The role of internal climate variability in projecting Antarctica's contribution to future sea-level rise

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Abstract

Retreat of the Antarctic ice sheet (AIS) is likely to be a major contributor to future sea-level rise (SLR). Current projections of SLR due to ice-sheet mass loss remain highly uncertain. Better understanding of how ice sheets respond to future climate forcing and variability is essential for assessing long-term risk of SLR. However, predictability of future climate is limited by uncertainties from emission scenarios, model structural differences, and internal climate variability (ICV) that is inherently generated within the fully coupled climate system. Among those uncertainties, the impact of ICV on the AIS changes has not been explicitly assessed. Here we quantify the effects of ICV on the AIS evolutions by employing climate fields from two large-ensemble experiments using the Community Earth System Model to force a three-dimensional ice-sheet model. We find that ICV of climate fields among ensemble members leads to significantly different AIS responses, and that most of the effect is due to atmospheric variability compared to oceanic. Our results show that ICV can cause about 0.08 m differences of AIS contribution to SLR by 2100 compared to the ensemble-mean AIS contribution of 0.38–0.45 m. Moreover, using ensemble-mean climate forcing fields as the forcing in an ice-sheet model significantly delays retreat of the West AIS for up to 20 years and significantly underestimates the AIS contribution to SLR by 0.07–0.11 m in 2100 and up to 0.34 m in the 2250's. This study highlights the need to take internal climate variability into account in assessing uncertainty associated with the AIS contribution in sea-level rise projections.

Keywords Ice sheets · Antarctica · Sea level · Climate variability · Climate model · Climate change

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1 Introduction

The Antarctic Ice Sheet (AIS) has the potential to cause substantial sea-level rise (SLR) but the evolution of the AIS in response to future climate change remains a major source of uncertainty in sea-level projections for the upcoming centuries (Church et al. 2013). To assess the contributions of ice sheet melt to SLR, physically-based numerical ice-sheet models have been adopted to investigate the behavior of ice sheet changes. However, various sources of uncertainties exist in the current ice-sheet modeling framework, including representations of physical processes in ice sheet changes, model parameter values, and climate forcings. The climate forcing fields, such as near-surface air temperature, precipitation and ocean temperature, incorporated in the ice sheet models are typically simulated from climate models. Therefore, the underlying uncertainty in future climate projections as estimated by climate models would eventually contribute into the overall uncertainty in projecting future ice sheet changes and their contributions to SLR.

Modeling future AIS changes is challenging because the AIS is a complex system interacting with both the atmosphere and ocean on distinct spatial and temporal scales. Several modeling studies (Winkelmann et al. 2012; Golledge et al. 2015; DeConto and Pollard 2016) have shown that greenhouse gas (GHG) emissions in the next few decades could strongly influence the long-term contribution of the AIS to global mean sea level (GMSL). Observational studies also suggest that recent retreat of ice shelves over the Antarctic Peninsula (AP) and the West Antarctic Ice Sheet (WAIS) (Paolo et al. 2015) can be attributed to massive and rapid regional warming in the atmosphere (Vaughan and Doake 1996; Turner et al. 2005; Steig et al. 2009) and the ocean (Liu et al. 2015; Cook et al. 2016). Variability and changes in the Antarctic climate can affect some key physical processes at the interface of ice sheet/ice shelf, atmosphere and ocean. For example, most of the WAIS is vulnerable to warm ocean water intrusion because it is grounded below sea level and its extensive ice shelves are in contact with the ocean (Joughin et al. 2012; Alley et al. 2015). The warmer ocean water increases oceanic basal melting under the floating ice shelves, leading to ice shelf thinning and reduction of buttressing that supports the interior grounded ice. If the grounding line retreats inland onto a deeper bed, the flux of ice across the grounding line will increase as the ice thickens, resulting in runaway ice-sheet retreat due to “Marine Ice Sheet Instability” (MISI) (Mercer 1978; Schoof 2007). In addition, the increased surface meltwater due to atmospheric warming can cause ice-shelf calving via hydrofractures with meltwater drainage into crevasses (Nick et al. 2010). This mechanism is partly responsible for causing the sudden break-up of the Larsen B ice shelf in 2002 (Scambos et al. 2000; Banwell et al. 2013) when the unusually warm summers preceded the event (Vaughan et al. 2003). After the ice shelf that provides buttressing is removed, the high subaerial cliff along the ice margin becomes structurally unstable, triggering catastrophic cliff failure into deep basins (Bassis and Walker 2011; Bassis and Jacobs 2013). The combined effects of hydrofracturing due to surface melt and large ice cliff failure at the grounding line are known as “Marine Ice Cliff Instability” (MICI), which could lead to accelerated disintegration of the WAIS occurring on centennial time scales (Pollard et al. 2015; DeConto and Pollard 2016). Therefore, better understanding of the interactions between ice sheets and climate system, and how ice sheets respond to future climate change and its variability, are essential for robustly estimating the AIS contribution to SLR.

Due to high computational costs of conducting fully-coupled climate-ice sheet simulations, currently most long-term studies with dynamical ice sheet models are driven by offline forcing fields, some provided from general circulation models (GCM) or high-resolution regional climate

models (RCM). However, various sources of uncertainties exist in future climate projections using GCMs, including emissions-scenario uncertainty, model uncertainty due to different model physics or resolutions (Tebaldi and Knutti 2007), and internal variability due to different initial conditions (Hawkins and Sutton 2009). The Coupled Model Intercomparison Project Phase 5 (CMIP5) (Taylor et al. 2012) uses multi-model ensembles to sample those uncertainties, providing a basis to quantify the responses of the climate system to plausible natural and anthropogenic forcing scenarios. Nevertheless, some studies (e.g. Knutti et al. 2009; Knutti and Sedlacek 2013) suggest that part of the model spread in climate projections is irreducible due to internal variability of the climate system even though the representations of relevant processes have been substantially improved in GCMs.

