Assessing the Radiative Effects of Global Ice Clouds Based on CloudSat and CALIPSO Measurements

YULAN HONG AND GUOSHENG LIU
Department of Earth, Ocean and Atmospheric Science, Florida State University, Tallahassee, Florida

J.-L. F. LI
Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California

ABSTRACT

Although it is well established that cirrus warms Earth, the radiative effect of the entire spectrum of ice clouds is not well understood. In this study, the role of all ice clouds in Earth’s radiation budget is investigated by performing radiative transfer modeling using ice cloud properties retrieved from CloudSat and CALIPSO measurements as inputs. Results show that, for the 2008 period, the warming effect (−21.8 ± 5.4 W m⁻²) induced by ice clouds trapping longwave radiation exceeds their cooling effect (−16.7 ± 1.7 W m⁻²) caused by shortwave reflection, resulting in a net warming effect (−5.1 ± 3.8 W m⁻²) globally on the earth–atmosphere system. The net warming is over 15 W m⁻² in the tropical deep convective regions, whereas cooling occurs in the midlatitudes, which is less than 10 W m⁻² in magnitude. Seasonal variations of ice cloud radiative effects are evident in the midlatitudes where the net effect changes from warming during winter to cooling during summer, whereas warming occurs all year-round in the tropics. Ice cloud optical depth τ is shown to be an important factor in determining the sign and magnitude of the net radiative effect. Ice clouds with τ < 4.6 display a warming effect with the largest contributions from those with τ ≈ 1.0. In addition, ice clouds cause vertically differential heating and cooling of the atmosphere, particularly with strong heating in the upper troposphere over the tropics. At Earth’s surface, ice clouds produce a cooling effect no matter how small the τ value is.

1. Introduction

Ice clouds play an important role in modifying Earth’s radiation budget via its so-called greenhouse-versus-albedo effects (Liou 1986). That is, it cools Earth by reflecting shortwave (SW) radiation back into space (the solar albedo effect), and at the same time, it warms Earth by reducing outgoing longwave (LW) radiation (the greenhouse effect). The balance of the greenhouse-versus-albedo effects determines whether an ice cloud has a net warming or cooling effect on the earth–atmosphere system, which is influenced by ice cloud microphysical properties such as crystal shape, effective radius, and optical thickness and by ice cloud macrophysical properties such as cloud location and extension in the atmosphere (Eliasson et al. 2011; Baran 2012). It has been well recognized that cirrus clouds, which are high, cold, and optically thin, have net warming effects on the earth–atmosphere system because their greenhouse effect is greater than their solar albedo effect (Liou 1986; Stephens et al. 1990; Khvorostyanov and Sassen 2002).

However, a wide spectrum of ice clouds exist in nature, including optically thick or low-level ice clouds in addition to high and optically thin cirrus. Investigation of the radiative effects across the whole ice cloud spectrum has been lacking, largely because of insufficient global and vertically resolved observations of ice clouds. Previous studies on ice cloud radiative effects have focused either on certain types of ice clouds such as thin cirrus or on ice clouds in certain regions such as in the tropics. For instance, Sun et al. (2011) studied subvisual ice clouds (optical depth τ < 0.3) using a synergy of measurements from the Clouds and the Earth’s Radiant Energy System (CERES), the Moderate Resolution Imaging
Spectroradiometer (MODIS), and the Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO). They demonstrated that these thin ice clouds have a diurnal mean SW radiative effect of $\sim 2.5\, \text{W m}^{-2}$ and that the LW radiative warming effect can reach to $15\, \text{W m}^{-2}$ for ice clouds with $\tau \approx 0.1$. Based on CloudSat and CALIPSO data, Berry and Mace (2014) found that ice clouds of the Asian summer monsoon produce a net warming radiative effect at the top of the atmosphere (TOA) of 21 W m$^{-2}$, and those ice clouds with ice water path (IWP) around 20 g m$^{-2}$ contribute the most significant heating to Earth. Satellite-based studies by Haladay and Stephens (2009) and Lee et al. (2009) have shown the importance of tropical thin ice clouds in the Earth radiation budget. However, whether ice clouds with a wide range of optical depth, globally as a whole, enhance or weaken the net radiation on Earth still remains unknown.

Reflected solar radiation and emitted thermal radiation by ice clouds largely rely on their optical depth (Baran 2012) because longwave effects saturate at an optical depth of 2–4, whereas shortwave albedo continues to vary with optical depth up to much greater values. It is interesting to ask at what optical depth the solar albedo and the greenhouse effects become balanced. Particularly, considering only high clouds in the tropics, the satellite-based study by Choi and Ho (2006) showed that ice clouds with $\tau < \sim 10.0$ warm the tropics. Otherwise, ice clouds cool the tropics. They also pointed out that the warming effect by thin ice clouds is predominantly greater than the cooling effect by thick clouds because of the large area fraction of the thin ones. However, their study mainly focused on certain types of ice clouds (i.e., high clouds) in certain regions (i.e., tropics).

Moreover, recent analyses found that a large disparity of global distribution and mean value of ice water path exists among state-of-the-art global climate models (GCMs), and ice cloud climatology derived from GCM models also shows large discrepancies in both magnitude and spatial distribution with that retrieved from satellite observations (Waliser et al. 2009; Eliasson et al. 2011; Li et al. 2012; Hong and Liu 2015). These discrepancies in turn cause biases in radiation fields. As indicated by Li et al. (2013), different GCMs produce significant regional biases of annual means of global fluxes. A full understanding of ice cloud radiative effects becomes necessary to improve the ice cloud representation in GCMs, and a global distribution of ice cloud radiative effects based on observations can also be used as truth to validate GCMs.

Observations by the cloud radar on CloudSat (Stephens et al. 2002) and the lidar on CALIPSO (Winker et al. 2003) provide an unprecedented opportunity to study ice cloud properties and their radiative effects on both global and regional scales. Vertical structures of ice clouds are resolved by these instruments, which are important in obtaining accurate radiative effects of clouds (Chen et al. 2000). In addition, a combination of the radar and the lidar measurements can retrieve ice clouds with a wide range of optical depth (Sassen et al. 2008), allowing radiative effects of all types of ice clouds to be examined. Based on these observations, Hong and Liu (2015) studied the climatological properties of global ice clouds with varying optical depths. Taking all seasons and regions into account, they found that all ice clouds are estimated to cover about 50% of Earth. Considering that cirrus clouds cover only about 20% of the globe (Liou 1986), noncirrus ice clouds should cover an even larger fraction; thus, their impact on Earth’s radiation balance can be significant.

Motivated by the lack of systematic analysis of the radiative effect of all ice clouds, the goal of this study is to assess the radiative effect globally and over the entire ice cloud spectrum by taking advantage of the availability of combined CloudSat and CALIPSO ice cloud observations. The investigation will include the assessment of the signs and magnitudes of ice cloud radiative effect as a global average. Spatial distribution, dependency on optical depth, and seasonal and regional variations of ice cloud radiative effects are also examined.

2. Data and methodology

a. Satellite observations and ice cloud retrievals

The most important dataset used for estimating ice cloud radiative effect is the cloud retrievals from CloudSat and CALIPSO. The CloudSat radar is operated at 94 GHz with a minimum sensitivity of $-30\, \text{dBZ}$. The radar-measured profile has a vertical resolution of 240 m with a 1.4-km cross-track by 1.8-km along-track footprint (Stephens et al. 2002). The CALIPSO lidar operates at two wavelengths of 0.532 and 1.064 $\mu$m with an average vertical resolution of 60 m (Winker et al. 2003). Using measurements from the radar and the lidar, it is able to retrieve global ice cloud properties including ice water content (IWC), effective radius $r_e$, and extinction coefficient for a wide range of ice clouds. It is well documented that the combined measurements from the radar and the lidar are able to derive physical properties from the thinnest ice clouds (detected by lidar only) to the thickest ice clouds because of the distinctive sensitivities to particle sizes of these two sensors (Sassen et al. 2008; Schwartz and Mace 2010). Lidar measurements are necessary to detect optically thin clouds in the upper troposphere, which the radar misses (Li et al. 2012). Although these missed ice clouds by the radar
contribute little to the total amount of atmospheric cloud ice mass, their radiative effects are significant, particularly for LW radiation (Haladay and Stephens 2009). For this reason, the CloudSat science team level-2B cloud water content radar-only product (2B-CWC-RO; Austin 2007), which derives ice cloud properties based on radar measurements only, is not used in this study. Instead, two other data products—that is, radar–lidar (DARDAR; Delanoë and Hogan 2008, 2010) and the CloudSat and CALIPSO level-2C ice cloud property product (2C-ICE; Deng et al. 2010)—are adopted to calculate ice cloud radiative effect, in which the radar and the lidar data are used jointly to derive cloud properties. The algorithms of DARDAR and 2C-ICE are briefly described as follows.

