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### Uncertainty in the 2 °C Warming Threshold Related to Climate Sensitivity and Climate Feedback

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ABSTRACT

Climate sensitivity is an important index that measures the relationship between the increase of Greenhouse Gasses and the magnitude of global warming. Uncertainties in the climate change projection and climate modeling are mostly related to the climate sensitivity. The climate sensitivities of coupled climate models determine the magnitudes of the projected global warming. In this paper, the authors have thoroughly reviewed the studies of climate sensitivities, and discussed the issues related to climate feedback processes, the methods used in estimating the equilibrium climate sensitivity (ECS) and transient climate response (TCR), including the TCR to cumulative CO2 emissions (TCRE). After presenting a summary of the sources that affect the uncertainty of climate sensitivity, the impact of climate sensitivity on climate change projection is discussed by addressing the uncertainties in 2 °C warming. Challenges that call for further investigations to the community in particular the Chinese community are discussed.

Key words: climate sensitivity, radiative forcing, climate feedback, 2 °C threshold, greenhouse gasses, climate model
气候敏感度、气候反馈过程与 2 °C 升温阈值的不确定性问题

ABSTRACT

气候敏感度是度量温室气体浓度升高和全球升温幅度关系的重要指标，当前气候模拟和气候预估中的很多不确定性问题，都直接和气候敏感度有关。气候敏感度的大小也决定着预估的气候变暖幅度的大小，直接影响到温室气体减排政策的制订。在简要回顾气候敏感度概念的提出和研究历史基础上，着眼于气候反馈分析，介绍了气候敏感度与辐射强迫和反馈过程的关系，总结了气候系统主要的反馈过程；根据大气层顶的能量平衡关系，利用 CMIP5 多模式结果介绍了平衡态气候敏感度和瞬态气候响应（包括累积碳排放的瞬态气候响应）的估算原理和方法，总结了气候敏感度不确定性的来源，并以“2 °C 阈值”问题为例，介绍了气候敏感度对预估结果不确定性的影响。随着观测资料的积累和气候模式的发展，继续减少气候敏感度的不确定性，估算包含碳循环的敏感度，利用地球系统模式规划最优碳排放路径是未来本领域主要的研究方向。

Key words: 气候敏感度, 辐射强迫, 气候反馈, 2 °C 阈值, 温室气体, 气候模式
1. Concept of climate sensitivity

The increasing of green-house gases, represented by CO$_2$, blocks part of the outgoing infrared radiation from the atmosphere-earth system (viz. green-house effect). Consequently, unbalanced downward radiative energy at the top of atmosphere (TOA) enter and heat the climate system. In a broad sense, climate sensitivity can be regarded as how fast and strong the climate system responses to such kind of net heating. Global mean surface air temperature ($T_s$) that is closely related to the radiation, is usually used as the climate response index. When the climate of the Earth fully responds under doubled pre-industrial CO$_2$ concentration and reaches to a new equilibrium state, the change of $T_s$ relative to the pre-industrial baseline is defined as “Equilibrium Climate Sensitivity (ECS)”, briefly called “climate sensitivity” in practice (Cubasch et al., 2001; Randall et al., 2007; Wang et al., 2012).

The study on climate sensitivity can be traced back to the end of the 19$^{th}$ century. Based on simple equilibrium thermal radiation theory, Arrhenius (1896) quantitatively explored the green-house effect of CO$_2$ at the earliest time and obtained a 4.4 K warming of $T_s$ under a doubled CO$_2$ concentration when the effects of vapor and snow-ice albedo were considered. The development of radiation theory and the birth of computer after the 1950s provided more powerful tools for climate research community. A one-dimensional thermal equilibrium model with convective adjustment was developed and used to study the sensitivity of surface temperature to radiative forcing induced by various factors (Manabe et al., 1964). Assuming an unchanged relative humidity, water vapors can double the $T_s$ increment purely caused by the CO$_2$ forcing (Manabe et al., 1967). One decade later, three-dimensional atmospheric general circulation model (AGCM) had been established and showed about 3 K surface warming under a doubled CO$_2$ with ideal land-sea distribution, fixed cloud cover and ignored ocean heat transport (Manabe et al., 1975). Charney et al. (1979) performed a set of coupled simulations with multi-AGCM coupled to mixed-layer ocean (or called slab ocean) and provided a possible range of climate
sensitivity within 1.5–4.5 K (so-called “Charney sentivivity”). After Charney
sensitivity was published, more sophisticated AGCMs coupled to slab ocean were
developed, including changeable cloud, prescribed ocean heat transport, higher
resolution etc. However, the climate sensitivity of these models changed little
compared to the Charney sensitivity (IPCC, 1990).

