Arctic Radiative Fluxes: Present-Day Biases and Future Projections in CMIP5 Models

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ABSTRACT

Radiative fluxes are critical for understanding the energy budget of the Arctic region, where the climate has been changing rapidly and is projected to continue to change. This work investigates causes of present-day biases and future projections of top-of-atmosphere (TOA) Arctic radiative fluxes in phase 5 of the Coupled Model Intercomparison Project (CMIP5). Compared to Clouds and the Earth’s Radiant Energy System Energy Balanced and Filled (CERES-EBAF), CMIP5 net TOA downward shortwave (SW) flux biases are larger than outgoing longwave radiation (OLR) biases. The primary contributions to modeled TOA SW flux biases are biases in cloud amount and snow cover extent compared to the GCM-Oriented CALIPSO Cloud Product (CALIPSO-GOCCP) and the newly developed Making Earth System Data Records for Use in Research Environments (MEaSUREs) dataset, respectively (with most models predicting insufficient cloud amount and snow cover in the Arctic), and biases with sea ice albedo. Future projections (2081–90) with representative concentration pathway 8.5 (RCP8.5) simulations suggest increasing net TOA downward SW fluxes (+8 W m⁻²) over the Arctic basin due to a decrease of surface albedo from melting snow and ice, and increasing OLR (+6 W m⁻²) due to an increase in surface temperatures. The largest contribution to future Arctic net TOA downward SW flux increases is declining sea ice area, followed by declining snow cover area on land, reductions to ice sea albedo, and reductions to snow albedo on land. Cloud amount is not projected to change significantly. These results suggest the importance of accurately representing both the surface area and albedos of sea ice and snow cover as well as cloud amount in order to accurately represent TOA radiative fluxes for the present-day climate and future projections.

1. Introduction

The Arctic region is warming at a rate nearly twice that of the global average (Serreze et al. 2009; Screen and Simmonds 2010). This “Arctic amplification” is a robust feature of global climate change, occurring not only in recent climate but also in paleoclimate records (Masson-Delmotte et al. 2014) and in climate model projections of the future (Winton 2006). Arctic amplification has been attributed to several processes and feedbacks, including the ocean–sea ice albedo feedback (Curry et al. 1995; Serreze and Barry 2011), the land–snow albedo feedback (Ghatak et al. 2010; Loranty et al. 2014), the water vapor feedback, cloud feedbacks (Screen and Simmonds 2010), and the lapse rate feedback (Pithan et al. 2014). Changes to poleward heat transport in the atmosphere and ocean may also contribute to Arctic amplification (Langen and Alexeev 2007; Graversen and Wang 2009; Döscher et al. 2010; Yang et al. 2010).

Modeling Arctic climate change that has occurred in recent decades and projecting future Arctic climate change is challenging because of the complexity of representing the processes associated with Arctic amplification and the...
sensitivity of radiative fluxes to differences in cloud, snow-extent, and sea-ice coverage and albedo. Hence, in order to accurately represent Arctic climate, models must accurately represent numerous components including surface type and albedo, cloud amount, and cloud phase. Additionally, the Arctic region suffers from a dearth of observations, particularly measurements at the surface. In recent years, satellite observations of Arctic basin–wide top-of-atmosphere (TOA) radiative fluxes and cloud amount have proven useful for climate model evaluation (Kay et al. 2011; Cesana et al. 2012; Barton et al. 2012; Cesana and Chepfer 2012; Xie et al. 2013; English et al. 2014). TOA fluxes are available from the latest version of the Clouds and Earth’s Radiant Energy System (Wielicki et al. 1996) product: the Energy Balanced and Filled data product (Loeb et al. 2009) (CERES-EBAF). Climate model comparisons to CERES-EBAF all-sky net TOA downward shortwave (SW) fluxes and outgoing longwave radiation (OLR) can identify overall energy balance biases, and comparisons of clear-sky and all-sky radiative fluxes can help identify the contributions of cloud and surface albedo biases. Although there remain uncertainties associated with CERES-EBAF clear-sky retrievals due to uncertainties in cloud clearing (Loeb et al. 2007), comparisons of modeled cloud amount with observations can add confidence to cloud radiative effect biases versus CERES-EBAF (English et al. 2014). However, there are challenges with cloud detection particularly by passive remote sensing instruments in the Arctic. Indeed, Stubenrauch et al. (2013) compared cloud detection by various satellites and found good agreement globally but significant disagreement in the Arctic. Chepfer et al. (2010) emphasized the utility of active remote sensing via the Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations–GCM-Oriented CALIPSO Cloud Product (CALIPSO-GOCCP), which is a part of the A-train (L’Ecuyer and Jiang 2010). Because of differences between modeled clouds and clouds observed by instruments, simulator packages such as the Cloud Feedback Model Intercomparison Project (CFMIP) (Bony et al. 2011) Observation Simulator Package (COSP) (Bodas-Salcedo et al. 2011), including a lidar simulator (Chepfer et al. 2008), are now commonly utilized to enable a more direct comparison of modeled clouds to observations. Assumptions are often made in the parameterizations that apply the simulator algorithms to model output, however, and there continue to be opportunities to improve the accuracy of using instrument simulators on model output. For instance, the Community Earth System Model version 1 (CESM1) version of COSP lidar simulator includes the effects of atmospheric snow crystals in the lidar simulator (Barton et al. 2012; Kay et al. 2012); the version used in the present study includes a correction for an error in the treatment of snow crystal size, which significantly reduced cloud amount in CESM1(CAM5) (English et al. 2014).

