Ocean Warming Pattern Effect On Global And Regional Climate Change

Shang-Ping Xie1

1Scripps Institution of Oceanography, University of California San Diego, La Jolla, CA, USA

Abstract  Anthropogenic emissions of greenhouse gases cause the planet to warm, and the ocean uptake of anthropogenic heat slows the warming, preventing the climate system from equilibrating with the increasing radiative forcing. The uneven ocean surface warming affects regional changes in tropical rainfall, El Niño, and the global climate sensitivity. Thus, the study of ocean warming patterns bridges the ocean-atmospheric dynamics community focusing on spatial patterns on one hand and the climate change community with a traditional emphasis on the planetary energy budget and radiative feedback on the other.

Plain Language Summary  The ocean surface warming pattern has emerged as a research frontier in climate science in relation to regional climate change and global climate sensitivity. In the high-latitudes, observations have largely confirmed the interhemispheric asymmetry in ocean warming climate models predicted between the North Atlantic and Southern Ocean, a pattern resulting from the pole-to-pole deep ocean overturning circulation. In the tropics, the ocean warming pattern is closely coupled with atmospheric convection and circulation, modulating tropical cyclone statistics and El Niño influence in faraway regions such as western North America. While climate sensitivity is often considered a constant inherent to a given model, recent studies show that it varies in time as the heat exchange between the surface and deep oceans causes the ocean surface warming pattern to evolve. This warming pattern effect highlights the challenges in estimating climate sensitivity from instrumental observations, which feature evolving radiative forcing (greenhouse gases vs. aerosols) as well as unforced internal variability.

1. Introduction

Three decades ago, then newly developed coupled atmosphere-ocean general circulation models (AOGCMs) were used to study climate response to gradual increase in atmospheric carbon dioxide (~1%/year) (Manabe et al., 1990; Washington & Meehl, 1989). The increased greenhouse effect causes the planet to warm overall and with distinctive spatial variations in the magnitude of warming. Recently, Stouffer and Manabe (2017) found that “the salient features of climate change distribution projected (by their model 30 years ago) are becoming evident in the observations.” Here we review advances made since in understanding the dynamics of surface warming patterns and the effect on the atmospheric circulation and regional climate change. We will discuss recent work that revealed a surprising pattern effect on the global-mean warming and hence climate sensitivity.

2. Asymmetry Between the Poles

Climate model simulations consistently show, and observations confirm, that surface warming over the Arctic is much larger than the rest of the globe. The warming melts snow and ice as illustrated by the rapid reduction in Arctic sea ice; the record minimum of September Arctic ice extent was set in 2007 and again in 2012, with 2019 matching 2007 as the second minimum. The reduced surface albedo allows more incoming solar radiation to be absorbed by the surface, and the enhanced surface warming melts even more snow/ice. In addition to this local snow/ice feedback, the increased atmospheric transport of water vapor/latent energy also contributes to the amplified warming over the Arctic.

Ocean circulation can cause surface warming to vary. Specifically, the deep water of the world ocean forms in the winter subpolar North Atlantic, flows southward, and upwells to the surface in the Southern Ocean (Marshall et al., 2014). The deep water that upwells in the Southern Ocean last sees the atmosphere...
hundreds of years ago and is free of the ongoing anthropogenic warming (dT = 0). The upwelling of the pristine water dampens the surface warming in the Southern Ocean (Bryan et al., 1988). The asymmetry between the amplified Arctic warming and the muted surface warming over the Southern Ocean (Figure 1a) appeared in the first AOGCM simulations and has been largely born out in observations. Consistent with the muted surface warming, sea ice change over the Southern Ocean is small compared with that in the Arctic.

Figure 1. Multimodel ensemble mean change in Coupled Model Intercomparison Project Phase 6 (CMIP6) runs where the atmospheric CO₂ concentration is increased 1% per year, calculated as the difference between Years 130–150 (the time of CO₂ quadrupling) and Years 1–20. (a) SST (°C). (b) Net surface heat flux (W/m², downward positive). (c) Percentage change in precipitation (color shading) and SST change (contours at intervals of 0.5°C) relative to the tropical mean (25°S–25°N), with a spatial correlation of 0.55 in the tropics.
The pristine water upwelled in the Southern Ocean comes to be in contact with the warming atmosphere, causing a strong downward net surface heat flux (Figure 1b). The surface water, while being heated, flows northward as the Ekman drift under the prevailing westerly winds and then subducts into the thermocline. In Argo observations of 2006–2013, most of the ocean heat storage is found south of 30°S (Roemmich et al., 2015), consistent with the strong heat uptake in the Southern Ocean. Here, the ocean heat uptake refers to surface heat flux change to distinguish from the heat storage of the water column; that is, heat storage = surface heat uptake + heat transport convergence.

