Relationships between Ice Water Content and Volume Extinction Coefficient from In Situ Observations for Temperatures from 0° to −86°C: Implications for Spaceborne Lidar Retrievals*

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ABSTRACT

An examination of 2 yr of *Cloud–Aerosol Lidar Infrared Pathfinder Satellite Observations* (CALIPSO) lidar observations and *CloudSat* cloud radar observations shows that ice clouds at temperatures below about −45°C frequently fall below the *CloudSat* radar's detection threshold yet are readily detectable by the lidar. The CALIPSO ice water content (IWC) detection threshold is about 0.1 versus 5 mg m^{−3} for *CloudSat*. This comparison emphasizes the need for developing a lidar-only IWC retrieval method that is reliable for high-altitude ice clouds at these temperatures in this climatically important zone of the upper troposphere. Microphysical measurements from 10 aircraft field programs, spanning latitudes from the Arctic to the tropics and temperatures from −86° to 0°C, are used to develop relationships between the IWC and volume extinction coefficient σ in visible wavelengths. Relationships used to derive a radiatively important ice cloud property, the ice effective diameter D_e , from σ are also developed. Particle size distributions (PSDs) and direct IWC measurements, together with evaluations of the ice particle shapes and comparisons with semidirect extinction measurements, are used in this analysis. Temperature-dependent $D_e(\sigma)$ and IWC– σ relationships developed empirically facilitate the retrieval of IWC from lidar-derived σ and D_e values and for comparison with other IWC observations. This suite of empirically derived relationships can be expressed analytically. These relationships can be used to derive IWC and D_e from σ and are developed for use in climate models to derive σ from prognosed values of IWC and specified PSD properties.

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1. Introduction

Ice and mixed-phase clouds cover 25% of the earth's surface, and because the difference between their solar and infrared radiative effects—the net radiative forcing—can be large and varies seasonally and latitudinally, they have a large and variable effect on the earth's radiation budget (Hartmann et al. 1992). The ice water content (IWC) directly influences net radiative forcing and is

FIG. 1. Summary of CALIOP-CloudSat joint cloud detection as a function of (left) temperature and (right) height vs latitude for temperatures $\leq 0^{\circ}\text{C}$ with percentage of clouds detected by (a),(b) CALIOP alone, (c),(d) both *CALIPSO* and *CloudSat*, and (e),(f) *CloudSat* alone. This figure is derived for 2 yr of data (2007–08) when CALIOP was directed at orientations of 0.3° and 3° off nadir, respectively—a factor that should not affect these results. The diamonds denote the levels that each instrument segment contributes to 50% of the ice cloud observation at certain temperatures.

a primary variable diagnosed or prognosed by climate models, although the diversity in model predictions is large (Waliser et al. 2009). The focus of this study is on improving the IWC estimates for the cold, often high clouds that are so important for determining the earth's radiation budget.

Satelliteborne active instruments now probing clouds from space furnish researchers with unprecedented atmospheric data previously unavailable with passive satellite instruments. Active remote sensing of clouds provides information about the vertical structure of cloud optical and microphysical properties necessary to model accurately the radiative transfer in cloudy regions and to assess the role of clouds in climate. The National Aeronautics and Space Administration (NASA) A-Train constellation of satellites flying in formation includes two active cloud-profiling instruments. The Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP) on the *Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations* (CALIPSO) satellite is a nadir-viewing, polarization-sensitive, elastic backscatter lidar, operating

at wavelengths of 0.532 and $1.064\ \mu\text{m}$, that measures laser light scattered from clouds and aerosols. The Cloud Profiling Radar (CPR), on the *CloudSat* satellite, is a 94-GHz, nadir-viewing radar measuring power backscattered by hydrometeors, clouds, and precipitation. Because of their different operating wavelengths and detection thresholds, CALIOP and CPR have different sensitivities to cloud; specifically, CALIOP, with its lower cloud detection threshold and the ability to identify phase and qualitative features of crystal shape, can reveal what *CloudSat* is unable to provide. Together, CALIOP and CPR provide data by which cloud occurrence and microphysical information can be characterized vertically, globally, and seasonally.

To identify the CALIOP and CPR regions of sensitivity to ice clouds, we have characterized the average fraction in percent of all clouds detected at the same position (height, latitude, and longitude) by CALIOP and CPR alone and together, from 2 yr (2007–08) of collocated daytime and nighttime data at temperatures $< 0^{\circ}\text{C}$, shown as a function of temperature and altitude

in Fig. 1a. Temperatures from the European Centre for Medium-Range Weather Forecasts ECMWF-AUX product, version 008, and the optimally merged profile-by-profile CPR and CALIOP dataset (GEOPROF product 2B_GEOPROF_LIDAR, version 004) are used in this analysis. Because CALIOP is more sensitive to small particles and is attenuated in optically dense clouds, CALIOP-only observations of cloud are the majority of points ($>50\%$ of the time) where temperatures are from $<-45^\circ$ to -50°C and heights are greater than from 6 km poleward to 12 km equatorward (see also Martins et al. 2011). These regions occupy the upper 4–5 km of ice clouds (Fig. 1b). CALIOP and CPR both can usually sense clouds where the temperatures are between -20° and -45°C (Fig. 1c). Convective lofting of optically thick cloud particle populations through deep levels limits overlapping data to temperatures from -30° and -50°C toward the equator. Because the lidar is attenuated by dense clouds, the CPR alone detects cloud with high probability at temperatures warmer than from -15°C (poleward) to -20°C (equatorward) (Fig. 1e).

CALIOP alone detects the majority of the lowest-altitude clouds near the poles and the highest clouds near the equator (Fig. 1b). CPR alone detects cloud in a 5-km-deep band above altitudes near the surface at $\pm 30^\circ$ to 5 km at the equator. The overlap region between CALIOP and CPR is less than 5 km deep (Fig. 1d).

A recent intercomparison of CALIOP and CPR retrieval techniques highlighted the need to refine the microphysical assumptions in the retrieval algorithms (Stein et al. 2011). In particular, it is important to focus attention on the retrieval of IWC because the IWC directly influences the net radiative forcing and it is a primary variable diagnosed or prognosed by climate models (Waliser et al. 2009). Here, we draw upon the analysis of Heymsfield et al. (2013, hereinafter H13) to characterize the microphysical properties of ice clouds. They used measurements from 10 field programs that spanned the latitudinal range from the Arctic to the tropics and included occurrences where clouds were formed by deep convection or by gentler updrafts at temperatures ranging from 0° to -86°C . As expected, the particle size distributions (PSDs) broadened with temperature (more larger particles) and had higher measured IWCs in the convectively generated rather than synoptically formed ice clouds. These results were quantified by fitting gamma-type functions to the ice concentration versus diameter measurements, averaged over 5 s, or about 1-km horizontal paths. Particle cross-sectional areas became more “filled in” by ice with increasing temperature, which, taken together with the broadening of the PSDs, suggests that aggregation and riming (especially for the convective occurrences) became more prevalent as temperatures increased.