Comparisons of climate models to Arctic observations over the past three decades have revealed persistent challenges simulating Arctic climate that have only partially been remedied in the latest set of models in phase 5 of the Coupled Model Intercomparison Project (CMIP5) (Taylor et al. 2012). Arctic cloud cover varies significantly among the CMIP5 models regardless whether comparing COSP lidar cloud amount (Cesana and Chepfer 2012) or native cloud amount (Karlsson and Svensson 2013). The majority of the CMIP5 models have biases in Arctic cloud phase partitioning, particularly excessive cloud ice (Li et al. 2012) and insufficient cloud liquid (Komurcu et al. 2014), regardless of model biases in cloud amount. The cloud amount and phase partitioning biases in turn cause too much surface cooling in winter and spring (Brutel-Vuilmet et al. 2013; Liu et al. 2011), excessively strong inversions (Medeiros et al. 2011; Pithan et al. 2014; de Boer et al. 2013; Barton et al. 2014), and biases in net TOA radiative fluxes (Xie et al. 2013; English et al. 2014). Analysis of one model—CESM1(CAM5)—reveals that SW biases due to insufficient cloud amount are compensated by SW biases due to excessive snow cover and/or snow albedo on land and sea ice (English et al. 2014). The CMIP5 models disagree in their representation of summer sea ice albedo, which has been found to significantly impact the cloud radiative effect (Karlsson and Svensson 2013) and Arctic surface temperatures (Koenigk et al. 2014). Likewise, while the CMIP5 models correctly capture declining Arctic land snow cover over the past three decades, the rate of snow cover loss is less than that observed (Derksen and Brown 2012; Brutel-Vuilmet et al. 2013). Projections of surface warming and sea ice loss in the future also have enormous intermodel spread even when considering internal variability. For instance, the year at which the Arctic is projected to become sea ice free under the RCP8.5 scenario ranges from 2010 to 2090 in the CMIP5 models (Liu et al. 2013). Studies have attempted to reduce the uncertainty in sea ice projections by weighting models on their ability to represent observations of sea ice extent in recent decades (Liu et al. 2013; Snape and Forster 2014). Despite this progress, there remain opportunities to quantify the relative contributions of clouds, snow cover, and sea ice albedos on present-day radiative fluxes and future projections.

The primary goals of this paper are to quantify present-day Arctic net TOA radiative flux biases in the CMIP5 models and projected radiative flux changes in the RCP8.5 simulations. Radiative fluxes and clouds are
quantified over various surface types (land, sea ice, open ocean, land areas with snow cover, and land areas without snow cover) to understand the contributions of surface type to present-day biases in projected changes in the future. The relative contributions of changes in surface area (e.g., declining sea ice coverage) versus changes in surface albedo (e.g., melting snow to reveal lower-albedo surfaces) to projected radiative flux changes in the future are also quantified. Relationships are explored between model fields and present-day biases and future projected changes (such as model snow cover extent and net TOA downward SW clear-sky biases over land, for example). The paper is organized as follows: Section 2 describes methods for models, datasets, and analysis. Section 3 evaluates present-day biases in radiative fluxes and cloud amount in the CMIP5 atmospheric models compared to CERES-EBAF and CALIPSO observations. Section 4 evaluates projected changes in radiative fluxes and cloud amount for the CMIP5 coupled models under the RCP8.5 scenario. Discussion and conclusions are covered in section 5.

2. Methods

a. CMIP5 models

CMIP5 is the latest collaboration among the global climate modeling community to provide intercomparisons among models (Taylor et al. 2012). We analyze monthly averages from the first model run of each ensemble (r1i1p1) for nine CMIP5 models that have the required model output for this analysis: TOA radiative fluxes, cloud amount in the CMIP5 atmospheric models compared to CERES-EBAF and CALIPSO-GOCCP observations. Section 4 evaluates projected changes in radiative fluxes and cloud amount for the CMIP5 coupled models under the RCP8.5 scenario. Discussion and conclusions are covered in section 5.

<table>
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<th>Model</th>
<th>Notes</th>
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<td>CanAM4/CanESM2</td>
<td>CanAM4 (atmospheric component of CanESM2) used for present-day simulations; CanESM2 used for RCP8.5 projections</td>
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<tr>
<td>CESM1-CAM5</td>
<td>Includes correction for an error in COSP cloud fraction (English et al. 2014)</td>
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<td>CNRM-CM5</td>
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<td>GFDL CM3*</td>
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<td>HadGEM2-AO/HadGEM2-A</td>
<td>HadGEM2-AO (atmospheric component of HadGEM2-A) used for present-day simulations; HadGEM2-A used for RCP8.5 projections</td>
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<td>IPSL-CM5A-LR</td>
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<td>MIROC5*</td>
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<td>MPI-ESM-LR</td>
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<td>MRI-CGCM3</td>
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* Excluded from evaluations of land masks with and without snow cover because of lack of availability of snow cover field from CMIP5 archive.

(2081–90 minus 2006–15). Present-day climate is investigated with CMIP5 Atmospheric Model Intercomparison Project (AMIP) simulations. AMIP simulations prescribe sea ice extent and sea surface temperatures (SST) to allow for a more detailed study of the impacts of atmosphere, cloud, and surface albedo biases on TOA radiative fluxes by eliminating possible errors from biases in sea ice extent or SST. Additionally, analysis of AMIP simulations allows a more fair comparison to observations by forcing sea ice extent and SST to match the conditions under which the observations were taken, mostly eliminating natural variability. RCP8.5 projections are investigated with CMIP5 coupled simulations, where sea ice extent and SSTs are allowed to freely evolve. Two models—GFDL CM3 and IPSL-CM5A-LR—did not have snow cover on land output available and were excluded from analysis of the land areas with and without snow cover (lnds and lndns, respectively). COSP-CALIPSO cloud amount in CESM1 (CAM5) includes a correction for an error in the treatment of snow crystal size (English et al. 2014).

b. Net TOA radiative fluxes (CERES-EBAF 2.8)

CERES-EBAF is the most reliable source of basin-wide TOA radiative fluxes in the Arctic over the past decade, and newer versions have advanced to distinguish clouds from underlying high-albedo sea ice and snow cover by utilizing cloud radiances from the collocated Moderate Resolution Imaging Spectroradiometer (MODIS) and sea ice concentration fields from the National Snow and Ice Data Center (NSIDC) (Hollinger et al. 1990). We utilize CERES-EBAF version 2.8, which is the latest release as of March 2015 (Loeb et al. 2014a). We quantify the contributions of clouds and surface albedos to all-sky biases by comparing model output to CERES-EBAF clear-sky and cloud forcing biases. CERES-EBAF instantaneous all-sky error is estimated...
to be 10 W m\(^{-2}\) in the SW and 3–5 W m\(^{-2}\) in the LW (Loeb et al. 2007). A more useful measure of uncertainty for this comparison would consider regional error across annual averages. The CERES-EBAF Ed2.8 data quality summary (Loeb et al. 2014b) estimates regional-average SW all-sky and clear-sky errors to be roughly 4 and 2.6 W m\(^{-2}\), respectively, and in the Arctic, annual-average SW all-sky and clear-sky errors are generally 1–2 W m\(^{-2}\) except in a few locations where SW clear-sky errors are 5–10 W m\(^{-2}\). For both CMIP5 models and CERES-EBAF observations, we average across the years 2000 to 2008. This was done because the CMIP5 AMIP simulations typically end in the year 2008, while CERES-EBAF observations begin in the year 2000.

