. In the early (late) winter, out of 71% (93%) of models showing positive correlation, only 7 (2) models overestimate the geopotential height anomalies over the NAE region, while the rest underestimate it. Furthermore, Fig. 5 shows the spatial correlation coefficients between reanalysis and models ENSO responses in the NAE region for early (December; blue) and late (February; red) winter. Out of 28 models, 19 show a positive correlation during both months. We define a threshold of 0.3 to identify models with substantial positive spatial correlations in the NAE

FIG. 2. Regression maps of omega anomalies at 500 hPa onto the standardized Niño-3.4 index for (a)–(e) NOAA-CIRES 20CR and (f)–(j) ensemble mean across all models (MME) for (top to bottom) November, December, January, February, and March. The black stippling indicates the grid points where the regression coefficients are statistically significant at the 90% confidence level, whereas the gray contours indicate the significance at the 95% confidence level. The unit of regression coefficients is Pa s^{-1} per standard deviation.

region, and 13 models, namely BCC-CSM1.1, BNU-ESM, CMCC-CMS, GFDL CM3, GFDL-ESM2G, GFDL-ESM2M, GISS-E2-H, INM-CM4, MIROC-ESM-CHEM, MPI-ESM-P, MRI-CGCM3, MRI-ESM1, and NorESM1-ME, fulfill this criterion for both December and February. These models are referred to as “good” models in the following. As a sensitivity test, the ensemble mean of good models shown in the following sections is recomputed by recategorizing the models based on the higher stringent thresholds (e.g., 0.4 and 0.5) (figures not shown). The MME results of good models based on the higher stringent thresholds are entirely consistent with the ones presented in this study, which shows that the results are robust and insensitive to the selection of threshold value.

Figure 6 shows the composite of December ENSO responses of those so-called good models for 200-hPa geopotential height (Fig. 6a), omega (Fig. 6b), and surface pressure (Fig. 6c). The ensemble mean of good models (MME Good) indeed shows a stronger dipole in the tropical Indian Ocean (Fig. 6b) compared to the MME mean (Fig. 2g). Moreover, the good models also show a stronger and more extended response in the subtropical SAJET region as well as a clear indication of a positive NAO-like response in the NAE region in the 200-hPa geopotential height (Fig. 6a compared to Fig. 1g), which indicates the importance of the ENSO forced tropical Indian Ocean response and its role as a “bridge” for the wave propagation into the NAE region in early winter.

The sensitivity of the results with respect to the ENSO flavors were also investigated in the reanalysis data. Specifically, the Niño-3 (Fig. S1 in the online supplemental material) and ENSO Modoki (Fig. S2) indices show a similar behavior to Niño-3.4 index in their response of the ENSO transition in 200-hPa geopotential height, omega, and surface pressure in the North Atlantic region, even though with some changes in magnitude and position. In summary, the early to late winter response transition is not sensitive to the ENSO index chosen, and this adds to the robustness of this transition.

To analyze the importance of the subtropical SAJET modification in the early winter ENSO response, a scatterplot with the 200-hPa geopotential height Niño-3.4 regression coefficients area-averaged over the subtropical SAJET region (25° – 35° N, 60° – 120° E) is plotted against the spatial pattern correlation of 200-hPa geopotential height Niño-3.4 regressions in the NAE region (20° – 80° N, 60° W– 40° E; Fig. 7) for December. There is a moderate but statistically significant negative relation (-0.38), indicating that the models that have a stronger negative 200-hPa geopotential height regression with ENSO in the subtropical SAJET region also show a larger spatial correlation of the regression pattern in the NAE region, consistent with the reanalysis. Approximately 68% of the CMIP5 models (i.e., 19 out of 28 models under consideration) reproduce the ENSO forced tropical Indian Ocean teleconnections with NAE (i.e., showing the positive spatial correlation over the NAE region).

