Neutral Mode Dominates the Forced Global and Regional Surface Temperature Response in the Past and Future

Fukai Liu1,2,3, Jian Lu4,5, and L. Ruby Leung4

1Physical Oceanography Laboratory, Frontier Science Center for Deep Ocean Multispheres and Earth System (FDOMES), Ocean University of China, Qingdao, China, 2Laboratory for Ocean Dynamics and Climate, Qingdao Pilot National Laboratory for Marine Science and Technology, Qingdao, China, 3College of Oceanic and Atmospheric Sciences, Ocean University of China, Qingdao, China, 4Atmospheric Sciences and Global Change Division, Pacific Northwest National Laboratory, Richland, WA, USA

Abstract Using a large suite of Green’s function perturbation experiments, we construct the emergent dynamical operator for surface temperature and extract the most excitable mode. The leading mode turns out to be the most dominant mode excited by the climate forcing of doubling CO₂, alone capturing 56% of the total spatial variances of the surface temperature response. The pattern of the leading mode is partly organized by the atmospheric annular modes in both hemispheres, and further modulated by the feedbacks from ocean currents. Though derived from a single model, the leading mode can capture the spatio-temporal variabilities of the global mean surface temperature over the past century in both the CMIP6 model simulations and observations. Moreover, the leading mode is most efficiently excited by radiative forcings from the midlatitude bands centered around 45°N and 25°S, where most human activities take place, suggesting the potency of human influence on the climate change.

Plain Language Summary Confidence in the regional features of climate change projections is a prized commodity for climate adaptation and mitigation. This study strives to decipher the pattern of the forced climate response using a dynamical mode-based framework such that certain confidence may be established for regional climate projections. The dynamical modes, also known as the neutral modes, are extracted through singular value decomposition of the linear response function of a coupled climate model using a large suite of Green’s function perturbation experiments. The leading mode turns out to be the most dominant mode excited by anthropogenic climate forcing, capturing the global mean surface temperature evolution over the past century in both observations and multi-model simulations, lending confidence to the regional climate change features captured by the leading mode. Coincidentally, the leading mode is most efficiently excited by radiative forcings from the midlatitude bands centered around 45°N and 25°S where human activities are most prominent.

1. Introduction

Barring the broadest spatial features, confidence in the regional details of climate change projections is still lacking (Neelin et al., 2006), which has been a longstanding challenge to the climate community (Beniston, 2013; Mlawer et al., 1997). The projected increase in heat waves and extreme precipitation has been largely premised on the thermodynamics-regulated temperature and moisture changes (Fischer et al., 2021; Madakumbura et al., 2021; Mantua & Hare, 2002), whereas the lack of robustness in the projected regional precipitation response by climate models has been blamed on the dynamics-regulated atmospheric and oceanic circulation changes (Biaisutti et al., 2018; Hall, 2014; Xie et al., 2015). Since no one lives in the global mean climate, improving the confidence in regional-scale climate change projections is critical for adaptation and mitigation planning at national and state (provincial) levels. As astutely pointed out by Xie et al. (2015), the challenge of unreliable regional climate change information cannot be addressed solely by increasing model resolution or regional downscaling; advancing understanding of the large-scale coupled dynamics between the ocean and atmospheric circulation might be foundational to narrowing the uncertainties on scales greater than 100s km.