c. Cloud amount (CALIPSO-GOCCP)

We use CALIPSO-GOCCP lidar observations (Chepfer et al. 2010) to assess cloud amount. CALIPSO-GOCCP is well suited for Arctic assessments and has been utilized in other model-observation comparisons (Kay et al. 2012; Barton et al. 2012; English et al. 2014). CALIPSO-GOCCP cloud detection is unaffected by the surface conditions or thermal structure of the atmosphere and is therefore more reliable than passive remote sensing in the Arctic. CALIPSO-GOCCP has 330-m horizontal resolution and 30-m vertical resolution from the surface to 8-km altitude, and 1-km horizontal resolution and 60-m vertical resolution above 8 km. Hence, CALIPSO-GOCCP is able to detect near-surface clouds as well as optically thin clouds, which are both common in polar regions. CALIPSO-GOCCP cannot detect clouds with a scattering ratio of less than 5 (roughly equivalent to a cirrus cloud with an optical depth of less than 0.1) (Chepfer et al. 2013) and cannot see the entire vertical extent of clouds if the optical depth is greater than 5 (Winker et al. 2009). We compare CMIP5 model output to CALIPSO-GOCCP observations by utilizing the corresponding COSP CALIPSO lidar simulator for each CMIP5 model. Sensitivity studies with CALIPSO-GOCCP suggest uncertainty in cloud amount ranges from 0% to 6% (Chepfer et al. 2010). However, since we apply the same algorithm that is used by the instrument to model output, errors when comparing CALIPSO-GOCCP observations to the corresponding CMIP5 lidar simulator should be minimized.

We average CALIPSO-GOCCP observations from 2006 to 2010 and compare to CMIP5 model output averaged from 2000 to 2008; comparison across similar years was not possible because model output for historical simulations ends at 2008 for most CMIP5 models, while CALIPSO-GOCCP observations began in 2006. Analyses with the CMIP5 models did not find significant differences in cloud amount when averaging 2000–08 compared to 2006–10: basinwide average cloud amount changed by less than 0.2%, and change over each surface mask was less than 1%. Additionally, net TOA radiative flux biases between the CMIP5 models and CERES-EBAF observations differed by less than 1 W m\(^{-2}\) when averaging the years 2000–08 instead of 2006–08. Hence, we conclude that the offset between the CALIPSO-GOCCP temporal average (2006–10) and CMIP5 models and CERES-EBAF (2000–08) does not introduce significant inconsistencies for analysis.

d. MEaSUREs Northern Hemisphere terrestrial snow cover product

For this study we use a newly developed Northern Hemisphere terrestrial snow cover extent product from the NASA Making Earth System Data Records for Use in Research Environments (MEaSUREs) program (Robinson et al. 2014). This product integrates three snow cover variables at 25-km spatial resolution for the period 1999 through 2012, derived from the Interactive Multisensor Snow and Ice Mapping System (IMS), MODIS Cloud Gap Filled Snow Cover, and passive microwave brightness temperatures, respectively. We average the years 2000 through 2008 in the merged dataset, which combines all three individual snow products and identifies an area as snow covered if any of the three products (IMS, MODIS, or passive microwave) reported snow for a particular day. The advantages of the MEaSUREs snow cover product over other widely used products, such as snow cover extent from NOAA’s National Climatic Data Center (http://www.ncdc.noaa.gov), include high spatial resolution and the integration of three independent snow observation sources. Both the IMS and the merged snow cover representations were regridded using a bilinear Earth System Modeling Framework routine from a 25-km Equal-Area Scalable Earth (EASE) grid to the AMIP model grid for comparison with the CMIP5 radiative flux model output.

e. Surface mask designations

CMIP5 model output and observations are compared over consistently defined surface masks (Table 2): the entire Arctic basin, all land surfaces, sea ice, open ocean, land areas with snow, and land areas without snow. The Arctic “basin” is defined as the region between 60° and 82°N, which includes polar continental landmasses (most of Alaska and Greenland and the northern half of Canada and Siberia) as well as most of the Arctic Ocean. The region north of 82°N is excluded to provide a fair comparison between model output and CALIPSO-GOCCP observations, which extend to 82°N. Restricting the latitude range from 60°–90° to 60°–82°N did not
significantly affect conclusions, consistent with previous Arctic studies with the CESM1(CAM5) model (Kay et al. 2012; English et al. 2014) or CERES-EBAF observations (Kay and L’Ecuyer 2013). Hence, for consistency between radiative flux and cloud comparisons we conduct the majority of our analyses across the region 60°–82°N. All CMIP5 model output and observational data are regridded to 1.9° by 2.5° horizontal resolution via piecewise linear interpolation using the NCAR Command Language (NCL) software (NCAR Command Language 2014).

We separate all surfaces using a binary mask with a cutoff of 0.5. This was done so that all Arctic grid boxes were evaluated; that is, none of the Arctic region was excluded from the analysis. Model grid boxes are designated “land” if land fraction is greater than or equal to 0.5. Model grid boxes are designated “sea ice” if sea ice fraction is greater than 0.5 and “open ocean” if sea ice fraction is equal to or less than 0.5. Since several of the CMIP5 models use a rotated pole grid for sea ice extent, the sea ice data are converted to the standard rectilinear atmospheric grid via bilinear interpolation using the NCAR Command Language (NCL) software (NCAR Command Language 2014).

In this section, we evaluate present-day (2000–08) net TOA radiative fluxes and cloud amount from the CMIP5 models with prescribed sea ice extent and sea surface temperatures. Model output is compared to CERES-EBAF observed net TOA radiative fluxes and CALIPSO-GOCCP cloud amount. Cloud amount comparisons utilize the COSP lidar simulator for a more consistent comparison between modeled clouds and observed clouds. Later, we will compare CMIP5 coupled model future projections of radiative fluxes and cloud amount compared to present-day simulations under the RCP8.5 scenario.

