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Estimations of climate sensitivity based on top-of-atmosphere radiation imbalance

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Abstract. Large climate feedback uncertainties limit the accuracy in predicting the response of the Earth's climate to the increase of CO_2 concentration within the atmosphere. This study explores a potential to reduce uncertainties in climate sensitivity estimations using energy balance analysis, especially top-of-atmosphere (TOA) radiation imbalance. The time-scales studied generally cover from decade to century, that is, middle-range climate sensitivity is considered, which is directly related to the climate issue caused by atmospheric CO_2 change. The significant difference between current analysis and previous energy balance models is that the current study targets at the boundary condition problem instead of solving the initial condition problem. Additionally, climate system memory and deep ocean heat transport are considered. The climate feedbacks are obtained based on the constraints of the TOA radiation imbalance and surface temperature measurements of the present climate. In this study, the TOA imbalance value of $0.85\,W/m^2$ is used. Note that this imbalance value has large uncertainties. Based on this value, a positive climate feedback with a feedback coefficient ranging from -1.3 to $-1.0\,W/m^2/K$ is found. The range of feedback coefficient is determined by climate system memory. The longer the memory, the stronger the positive feedback. The estimated time constant of the climate is large (70$\sim$120 years) mainly owing to the deep ocean heat transport, implying that the system may be not in an equilibrium state under the external forcing during the industrial era. For the doubled-CO_2 climate (or $3.7\,W/m^2$ forcing), the estimated global warming would be 3.1 K if the current estimate of $0.85\,W/m^2$ TOA net radiative heating could be confirmed. With accurate long-term measurements of TOA radiation, the analysis method suggested by this study provides a great potential in the estimations of middle-range climate sensitivity.

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1 Introduction

Large climate feedback uncertainties limit the ability of current general circulation models (GCMs) to predict the climate system change, including the response of the Earth's climate to the increase of CO_2 concentration within the atmosphere. Current estimates of global mean temperature increases for a doubled-CO_2 ($2\times CO_2$) atmosphere range from $\sim$1.0 K up to more than 10 K (IPCC, 2007), which has remained virtually unchanged for three decades. This wide envelope of climate projections is an obvious result of the intrinsic sensitivity of climate prediction systems to the climate feedback coefficient (Roe and Baker, 2007). Development of advanced methods to reduce the large feedback uncertainties is critical and urgent for both climate sciences and socioeconomic policies.

Most projections of future climate scenarios are based on GCM results. Complicated non-linear processes of the atmosphere, land, ocean, cryosphere, biosphere, and human activities, make the GCM simulated results difficult to understand. Incomplete knowledge of these processes, especially those related to clouds and precipitation, causes considerable differences in parameterizations and representations of physical, chemical and biological processes in individual GCMs. This, in turn, generates large differences in projected climate feedbacks and responses.

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Another way of predicting the climate change for a $2\times CO_2$ atmosphere is to build simplified or idealized models that focus on the variations of fundamental physical processes of the Earth's climate system. Among these simplified/idealized models, energy balance models based on perturbation theory have been used for decades. These idealized models generally solve linear differential equations (LDEs) that account for the basic climate mean state, forcing, and response. Early investigations (e.g., Dickinson, 1981; Hansen et al., 1984; Manabe et al., 1990) lacked specific information within the solutions of these LDEs owing to insufficient knowledge of the climate feedbacks, climate transition time constants, and heat reservoirs. Based on energy balance approach, Schwartz (2007) recently estimated the climate effective heat capacity, time constant and feedback coefficient from ocean heat storage and surface temperature measurements. The time constant of the climate system he found is about 8.5 years when an autoregressive technique was applied to the autocorrelation function of the detrended surface temperature measurements and the influence of very fast processes such as those from weather systems (or the sub-annual time constant of the climate system) was accounted for (Schwartz, 2008). This time constant of global surface temperature results in that the climate system may have a feedback slightly higher than that of blackbody emission to the external forcing mainly caused by the CO_2 increase in the atmosphere, because a relatively short equilibrium time period is needed. The short response can also be a direct result of an energy balance analysis that only considers a small heat capacity of the climate system (such as that of only the mixed layer of ocean). Although only the temperature of the ocean mixed layer is linearly related to surface temperature, the climate forcing driving the variations of the surface temperature, along with feedbacks, heats not only the ocean mixed layer but also the deep ocean owing to oceanic vertical heat transport. This heat transport process would significantly increase the time constant of the climate system. Actually, the autocorrelation function of the surface temperature measurements is non-integrable (or non-convergent) even when the time lag reaches as long as 20 years, indicating that the integral of the autocorrelation function cannot provide a reliable estimate of the time constant (Von Storch, 2004). The non-integrable feature of the autocorrelation of the observed temperature data implies that the surface temperature data series may be not long enough to describe the actual climate system under current transient conditions if an autoregressive technique is used. Besides this short time scale issue, previous energy balance analyses tried to solve an initial condition problem of LDEs, while the climate prediction for the increasing CO_2 atmosphere is clearly a boundary condition problem. Furthermore, the climate feedbacks should include not only short-term (including instantaneous) responses but also longer time scale (or historical) responses because the climate generally has certain memories, which are omitted in these energy balance models.