Internal climate variability, which is caused by internal processes and feedbacks within the coupled climate system, can substantially influence our understanding and interpretations of regional climate changes (e.g., Deser et al. 2012a; Wettstein and Deser 2014; Swart et al. 2015; Kay et al. 2015). In the climate modeling framework, the uncertainty caused by internal variability of a specific GCM can be probed through an ensemble of the fully-coupled GCM comprising dozens of ensemble members with different initial conditions (Deser et al. 2012b; Kay et al. 2015). However, only a few climate modeling groups have conducted large-ensemble experiments to estimate the internal climate variability of their GCMs (e.g., the Community Earth System Model, version 1 (CESM1), Kay et al. 2015; the Second Generation Canadian Earth System Model (CanESM2), Kirchmeier-Young et al. 2017). In addition, the role of internal climate variability has not been explicitly characterized in the current CMIP5 because many models only contributed one realization. As a result, inter-model spread in the CMIP5 is difficult to interpret and to attribute it correctly to either model structural uncertainty or internal variability (Knutti et al. 2009).

Several studies have investigated the response of the AIS to future climate trajectories under different levels of GHG emissions using the CMIP5 results (Golledge et al. 2015; Trusel et al. 2015). However, as the role of internal variability is not well represented within CMIP5, the uncertainty of projecting the AIS responses to future climate using the CMIP5 results might be underestimated. A recent study showed that the internal variability of GCMs can affect the Greenland Ice Sheet (GrIS) evolutions and its contribution to future sea level changes (Tsai et al. 2017). However, the effect of internal climate variability on the AIS changes for the upcoming centuries is still unclear. Therefore, considering that the representation of uncertainty associated with initial conditions will be improved in the future CMIPs, in this study, we quantify how much internal climate variability

may influence the AIS evolutions and the resulting sea level changes. Additionally, we explore the contribution of internal variability from atmospheric and oceanic fields on the AIS changes, respectively, to identify which component is more responsible for the overall impact of internal variability on the AIS changes. For the following content of this paper, we introduce the large-ensembles of GCM and the ice sheet model we use in Sects. 2.1 and 2.2. The experimental design for exploring the impact of internal climate variability on the AIS is described in Sect. 2.3. We present and discuss the internal climate variability over Antarctica and its role in affecting the AIS evolutions in Sect. 3. The goal of this study is to assess the sensitivity of the AIS response to the uncertainty caused by internal climate variability. Such assessments will become more feasible and prominent as the use of LEs with different initial conditions becomes more prevalent in future GCM studies and intercomparison projects. Better understanding the role of internal climate variability in influencing the AIS responses to future warming signal can improve the overall uncertainty quantification of ice sheet contributions to future SLR. In addition, this can aid decision-makers in adopting suitable adaptation strategies and the climate and cryosphere science communities in developing fully coupled climate-ice sheet models.

2 Methods

We follow the experimental design in Tsai et al. (2017), in which the impact of internal variability of the CESM on the GrIS is assessed. Here, we use the climate fields from two sets of large-ensemble (LE) experiments of the CESM, which represent the internal climate variability of this particular model, to force a three-dimensional Antarctic ice sheet-shelf model (Pollard and DeConto 2009, 2012).

2.1 Two CESM large initial-condition ensemble experiments

To incorporate the effects of internal climate variability on projecting future AIS changes, we use the climate fields from two LE experiments with the CESM, to study internal climate variability explicitly. In general, to sample internal climate variability of a single GCM, multiple ensemble members are generated with different initial conditions, but they all undergo the same external forcings. Due to the chaotic nature of the climate system, each ensemble member yields slightly different responses due to internally generated variability.

The first LE experiment (Srивer et al. 2015, hereafter the “SFK LE”) uses the fully coupled CESM 1 consisting of atmosphere, land, sea ice and ocean components. For this version of the CESM, the Community Atmosphere Model

(CAM) and Community Land Model (CLM) version 4 are run at spectral T31 grid resolution ($\sim 3.75^\circ \times 3.75^\circ$); the ocean model (the Parallel Ocean Program version 2 (Smith et al. 2010)) is run with horizontal resolution about 3° . The SFK LE is based on a 10,000-year preindustrial control simulation, using the fixed preindustrial forcing at 1850. Starting from the year 4200 of the control simulation, the initial conditions are obtained from the snapshots of climate fields every 100 years for branching 50 transient runs. This ensures that the climate system has reached a near equilibrium state. Unlike the experimental designs in other studies (e.g., Deser et al. (2012b); Kay et al. (2015)), the SFK LE samples the initial-condition uncertainty due to different states of all model components (i.e., atmosphere, ocean, land, and sea ice). Each of the 50 branch runs undergoes identical historical natural and anthropogenic forcings from 1850 to 2005 and the Representative Concentration Pathway (RCP) 8.5 (Moss et al. 2010; Meinshausen et al. 2011) from 2006 to 2100. To assess long-term contribution of internal climate variability to the AIS changes, the SFK LE has been extended to 2300 with Extended Concentration Pathways (ECP) following the RCP8.5 forcings (Tsai et al. 2017). In using a low-resolution CESM, the experimental framework of the SFK LE trades off resolution for better sampling of initial-condition uncertainty with more ensemble members.

For the second LE experiment, we use the output of the CESM Large Ensemble Project (Kay et al. (2015), hereafter the “NCAR LE”), in which the CESM version 1 is run with the CAM version 5.2 (CESM1(CAM5)) (Hurrell et al. 2013). The CESM1(CAM5) is also a fully coupled climate model that includes atmosphere, ocean, land, and sea ice models, and participated in the CMIP5 (Taylor et al. 2012). In this LE, the atmosphere and land models are run at $0.9^\circ \times 1.25^\circ$ finite volume grid and the ocean model is run at approximately 1° grid. Unlike the SFK LE, the initial states of NCAR LE members are distinguished only by small perturbations to the atmosphere and not the ocean; however as the simulations proceed, internal variability propagates into the ocean, on decadal time scales for the upper layers. Specifically, the NCAR LE begins with the 1850 control simulation with constant preindustrial forcing for a few centuries. After the climate in the control run has reached quasi-equilibrium state with the 1850 forcing, the first ensemble member was started with initial conditions extracted from a random date in the 1850 control run, and it was integrated from 1850 to 2100. However, the rest of the ensemble members are branched from a slightly different initial atmospheric state, in which the initial air temperatures are randomly perturbed at the level of round-off error (about 10^{-14} K). Following the protocol of LE experiments for studying internal climate variability, each ensemble member in the NCAR LE undergoes the same external forcing, with

the historical forcing applied during 1920–2005 (Lamarque et al. 2010) and RCP8.5 forcing applied from 2006 to 2100.

