A study of feedbacks and the formation of climate trends in the Arctic climate system

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Abstract. The role of the ocean in the response of the climate system to an increase in the atmospheric CO₂ concentration is investigated by using a system of numerical models, ICMMG-PlaSim. The results of this study are summarized as follows: a) the ocean, to some extent (up to 20%), contributes to the increase in the annual mean state and to the decrease in the amplitude of seasonal oscillations (by 2-3%), which ultimately leads to insignificant changes in the summer period and to a significant mitigation of winter, b) the ocean stabilizes the annual mean state of the Arctic oscillation, making it practically unchanged with increasing CO₂ concentration but, at the same time, contributes to the significant increase in the amplitude of the seasonal cycle of this oscillation, c) the ocean enhances the temperature (or thermal) component of the seasonal variation associated with the appearance of additional areas freed from ice cover, with an additional average increase in the temperature of the atmosphere at the ice edge. Besides, the ocean enhances the seasonal oscillations of this component, so that the summer manifestations become much stronger, d) our tests have revealed that the role of the Arctic dipole under global warming is insignificant. These conclusions, though, may undergo significant changes under a more detailed consideration of carbon cycles in the atmosphere, ocean, and land.

1. Introduction

The problem of future climate change in the Arctic is associated with a number of scientific problems, as well as socioeconomic and environmental ones. The projected effects of climate change on the natural environment of the Arctic are large and can have a significant reverse impact on the global climate, which means the global importance of climate change in the Arctic [1]. Most of the supposed feedbacks that are formed due to climate change are concentrated in the Arctic and are still poorly understood. The most significant consequences are brought about by changes in the cryosphere, in particular, a reduction in the area and volume of floating ice. The specificity of the Arctic region: the atmospheric polar vortex, the three-phase boundary layer, radiation-cloud connections, the formation of the thermohaline structure of the Arctic Ocean and North Atlantic, and a number of other features make the Arctic an extremely complex and interesting object of study.

Much attention of scientists is given to improving forecasts of climate change on a time scale from decades to centuries based on scenarios of carbon emissions, sulfur emissions, changes in land use and...
other anthropogenic (technogenic) factors affecting the climate. Closely related to this are studies devoted to explaining the observed climate change by various types of impacts, both man-made and natural.

The most significant climate changes occur when the global energy balance between the incoming energy from the Sun and the outgoing heat from the Earth is disturbed. There are a number of natural mechanisms that can upset this balance, for example, fluctuations in the Earth’s orbit, changes in the circulation of the ocean, and changes in the composition of the Earth’s atmosphere. The latter, as is obvious, is a consequence not only of natural processes, but also of anthropogenic pollution caused by emissions of greenhouse gases and aerosols. Each year, the absorption of CO\(_2\) by the ocean is approximately equal to a quarter of what is released annually into the atmosphere as a result of human activity. This is primarily controlled by the difference between the CO\(_2\) concentrations in the atmosphere and the ocean with slight fluctuations from year to year and due to changes in the circulation of the ocean and in the biology of the Earth. Due to the fact that anthropogenic emissions of CO\(_2\) and other greenhouse gases continue to grow, and also because they can remain in the atmosphere for decades or centuries (depending on the gas), we will have a warmer climate in the future. The IPCC [2] predicts an increase in global temperature by 2-6°C by 2100, while the temperature in the polar regions may rise even faster.

Melting due to global warming is expected to reduce the size and extent of polar ice caps, even after allowance for the accumulation of more snow and ice on ice sheets due to increased precipitation. The melting of polar ice and ground glaciers is expected to contribute to an increase in the sea level of approximately 0.2–1.5 m predicted by the IPCC in the 21st century [2], and a sudden collapse of the West Antarctic ice sheet would raise the sea levels by 5–6 m due to further warming.

Recently, a number of assessments of climate impact on the polar regions have been carried out, including a summary report on snow, water, ice and permafrost in the Arctic [3, 4], on the climate of the Antarctic [5, 6]; the results of the International Polar Year are presented in [7]. According to these studies, the surface temperatures in the Arctic have risen substantially since the mid-20th century, and the warming rate exceeds the global one. In the recent decades, the sea-ice extent of the summer minimum has decreased significantly, and the Arctic Ocean is predicted to become almost ice-free in summer during this century. The duration of the snow cover and its depth are decreasing in North America and increasing in Eurasia. Since the late 1970s the permafrost temperature has increased by 0.5 to 2°C. Since the 1950s the oceans absorbed 93% of the excess heat in the Earth's system, which accumulated as a result of an increase in the concentration of greenhouse gases in the atmosphere. A significant increase in the heat content was observed in the upper 2000 m of the ocean since the 1960s, when the surface layers of the ocean became warmer by about 0.7±0.1°C from 1900 to 2016.