With the problems in the estimation of climate feedbacks, especially for GCMs and idealized energy balance models, innovative methods in determining climate sensitivity are needed as called for by Aires and Rossow (2003). This study explores a potential method to reduce climate feedback uncertainties by considering the mean transient climate states and addressing previously mentioned issues within the idealized energy balance models. The climate model used here is both complicated enough to account for major physical processes of the climate with an increasing external forcing and simple enough to understand the physics of the results and analyzed physical processes. Owing to the limitation of the observational data length and the simplification of the modeled climate system, climate time scales either longer than multi-century involving deep ocean circulation and other even longer geological processes or shorter than those of weather phenomena are not considered in this study. Thus, this study more or less focuses on "middle-range" climate sensitivity, the key problem of the Earth's climate change caused by changing CO_2 amount of the atmosphere. The solution of the climate model is obtained using constraints from observations at the boundaries of the Earth's climate system, especially from the top-of-atmosphere (TOA) radiation imbalance measurements. The uncertainties in current feedback estimates are quantified. Because of the extreme importance of the climate energy imbalance for climate studies, long-term measurements of TOA radiation with both high precision and high absolute accuracy are required. With these measurements and the method suggested by this study, the uncertainty in climate feedback estimates could be significantly reduced.

2 Methodology

In an idealized energy balance model (e.g., Manabe et al., 1990; Schwartz, 2007), the response of the global mean climate to a radiative forcing F in a unit area can be expressed as:

$$Cp\frac{dT}{dt} = F + f_{\mathrm{tot}}T, \tag{1}$$

where T and t represent the small global mean surface temperature perturbation and the time, respectively; Cp is the climate heat capacity, assumed to be proportional to an effective depth of the ocean; and f_{tot} is the total feedback coefficient. A combination of the forcing and feedbacks represents the net radiation of the climate system that decides the change of global mean climate. The time constant in this case is Cp/f_{tot}. Were the climate in a normal state ($F=0$ for long time), any small temperature perturbation would cause at least a $3.3\,\mathrm{W/m^2/K}$ of radiative heat release to space mainly because of blackbody emission (Schwartz, 2007). Thus, the feedback coefficient for the normal climate f_n is $-3.3\,\mathrm{W/m^2/K}$. For a forced climate, any total feedback coefficient f_{tot} values larger (smaller) than f_n would be the result of positive (negative) climate feedbacks besides the blackbody thermal emission (Note: hereafter the meaning of "positive" or "negative" climate feedbacks is based on this blackbody emission concern). The evaluation of the idealized climate described by Eq. (1) is clearly focused on the two parameters Cp and f_{tot} as mentioned previously.

Since the climate system described by Eq. (1) only deals with short time scale (shorter than decade including instantaneous) feedbacks and climate states, no historical (or memory) climate state is involved. Thus, we call the feedback coefficients in the equation as short-term feedback f_S. For short-time scales and small climate perturbations, the changes in surface net radiation, mainly from emission, would be radiated to space at TOA as a result of too little atmospheric heat capacity and the short time for the climate to adapt to the surface change. Thus, the climate system may have a strong tendency that pushes itself back to its equilibrium state for small short-term climate perturbations mainly by radiation adjustment. Evaluating TOA radiation and surface data, a short-term feedback coefficient about $-6\,\mathrm{W/m^2/K}$ was found (Spencer and Braswell, 2009). This short-term feedback f_S can be considered as a result of the normal climate feedback f_n superimposed by other negative feedbacks of the climate system with a feedback coefficient f of $-2.7\,\mathrm{W/m^2/K}$, i.e., $f_S = fn + f$.

