On the angular effect of residual clouds and aerosols in clear-sky infrared window radiance observations
2. Satellite experimental analyses
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Received 7 June 2012; revised 6 December 2012; accepted 7 December 2012.

[1] This paper continues an investigation into the zenith angular effect of cloud-contamination within “clear-sky” infrared (IR) radiance observations commonly used in the retrieval of environmental data records (EDRs), which include “cloud-cleared radiances” (as is typical from hyper/ultra spectral IR sounders), as well as “cloud-masked” data (as is typical from imagers). The simple probability of clear line of sight (PCLoS) models and sensitivity studies of Part 1 (Nalli et al., 2012a) are corroborated with experimental analyses of environmental satellite data products as functions of sensor zenith angle, including sounder cloud-cleared radiances (CCRs) and retrieved effective cloud fraction, as well as narrowband imager cloud masking. Analyses of hyperspectral microwindow calc − obs are performed using MetOp-A Infrared Atmospheric Sounding Interferometer (IASI) CCR observations matched to dedicated radiosonde observations (RAOBs) during intensive validation field campaigns. The IASI calc − obs are found to exhibit a systematic positive bias with a strong concave-up variation with satellite zenith angle (i.e., an increasing positive bias symmetric over the scanning range) on the order of 1–3 K in magnitude, a signal associated with both residual clouds and dust aerosols. This is corroborated by analysis of the IASI retrieved effective cloud fraction product compared to the expected angular variations predicted by the PCLoS models, which show that the observed concave-up calc − obs variation may be the result of contamination by mid-to-upper tropospheric clouds. Finally, a corollary global analysis of the MetOp-A Advanced Very High Resolution Radiometer (AVHRR) cloud-mask shows concave-up variation that may be underestimating the angular variation for global ensembles containing clouds with vertical development (i.e., aspect ratios >0.5). The results presented in this work thus support the sensitivity studies of Part 1, indicating that contamination by residual clouds and/or aerosols within clear-sky observations can have a measurable concave-up impact on the angular agreement of observations with calculations.


1. Introduction
[2] As discussed previously in Nalli et al. [2012a], accurate satellite observations (obs) and calculations (calc) of clear-sky, top-of-atmosphere (TOA) spectral radiances are necessary for retrieval of environmental data records (EDRs) from satellite infrared (IR) sounder and imager remote sensing systems. Because IR-based EDR physical retrieval algorithms operate on the premise of minimizing clear-sky calc minus obs (calc − obs, or equivalently, obs − calc), it is important that differences between obs and calc be minimal under well-characterized conditions over the range of sensor scan angles. A systematic angular dependence in obs − calc may lead to undesirable scan-dependent artifacts and/or
errors in the calibration/validation (cal/val) of sensor data record (SDR, also referred to as Level 1B radiances) and EDR products (also referred to as Level 2 retrievals). EDRs from hyper/ultraspectral sounding systems [e.g., Smith et al., 2009] include atmospheric vertical temperature and moisture profiles (AVTP and AVMP, respectively), as derived from the new Joint Polar Satellite System (JPSS) Cross-track Infrared Microwave Sounding Suite (CrI MSS), as well as the MetOp-A Infrared Atmospheric Sounding Interferometer (IASI) [Cayla, 1993; Hilton et al., 2012] and Aqua Atmospheric Infrared Sounder (AIRS) [Chahine et al., 2006]. Imager EDR products include sea surface temperature (SST) from the JPSS Visible Infrared Imaging Radiometer Suite (VIIRS), and have been a key product from the legacy Advanced Very High Resolution Radiometer (AVHRR/3) flown on U.S. National Oceanic and Atmospheric Administration (NOAA) satellites [e.g., Nalli, 2004] as well as MetOp-A. Cal/val of the CrI MSS sounder and VIIRS imager systems onboard the Suomi National Polar-orbiting Partnership (NPP) satellite (launched in October 2011) have been a priority for ensuring products comply with mission requirements and have met global performance specifications [e.g., Nalli et al., 2012b].

In this set of companion papers, rather than attribute discrepancies in clear-sky calculation—obs solely to calc, we seek to examine how the obs themselves may contribute to such discrepancies as a function of zenith angle. It was emphasized in Part 1 [Nalli et al., 2012a] that global clear-sky observations are usually obtained from either cloud-clearing (in the case of sounders) or cloud-masking (in the case of imagers), and therefore themselves constitute “products.” Furthermore, cloud-clearing and cloud-masking algorithms are not generally designed to mask or correct for aerosols. Because of this, clear-sky observations will be subject to algorithmic errors beyond that of the sensor, a problem commonly referred to as cloud and/or aerosol contamination. Generally speaking, a small degree of residual clouds and aerosols usually remain in clear-sky window radiances, and these will lead to an obs that is cold-biased relative to the clear-sky calculation [e.g., Benner and Curry, 1998; Nalli and Stowe, 2002; Sokolik, 2002; Maddy et al., 2011, 2012]. Hypothetical models were derived in Nalli et al. [2012a] to estimate the angular sensitivity of residual clouds and/or aerosols on clear-sky IR window channel radiance observations. It was assumed that the probability for cloud-contamination behaves as a probability of a clear line-of-sight (PCLoS) for a very small absolute cloud fraction. Likewise, aerosol contamination was assumed to behave according to the increased slant-path arising from a small aerosol optical depth (AOD). Both these effects (PCLoS and mean slant-path) can be accounted for semitransparent clouds: Contamination by semitransparent clouds with vertical aspect ratios \( n < 1 \) (flatter clouds) will arise more from increased mean slant-path and less from decreased PCLoS, and vice versa for clouds with aspect ratios \( n > 1 \) (taller clouds). In all cases, it was found that very small levels of contamination can lead to measurable angular effects on the order of tenths of a Kelvin or more [cf. Nalli et al., 2012a].

In this second part, experimental analyses of environmental satellite data products as functions of sensor zenith angle are performed, including hyperspectral IR sounder cloud-cleared radiances (CCRs) and retrieved effective cloud fraction, as well as narrowband imager cloud masked observations. Section 2 overviews the radiative transfer model (RTM) used for high spectral resolution radiance calculations (calc) in this work. An overview of hyperspectral sounder cloud-clearing methodology (necessary for understanding the CCR product) is then given in Section 3, followed by analyses of microwindow calc—obs using MetOp-A IASI CCR observations matched to dedicated Vaisala RS92-SGP radiosonde observations (RAOBs) during NOAA Aerosols and Ocean Science Expeditions (AEROSE) [Morris et al., 2006; Nalli et al., 2011] in Section 4. It should be recognized that Vaisala RS92 rawinsondes are considered reference-quality measurements [e.g., Seidel et al., 2009], and dedicated sondes (i.e., sondes dedicated exclusively to satellite calc/val) thus provide the best possible temperature and water vapor profile measurements necessary for accurate state-of-the-art forward model calculations. Although sounder data is typically at coarser spatial resolution than imager data (e.g., AVHRR), high spectral resolution allows for surgical selection of channels to minimize the impacts of absorbing gas uncertainties in calc [e.g., Nalli and Smith, 2003] and thus place the tightest possible control on these variables. Similarly, ocean-based campaigns provide the tightest control on the surface variables [e.g., Nalli et al., 2006]. Section 5 then presents analyses of sounder and imager cloud products (effective cloud fraction and cloud mask, respectively) and compares them to the expected angular variations predicted by the PCLoS models discussed and developed in Part 1 [Nalli et al., 2012a]. In agreement with the sensitivity calculations in Part 1 [Nalli et al., 2012a], it is found that the likelihood of contamination by residual clouds and/or aerosols within clear-sky observations can have a measurable concave-up impact on the angular agreement with calculations, typically on the order hundreds of mK, but also as much as \( \pm 1 \) to \( \pm 3 \) K in sounder cloud-cleared radiances.

