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ABSTRACT

A proposed objective of the planned Aerosol–Cloud–Ecosystem (ACE) satellite mission is to provide constraints on climate model representation of aerosol effects on clouds by retrieving profiles of aerosol number concentration, effective variance, and effective radius over the 0.1–1-μm radius range under humidified ambient conditions with 500-m vertical resolution and uncertainties of 100%, 50%, and 10%, respectively. Shallow, broken marine clouds provide an example of conditions where boundary layer aerosol properties would be retrieved in clear-sky gaps. To quantify the degree of constraint that proposed retrievals might provide on cloud radiative forcing (CRF) simulated by climate models under such conditions, dry aerosol size distribution parameters are independently varied here in large-eddy simulations of three well-established modeling case studies. Using the rudimentary available aerosol specifications, it is found that relative changes of total dry aerosol properties in simulations can be used as a proxy for relative changes of ambient aerosol properties targeted by ACE retrievals. The sensitivity of simulated daytime shortwave CRF to the proposed uncertainty in retrieved aerosol number concentration is \( 2^{-15} \text{W m}^{-2} \) in the overcast limit, roughly a factor of 2 smaller than a simple analytic estimate owing primarily to aerosol-induced reductions in simulated liquid water path across this particular set of case studies. The CRF sensitivity to proposed uncertainties in retrieved aerosol effective variance and effective radius is typically far smaller, with no corresponding analytic estimate. Generalization of the results obtained here using only three case studies would require statistical analysis of relevant meteorological and aerosol observations and quantification of observational and model uncertainties and biases.

1. Introduction

The interactions of aerosols with clouds represent a leading source of uncertainty in quantifying anthropogenic radiative forcing of climate globally since pre-industrial times (Solomon et al. 2007). Clouds are also reported to constitute the largest source of uncertainty in climate sensitivity to radiative forcing in current coupled ocean–atmosphere climate models (Soden and Held 2006). In the tropics, differences in the predicted sensitivity of marine boundary layer clouds to CO₂ doubling have been specifically identified as the main source of uncertainty in cloud feedback (Bony and Dufresne 2005).

The contribution of low cloud properties to both anthropogenic radiative forcing of climate and climate sensitivity has motivated extensive efforts to characterize the interactions of aerosols with low clouds and improve their representation in climate models. Whereas in situ measurements are generally most capable of characterizing localized aerosol and cloud microphysical properties in detail, routine satellite-based measurements provide long-term global coverage that is useful for constraining climate models (e.g., Quaas et al. 2009). However, a common disadvantage of satellite retrievals relative to other measurements is reduced spatial resolution and measurement sensitivity. The process of determining whether specific proposed retrievals will be adequate to meet intended scientific needs is therefore critical.

In a recent U.S. decadal survey for Earth observations from space (National Research Council 2007), objectives for the National Aeronautics and Space Administration (NASA) Aerosol–Clouds–Ecosystem (ACE) satellite mission include retrieval of aerosol and cloud property profiles. Here we focus on the proposed retrieval of the number, effective variance, and effective radius of ambient aerosols in the 0.1–1-μm radius range at 500-m
vertical resolution with associated uncertainties of 100%, 50%, and 10%, respectively. These definitions represent nominal aerosol retrieval requirements established during ACE mission planning, which are intended for the purpose of better constraining the representation of aerosol–cloud interactions in climate models, to be distinguished from the proposed requirements for the retrieval of other aerosol properties more closely tied to the direct effects of aerosols on the atmospheric radiative budget. The retrieval of ambient rather than dry aerosol size distribution properties is considered to reduce the dependence of retrieval accuracy on collocated atmospheric state profiles that are not measured by the ACE mission (such as relative humidity) and size-resolved aerosol composition.

Given the proposed ACE uncertainty requirements placed on the retrieval of these and other aerosol and cloud property profiles, ACE mission planning efforts aim to systematically estimate relevant associated uncertainties in top-of-atmosphere radiative forcing as a common metric for considering the adequacy of each retrieval to meet intended scientific needs. It is our objective here to use large-eddy simulations with resolved aerosol and cloud particle size distributions to estimate the top-of-atmosphere cloud radiative forcing (CRF) associated with the specific proposed aerosol parameter uncertainties described above. We focus on broken low clouds in order to evaluate the sensitivity of cloud properties to aerosol properties below the cloud base, which are observable in cloud-free pixels among cloudy pixels by virtue of boundary layer mixing. We consider only marine clouds to limit the number of surface parameters.