Although the experimental designs are different between the SFK LE and the NCAR LE, they both provide ensembles of opportunity for studying the effects of initial-condition uncertainty on climate projections simulated by a single GCM. By using the climate fields simulated from the two LEs, our study specifically focuses on the AIS responses to internal variability under high greenhouse gas emissions scenario simulated from two differently configured, fully coupled climate systems. Other types of uncertainties, such as those from different climate model structures, resolutions, parameters, and forcing scenarios, are not considered in this study.

2.2 The 3-dimensional ice sheet-ice shelf model

We use a three-dimensional ice sheet-shelf model (Pollard and DeConto 2012, hereafter the “PSU ISM”) to estimate the AIS response to internal climate variability from two LEs. The model has been used extensively for studying past and future ice sheet changes over Antarctica (e.g. Pollard and DeConto 2009, 2012; Pollard et al. 2015; DeConto and Pollard 2016) and Greenland (Koenig et al. 2011). The PSU ISM uses vertically averaged ice dynamics as described in Pollard and DeConto (2012) and predicts the evolution of ice thickness, internal ice temperature, and the bedrock deformation in response to mass balance changes from precipitation, runoff, basal melting, oceanic sub-ice melting at vertical ice faces, and ice-shelf calving. The vertically averaged ice dynamics and parameterized grounding-line flux capture grounding-line migration while still allowing large-scale long-term runs to be computationally feasible. The model performs reasonably well in idealized inter-comparisons (Pattyn et al. 2012, 2013), although with somewhat larger differences on smaller space and time scales (Drouet et al. 2013; Pattyn and Durand 2013). The PSU ISM incorporates mechanism of MISI accounting for grounding-line migration due to reduced buttressing from ice shelves, which can cause rapid WAIS retreat due to sub-ice oceanic melting. Furthermore, additional physical mechanisms (i.e., MICI) have been added to the PSU ISM recently (Pollard et al. 2015; DeConto and Pollard 2016) to characterize (1) the increased ice-shelf calving from hydrofracturing due to surface melt and rainfall draining into crevasses on the ice shelf and (2) the large ice-cliff failure due to unbalanced overburden of ice above the ocean surface without ice-shelf buttressing. By incorporating these new physical treatments of MICI into the PSU ISM, the WAIS retreats more rapidly along with MISI, triggered by atmospheric warming.

For all ice sheet model simulations, the PSU ISM is run at spatial resolution of 20 km. The PSU ISM is driven by the transient climate fields simulated from two LEs, i.e.,

monthly 2m air temperature (T_{2m}), monthly precipitation (P), annual ocean temperature at 400m level (OT_{400m}) and the fixed surface elevations in the CESM. Ocean temperatures at 400 m are used to represent typical depths of warm Circumpolar Deep Water (CDW) water masses that flow onto the Antarctic continental shelf and under ice shelves (e.g., Pollard et al. 2015). These climate fields from the GCM are then interpolated onto the ice sheet model grid using bilinear spatial interpolation. To account for the differences between interpolated surface elevations in the CESM (h_s^{gcm}) and those in the PSU ISM (h_s^{ism}), the T_{2m} and P fields are lapse-rate corrected as follows:

$$T_{cor} = T_{gcm} - \gamma(h_s^{gcm} - h_s^{ism}) \quad (1)$$

$$P_{cor} = P_{gcm} \times 2^{\frac{T_{cor} - T_{gcm}}{\Delta T}} \quad (2)$$

where T_{cor} , P_{cor} are the lapse-rate corrected T_{2m} and P fields; T_{gcm} , P_{gcm} are the interpolated T_{2m} and P fields simulated from CESM; $\gamma = 0.008 \text{ } ^\circ\text{Cm}^{-1}$, which is the lapse rate accounting for the decrease in air temperature with elevated altitude; ΔT is $10 \text{ } ^\circ\text{C}$.

To remove potential bias in the climate fields from the CESM, the PSU ISM uses an anomaly method (Pollard 2010) to adjust the time-varying T_{2m} , P (T_{exp} , P_{exp}) and OT_{400m} fields from both LEs using the following equations:

$$T = T_{exp} + T_{obs} - T_{ctl} \quad (3)$$

$$P = P_{exp} \times P_{obs}/P_{ctl} \quad (4)$$

in which the ensemble mean averaged over 1920–2012 (T_{ctl} , P_{ctl}) from each LE is taken as the model modern control, and T_{obs} , P_{obs} are the observed modern climatology from the ALBMAP data set (Le Brocq et al. 2010). The OT_{400m} fields are adjusted in the same way as T_{2m} , using Eq. (3) and a modern ocean climatological dataset (Levitus et al. 2012).

To determine annual AIS surface mass budgets, the PSU ISM uses a simple surface mass balance model that involves seasonal cycles of zero-dimensional bulk quantities of snow and embedded meltwater, run through several years to equilibrium with a 5-day time step, driven by monthly T_{2m} and P interpolated in time to those time steps. A positive degree-day (PDD) scheme calculates the melt of snow or exposed ice at each timestep, with coefficients of 3 and 8 kg m^{-2} per degree (C) day for snow and ice, respectively, and with a uniform normal distribution representing diurnal cycles and synoptic variability. Precipitation is taken to be either rain or snow when air temperature T_{2m} is above or below $0 \text{ } ^\circ\text{C}$. After seasonal equilibrium is reached, net annual quantities are used to calculate refreezing of meltwater, and runoff of excess meltwater once the snow is saturated. More sophisticated 1-D column models of the snow-firn transition

are beginning to be used but are computationally expensive (e.g., Munneke et al. 2014); however, in future work, they should be incorporated to provide a more robust estimates of runoff and net surface mass balance. To calculate the oceanic sub-ice shelf melt rate, the PSU ISM uses a parameterization linking the oceanic melt rates to OT_{400m} at the nearest ocean model grid point. Sub-ice shelf melt rates are parameterized as a quadratic function of the difference between OT_{400m} and the pressure-melt point at the ice base (Pollard et al. 2015, Secti. S3, with the coefficient as in Pollard and DeConto 2012, Eq. 17 using $K = 3$ for all sectors). This parameterization accounts for the decrease of the melt point with depth, and results in higher sub-ice shelf melt rates at the grounding line.