### Table 2. Descriptions of surface masks analyzed.

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<thead>
<tr>
<th>Surface mask</th>
<th>Abbreviation</th>
<th>Gridbox requirements</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arctic basin</td>
<td>all</td>
<td>All grid boxes between 60 and 82°N</td>
</tr>
<tr>
<td>Arctic land</td>
<td>land</td>
<td>&gt;0.5 land fraction</td>
</tr>
<tr>
<td>Arctic sea ice</td>
<td>cice</td>
<td>≤0.5 land fraction and &gt;0.5 sea ice fraction*</td>
</tr>
<tr>
<td>Arctic open ocean</td>
<td>ocn</td>
<td>≤0.5 land fraction and ≤0.5 sea ice fraction</td>
</tr>
<tr>
<td>Arctic land with snow cover</td>
<td>lnds</td>
<td>AMPIM simulations (present-day with prescribed SSTs and sea ice extent): &gt;0.5 land fraction and &gt;0.5 MEaSUREs observed snow cover frequency* RCP8.5 simulations: &gt;0.5 land fraction and &gt;5-cm gridbox-averaged snow depth</td>
</tr>
<tr>
<td>Arctic land without snow cover</td>
<td>lnds</td>
<td>&gt;0.5 land fraction and ≤5-cm gridbox-averaged snow depth sea ice fraction</td>
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*CMIP5 present-day comparisons utilize consistent sea ice and snow cover surface masks for all models prescribed from Hadley Centre–observed sea ice extent and MEaSUREs snow cover frequency data, regardless of predicted snow cover from individual models. RCP8.5 comparisons utilize projected sea ice and snow cover extent from each model to determine the surface mask designation.
a. Seasonal cycles of Arctic net TOA radiative fluxes and cloud amount

The seasonal cycles of CMIP5 net TOA radiative fluxes across the Arctic basin (60°–82°N) are compared to CERES-EBAF in Fig. 1. A positive downward SW bias means that the CMIP5 models predict excess energy entering the Arctic system, while a positive OLR bias means that the CMIP5 models predict excess energy leaving the Arctic system. Therefore, the sign of OLR biases is reversed when calculating combined downward SW + LW biases. Monthly average net TOA downward SW all-sky biases are significantly larger than OLR biases (up to ±30 W m$^{-2}$ rather than ±10 W m$^{-2}$, respectively), despite the absolute magnitude of net TOA downward SW all-sky fluxes being smaller than OLR fluxes most of the year. Even in June, when net TOA downward SW all-sky fluxes are largest, their magnitude is only slightly larger than OLR fluxes (265 vs 229 W m$^{-2}$, respectively). Therefore, net TOA downward SW all-sky biases are generally larger than OLR biases both in absolute and in relative magnitude. The absolute magnitude of SW all-sky biases is largest in the spring and summer months where SW insolation is highest and the net TOA fluxes are sensitive to changes in surface albedo due to melting snow and ice. Net TOA downward SW fluxes are also very sensitive to sea ice extent, but in these simulations sea surface temperatures and sea ice extent are constrained to observations. As with SW all-sky biases, SW clear-sky and SW cloud forcing biases are up to 30 W m$^{-2}$. The CMIP5 models span positive and negative SW cloud forcing biases in roughly equal partitioning. However, most CMIP5 models have positive downward SW clear-sky biases in spring and summer, resulting in a positive downward SW all-sky bias averaged across models. This suggests that the CMIP5 models may have too little snow cover or surface albedos that are too low, or possibly CERES-EBAF retrieval errors. This is explored in more detail in sections 3b and 3c. The net TOA downward SW + LW biases (Figs. 1g–i) are largest in the spring and summer months. The CMIP5 model average downward SW + LW biases are positive throughout most of the year because of the contributions of negative OLR biases year-round (insufficient OLR leaving the system) and positive downward SW clear-sky biases in spring and summer (excess SW entering the system). Negative OLR biases are due in part to cold biases in surface temperature in the CMIP5 models, which are in turn driven by insufficient liquid water in Arctic mixed-phase clouds (Barton et al. 2014). SW biases can be caused by biases in cloud amount, cloud properties, surface type, and/or surface albedo.

To explore the possible causes of net TOA downward SW radiative flux biases, we first compare the seasonal cycles of CMIP5 total cloud amount using the COSP CALIPSO lidar simulator to CALIPSO-GOCCP observations (Fig. 2a). The majority of the CMIP5 models in this analysis have insufficient cloud amount throughout most of the year, especially in summer. While biases with high (Fig. 2b), middle (Fig. 2c), and low (Fig. 2d) clouds are generally similar to biases with total cloud amount, biases with low cloud most closely match total cloud amount biases. This is because low cloud is more common than medium and high clouds and because the CMIP5 models have a harder time representing low clouds, which are commonly mixed phase. The two exceptions are GFDL CM3 and MPI-ESM-LR; these two models have excessive cloud amount year-round, especially in winter. The causes of cloud amount variance among CMIP5 models are not clear, and there are many possible explanations, including circulation biases, humidity biases, and macroscopic- and microscopic-scale biases in the cloud representation. Since all of these simulations prescribe SSTs and sea ice extent, some of the larger-scale biases may be reduced. However, cloud microphysics processes remain independent and can vary significantly among models. GFDL CM3 and MPI-ESM-LR predict
mixing ratios of cloud liquid and ice and have explicit parameterizations of mixed-phase freezing processes (Donner et al. 2011; Stevens et al. 2013). Both IPSL-CM5A-LR and CESM1(CAM5) predict much lower cloud amount than GFDL CM3 and MPI-ESM-LR. CESM1(CAM5)’s cloud microphysics scheme, which predicts number concentrations and mixing ratios of cloud liquid and ice (Morrison and Gettelman 2008; Gettelman et al. 2010), is more advanced than the schemes in GFDL CM3 and MPI-ESM-LR; however, its