FIG. 3. Regression maps of SLP anomalies onto the standardized Niño-3.4 index for (a)–(e) NOAA-CIRES 20CR and (f)–(j) ensemble mean across all models (MME) for (top to bottom) November, December, January, February, and March. The black stippling indicates the grid points where the regression coefficients are statistically significant at the 90% confidence level, whereas the gray contours indicate the significance at the 95% confidence level. The unit of regression coefficients is hPa per standard deviation.

b. Response to the TWEIO in early winter

AKM2021 have shown that SAJET is strongly modulated by the tropical heating Indian Ocean anomalies. In this section,

FIG. 4. Taylor diagram of the spatial regression coefficients obtained by regressing geopotential height anomalies at 200 hPa onto the standardized Niño-3.4 index over the NAE region (20° – 80° N, 60° W– 40° E) for (a) December and (b) February.

the response to the pure TWEIO index (i.e., independent of ENSO) is analyzed for early winter (December). But before examining this, the model's fidelity in simulating the ENSO–TWEIO teleconnection is scrutinized through correlation analysis between Niño-3.4 index and TWEIO index for December, as shown in Fig. S3. Figure S3 reveals that all models except for two show the correct sign of the Niño-3.4–TWEIO relationship, and 20 out of 28 models yield statistically significant correlations with 95% confidence. However, the ensemble mean shows a weaker correlation than the reanalysis, which is consistent with a weaker impact of the TWEIO in the ENSO teleconnections. Figure 8 shows the regression of the pure TWEIO index onto the geopotential height anomalies at 200 hPa (Fig. 8a), omega at 500 hPa (Fig. 8b), and surface pressure (Fig. 8c) from NOAA-CIRES 20CR. Clearly, there is a strong modification of the subtropical SAJET, leading to a wave train that reaches the North Atlantic, projecting on the positive phase of the NAO. The 200-hPa height response in the subtropical SAJET region is consistent with Sverdrup balance (Rodwell and Hoskins 2001) and leads to the initiation of a wave propagation through the North Pacific into the NAE region.

To investigate the wave propagation from the Indian Ocean, Fig. 9a shows the wave activity flux at 200 hPa (Takaya and Nakamura 2001) for the reanalysis. The North Atlantic response is clearly connected to a meandering wave propagation initiated in the subtropical SAJET region, crossing the Pacific

FIG. 5. Spatial pattern correlation of Niño-3.4 geopotential height (200 hPa) regressions between NOAA-CIRES 20CR and individual CMIP5 models over the NAE region (20° – 80° N, 60° W– 40° E) for early (December) and late winter (February).

and entering the Atlantic in the Caribbean region, with a substantial reinforcement in the North Atlantic. The ultimate source of the wave is the heating dipole in the Indian Ocean, which generates an upper-level trough in the subtropical SAJET region through Sverdrup balance (AKM2021; Rodwell and Hoskins 2001). This trough is the starting point of the wave propagation. The waves initiated in the subtropical SAJET region resemble waves in the circumglobal teleconnection (CGT) first demonstrated by Branstator (2002). These waves are meridionally trapped and able to influence the whole Northern Hemispheric extratropics, including the North Atlantic region. The wave activity flux of the good models (Fig. 9b) also clearly shows a strong wave activity emerging from the SAJET with waves in phase with the reanalysis, which becomes weaker as the wave propagates into the North Pacific and Atlantic regions. This confirms the SAJET region as a possible relevant source to explain the NAE early winter response to ENSO/TWEIO.

c. Reason for model diversity

According to Figs. 4 and 5, there is a large diversity among the models in simulating the early-to-late winter ENSO response transition to the NAE region. We further analyze the behavior of the two extreme categories of models. It is noted that both the good models (MME Good; Figs. 10a–c) and the multimodel ensemble (MME; Figs. 10d–f) reproduce the main features of the observed 200-hPa geopotential height, 500-hPa omega, and SLP regressions onto the normalized pure TWEIO index (Figs. 8a–c). These results confirm that the TWEIO can trigger a positive NAO-type response independently of ENSO, and its coincidence with ENSO in early winter is responsible for the ENSO response transition in the NAE region discussed in section 4b. The CMIP5 models are generally able to trigger the wave train from the TWEIO, but they vary in their ability to show the covariability of TWEIO with ENSO in early winter. The pure TWEIO index (independent of ENSO) is not strongly related to local or remote SST forcing (for details, see Fig. S4).