A reduction in the sea ice affects the energy balance due to changes in the surface albedo and as a result of a reduction in the insulating layer between the ocean and the atmosphere [8]. These manifestations mean that the loss of sea ice initiates a feedback, which directly contributes to the Arctic amplification of near-surface warming [9]. In addition, the loss of Arctic sea ice was associated with climatic changes in the lower latitudes through changes in the oceanic and atmospheric circulation [10]. For example, large-scale atmospheric circulation in winter due to a reduction in the sea ice in September is characterized by an increase in the meridional structure of the flow resembling the negative phase of Arctic Oscillations (AOs), which supports cold winters in northern Eurasia. Rossby waves with anomalously large amplitudes penetrate the stratosphere in February and weaken the stratospheric polar vortex, which subsequently gives rise to negative AO anomalies [11]. However, the remote influence is still difficult to interpret, and its mechanisms are still under investigation.

In this paper, we are going to find out the role of the ocean in shaping climate trends associated with an increase in greenhouse gas concentrations, using CO\(_2\) as an example, and whether the ocean is a deterrent or, conversely, it strengthens the atmospheric trends.

In our previous work [12], which studied atmospheric forcing trends and their role in reducing the Arctic ice based on EOF decomposition, we found that among the first non-degenerate EOF decomposition modes (Figures 1, 3, 5, and 7) the third mode is a temperature mode, that is, it is absent
during decomposition using only surface pressure and appears only when a temperature field of 10 m is included in the state vector. An analysis of the spatial pattern of this EOF mode shows that its region of action coincides with zones of additional opening from the ice cover in the recent decades in summer and, therefore, it possibly arises as a result of the climatic trend. It would be interesting to find out the further role of this mode and of all the others in the formation of atmospheric and ocean dynamics in the 21st century.

2. Estimation of radiation time scales

If we assume that radiative transfer is the only process occurring in the Earth’s atmosphere, then the surface temperature will be determined solely by the total radiation in the upper part of the atmosphere. There are two fundamental time scales that determine how the atmosphere transfers heat. The first one is associated with atmosphere dynamics. The other one is the radiation time scale. It is the time taken by the atmosphere to reduce its temperature perturbation e-fold from the original value using radiation.

2.1. Atmosphere and terrain

To get started, we get a simple measure of the planetary radiation time scale. Consider an atmospheric column with surface pressure $p_s$. If there is no solar heating, the atmosphere will be cooled by thermal radiation. Assume that the entire column is initially in radiation equilibrium. Then at the terrain surface we have

$$ (1 - \alpha) S_0/4 + B = U, \quad (1) $$

where the first term is the mean incoming solar radiation ($S_0$ is the solar constant, $\alpha$ is the surface albedo), the second term is downward atmosphere thermal radiation, $U = \sigma T_e^4$ is the surface thermal radiation ($T_e$ is the surface temperature). For a column of atmospheric air, under the condition of hydrostatics, one can write the equation of the energy balance depending on time (see, for example, [13])

$$ \frac{c_p p_s}{g} \frac{dT_u}{dt} = \varepsilon U - 2B, \quad (2) $$

where $\varepsilon$ is the optical transmissivity of the atmosphere. Suppose that the temperature of the atmosphere can be represented as $T_u = T_{u_0} + \Delta T$, assuming that the temperature deviation from the temperature of radiation equilibrium is small. Substituting (1) into (2) to express $U$ and using the Taylor form in a first approximation, we obtain the following equation:

$$ \frac{c_p p_s}{g} \frac{d\Delta T}{dt} = \varepsilon \sigma T_e^4 - (\varepsilon - 2) \varepsilon \sigma (T_{u_0})^4 - 4(\varepsilon - 2) \varepsilon \sigma (T_{u_0})^3 \Delta T. \quad (3) $$