The DARDAR data product is developed at the University of Reading (Delanoë and Hogan 2008, 2010). The DARDAR algorithm is based on an optimal estimation framework, in which a state vector (extinction coefficient, extinction-to-backscattering ratio, and number concentration) for a single profile is first guessed to predict the observation vector (lidar backscattering and radar reflectivity factor). While Delanoë and Hogan (2010) demonstrated their algorithm could make use of infrared radiances, the DARDAR dataset uses only radar and lidar signals in the retrieval. The predicted observation vector based on a forward model is then compared to the actual observation. The state vector is iteratively adjusted until the differences between the observation vector and the corresponding predicted observations are minimized in a least squared sense. In the DARDAR algorithm, the unified particle size distribution of Field et al. (2005) is adopted to compute the bulk microphysical properties (e.g., IWC and \( r_e \)). For lidar signals, multiple scattering of atmospheric molecules and clouds in the lidar signal is accounted for using the fast multiple-scattering model of Hogan (2006). The mass–size relationship of Brown and Francis (1995) and the corresponding area–size relationship of Francis et al. (1998) are used to connect ice particle’s size, area, and mass. In the radar-only region, the DARDAR algorithm tends toward an empirical retrieval using radar reflectivity factor and temperature (Liu and Illingworth 2000; Hogan et al. 2006). Identification of cloud phase in 2C-ICE relies on CloudSat radar and CALIPSO lidar level-2B cloud classification product (2B-CLDCLASS-LIDAR; Wang et al. 2012), which identifies cloud phase and cloud type by taking advantage of the different sensitivities of the radar and lidar to cloud particle size. That is, the lidar is more sensitive to water droplet and the radar is more sensitive to ice crystals. Mixed-phase cloud is assumed to exist in regions with temperature ranging from \(-40^\circ\) to \(0^\circ\)C and where lidar echo experiences a strong increase initially, followed by a sharp signal decrease (Wang et al. 2012). 2C-ICE retrieval includes the ice in mixed-phase clouds.

An earlier study by Deng et al. (2013) suggests that 2C-ICE and DARDAR agree reasonably well with each other, and both retrievals are consistent with in situ observations. In this study, both DARDAR and 2C-ICE are, respectively, used to calculate radiative fluxes.

b. Radiative transfer modeling

Using ice cloud retrievals along with reanalysis and other satellite products to represent atmospheric components, we are able to conduct radiative transfer modeling of radiative fluxes at any level in the atmosphere. A radiative transfer model called libRadtran (library for radiative transfer; Mayer and Kylling 2005) is used in this study. LibRadtran provides flexible options related to surface and atmospheric components for users to set up. In all our calculations, we apply the discrete ordinate solver (DISORT) version 2 (Stamnes et al. 1988) with two streams as the radiative transfer scheme. The correlated-k distribution method developed by Fu and Liou (1992) is used for atmospheric absorption. The solar zenith angle (SZA), which is needed to determine incoming solar radiation, is computed based on Liou (2002).
Radiative fluxes are computed at pixel levels of DARDAR or 2C-ICE retrievals. Ice cloud properties are from DARDAR and 2C-ICE data. Liquid cloud properties are from CloudSat 2B-CWC-RO product, which provides water cloud properties such as liquid water content (LWC) and cloud drop effective radius (Austin 2007). To transfer cloud microphysical to optical properties, we use the parameterization of aggregate ice habit from Yang et al. (2000, 2005) for ice clouds and Mie theory for water clouds. Atmospheric temperature, ozone, pressure, and humidity are adopted from the CloudSat European Centre for Medium-Range Weather Forecasts (ECMWF) auxiliary products (ECMWF-AUX; Partain 2004). Land surface albedo is based on the MODIS land albedo product (MOD43B3) with 16-day temporal resolution in six wavelengths—0.47, 0.55, 0.67, 0.86, 1.24, and 2.13 μm (Moody et al. 2005). Ocean and sea ice albedos are assigned to fixed values at each wavelength, based on the database by Bowker et al. (1985). Where there are both sea ice and open water in a grid, we obtain the averaged albedo by weighting them with their respective area fractions, which allows smooth transition of surface albedo between open water and sea ice. Sea ice area fraction is from the sea ice concentration product, obtained from the National Snow and Ice Data Center (Peng et al. 2013). Tropospheric aerosols are represented by CALIPSO aerosol profile product (CAL_LID_L2_05kmAPro-Prov-V3-01; Young et al. 2008). Aerosol types are determined by CALIPSO vertical feature mask data (CAL_LID_L2_VFM-ValStage1-V3-01; Vaughan et al. 2005). For each aerosol type, we assign a value of aerosol asymmetric factor and single scattering albedo based on the database of Hess et al. (1998) for optical properties of aerosols and clouds. Precipitation is interpreted primarily based on CloudSat level-2C rain profile product (2C-RAIN-PROFILE; L’Ecuyer and Stephens 2002), which includes the liquid precipitation water content over ocean. When there are no retrievals in a rainy condition in CloudSat 2C-RAIN-PROFILE, such as rain over land, but the index for scene classification included in DARDAR (DARMASK_Simplified_Categorization product) indicates rain, we use a water content value of 0.15 g m⁻³ with effective radius of 25 μm. A similar treatment has been used by L’Ecuyer et al. (2008) for radiative flux calculation. The parameters used in the radiative transfer calculations are summarized in Tables 1–3.

Note that CloudSat and CALIPSO provide instantaneous measurements only twice a day, lacking diurnal variations of clouds (Rossow and Zhang 2010). In light of the diurnal cycle of solar insolation, a daily average of fluxes is obtained by performing the radiative transfer computations several times a day in 2-h increments to account all possible SZAs. For the purpose of calculating ice cloud radiative effect, for every pixel at each SZA, the model run will be performed twice, one for all sky (with ice cloud) and the other one for no-ice sky; the latter is the condition where all variables are kept the same but ice cloud is removed by setting IWC to be zero. Note that for the case of a mixed-phase cloud, only IWC is set to be zero while liquid water content is kept unchanged for no ice calculations. The fluxes and heating rates of all and no-ice skies are calculated for every CloudSat pixel, and then they are output with a vertical resolution of 1 km.

To ensure the model can simulate the Earth radiation reasonably, we check our method in two steps. First, instantaneous fluxes using 1-month data (January 2008) are calculated and compared to the level-2B atmospheric fluxes and heating rates product (2B-FLXHR-LIDAR, version P2), from the CloudSat Data Processing Center (L’Ecuyer et al. 2008; Henderson et al. 2013). The CloudSat 2B-FLXHR-LIDAR algorithm computed atmospheric instantaneous fluxes and heating rates product in both clear and cloudy skies using CloudSat, CALIPSO, and MODIS observations. The product is not used in this study directly for ice cloud radiative effect evaluations because it cannot separate the radiative effects of ice clouds when they occur above warm clouds. However, 2B-FLXHR-LIDAR is convenient to

<table>
<thead>
<tr>
<th>Wavelength (μm)</th>
<th>0.47</th>
<th>0.55</th>
<th>0.67</th>
<th>0.86</th>
<th>1.24</th>
<th>2.13</th>
</tr>
</thead>
<tbody>
<tr>
<td>Open water</td>
<td>0.026</td>
<td>0.031</td>
<td>0.042</td>
<td>0.100</td>
<td>0.060</td>
<td>0.020</td>
</tr>
<tr>
<td>Sea ice</td>
<td>0.900</td>
<td>0.900</td>
<td>0.900</td>
<td>0.854</td>
<td>0.368</td>
<td>0.052</td>
</tr>
<tr>
<td>Land</td>
<td>From MOD43B3 land product.</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

### Table 2. Single scattering albedo (SSA) and asymmetric factor g for the aerosol types used for radiative transfer modeling.