On the other hand, to further investigate why some models are unable to reproduce the ENSO teleconnection transition in the North Atlantic, in Fig. 11 the ensemble means of ENSO regressed fields for the poor models (MME Poor) in early winter are analyzed. Poor models are defined as the models

FIG. 6. Ensemble means of regression patterns, obtained by averaging the regression maps of (a) geopotential height (m) anomalies at 200 hPa, (b) omega (Pa s^{-1}) anomalies at 500 hPa, and (c) SLP (hPa) anomalies onto the standardized Niño-3.4 index across all good models (MME Good) for December. The black stippling indicates the grid points where the regression coefficients are statistically significant at the 90% confidence level, whereas the gray contours indicate the significance at the 95% confidence level. The units of regression coefficients are per standard deviation.

FIG. 7. Scatterplot of the spatial pattern correlation of Niño-3.4 geopotential height (200 hPa; GPH200) regressions over the NAE region (20°–80°N, 60°W–40°E) vs Niño-3.4 GPH200 regressions (unit is m per standard deviation) area averaged over the subtropical SAJET region (25°–35°N, 60°–120°E) for December.

having a negative spatial pattern correlation of 200-hPa geopotential height Niño-3.4 index regression in the NAE region for December (for details see Fig. 5). Out of 28 models, 8 models (CMCC-CESM, GISS-E2-R, HadCM3, HadGEM2-AO, IPSL-CM5A-LR, IPSL-CM5A-MR, IPSL-CM5B-LR, and MIROC5) fall into the “poor” category. The ensemble mean of poor models divulge a negative NAO-like pattern in the North Atlantic region (Figs. 11a,c). The TWEIO heating dipole is weaker and the negative node is shifted to the central Indian Ocean and it becomes more like a north–south pattern (Fig. 11b). In the 200-hPa geopotential height regression, the trough in the SAJET region is almost completely missing with just a weaker hint of it over East Asia (Fig. 11a). This result is similar to the ones obtained through pure ENSO (i.e., independent of TWEIO) regressions (see Figs. S5–S7). Clearly, in both reanalysis and the models in the pure ENSO regressions (Figs. S5–S7), the positive NAO-like signal in early winter (December) is reduced compared to the “full” ENSO response (Figs. 1–3). The late winter (February) transition is reduced only in the reanalysis significantly. To quantify this in more detail, the NAO index—defined as the difference of area-averaged regressed anomalies over the Azores high (20°–35°N, 50°W–0°) and Icelandic low (40°–70°N, 50°–5°W)—based on pure Niño-3.4 SLP regressions is computed (see Fig. S8) and compared with the one based on the full Niño-3.4 SLP regressions (see Fig. 12). It is clearly observed that for December the NAO response is reduced by about 50% in both reanalysis and models in case of pure ENSO regressions, compared to full ones. This means that there is an important impact of the Indian Ocean on the North Atlantic ENSO response, but also that the other 50% of the NAO signal might originate through different ENSO pathways, including the Caribbean pathway (Ayarzagüena et al. 2018). Furthermore, even the good models only reproduce about 50% of the NAO response seen in the reanalysis. This indicates that some important processes may be missing in the models.

To identify the reasons for the model diversity further, we have calculated the TWEIO index based on Niño-3.4 omega

500-hPa regressions (see Fig. 13). Models show in general a weaker response compared to reanalysis, particularly in early winter. However, for December MME of the good models does show a larger response (about 18%) compared to the

FIG. 8. Regression maps of (a) geopotential height (m) anomalies at 200 hPa, (b) omega (Pa s^{-1}) anomalies at 500 hPa, and (c) SLP (hPa) anomalies from NOAA-CIRES 20CR onto the standardized “pure” TWEIO index (independent of ENSO) for December. The black stippling indicates the grid points where the regression coefficients are statistically significant at the 90% confidence level, whereas the gray contours indicate the significance at the 95% confidence level. The units of regression coefficients are per standard deviation.