The objective of this paper is to improve the CALIOP, version 3, IWC retrieval parameterization derived from the volume extinction coefficient σ , drawing upon the H13 dataset and emphasizing ice clouds with temperatures colder than -50°C , where CALIOP alone detects ice cloud, and where few in situ measurements have been reported in the past. A secondary objective is to extend this new parameterization to temperatures where CALIOP and CPR overlap, providing a consistent IWC dataset regardless of whether CPR data are also available. Parameterizations developed in earlier studies for deriving the IWC from σ are discussed in section 2. Section 3 describes the dataset, and section 4 presents it, focusing on the low temperatures. Section 5 develops parameterizations of the composite dataset and evaluates them in different ways. Conclusions are drawn in section 6.

2. Ice water content derived from extinction

In this section, we discuss relationships between IWC and extinction developed in earlier studies. Heymsfield et al. (2005, hereinafter HWZ) developed the relationship

$$\text{IWC} = 119\sigma^{1.22}, \quad (1)$$

where IWC is in units of grams per meter cubed and σ is in units of inverse meters. The parameterization was developed using the fundamental relationship for visible wavelengths:

$$\sigma = 2A, \quad (2)$$

where A is the total particle cross-sectional area of the PSD. Details of the clouds sampled are given in HWZ. Values of σ and IWC were derived indirectly from PSDs using data from the 2D imaging probes (>50 – $100\ \mu\text{m}$) and small ($<50\ \mu\text{m}$) electronic particle probes, a combination of a video ice particle sampler (discussed below) and a 2DC (cloud) probe, or a balloonborne instrument. Additional data were obtained from a balloonborne instrument in midlatitude, synoptically generated cirrus (First International Satellite Cloud Climatology Project Regional Experiment: FIRE-2) near Coffeyville, Kansas, for T from -63° to -35°C .

Boudala et al. (2002) used PSDs from the 2D and small-particle probes to derive IWC and σ and used direct measurements of the IWC for evaluating the ice crystal densities they assumed. They expressed these results in terms of the effective particle diameter D_{ge} and developed temperature-dependent relationships for D_{ge} and IWC; together, these can be used to derive IWC from σ and T . They developed separate relationships with and without the inclusion of data from the small-particle probes.

0—ARM: ARM 2000, Atmospheric Radiation Measurement (ARM) Field Campaign, 2000

1: CF: CRYSTAL—FACE, The Cirrus Regional Study of Tropical Anvils and Cirrus Layers – Florida Area Cirrus Experiment, 2002

2: TC4: The Tropical Composition, Cloud and Climate Coupling (TC4) Field Campaign, 2007

3: NAM: NAMMA – The NASA African Monsoon Multidisciplinary Analyses Campaign, 2006

4: AIRS: AIRS_2, Alliance Icing Research Study II, 2003–2004

5: C3VP: Canadian CloudSat/CALIPSO Validation Program, 2006–2007

6: Rep: Replicator Observations, First ISCCP Research Experiment (FIRE)—2, 1991

7: SV, Subvisual: Experiments with CF and pre-AURA Validation Experiment, 2002 and 2004

8: SCT, SCOUT: Stratospheric Climate Links w/Emphasis on the Upper Troposphere/Lower Stratosphere, 2003

9: MPACE: Mixed-Phase Arctic Cloud Experiment, 2004

FIG. 2. Summary of temperatures (blue lines) and altitudes (yellow lines) sampled during the 10 field campaigns listed at the top and bottom. The thick horizontal lines divide the temperatures (-60°C) and the approximate altitudes (10 km) separating the colder- and warmer-temperature field programs.

Delanoë and Hogan (2008) developed a scheme based on optimal estimation theory that uses the combination of ground-based or spaceborne radar reflectivity, lidar backscatter, and infrared radiometer radiances for retrieving profiles of σ and IWC, from which the IWC is derived. The technique is developed using PSD data from a combination of 2D and small-particle probes for temperatures between about 0° and -50°C .

On the basis of measurements in wind tunnels and through the use of improved instrumentation, it has become increasingly accepted that ice particles shatter on the inlets of the instruments designed to measure them (see Korolev et al. 2011). Such shattering limits accurate PSD measurements. Not only have data from the Forward Scattering Spectrometer Probe (FSSP) and probes measuring in similar size ranges been compromised [as first identified in the very early study of Gardiner and Hallett (1985)], but those from the Particle Measuring Systems (PMS) 2DC and others as well (Field et al. 2006). The PSDs contaminated by shattering inadvertently factored into the analyses summarized above. Unfortunately, shattering factors into the derivation of σ and IWC nonuniformly, resulting in

more influence on σ than IWC (Boudala et al. 2002) and compromising these IWC estimation techniques.

3. Data sources

The present study emphasizes our observations at temperatures below where the HWZ IWC parameterization was developed. In this study, we carefully treat the contributions of small particles to the total IWC and σ , and evaluate these treatments (see the appendix). Furthermore, HWZ calculated the IWCs from the PSDs and were not constrained by measurements, whereas our study includes directly measured IWCs.

The data used in our study span the temperature range from -86° to 0°C (Fig. 2; see H13 for complete details). Data for temperatures between -86° and -55°C are from low-latitude clouds and the tops of midlatitude ones, providing correlative in situ observations where CALIOP detects cold, high cloud tops. In these situations, small, sub- $200\text{-}\mu\text{m}$ particles dominate the extinction and the IWC. Data for temperatures warmer than -55°C are from a combination of

in situ-generated cirrus in midlatitudes and convectively generated cirrus at low latitudes.

Particle size D in what follows is the approximate maximum diameter of a particle as found from the minimum diameter of a circle that fully encloses the projected 2D image of a particle, unless otherwise noted. For the 2D-size particles, we attempt to correct for shattering by using particle interarrival times to sort out single from shattered particles, as described in Field et al. (2006). Where necessary, we do not use data from the small-particle probes, an issue that is considered in the appendix.

a. Temperatures of -60°C and below

In situ-generated cirrus clouds were targeted in the FIRE-2 field campaign based out of Coffeyville, Kansas, in November and December 1991. For 3 days, balloonborne ice crystal Replicators provided vertical profiles of PSDs and detailed ice crystal habit information for ice crystals larger than $10\text{ }\mu\text{m}$ with $1.4\text{-}\mu\text{m}$ resolution (Miloshevich and Heymsfield 1997), during ascents through cirrus clouds at a rate of about 3 m s^{-1} (Fig. 2, see label Rep). Shattering on the probe's inlet was largely nonexistent; desirable, quasi-vertical profiles, rather than horizontal legs or spiral descents, were obtained through the cloud layers. Average properties are derived by dividing data from each cloud layer into 15 equal height layers; these layers range from 170 to 270 m deep, depending on the case. These data were used in the HWZ analysis, but were peripheral because temperatures warmer than -50°C were emphasized in that study.

Data were collected at low temperatures during three flights by the NASA WB-57 during the Cirrus Regional Study of Tropical Anvils and Cirrus Layers–Florida–Area Cirrus Experiment (CRYSTAL-FACE, hereinafter CF) and during a flight from Houston, Texas, to San Jose, Costa Rica, in the PreAura Validation Experiment (PA). The cold temperature CF observations, reanalyzed, and those from PA, referred to as subvisual, are based on the analysis reported in Schmitt and Heymsfield (2009). The data were limited to high-altitude nonconvective thin clouds with temperatures from about -60° to -86°C (Fig. 2), although one case was in strong gravity waves generated downstream of convection. PSDs obtained from the NASA WB-57 aircraft for both CF and PA were measured with the National Center for Atmospheric Research (NCAR) video ice particle sampler (VIPS) probe, which provides information about ice particles from 10 to $200\text{ }\mu\text{m}$ in a difficult-to-sample size range; the breakup of ice particles is readily observed and filtered out. Above this size range, 2D imaging probe data were used.