The first two terms in the right-hand side of (3) represent the equilibrium state and, therefore, give a total of zero. Finally, we have

$$ \frac{d\Delta T}{dt} = - \frac{4(\varepsilon - 2) \varepsilon \sigma g (T_{u_0})^3 \Delta T}{c_p p_s} \quad (4) $$

giving exponentially vanishing temperature perturbation with the time scale of the e-fold decrease,

$$ \tau_{rad} = \frac{c_p p_s}{4(\varepsilon - 2) \varepsilon \sigma g (T_{u_0})^3} \quad (5) $$
Assuming that $T_{ao} = 300\text{K}$, $\varepsilon = 0.6$, and $c_p = 1000 \frac{\text{J}}{(\text{K kg})^{-1}}$, the scale of radiation equilibrium in (5) will be about 18 days.

2.2. Upper ocean
A simple energy balance of the system, taking into account the heat capacity of the upper ocean layer, can be written as

$$C_s \frac{dT_s}{dt} = \Delta Q + \lambda_f \Delta T_s,$$

where $\Delta T_s$ is the sea surface temperature perturbation, $\Delta Q$ is the radiation forcing, $C_s$ is the heat capacity of the upper ocean layer (for the 200-m layer it is approximately equal to $8.36 \times 10^8 \frac{\text{J}}{\text{K m}^2}$), and $\lambda_f$ is the feedback parameter (approximately equal to $-1.0 \frac{\text{W}}{\text{K m}^2}$). Thus, an estimate of the time scale for the transition to an almost equilibrium state $\tau_{srad} = -C_s \cdot \lambda_f^{-1}$ leads to a value of approximately 22 years.

3. Description of the numerical model and testing experiments
The purpose of the present research is to elucidate the role of the oceanic and ice response to climate changes in the atmosphere and on land in the further development of these changes. According to the IPCC scenarios [2] we will associate climate change with an increase in the atmospheric CO$_2$. A similar approach is used quite often when trying to reproduce the features of climate change, in particular, to assess the sensitivity of climate models to an increase in the CO$_2$ concentration in the atmosphere (see, for example, [14] and the references therein).

Generally speaking, of the atmospheric gases water vapor is the dominant greenhouse gas. If H$_2$O were the only greenhouse gas present, the greenhouse effect of a clear-sky atmosphere in the mid-latitudes, measured by the difference between the infrared flux on the surface and in the upper atmosphere, would be about 60-70% of this value taking into account all gases. If only CO$_2$ were present, the corresponding value would be about 25%. Although, due to the overlap between the absorption bands of different gases, the calculation of percentages is not strictly additive.

The additional heat fluxes in 2025 relative to the pre-industrial value calculated according to the IPCC scenario A (“business-as-usual case”), as well as their percentage contribution for various greenhouse gases, are as follows: CO$_2$ - 2.9 Wm$^{-2}$ (63%); CH$_4$ - 0.7 Wm$^{-2}$ (15%); N$_2$O - 0.2 Wm$^{-2}$ (4%); CFCs and HCFCs - 0.5 Wm$^{-2}$ (11%); stratospheric H$_2$O - 0.2 Wm$^{-2}$ (5%). That is, the total additional heat flux is approximately 4.6 Wm$^{-2}$, which corresponds to the effective amount of CO$_2$ approximately twice the pre-industrial value.

As a research method we will use numerical modeling, and analyze the results of numerical experiments.

3.1. ICMMG-PlaSim climate system model
The PlaSim-ICMMG [15] system model was developed as a modular structure that allows one to create a range of models of the Earth system of intermediate complexity by choosing various options to describe the various components of the climate and carbon cycle. The model is able to integrate at different time scales. The modularity of its structure facilitates the connection of complex components with an increase in the computing power.

The system is based on the original version of the PlaSim model [16], in which the original modules for calculating the characteristics of the ocean and sea ice were replaced with more detailed models of the World Ocean (model SibCIOM [17, 18]) and ice dynamics (model CICE-3 [19, 20]). As part of the multi-modularity, an additional opportunity was realized, in comparison with PlaSim, to use an individual domain grid for each of the modules. An atmospheric module using the spectral approach is more convenient to describe in a spherical coordinate system (longitude, latitude) on isobaric surfaces, while it is undesirable for an oceanic module to have a singularity in the region of the north pole associated with these coordinates and, therefore, a three-pole coordinate system is used.
[21] with poles shifted to the continents. In the ice module, it is undesirable to use a grid covering the entire oceans, it is sufficient to confine the model domain to the subpolar and polar regions, discarding latitudes where there is almost no ice, and instead improve the resolution in those regions where ice-snow cover is really observed.