The actual climate system also has certain memories for climate states. For example, the soil moisture reservoir has a memory considerably longer than that of most atmospheric processes. The time scale of soil moisture memory is generally about half a year with significantly longer times (longer than 1 year) for deep soil (Wu and Dickinson, 2004). Other processes such as those in cold regions involving frozen soil, snow and ice and wind driven and thermohaline ocean circulations have much longer memories (Blender et al., 2006). Evaluating global mean surface temperature anomalies of the Goddard Institute for Space Studies (GISS; Hansen et al., 1996; updated at http://data.giss.nasa.gov/gistemp/) shows that significant memories of the climate system with 95% or higher confidence level can be detected from the autocorrelation function of the surface data with time lags shorter than about 8 years (Note that the estimated autocorrelation function can be found in Schwartz, 2007). Thus, a feedback term for system memory f_m is added to Eq. (1). Since this system memory feedback comes from a non-instantaneous response of the climate system, the feedback also represents long-term climate feedbacks, which has significant contributions to the middle-range climate sensitivity. In this study, we use the average of surface temperature perturbations during previous time periods to represent the effect of the climate feedbacks from memories.

An additional modification of Eq. (1) is required to separate the deep ocean from the surface heat reservoir so that the heat capacity Cp only represents the ocean mixed layer whose temperature tracks the surface temperature. The temperature variations of the thermocline and abyss of the ocean are different from those of the mixed layer and surface, but the net radiation into the climate system generally not only heats the ocean mixed layer but also transports into the deep ocean. Thus, a term to characterize this deep ocean vertical heat transport process is added to Eq. (1) as:

$$Cp\frac{dT}{dt} = F + fsT + \frac{fm}{tm}\int_{t-tm}^{t} T\,dt' - O, \tag{2}$$

where t_m is the system memory length, and O is the deep ocean component of the heat transport of TOA net radiation. Again, the net radiation is the combination of all radiative forcing and feedbacks (i.e., the summation of the first three terms in the right hand side of the Eq. (2) for current case). When memory length t_m approaches zero, the memory term reduces to $f_m T$, which can be merged to the short-term feedback fsT term, i.e., the system has no memory or long-term feedback effects. In this case, the difference between our analysis and previous studies is that the deep ocean heat transports are included here.

For oceanic vertical heat transport, a simple parameterization using an exchange coefficient in terms of mixed layer temperature or diffusive heat transfer is commonly used (Dickinson, 1981; Hansen et al., 1984; Lindzen and Giannitsis, 1998). Because of extreme complicated sea water vertical movement such as oceanic overturn and Ekman pumping, this simplification of heat transport may not be best representations of ocean vertical heat exchange (Dickinson, 1981; Hansen et al., 1984). This study assumes that the transported heat to deep ocean is proportional to TOA net radiation, i.e.,

$$O = \mu\left(F + fsT + \frac{fm}{tm}\int_{t-tm}^{t} T\,dt'\right), \tag{3}$$

where μ is the heat transport coefficient for the deep ocean. This assumption represents an integrated condition of the vertical heat transport over all ocean basins and may be more straightforward for a system with small perturbations. Also, it results in similar ocean heat transports and long-term feedbacks as those from the parameterization in term of mixed layer temperature from our simulations of the energy balance approach owing to nearly-linear relationship between TOA net radiation and temperature (c.f. Fig. 2 later).

Combining Eqs. (2) and (3), our energy balance model is derived as:

$$\frac{Cp}{(1-\mu)}\frac{dT}{dt} = F + fsT + \frac{fm}{tm}\int_{t-tm}^{t} T\,dt', \tag{4}$$

which can be rewritten as:

$$Cp'\frac{dT}{dt} = F + fsT + \frac{fm}{tm}\int_{t-tm}^{t} T\,dt', \tag{5}$$

with

$$Cp' = Cp/(1-\mu) = Cp/\eta = \rho SwD/\eta, \tag{6}$$

where η =1-μ is the factor of the net radiation that is trapped within the climate system before being transported into the deep ocean, and ρ, *Sw*, as well as D are water density, water specific heat, and ocean mixed layer depth, respectively. This $Cp^{'}$ value is assumed to be a constant once both the mixed layer depth and ocean heat transport coefficient are specified in our simulations.