The Atlantic meridional overturning circulation (AMOC) transports warm saline subtropical water into the subpolar North Atlantic, where it releases the heat and sinks to great depths. Under greenhouse warming, the increased poleward transport of water vapor causes precipitation to increase and salinity to decrease in high-latitude oceans. In the subpolar North Atlantic, the decreased surface salinity, together with the surface warming, reduces the water density at the surface, slowing down the deep-water formation and the AMOC. The reduced northward heat transport by the AMOC suppresses the ocean surface warming in the subpolar North Atlantic, and the disequilibrium with the warming atmosphere allows large ocean heat uptake (Figure 1b).

The AMOC slowdown predicted in early AOGCM simulations has not materialized in observations. These simulations considered only the greenhouse gas (GHG) forcing while anthropogenic aerosols are important in slowing Northern Hemisphere warming. In models that consider both GHG and aerosol effects, the AMOC stays stable through the historical period but is predicted to decelerate as the aerosol forcing peaks while the GHG forcing continues to rise (Delworth & Dixon, 2006). Anthropogenic aerosols reconcile the observed asymmetry in ocean heat uptake between the North Atlantic and Southern Ocean, but this asymmetry is expected to weaken, and the proportion of North Atlantic heat uptake is to increase (Shi et al., 2018) as societies act to curb air pollution. The recently deployed ocean observing systems are in place to observe this predicted AMOC deceleration (Smeed et al., 2018).

3. Tropical Patterns

Early global warming simulations predict a larger tropical ocean warming north than south of the equator and a northward shift of the tropical rain band called the intertropical convergence zone (ITCZ) (Manabe et al., 1990), patterns reaffirmed by state-of-the-art simulations in the Coupled Model Intercomparison Project (CMIP) archive (Figure 1). Observations, however, show a southward displacement of the ITCZ instead, especially in the Atlantic sector (Wang et al., 2016). With aerosols reflecting more solar radiation back to space in the Northern Hemisphere – also known as the global dimming of sunlight at the surface – the atmosphere needs to transport more energy from the Southern to Northern Hemisphere. This leads to a southward displacement of the ITCZ as the meridional energy transport is in the direction of the upper limb of the cross-equatorial Hadley overturning circulation (Hwang et al., 2013; Kang et al., 2008). The prolonged drought over the African Sahel from the 1950s to 1980s is due in part to the increase in aerosols over North America and Europe (Chiang & Friedman, 2012). The above energy theory relates cross-equatorial asymmetry in tropical rainfall and sea surface temperature (SST) to energy perturbations outside the tropics, due to anthropogenic aerosols in this case.

In GHG-forced AOGCMs, ocean surface warming is locally intensified over the eastern equatorial Pacific (Figure 1) due to the following effects. First, evaporation is the major mechanism to balance the strong downward radiation at the surface, and the evaporative damping coefficient is proportional to the mean evaporation (Xie et al., 2010). Over the eastern equatorial Pacific where the mean evaporation is weakest in the tropics due to low SSTs and weak winds, the SST response to a horizontally uniform radiative forcing would be locally enhanced. The second mechanism in favor of an enhanced eastern Pacific warming invokes the slowdown of the equatorial trade winds that flattens the equatorial thermocline (Vecchi & Soden, 2007), much as during El Nino. The atmospheric overturning circulation slows down in a warmer climate as a result of a slower increase in global precipitation than atmospheric moisture (Held & Soden, 2006).

The SST pattern has an important effect on atmospheric convection in the tropics. With weak horizontal gradients in tropospheric warming, the change in convective instability – say measured by the difference in moist static energy between the surface and upper troposphere – follows the SST warming pattern. To the extent that tropospheric temperature is determined by the tropical-mean SST, rainfall increases where the
local ocean warming exceeds the tropical mean, and vice versa (Xie et al., 2010; Figure 1c). This is equivalent to saying that the SST threshold for deep convection will increase with the tropical mean SST (Johnson & Xie, 2010) and rainfall change follows the pattern of \(dSST^*\), where the asterisk denotes the SST change relative to the tropical mean warming.