2.3 Experimental design

2.3.1 Exp. 1—Impact of internal climate variability on Antarctic Ice Sheet changes

To quantify the effects of internal climate variability on AIS response, we set up the first experiment (Exp1), in which we force the PSU ISM with the monthly T_{2m} , monthly P and annual OT_{400m} fields from two LEs for all individual years (i.e., 1850–2300 for the SFK LE and 1920–2100 for the NCAR LE, see the ensemble-mean of annual T_{2m} and annual OT_{400m} fields, and surface temperature, ice thickness and mass balance output from the PSU ISM in Figs. S1 and S2). Before incorporating the transient climate forcing fields from LEs, the PSU ISM is initialized with the first decadal mean of ensemble-mean climate forcing fields (i.e., 1850's for the SFK LE and 1920's for the NCAR LE) for 200,000 years (Fig. S3c) to equilibrate the ice-sheet states with these climates. Until the end of initialization, we note that both SFK and NCAR LEs exhibit slightly thicker ice over the WAIS and thinner ice over the EAIS compared to the modern observations (Fretwell et al. 2013); however, reasonably small departures from the modern state (Fig. S3b) are not critical because the main focus here is on the future impact of internal climate variability on the AIS, and not on providing highly accurate modern AIS simulations or future projections. For the transient ensemble runs forced by two LEs, all simulations are started from ice-sheet states at the end of initialization (year 200,000; Fig. S3a). In this way, each of the transient PSU ISM ensemble runs uses the identical quasi-equilibrated ice-sheet states as the initial conditions so that the differences between simulated AIS changes only reflect the effects of internal climate variability from two LEs.

After initializing the PSU ISM, we generate two ice-sheet-model ensembles using individual atmospheric and oceanic forcings representing transient climate changes with internal variability under historical and the RCP 8.5

forcings simulated from two LEs. Additionally, considering that the ensemble-mean climate typically represents the climate change signal from a specific GCM, we further investigate the ice sheet's response to the ensemble-mean climate forcings averaged over each LE. In all, the first ice-sheet-model ensemble consists of 51 members, with 50 members of climate fields from the SFK LE and the ensemble-mean climate; similarly, the second ensemble has 41 members from the NCAR LE (Table 1). By comparing the results of the ice sheet's response to the ensemble-mean climate forcing and the mean ice sheet response to the individual LE members, we can quantify the effect of internal climate variability on ice sheet evolution. Hence, in this study, we examine whether using the ensemble-mean climate forcing fields as a driving force in the ice sheet model are appropriate for representing the climate change signal from a given GCM. This is a salient research question as climate modeling groups are increasingly using large ensembles to characterize uncertainty associated with initial conditions.

2.3.2 Exp. 2—Contribution of atmospheric vs. oceanic internal variability on the Antarctic Ice Sheet changes

Considering that the PSU ISM is driven by both atmospheric and oceanic forcing, we assess the relative contribution of internal variability from atmospheric and oceanic fields on simulating the AIS changes (Table 2). In this experiment, we set up two large-ensemble ice sheet model simulations using climate fields from the SFK LE. Unlike Exp1, for each PSU ISM ensemble consisting of 50 members, the internal variability from atmospheric (T_{2m} and P) and oceanic (OT_{400m}) fields are considered separately in Exp2. For the first component in Exp2 (hereafter, "ExpATM"), we explicitly explore the impact of internal variability from atmospheric fields on the AIS changes. In the ExpATM, each of the 50 runs is forced with the individual atmospheric fields obtained from 50-member SFK LE; however, the oceanic forcing is identical across all 50 runs and it is provided from a randomly selected ensemble member in the SFK LE (ens#22). In the second component of Exp2 (hereafter, "ExpOCN"), the effects of internal variability from oceanic field are assessed specifically. As opposed to the experimental design of the ExpATM, all 50 PSU ISM runs in the ExpOCN are forced with corresponding 50 different oceanic forcings from the SFK LE but with the identical atmospheric forcings from ens#22. The results from ExpATM and ExpOCN will be compared with the results of SFK LE in Exp1 (hereafter, "ExpBAS") to assess the relative contribution of internal variability on AIS changes from atmospheric and oceanic fields.

3 Results and discussions

3.1 Internal climate variability over Antarctica

We explore the influence of internal climate variability on Antarctica's climate by examining austral summer (December–February, DJF) T_{2m} trends (Figs. S4 and S5), precipitation (*precip*) trends (Figs. S6 and S7) and annual OT_{400m} trends (Figs. S8 and S9) from 2006–2100 in two LEs. The internal climate variability of climate trends ($\Delta X'$) can be obtained by subtracting the ensemble-mean trend ($\overline{\Delta X}$) from the climate trend of each ensemble member (ΔX) individually:

$$\Delta X' = \Delta X - \overline{\Delta X} \quad (5)$$

where ΔX can be the trend of any climate variable. For T_{2m} , both LEs demonstrate small but distinct spatial patterns of $\Delta T'_{2m}$, deviating from the ensemble-mean forced response by up to $\pm 2^\circ\text{C}$ at some places over Antarctica (Figs. S4 and S5, also see Figs. S6 and S7 in the Supplementary Material for precipitation). Several places in Antarctica are subject to larger $\Delta T'_{2m}$ and $\Delta \text{precip}'$ as illustrated in Fig. 1a, b. For instance, spatial variabilities of $\Delta \text{precip}'$ trends are larger along the AIS margin and spatial variabilities of ΔT_{2m} from both LEs are larger around part of East Antarctica; however, we expect that the influence of internal climate variability of T_{2m} on the East Antarctic Ice Sheet (EAIS) changes during 2006–2100 might be negligible because the mean T_{2m} has not become high enough to drive significant melt. In addition, compared to $\Delta T'_{2m}$, we expect $\Delta \text{precip}'$ has little effect on the AIS because it maintains a close-to-linear relation with surface mass accumulation even if snow changes to rain as long as rain refreezes in the ice sheet, while $\Delta T'_{2m}$ is expected to have a larger effect on ice-sheet budgets due to non-linearity of surface melt verses air temperature. For both LEs, relatively modest internal variability of ΔT_{2m} occurs around AP, WAIS, Ross Ice Shelf, and along the George V coast (Fig. 1a). Moreover, the internal variability of ΔT_{2m} from the NCAR LE exhibits stronger pattern of variability over the east side of AP, Ronne Ice Shelf, George V coast and Amery Ice Shelf possibly due to its finer model resolution. Considering that WAIS could melt significantly in the future based on current “business-as-usual” projections of warming, the influence of internal climate variability of T_{2m} might play an important role in projecting future WAIS changes.