Derived properties for each particle observed by the VIPS and 2D probe include the projected area and maximum dimension, which are binned to produce PSDs.

Results are averaged over 5 s, corresponding to $\sim 1\text{-km}$ horizontal sampling resolution. Ice water content was measured by both the University of Colorado closed-path tunable diode laser hygrometer (CLH) and the Harvard University Lyman-alpha water photofragment-fluorescence hygrometer (HT) for CF and by HT for PA. The detection threshold of these instruments is about 10^{-4} g m^{-3} , and the precision of both instruments is about 10^{-4} g m^{-3} , although there are situations when this detection limit or precision may degrade somewhat (S. Davis 2012, personal communication).

We reanalyzed data reported in the de Reus et al. (2009) study of cold cloud measurements from the Stratospheric–Climate Links with Emphasis on the Upper Troposphere and Lower Stratosphere (SCOUT) field program based out of Darwin, Australia, in November–December 2005. The SCOUT samples used in this analysis were predominantly in the temperature range from -75° to -85°C . The measurements were acquired from flights on the following days: (a) 25 November, (b) 28 November, (c) 29 November, (d) 30 November–A, and (e) 30 November–B (see Vaughan et al. 2008). These data included four flights into the tops of “Hector storms” (intense deep thunderstorms) and one survey flight, when both particle spectrometers (see below) were operating [flights identified above as a–e]. Coincident lidar data were obtained (see the supplemental material available online).

PSDs were measured by a PMS FSSP-100 using Droplet Measurement Technologies (DMT) high speed electronics (SPP-100), with concentrations derived for seven size bins centered from 4 to $28\text{ }\mu\text{m}$, and by a DMT 2D-type cloud imaging probe (CIP), with concentrations derived in 19 size bins from $50\text{ }\mu\text{m}$ to $>1\text{ mm}$. Shattering contaminating the small particle observations is reduced by considering only periods where the largest measured particles were $\leq 200\text{ }\mu\text{m}$, but there were few instances where larger particles were sampled. The averaging methods and additional details of the data processing are presented in H13.

Total water content and water vapor content were measured for flights c–e, with two probes, the fast in situ hygrometer (FISH) and the fluorescent airborne stratospheric hygrometer (FLASH), respectively (see de Reus et al. 2009). The difference between the two measurements yielded the IWC. The detection threshold is about 10^{-4} g m^{-3} and the total uncertainty in the derivation of the IWC from the two hygrometer instruments is estimated to be 20%.

b. Temperatures warmer than -60°C

In this section we discuss the type of measurements acquired during the following set of seven field programs (see Fig. 2): the Atmospheric Radiation Measurement 2000 intensive observing period (ARM 2000 IOP,

3 cold cloud cases in synoptically generated ice clouds in Oklahoma); CF (largely convectively generated ice clouds sampled by the University of North Dakota Citation, 10 cases); the NASA African Monsoon Multidisciplinary Analyses (NAMMA, 12 mostly convectively generated clouds, many of which formed in or near dust layers); Tropical Clouds, Convection, Chemistry and Climate (TC4, 11 cases in Costa Rica, mostly convectively generated; see Toon et al. 2010); synoptically generated ice cloud layers with embedded convection and supercooled liquid water regions (filtered out of the dataset) from the Alliance Icing Research Study II (AIRS-2, 7 cases in the Toronto, Ontario, Canada, area; see Isaac et al. 2005); and the *CloudSat/CALIPSO* Validation Project (C3VP, 3 cases in Montreal, Quebec, Canada). Air temperatures during all field programs were measured with Rosemount temperature probes.

Particle probe details are given in H13. The PSDs were acquired using probes measuring particles sized from about $50\text{ }\mu\text{m}$ to $>1\text{ cm}$. In the appendix and as summarized in section 5, we use PSD data and semidirect measurements of σ to conclude that the extinction estimates for many of the periods of data at temperatures $> -60^\circ\text{C}$ are highly accurate without the inclusion of particles from the small-particle probes, at least above the detection level of about 0.01 m^{-1} for the semidirect method and for those cases where we had this method available. The inclusion of small particles where they occur in pockets can add up to 50% to σ , but far less than the factor-of-up-to-2–3-greater values of σ with the inclusion of the small particle data. For that reason, σ values from only the 2D-type probes are used where temperatures are above about -60°C . Total condensed water contents (TWCs)—ice plus liquid when present—were measured with a counterflow virtual impactor (CVI) or a cloud spectrometer and impactor for $\text{TWC} > 0.01\text{ g m}^{-3}$. Liquid water encounters were infrequent and have been filtered out of the dataset (see H13 for more details).

c. Ice water content and volume extinction coefficient calculated from the PSDs

In this section, we summarize (see also H13) how IWC and σ are derived from the particle probe measurements and the direct measurements of the IWC, how each parameter can be expressed analytically, and how they relate to each other. The IWC is measured and also calculated from the PSD, estimating masses m with a relationship given by a power law:

$$m(\text{g}) = aD^b, \quad (3)$$

where a is the coefficient (in cgs units), b is the exponent, and D is in centimeters. The coefficients chosen for the

warm cloud cases are based on a statistical comparison with the CVI measurements, yielding $a \sim 0.006$ (cgs) and $b = 2.1$ (H13), for $D > 69\text{ }\mu\text{m}$. For the CF and PA (VIPS) cases and for SCOUT, the exponential relationship developed by Schmitt and Heymsfield (2009) is cast in the form given by (3). That relationship is fitted to a power law yielding $a = 0.0175$ and $b = 2.51$, for $D > 12\text{ }\mu\text{m}$. The mass–diameter relationships are applied above the limits shown; below that size mass is taken to be that of a solid ice sphere. In H13, we use gamma functions fitted to the PSDs and the $m(D)$ relationship to yield analytic estimates of the IWC:

$$\text{IWC (g m}^{-3}\text{)} = aN_0 10^6 \Gamma(b + \mu + 1) / \lambda^{(b + \mu + 1)}. \quad (4)$$

In H13 these are shown to provide accurate representations of the measured IWCs. In (4), N_0 , λ , and μ are, respectively, the intercept, slope, and dispersion fitted by the gamma functions to each of the 2D PSDs.