We treat all of the aerosol as ammonium bisulfate to limit the number of aerosol composition parameters. The objective is to estimate aggregate CRF sensitivity to proposed retrieval uncertainties, including, but not limited to, any induced changes in cloud albedo, cover, and thickness.

We emphasize that quantifying the potential radiative impact of anthropogenic aerosols on broken marine cloud fields is not an objective of this work. Rather, we aim only to quantify CRF sensitivity to proposed uncertainties in retrievals of specific aerosol properties. The general application in mind here is not the observational correlation of aerosol with cloud properties, but rather the constraint of aerosol properties beneath broken cloud fields in climate models, which is considered in terms of associated CRF at the local scale (over individual broken cloud fields) and at the global scale (considering the global frequency and variety of broken cloud fields). For example, what is the possible CRF uncertainty associated with an uncertainty of 50% in the effective variance of aerosols in the 0.1–1-μm size range averaged over a 500-m layer at ambient relative humidity beneath a broken marine cloud field?

We are aware of no literature that quantifies the sensitivity of CRF over broken marine clouds to the specific aerosol properties targeted by ACE. There are two main barriers to adapting past work to meet the needs of this study. First, the aerosol properties targeted by ACE retrievals cannot be universally related to commonly considered aerosol properties, such as the total concentration of cloud condensation nuclei (CCN; e.g., McComiskey and Feingold 2008). CCN as a function of supersaturation is calculated during the course of the simulations here, but the inverse derivation of ACE aerosol size distribution properties from CCN requires the specification of detailed underlying aerosol and meteorological conditions. Second, observationally based studies generally rely upon the categorizing of observed conditions in terms of meteorological and cloud properties in order to isolate aerosol influences (e.g., Garrett and Zhao 2006), whereas the objective of the calculations presented here is to calculate the total CRF sensitivity to aerosol retrieval uncertainties, including the dynamical response, which cannot be assumed to be negligible (e.g., Garrett et al. 2009).

2. Simulations

The three cases of shallow, broken marine clouds developed for model intercomparison studies organized by the Global Energy and Water Cycle Experiment (GEWEX) Cloud System Study (GCSS) program are used as the basis for this study. We are aware of no other well-established modeling case studies. The cases are representative of trade wind cumuli observed during the Atlantic Trade Wind Experiment (ATEX; Stevens et al. 2001), the Barbados Oceanographic and Meteorological Experiment (BOMEX; Siebesma et al. 2003), and the Rain in Cumulus over Ocean (RICO) experiment (van Zanten et al. 2011).

Simulations are performed with the Distributed Hydrodynamic–Aerosol–Radiation Modeling Application (DHARMA; Stevens et al. 2002; Ackerman et al. 2004), a large-eddy simulation code coupled with size-resolved microphysics based on the Community Aerosol–Radiation–Microphysics for Atmospheres (CARMA) code (Ackerman et al. 1995; Jensen et al. 1998). Initial conditions and forcings follow the model intercomparison specifications for each case. The ATEX, BOMEX, and RICO simulations use uniform horizontal and vertical grid spacing of 100 and 40 m over a domain of 9.6 km × 9.6 km × 3 km (ATEX and BOMEX) or 12.8 km × 12.8 km × 4 km (RICO), a mass-doubling particle grid of 25 bins (ATEX and BOMEX) or 35 bins (RICO), and
a duration of 6–8 h. Radiative cooling follows intercomparison specifications, in which a horizontally uniform cooling rate is used for BOMEX and RICO simulations (cloud cover is less than 10%) and a Beer’s law parameterization of longwave cooling that depends on the condensed water profile in each column is used for ATEX. When offline shortwave radiative flux calculations are made, a solar zenith angle (θ) of 60° and a solar constant (S₀) of 1367 W m⁻² are used.