A promising way forward to address the challenge above may be through the lens of the intrinsic dynamical modes of the climate system. The rationale behind the mode approach is that the slow, large-scale evolution/response of the Earth climate is governed by sparse dynamics (Brunton, Proctor, et al., 2016), although climate is a complex, nonlinear, and high-dimensional system. Moreover, dynamical mode manifests the deterministic...
aspect of the system, more dominated by the well-understood dynamical processes than the less certain, parameterized physical processes. Indeed, mode behavior in climate change response has already been demonstrated in model configurations with different complexities. For example, in fully coupled climate models, the sea surface temperature (SST), surface air temperature, and precipitation response to the more regional aerosol forcing have been found to bear remarkable resemblance to the response to globally uniform greenhouse gases (GHGs; Boer & Yu, 2003; Xie et al., 2013). Moreover, aerosol forcings from different locations (North America, Europe, East, and South Asia, respectively) can excite temperature response with similar spatial patterns (Kasoar et al., 2018). In an idealized aquaplanet atmospheric general circulation model (AGCM), a preferred zonal-mean temperature pattern, featuring an enhanced warming in the tropical upper troposphere and a warming in the polar lower troposphere, can be recurrently excited by surface heating prescribed from different latitudinal bands (Kang et al., 2017). More encouragingly, the power of the most excitable mode has been demonstrated to explain the most prominent feature of the temperature response to CO₂ forcing—polar amplification (Alexeev, 2003; Alexeev et al., 2005; Langen & Alexeev, 2005, 2007). We are therefore emboldened to extend the neutral mode (NM) approach to build confidence in the more regional aspects of the climate change response.

For the Earth's climate, the surface temperature (TS) is governed by complex interplay among many different components constituting the climate and no single physical law-based equation can fully capture its response and evolution. Nevertheless, the linear response function (LRF) approach can encompass the dominant dynamics for the TS in a single matrix, and is useful in understanding the modal behavior of the climate system and identifying a causal relationship between the forcing and response (Liu, Lu, Garuba, Huang, et al., 2018; Liu, Lu, Garuba, Leung, et al., 2018; Lu et al., 2020). To compute the LRF and its corresponding NMs, we perform a large suite of Green’s function perturbation experiments with the fully coupled Community Earth System Model version 1 (CESM1) by perturbing the shortwave forcing at the top of the atmosphere from an array of cells covering the globe, one at a time. As will be shown later, the LRF exhibits considerable skill in predicting the global TS response given the climate change forcing. Notably, we found that the response to radiative forcing of different sources, such as GHGs versus aerosols, are predominantly the result of the excitation of the first NM of the derived LRF, lending confidence to many of the observed and simulated regional features of the climate warming.

2. Methods and Data

2.1. Linear Response Function and Neutral Modes

Assuming quasi-equilibrium, the slow-evolving state of TS (T) can be thought of as being maintained by

$$\mathcal{L} \delta T_i = \delta f$$

(1)

where $\delta T_i$ is the linear component of the TS response, $\mathcal{L}$ is the LRF that relates the linear response to the external radiative forcing, and $\delta f$ is the radiative forcing perturbation at TOA. There are multiple methods to estimate $\mathcal{L}$, among which the Green’s function approach has proved to be more accurate and effective (Dong et al., 2019; Hassanzadeh et al., 2016; Lin et al., 2021; F. Liu, Lu, Garuba, Leung, et al., 2018; Y. Wu et al., 2021), despite the computational cost.

As the surrogate of the governing dynamics, the constructed LRF can predict not only the TS response to arbitrary external forcing without running the climate model, but also express the response in terms of linear superposition of the dynamical modes of the system. By estimating the LRF, the dynamical modes can be derived as the leading singular right vectors of the LRF (Lu et al., 2020). The singular vector with the smallest singular value is least damped and hence most excitable, thus begetting its reference as the NM. The corresponding left singular vector, on the other hand, gives the corresponding optimal forcing of the mode. The NM decomposition of a climate response also serves to stratify the response into modes with different confidence, as the first neutral mode (NM1) tends to be least susceptible to noise and is well separated from other NMs, hence the most robust (see Supporting Information for further details on the LRF framework).