The σ are derived from

$$\sigma (\text{m}^{-1}) = 2 \sum N(D) A(D) = 2(\pi/4) \sum N(D) A_r(D) D^2, \quad (5)$$

where $N(D)$ is the concentration in each size bin. In (5), the area ratio A_r , the ratio of the particle area divided by the area of a sphere that completely encloses the projected area of a particle, $A_r = A/[(\pi/4)D^2]$, provides intrinsic information on particle shape. For sizes $> 50\text{ }\mu\text{m}$, the projected cross-sectional area of each particle A —whether from the VIPS or 2D imaging probes—is summed for all particles across the PSD to obtain the total cross-sectional area for the larger particles. For the cold temperature observations, except those from the Replicator, we add in the sub- $50\text{-}\mu\text{m}$ sizes by using the PSDs and relationship developed by Schmitt and Heymsfield (2009) from the VIPS probe from CF and PA for the sub- $50\text{-}\mu\text{m}$ area as a function of diameter. We use the actual images of particles from the Replicator for all sizes to find the total area.

The $A_r(D)$ data from each particle imaged during each 5-s PSD measured by the 2D probes were fitted to a power-law relationship (H13), $C_0 D^{C_1}$. For the cold temperature observations when particles below $200\text{ }\mu\text{m}$ dominated, we use a single area-ratio power law developed by Schmitt and Heymsfield (2009), fitted for $1 < D < 200\text{ }\mu\text{m}$ (with D in centimeters): $A_r = 0.25D^{0.21}$ for $D > 14\text{ }\mu\text{m}$ and $A_r = 1$ below this size. The original relationship was developed from high-resolution particle images, unavailable for most of the cold cloud sampling.

If we assume a gamma form to the PSDs and power-law-type area-ratio relationship form above, (5) can be integrated from 0 to ∞ to yield

FIG. 3. Extinction values derived from the 2D imaging probe PSDs and particle cross-sectional areas and from (6) using gamma functions fitted to the 2D PSDs and fits to the 2D probe cross-sectional areas (expressed in terms of the mean area ratio vs size using a power-law relationship). The 1:1 relationship is shown with a solid line, and data points show the median values in increments of even number of data points. The mean ratio of the derived to parameterized extinctions is indicated in each panel.

$$\sigma \text{ (m}^{-1}\text{)} = (\pi/2)C_0N_010^2\Gamma(C_1 + \mu + 3)/\lambda^{(C_1 + \mu + 3)}. \quad (6)$$

For the cold clouds, using either the Schmitt and Heymsfield (2009) $A_r(D)$ relationships with the measured PSDs or the single power law fit to the relationship or the gamma fits to the PSDs makes little difference in the derived values of σ . The median and mean-standard deviations of the ratios of σ derived from (6) to σ from

the measured quantities (5) for the cold cloud cases are as follows: 0.85, 0.75 $\pm$ 0.18 (Replicator); 0.98, 0.98 $\pm$ 0.03 (CF-PA); and 0.89, 0.90 $\pm$ 0.06 (SCOUT). For the warm cloud cases (Fig. 3), the median ratio of σ derived from (6) to σ from the measured quantities (5) is nearly unity. The analytic forms for IWC and σ [(4) and (6)] therefore provide useful information for all cloud cases. [In Fig. 3, data from the Mixed-Phase Polar Arctic

Cloud Experiment (MPACE) are omitted to reduce the number of panels in the figure; these ratios are 1.01 and 1.23 ± 1.21 .]

Note in Fig. 3 that there is an apparent overestimate in σ from (6) relative to the summed values using the actual $N(D)$ and $A_r(D)$ data at low σ 's for the warmer temperature observations; this is due to the integration of (6) beginning at a diameter of $0 \mu\text{m}$ rather than at $50 \mu\text{m}$, where the values of $A_r(D)$ are above 1. Although possibly a bias, this effect may actually provide useful information on the potential underestimate of σ by not including particles $< 50 \mu\text{m}$. The additional contribution to σ by the integration (6) was examined numerically and averages obtained for 10°C intervals of temperature between -60° and -20°C ; these are 0.09 ± 0.13 , 0.044 ± 0.07 , 0.03 ± 0.043 , and 0.014 ± 0.03 . Clearly, the contribution diminishes with increasing temperature. If the PSDs are continuous functions, as implied from the observations for the cold temperatures and from 2DS probe data in small particle sizes (see the appendix), these results suggest that the fitted PSD functions may provide a reasonable first estimate of the effect of not including sub- $50\text{-}\mu\text{m}$ particles.

From (4) and (6), the ratio IWC/σ becomes

$$\begin{aligned} \text{IWC}/\sigma \text{ (g m}^{-2}\text{)} \\ = [2a10^4/(\pi C_0)] [\Gamma(b + \mu + 1)/\Gamma(C_1 + \mu + 3)] \lambda^{(C_1 - b + 2)}. \end{aligned} \quad (7)$$

The relationship between IWC and σ is independent of N_0 .

The effective diameter D_e provides additional information for characterizing ice cloud radiative properties (Foot 1988):

$$D_e = (3/\rho) \text{IWC}/\sigma = 3.29 \text{IWC}/\sigma, \quad (8)$$

where ρ is the density of solid ice. In the supplemental material available online, we break down (7) into the product of three terms: P1, which is related to the a coefficient in the $m(D)$ relationship and the term C_0 in the area ratio term, both related to particle habit; P2, which is a function of the ratio of the two gamma terms, a function of the PSD dispersion and power in the mass dimensional and area ratio terms; and a third term, which is a function of the PSD slope. P1 is relatively constant for $T < -60^\circ\text{C}$ and decreases slightly with increasing temperatures above -60°C . P2 is largely independent of temperature. P3 increases more markedly with temperature. At a given temperature, P1, P2, and P3 have relatively little scatter. The net result is that the IWC/σ ratio and D_e increase with temperature but with relatively small variability at a given temperature.

Inserting appropriate, average coefficients $b = 2.51$ and $C_1 = -0.21$ from the mass and area ratio power-law exponents developed from the cold CF and pre-Atmospheric Variability Experiment (AVE) datasets into (7) and (8) yields the result that IWC/σ and D_e should be largely independent of λ for the colder-temperature observations. For the warmer-temperature observations with $b \sim 2.1$ and C_1 , which is temperature dependent (H13), the exponent in λ has a larger effect. Because λ is a strong function of temperature (H13), IWC/σ and D_e for the warmer temperature observations are largely dependent on the temperature. These temperature hypotheses are further examined in the next section.

4. Results

Here, our observations are partitioned into data from the colder and “warmer cases. Those clouds sampled within deep convection or in the immediate vicinity (excluding liquid water regions) during CF-warm, NAMMA, and TC4 (although with weaker updraft velocities) have larger maximum particle sizes than the clouds referred to as primarily stratiform sampled during ARM, AIRS, C3VP, FIRE-2, CF-cold, PA, and SCOUT (H13). The SCOUT observations were located outside of convection in anvil regions, or in situ generated cirrus. Because IWC and σ to a first approximation scale together through the parameter N_0 , convective and stratiform regions are grouped together in the discussion that follows.

The low-temperature observations are most relevant to CALIOP and are therefore emphasized. Note that with the exception of the few Replicator points (the cold CF data are reanalyzed using VIPS data), the low-temperature observations here are not included in the HWZ relationship, and the production of artifacts by ice shattering is largely not an issue because there are few large particles to shatter at these cold temperatures.

a. Summary of the IWC– σ relationship from the low-temperature observations

Relationships of the form $\text{IWC} = \alpha\sigma^\beta$ were fitted to the low-temperature observations (Fig. 4). When partitioned in 10°C temperature intervals (colored lines), values of β are mostly below 1.0 for the lowest temperature ranges (Fig. 4a). The small particles contribute more to σ than to the IWC. However, the low-temperature observations combined yield an exponent $\beta > 1$, indicating that, overall, D_e increases with temperature (Fig. 4b). The HWZ relationship also has a value of $\beta > 1$, although with a shallower slope but much larger pre-coefficient (Fig. 4b).