2. Radiative Transfer Calculation

For a window channel, the calculation has two primary components: The surface-leaving radiance (SLR) model, consisting of sea surface emission (emissivity and skin temperature) and quasi-specular reflection of downwelling atmospheric emission, along with the atmospheric transmittance and emission. Assuming a plane-parallel, non-scattering, clear-sky atmosphere (i.e., cloud and aerosol free) with azimuthal symmetry, the IR radiative transfer equation (RTE) for a down-looking sensor at zenith angle \( \theta \) (which is to be distinguished from nadir view angle, \( \vartheta \)) and quasi-monochromatic wave number \( v \) is given by

\[
R_v(\theta) = I_v(\theta)T_v(\theta) + I_v^d(\theta),
\]

where \( R_v(\theta) \) is the radiance measured by the sensor in mW/m^2sr^-1cm for quasi-monochromatic wavenumber \( v \) at the local zenith emission angle for the sensor field-of-view (FOV), \( \theta, I_v(\theta) \) is the spectral SLR (defined below), \( T_v(\theta) \) is the surface-to-observer path transmittance, and \( I_v^d(\theta) \) is the upwelling atmospheric-emitted radiance (we use the symbol for intensity, \( I_v \), in this paper to indicate modeled or derived component radiances).
[6] For a downlooking sensor located just above the surface operating in an atmospheric spectral window band, SLR is usually calculated as

$$I_{sr} (\theta) \approx e_r (\theta, \bar{u}) B_{sr} (T_s) + [1 - e_r (\theta, \bar{u})] I_{sa} (\theta),$$  

(2)

where $e_r (\theta, \bar{u})$ is the FOV mean emissivity for mean surface wind speed, $\bar{u}$, $B_{sr}$ is the Planck blackbody function, $T_s$ is the surface skin temperature, and $I_{sa} (\theta)$ is the downwelling atmospheric radiance. Note that equation (2) neglects solar contributions, and so strictly applies only under nighttime conditions, but is also employed for LWIR spectral windows where such contributions are small. We make of use two different emissivity models for computing $e_r (\theta, \bar{u})$, a conventional-type model [Masuda, 2006], and an “effective emissivity” type model [Nalli et al., 2008a, 2008b].

[7] To ensure accurate high spectral resolution forward radiance calculations in equation (1) (i.e., calc), the atmospheric transmittance, $T_{at}$, and radiance terms ($I_{sa}, I_{tr}$) are calculated using the Atmospheric and Environmental Research, Inc. (AER) rigorous Line-By-Line Radiative Transfer Model (LBLRTM), Version 11.7 [Clough et al., 2005]. Given accurate atmospheric input parameters (viz., profiles of temperature, water vapor and secondary absorbing gases), the variation of clear-sky slant-path transmittances based upon model physics is generally highly accurate, especially quasi-monochromatic transmittance, which has exact expression. The LBLRTM calculations performed in this work take into account absorbing species H₂O, CO₂, O₃, N₂O, CH₄, CFC-11, CFC-12 and CCl₄.

2.1. Microwindow Channel Selection

[8] To minimize uncertainties arising from gas absorption deviating from atmospheric state parameter inputs (including errors in RAOB-measured H₂O, as well as assumed values for fixed gases), we carefully selected 3 spectral microwindows (i.e., hyperspectral regions of high transmittance located between absorption lines) minimally impacted by absorbing species in the longwave IR (LWIR) region [cf. Nalli and Smith, 2003, Figure 1]. Although there are more transparent microwindows in the shortwave IR (SWIR) region (largely due to minimized H₂O continuum absorption), the use of daytime data in this paper for examining cloud and aerosol impacts precluded any possible exploitation of such channels.

[9] The peak of the LWIR H₂O continuum transmittance is located on the SWIR-side of the O₂ band (i.e., ~1080–1200 cm⁻¹). However, for a sensor onboard a satellite platform (i.e., located above the ozone layer), ozone absorption becomes significant even in the extreme wings; thus it is important to avoid the O₂ band as much as possible. Unfortunately, this coupled with peak dust aerosol absorption [cf., Nalli et al., 2006, Figures 9–10] reduces the usefulness of any microwindows in this spectral region. Thus we found the following LWIR microwindow channels, located just outside of this region, to be optimal for minimizing gas uncertainty for a sensor located at the TOA: $v_1 = [956.5, 958.5]$ cm⁻¹, $v_2 = [962.5, 964.5]$ cm⁻¹ and $v_3 = [1202.0, 1204.5]$ cm⁻¹. To provide an estimate of the sensitivity of these channels, Figure 1 shows the computed sensitivity in TOAcalc vs. obs to nominal systematic errors (ranging from −5 to +5%) in various input state parameters (H₂O, CO₂ and O₃, respectively), based upon LBLRTM calculations using equations (1) and (2). To provide a point of reference, calculations for the AVHRR split-window IR channels 4 and 5 (11 and 12 μm) are also shown. The significantly reduced sensitivity of hyperspectral microwindows to water vapor uncertainty is clearly evident.

3. Overview of IR Sounder Cloud-Clearing

[10] To facilitate a proper interpretation of the sounder experimental results to follow, a thorough review and expository of the cloud-clearing and cloud-parameter retrieval algorithms employed by hyperspectral sounders is warranted. The cloud-clearing methodologies for IASI and CrIMSS [e.g., the NOAA-Unique CrIS/ATMS Processing System, NUCAPS; Gambacorta et al., 2012], are essentially the same as that used for AIRS [Susskind et al., 2003], whereby a clear-column IR radiance spectrum is derived for a field-of-regard (FOR) consisting of a “golf-ball” of $K=2 \times 2$ (for IASI), or $K=3 \times 3$ (for CrIS/ATMS and AIRS), IR FOVs overlapping 1 microwave (MW) FOV, with a nadir spatial resolution of ~50 km. The cloud-clearing method basically assumes that differences among the IR FOV are solely due to clouds, then proceeds to extrapolate a clear-column spectrum (the cloud-clear ed radiance), $R_c$, from a linear combination of collocated FOVs, that is [Susskind et al., 2003]

$$\hat{R}_c (\theta_0) = \hat{R}_s (\theta_0) + \sum_{k=1}^{K} \eta_k (\theta_0) [\hat{R}_t (\theta_0) - R_{r,k} (\theta_0)]$$  

(3)

where $\theta_0$ is the central zenith angle of the FOR, $K$ is the number of FOVs (in case of IASI, $K=4$), $R_s$ is the mean radiance and $\eta_k$ is the cloud-clearing parameter (defined more below). For more specifics on cloud-clearing, the reader is referred to Susskind et al. [2003] or Maddy et al. [2011].