<table>
<thead>
<tr>
<th>Aerosol type</th>
<th>Marine</th>
<th>Dust</th>
<th>Polluted continent</th>
<th>Clean continent</th>
<th>Polluted dust</th>
<th>Smoke</th>
<th>Other</th>
</tr>
</thead>
<tbody>
<tr>
<td>SSA</td>
<td>0.997</td>
<td>0.888</td>
<td>0.892</td>
<td>0.972</td>
<td>0.836</td>
<td>0.209</td>
<td>0.90</td>
</tr>
<tr>
<td>g</td>
<td>0.772</td>
<td>0.729</td>
<td>0.698</td>
<td>0.702</td>
<td>0.764</td>
<td>0.336</td>
<td>0.80</td>
</tr>
</tbody>
</table>
use as a reference for checking our method. Figures 1a–d show comparisons of full-sky SW and LW radiative fluxes from 2C-ICE and 2B-FLXHR-LIDAR. The best agreement is for outgoing LW radiation at TOA with a mean difference of \(-1.7 \text{ W m}^{-2}\) and a root-mean-square (RMS) difference of \(4.6 \text{ W m}^{-2}\). For LW downward fluxes at the surface, we obtain lower values with a mean difference of \(-14.1 \text{ W m}^{-2}\) and an RMS difference of \(16.3 \text{ W m}^{-2}\). For SW instantaneous fluxes at TOA and the surface, RMS differences are relatively large compared to LW fluxes due to cloud scattering in the SW. The bias of SW downward flux is about \(-3.2 \text{ W m}^{-2}\), but it is about \(20.1 \text{ W m}^{-2}\) for SW upward flux at TOA because we set higher sea ice albedo in polar regions. After making sure the model produces reasonably instantaneous fluxes, we run it several times a day to obtain daily flux. The daily flux at the TOA is compared to the flux from CERES Single Scanner Footprint 1° global mean product (SSF1deg; Figs. 1e,f; Wielicki et al. 1996; Doelling et al. 2013). LW flux agrees better with a mean difference of \(-6.2 \text{ W m}^{-2}\) and an RMS difference of \(8.1 \text{ W m}^{-2}\). SW flux mean bias is about \(5.5 \text{ W m}^{-2}\), but RMS difference is relatively large (\(-15 \text{ W m}^{-2}\)).

### c. Computation of ice cloud radiative effect

Ice cloud radiative effect is defined as the net flux difference between conditions with and without ice clouds present as shown by

\[
IRE_{\text{TOA,surface}} = (F \downarrow - F \uparrow)_{\text{all-sky}} - (F \downarrow - F \uparrow)_{\text{no-ice}},
\]

(1)

where \(IRE\) is the ice cloud radiative effect at either TOA or Earth’s surface, and \(F \downarrow\) and \(F \uparrow\) represent downward and upward fluxes, respectively. Ice cloud radiative effect in the atmosphere is computed using atmospheric heating rate difference between the above two conditions:

\[
IRE_{\text{atmosphere}} = \left( \frac{\partial T}{\partial t} \right)_{\text{all-sky}} - \left( \frac{\partial T}{\partial t} \right)_{\text{no-ice}},
\]

(2)

where \(T\) is temperature, and \(t\) is time. For this definition, it considers warm clouds as background information, which reflect a large amount of solar radiation, enhancing ice cloud scattering and absorption of SW radiation. At the same time, warm clouds absorb thermal radiation from the surface, reducing absorption of LW radiation in ice clouds. For some previous research, an alternative definition of \(IRE\) is used—that is, flux difference between ice-only and clear skies, which is independent on warm cloud properties (Chen et al. 2000; Berry and Mace 2014). Comparisons of \(IRE\) from two definitions are discussed in appendix B.

Also, the effective ice cloud radiative effect at any location, \(5°\) longitude by \(5°\) latitude on the globe and \(5°\) latitude by 1-km height in the atmosphere, is calculated by

\[
IRE = \frac{\sum_{i=1}^{N} IRE_{i}}{N} = \frac{\sum_{j=1}^{M} IRE_{j}}{M} = \frac{\sum_{i=1}^{N} \sum_{j=1}^{M} IRE_{i,j}}{NM} = fIRE_{\text{conditional}},
\]

(3)

where \(N\) and \(M\) represent the numbers of total and ice-cloudy samples at the location, respectively. Thus, at any location the ice cloud radiative effect is determined by ice cloud occurrence frequency and conditional ice cloud radiative effect.

In interpreting the ice cloud radiative effect calculated from two ice cloud retrievals, we use the averaged value of \(IRE\) calculated from DARDAR and 2C-ICE as the “mean”—namely, \(IRE = 0.5(IRE_{\text{DARDAR}} + IRE_{\text{2C-ICE}})\) and the difference of \(IRE\) resulting from these two retrievals as uncertainties (i.e., \(Err = |IRE_{\text{DARDAR}} - IRE_{\text{2C-ICE}}|\)). The difference represents the uncertainties of ice cloud radiative effect due to the uncertainty in the retrievals of ice cloud properties.

### 3. Radiative effects by all ice clouds

Using one year (2008) of satellite and ancillary data, radiative fluxes are computed and ice cloud radiative effects at TOA, at the surface, and within the atmosphere are analyzed. As defined in section 2c, the averages of DARDAR and 2C-ICE for \(IRE\) are used in discussions as mean states, and their differences (absolute value) are
FIG. 1. (a)–(d) Comparisons of instantaneous fluxes calculated using 2C-ICE with CloudSat 2B-FLXHR-LIDAR product from January 2008. (e),(f) Comparisons of annual mean fluxes computed from 2C-ICE with CERES SSF1deg product from January to December 2008. Each data point is an averaged value in a 5° x 5° latitude–longitude grid box.
used to represent the uncertainty of the result. A more detailed comparison of DARDAR and 2C-ICE is discussed in appendix A.

a. Mean states

The global distribution of annually averaged SW, LW, and net ice cloud radiative effects at TOA and the surface are shown in Fig. 2. Figure 2 incorporates all possible measured spectrum and ice particles. At TOA, a large magnitude of SW radiative cooling (e.g., $\leq -20\,\text{W m}^{-2}$) occurs in the tropical convective and the midlatitude storm active regions [Fig. 2a(1)]. The radiative effect of ice clouds in the LW [Fig. 2b(1)] also has large values over the tropical (>$30\,\text{W m}^{-2}$) and midlatitude storm active regions. Relatively small values ($\leq 15\,\text{W m}^{-2}$) occur in the subtropics and near the poles.

The patterns of both SW and LW radiative effects are highly correlated to ice cloud occurrence frequencies and IWP as shown in Hong and Liu (2015).

In the tropics, although ice clouds have a large impact on both SW and LW radiative fluxes, cloud radiative effect in the LW generally dominates [Fig. 2c(1)], resulting in a net positive radiative effect (warming) with values exceeding 15 W m$^{-2}$ in deep convective regions. Negative values (cooling) of radiative effect occur over the Southern and North Pacific Oceans, where the cooling effect ranges from $-10$ to $-2$ W m$^{-2}$. The regional variation of the net radiative effect stems from the dependency of the radiative fluxes on ice cloud characteristics. As indicated in Hong and Liu (2015), tropical ice clouds mostly occur in high altitudes, whereas mid- and high-latitude ice clouds are mostly at

FIG. 2. Global distributions of (a) SW, (b) LW, and (c) net radiative effects of all ice clouds at TOA and the surface. (right) Zonal means of global ice cloud radiative effects at TOA and the surface for SW, LW, and net effects. Error bars represent the absolute value of ice cloud radiative effect differences between DARDAR and 2C-ICE.