The coupled AGCM/slab-ocean system is a popular tool in climate sensitivity
research for its cheap computation cost. The equilibrium state can be reached in
20–30 years after the forcing is imposed. Nevertheless, it may be not appropriate to
focus on long-term climate change, since the change of ocean heat transport is omitted,
such as the responses of the thermohaline circulation or more specifically the Atlantic
Meridional Overturning Circulation (AMOC; Zhou et al., 1998, 2000). With the
improvement of computer capacity, from the end of the 1990s fully coupled
atmosphere-ocean models have been used for long-term climate integrations (Stouffer
et al., 1999; Gregory et al., 2004; Danabasoglu et al., 2009; Li et al., 2013), however,
these kinds of fully coupled models were still too expensive for most modeling groups.
Therefore, an analysis based on transient response (viz. non-equilibrium) was
proposed to estimate the warming magnitude in the equilibrium state (Gregory et al.,
2004; Flato et al., 2013). The method was designed to use only a slightly longer than
100-year integration with fully coupled models to estimate the ECS. Compared with
the result from equilibrium simulation, the bias in the estimation of transient response
is with 10% and thus can be accepted (Li et al., 2013).

The focus of ECS is the final equilibrium state, regardless of how to reach it.
However, both the historical climate changes and future changes in different scenarios
are transient responses and cannot be regarded as equilibrium state. Thus, a new term,
Transient Climate Response (TCR), is introduced to describe the sensitivity of the
climate response in the transient state. It is defined as the change of $T_s$ relative to the
pre-industrial when the $CO_2$ concentration is doubled at an increasing rate of 1% per
year (Randall et al., 2007).

Recently, it is found that the linear relation between the $T_s$ change and the
cumulated CO$_2$ concentration in the atmosphere changes little with time and scenarios and reflects the time scale of warming from decadal to more than century (Matthews et al., 2009; Goodwin et al., 2015). A new sensitivity index, TCR to cumulative carbon emissions (TCRE), is defined as the change of $T_s$ by one unit cumulative carbon emission. The TCRE can provide the cumulative carbon emission in the atmosphere given a certain threshold of $T_s$ change (Collins et al., 2013). It is a bridge connecting the targets of controlling temperature and reducing carbon emission. Thus it is an important reference index for making emission reduction policies.

Climate sensitivity has received great attention because of its importance in describing the relation between increasing GHG concentration and the magnitude of global warming. The estimate of climate sensitivity has direct influence on the reliability of the projected further climate change and is thus a useful reference for policymakers in making decisions relevant to the reduction of carbon emission. As a review, the major objective of the paper is to summarize the recent progress in the studies of climate sensitivity, radiative forcing and feedback processes. We further discuss the source of uncertainty in the estimate of climate sensitivity and its influence on the projection with 2 °C warming threshold. Future research priorities related to climate sensitivity are also discussed and recommended.

2. Climate feedback

By introducing several terminologies in electronic engineering field, Hansen et al. (1984) proposed a linear feedback analysis which gradually became a standard method (Cubasch et al., 2001; Gregory et al., 2002, 2004; Roe 2009). After that, the issue of “forcing – response – feedback” (Fig. 1) was paid more attention. “Forcing” is the driver of a system evolving, that is the radiation perturbation at the TOA in terms of climate system, caused by various factors such as solar radiation change, aerosols emitted by natural process like volcanos, GHGs and aerosols emitted by anthropogenic activities. Besides the CO$_2$, the anthropogenic GHGs include other tracer gases, some of which can induce stronger radiative effect than CO$_2$ by one unit change in concentration (Shi 1991; Wang et al., 2000). For convenient calculation, the
concentration of all the GHG species is usually converted to the CO$_2$ that can induce the same radiative forcing, called “equivalent CO$_2$ concentration”.