The interaction of the system modules is carried out by a separate process, a coupler, in the framework of a multitasking parallel system (MPI), which provides the gathering of necessary information about the state of each of the modules at each time step and the scattering of the forcing fields for all components of the system on a spatial grid that is convenient for each module. The coupler allows the replacement of one module with another. In particular, we will take advantage of the fact that the real numerical model can be replaced by the so-called data model, that is, a module that is not a model of any component. Such a module emulates the operation of this component, and it is a program that sends the appropriate data set to the coupler at the right time, for example, read from files generated from observational data or from the results of other numerical experiments.

The atmospheric module of the system sets the atmospheric average CO$_2$ concentration as a parameter. This parameter plays an important role in calculating the optical properties of the atmosphere and in the formation of radiation fluxes. We will use this opportunity in numerical experiments simulating an increase in atmospheric CO$_2$ concentration similar to IPCC scenarios.

3.2. Numerical tests

In order to identify the potential role of the ocean and sea ice in the formation of climate trends resulting from atmospheric effects, the following series of numerical experiments with the PlaSim-ICMMG climate model is proposed:

- **test A**: Integration of the model at a fixed level of CO$_2$ concentration, approximately equal to the value in the mid-90s, 360 ppmv, with the full interaction of all modules before reaching the quasi-stationary mode, when most of the variability is reduced to seasonal changes and when every next year approximately repeats the previous one, it was about a hundred years, and in the future we will refer to this period of time as 1900-2000.

- **test B**: Restart of the complete system run from the quasi-stationary state of test A, with an abrupt increase in the CO$_2$ concentration value to the level of 450 ppmv. It seems to be a very close perspective of future changes in the atmosphere conditions since in May 2019 a CO$_2$ concentration of 415.26 ppmv was recorded at the Mauna Loa Observatory in Hawaii by researchers from the Scripps Institution of Oceanography. The value of 450 ppmv roughly corresponds to the mildest of the IPCC scenarios, RCP 2.6. This test is to show the reaction of the simulated climate system to a sudden change in the CO$_2$ concentration in the atmosphere.

- **test C**: Subsequent integration of the system, starting from the state of quasi-stationary state of test A, with an abrupt increase in the CO$_2$ concentration to 450 ppmv as in test B, but the ocean-ice system continues to reproduce the quasi-stationary state, which was recorded during the last year cycle of test A. This test is to show the reaction of the incomplete climate system to a sudden change in the CO$_2$ concentration in the atmosphere, the ocean and ice responses to this change are excluded from this test.

The time in tests B and C will be conditionally counted from 2000, the time of completion of the test A.

Thus, comparing the results of the three experiments, we can evaluate what processes arise from an increase in CO$_2$ in the coupled system, and from a comparison of the last two we can assess whether the ocean with sea ice (hereafter referred to as the ocean) enhances or weakens emerging trends, and is also possibly responsible for the formation of some of them. This study considers only the fast mode of atmosphere response, which is about 20 days’ time scale, but we can also trace the first tendencies of the slower mode associated with the upper ocean response. It should also be borne in mind that in this study we do not consider changes in the absorption of carbon dioxide by the ocean, believing that in a first approximation this process can be neglected.
4. Results

In [12], using the EOF analysis of atmospheric forcing, it was found that in the case where the state vector includes the surface pressure, temperature, and horizontal velocity components at a level of 10 m and in the absence of seasonal filtering, four non-degenerate modes can be distinguished: the first is associated with seasonal variations, the second with the AO phase [22], the third is the temperature mode associated with the introduction of the surface temperature field into the state vector, and the fourth representing the so-called Arctic dipole [23].

Figure 1 shows the structural function of the first mode in the form of a distribution of the surface pressure field and temperature at a height of 10 m. It can be seen that with the onset of the positive phase the temperature over the continents increases, and in the case of the negative phase, on the contrary. Changes over the oceans are minimal, but it can be seen that the temperature in the surface layer of the Arctic Ocean behaves more similarly to the continental part of the region. A similar picture can be seen in the surface pressure field; however, in contrast to the temperature above the ocean, significant variations are found over the oceans (except for the Arctic Ocean) acting in antiphase to variations over the continents.