FIG. 4. Synthesis of the CF, pre-AVE, and Replicator IWC and σ observations (a) in intervals of temperature and (b) for all data combined. The relationship developed by HWZ is plotted as a dashed line in each panel. Black dotted lines indicate values of effective diameter, as labeled.

b. Summary of the IWC- σ relationship from the higher-temperature observations

We now examine the data for the warmer ($0^\circ < T < -60^\circ\text{C}$) cases: ARM, CF (warm), TC4, NAMMA, AIRS, C3VP, and MPACE. For this comparison, we use the IWCs derived from the PSDs in order to reduce variability resulting from IWC hysteresis and other measurement errors and to provide a means of extending the IWCs to higher and lower values than those measured (of about 1.0 and 0.01 g m^{-3} , respectively) at these temperatures. As with the cold temperature observations, the $\text{IWC} = \alpha\sigma^\beta$ relationship yields exponents of about 1 when the data are partitioned into 10°C temperature intervals (Fig. 5a). This result indicates that D_e is approximately constant within a 10°C temperature interval. The exponent is >1 for all data from the warm temperature clouds combined (Fig. 5b), indicating that on average D_e increases with temperature. Relative to the low-temperature observations, there is much less variability in the range of IWC- σ (and D_e) values with temperature and hence less difference in the exponent between the 10°C temperature partitioned and combined data.

c. Synthesis of low- and high-temperature observations and sensitivity studies

Temperature-dependent IWC- σ relationships are developed in this section. The IWCs are derived using three methods:

- 1) from the direct IWC measurements (IWC_{meas});
- 2) using the fitted PSDs (IWC_{PSD}), with the (a , b) coefficients as given above;
- 3) a combined method (IWC_{comb}) that multiplies the a coefficient by a factor based on the ratio of $\text{IWC}_{\text{meas}}/\text{IWC}_{\text{PSD}}$ that accounts for a variation in $a(T)$ (H13). The b coefficient is left unchanged since it varies little with temperature (H13). This method provides a means of determining the IWCs when IWC_{meas} falls below the probe's detection level ($\sim 0.0001\text{ g m}^{-3}$ for $T < -60^\circ\text{C}$ and 0.01 g m^{-3} for $T > -60^\circ\text{C}$) and also partially compensates for measurement errors (see H13).

Figures 6 and 7 show the relationship between IWC and σ in 10°C temperature intervals for the first two methods, with several effective diameters plotted for reference. The cold ice cloud data for the temperature range from -50° to -70°C are combined into one interval to reduce the number of temperature ranges. Use of the direct IWC measurements produces relatively large scatter, and the PSD approach produces relatively little scatter. The fitted power-law relationships between IWC and σ listed at the top of each panel show a value of β mostly close to but below 1.0, indicating that the

effective diameter does not change substantially in each interval. From (7), this suggests that the PSD slope does not change appreciably in any temperature interval.

Figure 8 shows the relationship between IWC_{PSD} and IWC_{meas} , with the colder- and warmer-temperature observations identified. Curve fits that represent the systematic temperature deviations of the estimated IWCs from the direct measurements and estimates are shown. These fitted equations are used to derive modified coefficients a for the estimated IWCs and reduce the measurement errors caused by hysteresis and extend the lower detection limit for the direct measurements.

The combined method produces relatively little scatter compared to the direct measurements (Fig. 9), largely because of the hysteresis effect noted with the direct measurements. Given that the combined method likely produces the best estimate of the IWC- σ relationship that we can develop from our PSDs and area ratio data and also because of the much lower IWCs that it can generate, we use these coefficients to derive the temperature dependence of the α and β coefficients in later discussion unless otherwise indicated.

5. Discussion

We address the question of whether the in situ observations of extinction are comparable to what CALIOP retrieves, and where there are differences. We also examine the temperature dependence of the effective diameter. Various methods of deriving the IWC from σ are presented and evaluated.

a. Comparison of CALIOP and field program data

Here, we compare the temperature dependence of σ derived from global CALIOP retrievals to those from the PSDs. CALIOP acquires profiles of backscattered 532-nm laser light, from which cloud layers are detected and classified as water or ice (Winker et al. 2009); from these measurements the corresponding σ is retrieved (Young and Vaughan 2009).

CALIOP extinction retrievals are of two types. Constrained solutions represent cases where the lidar is able to directly measure the cloud-layer transmittance and this is used as a constraint on the extinction retrieval, providing the highest-quality CALIOP retrievals. Unconstrained solutions use an a priori extinction-to-backscatter ratio and allow retrievals when the cloud is optically thin and the transmittance cannot be measured accurately, or where the lidar beam cannot penetrate fully through the cloud. Figure 10a shows that σ increases by about a factor of 4 with temperature for constrained solutions of the lidar equation but by more than an order of magnitude for unconstrained solutions.

a: IWC vs σ by Temperature Intervals b: IWC versus σ , All Temperatures Combined

FIG. 5. As in Fig. 4, but for clouds designated as warm.

FIG. 6. The directly measured IWC vs σ derived from the particle probe data. The top of each panel shows the temperature interval and curve fits (orange solid lines) to the data. Colors identify the field program [see right legend in (a)]. The numbers on the left side of each panel show the number of 5-s PSDs associated with each field program. Lines of constant D_e are plotted with dotted lines, with the long dashed line corresponding to an effective diameter of $20 \mu\text{m}$ [scale shown in (b)].

Extinction from unconstrained retrievals is biased somewhat low due to the use of an extinction-to-backscatter ratio that has recently been found to be too small (Josset et al. 2012). Young et al. (2013) provide a detailed analysis of CALIOP extinction uncertainties and show that for

a fixed error in the cloud extinction-to-backscatter ratio, the magnitude of the subsequent errors in the retrieved extinction coefficients increases nonlinearly with cloud optical depth. Thus, one might expect the median extinction from unconstrained retrievals to be lower than from

FIG. 7. As in Fig. 6, except that the IWC is derived from the 2D probe PSDs.

constrained retrievals, especially in the warmer, lower portions of the clouds. However, at warmer temperatures (above -20°C) the unconstrained solutions become increasingly larger relative to the constrained solutions, likely because the unconstrained retrievals are able to retrieve into more of the lower, optically thick cloud.