The upwelling of cold water from a shallow thermocline keeps the eastern equatorial Pacific (e.g., Niño3 region) cool and below the convective threshold. In the current climate, only very strong El Niño events can push Niño3 SST above the threshold and cause substantive rainfall. Under greenhouse warming, the enhanced warming over Niño3 reduces the SST barrier to deep convection. Even if SST variability remains unchanged, extreme El Niño (defined as the total Niño3 SST exceeding the convective threshold) will become more frequent (Cai et al., 2014). While CMIP models do not agree how SST variability of El Niño-Southern Oscillation (ENSO) will change, a consensus emerged that convective variability over the eastern equatorial Pacific will increase, displacing the ENSO-induced teleconnection to the North Pacific and North America eastward (Zhou et al., 2014) with intensified ENSO influences on winter rainfall on the U.S. West Coast (Swain et al., 2018). Through atmospheric convective feedback in Niño3, the SST warming pattern further explains intermodel uncertainty in projected change in ENSO SST variability (Zheng et al., 2016).

While most CMIP6 models produce an El Niño-like SST pattern with enhanced eastern Pacific warming in the 1%/year CO₂ increase experiment (Figure 1), observations are equivocal on the SST pattern (Vecchi & Soden, 2007) because of large interdecadal variability, sparse sampling prior to the 1950s, and changes in measurement methods (e.g., from bucket, engine intake, to satellite) (Tokinaga et al., 2012). While mechanisms in favor of an El Niño-like warming seem to dominate in models, the mean upwelling in the eastern two-thirds of the equatorial Pacific damp the SST response to greenhouse warming, much as in the Southern Ocean. Seager et al. (2019) argued that this ocean upwelling damping limited the eastern Pacific warming in observations, although there are other ocean dynamic effects at work on equatorial warming (Coats & Karnauskas, 2018; Luo et al., 2017, and references therein). Uncertainties in aerosol effect and internal variability further complicate the interpretation of the observed SST pattern.

4. Global Climate Sensitivity

The globally integrated atmospheric energy budget may be cast as

\[
0 = R - \lambda T - D_O,
\]

where the overbar denotes the global mean and \(D_O\) the ocean heat uptake. We have linearized perturbation radiation at the top of the atmosphere (TOA) into a component due to composition change (e.g., CO₂) called the radiative forcing \(R\) and a component due to local surface temperature change \(T\) with \(\lambda\) called the radiative feedback parameter. Generally, \(R, \lambda, T,\) and \(D_O\) are all spatially variable. Specifically, let \(\lambda(x,y) = \lambda + \lambda^* (x,y)\) and \(T(x,y) = T[1 + P^* (x,y)]\). We assume that the surface warming pattern is scalable with global mean surface temperature (GMST), and \(P^*\) is the scaled pattern. By definition, \(\overline{P^*} = 0\) and \(\overline{\lambda^*} = 0\). Equation (1) becomes

\[
0 = R - \lambda_g T - D_O,
\]

where the global climate feedback parameter

\[
\lambda_g = \lambda + \overline{\lambda^* P^*}.
\]

When the ocean is fully equilibrated with the radiative forcing, \(D_O = 0\), and the equilibrium GMST is \(T_E \equiv \overline{R}/\lambda_g\). For a radiative forcing due to a doubling of atmospheric CO₂ (\(R_{2x} \approx 4\) W/m²), \(T_{2x} \equiv R_{2x}/\lambda_g\) is called the equilibrium climate sensitivity, which is inversely proportional to climate feedback parameter \(\lambda_g \approx 1.3\) W/m²/K for the multimodel ensemble mean).

Climate feedback is often decomposed into the components due directly to atmospheric temperature change, and indirectly to water vapor, cloud, and surface snow/ice. Local radiative damping \(\lambda\) is much larger...
in the tropics than polar regions (Armour et al., 2013). Equation (3) states that the global climate feedback is a function of the spatial covariance of the SST and local feedback patterns $\lambda P^*$. The global climate feedback is often treated as a constant that varies among models. Using experiments where atmospheric CO$_2$ is abruptly increased, recent studies found that in a given model, the global climate feedback increases considerably from the first few decades to a century into the integration (Armour et al., 2013). The ocean response to an abrupt CO$_2$ increase takes place in two stages. In the first decade or two, the ocean mixed layer (~100 m deep) warms up rapidly while the heat uptake by the subpolar North Atlantic and Southern Ocean slows the GMST increase. Following this fast-ocean response, GMST rises at much reduced rates, heated from below by the warming deep ocean. Because the subpolar North Atlantic and Southern Ocean switch roles from slowing to driving GMST increase, the warming pattern is more polar-amplified in the slow than fast ocean response (Long et al., 2014). Since the local radiative feedback is polar-amplified too, the covariance with the SST warming pattern and hence effective global climate feedback $\lambda_g$ are larger for the slow-ocean response.