FIG. 9. Regression map of geopotential height (m) anomalies at 200 hPa onto the standardized “pure” TWEIO index (independent of ENSO) for (a) NOAA-CIRES 20CR and (b) the ensemble mean across all good models (MME Good) for December. The vectors represent the wave activity flux ($\text{m}^2 \text{s}^{-2}$) calculated with regressed anomalies. The units of regression coefficients are per standard deviation.

MME of the poor models. This explains the overall weak North Atlantic response in all models compared to reanalysis, and the slight improvement in the good compared to the poor models. Additionally, the circulation anomalies in the North Atlantic,

as measured by NAO index in Niño-3.4 SLP regressions (shown in Fig. 12), clearly reveals that good models perform by construction much better in the early winter (December) response, as well as in the transition to late winter compared to poor models. We have also considered the relationship between the TWEIO index based on Niño-3.4 omega 500-hPa regressions and the NAO index based on Niño-3.4 SLP regressions, which is positive, but relatively weak (0.25). Furthermore, Fig. S9 shows the difference in ENSO regressed fields between the ensemble mean of good (MME Good; Fig. 6) and poor (MME poor; Fig. 11) models, confirming the changes in the Indian Ocean as one of the major candidates for the differences.

Overall, this result is consistent with the hypothesis that the ENSO modulated North Atlantic circulation anomalies in early winter originate in part from the subtropical SAJET region, where upper-level wind and geopotential height anomalies are forced by the TWEIO dipole heating anomalies.

5. Summary and conclusions

In this study the role of TWEIO heating dipole in the early-to-late winter ENSO-response transition in the NAE region has been revisited using a twentieth-century reanalysis and CMIP5 model outputs. In the reanalysis dataset, the transition from a response that spatially projects on the positive phase of the NAO to the canonical late winter ENSO response that spatially projects on the negative NAO phase is robust in the twentieth century. It is identified that TWEIO heating dipole

FIG. 10. Ensemble means of regression patterns, obtained by averaging the regression maps of (top) geopotential height (m) anomalies at 200 hPa, (middle) omega (Pa s^{-1}) anomalies at 500 hPa, and (bottom) SLP (hPa) anomalies onto the standardized “pure” TWEIO index (independent of ENSO) across (a)–(c) all good models (MME Good) and (d)–(f) all models (MME) for December. The black stippling indicates the grid points where the regression coefficients are statistically significant at the 90% confidence level, whereas the gray contours indicate the significance at the 95% confidence level. The units of regression coefficients are per standard deviation.

FIG. 11. Ensemble means of regression patterns, obtained by averaging the regression maps of (a) geopotential height (m) anomalies at 200 hPa, (b) omega (Pa s^{-1}) anomalies at 500 hPa, and (c) SLP (hPa) anomalies onto the standardized Niño-3.4 index across all poor models [MME Poor; having negative spatial pattern correlation of Niño-3.4 geopotential height (200 hPa) regressions over the NAE region] for December. The black stippling indicates the grid points where the regression coefficients are statistically significant at the 90% confidence level, whereas the gray contours indicate the significance at the 95% confidence level. The units of regression coefficients are per standard deviation.

plays a significant role in this transition present in the twentieth-century reanalysis. This Indian Ocean heating dipole is strong in early winter, but weakens considerably in late winter, and thus influences the ENSO response only in early winter significantly. We also noted that TWEIO can independently force the subtropical SAJET. The physical mechanism for the NAE region response to the TWEIO

heating dipole has also been revisited. It is consistent with a Gill-type response to the tropical Indian Ocean heating dipole that generates an upper-level trough in the subtropical SAJET region, which is the starting point for the Rossby wave propagation that traverses the North Pacific and enters the North Atlantic region. An analysis of wave activity flux confirms the origin of this wave train in the subtropical SAJET region. Furthermore, the early-to-late winter transition of the North Atlantic circulation anomalies is not sensitive to the exact ENSO definitions (i.e., Niño-3; ENSO Modoki).