For $-70^\circ < T < -50^\circ\text{C}$, the σ s for the CALIOP-constrained solutions and in situ particle probe datasets

are within a factor of 2 of each other, which is significant because this is where we expect CALIOP to provide the majority of the unique, global cloud-particle measurements, and hence our IWC- σ relationships are critical to evaluate in this temperature range. Below -70°C the in situ values are similar to the unconstrained solutions because the smaller optical depths typical of clouds at these temperatures do not allow as many constrained

FIG. 8. IWC derived from the PSDs in comparison with those measured with power-law curves (linear in the plot) fitted to the relationships.

extinction solutions. Increasingly large differences are found between both types of CALIOP extinction solutions and the in situ σ values as temperatures increase from -50° to -20°C because CALIOP is more frequently attenuated in clouds with higher optical depth associated with convection and warmer temperatures, and hence the σ values sampled by CALIOP are relatively low.

b. Effective diameters

The effective diameter D_e , proportional to the ratio IWC/σ [see (8)] is an important parameter for describing and quantifying the radiative properties of an ice cloud layer. Because of its importance, and to aid in the development of an IWC retrieval algorithm below, we have evaluated its temperature dependence. The D_e generally increases with temperature (Fig. 11a). The abrupt increase in D_e with temperatures from -60° to -50°C is partly circumstantial: 78% of the observations between -60° and -50°C are from the cases identified as “convective,” with correspondingly broader PSDs, higher IWCs, and more dense particles than the cloud

regions characterized as “stratiform” (from H13). A single power-law fit (red curve) provides good correspondence with the mean values as a function of temperature (large points and bars), with a correlation coefficient (r^2) value of 0.94 and with the largest differences noted for the lower temperatures. A three-segment exponential curve fits the data well for all temperatures, with r^2 values above 0.95.

Our D_e values are higher than those derived by Boudala et al. (2002) from 2D probe data (see Fig. 11a). Their IWC measurements, as derived from the Nevzorov probes, are likely to be biased low due to ice particles bouncing out of the probe’s sensing area (a problem that forced a redesign of the instrument sensing area); this could explain IWCs and hence D_e that are smaller than ours.

The inclusion of the small-particle probe data for the warmer ice clouds ($T > -60^{\circ}\text{C}$), as did HWZ, leads to about a factor-of-2 increase in σ relative to IWC and thus a reduction in D_e to about one-half the values shown in Fig. 11a (median and mean ratios of 0.51 and 0.46 ± 0.25 , respectively). This decrease in D_e is largely

due to the increase in σ (median and mean ratios of 1.59 and 3.81 ± 3.41 , respectively) relative to the increase in the IWC (median and mean ratios of 1.27 and 1.59 ± 0.34 , respectively). These increases in σ from the inclusion of small particles, however, are generally not realistic. In the appendix (see Fig. A4), we compare σ with and without the inclusion of data from the

small-particle probes with those directly measured by the diode laser hygrometer (DLH) instrument. This comparison shows that the small-particle probes generally overestimate σ relative to a semidirect coincident measurement by about a factor of 2. The absence of small-particle data for temperatures -60°C and above clearly omits some situations when small particles are

FIG. 10. Summary of CALIOP ice cloud extinction retrievals for the month of December 2009 and σ estimates derived from the in situ particle probe data. Median values of version 3.0, level 2, 5-km horizontally averaged nighttime CALIOP extinctions and the standard deviations are shown. CALIOP sampling statistics become increasingly poor at temperatures warmer than -20°C because of the inability of the lidar to penetrate to the base of optically dense ice clouds or to probe below water clouds. Only randomly oriented ice clouds are included here; clouds with oriented ice are filtered out. The data are for the (a) constrained, high-confidence retrievals, where the cloud transmittance can be measured directly based on the difference in magnitude of the molecular backscatter immediately above and immediately below the cloud layer, and (b) unconstrained solutions, which use a default value of the cloud extinction-to-backscatter ratio to retrieve extinction profiles when constrained solutions are not possible.

indeed present. For example, small, homogeneously nucleated particles were observed in a convective updraft during NAMMA at temperatures below -40°C (Heymsfield et al. 2009). This situation represents an

extreme case and one that the CALIOP lidar would not be able to penetrate. Other situations with small particles certainly do occur. However, overall, the analysis presented in the appendix suggests that our D_e values

FIG. 11. Effective diameters (a) from all sizes for temperatures below -60°C and 2D probe sizes above -60°C , with a curve fit shown and (b) in temperature intervals from the lowest temperature range (blue) to the warmest (orange), with the ratio of the D in that temperature interval divided by the mean value for that interval as a function of the σ divided by the midpoint mean for that temperature interval. Particle probe data are as in (a).

FIG. 12. Summary of the IWC- σ data from all field programs combined with dashed and dotted blue lines showing constant values of D_e . Individual colored points show data in the following temperature intervals: from -90° to -70°C and from -70° to -55°C for the cold-temperature clouds; and from -55° to -50°C , from -40° to -30°C , from -30° to -20°C , and from -10° to 0°C for the warmer clouds. Least squares fits to the IWC- σ data as derived from a single power law and a composite of two power laws are plotted and listed.

without the addition of small particles for $T > -60^\circ\text{C}$ are inclusive for the most part. In the supplemental material to this paper, we compare σ from the PSDs to values derived from coincident lidar data obtained during SCOUT (Fig. S2). For this dataset, the σ derived from the particle probes and derived from the lidar are within a factor of 2.

An additional point worth noting is that a given extinction value has much larger IWCs for the data for -60°C and above than for temperatures below -80°C . Because we have direct measurements of the IWC (see also H13) and have evaluated the σ values against the lidar data (Fig. S2), we believe this result is real. It can be accounted for by the PSD maximum diameter, which indicates that the largest particles are $10\text{--}80\text{ }\mu\text{m}$ below -80°C and that between -80° and -60°C the maximum diameter increases rapidly to above $200\text{ }\mu\text{m}$. This has the effect of rapidly increasing IWC relative to σ and thus $D_e(T)$.

Using the median values of σ (σ_{med}) derived in approximately 10°C temperature intervals from -85° to 0°C , Fig. 11b examines how D_e varies as σ increases or decreases from its midpoint value. There is a surprising constancy of the D_e values, suggesting that temperature is the central factor that controls D_e . Therefore, the exponential curves fitted to the temperature dependence in Fig. 11a can be used reliably to derive $D_e(T)$.

Figure 12 shows the composited IWC versus σ and D_e values (dashed and dotted sloping lines), with temperatures identified. This figure again shows that D_e increases with temperature; it is most notable at the lowest temperatures and then flattens with increasing temperature. Two sets of curves are fitted to the data: one power-law fit to all temperatures and two power-law fits over different ranges of σ . The slopes of the power-law fits are above 1 and differ markedly from those derived by temperature, a feature that is the result of combining all temperatures.

FIG. 13. Coefficients in fits to the temperature-dependent relationship $IWC = \alpha\sigma^\beta$ derived from the three methods used to obtain the IWC (Figs. 7–9): the direct measurements, the PSDs, and a blend of the two: (a) coefficient α and (b) exponent β . In (a) and (b), a best fit to the temperature dependence of each coefficient is shown.

c. Development of IWC retrieval algorithms

Drawing on the relationship $IWC = \alpha\sigma^\beta$ (in Figs. 7–9), Fig. 13 compares the temperature dependence of the α and β coefficients derived from the direct measurements, the PSDs, and the composite of the two. In all cases, the coefficient α tends to increase with temperature, a feature not considered in HWZ. The exponent β is relatively flat with temperature. Note also that at temperatures from -55° to 0°C , α increases much more slowly than for the lower temperatures. Curves fitted to

the coefficients derived from the composite method are shown and used in the discussion below.