Compared to the internal variabilities of T_{2m} trends, those of OT_{400m} trends from both LEs exhibit much smaller magnitude (Figs. 1c, S8 and S9 in the Supplementary Material), probably due to larger oceanic thermal inertia and the resulting weak sensitivity to distinct initial conditions. However, relatively stronger internal variability of OT_{400m} trends occur

in several places, such as Weddell Sea, Bellingshausen Sea, Ross Sea and Prydz Bay (Fig. 1c).

The areal-averaged summer T_{2m} anomalies over Antarctica from the two LEs during 2000–2099 (blue lines in Fig. 2a, b) show the model spread due to internal variability for up to $2\text{--}3^\circ\text{C}$ superimposed on the sustained warming trend. Compared to the CMIP5 results (pink lines in Fig. 2a, b), both LEs produce reasonable warming trends that fall within the range of the CMIP5 variability. While the trajectory of ensemble-mean T_{2m} anomaly from the SFK LE is similar to those from the CMIP5, the T_{2m} from the NCAR LE shows stronger warming trend during 2000–2099. Interestingly, the internal variability of T_{2m} does not change much as time evolves, while the multi-model variability of T_{2m} in the CMIP5 increases with the sustained warming trend. Over the first three decades, the internal variability from a single GCM can be more than half of the inter-model variability in the CMIP5 (vertical bars in Fig. 2a). However, the spread of T_{2m} due to internal variability appears to be independent of the warming trend, so that it would be relatively smaller compared to the increasing multi-model variability in CMIP5 after 2050. To isolate the impact of warming trend on interpreting T_{2m} variability, we subtract the warming trends defined by 3rd-degree polynomials from each of the ensemble members in both LEs and the CMIP5 (Fig. 2b). We notice that the magnitude of internal variability of T_{2m} does not decrease when the warming trends are removed. However, the magnitude of multi-model variability in the CMIP5 significantly decreases and becomes comparable to that of internal variability (comparing Fig. 2a, b). This implies that both multi-model differences and internal climate variability may contribute to the overall variability of near-future climate projections, at least until a strong warming trend emerges.

Knowing that the internal variabilities of two different versions of CESM persist and coevolve with the changing climate, we expect that the range of uncertainty of T_{2m} would ultimately widen the overall uncertainty if all climate models participating in the CMIP5 would perform additional initial-condition LE experiments. Although the multi-model uncertainty could be reduced through model evaluations and improvement, the uncertainty caused by internal variability of each single GCM could not be removed easily. Therefore, understanding the role of internal climate variability over Antarctica and quantifying its impact on simulating the AIS changes would improve uncertainty quantification of assessing future AIS mass loss and its contribution to SLR.

3.2 Responses of the Antarctic Ice Sheet to internal climate variability

Given that future warming of atmosphere and ocean can cause massive ice loss from the AIS, here we investigate

Fig. 1 Spatial variability (shown as one standard deviation) of 95-year trends of **a** DJF 2m air temperature (ΔT_{2m}), **b** precipitation ($\Delta precip$) and **c** ocean temperature (ΔOT_{400m}) from SFK LE and NCAR LE. Trends are calculated from 2006–2100 based on the linear least-squares estimation. The approximate locations of several ice sheets/glaciers and surrounding seas mentioned in the article are shown

Fig. 2 **a** Anomalous DJF T_{2m} time series from the SFK LE, NCAR LE (blue), 88 realizations from the CMIP5 RCP8.5 simulations (red) for Antarctica from 2000–2099 (Table 4 in Appendix A), and the ERA-Interim (black) from 2000–2016. The T_{2m} anomalies are with respect to the 1979–1999 mean and areal-averaged over the 60S–90S by 0–360E region. Vertical bars represent ± 1 standard deviation for each decade. **b** Time series of detrended Antarctica DJF T_{2m} among

two LEs and 88 realizations from the CMIP5 RCP8.5 simulations. The third-degree polynomial-fit of each ensemble member is subtracted from each original time series shown in **a**. The uncertainty (represented as vertical bars indicating ± 1 standard deviation for each decade) and spread of detrended T_{2m} time series between two LEs and CMIP5 are about the same magnitude

the role of internal variability of atmospheric and oceanic fields in the AIS changes from the two LE experiments. We estimate the spatial patterns of standard deviations of ice thickness (Fig. 4) from the SFK LE for different years, and from the NCAR LE for 2100. Although the impact of internal variability is mostly identified on short time scales (< 50 years), our PSU ISM results show that small internal climate variability can lead to different AIS responses and contributions to sea-level changes on decadal and

centennial scales. In 2100, the internal climate variability can cause different extents of ice loss over West Antarctica, such as Larsen C, Wilkins and George VI Ice Shelves at AP, Ronne Ice Shelf, Pine Island and Thwaites Glaciers, as well as several glaciers along the periphery of the EAIS. After 2150, the impact of internal climate variability on the AIS becomes more evident when the WAIS undergoes catastrophic collapse through MISI and MICI. For example, the variabilities of ice thicknesses over the Ronne Ice Shelf and

the ice shelves adjacent to Bellingshausen Sea, Amundsen Sea and Ross Ice Shelf are larger compared to previous years (Fig. 4). Unlike the WAIS, the inland EAIS is less affected by internal climate variability because most of the EAIS rests upon the bedrock above sea level and it is protected by sustained cold and dry climate. However, some EAIS submarine basins show sensitivity to internal variability after 2150, which is due to its impact on the MICI mechanism (DeConto and Pollard 2016).

The different trajectories of AIS contributions to global mean sea level changes (ΔGMSL) illustrates the integrated effect of internal climate variability from the two LEs on the AIS evolutions (Fig. 3a, d). Ranges of ΔGMSL estimated from the SFK LE and the NCAR LE demonstrate that the AIS is sensitive to variability from a fully coupled climate system. The internal climate variability can cause about 0.08 m differences in ΔGMSL until 2100 (Table 3 and Fig. 3c) for the SFK LE (orange) and the NCAR LE (light purple), which is about 18–21% of the total GMSL change during 2000–2100. Moreover, the spreads of AIS contributions to GMSL changes from both LEs are still slightly overlapped with each other until 2100. This implies that the impact of

internal climate variability could be comparable to climate model structural variability on the near-future (i.e., before 2100) AIS simulations using climate fields from differently configured CESMs. In sum, the results shown in Figs. 3 and 4 suggest that the small internal climate variability, combined with the warming trend, could trigger the onsets of ice retreat occurring in different years, resulting in different AIS contributions to GMSL among the ensemble members at a given year (Fig. 3 b, c, e–h).