Unauthenticated | Downloaded 08/14/22 05:07 AM UTC
lower levels. High ice clouds in the tropics produce a net warming effect because they are much colder than the surface, so they reduce outgoing radiation more efficiently. In contrast, in the midlatitudes, cooling occurs as a result of a weakened LW radiative effect caused by similar temperatures to the surface. In the high latitudes, the net warming effect is obtained largely as a result of the decrease of solar insolation, which limits the SW cooling effect of ice clouds.

Ice cloud radiative effects at the surface are also shown in Fig. 2. The distribution of SW radiative effect [Fig. 2a(2)] displays similar magnitude and pattern as those at TOA [Fig. 2a(1)]. Unlike LW radiative effect at TOA, the LW effect at the surface [Fig. 2b(2)] is weak (<2 W m\(^{-2}\)) within 40\(^\circ\)S–40\(^\circ\)N because water vapor and/or liquid water clouds mask the effect of ice clouds. Stronger radiative effect (>5 W m\(^{-2}\)) occurs in higher latitudes as the amount of water vapor decreases. Because of the small LW radiative effects in the tropics and midlatitudes, there exists a net cooling effect over these regions at the surface.

Zonal averages for the SW, LW, and net radiative effects by all ice clouds at TOA and the surface are shown in Figs. 2a(3), 2b(3), and 2c(3). The error bars reveal that relatively large uncertainties caused by the differences of ice cloud properties between DARDAR and 2C-ICE occur in the tropics for LW radiative effect at TOA (∼10 W m\(^{-2}\)). For SW radiative effect, the uncertainties are small.

On a global average, ice clouds reduce net solar radiation at TOA by 16.7 W m\(^{-2}\) with an uncertainty of 1.7 W m\(^{-2}\) and trap outgoing thermal radiation of 21.8 W m\(^{-2}\) with an uncertainty of 5.4 W m\(^{-2}\). As a result, ice clouds, on globally average, cause radiative heating of the earth–atmosphere system by 5.1 W m\(^{-2}\) with an uncertainty of 3.8 W m\(^{-2}\). At the surface, SW and LW radiative effects are about −17.3 and 5.4 W m\(^{-2}\), respectively, causing a net cooling effect of −11.8 W m\(^{-2}\) with an uncertainty of 1.5 W m\(^{-2}\). The relatively large uncertainty in the LW radiation term at TOA is attributed largely to the fact that DARDAR retrieves larger IWC values than 2C-ICE in the tropics (see appendix A).

Similarly, the ice cloud–induced heating or cooling within the atmosphere is also investigated, and the results are shown in Fig. 3. Ice cloud radiative effect in the atmosphere is defined using heating rate difference between all sky and no-ice sky as shown in (2). The SW radiative effect is positive in the upper and negative in the lower atmosphere (Fig. 3a), while the LW effect is the opposite (Fig. 3c). The SW radiative heating pattern can be explained as follows. 1) Ice clouds absorb SW near-infrared radiation, which occurs more in the upper atmosphere. 2) Ice clouds reflect SW radiation, which enhances the absorption by atmospheric molecules (e.g., water vapor and ozone) above the clouds. 3) Ice clouds reduce SW radiation reaching below ice clouds, which leaves less molecular absorption there (Chen et al. 2000). The first two mechanisms enhance warming effect in the upper atmosphere, while the third causes cooling below. Ice clouds absorb upwelling LW radiation and emit it back to the atmosphere at a lower cloud temperature, which causes warming at lower levels (greenhouse effect). This is especially evident in the tropics, where the warming effect extends up to 15 km (Fig. 3c). However, ice clouds enhance the LW cooling above clouds. This enhancement is weak in the tropics and relatively strong in the midlatitudes where cooling effect takes over starting at approximately 5 km. The latitudinal variation of ice cloud–induced heating rates largely arises from the differences among ice clouds in the tropics and midlatitudes. That is, tropical ice clouds occur frequently in the high altitudes, while the midlatitude ones often show in low altitudes (Hong and Liu 2015). Overall, LW radiative heating rates are much stronger in the atmosphere than those of SW because ice clouds absorb LW radiation but scatter far more SW radiation than they absorb (Haynes et al. 2013). Consequently, the pattern of net radiative effects basically follows that of LW (i.e., cooling above and warming below). In particular, tropical ice clouds with altitudes between 9 and 15 km show a positive enhancement for both LW and SW radiative effects, whereas in other regions the SW and LW radiative heating terms have opposite signs; therefore, they cancel with each other to some degree. Also, relatively large uncertainties also occur in the tropics (Figs. 3b,d,f), particularly for the LW radiative effect (see appendix A).

b. Seasonal and regional variations

The monthly zonal averaged SW, LW, and net ice cloud radiative effects at TOA and the surface are shown in Fig. 4. At TOA, the strongest seasonal variation of the SW radiative effect (Fig. 4a) occurs in the midlatitudes, ranging from −10 W m\(^{-2}\) in winter to −40 W m\(^{-2}\) in summer, whereas in the tropics and high latitudes, seasonal variations are much smaller. Seasonal variation for LW radiative effect (Fig. 4c) in the midlatitudes is weaker than that for the SW effect, mainly from 10 W m\(^{-2}\) in warm seasons to 20 W m\(^{-2}\) in cold seasons. Net radiative effect at TOA (Fig. 4e) in the midlatitudes shifts from warming to cooling in a 1-yr cycle. The net warming effect is 10–20 W m\(^{-2}\) during winter and net cooling effect can exceed −30 W m\(^{-2}\) in southern midlatitudes. In the tropics, strong net warming effect (10–20 W m\(^{-2}\)) persists over the whole year, mainly caused by high ice clouds. At the surface, the
Radiative effects are primarily determined in the SW with strong cooling in summer months in the mid-latitudes while variations in the tropics are mild. According to the study of Hong and Liu (2015), midlatitude ice clouds occur more frequently during winter than summer, which explains the stronger LW radiative effect in winter than in summer. The SW radiative effect in the midlatitudes, however, depends heavily on incoming solar radiation in addition to ice cloud occurrence, which largely determines the winter minimum (in absolute value) of the SW radiative effect.

We also computed atmospheric radiative heating rates due to ice clouds in four seasons: March–May (MAM), June–August (JJA), September–November (SON), and...
December–February (DJF) (figures not shown). Heating rates for tropical ice clouds vary with the shift of ITCZ for both SW and LW radiation terms. In the midlatitudes, while LW radiative heating rates do not have clear seasonal changes, SW heating rates are stronger during warm seasons and weaker during cold seasons.

4. Dependence of ice cloud radiative effect on optical depth

Studies by Hong and Liu (2015) showed that ice clouds have optical depth values in the visible ranging from near zero to about 100. The magnitude of ice cloud
Radiative effect should depend on optical depth for the following reasons (Choi and Ho 2006; Lee et al. 2009; Berry and Mace 2014; Hong and Liu 2015). First, ice clouds with different values of optical depth have different occurrence frequencies. Second, ice clouds with larger optical depth would have stronger radiative impacts on both SW and LW fluxes. The balance between the SW and LW radiative effects likewise depends on optical depth. Additionally, the characteristics of ice clouds are regionally dependent. For example, ice clouds with the same optical depth tend to occur at higher altitudes in the tropics than in the midlatitudes, which leads to a different impact on LW and SW radiation. This regional difference of cloud property also has a bearing on the dependency of ice cloud radiative effect on optical depth.

a. Global and annual means

The globally and annually averaged values of ice cloud radiative effects varying with optical depth are shown in Fig. 5, which are averaged by two methods. The conditional mean is the mean given an ice cloud present with the specified optical depth (Figs. 5a,c), while the mean is the conditional mean multiplied by ice cloud occurrence frequency lying in a particular range of optical depth (Figs. 5b,d). For the conditional means, SW radiative effect (negative) at TOA and the surface first monotonically strengthens as \( \tau \) increases from 0 to 20 before it turns to the opposite trend. The reverse of trend is because optically thick ice clouds are often associated with precipitation beneath them, which reflects more SW radiation than clear sky, resulting in a smaller difference of radiative fluxes between all sky and no-ice sky (see discussion in appendix B). The conditional means of LW radiative effect increase with \( \tau \) at TOA for the entire range of \( \tau \) (Fig. 5a), indicating that thicker ice clouds prevent more LW radiation energy from escaping to space. Meanwhile, at the surface, LW radiative effect only increases with \( \tau \) until \( \tau \) is about 20, followed by a

---

Fig. 5. Ice cloud radiative effects as a function of optical depth: conditional means of radiative effects at (a) TOA and (c) the surface and (b),(d) the mean ice cloud radiative effects. Error bars denote the absolute value of IRE differences resulted from discrepancies between DARDAR and 2C-ICE. The x axis is logarithmically scaled.
slow decrease (Fig. 5c). The explanation for this change of trend is that most of thick ice clouds ($\tau > 20$) occur in the storm active regions of tropics and midlatitudes (Hong and Liu 2015), where relatively rich water vapor in the lower atmosphere obscures ice cloud LW radiative effect at the surface.