2.2. Green’s Function Experiments

To construct the LRF that links external radiative forcing to TS response, we conduct a large set of Green’s function forcing experiments using CESM1 of the National Center for Atmospheric Research (NCAR), and the
model grid is T31_gx3v7 (nominal 3° resolution). The choice of the coarse-resolution configuration is due to the
large computational expense: about 25,000 years of simulation are required by Green's function experiments. The
atmospheric component of CESM1, the Community Atmospheric Model version 4 (CAM4) (Neale et al., 2013),
is vertically spaced in a hybrid pressure-σ coordinate with 26 levels, and has a resolution of 3.75° × 3.75° in
the horizontal. The land component Community Land Model version 4 (CLM4) (Lawrence et al., 2012), has
the same horizontal resolution as CAM5. The ocean component, the Parallel Ocean Program version 2 (POP2)
(Danabasoglu et al., 2012), has a nominal 3° horizontal resolution with meridional grid spacing decreasing to
~0.3° near the equator. Vertically, it has 60 unevenly spaced vertical layers with the highest resolution of 10m
near the surface. The sea ice component, the Community Ice CodE (CICE) (Hunke et al., 2015), has the same
horizontal grid as POP2.

Prior to the perturbation runs, we first integrate a long control (CTRL) for 500 years by forcing the CESM1
with GHGs, aerosols, and solar insolation at preindustrial level. The CTRL starts from an equilibrium state
that is available at NCAR (b40.x31 x 3.03). To facilitate the Green's function forcing experiments, we design an
array of 120 patches, covering the whole globe (as illustrated in Figure S1 in Supporting Information S1). The
widths of the rectangular patch in zonal and meridional directions are 45° and 12°, respectively. For each patch,
a pair of perturbation simulations are conducted, one with a positive and another with a negative shortwave radi-
ation perturbation at the top of the atmosphere, branching out from the 401th year of the CTRL run and being
integrated for 100 years. Only the last 50 years are used for analysis, with the prior years discarded as spin-up.
We caution that the perturbations runs are far from reaching equilibrium with the forcings, so the last 50 years
mainly capture the stabilized fast response component of the climate system (e.g., Held et al., 2010). Using the
last 30 years instead for the analysis does not lead to any qualitative change in the resultant leading NM and
optimal forcing (Figure S2 in Supporting Information S1). The solar perturbation is realized by multiplying the
baseline solar flux by a constant factor so that the maximum magnitude of the generated perturbation is ~30 W
m^{-2} (Figure S1 in Supporting Information S1). We perturb radiation at the TOA rather than the surface because
the TOA perturbation can maximize the simulated climate system's freedom to internally select the partitioning
of the energy transport response between the atmosphere and ocean (Yu & Pritchard, 2019). In addition, it is
found that the warming patterns induced by CO₂ forcing and a uniform surface radiative forcing are remarkably
similar, hinting that the TS responses are insensitive to whether the forcing is from the TOA or the surface (Figure
S3 in Supporting Information S1). With the 120 pairs of Green's function experiments, we can construct the LRF
relating the linear TS response to the TOA radiative forcing and obtain the NMs via singular value decomposition
(SVD) analysis. Note the TS responses are expressed in reduced dimension on the empirical orthogonal functions
of the TS in the long CTRL simulation (see details in Supporting Information).

In addition, the Green's function experiments, an extra doubling CO₂ experiment (2xCO₂) is also integrated for
100 years as an independent test case for evaluating the skill of the prediction made by the LRF.

2.3. Observations and CMIP6 Experiments

We use both observations and CMIP6 simulations to examine the contribution of NMs in modulating the
low-frequency TS variability. For observations, we use the NASA Goddard Institute for Space Studies Surface
Temperature Analysis (GISTEMP), which incorporates SST observations over oceans and surface air tempera-
ture observations over land. We mainly focus on the 1910–2020 period because observations are sparse in many
regions before 1910, and the limited missing values after 1910 are filled using spatial bilinear interpolation.