Because the ensemble-mean climate is commonly used to represent the forced climate change signals of a given GCM in other studies, we compare the results of the AIS response to the ensemble-mean climate forcing and the mean AIS response for two LEs. Similar to the findings in Tsai et al. (2017) for the GrIS, the AIS contribution to GMSL can be significantly delayed if we use the time-varying ensemble-mean climate forcing fields as the climate forcings in the PSU ISM (see Fig. 3a, d). However, the impact of internal climate variability on ice sheet contribution to GMSL is more significant for the AIS than the GrIS. The estimated AIS contribution to GMSL using ensemble-mean climate forcings falls out of the range of all ensemble members at

Fig. 3 Time series of AIS contribution to ΔGMSL since 2000 simulated from the PSU ISM with the SFK LE forcing (orange) and the NCAR LE forcing (light purple) until 2100 (a) and 2300 (d). Each light color line represents individual model runs; the two dark lines show the SLR contribution simulated with the SFK LE-mean climate forcing (navy) and NCAR LE-mean climate forcing (black); the dash lines represent 1-sigma uncertainty estimate of internal variability. The violin plots represent the kernel density estimate of total GMSL contributions from the AIS melt from 2000 until 2050 (b), 2100 (c), 2150 (e), 2200 (f), 2250 (g), and 2300 (h) using LEs. The three dotted lines within the distribution represent the quartiles of the distribution. The spacings in e–h are the same

Fig. 4 Spatial standard deviation of ice thickness (m) for two LEs in 2100 (upper) and 2150, 2200 (middle), 2250 and 2300 (lower) for SFK LE. Values less than 0.5 m are masked

early stage of simulations (before 2050 as shown in Fig. 3a, b). However, for the GrIS, although its contribution to GMSL with ensemble-mean climate forcings also falls out

of one-standard-deviation range before 2050, it has not fallen out of the range of all ensemble members until 2100 (cf. Fig. 2 in Tsai et al. (2017)). Compared to the mean Δ GMSL

of two LEs, the simulated Δ GMSL using ensemble-mean climate forcings are ~ 0.11 m less for the SFK LE and ~ 0.07 m less for the NCAR LE in 2100 (Table 3). Moreover, for both SFK LE and NCAR LE, the Δ GMSL using ensemble-mean climate forcing are significantly different from the individual trajectories of Δ GMSL since about 2010. These results suggest that internal climate variability has potential for significantly influencing the AIS changes and their contributions to GMSL in a decade and up to hundreds of years.

Therefore, using the smooth ensemble-mean climate forcings of the two LEs cannot capture the effect of internal climate variability on ice sheet changes and reproduce the mean ice sheet's response. Instead, the ensemble-mean climate forcing filters out internal climate variability, and using it as the climate forcing in the ice sheet model underestimates the AIS contribution to GMSL. The diverse AIS responses to individual and ensemble-mean climate forcings from the SFK LE persist through the whole simulation until 2300 (Fig. 3h). This is especially true for the WAIS (Fig. 5) where differences in ice thicknesses and grounding lines between the mean ice-sheet response (black line) and the ice-sheet response to ensemble-mean climate forcing (cyan line) over the WAIS are plotted. The red shading indicates the excess extent of simulated ice thickness using ensemble-mean climate as the forcing fields from the SFK LE. Starting from 2050, larger differences in ice thickness occur at western AP, Pine Island, Thwaites glaciers and several places along Wilkes Land (Fig. 5a and Movie S1 in Supplementary Material) when the ice starts to retreat over those places. Following that, the Ronne and Ross Ice Shelves, and interior WAIS undergo significant differences in ice thicknesses as time advances to 2250's (Movie S1). These results demonstrate that the WAIS is sensitive to internal climate variability, which can potentially shift ice sheet's response to warming signal and affect the timing of significant retreat of the WAIS.

Additionally, the rate of GMSL changes also show the delayed AIS response to ensemble-mean climate forcing and identify the discrepancies of ice sheet collapse events at particular places and timing between the mean AIS response and the AIS response to ensemble-mean climate forcing (Fig. 5b and the movie in Supplementary Material). The arrows in Fig. 5b indicate the approximate timing when the ensemble-mean ice sheets at particular locations have initiated collapse while the ice sheets simulated with ensemble-mean climate forcings have not. For instance, during 2170–2200, rate of Δ GMSL increases due to significant ice retreat over the Amundsen Sea Embayment and Ronne ice shelf. However, this event is delayed for about 20 years in the simulation using ensemble-mean climate forcing. Similar

delayed ice sheet responses also occur when the interior WAIS starts to significantly retreat during 2230–2260. In summary, our results show that using ensemble-mean climate as the driving force can significantly delay the dynamic ice-sheet response, particularly over West Antarctica. Even though the deviations of internal climate variability among all individual runs are small in the LEs, they have substantial impact on ice sheet changes, triggering the onsets of WAIS retreat at significantly different times.

Because the internal climate variability from a single GCM can affect AIS changes, it can also influence the time when SLR reaches a particular threshold (Fig. S10). Depending on which ensemble or time frame, our results show 6–14 year windows for the earliest and latest cases to cross specific thresholds of SLR caused by internal variability of a single GCM (Fig. S10). The length of time window (Δt in Fig. S10) of reaching modest rise of sea-level (i.e., 0.1 or 0.2 m rise with respect to year 2000) tends to be longer than that for detecting more severe cases in the future (i.e., up to 14 years before 2100, and about 6–8 years after 2100). This result suggests that the internal climate variability could play an important role in quantifying overall uncertainty in estimating Δ GMSL due to AIS melt before 2100. Similarly, using the ensemble-mean climate as the forcings can delay Δ GMSL reaching particular levels for up to 15 years (circles in Fig. S10) compared to the mean Δ GMSL estimate averaged over individual ensemble members (stars in Fig. S10). Thus, we highlight the importance of better representing underlying uncertainty caused by internal climate variability for future ice sheet changes because this information can aid decision makers in quantifying uncertainty of SLR estimates and determining adaptation strategies.