The only inputs that vary across simulations of each case are aerosol size distribution parameters, which are initialized uniformly throughout the domain. A diagnostic approach (Clark 1974) is used to avoid the need to specify unknown aerosol source terms and track core second moments (to restore aerosol dispersion upon droplet evaporation). For the ATEX and BOMEX cases, aerosols are assumed to be present in a single log-normal mode with baseline dry number concentration. Because νₑ reductions of 50% result in rₑ decreases of more than 10% (see dry total values in Table 1), we omit separate simulations for 10% changes in rₑ. Aside, we note that rₑ decreases achieved by reducing rₑ (rather than decreasing νₑ) decrease rather than increase activated droplet number concentration, but we find that the relative absolute magnitude of response is comparable in a typical simulation (not shown) and is far weaker than the response to uncertainty in number concentration, as described below.

<table>
<thead>
<tr>
<th>Case</th>
<th>Dry total νₑ (–)</th>
<th>Dry 0.1–1-μm νₑ (–)</th>
<th>Ambient 0.1–1-μm νₑ (–)</th>
<th>Dry total rₑ (μm)</th>
<th>Dry 0.1–1-μm rₑ (μm)</th>
<th>Ambient 0.1–1-μm rₑ (μm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ATEX baseline</td>
<td>0.18</td>
<td>0.13</td>
<td>0.17</td>
<td>0.15</td>
<td>0.16</td>
<td>0.20</td>
</tr>
<tr>
<td>ATEX narrow</td>
<td>0.09</td>
<td>0.06</td>
<td>0.09</td>
<td>0.12</td>
<td>0.13</td>
<td>0.16</td>
</tr>
<tr>
<td>BOMEX baseline</td>
<td>0.18</td>
<td>0.13</td>
<td>0.18</td>
<td>0.15</td>
<td>0.16</td>
<td>0.22</td>
</tr>
<tr>
<td>BOMEX narrow</td>
<td>0.09</td>
<td>0.06</td>
<td>0.10</td>
<td>0.12</td>
<td>0.13</td>
<td>0.18</td>
</tr>
<tr>
<td>RICO baseline</td>
<td>0.54</td>
<td>0.27</td>
<td>0.23</td>
<td>0.27</td>
<td>0.30</td>
<td>0.39</td>
</tr>
<tr>
<td>RICO narrow</td>
<td>0.27</td>
<td>0.10</td>
<td>0.12</td>
<td>0.15</td>
<td>0.18</td>
<td>0.25</td>
</tr>
</tbody>
</table>

3. Results

After model spinup (1–2 h), domain-averaged droplet number concentration is relatively constant in all simulations, but liquid water path (LWP) and cloud cover may evolve, depending upon the case. Figure 1 illustrates the cloud state at the end of each baseline simulation. We first consider the possible relationship of satellite-observable near-surface aerosol properties, which are viewable between broken clouds under ambient relative humidity conditions, to the dry aerosol properties carried in large-eddy simulation (LES) and climate models. Using the last four 3D output fields saved from each simulation (spanning the last 1.5 h of each simulation), we average aerosol number concentration in the 0.1–1-μm radius range over grid cells below 500 m in clear columns, defined as those where mid-visible optical thickness (τ, including cloud, haze, and gases) is less than 2.5 (using a threshold of 1.0 changes results negligibly). Figure 2 illustrates the results for three representative simulations. In the ATEX cases (monomodal aerosol), more aerosols are present in the 0.1–1-μm radius range under ambient versus dry conditions because hydration leads to growth into that size range. In the RICO case (bimodal aerosol), aerosol concentration in the 0.1–1-μm radius range is roughly identical under ambient versus dry conditions because the larger aerosol mode is already mostly in that radius range when dry.

For all cases, the doublings of the aerosol number in the 0.1–1-μm radius range under clear-sky ambient
conditions (the quantity targeted by satellite retrievals) correspond directly to the doublings of total dry \( N_a \) (the more fundamental underlying quantity). Results for \( n_e \) and \( r_e \) (summarized in Table 1) indicate a similar correspondence of relative changes in ambient clear-sky near-surface values in the 0.1–1-\( \mu \)m radius range with relative changes in dry all-sky values. Thus, for the purposes of this study, sensitivity to relative changes of total dry aerosol properties can be adopted as a proxy for sensitivity to relative changes of ambient humidified aerosol properties targeted by ACE retrievals. It is unknown to what degree this would be true under naturally varying aerosol conditions, as discussed further below.