In addition to the observation, we analyze twentieth-century historical (years 1850–2014), abrupt CO₂, quadrupling
(years 1–150), and the preindustrial simulations (years 1–150) in CMIP6. We also use historical simulations
from the Detection and Attribution Model Intercomparison Project (DAMIP) (Gillett et al., 2016), which separate
contributions from GHGs, anthropogenic aerosols, and other external forcings. For each model, only the first
member is used, all model outputs with different grids are interpolated to the nominal 3° grid that is used in our
Green's function experiments. The models included in this study are summarized in Supporting Information.
3. Results

3.1. The NM1 as the Most Excitable Response Pattern to CO2 Forcing

Figure 1a shows the nondimensionalized spatial pattern of the NM1 for global TS. Beyond the global mean warming component, the NM1 is overall characterized by an interhemispherically asymmetric pattern, with accentuated Arctic warming, enhanced warming over the NH land relative to the ocean, warming in the western boundary current regions over the midlatitude oceans, a warming hole in the subpolar North Atlantic, and lack of warming in the Southern Ocean. The NM1 pattern bears a marked resemblance to the pattern of the TS response induced by CO2 doubling (Figure 2a), accounting for 56% of the spatial variance of the latter. It is worth noting

![Figure 1. Spatial pattern of the leading neutral vector. (a) The spatial pattern of the NM1; (b) the corresponding optimal forcing; Shown in the side panels are the corresponding zonal mean profiles. Note the amplitude is all nondimensionalized as they are derived from SVD analysis. NM1, first neutral mode; SVD, singular value decomposition.](image)

![Figure 2. Predicting the TS response with the linear response function (LRF). (a) The TS response to 2xCO2 in CESM1. (b) Prediction of (a) with the LRF (in K). (c) The observed TS trend between 1966 and 2020 (c; in K/10 years). See Supporting Information for details on calculating the LRF-based prediction. Shown in the right side panels in (a)–(c) are the corresponding zonal mean profiles. (d) Scatter plot between the regression coefficient of the GMST anomalies at 4xCO2 against NM1 versus the climate sensitivity (in K) in CMIP6 models. The gray line represents an ordinary least squares regression fit to 39 CMIP6 models ($r = -0.97$). The CMIP6 models are color coded according to the order of the CMIP6 models listed in Supporting Information.](image)
that the power of the leading dynamical mode in capturing the pattern of the global warming response has been established previously in more idealized contexts (Alexeev, 2003; Langen & Alexeev, 2007), wherein the feature of polar amplification in the zonal mean temperature can be interpreted as the manifestation of the most excitable mode.

A dimensional prediction of the TS response to CO$_2$ doubling can be made by applying the inverse of the LRF to the direct radiative forcing at the TOA due to the CO$_2$ increase (Figure S4 in Supporting Information S1, see details in Supporting Information). The prediction (Figures 2a and 2b) reaffirms the dominance of the NM1 in capturing the CO$_2$-forced response, with all the afore-discussed spatial features present in both the predicted pattern and the CESM simulated response. Remarkably, the predicted pattern even captures many features in the observed TS trend pattern from 1966 to 2020 based on GISTEMP data (Figure 2c), with a global pattern correlation at 0.76. The spatial agreement lays a dynamical basis for the preferred warming over the northern high latitudes, northeastern American continent, Saharan Africa, and Middle East, and the lack of warming over the subtropical Pacific and Atlantic, the subpolar North Atlantic, and the Southern Ocean.