Fig. 1. Schematics of the “forcing-response-feedback” relation. $\lambda_X$, the feedback parameter for a certain feedback factor, means the induced perturbation of the radiative forcing at the TOA by the changes of the feedback factor per global mean surface temperature changes.

Idealized radiative forcing is the net flux at the TOA without any response anywhere after the forcing agents were imposed. In such a way, the doubled CO$_2$ concentration corresponds to about 4.37 W m$^{-2}$ (Ramaswamy et al., 2001). With the development of research, climate scientists began to know that the forcing responsible for the change of $T_s$ is the unbalanced radiation after rapid adjustments. For example, the stratosphere can adjust to radiative equilibrium within one month, whereas the more concerned changes in the troposphere are slower which is caused by the forcing after stratosphere adjustment. Taking this into consideration, the Intergovernmental Panel on Climate Change (IPCC) modified the definition of radiative forcing in the third assessment report (TAR) and the doubled CO$_2$ forcing is revised to 3.71 W m$^{-2}$ (Ramaswamy et al., 2001). Although the above concept of forcing was reserved, other methods to calculate forcing was proposed in the fourth IPCC report (AR4; Forster et
including more rapid processes (but slower than the stratosphere adjustment), such as the aerosol-related cloud changes (Jacob et al., 2005) and temperature adjustment in the troposphere (Hansen et al., 2005).

Multiple definitions of radiative forcing are summarized in the fifth IPCC report (AR5), including the old definitions shown above. A new definition called Effective Radiative Forcing (ERF) is proposed. The ERF is the doubled CO$_2$ forcing at the TOA but considering all kinds of rapid adjustments, including temperature changes in the troposphere and on land, aerosol-cloud interaction, changes in vertical structure of temperature and its effect on cloud (Myhre et al., 2013). The time scale of temperature change we concerned is longer than decades to century. Hence, the ERF can better represent the forcing agents which can impact the temperature change at longer time scales (Zhang et al., 2014). Because the understanding of the rapid adjustments is not sufficient, a large uncertainty is observed in the model-based estimation of the ERF (2.6–4.3 W m$^{-2}$; Flato et al., 2013).

The ultimate magnitude of the response is not only determined by the forcing, but also strongly influenced by various feedback processes. Stronger the positive feedback leads to higher climate sensitivity, and vice versa. The “forcing – response – feedback” represents a cyclic interaction to a new equilibrium state (Fig. 1). We further present a brief review of the main feedbacks that are recognized up to now.

2.1 Planck feedback

Based on the Stefan–Boltzmann law, the surface heated by the radiative forcing will emit more infrared energy outwards and reduce the net flux at the TOA. This basic negative feedback is called “Black-body radiation feedback” or “Planck feedback”. Some studies suggested that this process can be used as a reference system to measure other feedbacks rather than a solo “feedback” because it is the simplest and well-established relation between temperature and radiation (Roe, 2009). When contributions of different processes to the change of $T_s$ are focused without reference system, the cooling effect of Planck radiation can be regarded as an important feedback (Gregory et al., 2008; Chen et al., 2014; Pithan et al., 2014).
2.2 Water vapor feedback

Water vapor is the most important GHG that exerts the strongest warming effect. Based on the Clausius–Clapeyron relation, the water vapor in the atmosphere strictly depends on the temperature. Considering the short period of atmospheric hydrological cycle (about 10 d), the water vapor should be treated as “feedback” rather than “forcing”. The increased $T_s$ induced by external forcing will enhance surface evaporation and hold more water vapor in the air. More water vapor will block more outgoing radiation and increase the forcing at the TOA, which is the well-known “water vapor feedback” (Held et al., 2000).