![Figure 1](image1.png)

**Figure 1.** The structural function of the first mode presented as a distribution of the surface pressure field (a) and temperature at a height of 10 m (b).

![Figure 2](image2.png)

**Figure 2.** PC time series of the first EOF mode of decomposition of atmospheric forcing: (a) annual average values, (b) seasonal cycle averaged over years, (c) amplitude of seasonal cycle, (d) standard deviation from the seasonal variation. The black line is the results of test A, the blue one is test C, and the red one is test B. The blue arrow shows a jump in the transition to the mode of increased CO₂ concentration, and the red ones emphasize the correction introduced by the presence of active ocean.
When analyzing the trends of the first mode from 1980 to 2010, [12] noted an increase in the average value of the PC (principal component) time coefficient and a simultaneous decrease in the amplitude of the seasonal cycle, which in terms of temperature means the trend of warmer winters, with a relatively stable summer period. The results of tests A, B, and C are presented in Figure 2 as annual average values of the PC time series for the first EOF mode of the state vector. It can be seen that the trends associated with an increase in atmospheric CO2 concentration coincide with those noted in [12]. The average value of the time-depending coefficient of EOF at the end of test A turned out to be near zero, and with increasing concentration it increased noticeably in test C. In the case of inclusion of an active response from the ocean (test B in comparison with test C), this trend even increases somewhat (up to 20%). According to the results of [12], the seasonal variability of the first mode (Figure 2b) describes about 99% of the total variability of this mode. The amplitude of the seasonal cycle changes in the direction of its decrease (Figure 2c), and the jump is about 10% of its value, with the ocean adding another 2-3%. In terms of the values of the decomposition coefficient, a decrease from the level of 3200 is by 300-400 units, while the increase in the average annual value is also in this range. Thus, as before, we state that the summer state has changed insignificantly, and the winter one has noticeably approached the summer one. The standard deviation from the seasonal cycle of Figure 2b with an amplitude of Figure 2c is shown in Figure 2d, from which it can be seen that in the case of a fixed ocean the deviations from the seasonal course decrease by almost 2 times with an increase in CO2; however, in the case of an actively interacting ocean they remain approximately at the same level. Thus, the ocean supports intra-seasonal variability.

For the second mode (Figure 3), the previously noted trends of the period 1980–2010 consisted in a certain decrease in the average annual value of the time-dependent coefficient with a small increase in the amplitude of seasonal fluctuations, while the seasonal variability itself describes only the 20% variability of the second mode. The first means the strengthening of the positive phase of the AO, which was actually observed in the 90s [24]. However, Figure 4 shows that the atmosphere under the condition of a fixed ocean demonstrates the opposite tendency of an increase in the average annual value and a weakening of the AO, while the ocean strengthens the positive phase of this oscillation. In the classical view, the AO is the first mode of the EOF decomposition of the pressure field averaged during the period of January-March (or December-February). If atmospheric forcing used in [12] is used to construct the eigenfunction of the AO, then having obtained it and comparing it with our modes, one can notice that the second mode is 70% correlated with the eigenfunction of the AO, but at
the same time explains only 6% of the variability of atmospheric forcing, while the first mode correlates with the AO only by 5%, but at the same time explains 76% of the variability. Thus, the insignificant tendency of the average state of the second mode toward the amplification of the AO is excessively compensated by the reverse action from the trends of the first mode. As a result, the first two modes give a negative AO trend, which was also noted earlier in [11].

Figure 4. Same as Figure 2, but for the second mode.

Figure 5. The structural function of the third mode presented as a distribution of the surface pressure field (a) and temperature at a height of 10 m (b).

In Figure 4c one can also see that the amplitude of the seasonal cycle increased by almost 1.5 times, making the difference between the winter minimum (AO gain) and the summer maximum (AO attenuation) more significant. It is important that the ocean contributes to an increase in the amplitude by an order of magnitude greater than a decrease in the average state of this mode. The interseasonal variability, presented as the standard deviation from the seasonal variation in Figure 4d, as in the case
of the first mode, is somewhat weakened in test C, but due to the ocean reaction in test B it turns out even stronger than in test A.