Table 1 evaluates the different IWC estimation techniques using the HWZ relationship and the relationships from Figs. 11–13:

$$IWC \text{ (g m}^{-3}\text{)} = \alpha D^\beta; \quad (9)$$

$$\text{HWZ: } \alpha = 119, \quad \beta = 1.22; \quad (9a)$$

$$\text{Single power law: } \alpha = 527.0, \quad \beta = 1.32; \quad (9b)$$

TABLE 1. Evaluation of IWC retrievals from extinction and effective diameter: median and mean (with standard deviations) of the ratio of the IWC derived from each fit type to the IWC derived from the composite method, in different temperature intervals.

Temperature range (°C)	HWZ	New power-law fit	Two power-law fit	Temperature-dependent fit	From effective diameter
From -86 to 0	Median: 0.45 Mean: 0.43 ± 0.11	0.99 0.96 ± 0.32	0.94 0.91 ± 0.28	1.09 1.07 ± 0.26	1.08 1.08 ± 0.31
From -20 to -10	0.40 0.40 ± 0.12	0.94 0.92 ± 0.32	0.88 0.87 ± 0.29	1.06 1.05 ± 0.31	1.18 1.24 ± 0.46
From -40 to -20	0.42 0.40 ± 0.10	0.94 0.90 ± 0.31	1.08 1.09 ± 0.357	1.08 1.08 ± 0.23	1.07 1.08 ± 0.28
From -60 to -40	0.44 0.43 ± 0.12	0.96 0.93 ± 0.32	0.91 0.88 ± 0.29	0.97 0.99 ± 0.26	1.00 1.00 ± 0.28
From -86 to -60	1.83 1.73 ± 0.85	2.87 2.86 ± 1.53	2.92 2.87 ± 1.51	1.00 1.10 ± 0.68	0.98 0.95 ± 0.32

$$\text{Two power laws: } \alpha = 1867, \quad \beta = 1.48, \quad \sigma < 7.9 \times 10^{-4}; \quad \alpha = 186, \quad \beta = 1.15, \quad \sigma > 7.9 \times 10^{-4}; \quad (9c)$$

$$\text{Temperature-dependent fits: } a = 0.00532(T + 90)^{2.55}, \quad b = 1.31e^{(0.0047T)}; \quad (9d)$$

$$\begin{aligned} \text{From effective diameter: } D_e = \alpha e^{\beta T}, \quad \text{IWC} = \sigma(0.91/3)D_e, \quad \text{where } \alpha = 308.4, \quad \beta = 0.0152, \\ -56^\circ < T < 0^\circ\text{C}; \quad \alpha = 9.1744 \times 10^4, \quad \beta = 0.117, \quad -71^\circ < T < -56^\circ\text{C}; \quad \alpha = 83.3, \quad \beta = 0.0184, \\ -85^\circ < T < -71^\circ\text{C}. \end{aligned} \quad (9e)$$

The IWCs derived from the five methods have been evaluated against the composite dataset (Table 1). A median ratio of retrieved to measured IWCs of unity and a small standard deviation is ideal. The temperature-dependent method and especially the D_e approach excel for all temperature intervals; the HWZ power law underestimates the measured IWC by about a factor of 2 for most intervals; the new single power law is quite accurate except for the temperature range of most interest for the CALIOP observations; and the two power-law fit improves upon the single fit for the lowest temperatures and is reasonably accurate at the warmer ones except that the fit also overestimates the lower temperatures (not plotted). Therefore, the temperature-dependent and D_e approaches faithfully represent the observations; and the single power law, with caution for the lowest temperatures, provides a good fit to the observations.

6. Conclusions

In this study, temperature-dependent relationships are developed between the ice water content and extinction, and for the effective diameter as a function of temperature, for use in lidar retrievals of the IWC and D_e . The measurements draw on data spanning a wide range of temperatures, cloud formation mechanisms, and latitudes. The IWC is directly measured and also

derived from the PSDs using a number of different methods. The σ s are calculated from PSDs and from measured or derived particle cross-sectional areas. A comparison of σ values to those derived from a semi-direct method employing an in situ probe and derived from coincident lidar data suggests that our estimates of σ are reliable for those cases where these data are available.

The IWC/ σ ratio and the D_e , which is given by about 3.3 times this ratio, are found to increase with temperature. An examination of the IWC/ σ ratio and D_e is particularly useful because they do not depend specifically on the number concentration. We show that the increase in IWC/ σ and D_e with temperature results primarily from the PSD broadening relative to the contribution of the smaller particles to σ . These parameters increase more slowly at the warmer temperatures than at the colder ones. Although the circularity of the ice particles and their projected areas, and therefore their relative contribution to σ , increases with temperature, PSD broadening more than offsets these effects. The temperature-dependent fit and the effective diameter approach both perform much better than the other fits at temperatures below -60°C . The performance of these two methods is nearly identical between -20° and -60°C , but the effective diameter approach performs slightly better below -60°C . The analytical representations in the online supplemental material and in H13

FIG. A1. Numerical simulation of the scattering effect, the enhancement in the forward-scattering peak, for ice particle populations as a function of the ice particle effective radius $[(A1)]$. (left) The scattering effect as a function of the effective radius shown for different instrument fields of view. (right) The scattering effect as a function of the field of view of the detector shown for different effective radii. For reference, the DLH instrument has a field of view of 3 mrad (dotted vertical line in the right panel). For this field of view, the dotted vertical and horizontal lines in the left panel show the range of the scattering effect for a reasonable range of effective radii of 30 and 60 μm . (Calculations were prepared by Z. Zhang, NASA Goddard Space Flight Center.)

can also provide useful constraints on global modeling of ice cloud radiative properties.

The results presented here provide a method to characterize the ice water content from retrieved visible extinctions measured by CALIOP with increased accuracy at temperatures below about -50°C where *CloudSat* is usually unable to detect the (small) cloud particles. Joint CALIOP–*CloudSat* retrieval algorithms to derive these properties may also benefit from the results shown here.

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APPENDIX

Evaluation of Extinction Estimates from Particle Probes

We evaluate the accuracy and reliability of our extinction estimates calculated from (2) for temperatures warmer than -60°C where the PSDs are measured by the CIP, 2DC, and PIP imaging probes for particles $> 50\text{-}\mu\text{m}$

diameter and by the CAS and FSSP scattering probes for sizes $< 50\text{-}\mu\text{m}$. Cross-sectional areas of particles $> 100\text{-}\mu\text{m}$ are derived from the individual 2D particle images; particles $< 100\text{-}\mu\text{m}$ are assumed to be spheres. When available, we use data from the 2DS probe, detecting particles from 20 to $> 1280\text{-}\mu\text{m}$, with a resolution of $10\text{-}\mu\text{m}$.

Concentrations might be artificially enhanced when particles shatter on the probe's sensing area. Most shattered 2D-probe-detected particles are removed objectively based on particle interarrival times (Field et al. 2006). It is impossible to separate real particles from artifacts for the FSSP and CAS PSDs.