The key unknowns for the climate system described by Eqs. (5) and (6) are the coefficients f_m and μ. Although certain knowledge about mixed layer depth is helpful in understanding climate heat storage and transport processes, the solutions of feedback f_m and temperature T for these equations are not specifically dependent on the mixed layer depth. For example, a set of f_m, μ and T solutions that satisfy these equations for a depth $D=50$ m would have the same *fm* and T solutions as a climate system with $D=100$ m. The only change for this set of solutions is the η (or $1-\mu$) that is proportionally doubled since these cases with different mixed layer depths mathematically hold the same governing equations. So, once these equations are resolved with a specified mixed layer depth, the solution for any other depth is also resolved. This significantly reduces the complexity of the mathematical calculations. Furthermore, even though Eq. (5) is not in a normal form of LDE, the total feedback coefficient of the climate system can be obtained in a manner similar to a linear equation. For an example, the asymptotical solution provides: $T(t\rightarrow\infty)=-F(t\rightarrow\infty)/(fs+fm)$ $=-F(t\rightarrow\infty)/f_{\mathrm{tot}}$, where the total feedback f_{tot} is the combined result of *fs* and *fm*. This result implies the extremely important middle-range climate sensitivity, such as those for $2\times CO_2$ atmosphere, can be inferred after *fm* is estimated.

3 Results

To understand the basics of the change in climate states, we consider the mean temperature perturbation and climate forcing during the last 120 years. For the modeled climate system (Eq. 5), we assume that there is no temperature perturbation before time zero, or, T $(t<0)=0$. This $t=0$ can be considered as 120 years ago. Also, the forcing is set to be zero before this time, that is, F $(t<0)=0$, because this study mainly focuses on the climate change caused by anthropogenic processes, especially CO_2 change, whose forcing was minimal at pre-industrial time. Because of the system memory, these boundary conditions before $t=0$ are critical in solving the governing equation. After time 0, the forcing F is linearly increased to 1.8 W/m^2 at the end of the 120 years. This 1.8 W/m^2 external forcing is consistent with current estimates of the net change of effective forcing during the last 120 years (IPCC, 2007; Hansen et al., 2005; Schwartz, 2007). In this study, a short-term feedback coefficient *fs* of -6 W/m^2/K is used in Eq. (5).

As mentioned in the previous section, the coefficients for the system memory feedback and the deep ocean heat transport are the unknowns in Eq. (5). These unknowns, especially the feedback from system memory, are the key to the prediction of future climates. To solve these unknowns, we apply two constraints on Eq. (5) for the last 10 years of the 120-year studied period. The first constraint is the observed average of surface temperature increase T_L, which is about 0.6 to 0.7 K (Hansen et al., 2005). A T_L value of 0.65 K is used here. The uncertainty associated with this estimate is about 0.05 K. The other constraint is the imbalance of TOA radiation, Q_L, which has been measured by satellites for more than two decades. Decadal TOA net radiation records from satellite measurements show that the measurement precision is generally within about 1 W/m^2 for large-spatial-scale annual means (Wielicki et al., 2002; Lin et al., 2008). Unfortunately, there is a lack of high accurate absolute calibration for satellite TOA radiation measurements, and, thus, the absolute accuracy of the radiation measurements could be lower. Since there is almost no heat storage change within the atmosphere and land at annual time scales owing to their negligible heat capacity and temperature change, the TOA net radiation or the imbalance should be the same as ocean heat storage change. Actually, the measurements of interannual variations of TOA net radiation and ocean heat storage are found to be very consistent (Wong et al., 2006). From the ocean heat storage measurements, the TOA net radiation can be inferred as about 0.85 W/m^2 (Wong et al., 2006; Willis et al., 2004). Thus, the average value of 0.85 W/m^2 for the annual means of the last 10 years is used in this study, following Hansen et al. (2005) and Trenberth et al. (2009). Since the uncertainty of the annual means of the net radiation is about 0.4 W/m^2 (Wong et al., 2006), the uncertainty in the 10-year average of Q_L would be 0.13 W/m^2 if the year-to-year variations were independent. Considering potential physical processes beyond the year-to-year variability or decadal scale processes, an uncertainty of 0.2 W/m^2, which is about 50% higher than that (0.13 W/m^2) of the assumed independent variability, is assigned as the error bar for the random part (i.e., not including the systematic bias) of the errors associated with the 10-year TOA radiation imbalance (0.85 W/m^2) of current analysis. We emphasize that at present, the estimates of the absolute errors (or the systematic biases) associated with these ocean heat storage measurements (or the TOA radiation imbalances) are not available. Large errors as high as 1 W/m^2 potentially exist. This analysis uses those TOA imbalance and error bar values only as a relevant observation-based case for our novel method in estimating the climate sensitivity. Long-term TOA radiation measurements with high precision and high absolute accuracy in determining the TOA imbalance are critical for future climate predictions.