3.3 Responses of the Antarctic Ice Sheet to different sources of internal climate variability

As internal climate variability can lead to distinct ice sheet responses, we explore which component of internal variability (i.e., atmosphere or ocean) dominates the overall effect of internal climate variability on the AIS changes. Variability of oceanic warming and melting under the ice shelves, and variability of atmospheric warming and surface-meltwater-enhanced calving, may both have important influence on future AIS changes. Comparing the results of ExpBAS (i.e., the SFK LE in Exp1), ExpATM and ExpOCN, we find that the contribution of internal climate variability on future AIS changes is mostly attributed to the internal variability of atmospheric fields (Fig. 6). The two-sigma ranges of variability of Δ GMSL (Fig. 6a) are similar for ExpBAS (grey)

Fig. 5 **a** Differences between ensemble-mean ice thickness over WAIS using 50 SFK LE runs and the ice thickness simulated using SFK LE-mean climate fields for 2095, 2155, 2185 and 2239. Red shaded regions indicate excess ice thickness simulated using SFK LE-mean climate (see “Movie S1” in Supplementary Material for the complete simulation period for Antarctica). Also shown are the ensemble-mean grounding line (black) averaged over 50 runs, the

grounding line simulated using SFK LE-mean climate (cyan), and the grounding line in 2000 (grey). **b** Rate of ΔGMSL for two LEs and the approximate timings at particular locations when the ice thicknesses differ significantly between the mean AIS response (red/blue lines) and the ice sheet response to ensemble-mean climate forcing (dark red/navy lines)

Fig. 6 **a** Time series of AIS contribution to Δ GMSL for control experiment (ExpBAS, black), atmospheric variability experiment (ExpATM, red) and oceanic variability experiment (ExpOCN, blue) from 2000 to 2100. **b** Histograms of Δ GMSL from the AIS mass loss from 2000 to 2100 (upper) and from 2000 to 2300 (lower) for Exp-

BAS, ExpATM and ExpOCN. Standard deviation of ice thickness from ExpBAS, ExpATM and ExpOCN in 2100 over **c** Amundsen Sea embayment and **d** the coast of East Antarctica. Values less than 0.5 m are masked. The plots for all Antarctica in 2100, 2200 and 2300 are shown in Fig. S11

Table 1 Summary of large-ensemble CESM used in Exp. 1

	Number of ensemble members	Transient simulation period	CESM spatial resolution
SFK LE	50+1	1850–2300	$\sim 3.75^\circ$
NCAR LE	40+1	1920–2100	$\sim 1^\circ$

Table 2 The ensemble members of the SFK LE used in Exp. 2

	T2m	Precip	OT400
ExpBAS	All 50 runs	All 50 runs	All 50 runs
ExpATM	All 50 runs	All 50 runs	ens#22
ExpOCN	ens#22	ens#22	All 50 runs

Table 3 The estimates of mean AIS response, the simulation with ensemble-mean climate forcing, difference (mean AIS response–ensemble-mean climate forcing case) and ranges (within each LE) of the Antarctic Ice Sheet contributions to Δ GMSL (unit: m) for the SFK LE and NCAR LE in different years

Year	LE Exp	Mean AIS response	Ensemble-mean climate forcing	Difference	Range
2050	SFK LE	0.08	0.04	0.04	0.04
	NCAR LE	0.07	0.04	0.03	0.04
2100	SFK LE	0.38	0.27	0.11	0.08
	NCAR LE	0.45	0.38	0.07	0.08
2150	SFK LE	1.18	1.00	0.18	0.18
2200		2.62	2.36	0.26	0.21
2250		4.44	4.11	0.34	0.22
2300		6.07	5.78	0.29	0.19

Note that Δ GMSL are calculated with respect to the year 2000

and ExpATM (pink), whereas the Δ GMSL range in ExpOCN is much smaller. In addition, the ranges of Δ GMSL distributions in ExpBAS and ExpATM are comparable in 2100, 2200, and 2300; however, the range of Δ GMSL in ExpOCN is much narrower (Fig. 6b). This suggests that the internal variability from atmospheric fields dominates the overall contribution of internal climate variability on Δ GMSL due to AIS melt at least until 2300.

The spatial patterns of ice thickness variability from the three experiments illustrate the locations where the ice sheets are susceptible to the impact of internal variability from either atmospheric or oceanic fields (Fig. 6c, d and S11). Similar patterns of ice thickness variability between ExpBAS and ExpATM imply that the internal variability from atmospheric fields is the major contributor of overall internal climate variability over most of the WAIS and the periphery of EAIS. As atmospheric warming becomes the dominant driver of future Antarctic ice loss through the MICI mechanism (Pollard et al. 2015; DeConto and Pollard 2016), the AIS changes tend to be more sensitive to the internal variability from atmospheric fields rather than oceanic fields. Nevertheless, even though the atmospheric variability is mostly responsible for the overall impact of internal climate variability on AIS changes, the internal variability from oceanic fields can strongly influence the changes of marine-based ice sheets. For example, in ExpOCN, larger ice thickness variability occur around Western AP, Thwaites Glaciers, parts of WAIS and Amery ice shelf (Figs. 6c and S11 a), while ocean variability does not affect the ice sheet changes over most of the East Antarctica (Fig. 6d). By comparing the ratio of ice thickness variability in ExpOCN to that in ExpATM (Fig. S12), the contribution of internal variability from oceanic fields is apparent over some of the marine-based ice sheets in West Antarctica before 2150, and over the Ross Ice Shelf throughout the whole simulation

period. This is consistent with previous studies suggesting that the WAIS, Amery ice shelf and western AP are vulnerable to ocean forcing (Shepherd et al. 2004; Fricker et al. 2001; Cook et al. 2016). However, our results indicate that internal variability from atmospheric fields play a leading role on the overall effect of internal climate variability on ice sheet changes over most of the Antarctica (Fig. S12). This is consistent with DeConto and Pollard (2016), who found that the effect of atmospheric forcing on AIS evolution is larger than oceanic forcing and the EAIS basins are more sensitive to atmospheric fields under MICI mechanism after 2200.

4 Summary

Uncertainty in future climate projections, and insufficient understanding of ice-sheet dynamical response to the future climate, limit the ability to provide robust estimates of future sea-level changes and their associated uncertainty. A complete assessment of uncertainty in future SLR requires identifying and quantifying various sources of uncertainties in climate and ice sheet models. Many studies use climate fields from climate models as the forcing fields to project future ice sheet changes. Simulated climate fields contain uncertainty associated with differences in emission scenarios, model structures and internal variability, which can ultimately contribute to uncertainty in projecting future SLR due to the AIS mass loss. Therefore, quantifying those uncertainties and their impact on AIS changes is critical for robust assessment of future SLR because rising sea levels threaten large population in coastal regions and monsoon areas (Defrance et al. 2017). Although the model structural uncertainty could be reduced as models are improved in the future, the uncertainty caused by internal variability among each climate model is irreducible. However, the internal variability of each GCM could be quantified and compared (Hawkins and Sutton 2009; Yip et al. 2011; Sansom et al. 2013) if climate modeling groups conduct initial-condition LE experiments of their GCMs.