3.1. Basis and Development

[11] While the full IASI/AIRS and CrIMSS cloud-clearing methodology as summarized above has been developed and refined over a number of decades [e.g., Smith, 1968; Chahine, 1974, 1977; Susskind et al., 2003], fundamentally the approach is based upon the following physical reasoning. For simplicity of argument, we momentarily consider the approach using only $K=2$ adjacent FOVs and one cloud “formation” (i.e., a single cloud layer). The basic assumption in cloud-clearing is that the sole contribution to a change in radiance between adjacent FOR is the cloud fraction; surface and atmospheric state, hence the clear column radiance, $I_c$, are considered not to vary. For FOV $k \in \{1, 2\}$, equations for $R_{r,k}$ are given by

$$R_{r,1} (\theta) = [1 - N_1 (\theta_0)] I_1 (\theta_0) + N_1 (\theta_0) I_{c1} (\theta_0),$$  

(4)

and

$$R_{r,2} (\theta) = [1 - N_2 (\theta_0)] I_1 (\theta_0) + N_2 (\theta_0) I_{c2} (\theta_0).$$  

(5)

where $N_1 (\theta_0)$ and $N_2 (\theta_0)$ are the clear and cloudy radiances adjusted to the FOR central zenith angle (for an in-depth description of the local angle adjustment methodology, the reader is referred to the IASI Algorithm Theoretical Basis Document (ATBD), http://www.star.nesdis.noaa.gov/smed/spb/iosspt/qadocs/ATBD/L2-ATBD-8-30-06.pdf)
respectively. We note here that equations (4) and (5) are analogous to model equation (47) derived in Part 1 [Nalli et al., 2012a], with the effective cloud fraction equivalent to

\[ N(\theta) \equiv e_{\text{cc}}(\theta)[1 - P(\theta, x, \ldots)], \]  

where \( P(\theta, x, \ldots) \) is the probability of a clear line of sight (PCLoS) for a cloud layer with clouds having vertical aspect ratios \( x \) [Kauth and Penquie, 1967; Taylor and Ellington, 2008; Nalli et al., 2012a] and \( e_{\text{cc}}(\theta) \) is the cloud effective emissivity. Energy conservation for radiance incident upon the cloud requires that \( r_{\text{cc}}(\theta) + a_{\text{cc}}(\theta) + T_{\text{cc}}(\theta) = 1 \), where \( r_{\text{cc}}, a_{\text{cc}}, \) and \( T_{\text{cc}} \) are the cloud reflectance, absorptance and transmittance respectively. From Kirchoff’s law for radiance, \( e_{\text{cc}}(\theta) = e_{\text{cc}}(\theta) \), and assuming negligible scattering in the IR, \( r_{\text{cc}}(\theta) = 0 \), which then allows defining the cloud effective emissivity solely in terms of the cloud transmittance (which depends on cloud optical depth, \( \tau_{\text{cc}} \))

\[ e_{\text{cc}}(\theta, \tau_{\text{cc}}) \equiv 1 - T_{\text{cc}}(\theta, \tau_{\text{cc}}). \]

For more details on modeling equations (6) and (7), the reader is referred to the Part 1 paper [Nalli et al., 2012a]. Assuming \( e_{\text{cc}} \) does not vary appreciably with \( \nu \), equations (4) and (5) may be combined to eliminate \( r_{\text{cc}}(\theta) \) and solve for \( I_s(\theta_0) \). Adding the identity term \( R_{v_1} - R_{v_2}[(N_2 - N_1)/(N_2 - N_1)] \) and rearranging, it can be shown that the clear-column radiance can be derived as a linear extrapolation of the radiances from the two cloudy FOVs, that is

\[ I_s(\theta_0) = R_{v_1}I_s(\theta_1) + \eta(\theta_0)[R_{v_1}I_s(\theta_1) - R_{v_2}I_s(\theta_2)], \]  

where the channel-independent cloud-clearing parameter \( \eta \) is given by

\[ \eta(\theta_0) \equiv \frac{N_1(\theta_0)}{N_2(\theta_0) - N_1(\theta_0)}. \]

Given a forward calculation of the clear-column radiance in a single channel, \( I_s(\theta_0) \), an equation can be obtained based upon (8) for calculating \( \eta \) as

\[ \eta(\theta_0) = \frac{T_s(\theta_0) - R_{v_1}(\theta_0)}{R_{v_1}(\theta_0) - R_{v_2}(\theta_0)}. \]
3.2. Retrieval of Cloud Parameters

[13] The cloud parameters (viz., the effective cloud fraction and cloud top pressure) are retrieved with each CCR spectrum (for IASI, AIRS and NUCAPS) by means of a three-step retrieval algorithm involving an iteration of the retrieved FOR clear-column state vector, \(X\) (temperature and gas profiles): (1) \(X_{01}\) is obtained from the observed radiances via statistical regression and is used in the first cloud-parameter and cloud-clearing retrieval, (2) the CCR from step one are used to perform a second regression for obtaining \(X_{02}\), which is again used to retrieve cloud parameters, and (3) \(X_{02}\) is used as the first guess for the physical retrieval (a least square minimization between the derived cloud-cleared and computed clear-column radiances) of \(X_{11}\), which is used for a second cloud-parameter and cloud-clearing and a final physical retrieval, \(X_{12}\).

[14] The first and second steps retrieve cloud parameters on a single FOR. In step one, the cloud parameter first guess assumes two formations whose cloud top pressures are set to 350 hPa and 850 hPa, with corresponding cloud fractions of 0.1667 and 0.3333, respectively. The outputs from the first step are used as first guess for the second cloud parameter retrieval step. The outputs from the second step (FOR outputs) are uniformly set as first guesses (one for each respective layer) for all \(K\) fields of view. The third and final cloud retrieval step performs a cloud parameter retrieval in each FOR, up to 2 cloud pressures and 2K cloud fractions.

[15] The cloud parameter retrieval algorithm is based upon a multi-FOV equation analogous to the two-FOV equations (4) and (5) above, which in turn are analogous to the PClOoS model, assuming opaque clouds. Using the \(K\) calculated and observed radiances, \(R_{c, \lambda}(\theta_0)\) and \(R_{d, \lambda}(\theta_0)\), respectively, the \(j = 1, 2, \ldots, n_c \leq (K \cdot L + L)\) cloud parameters \(C_j\), consisting of \(K \times L\) effective cloud fractions, \(N_{A, j}\), plus \(L\) cloud pressures, \(p_l\), can be solved for. In practice this is done for two cloud formations (\(L = 2\)), and clouds are assumed to have unit emissivity (i.e., are considered opaque); thus \(n_c = 2K + 2\) cloud parameters are solved for. The solution is obtained via regularized least squares minimization using

\[
R_{c, \lambda}(\theta_0) - R_{d, \lambda}(\theta_0) = \frac{\partial R_{c, \lambda}(\theta_0, X)}{\partial C_j} \delta C_j = S_{\lambda, k, j} \delta C_j, (11)
\]

where \(\delta C_j\) is a finite perturbation to the cloud parameter being solved for and \(S_{\lambda, k, j} \equiv \partial R_{c, \lambda}(\theta_0, X) / \partial C_j\) is the sensitivity, which is estimated via finite differencing as

\[
S_{\lambda, k, j} = \frac{1}{\delta C_j} [R_{c, \lambda}(\theta_0, X + \delta C_j) - R_{c, \lambda}(\theta_0, X)]. (12)
\]

Note that for the retrieval of cloud parameters, only a LWIR subset of the set of channels used for the CCR step are employed.

[16] From the above it can be seen that the retrieved \(C_j\) are critical, “behind-the-scenes” components of the sounder’s final retrieved EDR products (i.e., temperature and water vapor profiles). The retrieved \(C_j\) include effective cloud fractions, \(N_{A, j}\), which have a direct physical correspondence with the PClOoS-based semitransparent cloud model, equations (6) and (7), described in detail in Part 1 [cf. Nalli et al., 2012a, §6]. Although they are determined through separate procedures, the retrieved cloud parameters and CCR are interrelated. The retrieval of \(C_j\) influences the computation of CCR by discarding from the original \(n_c\) channel set those channels where transmittance at the retrieved cloud top pressure exceeds a given threshold. On the other hand, the derived CCR provides the clear-column state vector \(X\) used in the forward calculations necessary for deriving \(C_j\). Errors in the cloud clearing procedure thus propagate through the atmospheric state \(X\) resulting in attendant errors in the cloud parameters. In situations where cloud clearing is deficient, a cold bias results in the CCR product; this cold bias will in turn propagate through the retrieved atmospheric state \(X\), resulting in colder temperature retrievals, thereby lowering the retrieved cloud fractions.