While some robust features in NM1 TS are rooted in the fundamental feedbacks of the climate system, such as the Arctic amplification due largely to the ice-albedo feedback and lapse-rate feedback, the ocean and atmosphere circulation associated with the NM1 also plays an important role in dynamically organizing the spatial structures of the NM1 TS pattern. The circulation features congruent with the NM1 can be computed by treating them as the responses to the optimal forcing and the results are presented in Figure 3 (left column). The NM1 pattern of sea level pressure (SLP; Figure 3a) projects strongly onto the positive annular mode in both hemispheres, as manifested in the elevated subtropics-to-midlatitude pressure (especially over the ocean basins) and the subdued pressure in higher latitudes. Dynamically consistent with these SLP structures is the meridional dipole in the zonal mean zonal wind pattern associated with the NM1, indicating a poleward shift of the eddy-driven component of the westerly jet (Figure 3b). In the ocean, the barotropic ocean circulation represented by the barotropic streamfunction is dynamically consistent with the overlying wind stress pattern as well (Figure 3c). In response to the poleward shift of the near-surface winds, the wind-driven ocean circulation gyres, especially the western portion of them, also shift toward the pole in both hemispheres. The synchronizing effect of the westerly jets over both the Atlantic and Pacific basins on the underlying ocean western boundary currents and SST has been recently noted in both observations and climate models, and identified to be an interbasin air-sea coupled mode at low frequency (Kohyama et al., 2021). The LRF analysis here further reveals that the interbasin air-sea coupled mode is an intrinsic mode of the coupled climate system. Moreover, the strong projection of the CO$_2$-forced response onto the NM1 mode (compare the lower panels with the upper ones in Figure 3) harbingers a more
positive synchronization mode in a warmer climate, with profound impacts on the marine ecosystems (Pershing et al., 2016; Saitoh et al., 1986).

The circulation features in both the ocean and atmosphere provide the dynamical underpinning for the many regional TS features discussed earlier. For example, the strong continental warming in northern Eurasia and northeastern North America is closely tied to Northern Annular Mode (Folland et al., 2009; Hurrell et al., 2003); the accentuated SST warming over the northwestern Pacific is the result of the enhanced warm advection due to the northerly shifted Kuroshio and its extension (L. Wu et al., 2012), likewise for the SST warming over the Gulf stream (Saba et al., 2016); the lack of warming in the midlatitude Southern Ocean can be attributed partly to the mean ocean upwelling, acting as a thermostat to the surface thermal perturbation (Armour et al., 2016), and partly to the enhanced and shifted wind-driven ocean heat uptake (W. Liu et al., 2018); the warming hole in the subpolar North Atlantic has long been noticed and it is the result of the weakening of the AMOC and the enhanced oceanic energy transport into the Arctic (Keil et al., 2020). Indeed, the weakening of the AMOC is found to be a concomitant feature of the NM1 here (Figure S5 in Supporting Information).

What is the origin of the circulation patterns of the NM1? Ocean dynamics or atmospheric dynamics? This question may be addressed by comparing the atmospheric circulation pattern of the NM1 with that derived from a similar Green's function perturbation approach with the atmospheric component of CESM coupled to a motionless ocean slab (F. Liu, Lu, Garuba, Huang, et al., 2018; Lu et al., 2020). The fact that the slab counterpart of the NM1 (Figure S6 in Supporting Information, see also Figure 2e of Lu et al. (2020)) exhibits similar annular mode structure as the NM1 here strongly suggests the orchestrating role of atmospheric dynamics in the wind and ocean current structures shown in Figure 3. The meridional dipole pattern in the atmosphere is a natural result of conservation of angular momentum and mass by fluid motions (e.g., Gerber et al., 2008).

### 3.2. The Dominance of the NM1 in the Global TS Evolution

The NM1 can successfully capture the temporal evolution of the global mean surface temperature (GMST) as well, regardless it arises from external climate forcing or from internal variability. Figure 4a displays the running 15-year trend of the multi-model ensemble (MME) mean GMST in the historical simulations by the participating climate models of CMIP6 (solid line) and the replication of the GMST trend by the NM1 (dashed line), the latter being achieved by only accounting for the TS component congruent with the NM1 for the TS each year simulated by each model when computing the GMST (see details in Supporting Information). The NM1-simulated time series is highly correlated with the CMIP6 simulated one, with a Pearson correlation $r = 0.99$. Apart from the dominance on GMST, the NM1 can also capture the TS spatial structures going along with GMST during different warming/cooling epochs of the twentieth century regardless of the forcing agents (bars in Figure 4), such as the extended warming period from 1920s–1940s and that post-1970s, and the warming hiatus during the period of 1940s–1970s (Figure 4a). The high skill is somewhat unexpected, since the NM1 used is derived solely from a coarse resolution CESM1, not necessarily representing the leading mode of other climate models. In Figure S7 in Supporting Information S1, we show that the NM1 of CESM1 can indeed capture the slow evolution of GMST, as well as the accompanying spatial structures, for every single one of the CMIP6 models, implicating that the NM1 here captures the leading dynamical mode common across the models.