At TOA, the conditional net radiative effect changes its magnitude and even its sign as optical depth varies (Fig. 5a). When $\tau$ is less than 4.6, the LW radiative warming effect is larger than the SW cooling effect, resulting in a net warming. The largest net warming effect of these ice clouds occurs at $\tau \approx 1.0$. When $4.6 < \tau < 70$, ice clouds show a negative net radiative effect at TOA because of the increase of SW radiative effect, and the largest cooling occurs at $\tau \approx 20$. As $\tau$ increases beyond 70, ice clouds turn back to a net radiative warming again due to a decreased SW radiative effect but an increased LW effect. At the surface, ice clouds are always cooling no matter how small the $\tau$ value is (Fig. 5c).

To obtain the ice clouds’ effective contribution, the mean radiative effects are computed by multiplying the conditional radiative effects by respective occurrence frequencies. The means of radiative effect at TOA and at the surface are shown in Figs. 5b and 5d, respectively. Although ice clouds have a peak occurrence frequency at $\tau \approx 1.0$ as shown by Hong and Liu (2015), the peak SW radiative effects at both TOA and the surface occur at $\tau \approx 5.0$ because the conditional SW effect at $\tau \approx 5.0$ ($\sim -60$ W m$^{-2}$) is much greater than that at $\tau \approx 1.0$ ($\sim -20$ W m$^{-2}$). Meanwhile, ice clouds with $\tau \approx 1.0$ have the strongest LW radiative effect at TOA because the conditional LW effect at $\tau \approx 1.0$ is large ($\sim 40$ W m$^{-2}$). As a result, for the net radiative effect at TOA, ice clouds with $\tau \approx 1.0$ contribute the peak warming, while the peak cooling is at $\tau \approx 10.0$. The net warming of optically thin ice clouds is much stronger than the net cooling of optically thick ice clouds, resulting in a total net warming to the earth–atmosphere system. Although ice clouds with $\tau > 70$ show large conditional net warming effect at TOA, since the frequencies of these clouds are small, their radiative contributions are small. However, these ice clouds may play important roles in convective cloud development because of their largely conditional radiative effects (Chen et al. 2000). At the surface, the peaks of SW, LW, and net radiative effects are located where $\tau$ is from approximately 5 to 10.

Global distributions of the ice cloud radiative effect at TOA for different categories of optical depth are shown.
in Fig. 6. Ice clouds are divided into five groups based on \( \tau \) values. They are subvisual (\( \tau < 0.03 \)), thin (\( 0.03 < \tau < 0.3 \)), opaque (\( 0.3 < \tau < 3.0 \)) ice clouds based on the classification of Sassen and Cho (1992) and two groups for optically thick ice clouds (\( 3.0 < \tau < 20.0 \) and \( \tau > 20.0 \)). Subvisual ice clouds [Fig. 6a(1)] display weak SW and LW radiative effects because of their infrequent occurrence (<5%; Hong and Liu 2015). Their SW radiative effects are close to zero globally, and their LW effects are smaller than 1 W m\(^{-2}\) in the tropics and close to zero elsewhere. For thin ice clouds, LW radiative effects are generally smaller than 5 W m\(^{-2}\) with larger values in the tropics, whereas SW effects are generally between about –1 and about –0.1 W m\(^{-2}\), resulting in a net warming globally. For opaque ice clouds, strong SW and LW radiative effects occur in the tropics and the midlatitude storm active regions, following the pattern of ice cloud occurrence frequencies (Hong and Liu 2015). The LW radiative warming effect exceeds SW cooling effect, causing a net warming globally. Particularly, in the tropical deep convective regions, the warming effect can be over 10 W m\(^{-2}\). Overall, ice clouds with \( \tau < 3.0 \) have a net warming effect globally based on the 1-yr averaged data with peak values being in the tropics and decreasing toward poles. These thin ice clouds frequently occur in the tropics at high altitudes (Hong and Liu 2015), leading to a strong net warming effect in the tropics. For ice clouds with optical depth greater than 3.0, the occurrence frequencies become small compared to the opaque ice clouds (Hong and Liu 2015), but they have strong SW cooling and LW warming radiative effects due to their large conditional values (Fig. 5a). In this case, SW radiative effect is greater than the LW radiative effect in the midlatitudes, leading to a net cooling effect there. In the tropics, a net warming (between about 1 and 5 W m\(^{-2}\)) occurs over land areas, but a net cooling effect (between about –5 and –1 W m\(^{-2}\)) occurs over oceans. High-latitude ice clouds always have warming effects no matter how large the \( \tau \) value is.

While it is not shown, at the surface these five groups of ice clouds cause a net cooling at most areas of Earth except near the polar regions. Table 4 summarizes the globally averaged ice cloud radiative effects for the five categories of ice clouds. Uncertainties are represented by the differences due to the discrepancy between DARDAR and 2C-ICE. At TOA, opaque ice clouds have a global SW radiative effect of about –5 W m\(^{-2}\) and a LW radiative effect of about 10 W m\(^{-2}\), resulting in a net warming effect of 5 W m\(^{-2}\). Thin ice clouds have a net warming effect of approximately 1 W m\(^{-2}\), with a LW radiative effect of 1.6 W m\(^{-2}\) and a SW effect of –0.6 W m\(^{-2}\). For subvisual ice clouds, both the averaged SW and LW radiative effects are close to zero. For ice clouds with \( \tau > 3.0 \), the magnitudes of SW and LW radiative effects are similar, which are about 7 W m\(^{-2}\) for ice clouds with \( 3.0 < \tau < 20.0 \) and about 3 W m\(^{-2}\) for ice clouds with \( \tau > 20.0 \). Although net cooling and net warming by optically thick ice clouds occur in different regions (Fig. 6), they are nearly offset with each other on a global average, resulting in a near-zero net effect. At the surface, all types of ice clouds show cooling effects.

Atmospheric heating and cooling induced by ice clouds with varying \( \tau \) are also assessed, and the results are shown in Fig. 7. The conditional SW and LW radiative heating rates generally increase with \( \tau \) (Figs. 7a and 7c, respectively). Net heating rate resulting from optically thin ice clouds with \( \tau < –1.0 \) is positive throughout atmosphere, while optically thicker ice clouds cause a dipole heating pattern in the vertical. The altitude at which the transition between heating and cooling occurs decreases with the increase of \( \tau \) when \( \tau < 10 \), but the trend is reversed for higher \( \tau \) values. Such a pattern is thought to be mainly caused by the regional characteristics of ice clouds. For instance, ice clouds with \( \tau \sim 10 \) occur more frequently at lower altitudes in the midlatitudes than in the tropics (Hong and Liu 2015), which can cause strong cooling at low altitudes.