4. IR Sounder calc — obs Analyses

[17] In this section, we consider operational satellite IR sounder data, namely NOAA-unique IASI Level 2 cloud-cleared radiance (CCR) granules produced by the NESDIS/STAR IASI Operational Product Processing System [e.g., Maddy et al., 2009]. Complete details of the IASI/AIRS processing system can be found in the Algorithm Theoretical Basis Document (ATBD) available at http://www.star.nesdis.noaa.gov/smcd/spb/iossppt/qadocs/IASI_Phase2/. Note that the NUCAPS system implemented at NESDIS/STAR is essentially the same processing system, but adapted to the JPSS CrIS/ATMS sensors currently being flown on Suomi NPP [Gambacorta et al., 2012].

4.1. Experimental Overview

[18] We conduct experimental analyses based upon marine correlative profile data obtained from NOAA Aerosols and Ocean Science Expeditions (AEROSE), a series of trans-Atlantic field campaigns on the NOAA Ship Ronald H. Brown [Morris et al., 2006; Nalli et al., 2011]. In the forthcoming analyses, IASI CCR granules have been matched with dedicated RAOBs launched over the tropical Atlantic Ocean during AEROSE campaigns, these providing a unique set of in situ correlative data (needed as state parameters for forward calculations) within outflows of Saharan dust and sub-Saharan smoke aerosols [Nalli et al., 2006, 2011].
Saharan dust aerosols are known to contaminate the CCR products of both IASI and AIRS [Maddy et al., 2012]. For additional specifics on the AEROSE campaigns, including RAOB locations and times (up through the 2010 campaign), the reader is referred to Nalli et al. [2011].

[19] Shown in Figure 2 are multi-campaign measurements of solar-spectrum AOD and derived Angstrom exponents, \( \tau_a \), obtained from Microtops handheld sunphotometers and processed using the standardized Maritime Aerosol Network (MAN) methodology [Smirnov et al., 2011]. These aerosol distributions are indicative of persistent elevated levels of aerosols over the tropical Atlantic due to outflows from the African continent. Bimodal distributions are evident in both the AOD and derived \( \tau_a \). The AOD modes of \( \geq 0.15 \) and 0.35 roughly correspond to marine-background and elevated dust/smoke outflows, respectively, whereas the \( \tau_a \) modes of 0.2 and 1.1 correspond to dust/marine-background and smoke, respectively. Note that a major dust outflow pulse was encountered in the 2007 campaign [Nalli et al., 2011] and this is clearly evident in the top plot as a long, high-magnitude tail. Mean \( \tau_a(\lambda) \) for \( \lambda \approx 0.87 \) μm for the 2007–2011 campaigns are 0.54, 0.14, 0.24, 0.19, and 0.26 respectively.

[20] During AEROSE, dedicated Vaisala RS92 rawinsondes are launched approximately 30 minutes prior to MetOp-A and A-Train ascending and descending overpasses [Nalli et al., 2011]. The nearest IASI CCR FOR within 200 km from each matchup granule is matched to RAOB launch locations from the 2007–2011 campaigns. Note that the 200 km threshold is merely to eliminate launches that occurred between successive IASI swaths in the tropics — the vast majority of matchups are actually well within 100 km of each other [cf. Nalli et al., 2006, Figure 12]. The RAOB lowest level measurements are first used for the surface wind speed, \( \bar{u} \), (a parameter for surface emissivities), and for simplicity, we set the sea surface skin temperature, \( T_s \), to the lowest level RAOB air temperature adjusted by the median air-sea temperature difference measured by the Marine Atmospheric Emitted Radiance Interferometer (M-AERI) FTS system onboard (determined to be \( \approx 0.88, 0.45, 0.63, 0.39 \) K for the 2007, 2008, 2010, 2011 cruises, respectively; 2009 M-AERI QA data are not available, so for simplicity they are taken equal to the 2010 values given the proximity of cruise tracks). To manage LBLRTM computation time without impacting the accuracy of the calculated radiances, the high vertical resolution RAOB sampling is then linearly interpolated to the standard 101 AIRS/IASI RTM pressure levels [Strow et al., 2003]. For the upwelling transmittance and radiance calculation at the satellite observer level, LBLRTM takes Earth curvature into account given the sensor zenith view angle at the observer height, which is the supplement of the nadir view angle, \( \vartheta \) (i.e., the sensor scan angle), that is \( \pi - \vartheta \). However, the downwelling radiance calculation, \( I_{\text{down}} \) at the surface (the LBLRTM “observer” at the surface looking up) must be at the surface incidence angle, which is equal to the sensor zenith angle, \( \theta \), which we calculate using the law of sines (assuming a spherical Earth),

\[
\theta = \arcsin \left( \frac{R_e + z_0}{R_e \sin(|\vartheta|)} \right),
\]

Figure 2. Histograms of shipboard (top) AOD and (bottom) derived Angstrom exponent, \( \tau_a \equiv -\ln[\tau_a(\lambda_1)/\tau_a(\lambda_2)]/\ln(\lambda_1/\lambda_2) \), as measured from Microtops sunphotometers (processed using the standardized MAN methodology), during the AEROSE 2007–2011 campaigns. Bimodal distributions are evident in both plots, these corresponding to marine-background and elevated dust/smoke (top), and dust/marine-background and smoke (bottom). The long, high-magnitude tail in the top plot is the result of the major dust outflow pulse encountered during the 2007 campaign documented in Nalli et al. [2011].
failure of the cloud-clearing (something that may be expected probably due to overt cloud contamination in the CCRs and especially apparent in the robust concave-up variation of convective clouds). Disregarding the outliers, a very strong

4.2. Results and Discussion

and CFCs as provided by LBLRTM.

for O3, uniform mixed CO2 set to the 2008 levels (384.8 ppm), upon what may be considered an optimal state specified by gas absorbers. It is also notable that the three microwindows exhibit similar angular dependencies, suggesting adequate control on the gas absorption variables as intended (i.e., §2). Thus, the LBLRTM clear-sky slant-path transmittance and radiance line-by-line calculations should be considered reasonably accurate. Furthermore, differences between the results using the two SLR (surface emissivity) models is an order of magnitude smaller than the observed variation, and indeed, practically unnoticeable at this scale, so the choice of SLR model also cannot be considered a culprit.

where \( R_s \) is the mean Earth radius and \( z_0 \) is the sensor height (here being the height of MetOp-A satellite orbit). Fixed gas inputs to LBLRTM are the standard Tropical Model profile for O3, uniform mixed CO2 set to the 2008 levels (384.8 ppm), and CFCs as provided by LBLRTM.

Figure 3. AEROSE calc – obs (LBLRTM calculation using RAOB matched with the nearest accepted IASI FOR) as a function of angle for 3 different microwindow channels: 956.5–958.5 cm\(^{-1}\) (a, d), 962.5.5–964.5 cm\(^{-1}\) (b, e), and 1202.0–1204.5 cm\(^{-1}\) (c, f); results using the conventional and effective emissivity SLR models are denoted with blue \( \times \) and red + symbols, with robust fitting curves (quadratic polynomials derived from robust linear least-squares) for each shown as lines of the same color. The bottom plots (d–f) show results after subtracting empirically derived \( \delta T_{B,a}(\theta, \tau_a) \) [equation (19) in Nalli et al., 2012a], using daily mean Microtops \( \tau_a \) measurements (within 24 hours of, and interpolated in time to, the IASI observation).

where \( R_s \) is the mean Earth radius and \( z_0 \) is the sensor height (here being the height of MetOp-A satellite orbit). Fixed gas inputs to LBLRTM are the standard Tropical Model profile for O3, uniform mixed CO2 set to the 2008 levels (384.8 ppm), and CFCs as provided by LBLRTM.