The SST pattern effect on global climate feedback is illustrated by imposing patches of SST anomalies in atmospheric GCMs (Zhou et al., 2017). Over the warm ocean conducive to deep convection (e.g., the western tropical Pacific), surface warming is strongly damped by TOA infrared radiation as the tropospheric warming is upward amplified and spread throughout the tropics by equatorial waves (Andrews & Webb, 2018). Over a cool ocean with mean subsidence (e.g., the eastern subtropical Pacific off California), by contrast, a warm patch enjoys a positive cloud feedback as the weakened capping inversion reduces stratiform cloud cover in the atmospheric boundary layer. While Equation (1) assumes that TOA radiation is a function of local surface temperature change, Dong et al. (2019) gave a generalized formulation considering the nonlocal radiative response to SST forcing.

Efforts to estimate climate feedback from observations are complicated by the uncertainties in radiative forcing (e.g., aerosols) and due to unforced internal variability (Gregory et al., 2020). Let GMST change $T = T_F + T_I$, where the subscripts $F$ and $I$ denote radiatively forced and internal variability, respectively. Since internal variability is not energy driven and dominated by spatial variations, the global-mean TOA net radiation is not well correlated with GMST on multidecadal timescales (Xie et al., 2016). By neglecting the internal variability effect, the TOA radiation response to temperature change is approximately $N_T \approx -\lambda_g T_F$, and the apparent climate feedback becomes

$$\tilde{\lambda} \equiv -N_T/T = \lambda_g/(1+T_I/T_F).$$

This explains why atmospheric models consistently overestimate the global climate feedback parameter by a factor of 2 or more during the satellite era (Andrews et al., 2018), a period with a slowdown of GMST increase ($T_I < 0$) due in part to an internal cooling of the tropical Pacific (Fyfe et al., 2016; Kosaka & Xie, 2013). Alternatively, if the tropical Pacific cooling relative to the increasing GMST were intrinsic to the greenhouse warming (Seager et al., 2019), it implies a subdued global warming and low climate sensitivity. The debate highlights the crucial link between the pattern and magnitude of greenhouse warming.

5. Discussion

Huge infrastructure required to develop and run AOGCMs limited the access to global warming simulations to a few national modeling centers 30 years ago. CMIP (Meehl et al., 2007) flattens the play field by coordinating model experiments across national boundaries and making the output freely available. Anyone with a computer on the internet can access to and analyze CMIP output, resulting in an explosive growth in climate change analytics.

Despite coarse resolution – 8° longitude ×4.8° latitude with 9 vertical levels in the atmosphere and 3.75° × 4° with 12 vertical levels in the ocean – the model of Manabe et al. (1990) provided a first glimpse into regional patterns of climate change. Increased resolution improves the representation of extreme events. At 0.5° resolution, global models begin to simulate explicitly key statistics of tropical cyclones (e.g., basin counts and track density) with skills in reproducing observed interannual variability. Ocean surface warming pattern accounts for the intermodel uncertainty in basin counts (Yoshida et al., 2017; Zhao et al., 2009). Clouds
were prescribed in Manabe et al. (1990), but interactive clouds introduce large uncertainty in climate feedback among models (Soden et al., 2008), due in part to the ocean-warming pattern effect (Zhou et al., 2017).

While atmospheric GHG concentrations are nearly uniform in the horizontal, the ocean surface warming features rich spatial variations. The ocean warming pattern affects the atmospheric circulation and regional changes in rainfall as well as ENSO and the global climate sensitivity. In addition to the distribution of radiative forcing (GHG vs. aerosols), the ocean circulation (e.g., global overturning circulation and upwelling) is important in shaping the SST-warming pattern. Further research is needed to advance predictive understanding of the SST pattern and to quantify various mechanisms for pattern formation (Collins et al., 2018).

Data Availability Statement

CMIP6 output is available (https://esgf-node.llnl.gov/projects/cmip6/).

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