With these aforesaid two constraints and other basic information mentioned previously for the climate system model, Eq. (5) can be solved numerically. Actually, an analytical solution for the climate system is also possible although this equation cannot be solved in the normal LDE framework.

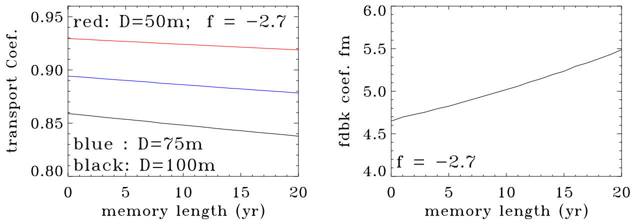

Fig. 1. The estimated coefficients of deep ocean heat transport μ (left panel) and climate feedback *fm* (in unit W/m^2/K, right panel) of the climate system for memory length t_m up to 20 years. The f parameter in this and following figures represents the feedback coefficient of short-term feedbacks other than blackbody emission (c.f. the text).

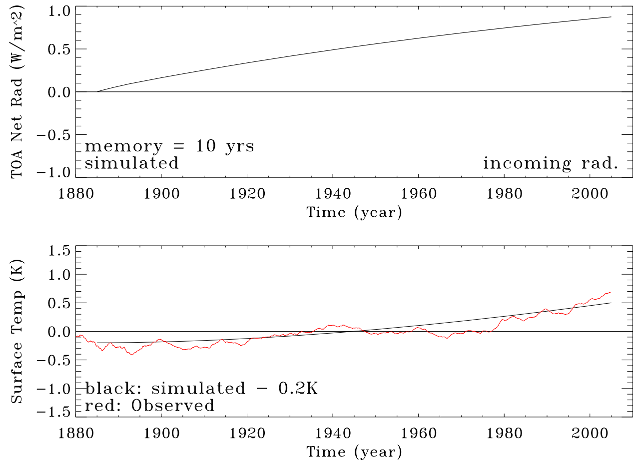

Fig. 2. Calculated results of TOA net radiation (top panel) and surface temperature (lower panel) during the last 120 years for a climate system with memory length of 10 years. For the surface temperature plot, calculated results (black line) are offset by 0.2 K to compare to the 5-year running mean of the observed GISS surface data (red curve).

But, to derive the analytical solution, a transcendental equation needs to be solved, which still requires numerical calculations in addition to complicated analytical efforts. This is beyond the scope of this paper. To focus on the understanding of physical processes of the climate system, numerical calculations are therefore directly applied to Eq. (5).