<table>
<thead>
<tr>
<th></th>
<th>Total</th>
<th>( \tau &lt; 0.03 )</th>
<th>0.03 &lt; ( \tau &lt; 0.3 )</th>
<th>0.3 &lt; ( \tau &lt; 3.0 )</th>
<th>3.0 &lt; ( \tau &lt; 20 )</th>
<th>( \tau &gt; 20 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>TOA</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SW</td>
<td>–16.72 ± 1.67</td>
<td>–0.02 ± 0.01</td>
<td>–0.59 ± 0.24</td>
<td>–4.65 ± 1.34</td>
<td>–7.76 ± 0.64</td>
<td>–3.71 ± 0.48</td>
</tr>
<tr>
<td>LW</td>
<td>21.78 ± 5.37</td>
<td>0.07 ± 0.04</td>
<td>1.55 ± 0.60</td>
<td>9.63 ± 3.88</td>
<td>7.27 ± 1.13</td>
<td>3.28 ± 0.21</td>
</tr>
<tr>
<td>Net</td>
<td>5.07 ± 3.80</td>
<td>0.05 ± 0.04</td>
<td>0.96 ± 0.37</td>
<td>4.98 ± 2.57</td>
<td>–0.49 ± 0.71</td>
<td>–0.43 ± 0.34</td>
</tr>
<tr>
<td>Surface</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SW</td>
<td>–17.25 ± 1.66</td>
<td>–0.02 ± 0.01</td>
<td>–0.52 ± 0.22</td>
<td>–4.67 ± 1.29</td>
<td>–8.21 ± 0.67</td>
<td>–3.83 ± 0.45</td>
</tr>
<tr>
<td>LW</td>
<td>5.41 ± 0.35</td>
<td>0.01 ± 0.01</td>
<td>0.21 ± 0.08</td>
<td>1.84 ± 0.27</td>
<td>2.73 ± 0.16</td>
<td>0.62 ± 0.07</td>
</tr>
<tr>
<td>Net</td>
<td>–11.84 ± 1.49</td>
<td>–0.01 ± 0.001</td>
<td>–0.31 ± 0.16</td>
<td>–2.83 ± 1.16</td>
<td>–5.48 ± 0.61</td>
<td>–3.21 ± 0.41</td>
</tr>
</tbody>
</table>
Mean ice cloud–induced heating rates (Figs. 7b,d,f) are obtained by multiplying the conditional heating rates by occurrence frequencies. Contribution by optically thick ice clouds to the heating or cooling is reduced because of their small occurrence frequencies. Ice clouds with $0.2 < \tau < 10.0$ contribute the most because of their frequent occurrence. In particular, ice clouds with $\tau \approx 1.0$ at altitudes of 10–15 km produce the strongest heating in the troposphere.

\[\text{b. Regional and seasonal variations}\]

As shown in the previous section, annual ice cloud radiative effect varies with ice cloud optical depth. However, the $\tau$ value that corresponds to zero net effect can vary with regions and seasons because of the evolution of solar isolation during the year (Harrison et al. 1990) and the variation of the regional characteristics of ice clouds.
The zonally averaged (weighted by occurrence frequency) ice cloud radiative effects at TOA as a function of optical depth are shown in Fig. 8 for four seasons as defined in section 3b. Overall, SW and LW radiative effects are primarily contributed by ice clouds with $0.3 < \tau < 20.0$ in the tropics and by ice clouds with $1 < \tau < 20$ in the midlatitudes. It is shown at TOA that the optical depth value corresponding to zero net radiative effect depends on seasons and regions (Fig. 8). In the tropics, the separation of warming and cooling happens at $\tau \approx 6.0$, and this value is nearly the same throughout the year. This separation value derived for all ice clouds is somewhat smaller than that of Choi and Ho (2006), who found that merely high ice clouds with $\tau < 10.0$ warm the tropics; otherwise, cooling effects occur. As discussed earlier, when $\tau$ is very large, net warming also occurs as a result of a decrease of SW radiative effect (section 4a). Here we show that this phenomenon primarily happens in the tropics as $\tau > 70$. In the midlatitudes, the balance of SW and LW radiative effects is more dependent on season than optical depth. For instance, midlatitude ice clouds have a warming effect during winter no matter how large the $\tau$ value is. But they become strongly cooling during summer especially when $\tau > 1$. In high latitudes, ice clouds have a warming effect overall except in the summer season when they show only a weak cooling effect. Seasonal variations of the ice cloud radiative effect shown in this section agree with results in section 3b. In addition, which types of ice clouds lead to dominant cooling or warming over the globe in different seasons is further examined here.

Besides optical depth, cloud height is another factor that influences ice cloud radiative effect. As mentioned above, strong cooling in the midlatitudes is mainly caused by low-level ice clouds, while tropical ice clouds with the same $\tau$ values have warming effect largely due to their high altitudes. Figure 9 shows the net radiative effect of ice clouds at TOA as cloud top and optical depth vary. Cloud top is determined by the highest level with ice water content greater than zero. Figure 9 shows results for the globe, tropics, midlatitude warm season (MLW), midlatitude cold season (MLC), high-latitude warm season (HLW), and high-latitude cold season (HLC).
Warming seasons are March through August in the Northern Hemisphere and September through February in the Southern Hemisphere, and cold seasons are March through August in the Southern Hemisphere and September through February in the Northern Hemisphere. The separation between the tropics and the midlatitudes is 30° and between the mid- and high latitudes is 60°.

For global mean (Fig. 9a), if \( \tau < 1 \), ice clouds have warming effect no matter how low the cloud-top height is. Net cooling occurs as \( \tau \) becomes larger; the warming–cooling transition starts at lower \( \tau \) values for clouds with

![Graphs showing the net radiative effect of ice clouds at TOA for different regions: (a) global, (b) tropics, (c) midlatitude warm season, (d) midlatitude cold season, (e) high-latitude warm season, and (f) high-latitude cold season.](image)

Fig. 9. Net radiative effect of ice clouds at TOA: (a) global and annual means, (b) tropics, (c) midlatitude warm season, (d) midlatitude cold season, (e) high-latitude warm season, and (f) high-latitude cold season.
lower cloud tops. For ice clouds with cloud top higher than approximately 15 km, there is always a net warming effect, which is mainly contributed by ice clouds in the tropics. In the tropics (Fig. 9b), ice clouds mostly produce warming except for those with cloud top <15 km and \( \tau > 10 \). In the midlatitudes, for warm seasons only those ice clouds that are thin and high (e.g., \( \tau < 3.0 \) and cloud top >8 km) produce net warming (Fig. 9c), while for cold seasons, ice clouds generally have a warming effect unless they are at very low altitudes (Fig. 9d). In the high latitudes, cold season ice clouds always produce a warming effect (Fig. 9f). But in warm seasons, they also produce cooling effects if they are low and optically thick (Fig. 9e). Results shown in Fig. 9 manifest which types of ice clouds, in terms of optical depth and cloud top, contribute the most warming and cooling to the earth–atmosphere system. Tropical ice clouds with \( \tau \approx 1.0 \) and cloud top of 13–16 km have the strongest warming effect, whereas the strongest cooling effects are contributed by ice clouds with \( \tau \approx 10 \) and cloud top below 13 km in the midlatitude warm seasons. In all regions and seasons, the strongest net warming effect comes from ice clouds with \( \tau \approx 1.0 \), and the largest cooling contribution is from ice clouds with \( \tau \approx 10 \).

5. Discussion and conclusions

Ice cloud properties for a wide spectrum of ice clouds were produced by both DARDAR and 2C-ICE algorithms, which utilize combined measurements from satellite radar and lidar. In this study, we use both DARDAR and 2C-ICE retrievals to investigate the radiative effects of the entire spectrum of ice clouds. Our analysis covers from global to regional scales, from annual average to seasonal variations and from subvisual to precipitating ice clouds.

A notable finding from this study is that the global net radiative effect by the whole spectrum of ice clouds is warming to the earth–atmosphere system. Globally, at TOA, their LW radiative warming due to greenhouse effect (\( \sim 21.8 \pm 5.4 \) W m\(^{-2}\)) exceeds their SW radiative cooling due to solar albedo effect (\( \sim 16.7 \pm 1.7 \) W m\(^{-2}\)). At the surface, the SW radiative effect (\( -17.3 \pm 1.7 \) W m\(^{-2}\)) exceeds the LW radiative effect (\( 5.4 \pm 0.4 \) W m\(^{-2}\)), causing strong cooling. Within the atmosphere, ice clouds generally cause radiative net warming at the lower and net cooling at higher levels.