Independent measurements and derived values of σ were needed to test the accuracy of our methods for deriving extinction from the 2D probe data and our assumption that small particles do not contribute *significantly* to the total extinction for $T > -60^{\circ}\text{C}$. We developed a method to derive σ directly from an open-path, 25-m-pathlength diode laser hygrometer (DLH) probe (Podolske et al. 2003) on the NASA DC-8. Although DLH primarily measures water vapor concentration, one channel provides information on transmission through the air; this transmission is directly affected by the presence of hydrometeors.

Using Beer's law, DLH transmission values are converted to extinction in visible wavelengths, corrected for particle forward scattering. The forward-scattering effect increases with increasing ice particle effective radius:

a: 20 Aug., NAMMA, Extinction Comparison, $T > 0^\circ\text{C}$
[No 2D particles]

b: LWC Comparison, $T > 0^\circ\text{C}$

FIG. A2. From penetrations of small cumulus in the boundary layer during NAMMA: (a) comparison of σ derived from the CAS with DLH and (b) comparison of the LWC measured by the CVI probe with the CAS values, above the CVI's $\sim 5\text{--}7\text{-}\mu\text{m}$ cut cut size and for all sizes.

FIG. A3. Data derived during two aircraft descents through the melting layer during the NAMMA field program, showing (a),(c) comparisons of extinction derived from particle probes and DLH and (b),(d) the corresponding vertical velocities and temperature.

FIG. A4. Summary of comparisons of particle probe extinctions without and with the inclusion of data from the small-particle probes and DLH for temperatures above about -60°C .

$$\text{Re} = (3/2\text{IWC})/0.91\sigma. \quad (\text{A1})$$

This occurs because the fraction of forward scattering increases with Re , with more scattered energy received by the detector (Fig. A1, left). We use calculations of the forward-scattered energy as a function of Re obtained from a new scattering library developed by Baum et al. (2011) and more recent unpublished results for severely roughened particles. Forward scattering is a function of Re and increases with the field of view of the detector (Fig. A1, right).

Extinction estimates from the particle probe data have been compared to the adjusted DLH estimates using the DLH probe's approximate 3-mrad field of view. This comparison dataset comes from two NASA DC-8 research flights during the NAMMA field program and three DC-8 flights during the TC4 campaign, where sampling was conducted both above and below 0°C . For TC4, data from the 2DS probe are also used for comparison.

When particles are small (cloud droplet sizes) and Re values are low, the DLH scattering effect is negligible (Fig. A1). Figure A2 compares σ derived from the CAS (sizing from about 2 to $30\text{ }\mu\text{m}$) with DLH and the CVI liquid water content (LWC) measurements in liquid clouds during NAMMA with mean droplet size $8.4\text{ }\mu\text{m}$ and when no particles $> 50\text{ }\mu\text{m}$ were sampled. The purpose of Fig. A2 is to show reasonable qualitative agreement between the values of σ in clouds with small

droplets and a small DLH scattering effect; the CAS probe is known to broaden the true spectra and is not used for the σ in the algorithm development, and the CVI has hysteresis that smooths out the condensed water content. There is excellent correspondence between σ from DLH and the PSDs as well as between the LWC derived from the CAS and that measured by the CVI (Figs. A2a and A2b, respectively). In the LWC comparison, we took both CAS droplets above the CVI probe's droplet "cut" size of between 5 and $7\text{ }\mu\text{m}$ and all particles for the same periods. Although the cut size clearly affects the LWC comparison (see CAS LWCs with and without the $<5\text{-}\mu\text{m}$ particles), Fig. A2 supports the view that the CAS was operating properly during NAMMA and that DLH σ values are reliable, at least when the scattering effect is negligible.

Figures A3a and A3c compare σ derived from the particle probes and DLH during descents from above, through, and continuing below the melting layer. In the ice region, the CIP appears to agree well with the DLH σ . The addition of σ from the CAS greatly overestimates the DLH σ . When about $+2^{\circ}\text{C}$ is reached (Figs. A3b,d)—and all of the ice is melted to droplets and drops with minimal shattering likely—the σ derived from the CIP underestimates the total extinction, whereas the addition of the CAS provides excellent comparison with the DLH. These comparisons suggest that almost all of the σ from the CAS when large ice particles are present is due to artifacts resulting from shattering and that the

FIG. A5. Extinction derived from DLH transmission corrected for the forward-scattering peak assuming heavily roughened particles in comparison with the values derived from the CIP + PIP (PP) and 2DS probes for periods at temperatures below 0°C from three flights during TC4.

imaging probe σ provides an excellent match to the DLH σ .

Data that are not shown demonstrate a similar comparison for TC4 during a climb through the melting layer into the ice region. At heights below the 0°C level, σ derived from the CAS + 2D probes are comparable to those from DLH (ratio ~ 1), but are much lower than those derived from the small-particle + 2D probes. This effect

changes at heights above the 0°C level, where shattered bits from the small-particle probes are likely to contaminate those PSDs.

A comparison of DLH and particle probe data from TC4 shows the results from an analysis of the CAS and SPEC, Inc., 2DS probe data. The CAS probe operated poorly in TC4, requiring repeated servicing by the manufacturer, and so should only be used in a qualitative

sense. The 2DS PSD and σ values are derived by SPEC. During a climb through cloud, σ values from the 2DS probe and DLH compare well in the liquid region and through to about the -15°C level, then the 2DS σ values are either below or above those from DLH in certain temperature intervals (Fig. A4). Because the 2DS does not detect particles below about $20\text{ }\mu\text{m}$, there must be little contribution from cloud droplets $< 20\text{ }\mu\text{m}$. The σ values from the CIP + PIP (used for the PSD $> 1000\text{ }\mu\text{m}$) agree reasonably well with DLH, suggesting (and in view of the NAMMA comparisons with the same probes) there is relatively little contribution by particles $< 50\text{ }\mu\text{m}$. For much of the climb, the PSD σ s are close to those from DLH, but the comparison in Fig. A4 suggests there are clearly pockets of small particles present.

Figure A4 summarizes the extinction comparisons between the particle probes and DLH (temperatures -60°C and above), with and without the inclusion of small particles. There is usually excellent agreement between DLH and the σ for these collocated datasets when only 2D probe data are used, especially where extinction values fall within a range where the CALIOP beam is not occulted within 1 km, $\sigma \sim 3 \times 10^{-3}\text{ m}^{-1}$ (shaded area). The σ values derived from the 2D and small-particle probes together significantly overestimate σ relative to DLH.

Figure A5 compares σ derived from the 2DS and 2D (CIP + PIP) probes (particle probes, labeled PP) to that from the DLH probe for the three TC4 flights in ice regions. The fitted curves indicate that PP and DLH σ s are almost the same, on average. The 2DS σ should be better than those for PP for the lower σ s because of its better resolution and lower size detection threshold; the PP should be better at estimating σ at the larger values due to its larger sample volume for large particles and the ability to measure larger particles. At temperatures $< -20^{\circ}\text{C}$, the PP and DLH σ 's compare favorably. The 2DS probe σ s are $\sim 50\%$ higher than the PP and DLH σ . Because there were no special front tips on the 2DS for TC4 (but now are), we suggest that occasional small particles produced by shattering that entered the field of view of the probe were not identified or removed using interarrival times but contribute significantly to σ for reasons related to depth-of-field considerations.

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