Figure 1 shows the numerical solution of the coefficients of the deep ocean heat transport (left panel) and memory feedback (right panel) for a memory length t_m of up to 20 years. As mentioned previously, the results for *fm* are not specifically dependent on mixed layer depth, so a 100 m mixed layer depth is used in solving *fm* and μ. Other μ values for different mixed layer depths are calculated based on the proportionality of η with the depth. The plotted μ values are for $D = 100$ m (black curve), 75 m (blue curve) and 50 m (red curve), respectively. These mixed layer depths are equivalent to the real Earth's ocean mixed layer depths of 141 m, 106 m and 70 m, respectively, since the ocean covers only about 71% of the Earth's surface. Although there are not enough measurements to determine the climatology of global averaged mixed layer depth, the actual global mean mixed layer depth would be generally within the range of the calculated depths. Thus, these calculated results constrained by observations provide certain information about the first order heat transports to the deep ocean. From Fig. 1, it can be seen that most of the heat (85 to 93% for $t_m < 10$ years) generated by the climate forcing and feedbacks is transported to the deep ocean. The shallower the mixed layer the larger the percentage of heat transported because the surface (or, mixed layer) temperature is constrained by current observations. In other words, with the same amount of net heat into the climate system as that used in this study (0.85 W/m^2), more heat has to be removed from the mixed layer to the deep ocean for a shallower mixed layer climate system to satisfy the observed temperature change. For a given mixed layer depth, changes in heat transport coefficients are generally small ($<\sim$2%). Small decreases of the coefficients with system memory length are caused by increases in the feedbacks for climates with longer memories (right panel).

Stronger memory feedback under the constrained net radiation condition means that more heat would be trapped in the mixed layer, thus, heat transport coefficient would be smaller to maintain the energy conservation. Generally, the long-term feedback coefficient *fm* from current estimation is positive and slightly increases with memory length from 4.7 to 5.0 W/m^2/K for t_m changing from 1 to 10 years, which results in the critical feedback coefficient f_{tot} varies from −1.3 to −1.0 W/m^2/K. This long-term feedback is so strong that it even changes the total climate feedback from negative owing to short-term feedbacks to positive, that is, $f_{\text{tot}} = fs + fm > fn$. This positive feedback is a direct result of the observed temperature increase and TOA net heating for the climate. Without the net heating, the system would be within its quasi-equilibrium state as shown by Schwartz (2007). Short-term feedback processes of the climate try to absorb external forcings applied to the system. The potential TOA radiation imbalance (or ocean heat storage change) implies that there are certain physical processes that provide persistent positive feedbacks to reduce the effect of short-term negative feedbacks and to let the climate system adapt to long-term forcing influences.

Figure 2 shows the simulated TOA net radiation and surface temperature during the last 120 years for a climate system with memory length of 10 years. The current 0.85 W/m^2 radiation imbalance could be built up from the entire history of the industrial era (top panel), and may continue to grow if no limits on the increase of CO_2 and other greenhouse gas emissions are made. For the surface temperature plot (bottom panel), our estimates (black line) are offset by 0.2 K in order to compare to the 5-year running mean of the observed

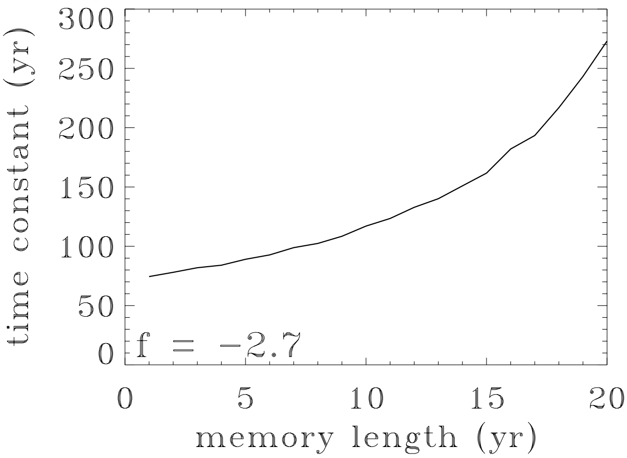

Fig. 3. The change of estimated climate time constant with climate system memory.

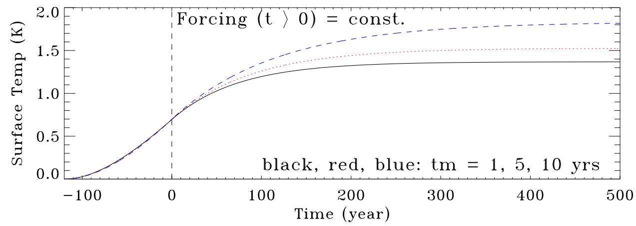

Fig. 4. The effect of big time constants on climate change. A fixed external forcing equal to the current value (or 1.8 W/m^2) is applied to the climate system from the present day (or time 0) forward. The black, red and blue curves represent the results for memory lengths of 1, 5 and 10 years, respectively.