2.3 Lapse-rate feedback

The lapse rate of tropospheric temperature will change when climate system warms. In the tropical regions, more water vapor condenses at the middle-to-upper troposphere and heats the local atmosphere, resulting in the warming in the upper layer stronger than the lower layer. This process is called “moist adiabatic adjustment” in which the moist adiabatic lapse rate decreases. The warmer upper layer is in favor of emitting more infrared radiation to the space and reducing the forcing at the TOA. That is the negative lapse-rate feedback. The warmer regions in the troposphere are usually filled with more water vapor, especially in the tropics. As a result, the positive water vapor feedback and negative lapse-rate feedback can partly cancel out and the net feedback is still positive (Cess, 1975; Held et al., 2000; Soden et al., 2006). Thus the two closely related feedbacks are unified into “water vapor – lapse rate feedback”. However, if we choose relative humidity instead of specific humidity as feedback agent, the compensation will be substantially reduced (Held et al., 2012; Ingram, 2013). In the mid-high latitude regions, warming is confined in the lower layer for a lack of moist adiabatic adjustment. The consequently positive lapse rate feedback mainly contributes to the “polar amplification” phenomenon (Colman, 2003; Pithan et al., 2014).

2.4 Snow-ice albedo feedback

The snow cover and sea ice in the high latitudes can rapidly respond to the
surface warming. Melted snow and ice decrease the surface albedo and shortwave reflected back to the space, and increase the forcing at the TOA. That is the positive snow-ice albedo feedback, which is one of the main contributors to the polar amplification (Pithan et al., 2014), as shown in the earliest study on the climate sensitivity (Arrhenius, 1986).

2.5 Cloud feedback

The cloud response is very complex under the background of climate warming. A variety of cloud parameters, such as cloud fraction, height, particle size, phase etc., all can impact the radiative flux at the TOA. One change of cloud attribute may bring about both the positive and negative feedbacks at the same time. If cloud fraction decreases with the surface warming, increased outgoing longwave (incident shortwave) will reduce (amplify) the TOA forcing, acting as a negative (positive) feedback. Then the sign of net cloud feedback is difficult to determine. The cloud in different altitude has different radiative effects. High cloud is more opaque to longwave, whereas low cloud mainly reflects shortwave. As a result, both the increase in high cloud and decrease in low cloud induced by surface warming can lead to positive feedback. In the tropics, the cloud top rises with the tropopause as the tropospheric temperature increases, resulting in positive feedback by reducing the emitting longwave. Under global warming, the storm track shift poleward with the expansion of the Hadley circulation. As a result, the area covered by frontal cloud is reduced and heated by more solar radiation, acting as a positive feedback. Most of the models used in IPCC AR5 show that the net cloud feedback may be positive (Boucher et al., 2013).

2.6 Other feedbacks

If the air-sea CO₂ exchange is considered, more CO₂ will be released into the atmosphere from the warmer ocean by reducing solubility and increase the radiative forcing at the TOA. That is the positive solubility feedback. Feedbacks become more complex when the changes of biosphere are involved. The response of vegetation cover to the warming could change the land albedo and produce “vegetation – albedo
feedback” (Zeng et al., 2009). Another example, ocean acidification due to CO$_2$ uptake may decrease the emission of dimethylsulphide (DMS) by marine organism which is the largest natural source of atmospheric sulphur. The decrease in atmospheric sulfate will affect the cloud formation and ultimately the radiative budget (Six et al., 2013). Therefore, in addition to the physical feedback mentioned above, the feedbacks involving the biogeochemical processes are also important. Thus the development of Earth system models that include biogeochemical cycles is the forefront of climate modeling community (Zhou et al., 2014).

The “forcing – response – feedback” relation describing the physical responses can be linearly expressed as

$$\lambda_X = \frac{\partial R}{\partial X} \frac{dX}{dT} = K_X \frac{dX}{dT}$$  \hspace{1cm} (1)

where $X$ is a certain feedback agent, such as water vapor; $R$ is the radiative forcing at the TOA; $T$ is the global mean surface air temperature; $K_X$ is called “Feedback Kernel” which describes the contribution of one unit change in $X$ to the $R$ and only depends on radiative transfer process; $(dX/dT)$ is the response of $X$ to surface warming; $\lambda_X$ is the feedback parameter in respect to $X$.