FIG. 1. Seasonal cycle of monthly-average present-day (a)–(c) net TOA downward SW, (d)–(f) OLR, and (g)–(i) net downward SW+LW all-sky (left), clear-sky (center), and cloud forcing (right) radiation biases in the Arctic (60°–82°N) for CMIP5 AMIP (atmospheric) model simulations minus CERES-EBAF observations. CMIP5 simulations and CERES-EBAF observations are 9-yr averages (2000–08). A positive net TOA downward SW bias means the model predicts excess SW fluxes entering the Arctic system. A positive OLR bias means the model predicts excess LW fluxes leaving the Arctic system. A positive net downward SW\_LW bias, calculated as downward SW bias minus OLR bias, means the model predicts excess net energy entering the Arctic system. CERES-EBAF regional-average SW all-sky and clear-sky errors are estimated to be 4 and 2.6 W m\(^{-2}\), respectively, in the CERES-EBAF Ed2.8 quality summary.
mixed-phase cloud representation is much simpler and is based on an empirical fit to midlatitude aircraft observations (Meyers et al. 1992). In contrast, IPSL-CM5A-LR’s cloud microphysics scheme and mixed-phase representations are simpler, predicting a single mixing ratio of total water and prescribing mixed-phase clouds using a linear expression (Hourdin et al. 2013). Arctic clouds are challenging to represent, particularly mixed-phase clouds, which are prevalent throughout much of the year (Morrison et al. 2012). Studies with CESM1(CAM5) found improvements to Arctic cloud amount when the mixed-phase cloud scheme is improved (Liu et al. 2011; English et al. 2014). Other studies also find large Arctic variability in cloud amount among CMIP5 models. When comparing native cloud amount to ERA-Interim reanalysis, Karlsson and Svensson (2013) found that most models predicted insufficient cloud amount, similar to our results. When Cesana and Chepfer (2012) compared CMIP5 cloud amount using the COSP lidar simulator to CALIPSO-GOCCP observations, they too found large variability but concluded that the CMIP5 models generally overpredicted cloud amount in the Arctic, in contrast to our results. Our analysis includes several CMIP5 models that were not included in their analysis.
A comparison of net TOA downward SW cloud forcing biases versus CERES-EBAF (Fig. 1c) with total cloud amount biases versus CALIPSO-GOCCP (Fig. 2a) reveals a generally inverse relationship in boreal summer, as expected. This is because a positive downward SW cloud forcing bias means that too much SW energy is entering the Arctic system, which would occur when there are insufficient clouds. In general, CMIP5 models with insufficient cloud amount (e.g., CNRM-CM5 and IPSL-CM5A-LR) produce too little radiative cooling from clouds, and CMIP5 models with excessive cloud amount (e.g., GFDL CM3 and MPI-ESM-LR) produce too much radiative cooling from clouds. There are some exceptions, however, and this is likely due to biases in modeled cloud optical depth and/or biases in the surface albedo underneath the clouds. These relationships are explored in more detail in section 3c. It is also possible that errors in CERES-EBAF clear-sky retrievals contribute to differences in SW cloud forcing between CERES-EBAF and the models, as was discussed in English et al. (2014) and will be discussed further in section 5.

b. Arctic spring/summer net TOA downward SW and cloud amount averages over different surface masks

Given that the largest net TOA downward SW biases occur in the spring and summer months, and the largest cloud amount biases also tend to occur in the spring and summer months, we next turn to a more detailed analysis of Arctic net TOA downward SW fluxes in the CMIP5 models using averages from March through September (MAMJJAS). Biases in net TOA downward SW clear-sky fluxes can be attributed to biases in surface albedo if there is negligible noncloud atmospheric absorption (which is reasonable in the relatively clean, dry Arctic atmosphere) and if incoming SW fluxes are comparable between the models and CERES-EBAF. In the Arctic, MAMJJAS incoming SW fluxes range from 324.9 to 327.6 W m$^{-2}$ across the nine CMIP5 models and CERES-EBAF observations, which is much smaller than the differences in net TOA downward SW clear-sky fluxes (213.9 to 227.8 W m$^{-2}$). Hence, biases in surface albedo are a primary driver of biases in net TOA downward SW clear-sky biases, and analysis of TOA flux biases over different surface types can be a useful tool to understand surface albedo biases and cloud forcing biases. The model spatial domains are separated into surface masks based upon surface composition (Table 2), and net TOA downward SW and cloud amount biases are plotted in Fig. 3. In both the CMIP5 models and CERES-EBAF observations, the surfaces with the highest net TOA downward SW all-sky and clear-sky fluxes are over open ocean and land areas without snow cover, due to lower surface albedo. Correspondingly, net TOA downward SW all-sky and clear-sky fluxes are lowest over sea ice and land areas with snow cover. Regions over land without snow cover have higher net TOA all-sky fluxes than regions over the ocean, while regions over ocean have higher net TOA clear-sky fluxes than areas over the land with snow cover. This difference is due to the fact that cloud amount over the open ocean is higher than that over land areas without snow cover in both the CMIP5 models (Fig. 3g) and CALIPSO-GOCCP observations (Fig. 3h), which causes a larger reduction from net TOA downward SW clear-sky fluxes to all-sky fluxes (i.e., a stronger cloud radiative effect).