Taking advantage of the DAMIP that discriminates external radiative forcings of different origins, we repeat the exercise of replicating the GMST trend with the NM1 for the MME mean response forced by GHGs alone (Figure 4b) and anthropogenic aerosols alone (Figure 4c), respectively. Over the past 120 years, the NM1-simulated trends achieved similarly high skill in capturing the global TS evolution for these individual forcing cases as for
the Historical CMIP6 ensemble. These results corroborate the finding (Xie et al., 2013) that different radiative forcings with different spatial distributions tend to excite climate response with similar spatial patterns, despite their distinct radiative effects, pinpointing the NM1 as the underlying reason behind their similarity. Another feature to note is that the NM1 component tends to underestimate the global mean warming trend since 1970s forced by GHGs alone. Further analysis (not shown) suggests that more NMs are needed to fully capture the global mean warming.

3.3. The Optimal Forcing for Global TS Warming/Cooling

A unique power of the NM analysis through the SVD decomposition of LRF is it pairs the mode pattern with the corresponding optimal forcing. Figure 1b presents the optimal forcing of the NM1, the loading at each patch of the forcing indicates the potency of producing the NM1 response (and hence the GMST warming, as the NM1 projects most strongly onto the GMST) per unit TOA energy perturbation. The global TS warming potency so represented shows considerable dependence on the geographic locations: the midlatitude TOA forcings centered around 45°N in the northern hemisphere and 25°S in the southern hemisphere—regions that happen to be most densely populated—are most efficient in driving the NM1 and hence global mean warming. In contrast, the tropical forcings over the Atlantic and the west Pacific warm pool turn out to be feeble in exciting the NM1 and GMST response, owing to the strong local and nonlocal negative feedbacks there, especially through the cloud processes (F. Liu, Lu, Garuba, Leung, et al., 2018; Park et al., 2022; Zhou et al., 2017). The weak efficacy of the tropical Atlantic forcing in driving global mean response has been previously noted (Hill & Ming, 2012). The feebleness of the western equatorial Pacific radiative forcing does not necessarily contradict results from Dong et al. (2019), who highlighted the importance of the warm pool region in driving global changes. Dong et al. (2019) specified SST forcings over the warm pool region, which provide an exogenous energy source insusceptible to the feedbacks, driving strong global warming. By contrast, the effect of the radiative perturbations employed in this region is strongly damped by cloud radiative feedbacks, and thus both the local and remote TS responses are limited (Park et al., 2022).

For the polar regions, radiative forcings there have little or even negative influence on global mean temperature, in contrast to an earlier study (Kang & Xie, 2014) that ascribes much greater importance to the high-latitude forcing in an idealized model without sea ice and ocean dynamics, wherein the forcing perturbation must be prescribed as an ocean heat flux convergence through an ocean slab. This apparent conflict is likely the consequence of the different experimental settings, one is forced from the TOA, the other is from the slab ocean, to which the atmosphere is coupled. The highly reflective polar surface due to sea ice and snow cover could be another reason for the meager sensitivity of the NM1 to the TOA radiative perturbation near the poles (Figure S8 in Supporting Information S1).