For different feedback agents $X$, how to calculate the corresponding $\lambda_X$ based on equation (1) is the central issue of feedback analysis. Readers are suggested to refer to Roe (2009) and Soden et al. (2008) for more detailed information on the feedback analysis method and “Radiative Kernel” approach to calculate the $\lambda_X$.

3. Principle of estimating climate sensitivity

The estimation of climate sensitivity is based on energy conservation,

$$N = F + E,$$  \hspace{1cm} (2)

where $N$ is net radiative flux at the TOA; $F$ is radiative forcing exerted by forcing agent; $E$ is the increased outgoing radiation after the atmosphere-earth system responds; positive direction is downward. To the first-order approximation, $E$ is expressed as the linear function in respect to the change of global mean surface air temperature $T'$,

$$E = \lambda' T',$$  \hspace{1cm} (3)
where \( \lambda \) is the net feedback parameter, the sum of all feedback components. The equations (2) and (3) are the fundamental of estimating climate sensitivity based on either observation or simulation.

### 3.1 ECS

Under fixed 2×CO\(_2\) concentration, fully coupled ocean-atmosphere model is integrated from the pre-industrial baseline to a new equilibrium. Then the \( T_s \) difference between the two states is the value of ECS. However, it is not easy to reach the equilibrium state due to the expensive computation cost. As an alternative, a slab ocean model is usually coupled with the AGCM. Though the computing cost is cheap, the simplified climate system cannot introduce the effect of ocean circulation. Technically, coupling a slab ocean model with AGCM is not easier than the development of a fully coupled model. Thus, an approximation method is proposed based on transient state from fully coupled model to estimate the ECS (Gregory et al., 2004). Combining the equations (2) and (3) we obtain

\[
N = F + \lambda T'.
\]

As CO\(_2\) concentration is fixed, \( F \) is a constant. Then \( N \) can be regarded as a function in respect to \( T' \). By linear fitting \( N \) against \( T' \), we obtain the \( F \) at the intercept of \( y \)-axis \( (T' = 0) \), and the equilibrium temperature (i.e. ECS under 2×CO\(_2\)) at the intercept of \( x \)-axis \( (N = 0) \). The slope of fitting line is the \( \lambda \).

Fig. 2. Schematics of transient response (a) and equilibrium response (b). \( N \), net
radiative flux at the top of the atmosphere; \( F \), radiative forcing induced by forcing factors; \( E \), increased outgoing radiative flux of the Earth system due to warming; \( U \), ocean heat uptake. For the transient response, \( N \) approximates to \( U \), both not zero. For the equilibrium response, \( E \) offsets \( F \) and a new equilibrium state reaches.

When the equilibrium is reached, the net flux \( N \) at the TOA is zero (Fig. 2). The ECS is expressed as

\[
\text{ECS} = \frac{F \times \Delta T}{\lambda},
\]

where \( F \), the forcing of \( 2 \times \text{CO}_2 \), the ERF estimated based on specific model or 3.71 W m\(^{-2}\) as commonly used before. The \( 1/(-\lambda) \) is called climate sensitivity parameter that describes the warming induced by 1 W/m\(^2\) forcing at the TOA.

![Figure 3](https://example.com/figure3.png)

Fig. 3. The relation between surface temperature change \( \Delta T \) and net radiative flux \( N \) at the top of the atmosphere under the \( 4 \times \text{CO}_2 \) scenario. \( \Delta T \) and \( N \) is the differences between abrupt\( 4 \times \text{CO}_2 \) and piControl runs. Gregory-style regression (Gregory, et al, 2004) is used to estimate the ECS. The outputs of the multi-model ensemble of 24 CMIP5 models are shown. Feedback parameter \( \lambda \) is evidently different in two response stages (roughly before and after the 20th year).