Seasonally averaged (MAMJJAS) CMIP5 median net TOA downward SW all-sky fluxes are about 4 W m$^{-2}$ higher than CERES-EBAF over the Arctic basin (Fig. 3c). Median biases are smallest over land (1 W m$^{-2}$) and largest over open ocean (6 W m$^{-2}$). However, individual model biases span a large range; net TOA downward SW all-sky biases range from −13 to +15 W m$^{-2}$, and three out of nine CMIP5 models have negative net TOA downward SW all-sky biases. Analysis of net TOA downward SW clear-sky biases and cloud amount biases can help provide insights regarding possible causes of all-sky biases over different surface types. While CMIP5 median net TOA downward SW clear-sky biases are roughly comparable to all-sky biases (6 vs 4 W m$^{-2}$), the range of biases is narrower for clear-sky fluxes than all-sky fluxes over all surfaces except sea ice and land areas with snow. The models span a large range of average cloud amount over all surfaces (Fig. 3g), which is likely a significant contributor to model-to-model variability in net TOA downward SW all-sky biases. Across the Arctic basin, CMIP5 median cloud amount is 14% lower than CALIPSO-GOCCP (60% vs 74%). CMIP5 median cloud amount biases do not vary much over different surface types but are smallest over the ocean and sea ice and largest over land areas (both with and without snow cover). The variability in net TOA downward SW clear-sky fluxes over sea ice is mainly due to the MIROC5 model, which is 10 W m$^{-2}$ lower than the next lowest model (not shown). MIROC5 net TOA downward SW clear-sky fluxes are too low (surface albedo is too high), likely from errors in representation of sea ice albedo and/or the albedo of snow and/or melt ponds present on sea ice. Since CMIP5-modeled sea ice extent is constrained to match observations in these simulations, clear-sky biases over sea ice for all of the
models are due to surface albedo errors, not sea ice extent errors. That there is a large spread in net TOA downward SW clear-sky biases over sea ice even when sea ice extent is prescribed illustrates the importance of getting both sea ice extent and sea ice albedo correct. The large spread in net TOA downward SW clear-sky biases among CMIP5 models, as well as their median bias, is consistent regardless of the criteria for designating a grid box as sea ice (Table 3). The large spread in the representation of Arctic sea ice albedo among the CMIP5 models and its impact on radiative fluxes is consistent with other studies (Karlsson and Svensson 2013; Koenigk et al. 2014).
SW clear-sky biases are positive over land areas with snow and negative over land areas without snow (Fig. 3f). This relationship occurs regardless of the cutoff criteria for designating grid boxes as land with snow (Table 3). This relationship also occurs regardless of whether Greenland is included in the analysis. When excluding Greenland, median clear-sky biases increase from 11.4 to 12.9 W m⁻² and average biases decrease from 9.0 to 6.5 W m⁻². This suggests that snow-covered land albedos are too high in the CMIP5 models, while snow-free land albedos are too low, and this bias is not limited to just Greenland. Recall that for this analysis, the MEaSUREs merged observed snow cover frequency dataset is utilized to identify which grid boxes are snow covered, regardless of what the simulated snow cover extent is in each of the models. Therefore, discrepancies between modeled snow cover extent and observed snow cover extent may contribute to net TOA downward SW clear-sky biases and surface albedo biases. Unfortunately, snow cover output fields are not available from all of the CMIP5 models, the fields are not consistent with one another (the models may output snow depth, snow area density, or snow area extent, with inconsistent conversions between one another), and none of the models output snow cover frequency of occurrence, which is the unit of measurement for the MEaSUREs dataset. Of the fields available, CMIP5 net TOA downward SW clear-sky flux biases have a low correlation with model snow depth (\( n = 4, r^2 = 0.41 \)) and a medium correlation with snow cover area extent (\( n = 4, r^2 = 0.64 \)) (not shown). The lack of correlation with snow depth may be impacted by Greenland’s deep snowpack. Across the Arctic basin (60°–82°N) over the years 2000–08 from March through September, the CMIP5 snow cover area extent for the four models available ranges from 31% to 37%. When converting model monthly-average output to snow frequency of occurrence by requiring 50% snow area extent as was done in other studies (Frei et al. 1999; Henderson and Leathers 2010), the modeled snow frequency of occurrence ranges from 31% to 39%, while the MEaSUREs snow cover frequency is 65%. This suggests that the positive CMIP5 net TOA downward SW clear-sky flux biases present over land areas with snow may be at least partially explained by insufficient snow cover extent. There is much uncertainty, however, in converting from modeled snow cover area extent to observed snow cover frequency of occurrence.

CMIP5 models with negative net TOA downward SW clear-sky flux biases likely have surface albedos that are too high. However, the previous analysis suggests that the CMIP5 models likely have insufficient snow cover over the regions designated as land areas without snow, which would result in surface albedos that are too low. It is possible that there are errors in the albedo representations of soil, trees, and vegetation in the CMIP5 models. It is also possible that the CMIP5 models have errors in the interactions between tree canopies and snow cover, which could cause biases over land areas with and without snow. Indeed, an analysis by Loranty et al. (2014) found that the CMIP5 models typically overestimate albedos in tree-covered regions and underestimate albedos north of the tree line.

c. Exploring correlations between radiative flux and cloud biases

A comparison of CMIP5 cloud amount bias (CMIP5 minus CALIPSO-GOCCP) versus CMIP5 net TOA downward SW cloud forcing bias (CMIP5 minus CERES-EBAF) reveals that SW cloud forcing biases are inversely related to cloud amount biases, as expected (Fig. 4). However, a zero cloud amount bias corresponds to a −15 W m⁻² SW cloud forcing bias (Fig. 4a), meaning that net TOA downward SW cloud forcing is too weak at a given cloud amount (clouds are not reflective enough). This offset is significantly influenced by two models (IPSL-CM5A-LR and MRI-CGCM3), suggesting they have biases in cloud optical depth and/or surface albedos. It is unclear whether biases in snow cover extent are influencing the cloud radiative effect for these two model outliers; snow output was not available for IPSL-CM5A-LR, and MRI-CGCM3 had the second-highest snow cover extent out of four models with available snow fields, so it was not an outlier. Over land areas with snow (Fig. 4e), however, all of the CMIP5 models but one have a negative offset from the origin, meaning that the models predict net TOA downward SW cloud forcing to be too low (downward SW clear-sky fluxes to be too high), even when accounting for biases in cloud amount. As this occurs only over one surface mask, the biases are likely due to biases in surface albedos rather than cloud optical properties. This is likely due to the models predicting insufficient snow cover area extent and/or the surface albedos from the interaction between snow cover and tree canopy producing surface albedos that are too low, consistent with the conclusions from section 3b.

4. RCP8.5 projections of clouds and radiative fluxes in CMIP5 models

With an evaluation of how well the CMIP5 models represent present-day radiative fluxes and cloud amount complete, we now turn to an assessment of future projections under the RCP8.5 scenario. We conduct a similar analysis as in section 3 but compare the output from
coupled simulations with freely evolving SSTs and sea ice extent rather than AMIP simulations with prescribed SSTs and sea ice extent. We compare an average of the years 2081–90 to an average of the years 2006–15).