A feature of the third mode of the EOF decomposition is that its structure in the temperature field identifies zones additionally released from ice in summer due to climatic changes: these are the area of the Barents and Kara Seas, the vicinity of the Bering Strait and the straits of the Canadian Archipelago (Figure 5). For the rest, you can see that the oceans and continents behave in different directions, but unlike the first fashion, the Arctic Ocean behaves like an ocean, not land. This is also a consequence of climate change, as a result of which a significant part of the Arctic Ocean in summer opens up for active interaction with the atmosphere.

Along with an increase in the CO₂ concentration, there is a sharp increase in the average annual value of the time-dependent coefficient of this mode, up to 30% compared to the amplitude of the annual cycle (Figure 6a), and the role of the ocean in this increase is positive but insignificant. The amplitude of the seasonal cycle increases in test C by about 15%, and in test B by 25% (Figure 6c). The seasonal cycle accounts for 45% of the total variability of this mode, which makes it second after the first mode in sensitivity to the season. Thus, the ocean plays an important role in the fact that in summer there is a significant decrease in the area of ice, and in winter it is more actively restored.

The Arctic dipole, represented by the fourth mode of the EOF decomposition (Figure 7), is formed as a result of opposite pressure changes over Greenland on the one hand, and over the Kara Sea and the Taimyr Peninsula on the other. In the positive phase, when the maximum pressure is over the Taimyr Peninsula, the prevailing winds counteract the transfer of ice to the North Atlantic, and vice versa, intensify the transpolar drift in the negative phase. In the negative phase, one can also note the warming of the coastal regions of the Arctic Ocean, the Baffin Sea, and the Hudson Bay.

The trends revealed by changes in the carbon concentration in the atmosphere are shown in Figure 8. The change in the average annual values of the decomposition coefficient for this mode (Figure 8a) in test C does not exceed the value of fluctuations of this value in test A. However, the inclusion of the active ocean increases the deviation of this value to the positive side by about 2 times, and then relaxes downward. Seasonality (the annual cycle) (Figure 8b) is weakly expressed and accounts for only 13% of the total mode variability. An increase in the CO₂ concentration leads to an almost 2-fold increase in the amplitude of the seasonal cycle in test C. However, in test B the ocean reduces the amplitude of seasonal changes (Figure 8c) back to unperturbed values. Deviations from the seasonal cycle for this
mode (Figure 8d) vary over a wide range and, although an increase in CO₂ in the atmosphere leads to a slight decrease in this deviation, in general, this decrease is in the range of usual changes of this value. Thus, the conclusion on the increasing role of the Arctic dipole during global warming made in [23] was not confirmed in the course of our experiments, which may mean, for a number of different reasons, that its strengthening is not directly related to an increase in the concentration of greenhouse gases in the atmosphere.

Figure 7. The structural function of the fourth mode presented as a distribution of the surface pressure field (a) and temperature at a height of 10 m (b).

Figure 8. Same as Figure 2, but for the fourth mode.

5. Conclusions
Summarizing the results of the numerical experiments, we can state that the role of the ocean in the response of the climate system to an increase in the atmospheric CO₂ concentration, to a first approximation, is as follows:

- the ocean contributes to some extent (up to 20%) to the increase in the annual mean state of the seasonal cycle and to the decrease in the amplitude of seasonal oscillations (by 2-
3\%), which ultimately leads to insignificant changes in the summer period and to a significant mitigation of the winter;

- the ocean reduces the changes in the annual mean state of the Arctic oscillation, making it practically stable with increasing CO\textsubscript{2} concentration but, at the same time, contributes to the significant increase in the amplitude of the seasonal cycle of this oscillation;
- the ocean enhances the temperature (or thermal) component of the seasonal variation associated with the appearance of additional areas freed from ice cover, with an additional average increase in the temperature of the atmosphere at the ice edge. Besides, the ocean enhances the seasonal oscillations of this component, so that the summer manifestations become much stronger;
- the role of the Arctic dipole under global warming is not significant, although the results of [23] show that this mode cannot be disregarded, since it can promote or restrict the Arctic ice export.

Thus, the direct response is that the ocean mainly enhances the trends resulting from an increase in CO\textsubscript{2} for the first three modes of the EOF decomposition. The third “temperature” mode actually arises due to the formation of a vast area free of ice.

However, these conclusions may undergo significant changes under a more detailed consideration of carbon cycles in the atmosphere, ocean, and land.

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