GISS surface data (red curve). The two temperature time series are very consistent. The detailed variability caused by climate processes such as volcanoes and El Nino/Southern Oscillation in the observed surface temperature is beyond the scope of the current analysis since this study only considers the changes in the mean climate state forced by long-term forcing. Near-linear trends of the estimated results for both radiation and temperature are observed, which is clearly related to the linear forcing F used in this analysis. Another reason for the near-linear results is the large time constant obtained by this study.

The time constant (Fig. 3) of the modeled climate increases with memory length and has a sharp change for memories beyond 10 years, especially for t_m>15 years. Even within the range of significant climate memories (1 to 10 years), the time constant is large, varying from 74 years for 1-year climate system memory to 117 years in the case of 10-year memory. This is much longer than the time scales of most atmospheric processes and climate system memory. Since the TOA radiation is significantly imbalanced, and the ocean mixed layer is relatively shallow compared to the net heating from the radiation, a large amount of heat (>85%) has to be transported to the deep ocean in order to satisfy surface temperature observations. This continuous process of dumping more and more heat to the deep ocean makes the surface and mixed layer temperatures hardly reach equilibrium, which, along with a much smaller absolute value of f_{tot} compared to that in normal conditions, increases the time constant of the climate system. Although our estimates of the time constant are obtained based on observational constraints, confirming these estimates is almost impossible because observations related to long time scale processes are presently not available. Coupled ocean-atmosphere GCM simulations (Hansen et al., 2007) show that the climate response for an instantaneous $2\times CO_2$ forcing reaches 60% of the equilibrium response after 100 years and 90% after 1000 years. The former system response corresponds to a time constant of 109 years, which is consistent with current estimates, while the latter indicates another even bigger time constant of the climate system of about 434 years. This longer time scale may be related to thermohaline circulations of the deep ocean, whose physical processes are beyond the scope of current study.

The effect of big time constant (~70 to 120 years) of the climate system on climate changes is illustrated in Figure 4 for a hypothesized scenario in which a fixed external forcing equal to the current value (or 1.8 W/m^2) is applied to the climate system after present day. In this figure, the present day is set to be 0, that is, the results for time >0 are projections. The black, red and blue curves represent the results for memory length of 1, 5 and 10 years, respectively, which may cover the entire range of possible memory length of the studied climate system. The constraints on the boundary conditions of both pre-industrial era and present days cause the three simulated results indistinguishable before time 0. A climate system with a shorter memory reaches its equilibrium state slightly earlier than those with longer memories. It can be seen that even with a stabilized forcing, the system may need a few hundred years more to reach its equilibrium state. The asymptotical temperature increase in this case varies from 1.4 to 1.8 K, because the total feedback coefficient f_{tot} ranges from −1.3 to −1 $W/m^2/K$. Thus, potential warming of the climate could be stronger than what has been observed in the last decade or so if the 0.85 W/m^2 TOA net radiation imbalance is confirmed.

With TOA radiation imbalance and surface temperature measurements, the key coefficients of climate feedback and deep ocean heat transport can be estimated, which may potentially reduce the uncertainty of estimated climate sensitivity. However, certain factors within the current energy balance model, including the choice of the estimated short-term feedback coefficient *fs* as well as errors in the constraints of the last 10-year averages of the temperature and net radiation used in our 120-year calculations, may influence the estimations of these coefficients. Changes in *fs* values directly

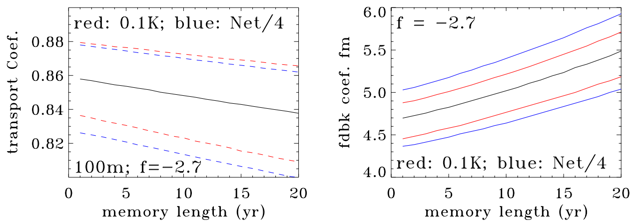

Fig. 5. Sensitivity tests of the deep ocean heat transport (left panel) and long-term climate feedback (in unit W/m^2/K, right panel) on the constraints of T_L and Q_L.

affect our solutions of *fm*, but the effect of different *fs* values on total feedback coefficient f_{tot} is small because the temperature and net radiation constraints force the modeled climate system to generate similar amounts of net heat and temperature increases to satisfy these boundary conditions. Thus, the basic conclusions about climate feedback and sensitivity would not be affected much by the choice of *fs*. This also means that the system constraints from the T_L and Q_L observations are the most important error sources in this middle-range climate sensitivity analysis.