To obtain more evident forced signal, \( 4 \times \text{CO}_2 \) forcing is usually used to drive the fully coupled model. Based on the empirical relation between \( \text{CO}_2 \) concentration and...
radiative forcing (Myhre et al., 1998), the forcing of 4×CO$_2$ is exactly as twice as that of 2×CO$_2$. Assuming the $\lambda$ is not changed, the ECS is half of the equilibrium temperature estimated from 4×CO$_2$ scenario. Figure 3 shows application of Gregory-style regression to obtain the ECS using the average of 24 CMIP5 models. The response in models shows two stages: a fast response in the first 20 years (green line) and a slow response later on (blue line), corresponding to different feedback parameter $\lambda$ (Chen et al., 2014). The mean $\lambda$ estimated by the ERF and ECS based on the two stages (dashed black line) is similar to the $\lambda$ calculated using the whole period (solid black line). The ECS estimated only by the slow response stage (the last 130 years) is 0.3 K higher than that derived from the whole period.

Besides model output, observational data can also be used to estimate the ECS in nature. However, in the real world, the $F$ evolves with time. Based on the equation (4), we should know the time series of $F$ contributed from the GHGs, aerosols, land use, solar perturbation, volcano activities and so on. The unbalanced flux $N$ at the TOA, and the surface temperature change $T'$ should also be known. Then we can fit the $(N - F)$ against $T'$ to obtain the value of $\lambda$ and then use the equation (5) to estimate the ECS (Forster et al., 2006).

3.2 TCR and TCRE

The TCR measures the sensitivity to CO$_2$ forcing in non-equilibrium state. Besides the feedback, the TCR is affected by the ocean heat uptake (OHU) (Figure 2). For transient response, the energy conservation is also satisfied, so we have

$$U = N = F + \lambda T',$$

where $U$ is the OHU, equal to the net flux at the TOA. Assuming the time-scale of OHU is much longer than that of the $T_s$ response, the U can be approximated as the first-order relation with the $T'$,

$$U = \kappa T',$$

where $\kappa$ is the efficiency of OHU with positive value. This linear approximation assumes that the ocean has infinite heat capacity and the OHU is regarded as one kind of “negative feedback”, which is applicable to the forcing moderately increasing
scenario (Gregory et al., 2008). Combining the equations (6) and (7), we obtain

\[ T' = \frac{F}{\kappa - \lambda}; \quad (8) \]

\[ \text{TCR} = \frac{F_{2x}}{(\kappa - \lambda)}. \quad (9) \]

It is evident that a strong OHU would can lead to a small TCR.

Based on the definition of the TCR, in practice, the value of TCR is calculated by using the change of T_s, a 20-yr mean state centered on the time of CO_2 doubling under the 1% yr^{-1} increasing scenario relative to the pre-industrial baseline. Using the same experiment, we can calculate the cumulative CO_2 emissions in the atmosphere C_e (unit: Pg C) before CO_2 concentration is doubled. Then, the TCRE is expressed as

\[ \text{TCRE} = \frac{\text{TCR}}{C_e}. \quad (10) \]

The unit of TCRE usually is converted to “K 10^{-3} Pg C”.

The emission-driven Earth System Model is another tool to estimate the TCR. The difference from conventional model is that the value of C_e is determined by carbon cycle and related feedbacks which may vary across models. Thus, the TCRE is the regression coefficient by fitting T_s against the cumulative CO_2 emissions (Collins et al., 2013).

4. Uncertainty in climate sensitivity

From IPCC TAR to AR5, extensive studies using paleoclimatic proxy data, historical instrumental observations and multi-model simulations, have not reduced the uncertainty of the ECS. The newly suggested possible range is 1.5–4.5 K, the same as the Charney sensitivity obtained in 1979 (Charney et al., 1979). Besides, divergent results are seen in different studies (Gregory et al., 2002; Forster et al., 2006; Andrews et al., 2012; Rohling et al., 2012; Olson et al., 2012; Masters, 2014). It is also believed that the relation between the feedback and ECS, which intrinsically determines the uncertainty, is hard to narrow (Roe et al., 2007).