We turn next to CRF. Because the simulated response of broken cloud fields to \( N_a \) changes can be small compared to internal variability (e.g., Xue and Feingold 2006; Xue et al. 2008), we employ the following run-time approximation of shortwave daytime CRF that does not rely on less frequent offline radiative calculations:

\[
\text{CRF} = -kS_o \cos(\theta)A = -kS_o \cos(\theta)f_c \frac{\tau_c}{\tau_c + 13} \tag{1}
\]

[Ramanathan 1987; his Eq. (14) with a clear-sky albedo of 0], where the ratio of top-of-the-atmosphere (TOA) broadband to cloud-top mid-visible albedo (\( k \)) is assumed to be 0.8 (Charlson et al. 1992) and the domain-averaged mid-visible albedo at cloud top (\( \hat{A} \)) is approximated as the fractional cloud cover (\( f_c \)) multiplied by a cloudy-sky albedo of \( \hat{A}_c/(\tau_c + 13) \) [Bohren 1987; his Eq. (14) with an asymmetry parameter of 0.85], where \( \tau_c \) is defined as the optical thickness of activated droplets treated as geometric scatterers (twice the total droplet cross-sectional area). By computing an albedo for the cloudy portion of the model domain from \( \tau_c \) averaged over the cloudy columns,\(^1\) we ignore any variability among the cloudy columns, analogous to the partly cloudy retrieval scheme of Coakley et al. (2005). By only differentiating between clear and cloudy columns such an approach does not fully address the plane-parallel albedo bias (Cahalan et al. 1994). However, the scene albedo thus computed on average is within 0.01 (relative errors within 10%) of that using the independent column approximation.

Figure 3 illustrates the CRF calculated for sequential doublings of aerosol number concentration using the baseline and 50% reduced \( n_e \) values in all cases. Also shown are the domain-averaged droplet number concentration (weighted by the mass mixing ratio of condensed water, \( N_d \)), LWP, and cloud cover. In the ATEX case, LWP decreases can be substantial despite increases in cloud cover, consistent with other ATEX and BOMEX studies (Xue and Feingold 2006; Xue et al. 2008).\(^2\) Relative changes in \( N_d \) induced by reducing \( n_e \) are always far less than those induced by doubling \( N_a \). Reducing \( n_e \) allows the smallest aerosols to be more easily activated, consistent with the modest \( N_d \) increases

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\(^1\) Defined as those with \( \tau_c > 2.5 \) through the rest of our analysis.

\(^2\) Owing to the complexity of the interactions between microphysics, radiation, and dynamics over a wide range of aerosol and precipitation conditions across three different cases here, it is beyond the scope of this study to attribute changes in various simulation diagnostics to specific mechanisms. However, it is reasonable to expect that behavior similar to that found in past studies using a similar modeling framework can be attributed to similar causes.
induced by decreasing $n_e$ for ATEX and BOMEX; the maximum response of $N_d$ to reducing $n_e$ is about 10% of that induced by doubling $N_a$. For RICO, all aerosols in the larger mode are activated regardless of $n_e$. In both cases, increased $N_d$ results in modestly decreased peak supersaturations (not shown), and $N_d$ thus increases somewhat less than linearly with $N_a$ in both cases.

The change in TOA cloud radiative forcing ($\Delta$CRF), the final metric for this study, is shown in Fig. 4. Over this handful of cases, the $\Delta$CRF response to progressive doublings of $N_a$ is similar for the baseline and reduced (narrow) $n_e$ values, and the sensitivity to halving $n_e$ (equivalent to greater than 10% change in $r_e$) is typically far smaller in absolute magnitude. For changes induced by doubling $N_a$, $\Delta$CRF tends to strengthen with increasing $N_a$.

Aggregating over all cases, $\Delta$CRF is seen in Fig. 5 to generally strengthen with increasing cloud cover, indicating that relative changes in cloud cover within each case are smaller than relative changes in cloud albedo. Also shown in Fig. 5 is an analytic estimate for the daytime Twomey effect induced by a doubling of $N_a$ for overcast conditions over a nonreflecting surface, assuming constant liquid water path and a constant relative dispersion of droplet size distributions,

$$
\Delta \text{CRF} = -kS_o \cos(\theta) \Delta A = -kS_o \cos(\theta) \frac{A(1 - A)}{3} \Delta \ln N_d,
$$