4.2. Results and Discussion

Figure 3 shows the 2007–2011 AEROSE IASI-RAOB matchup calc – obs results in terms of brightness temperatures (simply taking the difference of the inverse Planck functions for calc and obs) for 3 microwindows defined in Section 2, namely \( v_1 = [956.5, 958.5] \text{ cm}^{-1} \), \( v_2 = [962.5, 964.5] \text{ cm}^{-1} \) and \( v_3 = [1202.0, 1204.5] \text{ cm}^{-1} \). The large positive outliers are probably due to overt cloud contamination in the CCRs and failure of the cloud-clearing (something that may be expected from situations known to pose difficulties such as dust and convective clouds). Disregarding the outliers, a very strong concave-up variation of calc – obs with zenith angle is indeed observed in the main cluster of data points. This is especially apparent in the robust fitting curves overlaid (quadratic polynomials from robust linear least-squares, \( R^2 = 0.68, 0.67, 0.73 \) for \( v_1, v_2, v_3 \), respectively). The magnitude of the concave-up variation simply cannot be attributed to a deficiency in the forward model (i.e., calc). Our calc is based upon what may be considered an optimal state specification, namely coincident Vaisala RS92 RAOB profile matchups, and we have optimally selected microwindow channels minimally impacted by gas absorbers. In addition, it is notable that the three microwindows exhibit similar angular dependencies, suggesting adequate control on the gas absorption variables as intended (i.e., §2). Thus, the LBLRTM clear-sky slant-path transmittance and radiance line-by-line calculations should be considered reasonably accurate. Furthermore, differences between the results using the two SLR (surface emissivity) models is an order of magnitude smaller than the observed variation, and indeed, practically unnoticeable at this scale, so the choice of SLR model also cannot be considered a culprit.

Figure 3 shows the 2007–2011 AEROSE IASI-RAOB matchup calc – obs results in terms of brightness temperatures (simply taking the difference of the inverse Planck functions for calc and obs) for 3 microwindows defined in Section 2, namely \( v_1 = [956.5, 958.5] \text{ cm}^{-1} \), \( v_2 = [962.5, 964.5] \text{ cm}^{-1} \) and \( v_3 = [1202.0, 1204.5] \text{ cm}^{-1} \). The large positive outliers are probably due to overt cloud contamination in the CCRs and failure of the cloud-clearing (something that may be expected from situations known to pose difficulties such as dust and convective clouds). Disregarding the outliers, a very strong concave-up variation of calc – obs with zenith angle is indeed observed in the main cluster of data points. This is especially apparent in the robust fitting curves overlaid (quadratic polynomials from robust linear least-squares, \( R^2 = 0.68, 0.67, 0.73 \) for \( v_1, v_2, v_3 \), respectively). The magnitude of the concave-up variation simply cannot be attributed to a deficiency in the forward model (i.e., calc). Our calc is based upon what may be considered an optimal state specification, namely coincident Vaisala RS92 RAOB profile matchups, and
cold-bias artifact in the split-window multichannel sea surface temperature (MCSST) water vapor correction. The daily mean \( \tau_a \) were interpolated in time to the IASI observation (if there was no \( \tau_a \) measurement with 24 hours of the IASI observation, then the matchup was discarded from the analysis). Using the measured \( \tau_a \), agreement between calc and obs is improved and there is a small reduction in concave-up variation. However, in spite of the improvement, bias and concave-up variation visibly remain, as especially evident in the fitting curves, \( R^2 = 0.67, 0.67, 0.63 \), which is indicative of residual cloud contamination in the CCR product. We also note that the \( R^2 \) value for \( v_3 \) was reduced by 0.1 after the aerosol correction (unlike the other two channels), which may be indicative of the reduced applicability of the aerosol correction outside the spectral range it was derived under (viz., the 800–1000 cm\(^{-1}\) IR window).

[23] Maddy et al. [2011], using collocated AVHRR cloud-masked data, demonstrated that a median \( T_a \) bias of \(-1.40\) K in the IASI SST product (IASI retrieval relative to the European Centre for Medium-Range Weather Forecasts (ECMWF) model) is attributable to the residual-cloud effect (a reduced value of \( T_a \) bias \(-1.10\) K has subsequently been obtained). Their results are based upon a much larger and more globally diverse sample, whereas our results are based upon a smaller, but more geographically concentrated and homogeneous sample with relatively high levels of aerosols. Figure 4 shows histograms of the AEROSE calc — obs before and after the ad hoc “aerosol correction,” and aside from the difference in sign, our distributions in calc — obs agree closely with the SST obs — calc found by Maddy et al. [2011] (cf. Figure 8, op cit.). The median differences for the 3 channels are 1.35 K, 1.34 K and 1.64 K, respectively; the aerosol correction brings these to 0.7 K, 0.7 K and 0.86 K. While not conclusive, this suggests dust aerosols may account for up to \( \approx 50\% \) of the observed systematic bias in the AEROSE domain.

The left-hand plot (a) shows the geographic distribution of all ocean FOV within IASI granules matched with AEROSE RAOBs shown in Figure 5. The granules matched with AEROSE RAOBs are 1.35 K, 1.34 K and 1.64 K, with a large data population corresponding to the RAOB matchup domain. The mean sunphotometer AOD for the AEROSE data is \( \tau_a \approx 0.29 \), so we might expect error bounds as given in Part 1 [Nalli et al., 2012a, cf. Figure 5], lower middle plot. Our results have greater bias relative to \(-1.10\) K before aerosol correction and less bias after. This result is reasonable when considering that the AEROSE domain may be characterized as having higher aerosols but lower cloud fractions [e.g., Jacobowitz et al., 2003, cf. Figure 2]. However, considering the common occurrence of Saharan air layers (SAL) within the domain [Nalli et al., 2005, 2011], the associated low-level inversions may act to diminish \( \delta T_{s+a} \) thereby also plausibly leading to an “overcorrection” [Nalli et al., 2012a, cf. Figure 5, lower middle].

5. Cloud Product Analyses

[24] In this section, we examine the satellite cloud products themselves. We first consider the sounder-retrieved effective cloud fraction product (discussed in Section 3) for a large data population corresponding to the RAOB matchup sample used for calc — obs analyses in Section 4. These results are then followed by a corollary analysis of operational IR imager cloud-masking using the MetOp-A AVHRR in Section 5.2.

5.1. Sounder Retrieved Effective Cloud Fraction

[25] To examine whether concave-up angular variation in the calc — obs may be attributed to cloud contamination in the CCRs, we consider the full complement of the IASI granules matched with AEROSE RAOBs shown in Figure 5. The left-hand plot (a) shows the geographic distribution of all ocean FOV within IASI granules matched to RAOBs...
launched along the ship tracks shown in black; the right-hand plots (b–c) show the QA accepted and rejected FOR cases, respectively (the NOAA-unique IASI product accepts or rejects an entire retrieval profile according to the AIRS version 4 “mid-trop” quality flags, as defined in Tobin et al. [2006]). The distribution of accepted cases (b) may be thought of as the population ($\sum n = 525,843$) from which the FOR-RAOB matchup sample are taken (e.g., used for calc – obs in Figure 3). It can be seen in left plot (a) that the bulk of the observations occur between 0°–20°N and 15°–35°W, this being the result of PNE/AEROSE cruise tracks servicing PNE buoys along 20°N and 23°W [Nalli et al., 2011]. Note the reduction of accepted IASI cases within the vicinity of the ITCZ ($\approx 0$–10°N). Map projection is equal area.