The spatial distribution of the optimal forcing for the NM1 also bears some resemblance to the CO2 instantaneous radiative forcing (Figure S4 in Supporting Information S1): they are both characterized by an M-shape meridional distribution. Therefore, the CO2 radiative forcing can preferentially excite the NM1 by projecting strongly onto the NM1 optimal forcing; this also explains the dominance of the NM1 in the global warming response. Indeed, how much the CO2-forced response projects onto the NM1 tells a great deal about the climate sensitivity of the model. As shown in Figure 2d, about 94% of the inter-model spread in the global warming climate sensitivity, measured here by half of the GMST response to CO2 quadrupling, can be explained by the different projections of the CMIP6-simulated GMST response onto the NM1. When it comes to optimizing solar geoengineering forcing to counter the ongoing global warming, the optimal forcing pattern shown in Figure 1b may be a useful guide for deploying the forcing. Interestingly, Saharan Africa and Middle East are the two regions with the highest efficacy in exciting the NM1 and GMST warming. Building large-scale solar farms in these two regions might be a stone that can kill two birds: growing renewable clean energy and cooling the climate.

4. Summary and Discussion

By treating the forcing-response problem of the climate system as a linearized dynamical system, we have succeeded in extracting the most excitable modes—NMs—for global TS in a coupled climate model using a large suite of Green's function experiments. The resultant TS pattern of the NM1 bears great resemblance to the response to 2xCO2 as well as the observed TS trend since the 1960s, featuring elevated warming over the
Arctic, northern Eurasia, northeastern North America, the Saharan Africa and Middle East, and the western boundary current regions in the ocean, and muted warming in the North Atlantic and the Southern Ocean. The NM1 is found to be intrinsically organized by the atmospheric annular modes and the related wind-driven ocean dynamical feedbacks, explaining some of the above-mentioned regional TS features. The most notable finding from our results is that the NM1, though derived solely from CESM1, can well capture the slow TS evolution in both CMIP6 simulations and observations, implicating the relevance of the NM1 here to the mode behavior. The strong projection of the direct CO2 radiative effect on the optimal forcing of the NM1 underpins the dominance of the NM1 TS pattern in the climate change signal, giving confidence to some of the dynamically organized regional features. In the interim, the pattern of the optimal forcing of the NM1 can be a useful guide for where to put a cooling agent for solar geoengineering purpose.

Challenges abound as to applying the LRF approach for climate response and climate change mitigation purposes. It remains a challenge for the 3-degree climate model here to capture the features at subcontinental scales, not to mention the scale of 100s km as suggested in Xie et al. (2015). Consensus on the higher modes of the climate change projection remains to be established. If allowed computation-wise, it would be desirable to repeat the same calculation and analysis with the state-of-the-art climate models. Despite the skill of the NM1 in capturing the GMST evolutions in both CMIP6 simulations and the observations, we cannot rule out model dependence of the NM patterns, the robustness of which can only be assessed by inter-model comparison using similar experimental approach for different models. Another limitation here is that only the linear component of the response is considered in this study, while the real-world climate change response in a single realization can be affected by the intrinsic nonlinearity of the system (e.g., F. Liu et al., 2020; Lu et al., 2020). An even greater challenge is to predict accurately the climate forcing needed to cancel out the detrimental climate change impacts—equivalent to finding the optimal control for a dynamical system (Brunton, Brunton et al., 2016; Kravitz et al., 2016), a mathematical inverse problem entailing more innovative endeavors, such as combining Green's function experiments with advanced deep learning approaches. Notwithstanding the limitations above, confidence in many regional features of the forced climate change may be boosted through the lens of NMs of the climate system.

Data Availability Statement

The GISS Surface Temperature Analysis is freely available at https://data.giss.nasa.gov/gistemp/. The monthly CMIP6 model output is publicly available at https://esgf-node.llnl.gov/search/cmip6/. The code and data of the large suite of coupled patch experiments with CESM version T31_gx3v7 used for the analysis and data visualization in this study is released through the repository of Zenodo at https://doi.org/10.5281/zenodo.6768695.

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