From the equation (5), we have

\[ \Delta\text{ECS} = \frac{F_{2x}}{\lambda^2} \Delta\lambda, \quad (11) \]

It shows the relation between uncertainties of \lambda and ECS under the linearization assumption. If the \lambda is small, \Delta\text{ECS} will be large following one unit change of the \Delta\lambda.
It means that the uncertainty of ECS is inevitably large if the ECS itself is not small enough, which result in the upper limit of the probability hard to be constrained. It should be noted that these results are derived from the assumption of a linear feedback. Therefore, it is still controversial in understanding the relation between ECS and feedbacks (Roe et al., 2011).

The uncertainties of ECS estimated by models mainly come from feedback processes, especially the cloud feedback which contributes about 70% inter-model variance of the ECS. The shortwave feedback of low cloud in the tropical and subtropical regions (including the shallow convective cloud and stratocumulus) are highly uncertain (Randall et al., 2007; Klocke et al., 2011; Vial et al., 2013). The uncertainty of observation-based estimation of the ECS comes from the observational data itself, such as the net flux at the TOA, the forcing exerted by a variety of agents and the OHU. Although the ECS is the sensitivity to CO₂ concentration, the observed change of T₀ is the result from multiple forcing agents (Ma et al., 2005). Hence, to accurately estimate the climate feedback, the forcing from all the agents should be known. Besides the GHG forcing, aerosols is another important forcing agent which can exert forcing directly (direct effect) and also impact the radiative budget via interactions with cloud (indirect effect). The difficulty in estimating the aerosol forcing adds more uncertainties to the accurate estimation of the ECS/TCR.

The OHU plays an important role in the TCR/TCRE. A strong OHU delays the warming (Zhao et al., 1995; Stouffer et al., 2006). The simulation of eddy mixing intensity in ocean models can significantly affect the OHU in the vertical depths. Strong mixing in the Southern Ocean favors a more heat uptaken by the ocean which is further transported into the deep ocean through the meridional overturning circulation (Zhang et al., 2013). Large divergence of the spatial distribution of the OHU across models has been observed. Two typical distributions are prominent in the zonally mean pattern: high-latitude OHU and low-latitude OHU. They have different influences on the global warming. Following a more effective high-latitude OHU, we would witness a weaker warming (Winton et al., 2010; Rose et al., 2014). If the
carbon cycle is considered, the increasing rate of the cumulative CO$_2$ emissions is closely related to the carbon sources and sinks on the land and ocean. The uncertainties in the ecological processes and the interactions with temperature, precipitation, ocean circulation can further impact the magnitude of TCRE (Gillett et al., 2013).

5. Relation between “2 °C threshold” and climate sensitivity

The “2 °C threshold” issue is widely concerned by the public and research community. The 2 °C warming of $T_s$ above the pre-industrial level is considered as a warning threshold that indicates “dangerous anthropogenic interference” (Mann, 2009). Given the same radiative forcing and OHU, a larger ECS will shorten the time to reach a 2 °C warming. If the ECS is relatively large, the aim of ultimate warming below 2°C requires small forcing (CO$_2$ concentration), which would puts more stress on the emission reduction for human society. The issue can be understood based on the forcing-response relation.

Based on the equation (4), the ultimate equilibrium temperature $\Delta T$ is proportional to the forcing $F$. Assuming the constant feedback $\lambda$, we obtain

$$\frac{\Delta T}{ECS} = \frac{F}{F_{\lambda}} = \frac{\ln(C/278)}{\ln(2)}, \quad (12)$$

where $C$ (ppm) is the current concentration of CO$_2$ which is assumed to keep unchanged; 278 (ppm) is the pre-industrial reference CO$_2$ concentration. The CO$_2$ concentration $C$ can be expressed as the function of equilibrium warming $\Delta T$ and the ECS,