Figure 5 shows the results of sensitivity tests of the ocean heat transport (left panel) and long-term climate feedback (right panel) to the constraints of T_L and Q_L for the climate system discussed for Fig. 1. In these sensitivity tests, we add ±0.1 K temperature and ±25% (or 0.2 W/m^2) radiation uncertainties into the original T_L and Q_L values, respectively. This temperature uncertainty is about twice as large as current estimates, and the 0.2 W/m^2 uncertainty is about 50% higher than the assumed independent Q_L variability and consistent with the estimate of Kiehl (2007). We emphasize that the TOA net radiation uncertainty used here may be underestimated, as discussed previously. Although the actual errors for the absolute TOA radiation imbalance or ocean heat storage change are not available, this study still provides the essential information on the sensitivity of current method in the estimations of the climate feedback and middle-range climate sensitivity. The black, red and blue curves in the figure represent results of control, T_L and Q_L tests, respectively. The μ changes with both T_L and Q_L uncertainties are small and basically within about 2∼3%. Clearly, the changes in μ with T_L or Q_L changes are asymmetric because of the upper limit unity of μ values. Note that lower T_L constraints cause higher μ values than higher T_L constraints since more heat needs to be transported out for a constant heating condition. Unlike the μ values, the higher the T_L or Q_L constraints, the stronger the long-term feedbacks (or the bigger the *fm* values). Also, the same variations in T_L or Q_L generate almost the same *fm* changes. An additional critical point is that the *fm* changes caused by uncertainties in Q_L are generally significant, while the potential errors caused by T_L uncertainties may not be so severe. For example, the change in *fm* is smaller for a T_L uncertainty of 0.1 K than those from a 0.2 W/m^2 uncertainty in Q_L. Since the 0.1 K uncertainty added on T_L is already about 100% of the increases in the estimated T_L uncertainties, while the 0.2 W/m^2 uncertainty imposed on Q_L is only a 50% of the increases in the assumed independent Q_L variability and could still be underestimated owing to potential large bias errors in the absolute TOA radiation imbalance estimate. Thus, it is more likely that Q_L errors would create larger errors in the estimated *fm*. Based on the results of current sensitivity tests, we estimate that the error bars for *fm* are about ±0.4 W/m^2/K for the considered Q_L uncertainty of 0.2 W/m^2.

4 Conclusions

Since for the modeled climate system (or for the climate variability on time scales about a century) the climate memory is generally within 1 to 10 years, the estimated total climate feedback coefficient f_{tot} would be in the range of −1.3 to −1.0 W/m^2/K for the estimated 0.85 W/m^2 TOA radiation imbalance. Thus, for the 2×CO_2 climate (or 3.7 W/m^2 forcing), the estimated global warming would be in the range between 2.8 K to 3.7 K. Since the best estimated memory length of the climate system is about 4 years owing to the time lag of the maximum autocorrelation beyond 0 lag of the GISS surface temperature data, the best estimates of *fm* and f_{tot} would be 4.8 and −1.2 W/m^2/K, respectively, resulting in our estimated most likely warming of 3.1 K if the radiation imbalance used is confirmed. These results are clearly in alignment with previous projections around the peaks of climate sensitivity distributions obtained from GCMs (IPCC, 2007). The difference between current estimates and previous results is that our estimates provide very straightforward physics, and have a great potential to reduce the broad range of climate predictions from GCMs.

Because of the extreme importance of the climate energy imbalance for climate studies as shown in this report, long-term measurements of the TOA radiation with both high precision and high absolute accuracy are desperately demanded. These measurements will provide the key information to nail down the climate feedback and middle-range climate sensitivity. A great potential in accurate climate predictions, thus, could be realized. Furthermore, with long-term, accurate global energy imbalance measurements and the method suggested by this study, a physically-based tool for decisions related to global warming policies can be offered to the public and policymakers, which will have enormous socioeconomic impacts.

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