where we equate relative changes in $N_d$ with those in $N_a$ ($\Delta \ln N_d = \Delta \ln N_a = \ln 2$) and approximate $\Delta A/\Delta \ln N_d$ from the maximum value of $dA/d\ln N_d = A(1 - A)/3$ at $A = 0.5$ (Twomey 1991). A linear least squares fit through the broken cloud cases simulated here (ATEX, BOMEX, and RICO) gives an intercept near 0 and intersects the overcast limit at roughly $-15 \text{ W m}^{-2}$. The factor-of-2 difference between that and the roughly $-30 \text{ W m}^{-2}$ obtained from Eq. (2) indicates that offsets to the Twomey effect from simulated reductions in LWP are substantial across these three well-established broken cloud cases. We emphasize that these three cases cannot be expected to be representative of a meaningful climatology, as discussed further below. In Fig. 5, the wide range of $\Delta$CRF calculated from three overcast stratocumulus cases simulated by Ackerman et al. (2004) is representative of cases in which aerosol-induced changes in LWP are both positive and negative. It is unknown to what degree positive liquid water path changes would contribute to the climatology of broken cloud cases. The linear fit through the broken cloud cases presented here is seen to run through the weak end of the range of $\Delta$CRF calculated for stratocumulus cases, and much stronger (more negative) $\Delta$CRF is shown for other stratocumulus cases in which LWP increases with $N_a$ (see Ackerman et al. 2004).

Based on the results of the broken cloud cases in hand, however, we now briefly consider what global CRF uncertainty could be associated with uncertainties in
proposed retrieved aerosol properties. Broken marine cloud fields may be uniquely germane for the constraint of climate model representation of aerosol effects on cloud properties using satellite retrievals since warm low-level clouds are considered to be susceptible to aerosol influences (as discussed in section 1), the observable aerosols are relevant to the collocated clouds (which may not be the case for overcast stratocumulus, e.g.), and underlying marine surfaces provide superior retrieval conditions. Of the retrieval uncertainties considered here, the $D_{CRF}$ associated with $N_a$ uncertainty is the largest and therefore would dominate. The mean frequency of cumulus from surface observations is reported as 34% over global oceans, with a mean cloud cover of 38% when present (Warren et al. 1988), which is represented by the dashed line in Fig. 5. Intersection with the linear fit at roughly $-6$ W m$^{-2}$ provides a global equivalent uncertainty in diurnal shortwave CRF of roughly 0.7 W m$^{-2}$, which is arrived at by multiplying 6 W m$^{-2}$ by the mean global frequency of 0.34, the fraction of the earth surface covered by ocean of 0.7, and a fraction of 0.5 to account for the fraction of the globe that is illuminated. Thus, in a simplistically considered global model, if the number concentration of 0.1–1-$\mu$m near-surface aerosols at ambient relative humidity representative of clear-sky conditions in broken marine clouds fields were 100% greater than perfectly known values, simulated global CRF could diverge by at least 0.7 W m$^{-2}$ from a correct
value. This is a lower limit to uncertainty in part because it assumes that the relationship of the retrieved aerosol property to the total dry aerosol size distribution is also known, but it serves to quantitatively illustrate the degree to which uncertainties in currently proposed aerosol property retrievals may be associated with limitations in their ability to constrain global CRF in climate model results with respect to one relevant cloud type.

Generalizing this study regionally or climatologically would require statistical evaluation of the relevant meteorological and aerosol conditions and synthesis of an appropriately representative ensemble of case studies, in addition to quantification of uncertainties and biases, which cannot be assumed to be insignificant for either the models (e.g., Stevens and Seifert 2008) or observations (e.g., obtaining statistically representative coverage diurnally and regionally).

4. Summary and discussion

NASA’s ACE satellite mission proposes to retrieve vertical profiles of aerosol number concentration, effective variance, and effective radius over the 0.1–1-μm radius range under humidified ambient conditions with 500-m vertical resolution and uncertainties of 100%, 50%, and 10%, respectively. Shallow, broken marine clouds provide an example of conditions where boundary layer aerosol properties would be retrieved in clear-sky columns. The degree to which proposed aerosol profile retrievals would contribute to constraining climate model representation of top-of-atmosphere cloud radiative forcing (CRF) under such conditions is quantified using large-eddy simulations of three well-established model intercomparison case studies.