Figure 5. Geographic histograms (100 × 100 km bins) of 2007–2011 AEROSE-domain IASI observations consisting of granules matched with AEROSE RAOBs: (a) all ocean granule matchup FOV ($\sum n = 718,171$), with the 2007–2011 AEROSE cruise tracks overlaid in black, (b) QA accepted cases, $\sum n = 525,843$ (total yield $\approx 73.2\%$), accepted cases being the population considered by analyses in this paper, and (c) QA rejected cases, $\sum n = 192,328$ (for clarity, the colorscale here is divided by two). The high density area between 0°–20°N and 15°–35°W (left plot) is the result of the PNE/AEROSE cruise tracks servicing PNE buoys along 20°N and 23°W [Nalli et al., 2011]. Note the reduction of accepted IASI cases within the vicinity of the ITCZ ($\approx 0$–10°N). Map projection is equal area.
The relatively small cloud fractions \( \sum N_{l} < 0.5 \) in the vicinity of the ITCZ \((\approx 5 - 10^\circ N)\) are partially a sampling artifact resulting from the fewer accepted cases in convective regions (cf. Figure 5). The bottom right (d) shows the normalized histogram of \( \sum N_{l} \) for accepted ocean cases, with a \( \beta \)-distribution fit (\( \beta \) parameters \( a = 0.02 \) and \( b = 0.2 \)) shown in red.

Figure 6. Geographic distribution (1° lat/lon binned means, accepted ocean cases) of IASI effective cloud fractions for formations (a) \( l = 1 \), (b) \( l = 2 \), and (c) total, for the AEROSE domain shown in Figure 5.

The relatively small cloud fractions \( \sum N_{l} \) in the vicinity of the ITCZ are partially the result of the sampling artifact associated with the fewer accepted cases in the region of deep convection.

[27] Figure 6d shows a normalized histogram of the total effective cloud fraction distribution. The observed distribution of IASI-retrieved \( \sum N_{l} \) is clearly non-Gaussian, exhibiting an unusual asymmetric U-shape. As was pointed out by Falls [1974], the distribution of cloud cover amounts may be properly characterized as a beta distribution. The beta distribution can assume a variety of patterns, including U-shapes and bell shapes, depending upon the two parameters \( (a \) and \( b \)) used to define it; a U-shaped pattern results if \( a < b < 1 \) [Falls, 1974]. Overlaid on the histogram in Figure 6d is a beta distribution fit obtained by setting the \( \beta \)-parameters \( a = 0.02 \) and \( b = 0.2 \). Physically speaking, the type of beta distribution pattern (i.e., U-shaped or bell-shaped) is a consequence of the size of the satellite FOV relative to the size of the clouds [Falls, 1974], which in turn depends on the cloud type and regime. A U-shape arises from an FOV size being of comparable magnitude (or smaller) than the average size of the clouds, as most FOVs will be either completely cloudy or clear, with few “partially cloudy” cases. In our case, however, we do not attribute the observed distribution solely to these considerations, as the IASI FOV are \( \approx 12 \) km at nadir. For example, while the AEROSE main observation area may in fact have reduced overall cloudiness [cf. Jacobowitz et al., 2003; Wylie et al., 2005], due in part to the persistence of SALs in the region, it is doubtful that there are as many completely cloud-free \( (\sum N_{l} = 0) \) IASI FOV as is shown. We also note that the observed distribution at large cloud amounts departs from the beta distribution, this being partially the result of deep-convection and precipitating cloud regions not passing QA.

[25] Figure 7 summarizes the IASI FOV cloud parameter retrievals, \( C_{l} \), for the AEROSE IASI granule population discussed above. The vertical distributions of retrieved cloud pressures, \( p_{l} \), for the \( l \in \{1, 2\} \) (upper and lower troposphere) formations are shown in Figure 7d. The combined cloud pressure distribution is clearly seen to be bimodal, with modes at the tropopause (200 hPa) and MBL (750 hPa). The angular dependence of the retrieved effective cloud fraction, \( \overline{N}_{l} \), is shown in Figure 7a–b for formations \( l = 1 \) and \( l = 2 \), respectively as \( 5^\circ \delta \theta \) binned means, with the combined result for both formations, \( \sum N_{l} \), given in Figure 7c. Variability about the means are designated with error bars encompassing ±1 mean absolute deviations. The upper formation retrievals (Figure 7a, \( l = 1 \)) are observed to exhibit a flat to slightly concave-down variation, whereas a slightly concave-up
Figure 7. IASI cloud parameter retrievals for the 2007–2011 AEROSE domain (cf. Figure 5): (a–c) binned mean (5° δθ bins) effective cloud fraction, Nl, as a function of θ for formations l = 1, 2, and combined, ∑Nl, respectively (variability about the means are shown with error bars designating ±1 mean absolute deviation); and (d) histogram of the cloud top pressures, p1, p2 for the L = 2 cloud formations. For comparison, shown in dashed colored lines in (a–c) are the angular variations of effective cloud fraction predicted by the generalized PCLoS model equation (6) for various combinations of cloud aspect ratios (x) and IR optical depths (τc). Annotations in (a–b) include the assumed cloud shape for the PCLoS cloud model calculation, Ts (i.e., the mean cloud top temperature for the layer, giving an indication of the relative impact on brightness temperature), and the formula used for estimating the absolute cloud fraction, Nl (required for the PCLoS calculation), which is taken to be the IASI effective cloud fraction near nadir, Nl(0), divided by assumed values of nadir cloud effective emissivity appropriate to the cloud formation, that is Nl ≡ 1 − P(0) = Nl(0)/εc,l(0).

Overlaid in Figure 7a–c (colored dashed lines) are the angular variation of clouds predicted by equations (6) and (7), for various combinations of cloud aspect ratios (x) and IR optical depths (τc). In these calculations, the absolute cloud fraction, Nl, is taken to be the IASI retrieved effective cloud fraction near nadir, Nl(0), divided by assumed values of nadir cloud effective emissivity, that is Nl ≡ 1 − P(0) = Nl(0)/εc,l(0). In Figure 7a–b it is found that the IASI Nl(0) ≈ 0.095, 0.098 for formations l = 1 and 2, respectively. Assumed values of τc,l are used for deriving the εc,l(0) using equation (7) and the methodology outlined in Part I [cf. Nalli et al., 2012a, §2.1 and 6]. We obtain τc,l estimates based upon the 22-year IR sounder observations of Wylie et al. [2005], who categorized clouds according to nlay = 3 altitude layers (“high”: p < 440 hPa; “middle”: 440 < p < 700 hPa; and “low”: p > 700 hPa), and 3 opacities (“thin”: τc < 0.7; “thick”: 0.7 < τc < 3.0; “opaque”: τc > 3.0 (Table 2b op. cit.). Generally speaking, Wylie et al. [2005] observed thin and thick clouds at mid-to-high altitudes, and opaque clouds at low-to-mid altitudes, as might be expected. Based on this, we simply assume τc,l = 0.7 for high clouds (l = 1), τc,l = 1.9 (i.e., the mean of 0.7 and 3.0) for the mid-level clouds (l = 2) and τc,l = ∞ for low clouds (l = 3), and then calculate εc,l(0) = 0.5, 0.85 and 1, respectively, from equation (7). We then obtain εc,l(0) and τc,l for the L = 2 cloud formations via simple finite differencing by first assuming 4 pressure boundaries for the nlay = 3 cloud altitudes, p = 50, 400, 700 and 1000 hPa, then

\[
ε_{c,l}(0) = \frac{ε_{c,l-Δl}(0)}{δ_c ln(p)} + \frac{ε_{c,l-l-1}(0)}{δ_c ln(p)} + δ_c ln(p).
\]

where δ_c ln(p) = ln(p_{l+1})/ln(p_l). Plugging the assumed values of εc,l(0) and cloud boundaries into (14), we arrive...
retrieved a small and opacity. Although the lower formation (a cloud shapes for formations assuming trapezoidal (anvil) and semiellipsoidal (cumulus) perform calculations allowing the assumed variation. tions (Figure 7a,c) do not capture the expected concave-up (depth 2.2 (Figure 7b, green dashed line), the upper formation modeled clouds with a smaller aspect ratio 0.5 and optical except the we only consider FOV with 0 or 1 retrieved cloud cave-up variation in Based on these inputs, it can be seen that a non-negligible con-

We take the thinnest and thickest clouds of each formation to be (increasing vertical to horizontal proportions) lead to increasing optical depths. We take the thinnest and thickest clouds of each formation to be \( \tau_{rc} \pm 0.4 \) corresponding to \( x = 0.25 \) and 1.25, respectively. Based on these inputs, it can be seen that a non-negligible concave-up variation in \( \mathcal{N}(\theta) \) occurs, even for clouds with very small \( x \), with the magnitude increasing with jointly increasing \( x \) and opacity. Although the lower formation (\( l = 2 \)) IASI retrieved \( \mathcal{N}_{l}(\theta) \) shows a concave-up variation consistent with modeled clouds with a smaller aspect ratio 0.5 and optical depth 2.2 (Figure 7b, green dashed line), the upper formation (\( l = 1 \)) retrievals, and consequently the combined cloud fractions (Figure 7a,c) do not capture the expected concave-up variation.