At TOA, the zonally averaged LW radiative effect in the tropics is about 40 W m\(^{-2}\) while SW radiative effect is around -25 W m\(^{-2}\), resulting in a net warming effect by ice clouds of more than 10 W m\(^{-2}\) throughout the year. In the midlatitudes, primarily because of variation of solar insolation, SW radiative cooling effect by ice clouds reaches to \(-40 \) W m\(^{-2}\) in summer and decreases to about \(-10 \) W m\(^{-2}\) in winter. The LW radiative effect, on the other hand, varies from 10 to 20 W m\(^{-2}\), being weaker in summer and stronger in winter. Thus, the net radiative effects by ice clouds change signs from positive (warming) in winter to negative (cooling) in summer. In high latitudes, the annual cycle of the radiative effects has a similar trend to that in the midlatitudes, but with smaller amplitude.

Another outcome of this study is the identification of how different types of ice clouds (in terms of optical depth) contribute to the SW and LW radiative fluxes at TOA, at the surface, or in the atmosphere. When averaged globally, ice clouds with \( \tau < 4.6 \) warm the earth–atmosphere system with the largest contribution by those with \( \tau \approx 1.0 \), while those with \( 4.6 < \tau < 70 \) cool the earth as a result of a strong solar albedo effect. When \( \tau > 70 \), the radiative effect turns back to positive (warming) again, primarily because optically thick ice clouds are often associated with rain below them (see appendix B). When we focus on the tropics, the value that separates the radiative warming and cooling effects is \( \tau \approx 6 \), and it does not vary much through the year. However, the separation \( \tau \) value changes greatly in the mid- and high latitudes as the season changes.

Global distributions of ice cloud radiative effect indicate that ice clouds with \( \tau < 3.0 \) show net warming effect everywhere. Particularly, opaque ice clouds (\( 0.3 < \tau < 3.0 \)) contribute the most to net warming effect (\( \sim 5 \) W m\(^{-2}\)) among all ice cloud types. For optically thicker ice clouds (e.g., \( \tau > 3.0 \)), net cooling effects occur in the midlatitudes and tropical oceanic regions (between about \(-5 \) and \(-1 \) W m\(^{-2}\)), while net warming effects occur elsewhere (between about 1 and 5 W m\(^{-2}\)). As a global average, these ice clouds show the LW radiative effect offsets the SW effect, resulting in a nearly zero net effect. At the surface, globally averaged radiative effects show all ice clouds produce cooling. In the atmosphere, ice clouds produce heating in the lower and cooling in the upper levels. The strongest heating signature is found at the upper troposphere of 10–15 km in the tropics, with ice clouds with \( \tau \approx 1.0 \) contributing the most significantly to the warming.

The net radiative effect of ice clouds at TOA clearly depends on cloud-top height as well. Optically thin ice clouds tend to have a net warming effect even though their altitudes are low. Optically thick ice clouds need to be located at high altitudes to have the SW and LW radiative effects in balance. On a global average, ice clouds with \( \tau \approx 1.0 \) have a warming effect if the cloud top is higher than 5 km; otherwise, they have a cooling effect. Meanwhile, ice clouds with \( \tau \approx 10 \) tend to have a cooling effect unless their tops are higher than 15 km.
The altitude at which transition between net warming and net cooling occurs also differs in different seasons. In the midlatitude warm season, optically thick (e.g., $\tau > 3.0$) ice clouds tend to have a cooling effect no matter how high the cloud top is. But in cold season, they have a warming effect if the cloud top is higher than 8 km. In high latitudes, ice clouds always have a net warming effect regardless of cloud-top height during cold season, while they have a cooling effect during warm season if the cloud top is lower than 10 km and optical depth is large enough (e.g., $\tau \approx 10$).

Observations of the CloudSat radar and the CALIPSO lidar provided unprecedented ice cloud properties with detailed vertical structures. Taking advantage of these state-of-the-art measurements, the ice cloud radiative effects are characterized at TOA, at the surface, and in the atmosphere. The global mean state, seasonal and regional variations, and dependency on cloud optical depth of the ice cloud radiative effects enhance our understanding of ice cloud–radiation interactions. The results presented in this study can be useful for validating the impact of ice clouds on corresponding radiation fields in GCMs such as in CMIP5.

Finally, readers should be cautioned when interpreting the radiative effect values due to the uncertainties associated with ice particle shapes and their optical properties. In this study, we only adopted one ice habit (aggregate) for calculating ice cloud radiative effect because this ice habit covers a relatively broad size range (Heymsfield et al. 2002). Different parameterizations of ice optical properties will alter the values of ice cloud radiative effect. For example, Key et al. (2002) showed various ice habits induce over 15% of differences in SW radiative fluxes. Baran (2012) and Baran et al. (2014) showed that GCM-simulated LW and SW radiative fluxes differ by 15% of differences in SW radiative effect. For example, Key et al. (2002) showed ice optical properties will alter the values of ice cloud radiative effect locally by over 10 W m$^{-2}$.

APPENDIX A

Comparisons of Ice Cloud Properties and Ice Cloud Radiative Effects between DARDAR and 2C-ICE

Ice cloud radiative effects shown in this paper display relatively large uncertainties in the LW radiation at TOA [Fig. 2b(3)] and in the tropical atmosphere around 10–15 km (Fig. 3d). To understand these uncertainties, we compare ice cloud properties and ice cloud radiative effects derived from DARDAR and 2C-ICE.

Figure A1 shows the probability distribution functions (PDFs) of ice cloud optical depth, IWP, IWC, and $r_c$ in the tropics, midlatitudes, and high latitudes. Globally averaged $\tau$ and IWP distributions [Figs. A1a(4) and A1b(4)] show similar curves between DARDAR and 2C-ICE with the mode of $\tau$ around 1.0 and the mode of IWP around 25 g m$^{-2}$. The 2C-ICE data indicate a secondary mode at smaller $\tau$ and IWP, whereas DARDAR shows sharper distributions. For IWC [Fig. A1c(4)], the mode of DARDAR IWC (0.005 g m$^{-3}$) is twice that of 2C-ICE (0.0025 g m$^{-3}$), and the width of 2C-ICE IWC distribution is wider than that for DARDAR. For $r_c$ [Fig. A1d(4)], the mode of DARDAR $r_c$ (35 $\mu$m) is slightly greater than that of 2C-ICE (33 $\mu$m).

In different regions, the most noticeable differences between DARDAR and 2C-ICE happen in the tropics. First, double-mode patterns of IWP and $\tau$ distributions from 2C-ICE are evident for the tropical ice clouds. As...
shown in Fig. A1a(1), τ distribution of 2C-ICE manifests two equally important modes (i.e., τ = 0.05 and 1.0), and, in Fig. A1b(1), two modes of IWP are evident around 0.8 and 15 g m^{-2}. However, both τ and IWP distributions of DARDAR in the tropics merely show one evident and sharp mode. In terms of IWC in the tropics, the mode of IWC distribution is around 0.8 mg m^{-3} and the one of DARDAR is about 5 mg m^{-3}. For r_e in the tropics, both DARDAR and 2C-ICE generally agree well with each other. Note that the modes of IWC and r_e in the tropics mainly result from ice clouds between 10–15 km (Hong and Liu 2015).

Outside the tropics, DARDAR and 2C-ICE are very consistent with each other in both IWP and τ distributions. The differences in IWC distributions are much smaller than those in the tropics. For r_e, DARDAR generally agrees with 2C-ICE although there is some disagreement for r_e < 40 μm.

The PDFs of the differences between the ice cloud radiative effects when using DARDAR and 2C-ICE are shown in Fig. A2 with a monthly scale 5° grid in the tropics, mid-latitudes, and high latitudes. At both TOA and the surface, the difference PDFs peak around 0 W m^{-2} for the SW, LW, and net radiative effects. The radiative effects of ice clouds resulted from DARDAR and 2C-ICE agree the best at the tropical surface where the histograms show the narrowest width [Fig. A2b(2)]. The reason is that the LW surface radiation depends strongly on warm clouds and humidity in the tropics which are identical in the DARDAR and 2C-ICE calculations. The largest discrepancies occur in tropical TOA with SW differences over ~10 W m^{-2} [Fig. A2a(1)] and LW differences over
Differences outside the tropics usually fall within 10 W m\(^{-2}\). Thus, relatively large uncertainties of ice cloud radiative effect discussed in section 3a are caused by relatively large discrepancies of ice cloud properties between DARDAR and 2C-ICE in the tropics and by larger ice cloud occurrence of DARDAR than 2C-ICE (see Table B1). These ice clouds mostly occur at altitudes of 10–15 km, which primarily impact on the LW fluxes.