The ENSO-response transition is to some extent reproduced by CMIP5 models, although with smaller amplitude and with varying pattern correlations in the North Atlantic region. Generally, the late winter response is better reproduced than the early winter response to North Atlantic circulation anomalies, because of the model's ability to simulate the direct ENSO response, which is stronger in late compared to early winter. Models that show a larger pattern correlation in the North Atlantic response in early winter also show a stronger trough response in the subtropical SAJET region with stronger Indian Ocean heating dipole. Instead, models with a negative ENSO response pattern correlation in the North Atlantic region also fail to show a strong response in the subtropical SAJET region. Overall the early winter TWEIO heating dipole response to ENSO in the Indian Ocean in the CMIP5 models is substantially weaker than that of the reanalysis. Furthermore, the difference between good and poor models in this response is rather small. Even though there is a weak relationship between the TWEIO response to ENSO and the North Atlantic response to ENSO, it is likely that also other factors are important in explaining the differences between good and poor models in their NAE response to ENSO. For example, the simulations of ENSO teleconnection pathways through the tropical North Atlantic and Caribbean regions or the simulation of ENSO itself could be candidates for explaining parts of the differences. However, it is beyond the scope of this study to analyze these hypotheses in detail. On the other hand, CMIP5 models reproduce overall well the North Atlantic response in early winter to TWEIO heating dipole independent of ENSO. This study further highlights the importance of the ENSO–interbasin teleconnections and

FIG. 12. Bar plot of NAO index, defined as the difference of area-averaged Niño-3.4 SLP regressions (obtained by regressing SLP anomalies onto the standardized Niño-3.4 index) over the Azores high (20° – 35°N , 50°W – 0°) and Icelandic low (40° – 70°N , 50° – 5°W).

FIG. 13. Bar plot of TWEIO index based on Niño-3.4 omega 500-hPa regressions, obtained by regressing omega (500 hPa) anomalies onto the standardized Niño-3.4 index.

their response to the NAE circulation anomalies in the coupled models.

The findings of this study are complementary to those of Ayarzagüena et al. (2018), who also found a precipitation response and, therefore, heating dipole response to ENSO to be relevant for the early winter ENSO teleconnection to NAE region. They analyzed the influence of a Rossby wave source induced by a precipitation dipole between the Caribbean Sea and the Gulf of Mexico. In the present study, the importance of an Indian Ocean heating dipole has been analyzed. It is likely that both influences are important, and one or the other may dominate the early winter ENSO teleconnection pathway in individual events.

Acknowledgments. Manish K. Joshi gratefully acknowledges the Director of IITM for support and encouragement. IITM is fully supported by the Ministry of Earth Sciences, Govt. of India. Manish K. Joshi also thanks ICTP, Trieste, Italy, for providing the facilities during his visits to the Centre under the Junior Associateship award received from ICTP. The authors thankfully acknowledge the World Climate Research Programme's Working Group on Coupled Modeling that is responsible for CMIP, and also the climate modeling groups for producing and making their model outputs available. For CMIP the U.S. Department of Energy's Program for Climate Model Diagnosis and Intercomparison provides co-ordinating support and led development of software infrastructure in partnership with the Global Organization for Earth System Science Portals. Support for the Twentieth Century Reanalysis Project V2c dataset is provided by the U.S. Department of Energy, Office of Science Biological and Environmental Research (BER), and by the National Oceanic and Atmospheric Administration Climate Program Office. The authors thank the three anonymous reviewers and the editor for their constructive and fruitful comments/suggestions, which helped us in improving the manuscript substantially.

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