$$C = 278e^{\frac{\Delta T}{ECS}}, \quad (13)$$
Fig. 4. Relation between the equivalent CO₂ concentration and ECS constrained by certain warming threshold. The range of ECS is from the new estimation in the IPCC AR5. The curves under “1.5 °C threshold (blue)”, “2 °C threshold (black)” and “3 °C threshold (magenta)” are shown. For the “2 °C threshold”, corresponding equivalent CO₂ concentration is about 450 ppm when the ECS is the median estimated value. The relation between the C and ECS given ΔT = 1.5 °C (blue line), 2 °C (black line) and 3 °C (magenta line) is shown in Figure 4. The range of ECS is from the IPCC estimation, i.e. 1.5–4.5 K. When ΔT = 2 °C and the value of ECS is near the median of range (about 3 K, 50% probability higher or lower), corresponding CO₂ concentration is about 450 ppm. It is the base of the statement that “atmospheric CO₂ concentration should not exceed 450 cm³ m⁻³ if the warming is intended to be below the 2 °C threshold” (Schneider et al., 2007; Calvin et al., 2009; Wang et al., 2013; Oppenheimer et al., 2014).

It should be noted that the uncertainty in the ECS extensively impacts the CO₂ concentration under a certain temperature target. If the ECS is 1.5 K, the allowed CO₂ concentration can be as large as 700 ppm (Fig. 4). However, based on current knowledge, the probability of ECS below 1.5 K is very small (less than 0.05; Stocker et al., 2013). That is why Mann (2014) emphasized the importance and urgency of reducing GHGs emission, although we have experienced a flat warming period called global warming hiatus during the recent decade.
6. Research prospects

The history of the research on the climate sensitivity can be traced back to 100 years ago. Following the increasing of observational data, the developments of fully coupled climate system or even Earth System models, and the improvement of approaches for feedback analysis, our understanding on the issue has been greatly improved. Nevertheless, there still exist great challenges. For example, high-quality observational data is too short to detect climate change signal, especially in the Southern Ocean where the OHU is substantially large. In addition, the parameterized processes are still not perfect in the current state of the art climate models. This further reduces the reliability and limits the application of model output. The difficulty in reducing the uncertainty of climate sensitivity can be either due to the intrinsic climate system or the deficiency of our current knowledge. Based on our review of the recent proceedings in this field, the following research priorities are recommended to the climate sensitivity research community, in particular the Chinese community where the climate sensitivity studies remains to be weak.

(1) Non-linear interaction of feedbacks

Linear feedback analysis has been a matured method. “Radiative Kernel” approach based on the linear feedback analysis can provide spatial distribution of different feedback processes. However, large gaps between the sum of individual feedback and total feedback are found in many models, indicating that the strong non-linear interactions are not negligible (Via et al., 2013). The climate sensitivity is determined by the feedbacks. Therefore, the interactions among different feedback processes should be highlighted in the future research.

(2) Constraining climate sensitivity by combining the new observation and model development

One precondition for a reliable climate sensitivity of a model is that the historical climate change should be reasonably reproduced by the model, such as the warming trend in the 20th century. Many models still show limitations in the simulation of the 20th century global warming (Zhou and Yu 2006; Zhou et al., 2013). The uncertainty
of climate model’s sensitivity may be reduced if the model has been sufficiently constrained by observation (Jackson et al., 2008). The cloud-related convection is one important source of uncertainty in climate sensitivity (Sherwood et al., 2014). We need to improve the space-time resolution of monitoring cloud and convection at global scale, promote the understanding of interactions between cloud, convection and large-scale circulation, and properly parameterize these processes in climate models (Stevens et al., 2013). In addition, it remains a great challenge for climate model to simulate the abrupt change recorded in the paleoclimate archives, which is a strict criteria to test the performances of climate models (Wang et al., 2013).

(3) Estimating Earth system sensitivity

Under the concept of traditional physical climate, the focus of climate sensitivity research is “forcing – response – feedback” process following the increase of atmospheric CO$_2$. The responses of land and ocean carbon repositories were not taken into account. In nature, the carbon cycle including the biological effects can further feed back to the increasing atmospheric CO$_2$ concentration and surface warming. This kind of processes can influence the change of surface temperature at time scales from decadal to millennium, and impact the estimation of the ECS. The carbon cycle and related feedbacks determines the increasing rate of cumulative CO$_2$ emissions, further adding uncertainties to the TCRE. Thus, more efforts should be devoted to the estimation of Earth system sensitivity and the development of optimal emission path with the concept of TCRE.

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