**APPENDIX B**

**Influence of Water Cloud and Rain on Ice Cloud Radiative Effects**

In section 4, it is noticed that when \( \tau > 20 \), ice cloud SW radiative effect reverses its trend from increase to decrease with \( \tau \) (Fig. 5a). The trend reversal is a result of the association of liquid water clouds or rain with thick ice clouds. Since ice cloud radiative effect is defined as the flux difference between all and no-ice skies, the increase of SW reflection by liquid water clouds or rain (relative to clear sky) can reduce ice cloud SW radiative effect. To verify this explanation, we first classify all ice clouds into four groups: ice cloud only, ice cloud with liquid water cloud below, ice cloud with rain below, and ice cloud with both liquid water cloud and rain below. We then also assess how different the ice cloud radiative effect would be when using an alternate definition which is independent of warm clouds (i.e., flux difference between ice-only and clear skies).

Figure B1a shows the number of occurrence of the above four groups for different ranges of optical depth derived from DARDAR product (results from 2C-ICE are similar, not shown). For optically thin ice clouds (e.g., \( \tau < 3.0 \)), the most frequent group is ice cloud alone, followed by ice cloud associated water cloud below. Very few ice clouds occur with rain. As optical depth increases, the number of ice cloud occurring with water cloud decreases sharply, while the number of ice cloud with rain increases instead. When \( \tau > 20 \), the number of ice cloud with rain is even greater than that of ice only.

The computed conditional SW and LW radiative effects based on DARDAR are shown in Figs. B1b and B1c, respectively, for the above four cloud groups and an “all” group which combines all the four groups. The SW radiative effect for the group of ice cloud alone is stronger than any other groups, indicating that water cloud or rain below ice clouds cause small SW radiative effect because liquid water reflects a large amount of solar radiation. In particular, when \( \tau > 20 \), the overall
SW radiative effect roughly follows the trend for the ice cloud with rain group, primarily due to the relatively large number of the latter group. It is noted that the SW radiative effect of ice clouds with rain (green line) or liquid water cloud (blue line) shows a decrease in absolute value when ice optical depth value becomes larger than approximately 30. This reduction can be explained as follows. SW reflection becomes nearly leveled (no change with optical depth) for full-sky conditions when ice clouds is thick (optical depth $\tau > 30$). However, a thicker ice cloud is often associated with a deeper rainwater layer below so that the no-ice sky upward SW radiative flux still keeps increasing with optical depth. As a result, for ice optical depth larger than approximately 30, the difference between upward SW radiative fluxes of full sky and no-ice sky decreases as optical depth increases. Ice cloud with water cloud or rain below generally produces weak LW radiative effects when ice cloud optical depth is small (e.g., $\tau < 1$) because water clouds below absorb surface emission and emit less LW radiation. When $\tau$ is large, ice cloud LW radiative effect for the all group also follows the trend of ice clouds with rain below.

In some literature, ice cloud radiative effect is defined by the flux difference between ice-only sky and clear sky (no cloud and no precipitation) rather than the definition we used in this study (i.e., flux difference between full sky and no-ice sky). For readers being able to make comparisons to earlier studies, in the following we show the differences in ice cloud radiative effect by the two definitions. We call the radiative effect defined by the current study IRE1 for full-sky minus no-ice-sky fluxes and IRE2 for ice-only sky minus clear sky fluxes. For IRE2 calculations, all atmospheric variables are kept the same as those in IRE1 calculations, but liquid water content and rainwater content are set to be zero. The global average of ice cloud radiative effects at TOA and the surface for all ice clouds combined and separately for each of the four ice cloud groups are summarized in Table B1. These results are based on DARDAR and 2C-ICE data during 2008.

Ice clouds that occur alone have a globally averaged SW radiative effect of about $-13.1$ W m$^{-2}$ and LW radiative effect of $15.3$ W m$^{-2}$ at TOA. Since these ice clouds occur alone, their IRE1 and IRE2 are the same. These clouds contribute the strongest to the total radiative

**Table B1.** A summary of ice cloud radiative effects (W m$^{-2}$) by two definitions: ice cloud radiative effect defined as flux difference between full sky and no-ice sky (IRE1) and ice cloud radiative effect defined as flux difference between ice-only sky and clear sky (IRE2).

<table>
<thead>
<tr>
<th>Cloud scenes</th>
<th>Occurrence (%)</th>
<th>SW$_{TOA}$</th>
<th>SW$_{surface}$</th>
<th>LW$_{TOA}$</th>
<th>LW$_{surface}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>DARDAR</td>
<td>2C-ICE</td>
<td>IRE1</td>
<td>IRE2</td>
</tr>
<tr>
<td>Ice clouds</td>
<td>34.7</td>
<td>31.7</td>
<td></td>
<td>-13.1</td>
<td>-13.1</td>
</tr>
<tr>
<td>Ice with water clouds</td>
<td>12.4</td>
<td>11.6</td>
<td></td>
<td>-0.4</td>
<td>-2.6</td>
</tr>
<tr>
<td>Ice with rain</td>
<td>5.1</td>
<td>4.8</td>
<td></td>
<td>-3.7</td>
<td>-9.2</td>
</tr>
<tr>
<td>Ice with water clouds and rain</td>
<td>1.5</td>
<td>1.2</td>
<td></td>
<td>-0.01</td>
<td>-0.5</td>
</tr>
<tr>
<td>All ice clouds</td>
<td>53.7</td>
<td>49.3</td>
<td></td>
<td>-17.2</td>
<td>-25.4</td>
</tr>
</tbody>
</table>
effects because of their large occurrence frequencies (i.e., 34.7% for DARDAR and 31.7% for 2C-ICE). Ice clouds associated with water cloud below account for about 10% in occurrence frequency globally. These ice clouds generate small IRE1 radiative effect (−0.4 W m$^{-2}$ for SW and 1.7 W m$^{-2}$ for LW at TOA). For IRE2, the globally averaged SW radiative effect is about −2.6 W m$^{-2}$ and LW effect is about 3.2 W m$^{-2}$, indicating that, because of different definitions, liquid water clouds may cause about 2 and 1.5 W m$^{-2}$ differences in globally averaged SW and LW radiative effects of ice clouds. Similarly, the group of ice cloud with rain below produces an IRE1 SW radiative effect of about −3.7 W m$^{-2}$ and LW effect of about 4.5 W m$^{-2}$. These ice clouds cover about 5% global area. Although its coverage is only about half of that of ice clouds with water clouds below, ice clouds with rain below show stronger SW and LW radiative effects mostly due to their large conditional radiative effects. For IRE2, the SW radiative effect is even larger (−9.8 W m$^{-2}$), and LW effect is 5.5 W m$^{-2}$; the differences of IRE1 and IRE2 are about 6 W m$^{-2}$ for SW and 1 W m$^{-2}$ for LW radiative effects. Finally, for ice clouds with water clouds and rain, their impacts on radiation are quite small as a result of their low occurrence frequencies (−1%).

If ice cloud radiative effect is defined as flux difference between full and no-ice skies, we obtain a net warming effect of 5 W m$^{-2}$ at TOA, indicating that ice clouds are warming Earth. If an alternative definition is adopted (i.e., flux difference between ice-only and clear skies), ice clouds on a global average could either warm or cool Earth (net effect is −0.7 W m$^{-2}$) considering the data uncertainties.

REFERENCES


Unauthenticated | Downloaded 08/14/22 05:07 AM UTC


Wang, Z., D. Vane, G. Stephens, and D. Reinke, 2012: Level 2 combined radar and lidar cloud scenario classification product