[30] Figure 8a–c shows the angular dependence of the IASI retrieved effective cloud fraction, \( \mathcal{N}_{l} \), similar to Figure 7a–b, except the we only consider FOV with 0 or 1 retrieved cloud formations (i.e., restricted to FOV designated as cloud-free, \( \sum_{l} N_{l} = 0 \), or those with only one retrieved cloud “layer”) confined to 3 layers defined by the 4 pressure boundaries assumed above, namely 50–400 hPa, 400–700 hPa, and 700–1000 hPa. This is to restrict ourselves to cases more directly comparable to the PCLoS model assumption of a single layer, and then examine how the retrievals compare with calculations for the three tropospheric layers [e.g., Wylie et al., 2005] considered above. As before, the dashed colored lines again show the modeled effective cloud fractions using equations (6) and (7) for combinations of \( \tau_{rc} \) and \( x \). Given assumed cloud layer parameters (cloud optical depths 0.7, 1.9 and \( \infty \), leading to effective emissivities 0.5, 0.85 and 1), absolute cloud fractions are derived as \( N_{l} = 0.13, 0.14 \) and 0.09, respectively. We assume cloud shapes for high and low level clouds as trapezoid (anvil) and semiellipsoid (cumulus), respectively [e.g., Nalli et al., 2012a]; because we do not have a good sense for what the shapes of midlevel broken clouds typically are, we simply assume ellipsoids as a “generic” cloud shape. The parameters \( \tau_{rc} \) and \( x \) are then varied jointly as in Figure 7. The bottom plots (d–f) show the corresponding angular differences between retrieval and model. Here it can be seen that

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**Figure 8.** Angular variation (\( 5^\circ \cdot 5^\circ \) binned means) of IASI retrieved effective cloud fraction \( \mathcal{N}_{l} \) for FOV constrained to 0 or 1 retrieved cloud formations. Layer averages are computed for (a) upper, 50–400 hPa (b) middle, 400–700 hPa and (c) lower, 700–1000 hPa layers. As in Figure 7, dashed colored lines show the angular variations of effective cloud fraction predicted by the generalized PCLoS model equation (6) for various combinations of cloud aspect ratios (\( x \)) and IR optical depths (\( \tau_{rc} \)), with the bottom plots (d–f) showing the corresponding angular differences between retrieval and model. Annotations in (d–f) are as in Figure 7, but also include \( n \), which is the number of FOVs in the sample.
for situations limited to a single retrieved cloud formation, the angular variation of the retrievals appear to be reasonable for a lower tropospheric layer (700–1000 hPa) of opaque clouds (and consistent with clouds having $\alpha \approx 1$), when there are no cloud layers above it. However, there are non-negligible discrepancies for situations involving mid-to-upper tropospheric cloud layers. For the topmost layer (50–400 hPa), the results are mixed, depending on the assumed opacity and aspect ratios. The retrieved cloud fraction angular variation is consistent with clouds having small aspect ratios and high transparency ($\alpha \lesssim 0.75$, $\tau_{\text{in}} \lesssim 0.7$), but not so for thicker, more vertically developed clouds. There is also a small degree of angular asymmetry, which may be associated with the space-time distribution of tropical anvil clouds [e.g., Gong and Wu, 2011]. For mid-tropospheric cloud layers (400–700 hPa), the retrievals do not show any concave-up tendencies, and in fact are slightly concave-down, thereby leading to notable discrepancies at larger angles. It may therefore be cloud contamination arising from difficulty with clouds in the mid-to-upper troposphere, these being located between the bimodal vertical distribution associated with $L=2$ allowable cloud formations (Figure 7d), that may be contributing to the concave-up calc — obs discussed in Section 4.2, at least for situations involving only single cloud layers.

5.2. Imager Cloud Mask

As a corollary to the sounder cloud analysis in Section 5.1, we here produce results obtained from the Clouds from AVHRR–Extended (CLAVR-x) cloud-mask product [e.g., Stowe et al., 1999] employed by the AVHRR/3 imager onboard the MetOp-A satellite. Complete details of CLAVR-x can be found in the CLAVR-x Cloud Mask Algorithm ATBD available at http://cimss.ssec.wisc.edu/clavr/clavrx_docs.html. Following the procedure described in Maddy et al. [2011], our analysis relies on semi-orbital MetOp-Â AVHRR data that have been matched with IASI granules over oceans for 8 March 2011. The results of the CLAVR-x cloud mask for one minus “confidently clear” (mask = 0) and one minus “confidently and probably clear” (mask = 0 and 1) for this day are given in Figures 9 and 10, respectively. In these figures, results are delineated into four regions; (a) AEROSE domain (tropical Atlantic Ocean), (b) tropical zone, (c) tropical-midlatitude zone, and finally (d) global. Calculations from the PCLoS model, assuming semi-ellipsoidal (hemispheroidal) clouds are also shown for various aspect ratios, taking the absolute cloud fraction, $N = 1 - P(0)$, to be 1 minus the smoothed CLAVR-x clear value at nadir ($\theta = 0$).

A number of interesting features are evident in these figures and worthy of comment. First off, we note that the CLAVR-x algorithm does indeed show a general increase in cloudiness (i.e., $1 - P$) with zenith angle. Cloudiness also increases, as does angular symmetry, as we advance from regional (a) to global zones (b–d); angular asymmetry...
is most apparent in the tropical domains (a–b). This makes sense in that as these samples (a–b) are more space-time localized, and thus less likely to the conform to the assumption of a Poisson distributed population of clouds (as assumed by the PCLoS model). In the case of the tropics, the asymmetry may also be an aliasing phenomenon associated with the space-time distribution of tropical anvil clouds [e.g., Gong and Wu, 2011] as mentioned above. This also may explain why the greatest angular variation in non-clear FOV is found in the AEROSE domain results (a), where the increase with angle is consistent with PCLoS clouds having greater aspect ratios (i.e., $\alpha \approx 0.75$–1.0). In the remaining zones, however, the angular increase is generally $\approx PCLoS \times 0.5$). Furthermore, the angular discrepancies between observation and model are seen to be greater in the “confidently and probably clear” (mask = 0 and 1, Figure 10). Use of mask = 0 and 1 implies a greater probability of residual cloud contamination in the cloud-masked product than the “confidently clear” product (mask = 0, Figure 9), and this contamination indeed leads to greater angular discrepancies from the PCLoS predicted values. These results support the hypothesis of residual cloud contamination conceivably becoming exacerbated at larger zenith angles in global clear-sky observations obtained from narrowband imagers (e.g., AVHRR).