Using two LEs of CESM that are designed for sampling initial-condition uncertainty, we assess the uncertainty in projecting future AIS changes caused by internal climate variability of the CESM. The internal climate variability, though it is relatively small compared to the ensemble-mean climate signal, exhibits notable deviation from the mean climate trends over some parts of Antarctica. Our results demonstrate that the internally generated climate variability of a single fully-coupled GCM can lead to significant differences in AIS responses. In our simulations, Δ GMSL estimates due to AIS loss vary because the internal climate variability triggers runaway collapse of submarine basins at different times, due to the threshold behavior of retreat (Pollard et al. 2015; DeConto and Pollard 2016). The ensemble spread of Δ GMSL trajectories indicates

that the overall uncertainty in the AIS contributions to SLR projections could broaden if the internal variability of each GCM is taken into account. From decision-making perspective, the uncertainty due to internal climate variability also affects timing of reaching specific ΔGMSL values due to AIS loss. Furthermore, atmospheric variability, rather than oceanic variability, dominates the impact of internal climate variability on the overall AIS changes, except for the marine-based WAIS where the contributions from both variabilities are roughly comparable, at least till 2100. While the ensemble-mean climate of a climate model LE is widely regarded as appropriate forcing for ice sheet model, we emphasize that employing the relatively smooth ensemble-mean climate as the forcing fields in this way can significantly underestimate the AIS contributions to SLR.

Our study aims to demonstrate that future AIS responses can be sensitive to relatively small internal climate variability of a GCM. However, conducting a complete assessment of uncertainty caused by internal climate variability in the CMIP5 models, for instance, is challenging because it would require multiple simulations characterizing internal variability of each GCM participating in the CMIP5. Because we only consider the internal variability of two CESM LEs, our results cannot represent the full range of possible responses caused by internal variabilities of all GCMs in the CMIP5. Additionally, we emphasize that the magnitude of effect of internal variability on SLR projections due to AIS mass loss is based on the estimates of a single but well established ice sheet model, which implemented the updated MICI mechanism for Antarctica. In both Pollard et al. (2015) and DeConto and Pollard (2016), the MICI mechanism is dominant (compared to slower-acting sub-ice oceanic melt) in warmer climates as the ice sheet responds sensitively to air temperatures (via surface melt causing hydrofracturing of ice shelves). With the MICI mechanism implemented in the ice sheet model, the effect of future ICV is manifest. Other ice sheet models without MICI might be less sensitive to internal variability and might produce smaller spread of SLR projections. Hence, the results here should be regarded as upper-bound estimates of the effects of ICV on future Antarctic retreat; if MICI is not active, then ICV would have lesser effects mainly via ocean temperatures and sub-ice ocean melting. Some caveats exist in our study, such as the coarse resolution and bias of climate fields in the CESM. Also we use one-way coupling instead of a fully coupled ice sheet-climate modeling framework, and so neglect important feedbacks, for instance, between melt-water discharge and ocean dynamics (Vizcaíno et al. 2010), between ice-sheet configuration and atmospheric circulation (Holden et al. 2010) and sea-level feedback (Gomez et al. 2015). However, we seek to initiate better assessment of uncertainty in future SLR due to the range of AIS responses to internal variability of different climate models. As other studies (e.g., Fyke et al. 2014; Vizcaino et al. 2015) have shown that variability of ice sheet surface mass balance may

increase with future warming climate, our study provides a complementary assessment of AIS contributions to future sea-level changes reflecting the combined effect of increasing ice sheet variability and internal climate variability represented by large-ensemble simulations. Considering that both climate and glaciology communities have been developing coupled ice-sheet-climate models (Vizcaíno et al. 2013; Vizcaíno 2014; Ziemer et al. 2014; Lenaerts et al. 2016; Nowicki et al. 2016), understanding the role of internal climate variability on ice sheet evolutions is important because coupled climate models with interactive ice sheet component are also subject to the uncertainty caused by internal climate variability. In our study, we seek to provide insightful uncertainty estimates related to internal climate variability for ice-sheet-sourced sea-level projections and we recommend that the impact of internal climate variability on ice sheet changes should be considered in future Ice Sheet Model Intercomparison Project and CMIP. In addition, quantifying the contribution of internal climate variability to the AIS changes and its associated uncertainty also helps policy makers conduct more robust estimates of SLR and identify suitable adaptation strategies.

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Appendix A

Table 4 List of the CMIP5 coupled models used in this study

Model	Modeling center	Ensemble members
ACCESS1.0	CSIRO-BOM	1
ACCESS1.3		1
BCC-CSM1.1	BCC	1
BCC-CSM1.1(m)		1
BNU-ESM	GCESS	1
CanESM2		5
CCSM4	NCAR	6
CESM1(BGC)		1
CESM1(CAM5)	NSF-DOE-NCAR	3
CESM1(WACCM)		3
CMCC-CESM	CMCC	1
CMCC-CM		1
CMCC-CMS		1
CNRM-CM5		5
CSIRO-Mk3.6.0	CSIRO-QCCCE	10
EC-EARTH		2
FGOALS-g2	LASG-CESS	1
FIO-ESM		3
GFDL-ESM2G	NOAA GFDL	1
GFDL-ESM2M		1
GISS-E2-H	NASA GISS	5
GISS-E2-H-CC		1
GISS-E2-R		5
GISS-E2-R-CC		1
HadGEM2-AO	NIMR/KMA	1
HadGEM2-CC		3
HadGEM2-ES	MOHC	3
INM-CM4		1
IPSL-CM5A-LR	INM	4
IPSL-CM5A-MR		1
IPSL-CM5B-LR	IPSL	1
MIROC-ESM		1
MIROC-ESM-CHEM	MIROC	1
MIROC5		3
MPI-ESM-MR	MPI-M	1
MPI-ESM-LR		3
MRI-CGCM3	MRI	1
MRI-ESM1		1
NorESM1-M	NCC	1
NorESM1-ME		1

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