For the rudimentary aerosol representations in each case, it is first found that relative changes in total dry aerosol properties can be used as a proxy for relative changes in ambient aerosol properties targeted by ACE retrievals. Namely, the fraction of humidified aerosol in the 0.1–1-μm radius range at 0–500 m in clear columns doubles along with total aerosol number concentration \( N_a \). Similarly, the clear-sky ambient aerosol effective variance \( \nu_e \) in the 0.1–1-μm radius range is roughly halved along with a 50% reduction in dry \( \nu_e \), and corresponds to greater than 10% changes in either dry or clear-sky ambient effective radius \( r_e \), see Table 1.

The sensitivity of daytime shortwave CRF to the proposed 100% uncertainty in retrieved aerosol number concentration is found to be \(-15\, \text{W m}^{-2}\) in the overcast limit. This is roughly a factor of 2 smaller in magnitude than an analytic estimate of the Twomey effect owing primarily to aerosol-induced reductions in liquid water path across the small set of case studies considered here. It is unknown to what degree aerosol-induced increases in

![Fig. 4. For (left to right) ATEX, BOMEX, and RICO change in shortwave daytime CRF induced by progressive doublings of \( N_a \) at baseline \( \nu_e \) (solid lines) and at 50% reduced \( \nu_e \) (dotted lines; "narrow"), and by 50% reduction of \( \nu_e \) at a given \( N_a \) (dashed lines).](image1)

![Fig. 5. The ΔCRF from doubling \( N_e \) vs cloud cover for all cases. A least squares linear fit through ATEX, BOMEX, and RICO cases has an intercept of \(-0.7\, \text{W m}^{-2}\) and a slope of \(-15\, \text{W m}^{-2}\). An analytic estimate of the Twomey effect [Eq. (2)] gives a value of \(-33\, \text{W m}^{-2}\) for overcast conditions. By comparison, simulations of stratocumulus (Ackerman et al. 2004) give values from \(-6\) to \(-73\, \text{W m}^{-2}\) in the overcast limit. Cloud cover of 0.38 (dashed line) is global mean over oceans when cumulus is present based on surface observations compiled by Warren et al. (1988).](image2)
liquid water path, demonstrated to occur for overcast stratocumulus (e.g., Ackerman et al. 2004), would contribute to a climatology of broken cloud cases, thereby increasing CRF sensitivity. In the three cases considered here, CRF sensitivity to the proposed uncertainty of 50% in retrieved \( v_r \) (equivalent to changing retrieved \( r_s \) by more than the proposed uncertainty of 10%) is typically far weaker than the sensitivity to proposed uncertainty in retrieved aerosol number concentration. These results are intended to aid in the evaluation of the proposed ACE aerosol retrieval requirements.

A notable aspect of these case studies is the large difference between the bimodal aerosol size distribution shape for the RICO case and the monomodal shape for the other cases. As a consequence of the shape difference, the number of aerosols in the 0.1–1-μm radius range is a small fraction of those serving as cloud condensation nuclei in the RICO case and exceeds 25 cm\(^{-3}\) only at the highest \( N_s \) value considered (see Fig. 2). Such low concentrations near the surface could be challenging to detect via satellite. In addition, we have assumed that aerosol numbers in the accumulation mode are perfectly correlated with aerosol numbers in the Aitken mode, but no such correlation is guaranteed in nature. The lack of such correlation would prevent the use of relative changes in humidified aerosol properties in a limited radius range as a reliable proxy for relative changes in total dry aerosol properties, for instance. Thus, to obtain regionally applicable results, the climatology of aerosol number size distribution parameters may be at least as important as thermodynamic conditions, and their co-variability may also be important. Finally, although our assumption of ammonium bisulfate aerosol composition is expected to provide a reasonably accurate representation of aerosol activation properties under a relatively wide range of natural conditions (e.g., VanReken et al. 2003), and the variability of composition is expected to be of secondary importance compared with the variability of aerosol number size distribution properties (e.g., Dusek et al. 2006), the covarying climatology of size-distributed aerosol composition may also be important.

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CORRIGENDUM

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In Fridlind and Ackerman (2011) we mislabeled the panels of Fig. 1. A corrected Fig. 1 is shown below.

**REFERENCE**


**Fig. 1.** Top-of-atmosphere albedo at the end of the simulation with the lowest aerosol number concentration and baseline effective variance for each case.

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