**a. Projected changes in seasonal cycles of Arctic net TOA radiative fluxes**

The seasonal cycles of RCP8.5 projected changes to net TOA downward SW and OLR fluxes (2081–90 minus 2006–15) in the CMIP5 models are illustrated in Fig. 5. Monthly-average projected radiative flux changes to the Arctic system are dominated by approximately 40 W m$^{-2}$ increases in net TOA downward SW clear-sky fluxes in the summer months, due to a reduction in high-albedo surfaces including sea ice and snow cover on land and sea ice. The CMIP5 models also predict a decrease in downward SW cloud forcing (Fig. 5c), which translates to a stronger SW cloud forcing. This is because downward SW cloud forcing is negative and the CMIP5 models project larger negative values in the future and is due to a decrease in surface albedo from melting snow and ice, causing a larger albedo contrast between clouds and the underlying surfaces, making the cloud radiative effect stronger. Cloud amount is not contributing to changes in downward SW cloud forcing, as it is projected to have changes of less than 2% (Fig. 6i). However, individual models span a larger range. CMIP5 models also project increases in OLR fluxes year-round, particularly clear-sky OLR fluxes (Fig. 5e) in winter due to higher surface temperatures, despite known decreases to OLR that are associated with increasing concentrations of greenhouse gases. The net effect from a monthly-average standpoint is increased energy into the Arctic system in the spring and summer months and decreased energy into the Arctic system (e.g., more energy leaving the Arctic system) in the winter months. From an annual average standpoint, the
increase in net TOA downward SW all-sky fluxes from 2006–15 to 2081–90 (+8 W m\(^{-2}\)) is nearly compensated by an increase in OLR (+6 W m\(^{-2}\)). The net change is +2 W m\(^{-2}\) into the Arctic system. This analysis does not consider changes to northward energy transport from lower latitudes.

\textit{b. Projected changes over different surface masks in spring/summer}

Projected changes to net TOA downward SW radiative fluxes and clouds in MAMJJAS over different surface masks are provided in \textit{Fig. 6}. Downward SW all-sky radiation fluxes and clouds in MAMJJAS are provided in \textit{Fig. 6}.
and clear-sky fluxes (2081–90 minus 2006–15) are projected to increase over all surface masks, but with varying magnitudes. Changes to model median cloud amount are less than 2% over all surface masks (Fig. 6i), suggesting limited contributions of cloud changes to net TOA downward SW all-sky changes. However, individual models span a larger range. The largest net TOA downward SW all-sky and clear-sky increases per unit area are projected to occur over sea ice owing to its decreasing surface albedo as snow cover melts on sea ice and melt ponds form and as the fractional sea ice cover decreases (in this analysis, to be considered sea ice a grid box must have fractional sea ice cover of greater than 0.5). Relatively large increases in median net TOA downward SW clear-sky fluxes are projected regardless of whether the sea ice cutoff criteria ranges from 0.3 to 0.7 (Table 4). Hezel et al. (2012) found that part of the projected decline of spring snow depth on Arctic sea ice is due to later autumn freeze-up of Arctic sea ice, reducing the length of time that falling snow has to build up on sea ice. Downward SW all-sky fluxes over ocean and land areas with snow cover are projected to increase.

Fig. 6. As in Fig. 3, except MAMJAS averages of projected changes for the RCP8.5 scenario (2081–90 minus 2006–15) instead of present-day biases vs observations.
by less than 5 W m\(^{-2}\). This is not surprising since the albedos of ocean and snow are not projected to change significantly. This is supported by the relatively small projected change in net TOA downward SW fluxes over land areas with snow over grid boxes with at least 95% snow cover (Table 4). However, since the surface area of land with snow cover is projected to increase, over all land areas net TOA downward SW all-sky fluxes are projected to increase by about 10 W m\(^{-2}\) (Fig. 6c), and clear-sky fluxes are projected to increase by about 12 W m\(^{-2}\) (Fig. 6f).

The range in net TOA downward SW all-sky fluxes among individual CMIP5 models is roughly comparable in magnitude to the range in clear-sky fluxes, suggesting that clouds are not driving the variability. Rather, differences in sea ice extent, snow cover, and surface albedo parameterizations are driving the changes. The variability is highest over sea ice, where differences in snow/sea ice/melt pond albedo parameterizations contribute to variability, as well as significant variability in the projected surface area of remaining sea ice, which varies by more than a factor of 2 across the nine models studied (average March through September loss in sea ice extent ranges from \(1.8 \times 10^6\) to \(4.0 \times 10^6\) km\(^2\)) (Fig. 7c). The projected loss of sea ice area is roughly twice that of the projected loss of land area with snow (CMIP5 model median loss of sea ice is \(2.6 \times 10^6\) km\(^2\) and land area with snow is \(1.4 \times 10^6\) km\(^2\)).

When incorporating both the changes in radiative fluxes (W m\(^{-2}\)) with the changes in surface areas (km\(^2\)) for each mask, net TOA downward SW all-sky and clear-sky fluxes are projected to increase by 0.38 and 0.55 PW, respectively. Given that cloud amount is not projected to change significantly, changes in TOA downward SW clear-sky radiative fluxes have similar relative differences as changes in all-sky radiative fluxes, albeit larger magnitudes. Increases in net TOA downward SW clear-sky fluxes over the open ocean are the biggest contributor, owing mainly to an increase in surface area due to loss of sea ice (Fig. 7c), but also to a slight increase in W m\(^{-2}\) due to a loss of fractional sea ice cover in the grid boxes designated as open ocean across the Arctic basin (Fig. 7f). Likewise, an increase in net TOA downward SW fluxes over land areas without snow cover contributes approximately half as much SW to the Arctic system as increases over the open ocean, due mostly to an increase in land areas that are not snow covered, but partly to more loss of fractional snow cover in grid boxes designated as land areas without snow cover.

To quantify the relative contributions of changes in mask area versus changes in surface albedo to changes in TOA fluxes over the different masks, we calculate the contributions of each to the total TOA energy budget in the Arctic averaged annually (Fig. 8). A decrease in sea ice extent is the largest contributor to projected increases in net TOA downward SW all-sky and clear-sky fluxes in the decade 2081–90 compared to 2006–15. Decreases to snow cover area extent on land contribute the second largest contribution, followed by changes in sea ice albedo of the sea ice that does remain. Changes to snow cover albedo on land contribute insignificantly to projected increases to net TOA downward SW fluxes in 2081–2090.