6. Summary and Conclusions

[33] Physical retrieval algorithms for EDR products (including AVTP, AVMP, skin SST, etc.) typically operate by minimizing differences between clear-sky (i.e., cloud-cleared or cloud-masked) IR observations and forward model calculations (i.e., obs − calc). It is therefore extremely important that differences between obs and calc be minimal under well-characterized conditions. Comparative analyses of calc and obs are thus often employed to improve the forward model physics, or otherwise “tune” the model for a specific sensor. In performing analyses of this sort, however, there is an implicit assumption of accurate “clear-sky” observations. We have revisited this assumption recognizing that global clear-sky observations themselves are usually obtained from either cloud-masking (in the case of imagers) or cloud-clearing (in the case of sounders) algorithms, and therefore in actuality constitute “products.” Furthermore, cloud-masking and cloud-clearing algorithms are not generally designed to mask or correct for aerosols. Because of this, clear-sky observations (including sensor data records or SDRs) will be subject to algorithmic errors beyond those of the sensor itself, a problem known as residual cloud and/or aerosol contamination. Generally speaking, residual clouds and aerosols remaining in clear-sky window radiances will lead to an obs that is cold-biased [e.g., Benner and Curry, 1998; Nalli and Stowe, 2002; Sokolik, 2002; Maddy et al., 2011, 2012; Nalli et al., 2012a].

[34] We have carefully explored this issue under the working hypothesis that residual cloud/aerosol contamination may depend upon the zenith view angle of the sensor. In this companion paper, we have conducted experimental analyses to corroborate the simple physical models and sensitivity analyses of Part 1 [cf. Nalli et al., 2012a]. We began simply by performing calc − obs analyses using MetOp-A IASI sounder cloud-cleared radiances (CCRs) matched with dedicated Vaisala RS92 RAOBs obtained from multiple NOAA AEROSE campaigns conducted over the tropical Atlantic Ocean downwind of Saharan dust and biomass burning smoke aerosols [Morris et al., 2006; Nalli et al.,
Sensitivity analyses detailed in the Part 1 paper may be on the order of hundreds of mK as shown in the EDR retrievals are more likely to fail) have been identified. Spheric gases (Figure 1) are less than that predicted by evidence arising from background uncertainties in atmospheric gases. We found the AEROSE-IASI calculations to exhibit a strong concave-up variation with zenith angle on the order of 1 K or more in magnitude in all three microwindows. It is believed that residual clouds and dust aerosols were the primary culprits, these both being known sources of error in hyperspectral CCR observations [e.g., Maddy et al., 2011, 2012]. This was corroborated by our subsequent analysis of the IASI retrieved effective cloud fraction product (for a large population of AEROSE match-up granules), which we compared to the expected angular variations predicted by the PCLoS models discussed and developed in Part 1 [Nalli et al., 2012a]. It was found (for cases limited to one or zero cloud formations) that although the retrieved cloud fractions in the lower troposphere conformed well to the angular variation predicted by PCLoS, the results were mixed for the upper troposphere, depending on the assumed aspect ratio and transparency of the clouds. However, the retrievals for the mid-troposphere did not show any concave-up cloud tendencies as expected. Thus, the observed concave-up variation in calc – obs may primarily be the result of difficulty with mid-to-high level tropospheric clouds; mid-level clouds in particular are typically located between the pressure levels of the two retrieved cloud formation. Finally, we performed a similar analysis of the MetOp-A AVHRR CLAVR-x cloud-mask product for the AEROSE domain, as well expanded to the tropics, tropics–midlatitudes and global zones, and while the cloud-mask does show concave-up cloudiness with zenith angle, globally the variation is only consistent with PCLoS predictions assuming aspect ratios of 0.5 or less (i.e., flatter clouds); it is therefore conceivable that for global cloud ensembles with mean z > 0.5 the cloud mask may underestimate the angular variation in cloudy lines of sight. The IR sounder results presented in much of this paper focused on a region where dust is dominant; we suspect that clouds play the more dominant role at global scales.

6.1. Implications for EDR Retrievals and Future Work

In this paper, we showed that the angular dependence arising from background uncertainties in atmospheric gases (Figure 1) are less than that predicted by our PCLoS/aerosol models. Assuming near-unity transmittance in surface-sensitive microwindow channels (e.g., the AIRS 2616 cm−1 “superwindow”), the angular bias of surface skin/air temperature retrievals from contamination in IASI cloud-cleared radiances will be on the order of the calc – obs results shown in Figure 3, that is \( \approx \pm 1 \) to \( \pm 3 \) K for \( \theta \leq 65^\circ \), in agreement with the median bias of \( \approx 1.4 \) K found by Maddy et al. [2011]. Analogous bias from contamination in generic “clear-sky” observations may be on the order hundreds of mK as shown in the sensitivity analyses detailed in the Part 1 paper [Nalli et al., 2012a].

Because these specific situations (where IR Level 2 EDR retrievals are more likely to fail) have been identified, the findings can provide guidance toward improving future EDR products. One possibility would be adopting an angle-dependent QA on the retrievals. Specifically, because we may expect that \( P(\theta_0) < P(\theta_1) \) for zenith angles \( \theta_0 > \theta_1 \), tighter rejection thresholds could be adopted for cloud-cleared retrievals at larger \( \theta \). Another possibility would be to use the PCLoS modeling to generate a first guess for the retrievals. For instance, given an observation of nadir cloud fraction \( N \) from an imager, sounder, other correlative measurement (e.g., lidar) or ensemble of measurements, the first guess could be assigned as a function of angle based on the PCLoS model. Uncertainty in the guess could be estimated by the dispersion caused by perturbing the cloud aspect ratios.

[37] The results presented in this paper support our general hypothesis regarding the expected angular effect for ensembles of window channel radiances observations contaminated by residual cloudiness and/or aerosol, as is commonly the case for cloud-cleared or cloud-masked data. However, even if such algorithms are not used (e.g., observations taken from a field experiment during clear conditions), there still exists a possibility for cloud contamination (e.g., from sub-pixel clouds); this will be the subject of future research.

Acknowledgments. This research was supported by the NOAA Joint Polar Satellite System Office (NJO), NASA Research Announcement (NRA) NNH09ZDA001N, Research Opportunities in Space and Earth Science (ROSES-2009), Program Element A-41 (The Science of Terra and Aqua), the GOES-R Algorithm Working Group (AWG) (W. Wolf and T. Schmit, STAR AWG leads), and the STAR Satellite Meteorology and Climatology Division (SMCD) (M. D. Goldberg, SMCD Division Chief). AEROSE is supported by the NOAA Educational Partnership Program grant NA15AE1625, NOAA grant NA17AE1623 to establish the NOAA Center for Atmospheric Science (NCAS) at Howard University. We are grateful to the following individuals for their contributions in support of this work: D. Wolfe (NOAA/ESRL), supporting the Vaisala sounding system onboard the NOAA Ship Ronald H. Brown; A. Flores, M. Oyola (Howard University) and numerous other students, along with A. Smirnov (NASA/GSFC), for supporting AEROSE Microtops measurements; A. Heidinger and W. Straka (UW/CIMSS) for AVHRR CLAVR-x data; P. J. Minnett and M. Szczodrak for M-AERI support; S. DeSouza-Machado, for consultation on aerosol radiative properties; A. Ignatov and X. Liang (NEDIS/STAR), and CRTM investigators P. van Delst (NWS/NCEO), Y. Han (STAR) and Y. Chen (JCSDA), for meetings and discussions pertaining to MICROSC CRTM results (which led us to our original finding and application of the PCLoS model to the problem of cloud contamination), including their subsequent feedback on our initial results in presented at STAR in March 2011. We also thank the two anonymous reviewers who provided constructive feedback that we used to strengthen the paper. NOAA-unique IASI retrieval products from the NEDIS/STAR IASI Operational Product Processing System are available on the NOAA CLASS website: http://www.nesdis.noaa.gov/saa/products/search/datatype_family=IASI_OI_GRHST_products are available at http://powder.jpl.nasa.gov/.

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