Finally, comparisons between projected changes to net TOA downward SW all-sky fluxes and biases in present-day fluxes are provided in Fig. 9. Over the entire Arctic basin (Fig. 9a), there is very little correlation between an individual model’s downward SW all-sky flux bias and its projected change to downward SW all-sky fluxes in the future (there is large scatter and the slope is near zero). This suggests that there is not a relationship between a model’s ability to represent present-day SW all-sky fluxes and its projection of future changes to SW all-sky fluxes. There is a relationship between present-day bias and future change over sea ice (Fig. 9b); the CMIP5 models with near-zero present-day biases representing downward SW all-sky fluxes over
sea ice tend to project changes to downward SW all-sky fluxes over sea ice of 5–20 W m⁻².

5. Discussion

Compared to CERES-EBAF, the CMIP5 models have larger monthly biases representing net TOA downward SW fluxes than OLR in the Arctic. Both SW clear-sky and SW cloud forcing biases contribute to the SW all-sky biases. Most models have positive downward SW clear-sky biases, while both positive and negative downward SW cloud forcing biases are present. The SW cloud forcing biases are mainly due to cloud amount biases, supported by a comparison of CMIP5 cloud amount to CALIPSO-GOCCP observations using each model’s respective COSP CALIPSO lidar simulator. The SW clear-sky biases are present over all surface types (sea ice, open ocean, land with snow cover, and land without snow cover), suggesting several contributions to SW clear-sky model biases. SW clear-sky biases are largest over land areas with snow cover. These biases may be due to insufficient snow area extent in the models, although it is difficult to compare modeled snow area extent to observed snow frequency.
of occurrence. The biases are not related to snow depth, as CMIP5 snow depth has a low correlation with CMIP5 SW clear-sky biases. Other possible contributions include errors in the albedo parameterizations of the

interactions between snow cover and land vegetation canopy/albedo, as supported by other work (Loranty et al. 2014), and/or errors in the subgrid-scale fractional representation of snow cover albedo. Despite having

FIG. 8. Annual averages of contributions to CMIP5 projected net TOA downward SW (a) all-sky and (b) clear-sky changes (2081–2090 minus 2006–2015). The contributions include changes to sea ice extent \(c_{\text{area}}\), sea ice albedo \(c_{\text{alb}}\), snow area extent on land \(s_{\text{area}}\), and snow albedo on land \(s_{\text{alb}}\).

FIG. 9. Scatterplots of net TOA downward SW all-sky fluxes. y axis: MAMJJAS averages of CMIP5 projected net TOA downward SW all-sky change (2081–90 minus 2006–15) (W m\(^{-2}\)). x axis: MAMJJAS averages of present-day (2000–08) net TOA downward SW all-sky bias (CMIP5 minus CERES-EBAF) (W m\(^{-2}\)). Linear trendline illustrates a hypothetical one-to-one relationship between projected all-sky change and present-day all-sky bias.
prescribed sea ice extent, CMIP5 models have a large range of biases in SW clear-sky fluxes over sea ice. This suggests that several of the CMIP5 models have errors in their representation of the albedo of snow on sea ice and/or melt ponds, or possibly the albedo of sea ice itself, highlighting the importance of getting not only sea ice extent correct but also sea ice albedo. Cloud amount biases are similar across the surface masks, suggesting that cloud biases are not driving radiative flux biases more over one surface mask than another. Figure 1 indicates that both cloud forcing and clear-sky biases are about 30 W m\(^{-2}\). There is an offset between modeled COSP-CALIPSO cloud fraction biases compared to CALIPSO-GOCCP and downward SW cloud forcing biases compared to CERES-EBAF, which suggests that model biases in surface albedo being too high are causing errors in the cloud radiative effect being too weak in the models. Therefore, this analysis suggests that with increased Arctic cloud amount and/or cloud optical thickness and increased snow area extent and/or improved snow-surface albedo parameterizations, many CMIP5 models would more accurately represent net TOA downward SW clear-sky and SW cloud forcing radiative fluxes in the Arctic.

It is somewhat surprising that the CMIP5 models have a positive downward SW clear-sky bias over open ocean (the median bias is 5 W m\(^{-2}\)), as it should be relatively simple to prescribe ocean albedo. This bias is not explained by fractional sea ice cover; when changing the cutoff criteria for grid boxes designated as open ocean from \(\leq 0.5\) sea ice fraction to \(\leq 0.05\) sea ice fraction, the median bias is 4 W m\(^{-2}\) (not shown). It is possible that these downward SW clear-sky differences are attributable to errors and uncertainties with CERES-EBAF retrievals. There is much difficulty obtaining accurate high-latitude retrievals of clear-sky radiative fluxes owing to low albedo contrasts between clouds and snow combined with high-latitude zenith angles, and prior analysis of SW clear-sky biases using the CESM1(CAM5) model suggested CERES-EBAF biases in SW clear-sky retrievals over some areas at latitudes greater than 60\(^\circ\)N (English et al. 2014). However, analysis over different surface types combined with cross-comparison with cloud amount biases versus CALIPSO-GOCCP adds confidence that the majority of the SW clear-sky biases between the CMIP5 models and CERES-EBAF observations are consistent and can be otherwise explained.

The CMIP5 models project significant increases to net TOA downward SW fluxes in the spring/summer under the RCP8.5 scenario, which are partially compensated by increases to OLR fluxes year-round, with a net increase of 2 W m\(^{-2}\) into the Arctic basin. A breakdown of the fluxes over surface masks suggests that declining sea ice area is the largest contributor to an increase in net TOA downward SW fluxes, but that changes to sea ice albedo and snow cover area extent on land contribute as well. These results suggest that it is important to accurately represent both surface area and albedos of sea ice and snow cover to get present-day net TOA downward SW fluxes correct and to accurately predict the Arctic radiative budget in the future. Because of the variations in net TOA downward SW fluxes among CMIP5 models over the different surface masks, representation of these surface processes continues to be a challenge. Finally, while the CMIP5 models have large variability in cloud amount in present-day climate, with most models predicting insufficient cloud amount versus CALIPSO-GOCCP over all surface masks, cloud amount is not projected to change significantly in the future over any of the surface masks.

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